

IN AN ALTERNATING MARINE AND NON-MARINE
DEPOSITIONAL SETTING, WHERE AND HOW ARE
EARLY CARBONIFEROUS TETRAPODS PRESERVED?

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In an alternating marine and non-marine depositional setting, where and how are early Carboniferous tetrapods preserved?

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During the Tournaisian tetrapods were adapting to terrestrial environments. Three sites exposing the Ballagan Formation in the Northumberland and Tweed basins were logged and sampled in detail to determine the sedimentological processes and depositional environments preserving tetrapod and associated vertebrate fossils.

The Coldstream site is interpreted as a marginal marine bay with fluvial input, whereas Coquetdale and Whitrope Burn are interpreted as marginal marine, lagoonal environments without direct fluvial input. Dolostone beds typical of the Ballagan Formation are present where there is a degree of restriction but absent in the more open marginal marine bay site. Packages of dolostones indicate marine transgressions: the repeated close vertical association of dolostones with siltstones with roots and desiccation cracks indicates a location close to base level subject to repeated changes in marine water depth.

Vertebrate and most invertebrate fossil material was transported in silty flows into marginal marine basins. Early pyritisation or dolomitisation has preserved uncrushed fossils. At Coldstream tetrapod and lungfish material was transported directly into the bay. At Coquetdale and Whitrope Burn marginal marine species have been transported into and around the more restricted, lagoonal sites. At Whitrope Burn chondrichthyan teeth, tetrapod fragments and lungfish toothplates were transported within the lagoon in a density flow.

Total organic carbon (TOC) values of generally <1% are typical of the Ballagan Formation. Bulk (-20 to -22‰) and specific (average 22.2‰) $\delta^{13}\text{C}$ org. carbon isotope values compare well with other Ballagan successions but are less negative than published values for the early Carboniferous. The similarity of the bulk and specific values indicate dominance by terrestrial material.

The preservation of tetrapod and other vertebrate fossils in the marginal shallow marine basins in this study contrasts with finds in the Tweed Basin, where most of the significant tetrapods have been preserved in alluvial overbank settings: floodplains or ephemeral floodplain lakes.

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Table of contents

1	Introduction	1
1.1	Background	1
1.2	Stratigraphic and geological context.....	2
1.3	The study sites.....	5
1.4	Thesis Structure.....	5
2	Methodology.....	7
2.1	Logging and sample selection	8
2.2	Ground rock surfaces	8
2.3	Thin sections.....	8
2.4	Scanning Electron Microscope work.....	9
2.5	Carbon isotope analysis	9
2.6	Facies and facies assemblages	11
3	Coldstream.....	12
3.1	Introduction	12
3.2	Coldstream facies assemblages	14
3.2.1	Bay-fill Facies Assemblage	14
3.2.2	Vegetated Floodplain Facies Assemblage.....	22
3.2.3	Shallow Marginal Marine Bay Facies Assemblage.....	26
3.2.4	Shallow Marine Near-Shore Facies Assemblage.....	32
3.3	Isotope analysis.....	34
3.4	Sedimentological development of the site at Coldstream	34
4	Coquetdale.....	37
4.1	Introduction	37
4.2	Coquetdale Facies Assemblages	39

4.2.1	Shallow Marginal Marine Bay Facies Assemblage	39
4.2.2	Bay-fill Facies Assemblage	54
4.2.3	Vegetated Floodplain Facies Assemblage.....	59
4.2.4	Shallow Marine Near-Shore Facies Assemblage.....	62
4.3	Isotope analysis	67
4.4	Sedimentological development of the site at Coquetdale	67
5	Whitrope Burn	71
5.1	Introduction	71
5.2	Lower, thin beds.....	73
5.3	Fossiliferous Bed	74
5.4	Bioturbated bed	75
5.5	Siltstones	77
5.6	Summary	78
6	Stable Carbon isotope values for Coldstream and Coquetdale.....	80
6.1	Introduction	80
6.2	Methodology.....	81
6.3	Results	81
6.4	Bulk carbon isotope data compared to data from Ballagan Formation mudstones	87
6.5	Discussion.....	89
6.6	Conclusions	93
7	Discussion and conclusions.....	95
7.1	Comparison of the depositional environments at the study sites.....	95
7.2	Palaeogeography.....	100
7.3	Carbon isotopes	103
7.4	Preservation of fossil material	104

7.5	Conclusions	107
Appendix A - Sites images.....		111
Coldstream.....		111
Coquetdale.....		112
Whitrope Burn		113
Appendix B - XRD results		114
Coldstream.....		114
Coquetdale.....		114
Whitrope Burn		117
Appendix C - Coldstream block for detailed study.		119
Appendix D - Carbon isotope values.....		126
Bibliography		128

List of Tables

Table 3.1: Details of fossil finds in beds from 33-40 cm on the logged section	17
Table 6.1: Bulk and specific carbon isotope values for Tournaisian sites in this study	82
Table 7.1: Facies and depositional processes found at Coldstream, Coquetdale and Whitrope Burn	97
Table 7.2: Facies assemblages found at Coldstream, Coquetdale and Whitrope Burn	98

List of Figures

Figure 1.1: Stratigraphic framework of the Scottish Borders and Northumberland Basin.....	3
Figure 1.2: Palaeogeography of southern Scotland and northern England in the Tournaisian (early Carboniferous).	4
Figure 2.1: Outline map of Scotland and Northern England with detail of site locations.	7
Figure 3.1: Log of Coldstream showing facies, facies assemblages and carbon isotope curves.....	13
Figure 3.2: Bay-fill facies assemblage.	16
Figure 3.3: Bay-fill facies assemblage, facies 5 and 6.	19
Figure 3.4: Vegetated floodplain facies assemblage.	24
Figure 3.5: Cross-stratified midstream bar deposits.	26
Figure 3.6: Shallow marginal marine bay assemblage.	28
Figure 3.7: SEM images of dolomite growth.	29
Figure 3.8: Shallow marine near-shore facies assemblage.	33
Figure 3.9: Development of the facies assemblages through time.....	35
Figure 4.1: Log of Coquetdale section showing facies, facies associations and carbon isotope curves.....	38
Figure 4.2: Dolostone package development. A: view upstream, towards left bank; foreground is 3 m across.	42
Figure 4.3: Detail of dolostone package development.....	43
Figure 4.4: Internal details in dolostone subfacies 2.	44
Figure 4.5: Nodular surface of a dolostone package. Left bank of River Coquet, view upstream (west).....	46
Figure 4.6: Fossil and pyrite occurrence in dolostone packages.	47
Figure 4.7: Coquetdale bay-fill facies assemblage.....	55
Figure 4.8: Reddened horizons and roots in the bay-fill facies assemblage.	58
Figure 4.9: Vegetated floodplain facies assemblage.	60

Figure 4.10: Shallow marine near-shore facies assemblage (A-D) and shallow marginal marine bay assemblage (E).	63
Figure 4.11: A suggested palaeogeographic development of the site at Coquetdale through time.	66
Figure 5.1: Log of Whitrope Burn section.....	72
Figure 5.2: Internal structure of beds at Whitrope Burn.....	75
Figure 5.3: Dolomite formation in rocks at Whitrope Burn.	76
Figure 6.1: Coldstream log showing facies and facies associations with bulk and specific $\delta^{13}\text{C}$ values.	84
Figure 6.2: Coquetdale log showing facies and facies associations with bulk and specific $\delta^{13}\text{C}$ values.	85
Figure 6.3: Coquetdale 'plant bed' section showing facies, facies associations and carbon isotope curves.....	86
Figure 6.4: Bulk carbon isotope values for the Ballagan Formation sites.	88
Figure 6.5: Bulk $\delta^{13}\text{C}$ org. values for the Ballagan Formation sites, to 4% TOC.....	89
Figure 7.1: Model of likely site locations illustrating their comparative positions and suggested marine/terrestrial influences.	102

1 Introduction

1.1 Background

This study investigates the sedimentological context of early Carboniferous vertebrate fossils, specifically tetrapods and lungfish, at three key sites in Northumberland and Southern Scotland. The aim is to determine the depositional environments of the Ballagan Formation.

The study forms part of the TW:eed project. The TW:eed project is a NERC funded, multi-disciplinary consortium project entitled 'The Mid-Palaeozoic biotic crisis: setting the trajectory of tetrapod evolution'. It is investigating the evolution of tetrapods, from aquatic and fish-like early limbed vertebrates to those adapted to walk on land. This is set into the context of the re-population of ecosystems in the earliest Carboniferous, following the end-Devonian Mass Extinction. The multi-faceted nature of this investigation lends itself well to a consortium project where different institutions contribute their expertise. It is led by a team of palaeontologists from the University of Cambridge, supported by the University of Southampton, the University of Leicester, the British Geological Society and the National Museums of Scotland.

The University of Leicester is providing the expertise to interpret the sedimentological settings within which the fossils are found and, more broadly to interpret the environment of deposition and the climate variability at the time. One of the TW:eed project objectives is to understand how the depositional environments that existed between 360-347 million years ago contributed to the preferential preservation of early tetrapods.

During the Devonian, tetrapods were aquatic but by the mid Viséan they had become fully terrestrial, although little is known about the how and where this evolution occurred. There was an apparent gap in the record of tetrapod evolution following the Hangenberg crisis and the End Devonian Mass Extinction which lasted until the mid Viséan, this gap being known as 'Romer's Gap' (Romer, 1956; Smithson *et al*, 2012; Kaiser *et al*, 2015). This gap had hampered our understanding of how tetrapods

developed during this crucial time. However, following some significant finds by the team at the University of Cambridge (Clack, 2002; Clack and Finney, 2005; Smithson *et al*, 2012) the TW:eed project was established to investigate further the evolution of tetrapods, and the environment in which this occurred.

Until 2007 tetrapod fossils had only been found at two sites in the Tournaisian (earliest Carboniferous): an articulated skeleton in Scotland and isolated bones from a site in Nova Scotia (Smithson *et al*, 2012). Following extensive and dedicated searching and collecting in southern Scotland and Northumberland, by the late Stan Wood and the team from the University of Cambridge, more material has since been found. This suggests that the previous gap in the fossil record at this time, was in fact a collecting artefact.

1.2 Stratigraphic and geological context

The TW:eed project concentrates on fossils from the Tournaisian Stage of the earliest Carboniferous and all the sites in this study are from the Ballagan Formation, part of the Inverclyde Group (Fig. 1.1). The Ballagan Formation is estimated to represent approximately 12 million years of deposition (Waters *et al*, 2011) and is characterised by inter-bedded sandstones, mudstones, and argillaceous dolostones (colloquially known as 'cementstones').

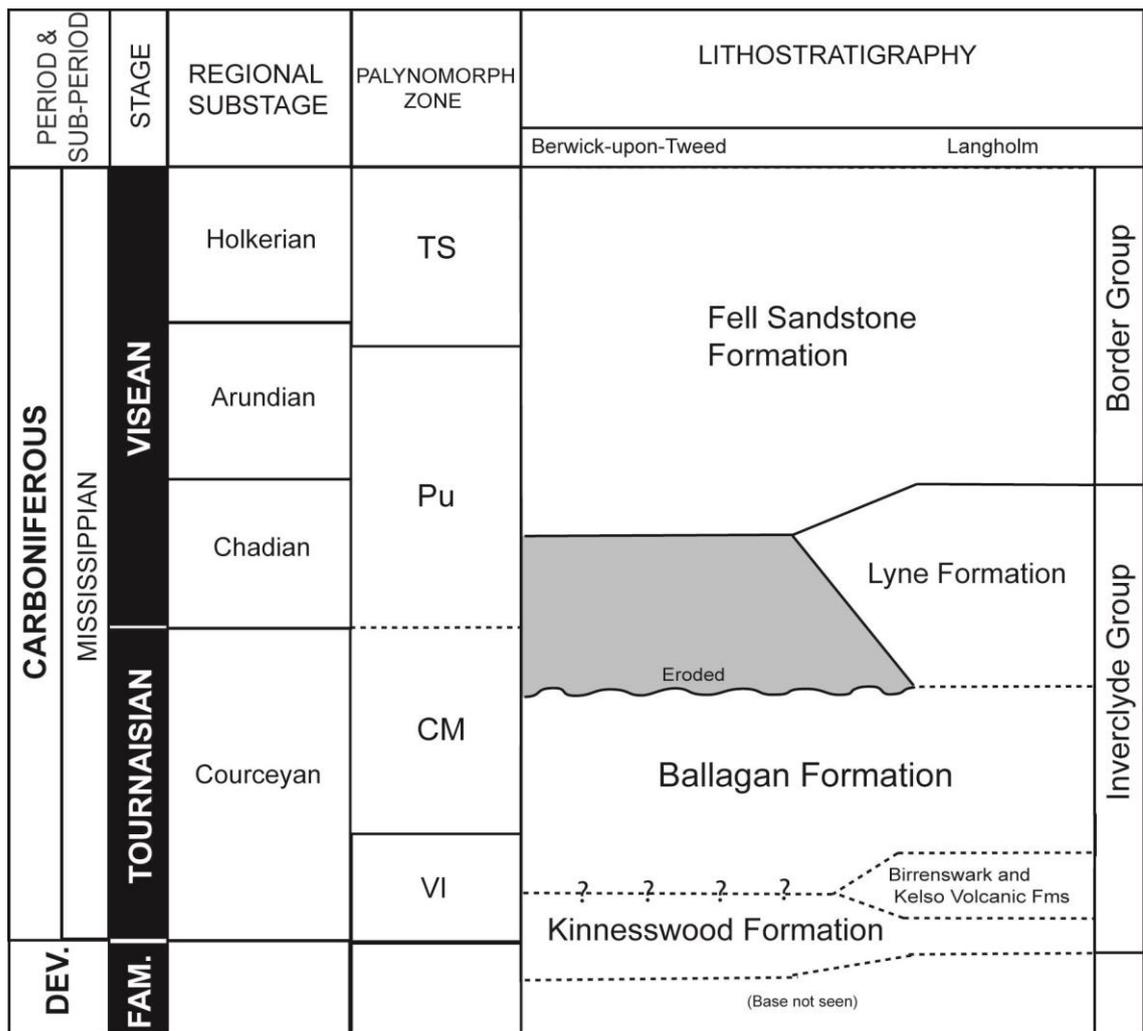


Figure 1.1: Stratigraphic framework of the Scottish Borders and Northumberland Basin. Adapted from Kearsy *et al* (2016). Palynomorph zones after Stephenson *et al*, 2002, 2004; Williams *et al*, 2005.

During the early Carboniferous the British Isles were lying about 4° S of the equator (Scotese and McKerrow, 1990). Climatic conditions at the time were likely to be semi-arid rather than truly arid, with seasonal monsoonal rainfall (Falcon-Lang, 1999; Kearsy *et al*, 2016). Evaporites found in the Tweed and Northumberland Basins, but not at the sites examined for this study, indicate periods of higher aridity.

The Tweed and Northumberland Basins are part of a series of grabens and half grabens formed by reactivation of late Caledonian basement structures and north-south extension during the late Devonian and early Carboniferous. The Northumberland Basin is a half-graben, formed along the inferred line of the Iapetus Suture, with its thickest rock succession being along its southern edge. Early syn-rift

sedimentation was dominated by fluvial deposits intercalated with basaltic lavas and dolostones (Guion *et al*, 2000). During the early Carboniferous the Tweed Basin was separated from the Northumberland Basin by the Cheviot Block (Stone *et al*, 2010), although there may have been at least an intermittent connection with the Northumberland Trough. In the early Carboniferous the Northumberland Basin was connected to the marine Peel and Solway basins to the west (Stone *et al*, 2010).

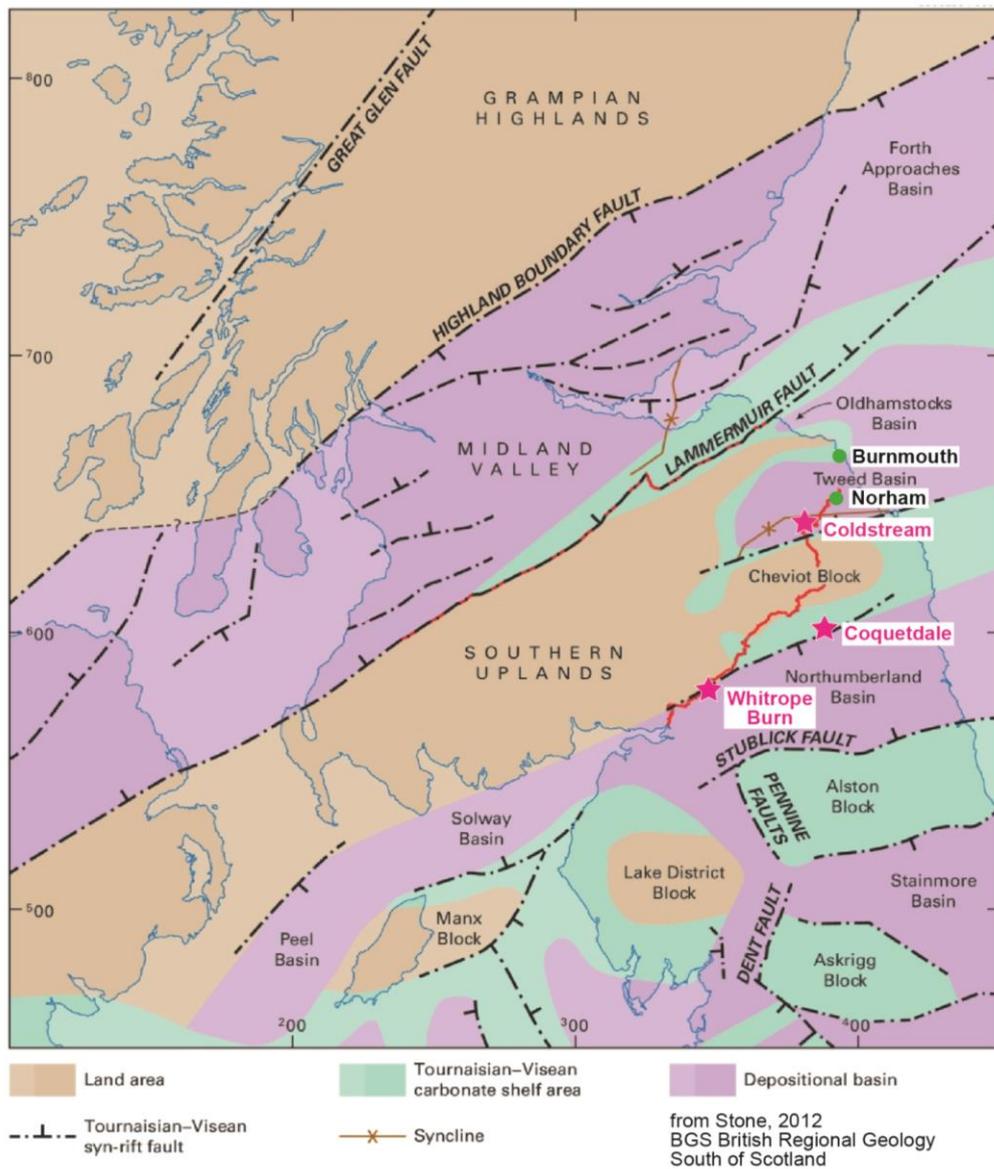


Figure 1.2: Palaeogeography of southern Scotland and northern England in the Tournaisian (early Carboniferous). Location of sites in this study (pink) and comparative sites investigated as part of the Tweed project (green). Adapted from Stone, 2012.

Deposition of the Ballagan Formation in the Tweed Basin is reported to have been on a low-lying coastal alluvial plain (Anderton, 1985; Scott, 1986; Andrews and Nabi, 1998; Andrews *et al*, 1991; Bennett *et al*, 2016). However, contemporaneous deposition in the Northumberland Basin is understood to have been into a shallow marine gulf (Belt *et al*, 1967; Leeder, 1974) with marine incursions from the west (Stone *et al*, 2010). The shallow marine waters of the Northumberland Basin would have been well-lit and warm due to their geographical position.

1.3 The study sites

As part of the TW:eed project much of the sedimentological research has focussed on the coastal section at Burnmouth and the core from a borehole at Norham (BGS borehole number NT94NW20) in the Scottish Borders (Fig. 1.2). Other sites both inland and around the coast have also been investigated (e.g. Smithson *et al*, 2012; Smithson *et al*, 2016).

For this part of the project three inland sites, identified as significant for their fossil finds, were selected for sedimentological investigation to determine the environment at the time of deposition. The site at Coldstream is on the southern edge of the Tweed Basin, and the sites at Coquetdale and Whitrope Burn are on the northern edge of the Northumberland Basin (Figs. 1.2). The sites were logged on a centimetre scale and samples examined in detail in order to put the fossil finds into context, both locally and within the wider development of the basins at this time.

1.4 Thesis Structure

Following a description of the techniques and methods used during the research (Chapter 2), each of the sites has been described in detail in individual chapters (Chapters 3-5). Samples were taken for bulk organic carbon isotope analysis from the Coldstream and Coquetdale sites and the results are presented in Chapter 6. Finally

the findings from all three sites are discussed within the wider basin context. The facies and facies assemblages are tabulated in the discussion (Chapter 7) to facilitate comparison across the three sites. Conclusions are drawn about the presence and preservation of tetrapod and other fossil material, and how these sites compare to other sites in the TW:eed project.

2 Methodology

Sections examined for this study were selected on the basis that tetrapod or lungfish samples had been found at the sites, and that they had potential for more fossil finds. Multiple techniques were employed to understand the sedimentology of the sections.

All three sections (Fig. 2.1) are exposed in river banks. The section at Coldstream (NT 851 402) is exposed in a bank of the River Tweed near Coldstream, the base of the section is not visible and the top is covered by soil (Appendix A, Fig. 1). The Coquetdale site (NT 900 062; Appendix A, Fig. 2) is in the bank of the River Coquet to the west of Alwinton in Northumberland, here the base of the section is in the river and the top is vegetated. The site at Whitrope Burn is a short section in the banks of the Whitrope Burn, north of Newcastleton in the Scottish Borders (NY 507 965; Appendix A Fig. 3).

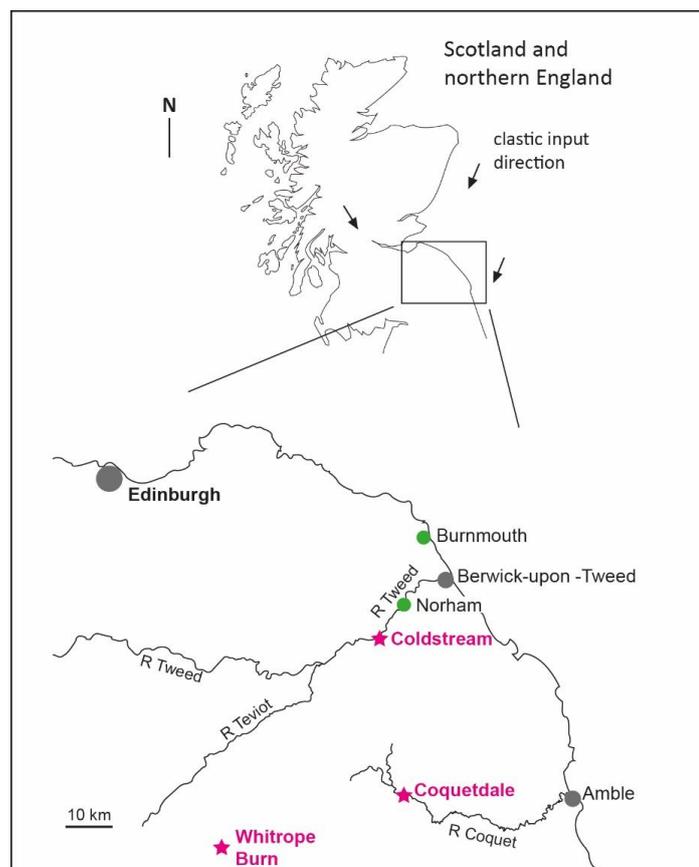


Figure 2.1: Outline map of Scotland and Northern England with detail of site locations. Pink sites represent those investigated in this study, green sites are other localities investigated as part of the TW:eed project.

2.1 Logging and sample selection

All sections were measured and logged on a mm-cm scale. Rock colour was assessed using the Munsell Rock Color Book.

At Coldstream a 3.7 m section was logged. Twenty individual samples were taken, and a block 11 cm thick from a particularly fossil rich horizon (31-42 cm on log) was extracted for detailed analysis. All samples were labelled with their orientation and, if fragile were wrapped in foil. Mudstone and siltstone samples were taken to determine structure, bulk organic carbon isotope values and fossil content; some samples were taken, for example, so that very small-scale structures identified in the field could be confirmed under the binocular microscope or in thin section.

At Coquetdale a 10 m section was logged. Fifty-two samples were taken, comprising dolostones, sandstones, siltstones and mudstones. Ten samples were taken from the Whitrope Burn site, seven of which were dolostones, and three of which were mudstones from the upper part of the section.

2.2 Ground rock surfaces

Samples were cut with a rock saw, and ground to give a smooth surface and reveal detail of the internal structure. Surfaces were then scanned at 1600 dpi using an Epson Perfection 3200 photo scanner.

2.3 Thin sections

Ultra-thin (20 μm) uncovered and polished thin sections were made, the ultra-thin nature of sections allowing fine detail of sedimentary structures in mudstones to be seen. Some fragile specimens were soaked in araldite to stabilise them before being cut for thin sections. All sections were made by Wagner Petrographic. Seventeen thin sections were made from the Coldstream section, and 6 from the extracted block; 28

thin sections were made from Coquetdale, and 5 from Whitrope Burn. Thin sections were scanned using an Epson Perfection V750 Pro to reveal the fabric of the samples and then examined using an Olympus BH-2 petrographic microscope fitted with a Canon EOS 5D camera.

2.4 Scanning Electron Microscope work

Where greater detail was needed to assess aspects of the rock samples, thin sections were examined on a Hitachi S-3600N Environmental Scanning Electron Microscope, and an Oxford INCA 350 EDX system was used to determine the elements present. The scanning electron microscope was also used to take images of microfossils.

2.5 Carbon isotope analysis

Carbon isotopes were analysed only for samples from Coldstream and Coquetdale. Samples were prepared at the University of Leicester and then sent to the Stable Isotope Facility at the British Geological Survey (BGS), a node of the NERC Isotope Geosciences Facilities. $^{13}\text{C}/^{12}\text{C}$ analyses were performed by combustion in a Carlo Erba 1500 online to a VG TripleTrap and Optima dual-inlet mass spectrometer. $\delta^{13}\text{C}$ values were calculated to the VDPB scale using a within-run laboratory standard calibrated against NBS-19 and NBS22. The elemental analyser has a precision of $<0.1\%$. Total organic carbon (%TOC) was also measured as part of this analysis. Mudstone and siltstone samples were selected for organic carbon isotope analysis of both bulk material and, separately, of specific woody fragments. The bulk samples give the $\delta^{13}\text{C}$ org. value of organic carbon material for the complete sample, whether of terrestrial or marine origin. The specific samples give the values only for the picked woody, terrestrial material. Siltstones were selected on the basis that they would contain enough organic matter to determine a value. Sandstone samples were not selected because it was judged that their organic carbon content was likely to be too low. Eleven samples from Coldstream were selected for bulk analysis, and nine for specific

terrestrial material analysis (three of the bulk samples being very close together). At Coquetdale seventeen samples were selected for bulk analysis and 18 for specific values. An additional eight samples were also taken from a nearby short section at Coquetdale, each of which were analysed for both bulk and specific values.

Preparation - bulk material:

Selected samples were crushed with a fly press, then about 5 grams of the sample were milled in a planetary mill. Milled samples were then analysed for total carbon and total sulphur values using a Leco CS230 Carbon/Sulphur Determinator. Samples with less than 3% carbon were deemed acceptable for bulk carbon isotope analysis. The total carbon content of samples at this stage is a combination of organic and inorganic carbon. Samples with values higher than 3% were usually discarded because analyses from the larger TW:eed project (C. Bennett pers. comm.) found that the inorganic carbon was likely to be dolomite. Whilst calcite was removed by NIGL prior to isotope analysis dolomite is very difficult to remove without potentially damaging the organic carbon. To ensure these analyses were consistent with the larger TW:eed dataset, the same processing techniques were used. Occasionally it was obvious that samples were very rich in plant material and these were then also analysed.

A few samples with a carbon content greater than 3% (and with obvious shelly content) were analysed by X-Ray diffraction (XRD) on a 'Bruker D8 Advance with DaVinci' powder diffractometer to ascertain the nature of the carbon content (for XRD traces see Appendix B). Where no dolomite was found to be present samples were also sent for carbon isotope analysis because any calcite present would be removed.

Specific material:

Samples prepared and sent for bulk carbon analysis were also prepared for picking of terrestrial fragments. Determining the $\delta^{13}\text{C}$ value for specific woody fragments from a given sample would inform the results from the bulk data in that part of the value for the bulk data may have come from a marine source. Understanding the values for the terrestrial data also gives insight into the atmospheric carbon content of the time period; this relationship is discussed in chapter 6.

Crushed material was soaked in a 5% solution of HCl overnight in a glass beaker, with sufficient HCl to just cover the sample. The beaker was then filled with de-ionised water, stirred, and allowed to settle for 24 hours. The water was then gently tipped off so as not to lose any of the sediment. De-ionised water was added again, left to settle for another 24 hours and then poured away again; this was repeated again such that each sample had three rinses in de-ionised water to remove any remnant HCl. Where sediment was found floating on the surface the washing was filtered through 64 µm sieve material and retained.

The washed sediment was then sieved through nylon mesh material, using acrylic sieve rings and new sieve material for each fraction of each sample to ensure that there was no contamination between samples. Samples were sieved at 250, 125 and 64 µm. Wet sieving using de-ionised water was employed since it was found to be gentler than dry sieving. Each fraction of material was then dried separately and packaged for picking.

Each fraction was picked separately under a binocular microscope taking care not to touch any of the material. Picked material was put into 5 x 3.5 mm silver capsules which were then put into 8 x 5 mm tin capsules, the tin being used to create a flash in the burning process used to ascertain $\delta^{13}\text{C}$ values. Capsules were then sent to the Stable Isotope Facility at the BGS.

The carbon isotope data for both bulk and specific values was converted into curves and fitted to height on the logs. These can be seen in Chapters 3, 4 & 6.

2.6 Facies and facies assemblages

Logged sections were divided into facies according to the lithology, sedimentary characteristics and fossil content. The facies were then grouped into assemblages to characterise the environment at the time of deposition. Facies and the assemblages are recorded on the logs (Figs. 3.1, 4.1 & 5.1) and detailed on the legend. Facies assemblages are described and interpreted in Chapters 3 - 5.

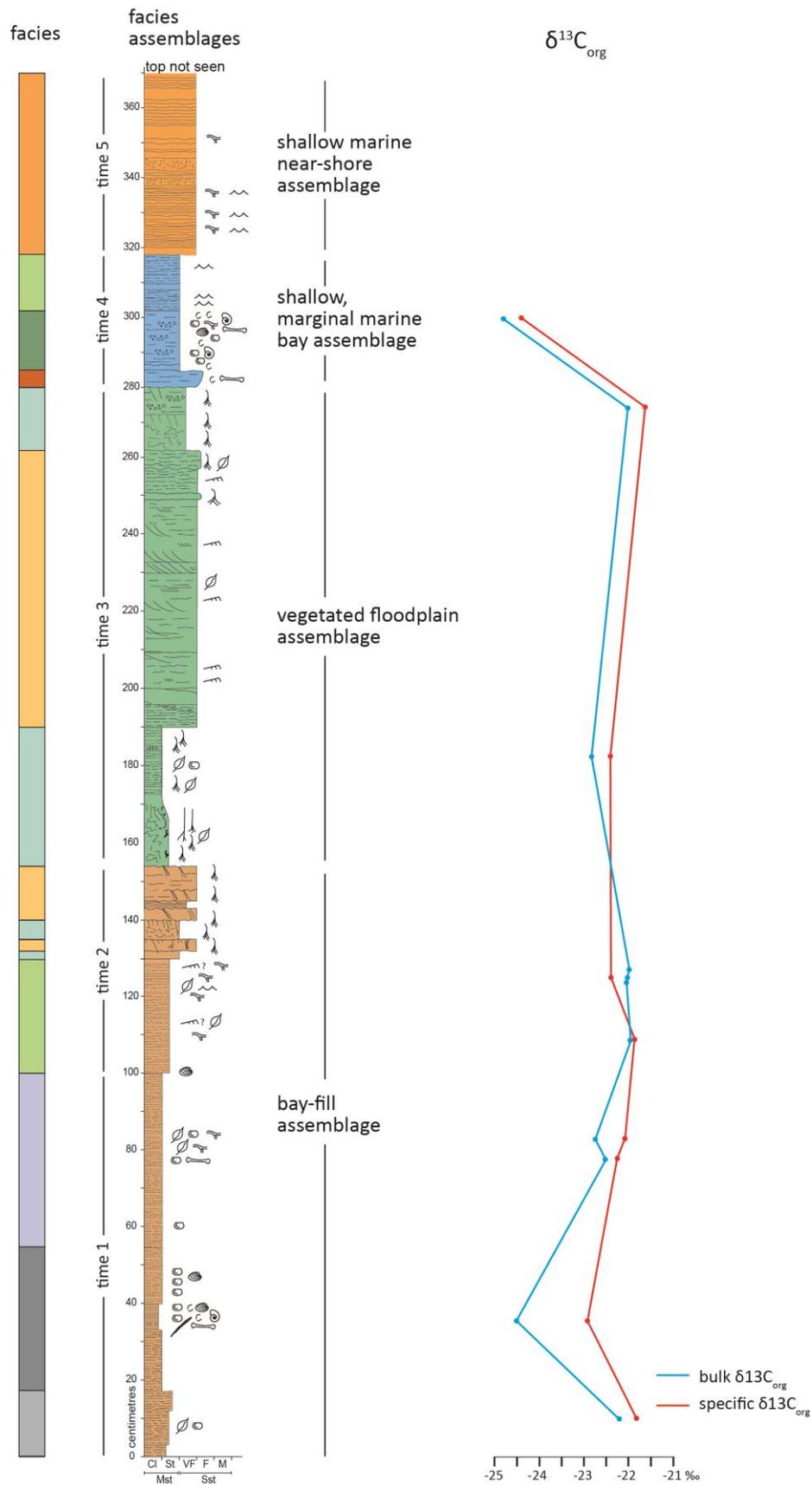


Figure 3.1: Log of Coldstream showing facies, facies assemblages and carbon isotope curves.

3.2 Coldstream facies assemblages

Analysis of the section at Coldstream indicates development from a marginal marine bay to a vegetated floodplain crossed by fluvial channels, a transgressive event and the establishment of shallow, marginal marine conditions, increasing water depth and mouth bar facies.

3.2.1 Bay-fill Facies Assemblage

Description

The base of the logged section (facies 1, 0-17 cm on the log; Fig. 3.2A,) comprises medium grey, coarsening-upward siltstones with sub-angular quartz and feldspar grains and rounded mud clasts. The siltstones are poorly sorted, calcite-cemented, and contain bone and scale pieces, shell fragments and organic material (Fig. 3.2A).

The overlying micaceous silty mudstones (17 - 55 cm, facies 2; Fig. 3.2B) are laminated, contain occasional outsized, rounded quartz grains, and are rich in ostracods, bone and scale fragments and plant remains. Finely dispersed pyrite is common throughout, and many of the organic fragments are pyritised. Pyrite nodules found on plant material, and commonly incorporating faunal fragments (Fig. 3.2C) are also present and can be up to 15 mm in size. There is no evidence of bioturbation. XRD analysis reveals that chlorite is present in these siltstones (Appendix B, Fig. 1).

A detailed study of a series of particularly fossil-rich beds, underlain by a layer with concentrated plant straps up to 10 cm wide, includes the mudstones between 33 and 40 cm on the log and reveals alternating laminae of coarser and finer mudstones. The coarser laminae are very rich in fossil and plant fragments (Fig. 3.2C). Faunal remains include ostracods (*Shemonaella* sp., *Cavellina* sp. and a possible specimen of *Carbonita*; C. Bennett, pers. comm.); the bivalve *Modiolus* (Wilson, 1965) (occasionally in the 'butterfly' position); 'spirorbiform microconchids' ('spirorbids') (Taylor and Vinn, 2006); bone fragments; scales and teeth of sarcopterygians (rhizodonts and lungfish), actinopterygians, possible chondrichthyans (*Ageleodus*); and some fragments of eurypterid carapace (Hibbertopteroidea, A. Ross, pers. comm.). Other important fossil

material, including tetrapod skull bones and vertebrae, *Gyracanthus* spines and scorpion fragments have also been found in these beds (Smithson *et al*, 2012). Ostracods are poorly preserved and are either moulds or retain only fragments of carapace. Plant material is represented as recognisable compressed straps, as vitrain (compressed gellified plant) and occasionally as fusain (charcoal) (Scott, 2010); spore fragments are also common. See Table 3.1, and Appendix C for details of beds and fossil abundances. The preferred environmental conditions for each of the fauna are detailed in the interpretation below and in Appendix C.

Fine-grained laminae contain very few fossil fragments, although small pieces of plant material (<4 mm) are present. There are occasional coarser-grained scours containing very small (<0.5 mm) ostracod valves in these fine-grained laminae, and a few paler grey, well-cemented nodules.

Ostracods and spirorbids occur almost throughout, although the ostracods vary in abundance according to the grain size of the laminae, with greater abundance in the coarser laminae. The very fine-grained beds are sparsely populated with ostracods (0.5-1 mm in size), and one bed is unique in that it contains no ostracods (or other fauna); coarser beds contain ostracods up to 2 mm in size.

It is interesting to note that the pyrite nodules mainly occur in coarser-grained laminae either side of a 15 mm thick, almost barren, very fine-grained bed which has compressed plant material at its base. The coarser beds either side of the central fine-grained bed are also where plant material is most abundant and the majority of the bone and scale fragments occur (Fig 3.2C).

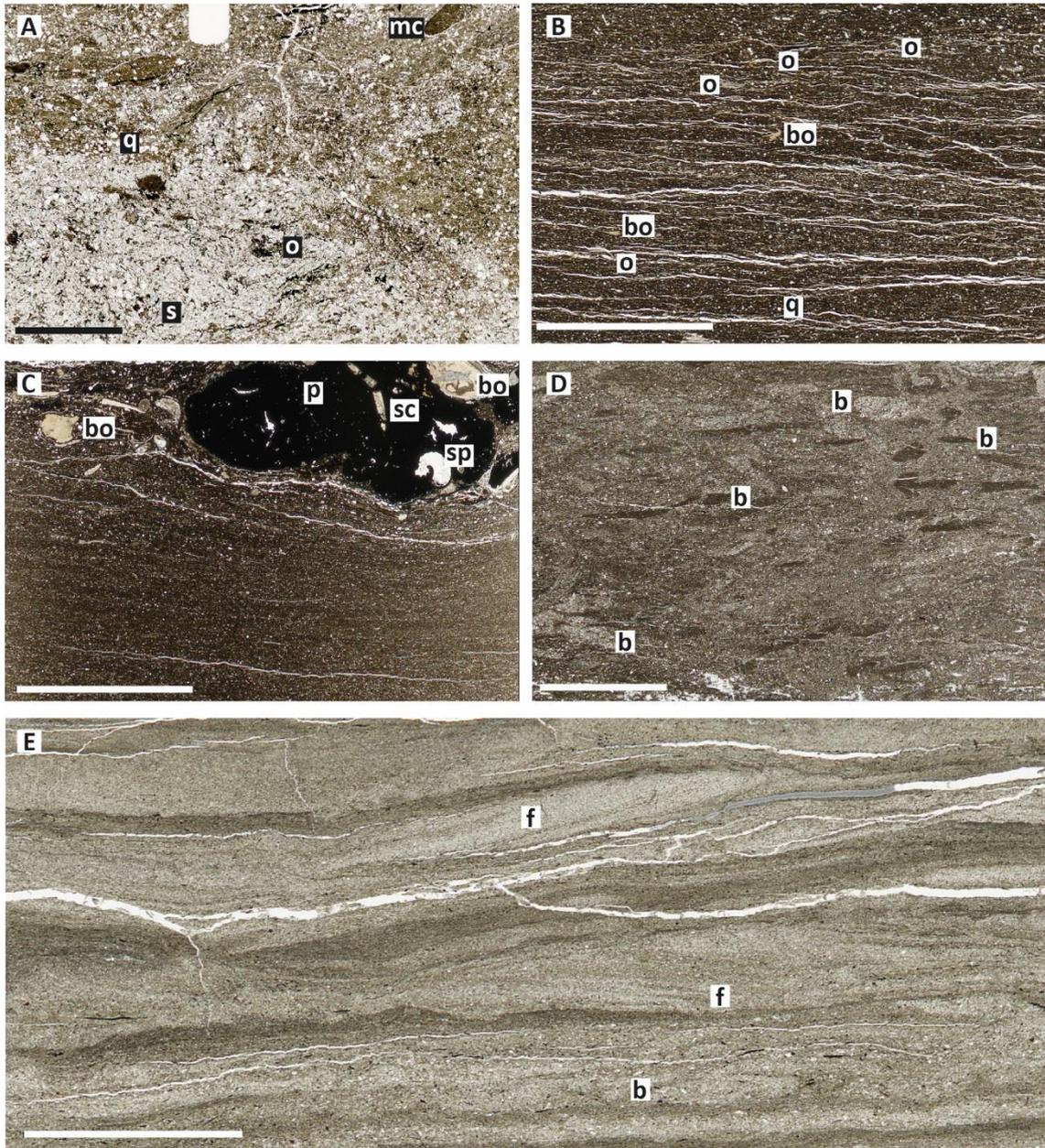


Figure 3.2: Bay-fill facies assemblage.

All scale bars = 5 mm. A. facies 1, siltstones with sub-angular quartz (q), mud clasts (mc), bone, scale and shell (s) fragments and organic material (o); B. facies 2, micaceous silty mudstones with occasional out-sized quartz grains (q), with bone fragments (bo), rich in ostracod valves (o); C. facies 2, pyrite nodule (p) and faunal fragments, including spirorbids (sp) and bone (bo) and scale (sc) pieces in fine-grained mudstone; D. facies 3, micaceous siltstones with larger (1.5 - 3 mm diameter), pale, quartz-rich burrows; and smaller (0.5 - 1.5 mm), darker, clay-rich burrows (b); E. facies 4, micaceous siltstone to very fine sandstone with ripple foresets (f) dipping in two directions and some bioturbation (b).

Bed	Lithology	Plant material	Plant, cubic, shiny - vitrain	Charcoal – fusain	Megaspores	Megaspore fragments, or poss. Scorpion cuticle	Spirorbids	Ostracods	Bivalve - <i>Modiolus</i>	Lungfish scales	Rhizodont scales	Actinopterygian scales	Actinopterygian tooth	<i>Ageleodus</i> tooth	Lepidotrichia	Bone	Eurypterid
Layer a 15 mm	Silty m'st, not laminated	✓				✓	✓	✓✓✓	✓		✓	✓			✓		
Layer b 18 mm	Silty m'st, not laminated	✓✓					✓	✓✓✓	✓✓ artic.			✓✓		✓			
Layer c 5 mm	Silty m'st, not laminated	✓✓✓			✓	✓?	✓✓	✓✓✓	✓✓	✓✓	✓✓	✓✓	✓ (x2)		✓	✓	
Layer d 12 mm	Silty m'st, not laminated	✓✓✓				✓✓	✓✓	✓✓✓	✓ (one)		✓✓	✓✓			✓	✓	
Layer e 20 mm	Top part, = layer d?	✓					✓	✓✓✓			✓✓	✓✓					
Layer e	Dark, fine-grained m'st, not laminated	✓✓ some, smaller				✓?	✓	✓✓ generally smaller									
Layer e	Layer f?, slightly siltier	✓✓✓	✓	✓		✓✓		✓✓✓ all sizes									
Layer f 5-10 mm	Silty m'st, not laminated	✓✓✓	✓✓✓		✓✓	✓✓?	✓✓	✓✓✓ some pyritised	✓		✓	✓			✓	✓	
Layer g 6-10 mm	Micaceous m'st	✓✓✓	✓✓		✓✓		✓	✓✓✓ also in patches				✓?					✓
Layer h 5 mm	Stuck to base – uneven base	✓✓ with nodes, some strap-like			✓✓			No ! ostracods									
Layer i 16 mm	?=bed below the block	✓ one piece					✓✓	✓✓✓	✓							✓ one piece	

Table 3.1: Details of fossil finds in beds from 33-40 cm on the logged section. Layer a is the top layer, layer l the base. ✓ = rare; ✓✓ = common; ✓✓✓ = abundant. Blue area indicates the fine-grained central bed.

From 55 – 130 cm on the log (facies 3 and 4) the micaceous siltstones and very fine-grained sandstones become progressively well laminated and the colour grades from medium grey to medium light grey. Chlorite grains likely contribute to the slight greenish tinge to the beds of the upper part of this section. The abundance of ostracods and bivalves (*Modiolus*) reduces; bioturbation is common, particularly in some beds; plant and spore material remains common; and towards the top there is some indication of wave action. Fine pyrite is dispersed throughout the beds from 55 – 100 cm, and is still detected towards the top of this section but is increasingly less common.

Bioturbation, from 55 – 100 cm on the log (Fig. 3.1), is of two types: larger (1.5 - 3 mm diameter), pale, quartz-rich burrows; and smaller (0.5 -1.5 mm), darker, clay-rich burrows with less pyrite than the host sediment (Fig. 3.2D). There are some bone/scale fragments, common plant material, and some spore fragments. Much of the plant material is pyritised but un-pyritised vitrain (Scott, 2010) is also present. Ostracods (indet.) and *Modiolus* occur but are less common than between 33 – 40 cm, and become increasingly less common up section.

The section between 100 and 130 cm (Fig. 3.1) is a coarsening-upward, rippled, micaceous siltstone to very fine sandstone with discontinuous laminae and occasional chlorite grains. Ripples (tentatively observed in the field, and confirmed in thin section) have foresets in two directions (Fig.3. 2E) and are probably symmetrical ripples. Plant material and organic matter is abundant and is particularly concentrated in slightly darker laminae. Some bioturbation is present in the siltier laminae. *Modiolus* occurs around 100 cm but then disappears, as do ostracods.

The uppermost part of this bay-fill assemblage (130 – 154 cm, facies 5 and 6) is composed of interbedded fine to very fine, fissile and micaceous, rooted sandstones (Fig. 3.3A). The beds are light grey to medium light grey with greenish grey patches. Coarser beds have erosive, undulating bases and highly irregular tops, and bed thickness varies laterally. Beds sometimes form small lenses which pinch out laterally (Fig. 3.3A). Roots are common, and beds are rich in organic matter. Some beds show mm-scale discontinuous laminae and possible ripple cross-lamination.

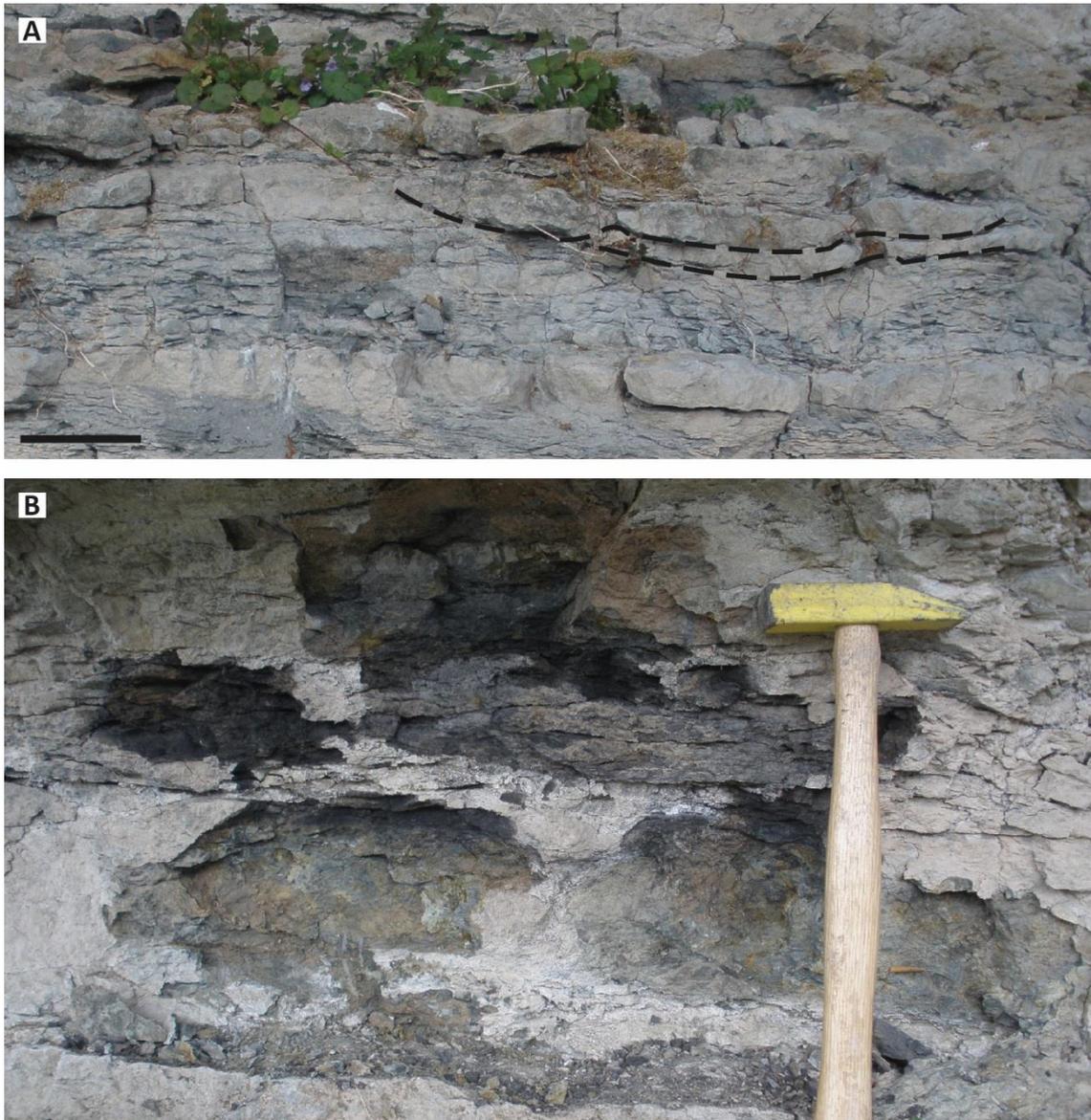


Figure 3.3: Bay-fill facies assemblage, facies 5 and 6.
A. scale bar = 10 cm. Facies 5 and 6, fine-grained rooted sandstones with scour channels, coarser beds have erosive bases and highly irregular tops; B. facies 5, greenish-grey siltstones and olive-black mudstones of a gleysol overlain by carbonaceous material.

Interpretation

This facies assemblage between 0 – 154 cm (fawn on Fig. 3.1) represents the filling of a shallow bay area, with an upward-coarsening suite of mudstones and siltstones which display increasing evidence of shallow water features.

The poorly sorted calcite-cemented beds at the base of this bay-fill assemblage represent transported material, probably by a small mass flow deposit into the bay.

Strap-like leaves, found at the base of the overlying bed, settled out after the main part of the mass flow.

The overlying fossiliferous beds contain transported faunal remains, all of which are marginal marine to non-marine species, the presence of bivalves in the 'butterfly' position indicating that material has not been transported far. These fauna represent a diverse community living in normally oxygenated conditions and resident nearby at the time. Alternating coarser, fossil-rich beds and finer-grained, fossil-poor beds indicate fluctuating siliciclastic deposition. Fine-grained beds occur only occasionally and probably represent settling from the tail end of a flow, sorting leading to the presence of only small plant fragments, small ostracods and occasional spirorbids.

Ostracods (mainly marginal marine to brackish water species, with one brackish to freshwater species; see Appendix C), and the 'spirorbids', here interpreted as brackish water species by faunal association (Schultze, 2009; Taylor and Vinn, 2006; Flugel, 2010; Carpenter *et al*, 2011, 2014) , are present throughout, but are more numerous in the coarser beds, interpreted as better oxygenated conditions. The ostracods are generally smaller in the very fine-grained beds but include larger specimens in the coarser-grained beds, suggesting either that the ostracods were quickly able to utilise the improved environment for development or that larger ostracods were transported from closer to shore. The latter is supported by the observation that fine-grained beds contain slightly coarser-grained scours containing ostracod valves, another indication of flows with varying erosive power.

The bone, scale and teeth fragments are all found in the coarser beds, suggesting that they too were probably transported to, or concentrated in, the site. It is likely that the flora and fauna are local since some plant fragments are preserved with good fidelity.

Plant material is ubiquitous but is especially prevalent in the coarser beds. The presence of strap-like leaves and spore material supports a location close to shore; occasional pieces of charcoal may also suggest a near-shore location but since it is known that charcoal can float for a long time (Nichols and Jones, 1992) and may travel considerable distances (Scott, 2010) this is not diagnostic. Although some of the plant

pieces have been pyritised, other material is preserved as vitrain; this material was either compressed early before pyritisation occurred, or perhaps the gelified nature prevented pyritisation, since the more 'open-weave' charcoal fragments can become pyritised. Gelified plant material with no pyritisation represents plant material that has remained in the water column for a long time before burial (Bak *et al*, 2015).

Framboidal pyrite and pyritisation of bioclastic material indicates a likely marine setting (Raiswell, 1982; Wignall and Newton, 1998; Taylor and Macquaker, 2000; Ferreira *et al*, 2015); the abundant plant material contributing to microbial sulphate reduction, and the production of authigenic pyrite, the resulting hostile environment contributing to the lack of bioturbation. The pyrite nodules, occurring in specific horizons and preserving uncrushed plant material indicate that pyritisation of fossil fragments occurred soon after deposition. Bone and scale fragments are sometimes into the pyrite nodules incorporated (Fig. 3.2C) and it may be this early pyritisation that helped to preserve the fauna in these beds. It is interesting to note that the pyrite nodules, and the majority of the bone and scale fragments, occur on either side of a very fine-grained bed. It is likely that the abundance of plant and faunal material contributed to the speed and extent of pyritisation and hence preserved more fauna, and that it was the presence of pyrite nodules that attracted attention to these fossiliferous beds, revealing their important contribution to our understanding of vertebrate populations at this time. Pyrite is common in the lower part of this bay-fill facies assemblage but then gradually reduces in abundance towards the top, indicating fresher conditions.

The appearance of the semi-infaunal bivalve *Modiolus* just below 40 cm on the log, and its co-occurrence with spirorbids indicates a brackish environment (Wilson, 1965; Wilson, 1989; Ballèvre and Lardeux, 2005; Taylor and Vinn, 2006; Williams *et al*, 2006; Schultze, 2009; Carpenter *et al*, 2011). Conditions must then have been sufficiently oxygenated to support larger invertebrates such as *Modiolus*. The two types of bioturbation present at around 80 cm on the log, and its pervasive nature, indicate increasing oxygenation and conditions more conducive to infaunal activity. However, diagenetic pyritisation of fossil fragments, dispersed pyrite and pyrite framboids,

confirms a continuing marine influence, degradation of plant material and dysoxic conditions in the sediment soon after deposition.

As grain size increases to silt to very fine sand-size (100 – 130 cm) wave ripples become evident, suggesting shallower water. The increased abundance of plant and other organic material testifies to the proximal nature of the source and a position closer to shore. Bioturbation is evident, and the presence of pyrite indicates a continuing marine influence. Detrital chlorite grains contribute to the slight greenish tinge of some of the beds and are likely derived from nearby igneous or metamorphic sources.

The bay-fill assemblage is completed by a series of very fine to fine-grained beds with roots. Frequent sub-aerial conditions are indicated by multiple levels of plant colonisation, and roots are apparent in most of the beds, colonisation occurring during hiatuses in deposition. Small coarser grained scours cut through these edge-of-bay beds (Fig. 3.3A). These uppermost beds of the bay-fill assemblage depict a transitional time between very shallow water and sub-aerial conditions, between the filling of the bay and the overlying floodplain environment.

This assemblage probably represents a protected, marginal marine to brackish-water environment, with a limited connection to an open seaway. Over time the shoreline prograded, the bay was infilled (time 1 - 2 of palaeogeography, Fig. 3.9), and vegetation colonised the very shallow-water or sub-aerial sediments. The abundant fauna suggest a generally hospitable environment hosting a diverse range of marginal marine animals.

3.2.2 Vegetated Floodplain Facies Assemblage

Description

This facies assemblage occurs between 154 and 280 cm (facies 5 and 6) and comprises poorly cemented micaceous mudstones and siltstones and packages of coarse- to fine-grained sandstones.

Between 154 and 190 cm greenish grey mudstones and siltstones occur (Fig. 3.3B). Organic matter and plant material are abundant and roots are common throughout; any original lamination has been destroyed by rooting. The olive black to greenish-black mudstones at 170 - 182 cm are particularly organic rich. The overlying medium light grey mudstones complete the succession, are still rich in organic matter, are finely bedded, and contain some poorly preserved ostracods and occasional small (0.75 mm) spirorbids. The ostracods are very small (0.1-0.5 mm) and concentrated in discrete patches. Some of the plant material has been pyritised.

A similar, but siltier, package is present from 262 to 280 cm, with a greeny-grey gleysol at the base overlain by rooted siltstones. In contrast to the succession between 154 and 190 cm however, this upper interval contains no fauna and only pyritised remains of possible roots.

Planar cross-stratified sandstones with ripple laminae have a sharp erosional contact with the fine-grained deposits at 190 cm. These fine-grained sandstones comprise a unit 70 cm thick with low angle cross-bedding dipping 4-6° towards the south (Fig. 3.4). Multiple reactivation surfaces are present within the package, often cutting out previous beds. Laterally the beds dip more steeply at about 18° but no ripples were observed on these steeper surfaces. The steeper cross-bedding grades laterally into that at the shallower angle (Fig. 3.5). Nearby cross-stratified beds display cross-bedding at an angle of 8°, in the same direction. The cross-bedded sandstones appear to lack bioturbation or evidence for roots and finish abruptly at 262 cm. However, some roots and more organic matter occur around 250-262 cm. A thin (2 cm) fine sandstone bed at 250cm, and the topmost bed at 257-262 cm have undulose bases and tops and contains root traces. The upper surface of this package of cross-bedded sandstones at 262 cm is sharp.

Interpretation

The floodplain facies assemblage represents a waterlogged environment with abundant vegetation and cut by a distributary channel.

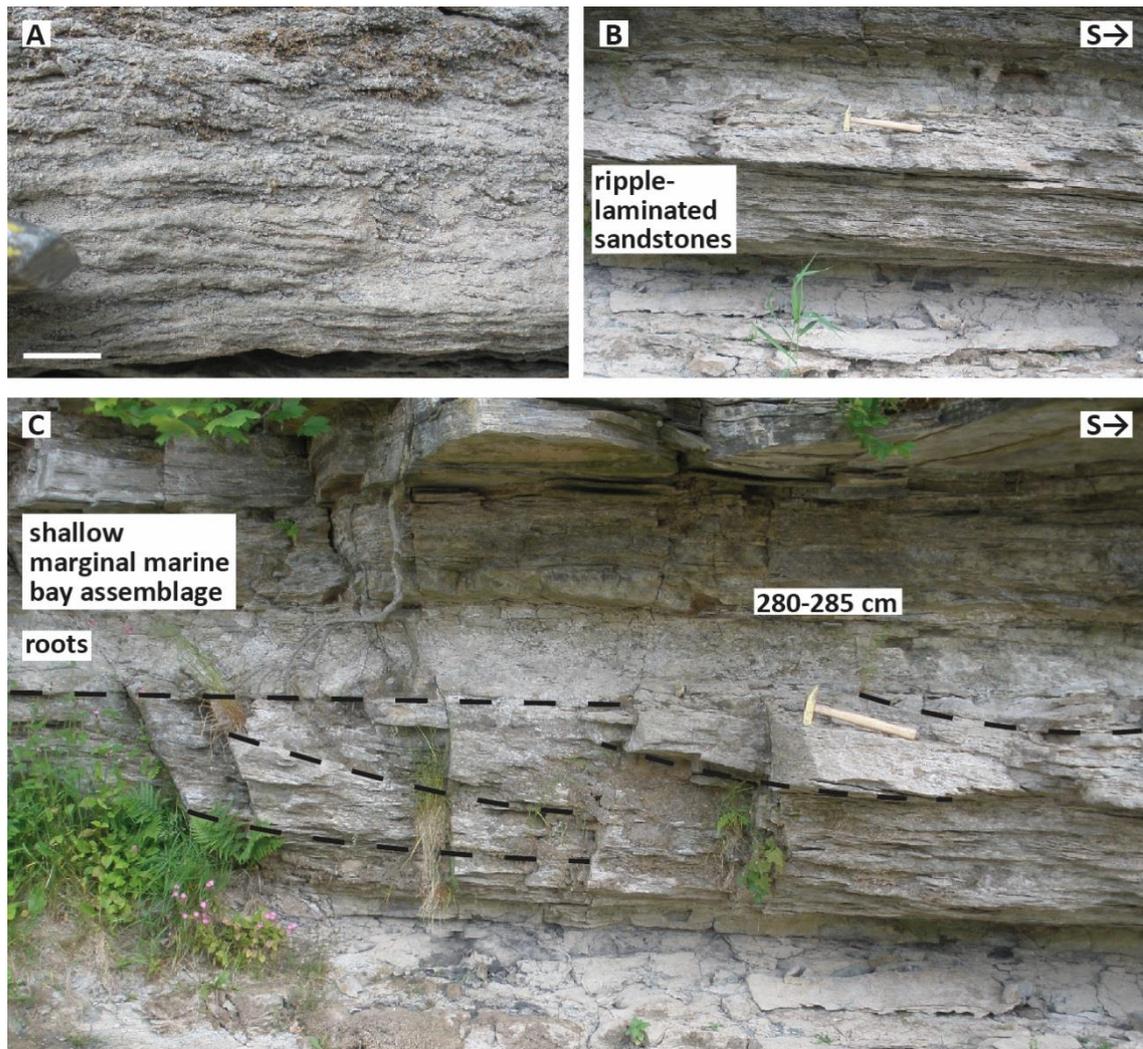


Figure 3.4: Vegetated floodplain facies assemblage. A. Scale bar = 2 cm. Ripple foresets dipping south; B. ripple-laminated sandstones with sharp erosional base, prograding towards the south and overlying the rooted gleysol and dark organic-rich mudstones of the floodplain facies (see interpretation); C. Cross-stratified sandstones showing multiple reactivation surfaces (dashed lines) and sharp top. The fine-grained sandstone at 280 – 285 cm terminates the rooted floodplain facies assemblage and initiates the shallow, marginal marine restricted bay assemblage.

The rooted greenish-grey mudstones and siltstones at the base of this assemblage are interpreted as a gleysol indicating wet, reducing conditions when the water table was high (Sheldon and Tabor, 2009; Mack *et al*, 1993). The dark colour and the organic-rich, carbonaceous nature of the bed suggest the incomplete development of an early coal-like bed and an environment similar to that of the later coal-forming swamps of the Viséan.

The presence of ostracods indicates that there were oxygenated pools which they could colonise. However, the size and discrete occurrences of the ostracods suggest

winning and concentration during mud deposition. The poor preservation, with only partial carapaces remaining, may be a result of transportation or it may be because of acidic post-depositional conditions. Brackish water/freshwater ostracods tend to have thin shells and are therefore particularly prone to damage and dissolution. Pyrite is present in clumps in this bed; some of this being pyritised plant material.

The upper package of floodplain deposits (262 -280 cm, time 3 in the palaeogeography (Fig. 3.9)) also commences with a greeny-grey, rooted gleysol but here the beds are siltier. The lack of fauna in these beds may be a preservational bias because pyritisation is extensive.

The cross-bedded sandstones are interpreted as a midstream bar in a distributary channel cutting through the floodplain. The channel-fill deposits are cross-stratified and have basal erosion surfaces (e.g. Bridge *et al*, 1986; Sambrook Smith *et al*, 2006). The homogeneous sandstone character (all fine-very fine sand sized, with no finer-grained laminae between) and the numerous reactivation surfaces suggest frequent, repeated bedform migration at high-flow stages (Rygel and Gibling, 2006). Low-angle cross stratified prograding strata dipping 4-6° towards the south to southwest indicate the palaeoflow direction. Apparent dip, measured on the ripple foresets seen within the dipping strata, is in a similar direction confirming that these sandstones represent the true downstream progradation direction of the package rather than lateral migration, thus suggesting a migrating mid-stream bar within a channel body (Bridge *et al*, 1998).

Variations in dip may be due to an undulating bar top migrating downstream with the angle of dip varying within the profile of the bar (Bridge *et al*, 1998). Roots towards the top of this bar feature suggest colonisation of the bar top, and that the migrating bar may have completely filled the channel.



Figure 3.5: Cross-stratified midstream bar deposits. Cross-stratified bar deposits migrating towards the south and displaying basal erosion surfaces and ripple foresets on the terminal, lower-angled edges. Note hammer for scale in the centre.

The top of the cross bedded package has been eroded and the sharp, but undulating, base of the uppermost bed (257-262 cm) suggests that it was a moderately erosive event. The slightly coarser nature of this uppermost rooted bed which temporarily inundated the area may represent a sheet flood, followed by plant colonisation (Fig. 3.4 C). The return to silty sediments indicates that floodplain conditions were re-established in this location with sediment supplied from coeval channels on the floodplain.

3.2.3 Shallow Marginal Marine Bay Facies Assemblage

Description

The fine-grained sandstone bed at 280-285 cm (facies 11; Fig. 3.4C) terminates the rooted floodplain facies assemblage and initiates the shallow, marginal marine bay assemblage. The bed lacks ripple cross-lamination, has an erosive base and a sharp top, contains abundant bone and shell fragments and has been bioturbated so that any original lamination has been destroyed.

The facies assemblage between 285 and 318 cm comprises siltstone (quartz, feldspars, micas and clay minerals) and patches of microcrystalline calcite and dolomite (facies 7 and 4). Lacking lamination and mottled black, the siltstone, with some bone and scale fragments, and the calcite/dolomite patches have been mixed by bioturbation. The lower part of the assemblage contains crushed ostracod valves (0.5 – 1 mm in size)

and some bone and scale fragments in mixed clays and silts; these are overlain by a 2 cm-thick layer of spirorbids (Fig. 3.6A). As the density of spirorbids declines abundant small (0.5 – 1 mm) articulated ostracods predominate. These ostracods, in discrete patches and mixed with silts and carbonate mud (Fig. 3.6B), are then overlain by heavily bioturbated mixed siltstone and carbonate mudstone with both articulated and disarticulated ostracods and some bone and scale fragments.

Centimetre-scale symmetrical ripples are visible in outcrop from 302 -318 cm (facies 4) and a bed at the base of these rippled siltstones contains abundant disarticulated ostracod valves ranging in size from 0.5 to, occasionally 3 mm, but predominantly 0.75 -1.5 mm. Most valves are in 'normal' orientation (valves facing upwards and downwards) indicating a very low energy environment, although some are at higher angles, representative of a degree of bioturbation (Guernet and Lethiers, 1989). Many (at 300 cm on the log) are also stacked as in the 'cup in cup' ostracod valves of Wakefield (1995), or 'en piles d'assiettes' of Guernet and Lethiers (1989) (Fig. 3.6C).

A key feature of the assemblage is the abundance of spirorbids and the monospecific ostracods. The lower part is dominated by spirorbids (Fig. 3.6A) overlain by abundant, very small (0.5 – 1 mm) articulated ostracods (Fig. 3.6B), whereas the upper part contains dense bioturbation overlain by abundant monospecific disarticulated ostracod valves (Fig. 3.6C). In this assemblage the well-preserved spirorbids and ostracods retain the calcite of their shells.

Dolomite is first noted in this assemblage and is present in two types. Within the siltstones dolomite forms as euhedral to subhedral grains, sometimes with higher magnesium centres, grains are up to 50 μm in size, and some calcite is also present. In the carbonate mudstone patches the dolomite is microcrystalline (5-20 μm ; (Folk, 1959), anhedral and is mixed with calcite (Fig. 3.7A-C).

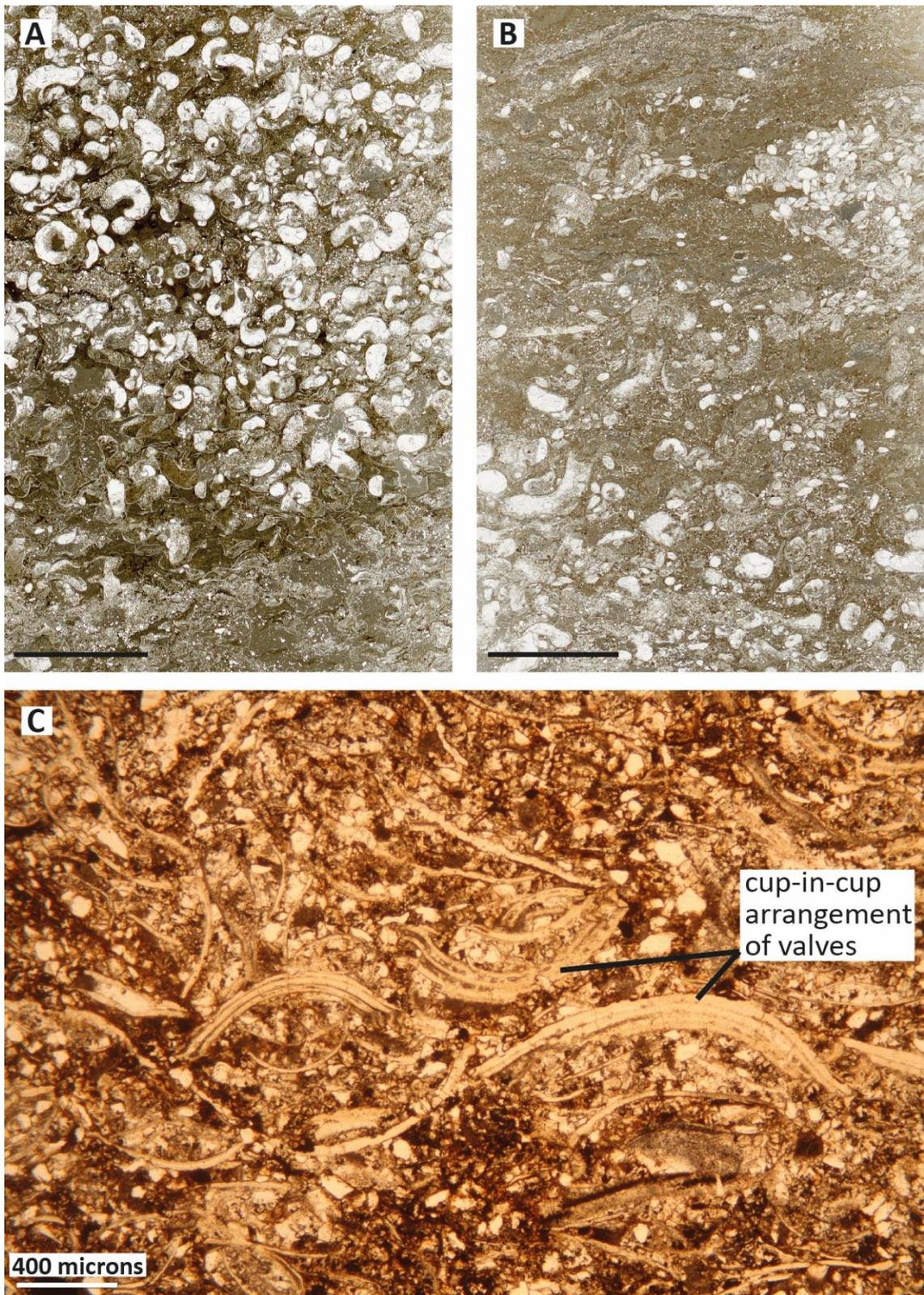


Figure 3.6: Shallow marginal marine bay assemblage. Scale bars = 5 mm unless otherwise labelled. A. Abundant spirorbids in mixed silts and clays; B. Ostracod carapaces and valves, both dispersed and concentrated into patches overlying the spirorbids bed; C. cup-in-cup arrangement of monospecific ostracod valves indicating shallow water shoreline conditions.

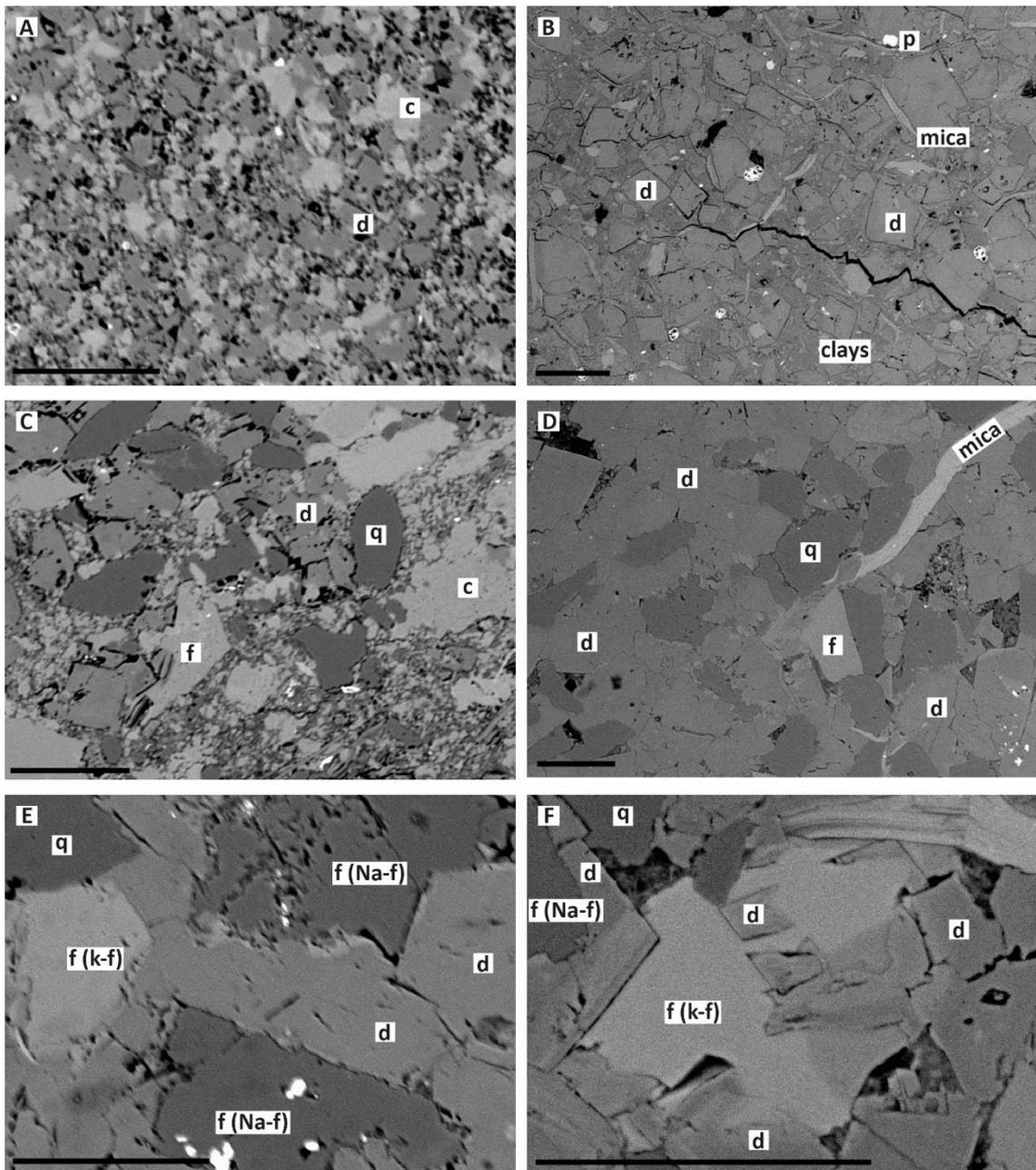


Figure 3.7: SEM images of dolomite growth.

All scale bars = 50 μm . A. microcrystalline calcite (c) and calcite partially replaced by dolomite (d); B. rhombic dolomite growing freely in clay-rich silty and calcitic fractions (facies 7) (p - pyrite); C. subhedral to euhedral dolomite in mixed silty fraction, with calcite, facies 7 (q - quartz, f - feldspar); D. subhedral (coalesced) dolomite replacing/overgrowing very fine sandstone of facies 8; E. dissolution of sodium feldspar grains and corrosion of crystal boundaries in facies 8 (k-f - potassium feldspar, Na-f - sodium feldspar); F. intergrowth of dolomite and potassium feldspar crystals, either dolomite growing into and replacing the feldspar or overgrowth of authigenic feldspar (facies 8).

Interpretation

The erosive-based, fine-grained sandstone bed with abundant bone and shell fragments at 280 – 285 cm is interpreted as a transgressive lag deposit with a

ravinement surface at its base (Plint *et al*, 2012). It represents a local marine transgression which re-established the shallow-water environment (Fig. 3.4C). The reworked bioclastic material must have been eroded from adjacent fossil-rich areas, perhaps the underlying bay-fill facies, and concentrated in the lag deposit.

The overlying fine siltstones and microcrystalline calcite and dolomite of this assemblage represent a shallow, brackish-water, restricted bay setting.

The abundant, and yet low diversity fauna indicates that environmental conditions severely limited the species which could populate the area. This is characteristic of brackish-water environments (e.g. Hudson, 1963). In very restricted environments individual species adapted to the particular conditions can become highly abundant. In this case the brackish-water spirorbids and monospecific ostracods (indet.) have thrived. Low diversity can also be related to soft substrates (Brenchley and Harper, 1998) but is unlikely to be the limiting factor here because overlying burrows are not compacted, indicating a degree of competence, and early lithification. Despite the predominance of a low-diversity fauna the influx of occasional bone and scale fragments within silty fractions indicate that other fauna did, however, exist nearby.

Brackish-water conditions are associated with lagoons, estuaries and a variety of deltaic environments (Brenchley and Harper, 1998) where there is freshwater dilution of marine waters and/or imperfect tidal exchange (Wakefield, 1995). At Coldstream the brackish water environment is likely due to some restriction of the bay, perhaps by mouth bars, or very shallow water, limited exchange of marine waters, and to freshwater input from a prograding fluvial system (time 4 of Fig. 3.9)

Brackish-water can also restrict the size of fauna but the small size of the ostracods here may represent juveniles only (Hudson *et al*, 1995), suggesting either that conditions were not conducive to further development, or that a degree of hydrodynamic size sorting occurred (Wakefield, 1995). The latter suggestion is supported by the presence of discrete concentrations of articulated ostracods of similar sizes at 290 cm; the articulation being indicative of rapid burial.

However, most of the ostracods are preserved as disarticulated valves, the diverse angles of the valves suggesting bioturbation (Guernet and Lethiers, 1989) and/or mixing of the sediment by storm or shoreline hydrodynamics. The unusual 'cup in cup' arrangement of ostracod valves, found in situations with abundant, monospecific fauna and characteristic of lacustrine or lagoonal environments, is indicative of the gentle but constant agitation of very shallow water found at the shorelines of lakes or lagoons (Guernet and Lethiers, 1989; Wakefield, 1995). This supports the sheltered nature of the bay at this time. Young (1896) also described nest-like groups of ostracod valves from a shoreline environment in the Dinantian Campsie Limestones.

Despite the semi-restricted nature of the bay, the mixed patches of silt-sized grains and microcrystalline calcite and dolomite, the abundant shelly material, and the bioturbation evident at 300 cm indicate sufficient oxygenation to support an infaunal/epifaunal biota. This is attributed to continued freshwater input.

This facies contains the first noted occurrence, in this section, of inorganic calcite and dolomite. Calcium carbonate within the sediment may have been derived from bay waters saturated with calcium carbonate, or from the dissolution and re-precipitation of calcite from the abundant shell material present. Calcite is mainly present as authigenic growth within silty fractions but it is also seen as microcrystalline calcite, formed in shelter porosity, perhaps where saturated waters precipitated calcite directly (Fig. 3.7A).

Dolomite is present in both the silty and microcrystalline patches and is likely to be a diagenetic alteration of the calcite, the magnesium for the dolomite being supplied either from the mixing of fresh and marine waters within the bay (Humphrey, 2000), by subsequent marine flooding as relative sea level rose, or by later percolation down from the overlying shallow marine sandy mouth bar deposit. In places calcite has been completely replaced by dolomite. Dolomite in the silty fractions is euhedral to subhedral, implying space to grow, and is anhedral in the microcrystalline fractions. In the silty fractions dolomite crystals appear to have grown between the clay minerals, quartz and feldspar grains or to partially replace grains such as feldspars. Junctions

between grains often appear to be ragged, suggesting that dissolution and/or corrosion of grains was occurring as dolomite formed.

3.2.4 Shallow Marine Near-Shore Facies Assemblage

Description

This assemblage (318-370 cm on log, facies 8) consists of sharp-based, flaggy, wave-rippled, fine- to very fine-grained sandstones with undulating upper and lower surfaces and fine internal lamination (Fig. 3.8A). Wave ripple foresets can be seen in both hand specimen and in thin section (Fig. 3.8B). Occasional articulated ostracods and some fragmentary pieces of ostracod carapace are present in the slightly more organic-rich laminae. Bioturbation is common, although much of it is on a mm-scale so internal lamination and wave ripple foresets are still apparent (cryptobioturbation of Pemberton and Gingras, 2005 and Pemberton *et al*, 2008). Some beds contain larger burrows (1-3 mm wide) and are more disturbed; fallen blocks found locally display abundant bioturbation on lower surfaces. Some pyrite is present. These sandstones are primarily composed of quartz with grains of Na- and K- feldspars, biogenic allochems and Mg-bearing clays, with some intergrowth of dolomite (Fig. 3.7, D-F).

Interpretation

The flaggy nature of the wave-rippled sandstones is characteristic of mouth bars (Elliott, 1974; Shiers *et al*, 2014), or the finer deposits of a fluvial dominated delta (Bristow, 1988; Kane, 2010), the symmetrical ripples indicating wave reworking of material in shallow water. Here the flaggy sandstones are interpreted as mouth bars since they are horizontal (parallel to the thin sandstone which cuts off the top of the cross-bedded sandstones) and there is no remaining evidence of current ripples. The underlying facies assemblages, and these wave-rippled mouth bar sandstones indicate that a standing body of water developed that precluded vegetation, although water depths were still shallow. Some dolomite growth within the sandstones appears to have occurred later since dolomite is seen to be overgrowing and corroding feldspar and quartz crystals (Fig.3.7, D-F), (replacement cf. Al-Awadi *et al.*, (2009).

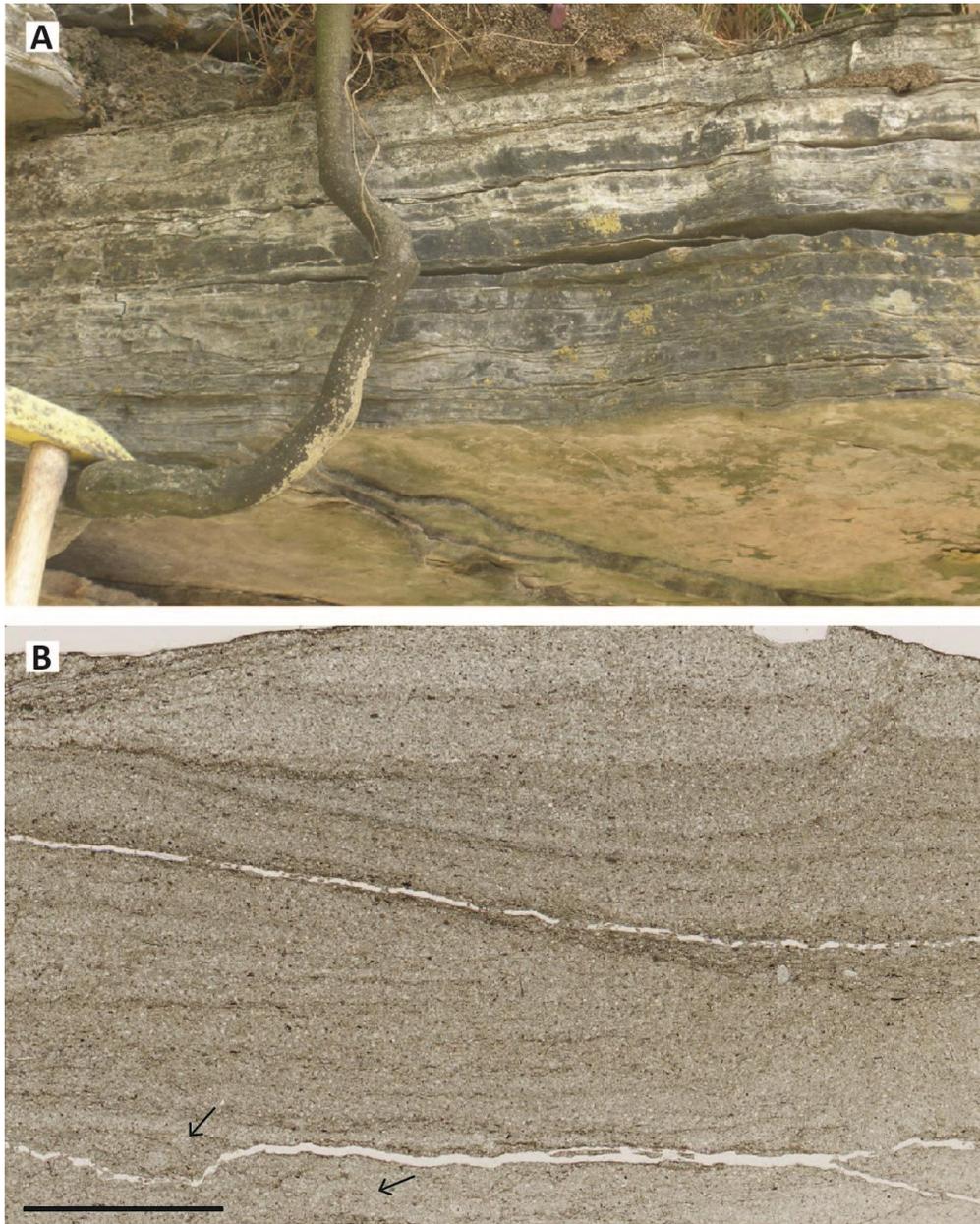


Figure 3.8: Shallow marine near-shore facies assemblage. Scale bar = 5 mm. A. flaggy wave-rippled fine-grained sandstones of the mouth bar assemblage; B. wave-ripple foresets as seen in thin section, with cryptic bioturbation towards the base (arrowed).

There is no evidence of the low-diversity brackish water fauna found previously which suggests an increase in marine influence and a less restricted bay environment; the rare ostracods and increased amounts of organic matter found along certain foresets have likely been transported, suggesting a complex interaction between fluvial and wave processes. The bioturbation suggests that it was a well oxygenated environment. Elliott (1974) observed that facies associations similar to these mouth bars may also be

produced by minor river deltas discharging into lagoons so it is likely that, despite transgression, the bay was not yet fully open to a seaway. This facies assemblage is represented by time 5 on Fig. 3.9.

3.3 Isotope analysis

Mudstone and siltstone samples were taken for carbon isotope analysis and were spread throughout the section (Fig. 3.1). Eleven samples were taken for bulk $\delta^{13}\text{C}_{\text{org}}$ analysis and nine samples of picked woody material were also analysed for specific $\delta^{13}\text{C}_{\text{org}}$ values (see Chapter 2 for sample preparation). At Coldstream the bulk values range from -21.9 to -24.8‰ whereas the specific samples vary between -21.6 and -24.4‰. However, at 36 cm (facies 2) the bulk value drops to -24.51‰ and the specific value drops to -22.91‰. This change coincides with the fauna-rich beds. The isotope profiles are discussed further in Chapter 6.

At Coldstream the specific values are typically lower than the bulk values but they also track the bulk values very closely which is unusual and not fully understood (see Chapter 6 for further explanation and interpretation).

3.4 Sedimentological development of the site at Coldstream

The exposure commences with laminated, marine bay fill mudstones and siltstones where fossils are preserved in the coarser laminae. Fossil fragments represent marginal marine to non-marine fauna that must have been resident nearby at the time. Lungfish and tetrapod bones have been found in these rocks, including tetrapod skull bones and vertebrae. The plant straps underlying the fossil-rich beds, and the preservation of fossil material in the coarser laminae, indicate that they have been transported to the site, and that the shoreline was nearby (T1 of Fig. 3.9).

Fossil content diminished and the abundance of wave-ripples increased as the bay fill assemblage coarsened upward and water depths become shallower. The top of the

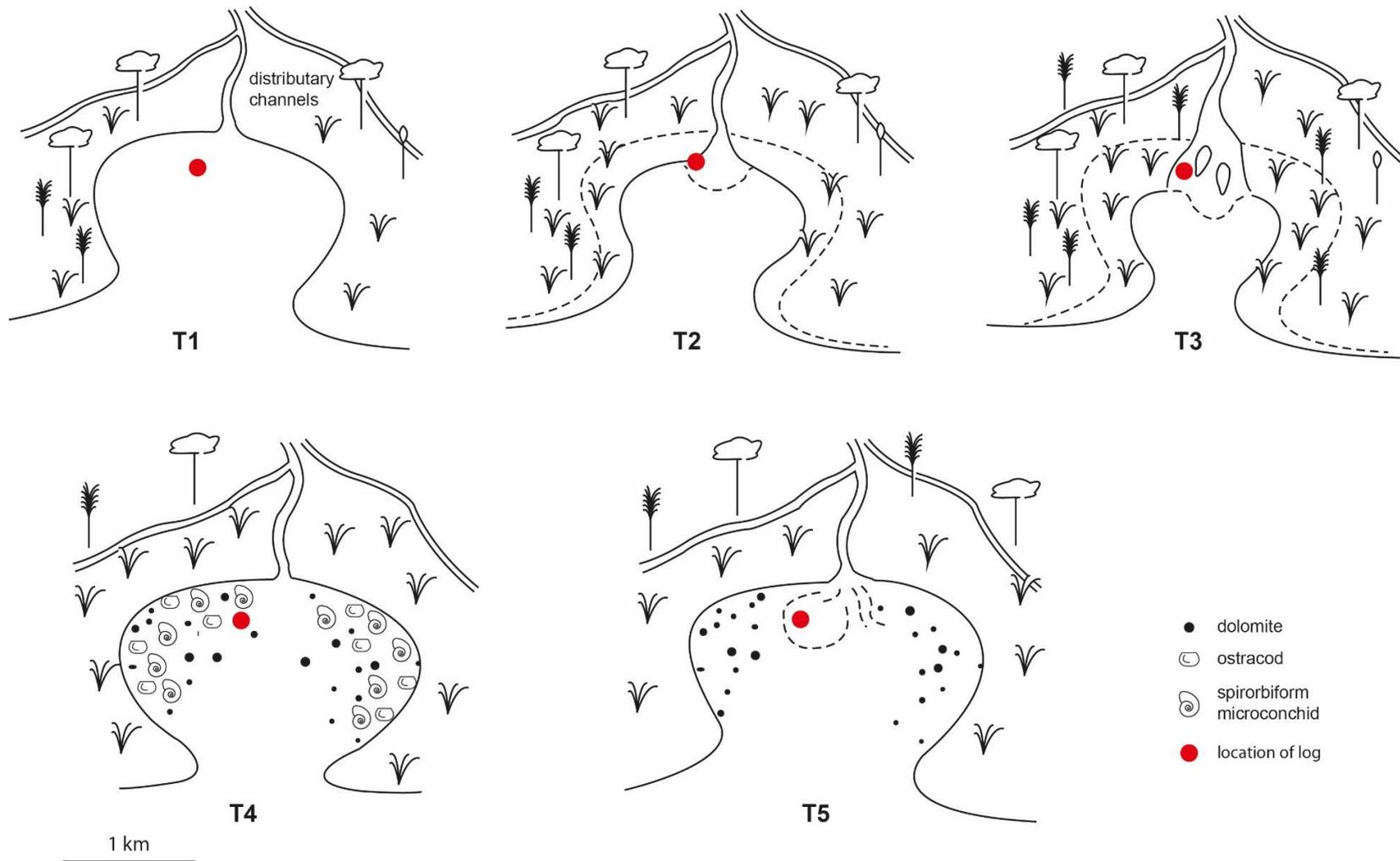


Figure 3.9: Development of the facies assemblages through time.

T1 = time 1- the lower bay fill facies assemblage; T2 represents the later stages of bay fill, with some plant growth; T3 represents the development of floodplain conditions; T4 and 5 illustrate transgression and the return to shallow marginal marine conditions, later with fluvial influence and mouth bars.

bay-fill assemblage contains roots where intermittent subaerial conditions prevailed (T2 of Fig. 3.9).

The protected, semi-restricted, marginal marine to brackish water environment of the bay fill assemblage is superseded by a waterlogged floodplain with abundant vegetation, dissected by small distributary channels (T3 of Fig. 3.9).

A marine transgression, represented by a fossil-rich transgressive sandstone bed, terminated the floodplain conditions and re-established a shallow-water environment with brackish water and a restricted fauna (T4 of Fig. 3.9). A limited diversity (only two species) of fauna is present, but individual species occur in great abundance and this abundance may have contributed to the arrangement of 'cup in cup' ostracod valves, an indicator of very shallow shoreline conditions. Here the shallow marginal marine bay assemblage represents a semi-restricted, very shallow water environment which maintains a connection to the sea. The siltstones of this shallow water environment contain authigenic calcite, and dolomite formed by later diagenesis. Although dolomite is present here it is not in the form of the typical dolostone beds of the Ballagan Formation, perhaps because dolomitisation is only partial.

The wave-rippled mouth bar sandstones represent increased fluvial input to the bay at this location and may suggest an increase in water depth, relative to the ostracod cup in cup facies, and a less restricted bay position (T5 of Fig. 3.9). Dolomite rhombs present within the siliciclastic deposits are diagenetic. The flaggy sandstones of this upper part of the section, although containing some dolomite still look like the flaggy, wave-rippled sandstones that they were deposited as. Perhaps here, with greater fresh water input and a more open marine environment with available sulphate, the reduced effect of dolomitisation means that the typical dolostones of the Ballagan Formation did not develop.

4 Coquetdale

4.1 Introduction

The section at Coquetdale was chosen for sedimentological analysis because lungfish material had been found nearby and fossils found in the section suggested the sedimentology warranted further investigation (Smithson pers. comm., 2012).

The section at Coquetdale is 10 metres thick and was logged on a millimetre scale. The logged section (Fig. 4.1) was divided into facies according to lithological and sedimentary characteristics, including fossil content, of the rocks. These facies were then grouped into facies assemblages (Fig. 4.1) that were used to interpret the environment and its evolution through time.

facies		legend					
f1	siltstone/sandstone with mud clasts and shell fragments	f9	mudstone, significantly disturbed by roots	●	pyrite nodule	○	ostracod
f2	mudstone, laminated, sometimes fossil-rich	f10	carbonate cemented siltstones with faunal fragments	~	wave ripple	⊖	bivalve
f3	siltstone, bioturbated	f11	fine to medium sandstone, erosional, sometimes with HCS	~	ripple cross-lamination	⊖	broken shells
f4	siltstone, laminated, sometimes current or wave-rippled			=	horizontal lamination	⊖	bioturbation
f5	mudstone and siltstone, with some roots		facies assemblages	○°○°	mottled fabric	⊖	bone/scale
f6	very fine to fine sandstone, sometimes cross-bedded, sometimes rooted		bay fill assemblage	⊖	planar cross bedding	⊖	articulated bones
f7	mixed silts and microcrystalline calcite/dolomite (fossil rich)		vegetated floodplain assemblage	⊖	rootlets	⊖	teeth
f8	sandstone, planar-bedded, wave-rippled		shallow marginal marine bay assemblage	⊖	strap-like leaves	⊖	spirorbiform microconchid
			shallow marine near-shore assemblage	⊖	plant material	⊖	gastropod
				⊖	intraclasts	⊖	convolute bedding

Legend for log below

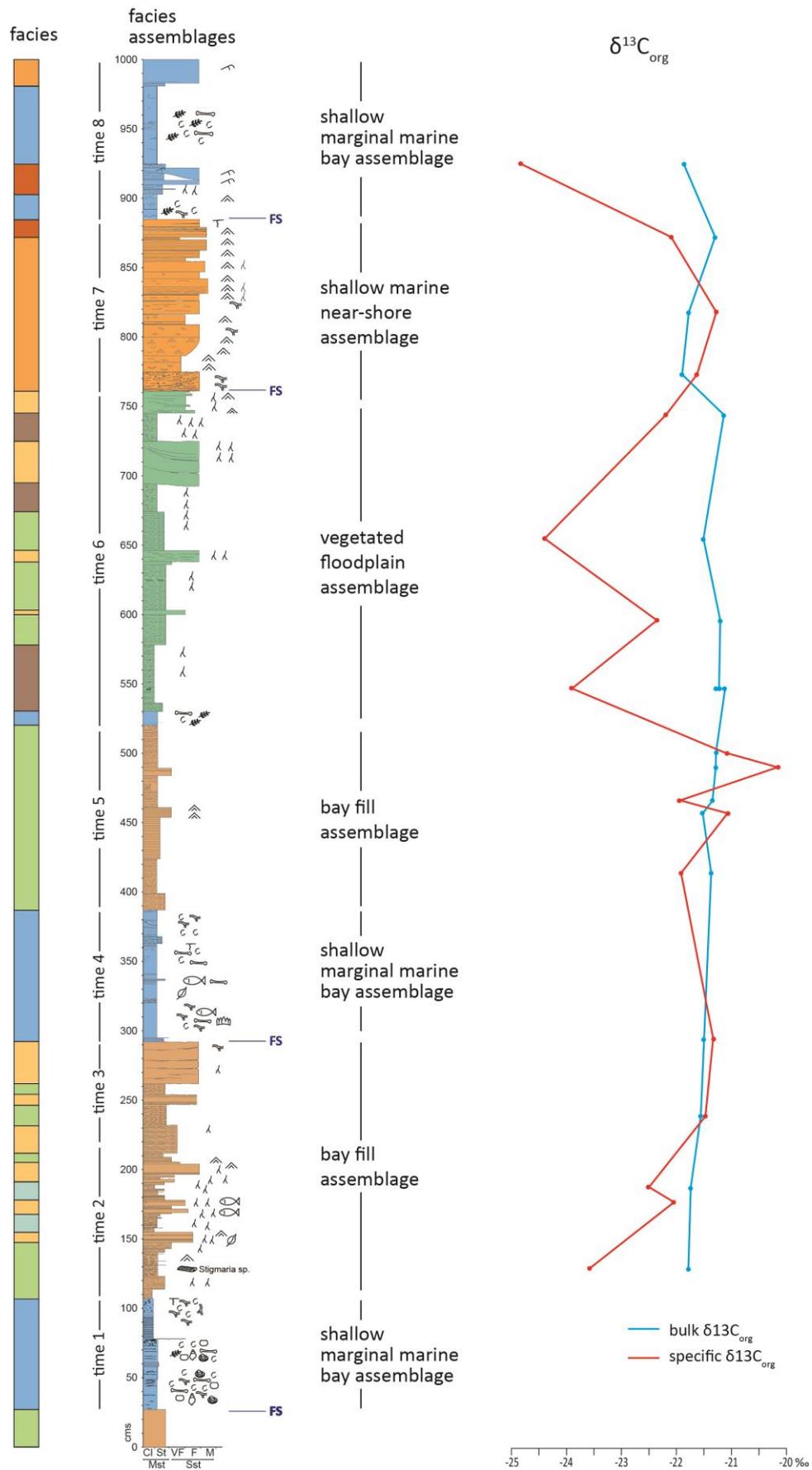


Figure 4.1: Log of Coquetdale section showing facies, facies associations and carbon isotope curves.

4.2 Coquetdale Facies Assemblages

4.2.1 Shallow Marginal Marine Bay Facies Assemblage

Description

This assemblage is almost entirely represented by facies 10 (dolostones), except at the top of the measured section where it includes facies 8 (wave-rippled sandstones) and facies 11 (fine-medium grained sandstones, with hummocky cross-stratification). The dolostones (locally known as 'cementstones') are light to dark grey in colour internally, weathering to a characteristic yellowish colour on exposed surfaces. The dolostones contain some fine-medium-sized siliciclastic grains, and in one example, cm-sized pebbles are identified.

There are three dolostone packages at Coquetdale. Each bed within a package either immediately overlies its predecessor or is separated by very thin (1-5 cm) mudstone/fine siltstone beds. This arrangement of beds appears to be characteristic of the Ballagan Formation at this location. Only rarely are dolostones seen as single beds (e.g. at 520 cm). Although dolostone beds are commonly reported in the literature (Belt *et al*, 1967); (Anderton, 1985); (Andrews *et al*, 1991); (Andrews and Nabi, 1994); (Leeder, 1974) their occurrence as packages, which may be significant, has not been reported.

The lower dolostone package (28-106 cm) commences with a bed containing rounded clasts of micritic dolomite and lies at the base of the first coarsening upward succession (Fig. 4.1). The second package (295-387 cm) underlies a 133 cm thick package of siltstones. The uppermost package and the single dolostone bed that is interbedded with facies 4, differ significantly from the lower packages and will be considered later.

Individual beds differ in their siliciclastic content and structure within the dolostone packages. These dolostones have been divided into two subfacies that are related to the facies defined at Burnmouth (Bennett *et al*, 2016). However, they differ from the

dolostones described from Burnmouth and the Norham core by Bennett et al. (in review) in their internal character and content.

Dolostone sub-facies 1 contains siliciclastic grains, predominantly quartz, and sedimentary structures from the original sandstone or siltstone are visible. In thin section the siliciclastic grains are dominant, dolomite rhombs are present and there is evidence of dolomite fossil recrystallization.

Dolostone sub-facies 2 is thinly bedded, or has soft sediment deformation within siltstone and dolomite units, which are commonly homogenised by bioturbation. In thin section silts are cemented by dolomite rhombs. Fine-grained laminae contain micritic dolomite and fossils are often at least partially dolomitised.

The first and second dolostone packages show similar stacking patterns, with an evolution of bed type from the base to the top of the package indicating decreasing siliciclastic input (Fig. 4.2). This pattern is also seen in a pair of dolostone beds at 885-903 cm (Fig. 4.10E). The lower beds consist of subfacies 1 (Fig. 4.2F-H; Fig. 4.3C&D), with greater proportions of siliciclastic material and all retain more of the original sedimentary structures compared to the subsequent beds. These are overlain by subfacies 2 dolostones (Fig. 4.2B-E; Fig. 4.3A-B; Fig. 4.4A; Fig 4.5). As the siliciclastic component decreases in the lowest package the upper beds become finely laminated (Fig. 4.2D-E; Fig. 4.3B), and then intensely bioturbated (Fig. 4.2B-C; Fig. 4.3A). The uppermost dolostone bed in the package at 295-387 cm displays soft sediment deformation with minor bioturbation (Fig. 4.4A) and contains calcispheres and calcifying green algae, including dasyclad sections (Fig. 4.4B-D). The calcispheres are small (100-500 μm), smooth, thin-walled, probably algal cysts.

In the pair of dolostones (885-903 cm) the bed at 885-892 cm contains more siliciclastic material (and bioturbation) compared to the upper bed, with the upper dolostone (892-903 cm) containing more microcrystalline dolomite within laminae. The coarser grained laminae contain ostracod valves, organic material and calcispheres. In the only example from this location, the upper dolostone exhibits

cracks that are infilled with coarser material including ostracod valves and carapaces, calcispheres and organic material, some of which has been pyritised.

The single siliciclastic-rich dolostone at 520-530 cm is a dolomite-cemented sandstone and contains rounded mud clasts, bone fragments, ostracods and dasyclad sections from calcifying green algae (Fig. 4.4E).

The uppermost dolostone package, (885 and 981 cm) is interbedded with erosive-based fine to medium grained sandstone beds displaying hummocky cross stratification and gutter casts (Fig. 4.10E) similar to those reported by Plint, Macquaker and Varban (2012). The uppermost bed (925-981 cm) is a dolomitised sandstone, comprising a mixture of siliciclastic material and microcrystalline dolomite, rich in bone and shell fragments, ostracod valves, and one observed spirorbiform microconchid, much of which is caught up in fine-grained dolomite strands and wisps and has likely been transported. Bioturbation is common.

Slightly undulose bases and tops are common in the dolostones (e.g. at 520 cm). Some beds appear to load into the underlying mudstones/siltstones. Most upper bed surfaces are fairly flat but may be undulose, perhaps due to differential compaction (e.g. Fig. 4.5). Occasionally thin, silty dolomite-cemented beds containing abundant faunal fragments are found on top of thicker dolostones, and appear to be genetically linked to the underlying dolostone bed. A thin bed at 334 cm contains framboidal pyrite and abundant faunal fragments including bone fragments, rhizodont scales, eurypterid cuticle and an actinopterygian jaw.

The best example of fine lamination occurs in the lowest package (Fig. 4.2D-E; Fig. 4.3B). This bed contains some post-depositional fracturing and bioturbation.

Most of the dolostones are bioturbated, sometimes intensively. The top bed of the first dolostone package has a particularly diverse ichnofauna, including *Teichichnus* (Fig. 4.2B-C; Fig. 4.3A). Beds lacking bioturbation tend to be dominated by siliciclastic material, e.g. the thin (1 cm), fossil-rich bed at 334 cm and the single dolostone bed at 520-530 cm.

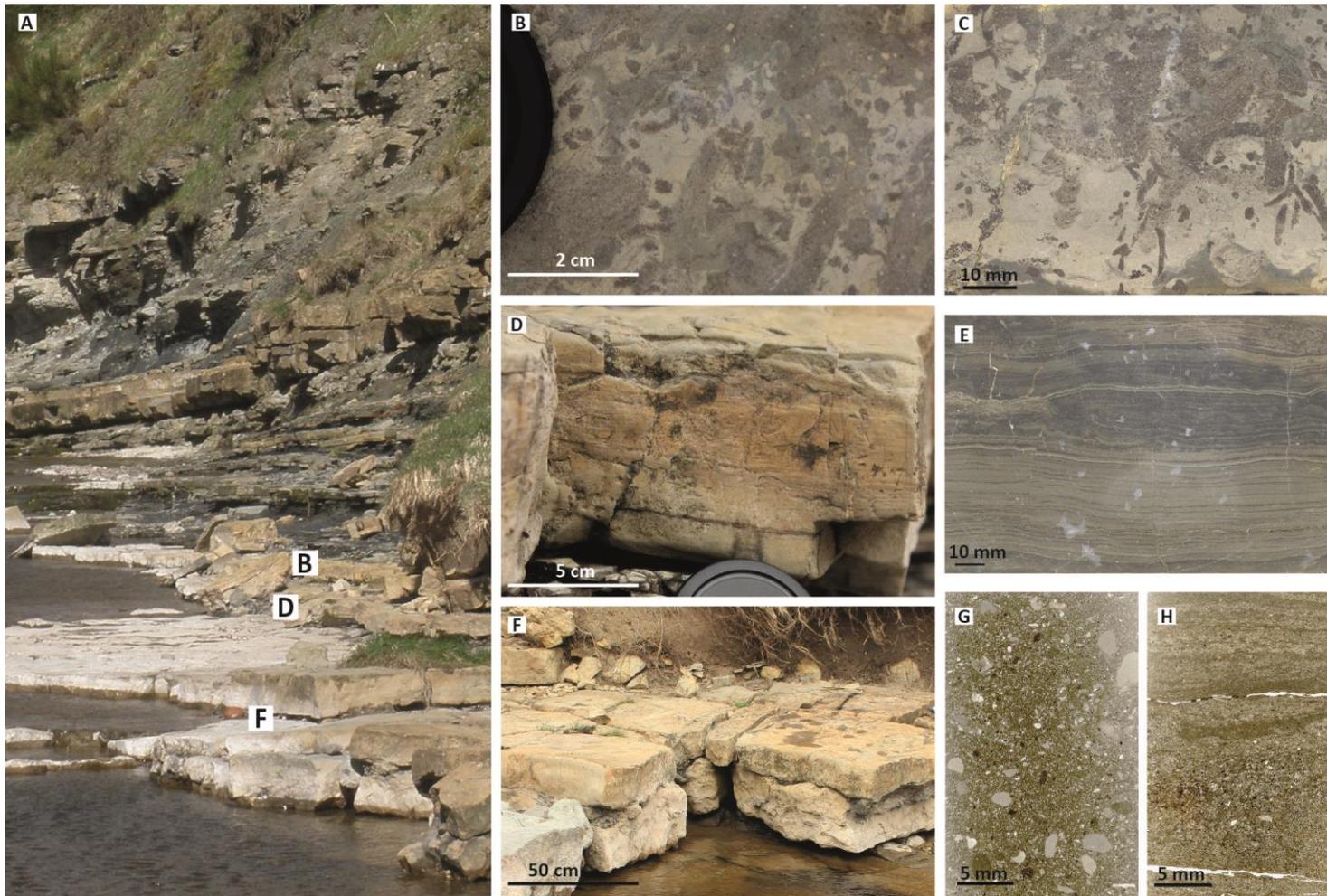


Figure 4.2: Dolostone package development. A: view upstream, towards left bank; foreground is 3 m across. Showing the development of the first dolostone package from subfacies 1 (A - bed F; F-H) to subfacies 2 (A - bed D; D-E; A - bed B; B-C). Lowest bed contains rounded clasts and indicates ravinement processes; upper beds show increasing lamination with diverse bioturbation at the top, indicating quiet conditions and time for the development of an infauna before deposition of the overlying bay-fill assemblage.

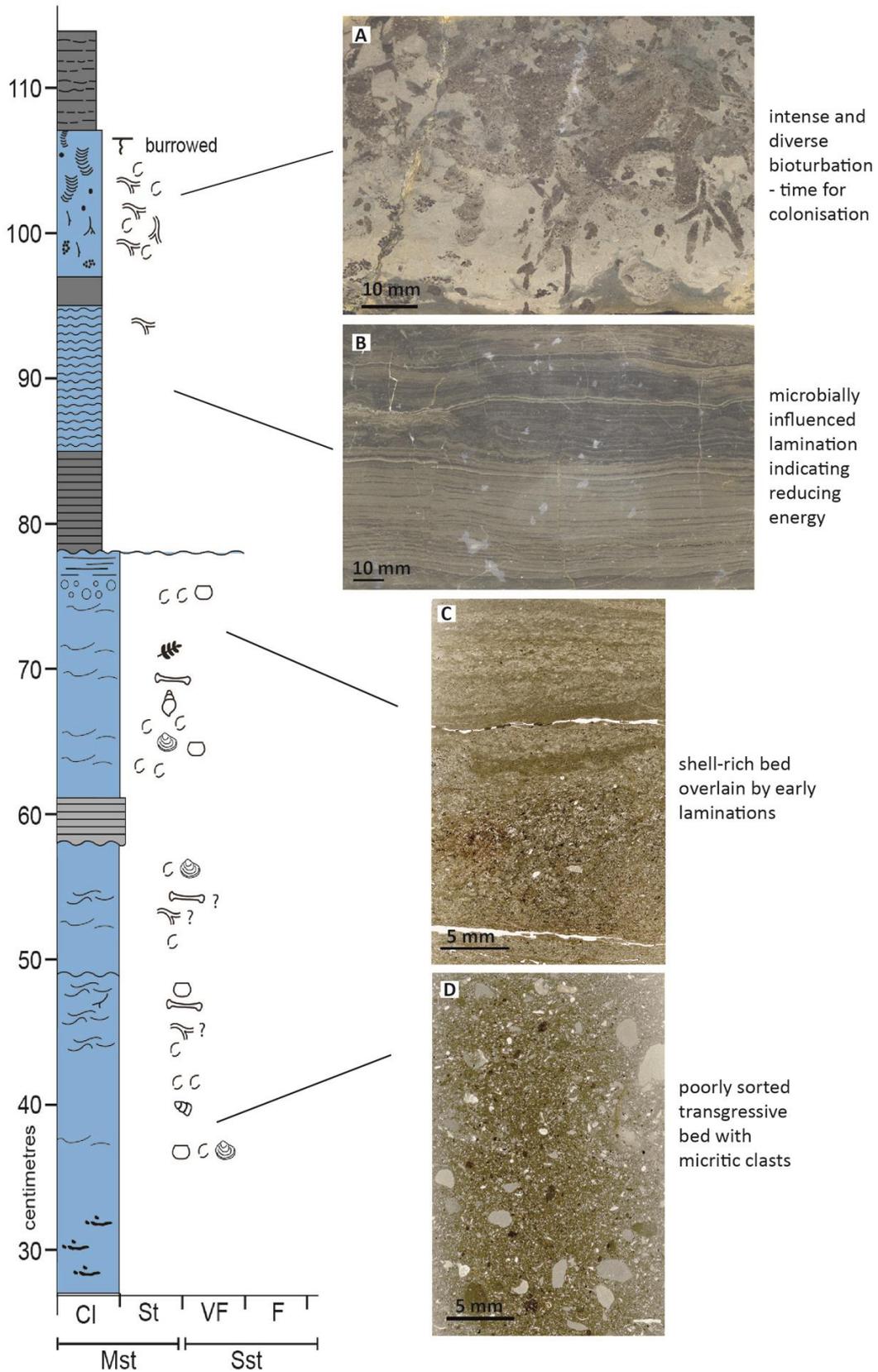


Figure 4.3: Detail of dolostone package development. Lower beds (D & C, subfacies 1) contain more siliciclastic material and retain more of the original sedimentary structures. Upper beds (B & A, subfacies 2) have decreasing siliciclastic content, become finely laminated and then intensely bioturbated as conditions allow.

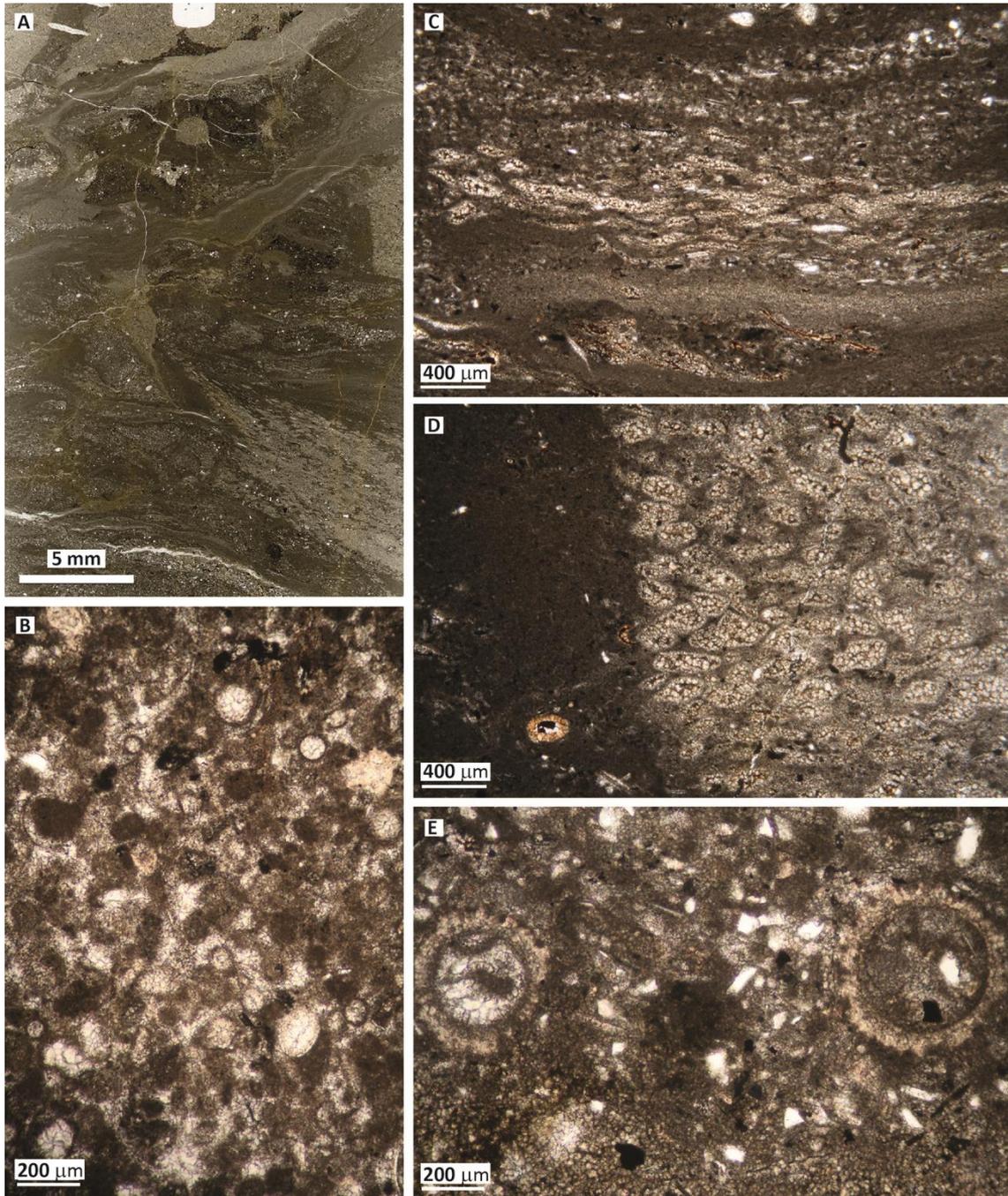


Figure 4.4: Internal details in dolostone subfacies 2.

A. dolostone subfacies 2 showing soft sediment deformation, bioturbation and calcifying green algae (lower right); B. abundance of round calcispheres; C-D. calcifying green algae found in upper beds in the second dolostone package; E. dasycladacean sections found at the top of the second package of dolostones and in the single dolostone bed (520-530 cm).

A bed at 355cm with limited internal bioturbation, but with obvious bioturbation on the top surface, is internally brecciated, a feature noted only once in the dolostones.

All dolostones are fossil-rich; a qualitative analysis of fossil type and occurrence reveals differences between dolostones and throughout the section (Fig. 4.6). Fossil material includes eurypterid cuticle; an actinopterygian jaw and, separately, scales; a bradyodont toothplate; rhizodont scales; occasional bivalves (filaments in nodule), and escape traces; 'spirorbids'; ostracods and the trace fossil *Teichichnus*. Bone and scale fragments are found in almost all samples but are particularly abundant in the second package (295-387 cm). The fish remains identified are from this package, but well-preserved remains have been observed on many dolostone surfaces. Ostracods are common although their preservation is too poor to enable identification; bivalve shells are sometimes observed and 'shelly material' other than ostracod valves is commonly present. Plant/organic material is common, particularly towards the base of the section. Calcispheres first appear in the uppermost bed of the second package (380 cm), (Fig. 4.4B; Fig. 4.6), in the same sample as calcifying green algae (Fig. 4.4C-D). Calcispheres and dasyclad sections (Fig. 4.4E) are both present in the single dolostone at 520-530 cm and the fractured dolostone at 900 cm.

In many of the fossiliferous beds dolomitisation appears to have occurred as a diagenetic alteration since many of the shelly fragments are either completely dolomitised or retain some of their original calcite, with dolomite overgrowing the edges.

The upper surface of the second package (387 cm) has several nodules of varying shapes and sizes, typically 15-20 cm across, on its surface (Fig. 4.5A). A cut and polished surface reveals a layer of fine-grained matrix with bivalve filaments and ostracods with a gradational transition into a layer with rounded and, in some examples, bored clasts, angular clasts of smaller lithic clasts, abundant ostracods of all sizes, bivalve filaments, and bone and scale fragments (Fig. 4.5B).

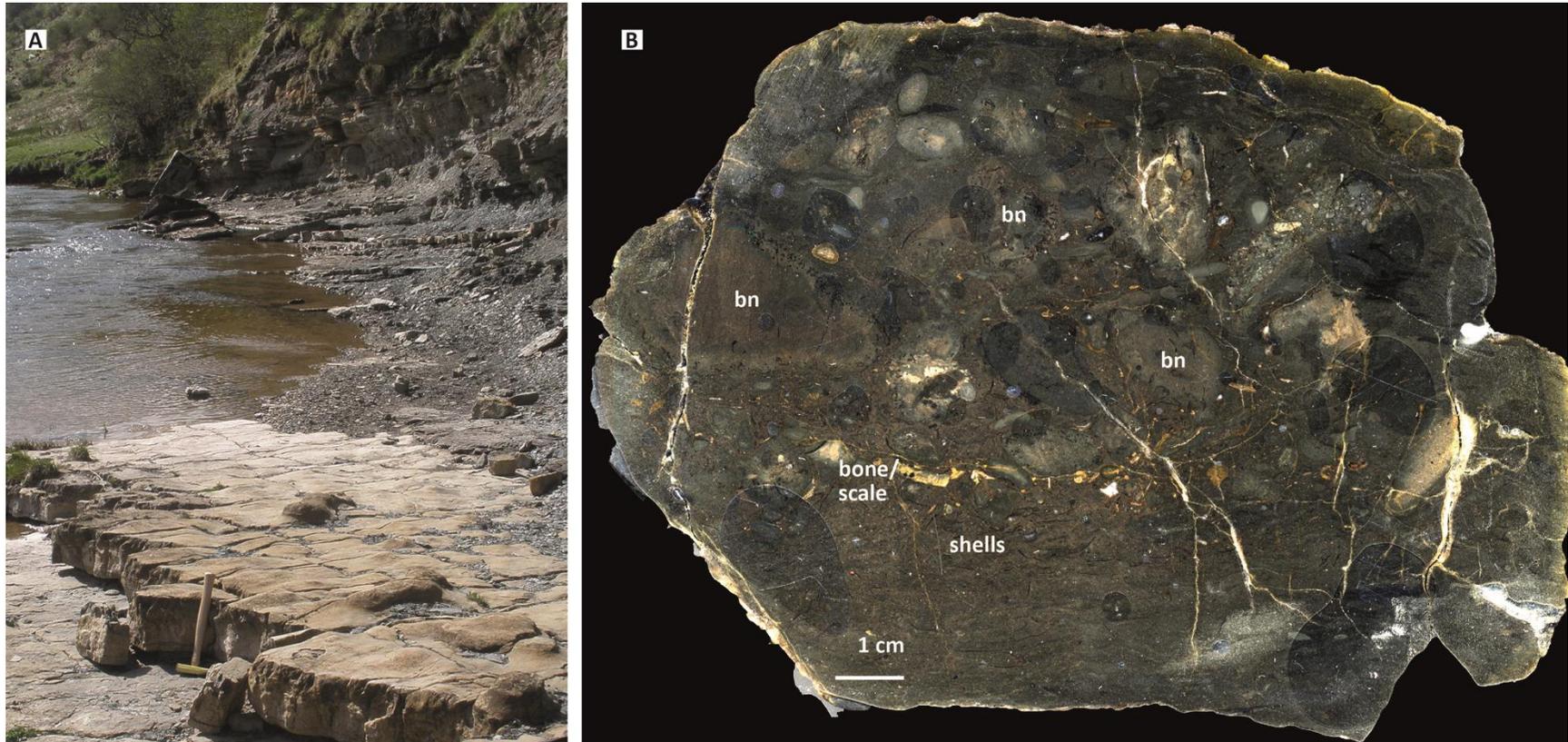


Figure 4.5: Nodular surface of a dolostone package. Left bank of River Coquet, view upstream (west).
Top of second dolostone package showing nodules in situ (5A) and polished cut section of nodule (5B); note rounded, bored clasts (bn) and transported and concentrated fossil fragments.

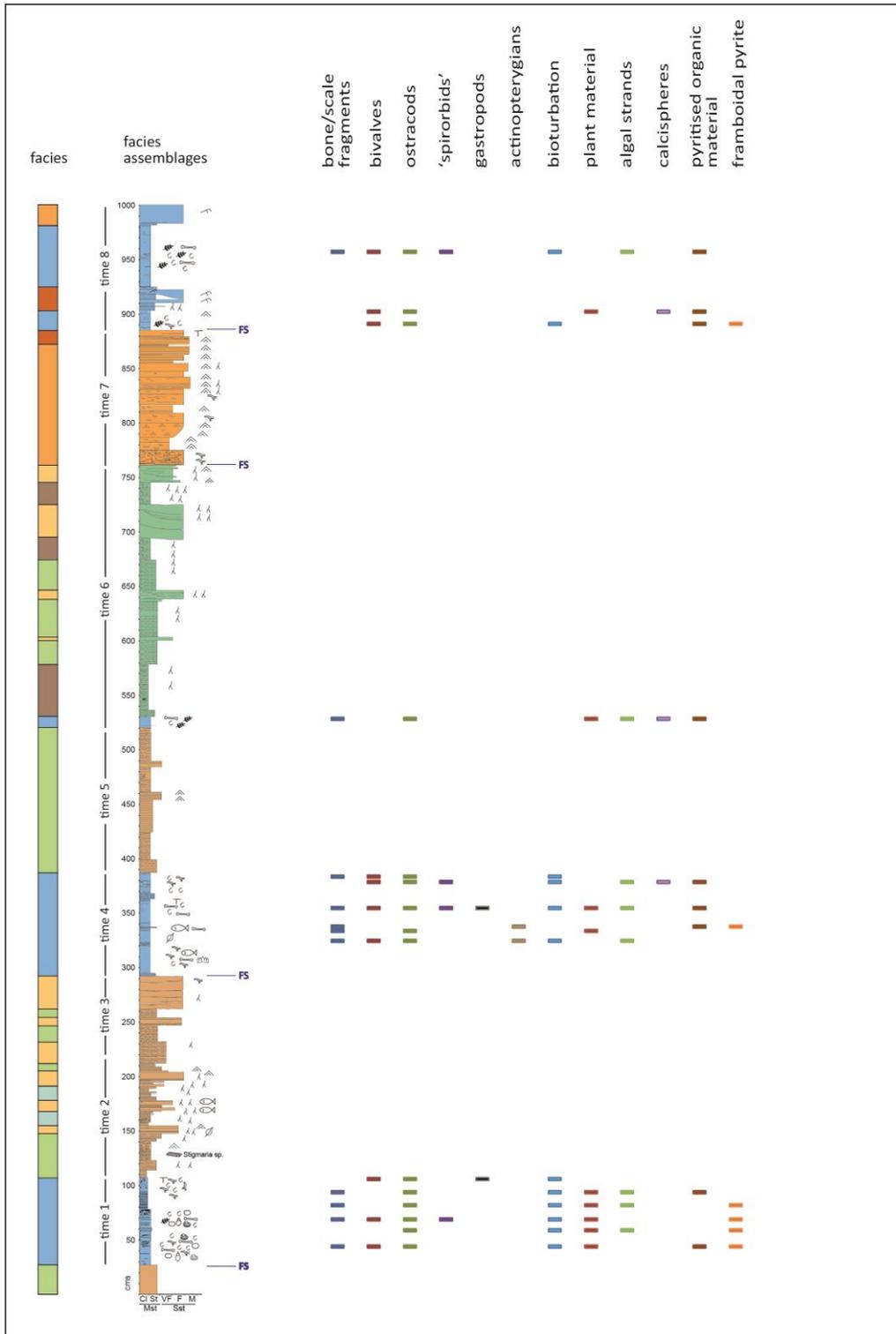


Figure 4.6: Fossil and pyrite occurrence in dolostone packages. Occurrence of fossil types and framboidal pyrite in the dolostones. Note that framboidal pyrite signalling marine deposition is most common in the lower dolostones and in the lower beds of packages, and that calcispheres (marginal marine, shallow water) do not appear until the top of the second package of dolostones.

Interpretation

Dolostones are common in, and characteristic of, the Ballagan Formation (Belt *et al*, 1967; Leeder, 1974; Browne, 1980; Browne *et al*, 1999; Andrews *et al*, 1991).

In Coquetdale the base of each dolostone package is interpreted to represent a marine transgression. The rounded, transported clasts found at the base of the first package were generated by ravinement during this transgressive event. The relatively low siliciclastic content of these beds, and the thickness of individual beds suggests that carbonate production in the water column was high during formation. The primary calcium carbonate was later dolomitised with the necessary magnesium supplied from sea water and sulphate reduction then allowing the precipitation of dolomite (Eugster and Hardie, 1975). Dolostones are represented by times 1, 4 and 8 on the palaeogeography (Fig. 4.11).

The variable siliciclastic content was controlled by the proximity of the shoreline and the amount of energy in the system, the common plant and organic material towards the base of the section corresponding with greater siliciclastic input and fluvial input. Transgression, and a receding shoreline gave rise to the stacking pattern seen in the dolostone packages. Fine-grained dolomite (or fine-grained calcium carbonate micrite, later dolomitised) formed an increasing part of the mix in the upper beds as siliciclastic input was reduced. The stacked beds of the packages evolved from the initial siliciclastic-rich beds deposited during the early phase of a transgression to the fine-grained dolomite/calcite-dominated upper beds as the shoreline receded. This transition is also associated with the presence of lamination, bioturbation and more soft-sediment deformation, the early-formed dolomite initially forming as fine-grained, unlithified sediment prone to deformation by mixing. Soft sediment deformation is particularly intense in the top bed of the second package (295-387 cm), (Fig.4.4A). The uppermost beds of the packages suggest shallow water, sometimes lagoonal conditions, with calcifying green algae present nearby.

Microbial lamination and the abundant and diverse bioturbation observed in the upper beds of packages coincides with a reduction in siliciclastic input, a reduced rate

of deposition, and an environment conducive to longer-term colonisation such that microbial mat formation and/or long-term bioturbation became possible.

The best example of fine lamination occurs in the lowest package (Fig. 4.2D&E; Fig. 4.3B). It is suggested that the laminated sediments were bound by microbial veneers, or mats, and indicate a very quiet environment (Martin Pedley pers. comm.; Schieber, 1998; Schieber, Southard and Thaisen, 2007). They are similar to the biolaminites found in shallow water and lagoonal facies, or the planar stromatolites of Slowakiewicz and Mikolajewski (2011). The limited bioturbation in this bed suggests that conditions at the time of deposition were probably less conducive to infaunal colonisation, perhaps due to lower oxygen or altering salinity, lamination thus being retained. Similar laminated sediments in the Permian have been interpreted to suggest that salinity was increased, based on the limited variety of the fauna within (Slowakiewicz and Mikolajewski, 2011), but the brackish/marginal marine nature of the flora and fauna present suggest that this is not the case here. Disruption of the laminae by bioturbating organisms must have occurred when the mat-bound sediment was competent enough to support the burrowing, and yet still soft enough for some deformation. Occasional post-depositional fracturing cross-cuts the lamination.

Coarser grained parts of the dolostones, and some of the fine-grained muddy parts, with their mix of calcite and dolomite, likely formed as limestones and were then subject to post-depositional dolomitisation (Tucker and Wright, 1990; Al-Awadi *et al*, 2009) in a similar way to the contemporary dolomitised marine limestones of County Armagh in Northern Ireland (Somerville *et al*, 2001). There is much discussion as to the formation of dolomite but Andrews *et al* (1987) maintain that even where dolomite appears to have precipitated as the original carbonate form there is often a calcium carbonate precursor. It is possible that some of the finest grained fractions (microcrystalline dolomite, and possibly where dolomite rhombs have formed in clays) may have been formed by very early precipitation within the sediment (Vasconcelos *et al*, 2006; Illing *et al*, 1965; McKenzie, 1981; Gregg and Sibley, 1984). Eugster and Hardie (1975) also suggest that some primary dolomite existed in dolomitised mudflats around lake margins. However, today direct precipitation of dolomite is rare, with most dolomite forming as replacement of pre-existing calcite or aragonite (Baker

and Kastner, 1981; Tucker and Wright, 1990; Al-Awadi *et al*, 2009). In the anoxic sludge of the bottom of the Lagoa Vermelha, Brazil, there are three carbonate minerals: calcite, high Mg-calcite and Ca-dolomite (Vasconcelos and McKenzie, 1997), with the dolomite being precipitated only in the dry season when Mg concentrations are raised by evaporation and oxygen levels are reduced, leading to hypersaline conditions. Vasconcelos *et al* (2006) found that high Mg-calcite is forming in the upper, oxygenated layers of microbial mats in the Lagoa Vermelha, whereas in the underlying anoxic parts of the mats dolomite is forming, lending support to the idea that dolomite might (only?) be forming where oxygen is reduced. It may be that as the upper part of the mat is buried by further growth and oxygen levels reduce the Mg-calcite is being diagenetically altered to dolomite, rather than the assumption that the dolomite is forming there as primary precipitate. Vasconcelos and McKenzie (1997) found that sulphate reduction in the anoxic layers promoted the conversion of unstable high Mg-calcite to Ca-dolomite, and then to stoichiometric dolomite with increasing depth in the sediment. This ageing and alteration of metastable Ca-dolomite with depth may be part of the process that generates dolomite.

At Cockburnspath in the Scottish Borders, Andrews *et al* (1991) found a contemporaneous nodular calcrete bed which is now dolomite. This must have been diagenetically altered from calcium carbonate, and although calcretes are associated with arid or semi-arid conditions no trace of evaporites were found here, in common with the sites studied in the Northumberland Basin, suggesting that highly saline waters were not associated with these dolostones and therefore that conditions were more likely to be semi-arid than truly arid.

Whatever the process of formation of dolomite it is clear that all dolomitisation at Coquetdale must have occurred early because faunal fragments remain largely uncrushed and the volume of beds appears to have been maintained.

The body and trace fossils (including spirorbiform microconchids, *Teichichnus*, and the chondrichthyans) indicate that the dolostones at Coquetdale are marginal marine in origin. Conodonts found in the dolostones at Coquetdale several years ago were of the most marginal marine type, and are similar to those found in the brackish-water,

Lower Carboniferous Granton Shrimp Bed of Edinburgh, Scotland (Briggs *et al* 1987), (Mark Purnell, pers. comm.).

Eurypterid material found on the surface of one dolostone bed was probably transported, particularly as eurypterid material is likely to be moulted exuviae. At this time it is likely that eurypterids were confined to freshwater-dominated settings (Lamsdell and Braddy, 2010; Braddy, 2001); although their habitats are still under review. Actinopterygians were found in environments ranging from freshwater to marine (Falcon-Lang *et al*, 2006; Stamberg, 2006; Carpenter *et al*, 2011); rhizodonts were found in fresh to brackish water environments (Carpenter *et al*, 2014) and bradyodont sharks (holocephalans) were marine at this time (Friedman and Sallan, 2012), but may sometimes have ventured into non-marine environments (Tim Smithson, pers. comm.). Spirorbiform microconchids ('spirorbids') are brackish water indicators (Williams *et al*, 2005; Schultze, 2009; Taylor and Vinn, 2006; Flügel, 2010, p 530; Carpenter *et al*, 2011; Carpenter *et al*, 2014). The bivalve filaments were not identifiable and the ostracods were too poorly preserved for identification so could not be used for environmental analysis. *Teichichnus*, found in association with other as yet unidentified trace fossils, is prevalent in shoreface or lagoon/bay environments with brackish water (Pemberton *et al*, 2001).

Calcspheres, calcifying green algae and sections from dasycladacean green algae are found in several dolostone beds and indicate a shallow shelf or lagoonal environment relatively close to shore (Fig. 4.4A-E). Calcspheres (interpreted as being algal cysts) (Fig. 4.4B) were abundant in restricted to semi-restricted lagoonal environments during the Carboniferous (Flügel, 2010; Marszalek, 1975). Today calcifying green algae are found in shallow marine, warm, low-moderate energy environments and tropical lagoonal areas (Flügel, 2010, p 436) and were found in similar environments in the Permian (Slowakiewicz and Mikolajewski, 2011); it is therefore likely that similar conditions prevailed here in the early Carboniferous (Flügel, 2010, p.430). Calcifying green algal fragments (Fig. 4.4C-E) and other fragments seen in thin sections are similar to those attributed to dasycladacean green algae in Flügel (2010), and are characteristic of semi-restricted lagoonal environments. They are remarkably like *Mizzia*, a Permian dasyclad species (Slowakiewicz and Mikolajewski, 2011), and may

also have been present in the early Carboniferous. Calcispheres and dasyclad sections are present in the single dolostone bed at 520-530 cm and in the desiccated bed in the uppermost group of dolostones and suggest a shallow, warm, well-lit, semi-restricted environment close to shore.

The uppermost dolostone group (885-981cm) also contains calcispheres, suggesting a likely position close to the shoreline; the presence of framboidal pyrite dispersed within the sediment confirming these beds as being marine or quasi-marine.

The presence of bioturbation in most dolostones indicates well oxygenated conditions (possibly with the exception of the laminated dolostone bed at 95 cm), with a rate of deposition which allowed sufficient time available for a diverse infauna to develop. The abundant and diverse bioturbation present is characteristic of a marine environment, with the presence of *Teichichnus* defining it as being a shoreface or lagoon/bay environment with brackish water (Pemberton *et al*, 2001). This, in combination with the presence of the calcareous green algae, supports the interpretation that the dolostone beds were deposited in a shallow, well-lit lagoon or bay environment.

The isolated dolostone at 520-530 cm containing rounded mud clasts, bone/scale fragments, algal strands and dasyclad-like green algae is at the top of a bay-fill assemblage, where water depths were likely to have been very shallow, and is overlain by the first palaeosols of the floodplain assemblage. This dolostone bed is interpreted as a brief marine incursion that altered or relocated the siliciclastic source to this part of the basin, and initiated the subsequent floodplain environment.

Fossil-rich nodules on the surface of the uppermost bed of the second package of dolostones (387 cm) are interpreted as transported concentrations of clasts, shelly material and bone/scale fragments (Fig. 4.5B). Many of the clasts within the nodule are rounded and some have been bored indicating that rounding and boring occurred prior to transport to this location. The rounded clasts probably originate from a hardground or shoreline nearby, the abundant bone fragments and shells in the nodule are likely to have been derived from close by because the shells are often

unbroken. Lithification must have been rapid since material is preserved uncrushed. It is possible that these clasts and fossil fragments have been concentrated by flow processes into depressions in the upper surface of the dolostone bed, differential compaction then compressing the soft dolomite-dominated bed below more than the more competent clasts and fossils. The nodule examined is a rich source of fossil material.

Some of the dolostones have thin silty, fossil-rich, dolomite-cemented beds on top, e.g. at 334 cm. Thin siltstones between some of the dolostone beds are likely deposited from plumes of fine material intermittently transported across the shelf. Fossil material, commonly found on the top surfaces of beds, was perhaps transported in these silty plumes. Beds derived from these late surges of material and the silty plumes are the main source of fossil material at Coquetdale.

Beds that lack bioturbation tend to be dominated by siliciclastic input (e.g. the thin, 1 cm, fossil-rich bed at 334 cm and the single dolostone bed at 520-530 cm) or show evidence for syneresis or desiccation processes. A bed at 355cm containing very limited bioturbation internally (but with obvious bioturbation occurring on the top surface) is internally brecciated. Brecciation likely occurred because of shrinkage of the sediment either due to desiccation or syneresis (Plummer and Gostin, 1981), the latter seeming more likely here since these beds appear to have been deposited in water. Syneresis processes bring about volume reduction due to subaqueous or subsurface dewatering in response to salinity changes (Nelson, 1979) and form the three dimensional 'chicken-wire' patterns seen in brecciated beds. Salinity changes involved in brecciation were likely less favourable for infauna, hence the lack of bioturbation evident in these beds. Cracks extending down from a distinct horizon at 900 cm however, are interpreted as desiccation cracks and indicate sub-aerial exposure.

The uppermost dolostone (Fig. 4.1) is interpreted as a diagenetically dolomitised sandstone. The sand appears to have been deposited during one event in places but, laterally the presence of several beds suggests repeated erosion and deposition of sand, perhaps from changing sources as sandy material was transported into and

around the system, sometimes by storm events; faunal fragments have been transported, but still indicate a near-shore position. The rooted upper surface of this bed indicates sub aerial exposure by the time the top of the bed was deposited. The fact that these upper beds are cut by fine-medium grained sandstones displaying hummocky cross-stratification (HCS) and gutter casts indicates stormy conditions at the time of deposition and supports the view that the thick dolostones found in this part of the succession are likely to have been sandstone bodies rapidly deposited into shallow water, following another transgression, which have later been dolomitised.

4.2.2 Bay-fill Facies Assemblage

Description

The bay-fill assemblage begins at 106 cm above the base of the measured section, the mudstones/siltstones and symmetrically-rippled siltstones and sandstones of this assemblage are organised into coarsening upward packages. These interbedded siltstones to fine-grained sandstones typically range from medium dark grey to light grey, with some greenish grey to olive grey beds. The coarser siltstones and sandstones are typically medium grey to light grey in colour and the sandstones in this assemblage are calcite-cemented.

The lower package of siltstones (106-148 cm, facies 4) are mainly dark grey to medium dark grey, weathering to brownish grey to greenish grey with a platy to blocky character. The coarser siltstones are thinly bedded with internal laminae, contain abundant organic matter (particularly on lamina planes), lenses of finer mudstone with roots, and one *in situ Stigmaria* root was observed. Symmetrical ripples first appear towards top of the lower package of siltstones (134 cm).

The overlying micaceous, interbedded mudstones and siltstones, (facies 5) and the fine sandstones of facies 6 (Fig. 4.7, 148-292 cm) are commonly rooted and symmetrically rippled (Fig. 4.7B&D). Siltstone beds, 3-10 cm thick, are laterally continuous with consistent thicknesses; sandstones are more variable in thickness laterally, beds being typically 2.5-20 cm thick, with one bed up to 30 cm in thickness. Framboidal pyrite is present within these fine-grained bay fill sediments.

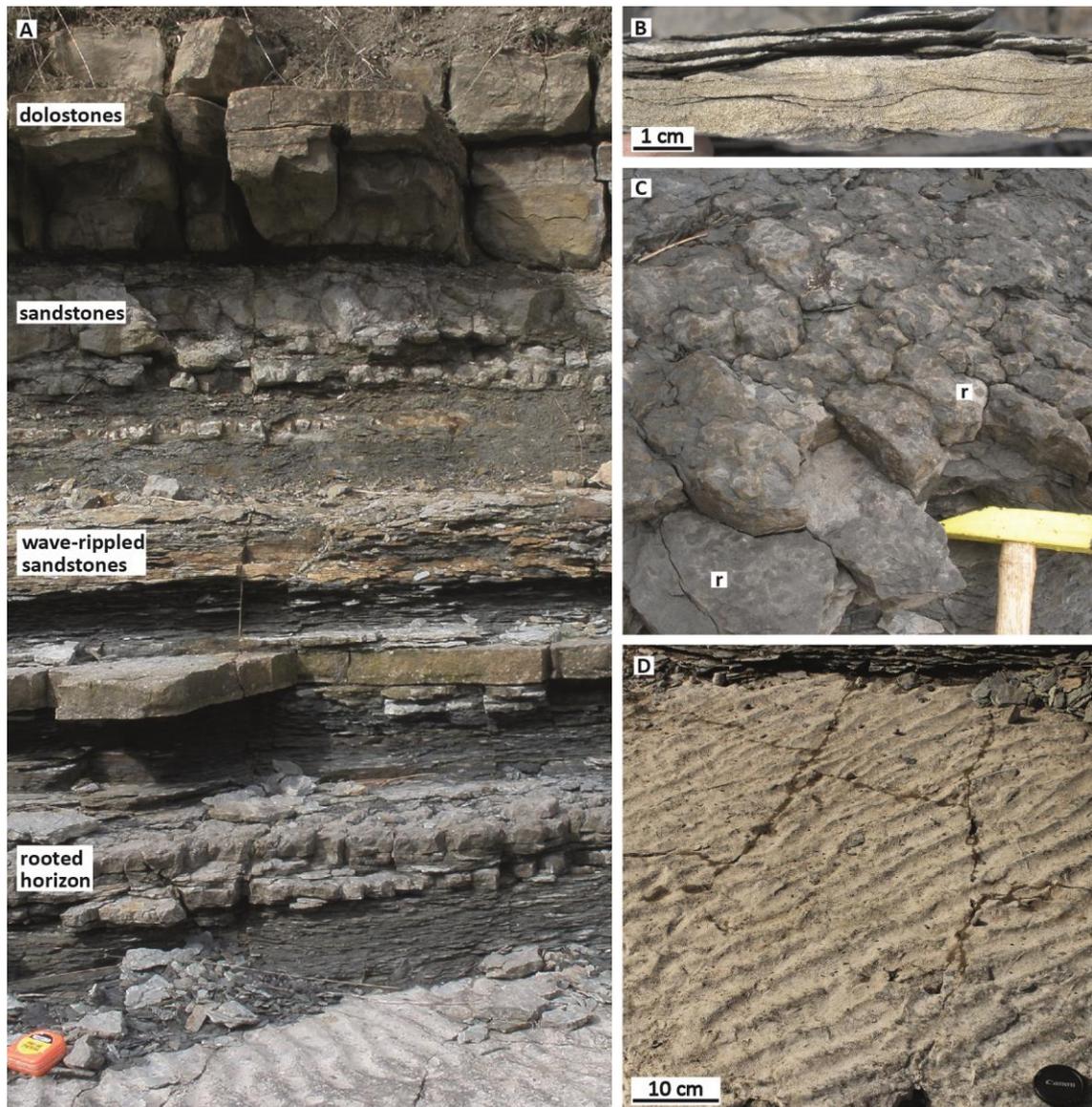


Figure 4.7: Coquetdale bay-fill facies assemblage. Fluctuating water levels of the bay-fill assemblage illustrated by the alternation between symmetrically-rippled sandstones (B & D), rooted siltstones (C, r = roots) and coarser planar-bedded or loaded sandstones (A).

The generally flat-bedded and parallel-laminated siltstones have undulating tops and bases, contain rootlets, occasional *Stigmara* and a single find of an actinopterygian tooth. Some of the siltstone beds display a reddish colouration (Fig. 4.8). The sandstone beds are carbonate-cemented, rooted, and with undulose tops and bases. One bed contains bivalve escape traces. Roots are occasionally up to 25 cm deep. One particular bed is intensely rooted and has an undulose, desiccated upper surface (Fig. 4.7C). A 30 cm sandstone (262-292 cm) is variable in thickness along its lateral extent,

comprises several individual beds, contains abundant organic matter, including plant material, and has an intensively bioturbated upper surface. Occasional shark spines (gyracanth) and a cladodont-type fish (chondrichthyan) tooth have been found in the fine-grained sandstones (Smithson pers. comm., 2016).

Rocks of the bay-fill assemblage-type recur at 387-520 cm (Fig. 4.1; T5 of Fig. 4.11), range from very fine to coarse siltstones, with symmetrical ripples in discrete beds of coarser siltstone (Fig. 4.7B), and vary in colour from medium dark grey to medium light grey, with the coarser siltstones being a light olive grey.

Interpretation

The initiation of bay-fill deposition is the point at which the sediment supply exceeded the effect of transgression or subsidence, and the point where siliciclastic input overwhelmed carbonate production. The intercalation of wave-rippled surfaces and horizons with roots seen in the bay-fill assemblage indicates fluctuations between shallow water and sub-aerial exposure.

Few fossils apart from roots and *Stigmaria* have been found in the bay-fill sediments but this may be a collecting artefact; the one cladodont-type tooth in this assemblage suggests at least intermittent marine influence (Sequeira and Coates, 2000).

Gyracanthus inhabited brackish to marginal marine settings (Carpenter *et al*, 2014; Falcon-Lang *et al*, 2006; Turner, Burrow and Warren, 2005) and actinopterygians occurred in environments ranging from freshwater to marine (Falcon-Lang *et al*, 2006; Stamberg, 2006; Carpenter *et al*, 2011). These fossils suggest a shallow water, marginal marine environment; the framboidal pyrite in the bay-fill sediments and in the dolostone packages above and below supporting the marine influence (e.g. Taylor and Macquaker, 2000; Ferreira *et al*, 2015). Fragments of arthropod cuticle and plant matter on bedding planes suggest a position not too far from the shoreline. The fossils in the bay-fill assemblage suggest fluctuating salinities of this part of the section with the overall coarsening upward profile from mudstone to fine to medium-grained sandstone indicating a gradual infilling of a bay or shallow shelf.

The mudstones and laminated fine siltstones of this assemblage represent deposition in a standing body of water (T2 & T5 of Fig. 4.11). Symmetrical ripples observed in the intercalated coarser siltstones and sandstones indicate a shallow environment above wave base. The repeated coarsening-upward packages (123-204 cm) are interpreted as shallowing cycles as sediment was supplied to the bay representing progradation of a nearby sediment source and variations in water depth. Once a period of deposition finished, the sediment compacted and finer deposition resumed before the next input of coarser sediment. Each of the coarser, wave-rippled siltstones or sandstones represents the upper part of small shallowing-upward cycles, and the trend for increasing proportions of coarse siltstone and sandstone beds upwards through the bay-fill assemblage represents the gradual filling of the bay although water level may have been fluctuating through time.

Bivalve escape traces in a sandstone bed (250 cm) and the increase in grain size from silt to fine sandstone suggest a sudden influx of coarser material, and that periodically there was rapid deposition (T3 of Fig. 4.11). The presence of bivalves also indicates that the environment present prior to this event was sufficiently well oxygenated to support macrofauna.

The abundant roots (Fig. 4.7C; 4.8B), including occasional thin deep-reaching roots (up to 25 cm), and associated fractures along with desiccation cracks seen at 177 cm (Fig. 4.7C) and the red colouration which is probably the result of early palaeosol formation (Fig. 4.8A) (Kearsey *et al.*, 2016) suggest a varied vegetation growing in a marginal bay-edge setting in very shallow water with at least occasional subaerial exposure. The presence of three dimensional plant pieces (Fig. 4.8C, stem?), roots of various sizes, *Stigmaria* (Fig. 4.8D), and possible lycopod leaves, indicate that established communities of vegetation grew nearby.

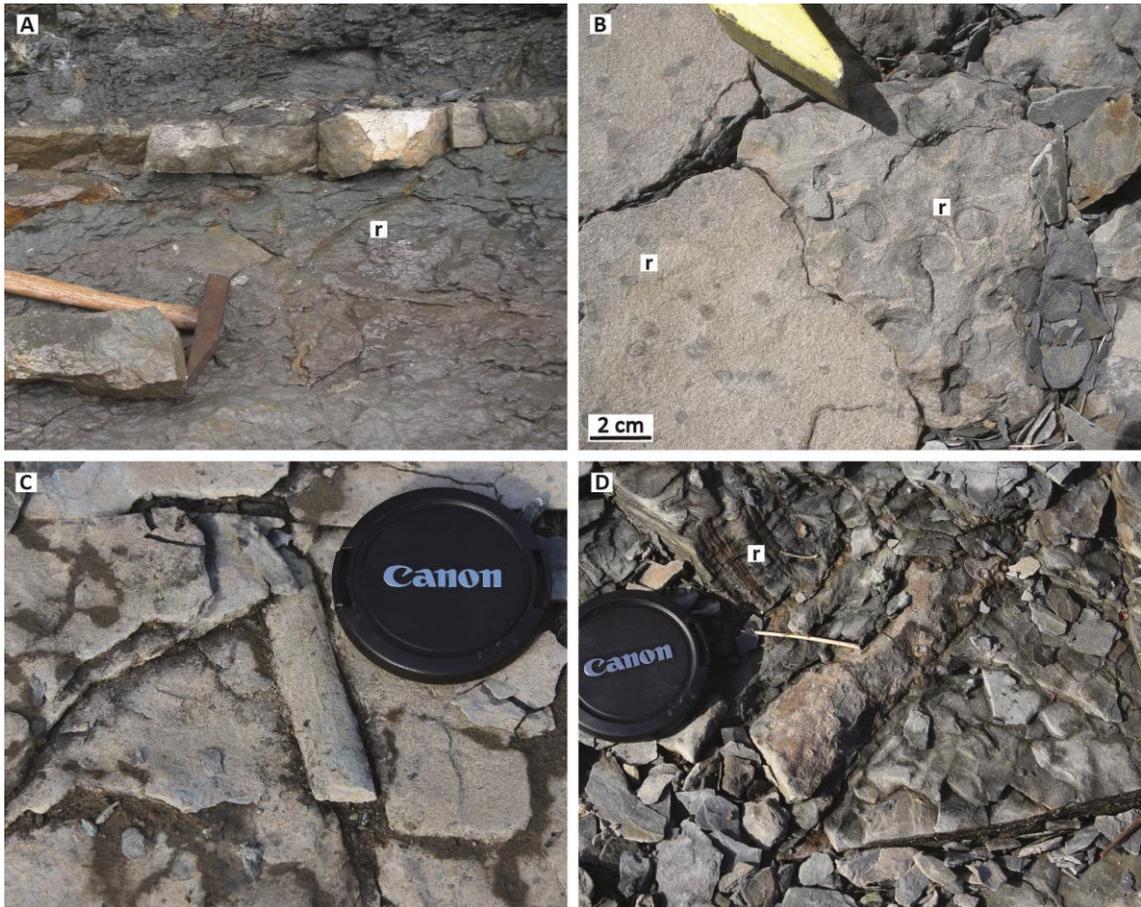


Figure 4.8: Reddened horizons and roots in the bay-fill facies assemblage. A. reddened horizon with roots (r); B. rooted sandstone; C. ?stem, 3D preservation indicating early lithification; D. *Stigmaria* in situ, also preserved in 3D. Lens cover is 6 cm diameter.

Roots are less common higher in the bay-fill succession as sandstones become more dominant, suggesting that the beds were deposited more rapidly as the bay filled, so that vegetation was rarely established. In addition, the top of the thick sandstone that caps the bay-fill assemblage (262 – 292 cm), is bioturbated, suggesting increasing water depths and time for colonisation. This bioturbated top to the thick sandstone marks a switch in the local regime from bay-fill back to the shallow marginal marine bay assemblage (dolostones), the coarser-grained sediment and the repeated depositional events indicating increased energy in, and diversion of, the system. Subsequently siliciclastic input was greatly reduced and a transgression resulted, although the effect of a local sea level rise cannot be dismissed.

4.2.3 Vegetated Floodplain Facies Assemblage

Description

The vegetated floodplain assemblage comprises laminated, sometimes symmetrically or asymmetrically rippled siltstones (facies 4), very fine- to fine-grained sandstones that are sometimes cross-bedded and sometimes rooted (facies 6), and palaeosols (facies 9). The assemblage occurs between 530 and 761 cm (Fig. 4.1).

Roots are common in this assemblage and are present at the top of beds of all lithologies (mudstones, siltstones and sandstones). Roots become more abundant upward, and exposures have a broken, 'hackly' texture that is characteristic of palaeosols at this location (Fig. 4.9A). Visible roots are three dimensional, up to 0.75 cm thick and penetrate up to 20 cm. The uppermost rooted bed, at 725-745 cm, is particularly densely rooted, although roots extend only a few centimetres into the underlying lithology.

The fine-grained, laminated siltstones at 536-578 cm (facies 9) are medium dark grey to dark greenish grey with patches of a moderate reddish brown colour and contain abundant comminuted organic material and occasional fragments of arthropod cuticle on partings. The upper part of this bed is rooted.

Facies 4 siltstones from 578-674 cm are neither laminated nor symmetrically-rippled. Some have sharp erosive bases (590 cm, Fig. 4.9B) and some remnant lamination, but are largely structureless. Multi-directional ripple foresets are identified in the micaceous siltstones (603-636 cm, Fig. 4.9C). There is no parallel lamination. The upper part of this bed contains evidence of some rooting; roots are deep red and occur occasionally, but throughout.

The siltstones are cut by micaceous scour-based fine-medium sandstones (facies 6) that are up to 30 cm thick in places and display crossbedding. There are several sub-beds within a scour (Fig. 4.9E) and scour bases are bioturbated. The bed tops are rooted and each has thin lateral extensions which are more densely, but shallowly (cms) rooted (Fig. 4.9D).

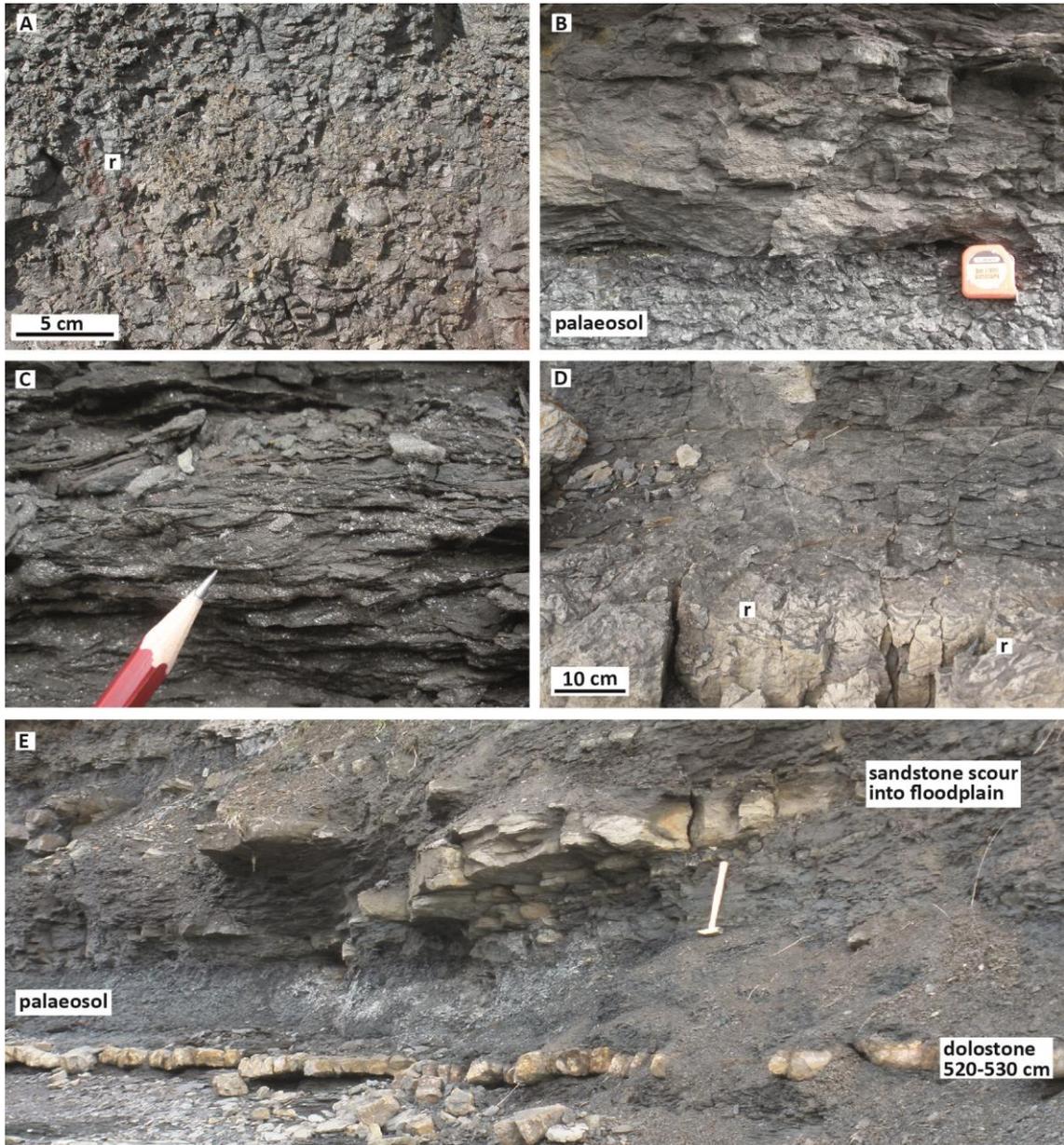


Figure 4.9: Vegetated floodplain facies assemblage.

A. 'hackly' texture of palaeosol, with red roots; B. siltstone with sharp erosive base eroding into the underlying palaeosol; C. ripple foresets seen in some of the siltstones from intermittent overbank flows; D. rooted (r) overbank sandstones, lateral extensions of sandstone scours seen in E.

Towards the top of this assemblage the first symmetrical ripples appear. Ripples are mud-draped. This uppermost bed of the assemblage contains both symmetrical and asymmetrical, ripples (Fig. 4.10A).

Interpretation

This assemblage represents a floodplain environment with increasingly pervasive vegetation (T6 of Fig. 4.11). Roots and the absence of the wave rippled surfaces, that are so common elsewhere in the section, indicate sub-aerial conditions.

Beds with roots represent palaeosols; these are thick but poorly developed and are interpreted as entisols, retaining some of their original bedding (Retallack, 1988, 2001). Entisols lack pedogenic horizons, indicating that they formed quickly in areas of active erosion or deposition, including floodplains (Retallack, 2001; Kearsley *et al*, 2016). The broken 'hackly' texture of the palaeosols indicates that roots have caused considerable disturbance of any original lamination in the finer-grained packages. Roots are increasingly common up through the assemblage suggesting an increasingly vegetated floodplain.

Asymmetrical ripple foresets observed in some floodplain siltstones are probably derived from overbank flows transporting fine-grained material, and suggest intermittent inundation of the floodplain environment from the river systems. The 'patchy' deposition, and multiple flow directions between ripple sets indicate irregular individual flows across the floodplain that filled in local depressions (Fig. 4.9C).

Periodically flows scoured the floodplain, developing small, multiple-fill scours (Fig. 4.9E), with their thin lateral extensions being interpreted as overbank deposits (North and Davidson, 2012). Scours likely occurred at times of increased run-off, perhaps initiated by increased rainfall. Few roots are associated with the scours, limited by repeated depositional phases with little time available for plant colonisation before the next influx of sediment. However, the sand-grade overbank deposits are more intensely, but shallowly rooted (a few cm) (Fig. 4.9D), suggesting floodplain colonisation by opportunistic species. The roots in these overbank deposits are similar to the rootmats documented by Kearsley *et al* (2016). Bioturbation present on the bases of some of the scours indicates that the floodplain environment was well-oxygenated and hospitable to an infauna between coarser-grained incursions.

Towards the top of this assemblage the first symmetrical ripples appear in siltstones and sandstones, indicating wave action in standing water bodies and perhaps more frequent inundation of the floodplain. Ripples are mud-draped, suggesting that there was still a considerable amount of very fine-grained material in the system as well as increasing proportions of sand-grade sediment. The presence of wave ripples and current ripples in the uppermost bed of this assemblage suggests wave reworking of initially current rippled deposits (Fig. 4.10A). This uppermost bed appears to be transitional between the floodplain assemblage and the overlying shallow marine near-shore assemblage with a mixture of processes as the floodplain assemblage is overtaken by the predominantly wave-worked overlying shallow foreshore facies.

4.2.4 Shallow Marine Near-Shore Facies Assemblage

Description

The shallow marine near-shore assemblage comprises planar-bedded, symmetrically-rippled fine to medium grained sandstones from 761-885 cm (Fig. 4.1, facies 8). Beds are 4-13 cm thick, planar bedded sandstones with abundant bivalve escape traces and occasionally either rooted or bioturbated tops. The sandstones are not carbonate-cemented whereas many of those lower in the section have a carbonate cement.

A package of sharp-based, symmetrically-rippled, parallel-bedded micaceous fine to medium sandstone beds at 775-885 cm (Fig. 4.10B) comprises 5-10 cm thick beds with slightly undulating erosional bases. Symmetrical ripples in the lower beds are sometimes mud or silt-draped, particularly towards the base of this assemblage. Upper beds are slightly coarser-grained, cleaner sandstone, and are planar bedded. The topmost beds (874-885 cm) display distinct wave-forms with undulating tops and bases and symmetrically or asymmetrically rippled laminae. Small-scale hummocky cross-stratification (HCS) is observed. These uppermost beds lack the silty laminae of the lower beds.

The lowest part of the assemblage (761-763cm) has an erosive base and contains black organic matter and brick-red bone, scale fragments and an actinopterygian tooth. Organic matter is abundant and typically lies parallel to sub-parallel to bedding. This is

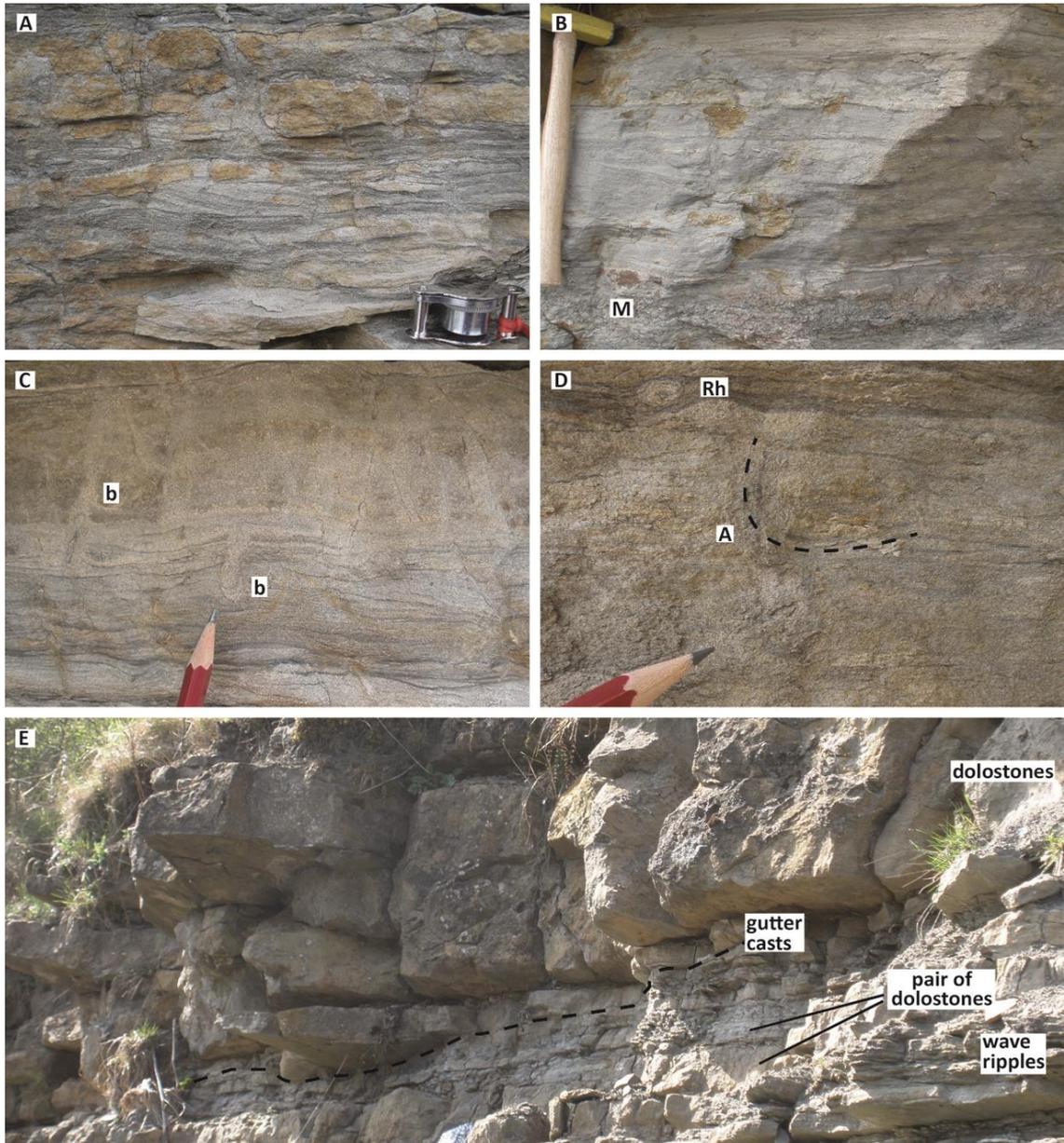


Figure 4.10: Shallow marine near-shore facies assemblage (A-D) and shallow marginal marine bay assemblage (E). A. mud-draped current and wave rippled sandstone at the base of the assemblage; B. sharp-based, wave-rippled, parallel-bedded shoreface sandstones, with intense bioturbation by *Macaronichnus* at the base; C. bivalve escape traces (b) in the wave-rippled sandstones indicating rapid deposition; D. *Arenicolites* (A) and possible *Rhizocorallium* (Rh) in the shoreface sandstones; E. interbedded sandstones and dolostones of the shallow marginal marine bay assemblage, showing gutter casts and overlying thick siliciclastic-rich dolostones.

overlain by a very mottled fine-medium grained sandstone bed (763-775 cm); the mottling is caused by abundant bioturbation. *Macaronichnus*-like burrows are monospecific, filled with clean quartz grains of a similar grain size to the surrounding fabric and have dark micaceous linings.

The sandstones from 775-885 cm are bioturbated. Bivalve escape traces are common (Fig. 4.10C) and there is also one recognisable example of *Arenicolites* (Fig. 4.10D) (Pemberton *et al*, 2001; MacEachern *et al*, 2007), and possible *Rhizocorallium* (Bann and Fielding, 2004; MacEachern *et al*, 2007).

These sandstone beds are overlain by dolostones and beds with intermittent hummocky cross stratification and gutter casts.

Interpretation

The sedimentology and ichnotaxa of this assemblage suggests a nearshore, wave-dominated environment (T7 of Fig. 4.11). The concentration of fossil material (organic matter, bone and scale fragments and an actinopterygian tooth) and the sharp, erosive base to the lowermost bed in this assemblage suggests reworking during a marine transgression of the underlying floodplain. The presence of a transported actinopterygian tooth does not give an indication as to environment. Actinopterygians are found in facies ranging from freshwater to marine, and some species may have been euryhaline (Falcon-Lang *et al*, 2006; Stamberg, 2006; Carpenter *et al*, 2011, 2014).

The intensely bioturbated fine to medium-grained sandstone fabric overlying the basal bed, and the monospecific *Macaronichnus*-like burrows are characteristic of a high energy foreshore to upper shoreface environment (Pemberton *et al*, 2001; Pearson *et al*, 2013; Bann and Fielding, 2004) where sediments are highly oxygenated. The low diversity of the ichnofauna is characteristic of high energy conditions and continuously migrating bedforms, which are challenging for endobenthic organisms to colonise (Pemberton *et al*, 2001). Preservation of the *Macaronichnus*-like burrows in these rocks is due to the deep tier position of the trace-maker (Pemberton *et al*, 2001; Bann and Fielding, 2004). *Macaronichnus* is common in high energy near shore or foreshore settings (Pemberton *et al*, 2001).

A gradual deepening is indicated by the succession from mud-draped wave ripples into cleaner, tabular, planar-bedded, well-sorted fine- to medium-grained sandstones with

wave-ripples, hummocky cross-stratification, bivalve escape traces and occasional vertical burrows.

Sharp, erosive-based beds represent repeated, regular influxes of sediment. Bivalve escape traces, typically found in shallow-water shoreface to foreshore environments and present in the planar-bedded, wave-rippled beds indicate rapid deposition of sediment and well-oxygenated, nutrient-rich conditions. The traces represent equilibrium activity by the trace-maker, as seen in Kane (2010) and Zonneveld and Gingras (2013). Bivalve escape traces and *Skolithos* have also been found in similar shallow high-energy environments of the Lower Triassic of Svalbard (Worsley and Mork, 2001), and in the upper shoreface setting of Raychaudhuri *et al* (1992).

Arenicolites is characteristic of the upper shoreface to foreshore, and the low energy shoreface environment (Pemberton *et al*, 2001). *Arenicolites* and *Skolithos*, likely the burrows of suspension feeders with deeply penetrating domiciles, are preserved in this environment because of their depth. Bann and Fielding (2004) also find both *Arenicolites* and *Skolithos* in their facies C2 'Interbedded massive and laminated sandstone', a high energy wave-dominated environment similar to that found here, and suggest deposition above fair-weather wave-base. The combination of the sedimentological detail and these ichnofossils support the suggestion that these beds were deposited in a foreshore to shoreface environment.

The coarser, upper sandstone beds generally indicate an increase in wave energy and wave-reworking, although occasional beds appear massive at outcrop and are perhaps the result of rapid deposition. These uppermost beds with erosional, undulating wave-rippled tops and bases and hummocky cross-stratification are interpreted to have been deposited above storm wave-base, where deposition is dominated by increased energy and intermittent storm wave processes; these structures being characteristic of shoreface deposits (Reineck and Singh, 1980; Bann and Fielding, 2004; Hampson *et al*, 2008; Plink-Bjorklund, 2008; Plint *et al*, 2012). These deposits are the likely precursors to the increasingly storm-derived sediments of the overlying shallow marginal marine assemblage (Fig. 4.10E, lower right).

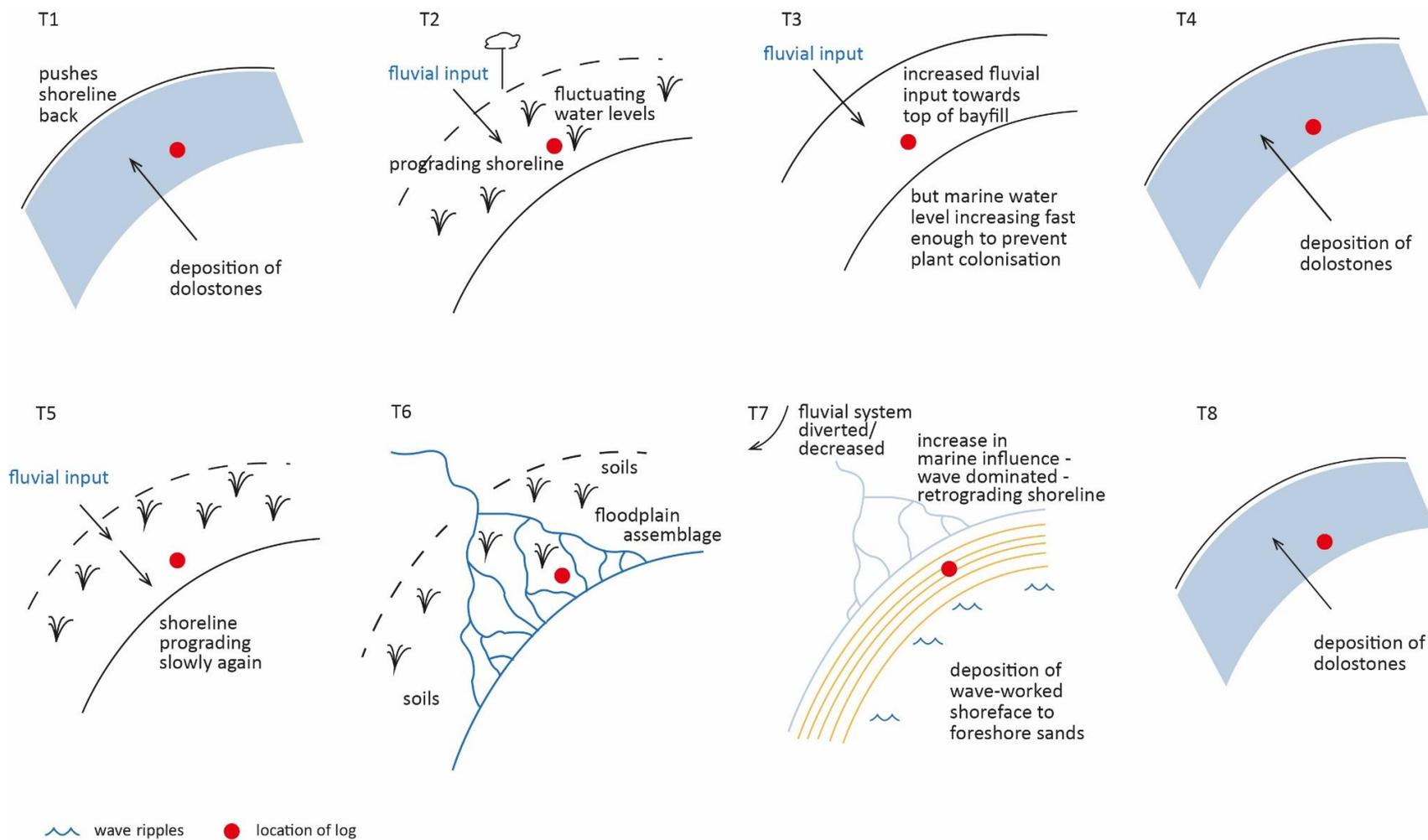


Figure 4.11: A suggested palaeogeographic development of the site at Coquetdale through time. Time 1 to time 8 (T1-T8) are referred to in the text and on the log (Fig. 1). Fossil evidence suggests that shallow water conditions prevailed and that the site was part of a lagoon at least part of the time, with development of a short-lived floodplain on one occasion.

The absence of carbonate cementation in these sandstones may be due to an increased fluvial, freshwater input, or to a more open marine setting where wave-action was more important than at times when the dolostones were being deposited.

This shallow marine near-shore sandstone facies assemblage is overlain by the dolostones, with interbedded hummocky cross stratified sandstones of the shallow marginal marine bay assemblage.

4.3 Isotope analysis

Mudstone and siltstone samples were taken for carbon isotope analysis and were distributed as evenly as possible throughout the logged section (Fig. 4.1). Seventeen samples were analysed for both bulk and specific, picked woody $\delta^{13}\text{C}$ org. values, plus one extra specific sample (see Chapter 2 for sample preparation).

The bulk samples vary between -21.12 and -21.9‰ whereas the specific samples range from -20.14 to -24.84‰ and swing back and forth, many of the values being more negative than their equivalent bulk values. The most negative specific $\delta^{13}\text{C}$ org. values seem to be associated with the most organic-rich bed, with woody material available for picking; these values are contrary to expectations. It is not understood why the specific $\delta^{13}\text{C}$ org. values at Coquetdale should be so variable, or more negative than their equivalent bulk values, but some possibilities are explored in Chapter 6.

4.4 Sedimentological development of the site at Coquetdale

During deposition of the Ballagan Formation, the Northumberland Basin was a marine gulf (Belt *et al*, 1967; Leeder, 1974; Anderton, 1985). The dolostone packages are a key feature at Coquetdale, with packages interpreted as transgressive marine flooding events that repeatedly established marginal marine bays at this location. Dolostone packages are then overlain by bay fill assemblages which developed into short-lived floodplain environments.

Dolostone packages record reducing siliciclastic supply and quieter conditions during deposition, and the fossil content indicates a shallow marginal marine environment. Thin silty dolomite-cemented beds with abundant faunal fragments topping thicker dolostones represent late surges of material into the basin.

It is suggested that the shelf was very shallow-dipping at this time and that a moderate increase in water depth could effectively reduce siliciclastic input by establishing widespread shallow water conditions and retreating shorelines. This shallow water would have been well-lit and warm as during the early Carboniferous the Northumberland Basin was lying about 4° S of the equator (Scotese and McKerrow, 1990). Warm shallow water would have been conducive to algal growth (biolaminites and calcareous algae) and the shallowness of the shelf would also have contributed to the 'restriction' (reduced exchange of marine waters) of the bay. A degree of 'restriction' may have contributed to the 'quieter', less energetic conditions necessary for the formation of biolaminites, and time for colonisation of the sediment. Bioturbation, soft-sediment deformation, calcifying green algae, and nodules in/on the uppermost beds of the dolostone packages indicate proximity to shore and a likely semi-restricted lagoonal setting at this stage in the development of the environment.

Dolostone packages are overlain by bay fill sediments, with framboidal pyrite indicating a marine setting (Raiswell, 1982; Taylor and Macquaker, 2000; Ferreira *et al*, 2015). Following the first package of dolostones the marginal marine bay fill sediments coarsen upward and roots are increasingly extensive, representing the gradual infilling of a shallow marine bay. Conditions fluctuated between a shallow wave-influenced marine bay and a very shallow to emergent silt-dominated floodplain that occasionally dried out to leave desiccation cracks across the vegetated surface.

Bay fill deposition following the second package of dolostones remained fine-grained, suggesting a shallow muddy shoreline with only intermittent coarser input which may have been dependant on inland pluvial events. A brief marine incursion occurred when the solitary dolostone bed at 520-530 cm was deposited; again its fossil content (dasycladacean green algae) indicating proximity to shore. Following this the fine-grained bay fill gradually developed into a floodplain with immature, rooted

palaeosols, dissected by a fluvial system which occasionally over-spilled and formed the shallow-rooted, sandy overbank deposits.

The wave-rippled shoreface and foreshore sediments overlying the floodplain at 761-885 cm indicate a significant change in the depositional regime. A transgressive event established a coarser sand-dominated system with increased wave action indicating that the bay was more open, the planar-bedded, wave-rippled sandstones representing shoreface sands, as in Amorosi *et al* (1999) and the overlying storm beds supporting the suggestion that the bay was more open by that time. Alternatively, the wave-rippled sands may have been part of a barrier island system as in Kieft *et al* (2011), but this seems less likely as there is no evidence of there having been a back-barrier lagoon at this point in the succession. The sedimentology supports the suggestion that the bay was more open at this time but these wave-rippled sands are overlain by dolostone beds. The formation of dolomite seems to occur in more restricted locations so it is likely that by the time the overlying dolostones were being deposited the bay had again become restricted, perhaps as a result of the movement of the sand in the system.

The ichnofauna present in these sandstones confirm a marine setting. As water levels increased the planar bedded, wave-rippled sandstones evolved into undulating, increasingly storm-influenced, sandstones.

Carbonate production in this shallow marine setting at Coquetdale seems to have been possible when siliciclastic input did not overwhelm it. Generally the siliciclastic rocks are not carbonate-cemented although the sandstones of the bay fill assemblage are calcite-cemented. The source of the cement may have been from carbonate production in the water column when the sandstones were deposited, or the calcite cement may have developed much later in the diagenetic history. However, the foreshore/bar sandstones towards the top of the section do not appear to be carbonate-cemented, and yet the dolostone beds just above are. It seems that where siliciclastic input was increased, so the effect of carbonate production in the water column was reduced or eliminated.

Repeated marine transgressions and deepening water allowed for a resumption of carbonate production and the formation of the dolostone beds. In this study the dolostones are interpreted as originating as calcium carbonate-influenced beds which were penecontemporaneously, or later, dolomitised. A possible analogy for the presence of both calcite and dolomite in some beds at Coquetdale is the Dinantian Ardross Limestone of East Fife, Scotland (Searl, 1990) where some beds have been incompletely dolomitised and where it is thought that calcite was stable in dolomitising fluids, with patches of dolomite replacing meta-stable calcite.

The fossil content of the Coquetdale section, supported by the sedimentological evidence, depicts a shallow, marginal marine environment at times when the dolostones were deposited, with fluctuating water levels, and probably salinities, in the intervening bay environments. Some of the fossil material has been transported but the sedimentology suggests that this is not likely to have been far. Fossil fragments settled out of the water column to be preserved on the top of dolostones, were transported into the basin in silty plumes, and were also concentrated at times to form the fossil-rich nodules.

The above data depict the development of the marginal marine site at Coquetdale from a shallow lagoonal environment with intermittent transgressions, to a floodplain, and then to more open shallow-water foreshore environment.

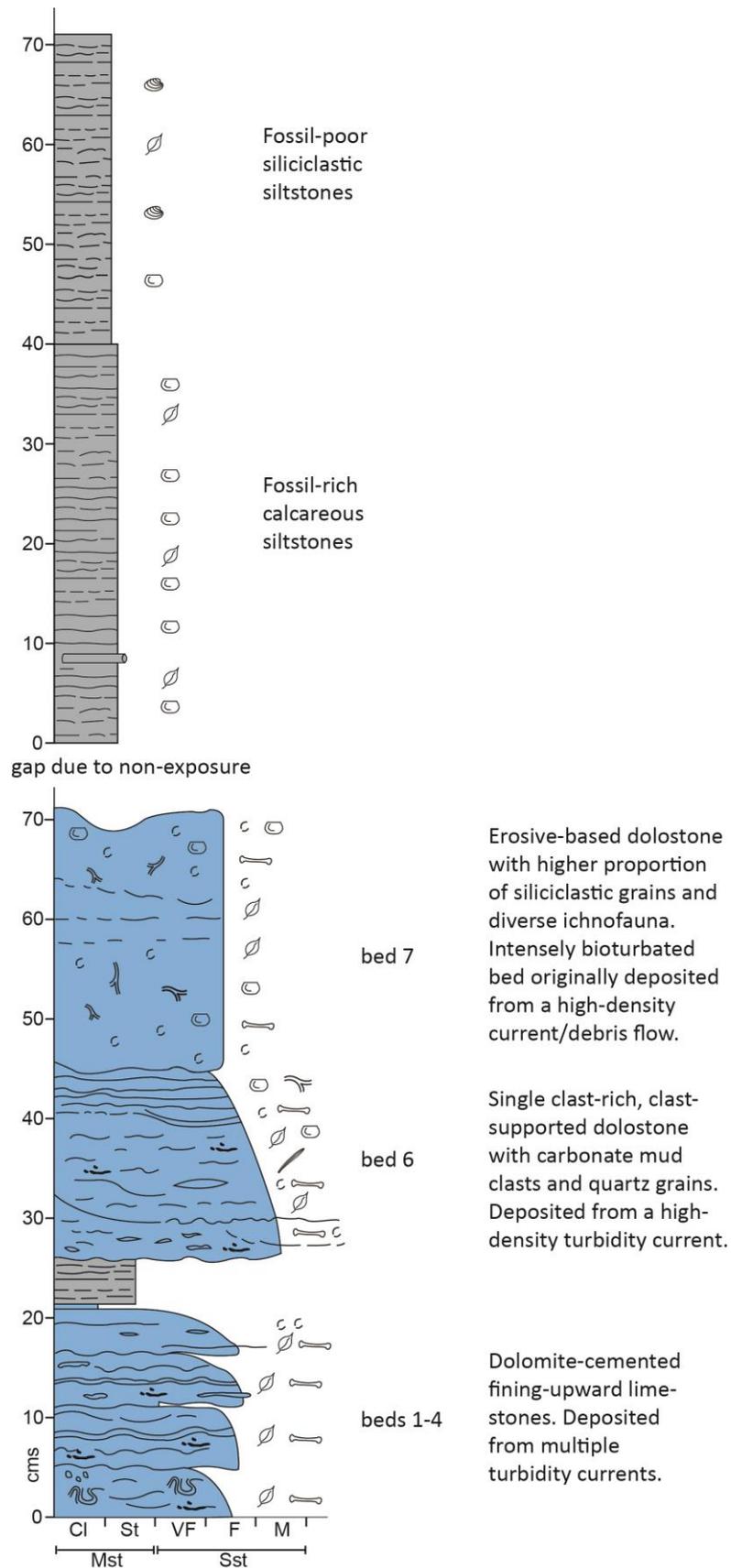


Figure 5.1: Log of Whitrope Burn section. Log showing thin dolomite-cemented beds at the base, the fossil-rich density current deposit, the bioturbated bed and the overlying, but separate, siltstone beds.

The following beds are part of the **shallow, marginal marine bay assemblage**, as defined at Coldstream and Coquetdale (Chapter 3 & 4).

5.2 Lower, thin beds

At the base of the succession is a series of four, thin (5-7 cm) dolomite-cemented, fining-upward limestone beds with undulose laminae (Fig. 5.1, beds 1 – 4; Fig. 5.2A-B). Each bed contains lithic clasts (generally 1-3 mm in size, but, in the lowest bed, up to 2 cm in size) and has a micritic dolomite top. The carbonate mud clasts, some of which are laminated, are suspended in a matrix composed mainly of micritic dolomite with quartz and larger calcite grains (typically fine sand-sized), and some fossil fragments. These beds contain 9.49 – 12.75% non-carbonate material. Dolomite rhombs are seen to be growing on/in calcite in parts of these beds (Fig. 5.3C-F) whereas the dolomitic micrite tops to beds comprise euhedral to subhedral rhombs of dolomite (sizes 2-10 μm) in a clay matrix. Fossil fragments are common, but not abundant, in these lower beds and include plant material (up to 8 mm), shelly and skeletal fragments including occasional incomplete ostracods, a few actinopterygian teeth, and some small bone and scale fragments. Fossil material is more abundant in the base of the topmost bed, and this bed also has a thicker micrite cap.

These thin beds are interpreted as siliciclastic-rich dolomitised limestones. Highly erosive flows transported and concentrated material derived from the margin of the basin. The multiple internal erosional surfaces suggest repeated relatively small surges of slightly coarser material within the main flow. The undulose internal laminae may be the result of soft-sediment deformation and/or deposition onto a very soft substrate deposited by the earlier flow. Uncrushed fossil material indicates that calcite formation and partial replacement by dolomite within the coarser parts of beds, and dolomitisation of the very fine-grained tops of beds must have occurred early.

The thicker dolomitic mud top to the uppermost bed may indicate a prolonged period of settling of the tail end of the flow and dolomite precipitation or cementation before

the deposition of the overlying richly fossiliferous bed and an intervening thin (4 cm) siltstone.

5.3 Fossiliferous Bed

A bed with a particularly rich and diverse assemblage of cartilaginous and bony fish and some tetrapod remains occurs 26 cm above the base of the section. This 20 cm bed (Fig. 5.1, bed 6; Fig. 5.2C) is clast rich and clast supported with a matrix of fine-medium quartz and carbonate mud. In common with the other beds in this section it appears to have been dolomite-cemented, although petrographic analysis has not been completed for this bed. Although larger clasts (0.5-1 cm) typically occur near the bed base, overall the clasts appear unsorted. The diverse clast assemblage includes lithic fragments, plant straps (5-10 cm wide), well preserved vertebrate fossil fragments (including lungfish tooth plates, remains of cartilaginous and bony fish, and some tetrapod fragments), actinopterygian teeth and shelly material (predominantly ostracods and 'spirorbids'). Twenty different chondrichthyan tooth taxa were identified from this one bed, one spine taxon and one dermal denticle taxon (Richards *et al*, in review); this is an unusually rich collection of faunal fragments found within one bed.

The complex internal structure of the bed is characterised by multiple erosional surfaces that are overlain by fine-grained laminae. Each erosive surface is associated with abundant clasts, and concentrations of tooth plates and other fossil material.

This bed is interpreted as a density current deposit where material has been transported, and perhaps concentrated, from the basin margin. The limited sorting suggests deposition from a debris flow, transporting material *en masse* downslope, or from a high-density turbidity current, where the flow concentrated fossil material and mixed with carbonate mud, but did little in the way of sorting. In either case the finer-grained laminae towards the top were probably deposited as part of the waning flow. Large plant straps indicate that the origin of the flow must have been relatively close

to shore, or that the large plant material was temporarily stored on the shelf before being transported into the deeper water and more distal reaches of a lagoon.

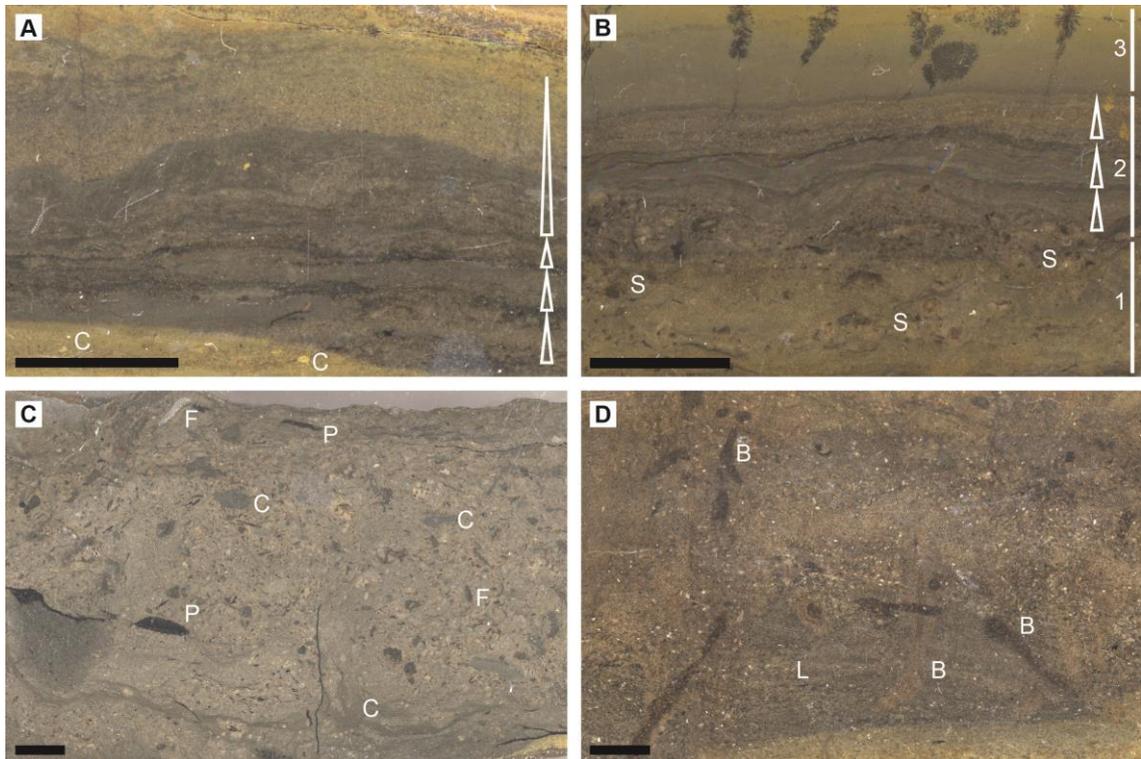


Figure 5.2: Internal structure of beds at Whitrope Burn.

All scale bars are 1 cm. Labels = C - transported lithic clasts; S - shelly fragments; P - plant material; F - fish fragments; B - burrow; L - laminae. A. Lower thin bed showing small fining upward laminae (arrows), transported lithic clasts (C) and the undulating nature of the laminae; B. coarser bed base with shells, occasional ostracods, scale fragments and small plant material, part 1 is mixed and unsorted with bioclastic material, part 2 is laminated, arising from pulsed, coarser input within the flow, part 3 is composed of very fine carbonate material representing the finest part of the flow; C. Bed 6, clast-rich, fossil-rich density current deposit; D. Bed 7, bioturbated, with some remnant lamination.

5.4 Bioturbated bed

Overlying the fossiliferous bed is a bioturbated, dolomite-cemented bed with an erosive base and greater siliciclastic content (29% non-carbonate). XRD analysis shows that this bed contains dolomite, but no calcite. This 25 cm bed (Fig. 5.1, bed 7; Fig. 5.2D) has been intensively bioturbated, with an ichnofauna that includes *Teichichnus*-like burrows (Bromley, 1996; Pemberton *et al*, 2001), but retains some remnant lamination (Fig. 5.2D). Fossil fragments are abundant, particularly ostracods; lungfish

material, bone and scale fragments, and small (1-2 mm) pieces of plant and spore material are also present. Fossil fragments do not appear crushed, and many are pyritised. Cavity-filling kaolinite and distributed barite are observed (Fig. 5.3 A-B).

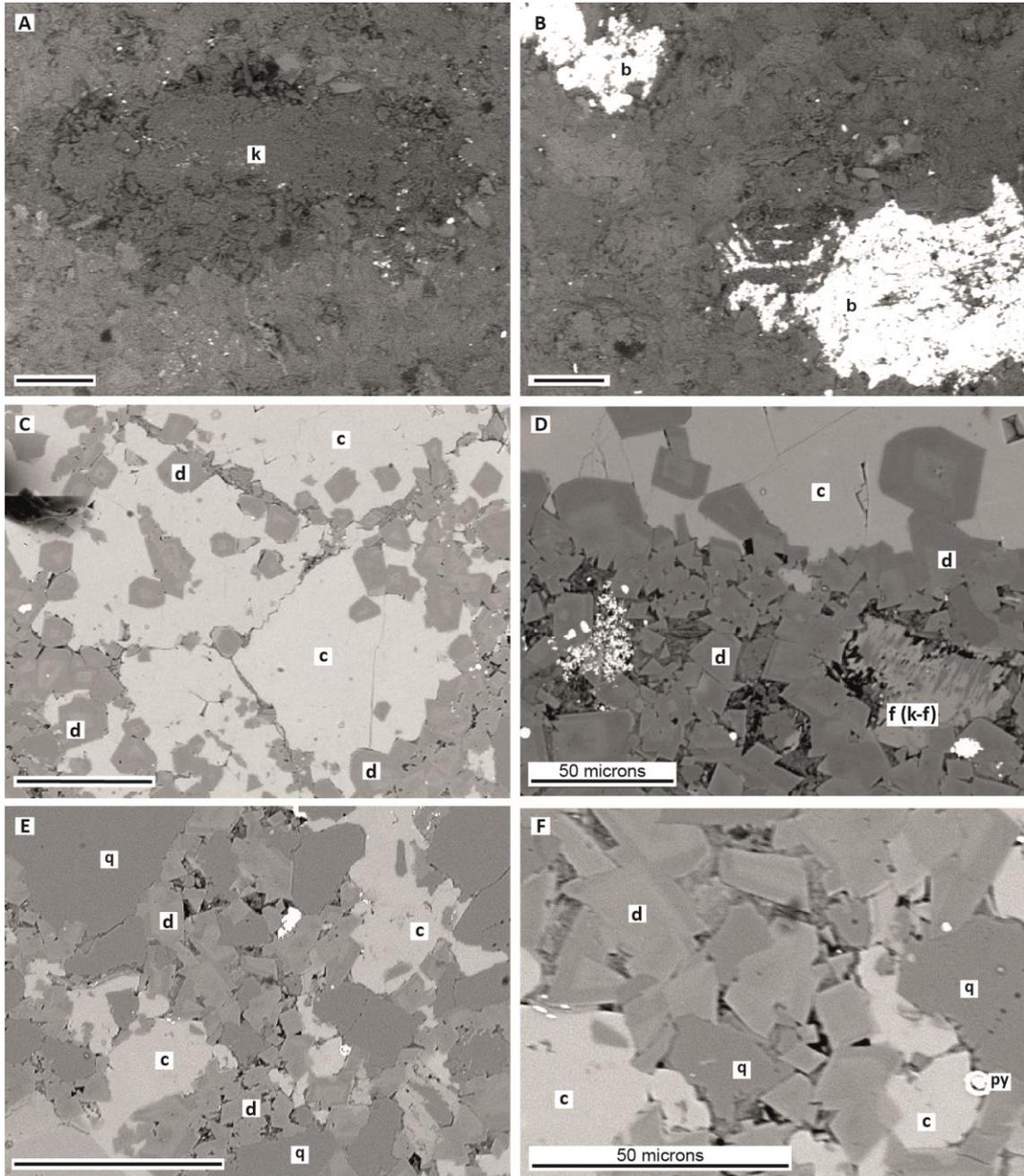


Figure 5.3: Dolomite formation in rocks at Whitrope Burn.

All scale bars = 100 microns unless otherwise shown. A. kaolinite (k) filling spaces (bed 7); B. barite (b) in a siliciclastic-dolomite matrix (bed 7); C. dolomite rhombs growing on calcite shell, uppermost thin bed at the base of the section; D. dolomite rhombs with higher Mg rims growing on calcite, lower thin beds, pyrite on left of image; note k- feldspar on lower right of image inter-grown with darker Na-rich parts; E-F. dolomite over-growing calcite formed in a siliciclastic matrix, uppermost thin bed.

The more abundant siliciclastic material suggests increased land-derived sediment. This bed may have been deposited by similar processes as the underlying units, but the bioturbation has removed much of the internal structure suggesting that, after deposition, there was time to colonise the sediment surface. The presence of *Teichichnus*-like burrows supports a lagoon/bay facies characterised by brackish water (Pemberton *et al*, 2001) or a shoreface to offshore environment. The lack of calcite suggests that the dolomite formed either as primary dolomite or as a complete replacement for original calcite mud; dolomitisation must have occurred early because fossil fragments appear uncrushed. Pyrite, kaolinite and barite in bed 7 are products of diagenesis.

5.5 Siltstones

These beds are part of the bay fill assemblage.

The dolostone beds are overlain by at least 70 cm of micaceous siltstones. The lower, 40 cm thick, siltstones are bedded, contain carbonate and are fossil rich. Small ostracods (0.5 – 1 mm), too poorly preserved to be identified but considered to be juveniles, are abundant and are sometimes concentrated into masses with spirorbiform worm tubes. Small pieces of plant and spore material are also common.

The upper siltstones are less well-bedded, do not contain carbonate, and are at least 30 cm thick; the top is unseen. Plant material is small (up to 8 mm) but common; rare ostracods, all *Cavellina* sp., and some bivalves (*Naiadites*) are present, but neither have retained their shells.

The presence of siltstones indicates an increase in silt- and clay-sized siliciclastic material delivered to this part of the basin that may have induced a switch-off in carbonate production. The upper siltstones do not contain carbonate, indicating that carbonate production had ceased by then.

The abundant, sometimes concentrated, juvenile ostracods of the lower siltstones may indicate winnowing, or that localised conditions were not conducive to ostracod

development. The occasional *Cavellina* ostracods found in the upper siltstones are likely a eurytopic species (Bennett *et al*, 2012; Williams *et al*, 2005) so cannot define the environment. However, in the nearby Roughley Burn, the bivalve *Naiadites* is found in association with *Modiolus latus* and the marine species *Sanguinolites* (Brandon *et al*, 1995; Edwards and Stubblefield, 1947) in the equivalent succession. It is thought that the species of *Naiadites* found here in the Ballagan Formation may have been marine (Peter Brand, pers. comm., 2015). The presence of *Naiadites* at Whitrope Burn therefore suggests a marine environment for the Whitrope Burn site.

5.6 Summary

The exposure at Whitrope Burn depicts a shallow-water marine environment, probably a sheltered lagoon. Initially carbonate production appears high and was followed by increased siliciclastic input.

The thin limestones at the base suggest deposition from small, pulsed, flows, each followed by siliciclastic silt deposition, and terminated by deposition from a more significant density current. The overlying bioturbated dolostone indicates continuing high carbonate production at a time of greater siliciclastic input. The fossils probably originated close to shore as suggested by the presence of tetrapod skeletal material and the large number of plant straps. In all beds minimal compaction of fossil fragments and burrows suggests early lithification. Dolomite must therefore have formed early post-deposition, either as primary growth within clays or as secondary dolomitisation of existing carbonate sediments. The absence of wave-produced structures indicates deposition below wave base.

The overlying micaceous siltstones indicate that siliciclastic input continued and carbonate production largely ceased. The fauna preserved in the siltstones indicates marginal marine conditions at a time when siliciclastic deposition appears dominant.

The environment at Whitrope Burn is defined as being a marine lagoon or bay with spatially or temporally restricted access. There is evidence for a degree of endemism

or niche separation within the chondrichthyan fauna and this supports the premise that the environment was somewhat restricted (Richards *et al*, in review). Bradyodont teeth found at Whitrope Burn differ from those found at Coquetdale (T.Smithson, pers. comm., 2015), either supporting the premise of local populations, perhaps because of environmental differences, or because the site at Whitrope Burn appears to be slightly younger than that at Coquetdale. The marine Northumberland Trough must have contained lagoons and restricted bay environments, at least along its northern edge, during the late Tournaisian to early Viséan.

6 Stable Carbon isotope values for Coldstream and Coquetdale

6.1 Introduction

By the early Carboniferous plants were widely established on the land (Rex and Scott, 1987; Falcon-Lang, 1998; Peters-Kottig *et al*, 2006), and were recovering following the Hangenberg crisis of the Late Devonian (Kaiser *et al*, 2015). Plants reflect the carbon isotope composition of the atmosphere (Farquhar *et al*, 1989) since the $\delta^{13}\text{C}$ value found in plant material is influenced by the $\delta^{13}\text{C}$ value of atmospheric CO_2 . Terrestrial organic matter is isotopically less negative than bulk material (with its marine content) due to its exposure to atmospheric CO_2 (Peters-Kottig *et al*, 2006) since plants preferentially absorb more of the lighter ^{12}C isotope than ^{13}C from the atmosphere. During the Carboniferous plant material therefore has a less negative $\delta^{13}\text{C}$ org. signal than marine material (Peters-Kottig *et al*, 2006). Fluvial systems transport plant debris to marine shorelines and shelves where it can be preserved in the marine realm. The proportion of terrestrial plant material versus organic matter generated in the marine water column (e.g. algae), and the resulting $\delta^{13}\text{C}$ value of a bulk sample therefore gives an indication as to the relative proportions of terrestrial to marine influence in a given sample (Könitzer *et al*, 2014).

From the Silurian to the Permian there is an observed trend towards less negative $\delta^{13}\text{C}$ org. values for terrestrial organic material, and an average $\delta^{13}\text{C}$ org. value of -23‰ is ascribed to plant material from the Early Carboniferous *et al*, 2006). The more ^{13}C -enriched (less negative) values of Permo-Carboniferous terrestrial organic matter, compared to earlier samples, have been interpreted to reflect the increased burial of terrestrial carbon and a resulting increase in the isotopically light sedimentary organic carbon pool (Peters-Kottig *et al*, 2006). The Carbon isotope analyses were conducted to assess the differences between specific terrestrial material and the bulk values selected from the same sample, and to compare data with those from other TW:eed project sites (Burnmouth and Norham).

6.2 Methodology

Carbon isotope values were measured for both bulk organic matter and specific, selected woody material from both Coquetdale and Coldstream. Bulk samples were derived from the crushed, complete rock samples. The specific samples were taken from organic matter (preferably visibly obvious plant material) picked either directly out of rock samples where organic material was obvious, or from crushed material from which carbonate had been removed by dissolving in 5% HCl and then washing thoroughly in de-ionised water (see Chapter 2 for preparation techniques). All samples were sent to NERC Isotope Geosciences Laboratories (NIGL) at the British Geological Survey for carbon isotope analysis; the total organic carbon (TOC) values used in this study are those supplied by NIGL.

Bulk samples give the $\delta^{13}\text{C}_{\text{org}}$ value for the complete sample of organic matter, whatever origin. Results from the specific material only indicate the $\delta^{13}\text{C}_{\text{org}}$ value for terrestrial material. The $\delta^{13}\text{C}_{\text{org}}$ data from bulk and specific analyses from both Coldstream and Coquetdale were plotted against facies and facies assemblages for each of the locations. These are relatively short sections (Coquetdale – 10 m, and Coldstream – 3.7 m) and samples were spaced at 10 -50cm and were selected based on lithology, samples being taken from mudstones or fine siltstones. Samples were also taken from a 90 cm plant- and vertebrate and invertebrate-rich section upstream from the main section in Coquetdale.

6.3 Results

The carbon isotope values for the bulk samples and specific organic matter are summarised in Table 6.1 for the main Coquetdale section, the Coquetdale ‘plant bed’ section, and for Coldstream; detailed results are available in Appendix D. At Coquetdale, the bulk samples are characterised by fairly consistent values in a narrow range, whereas $\delta^{13}\text{C}$ values for the specific wood debris have a greater range and more negative values. The Coquetdale ‘plant bed’ bulk and specific samples are more similar in range, with the specific samples again usually being a little more negative than the

bulk samples. At Coldstream the $\delta^{13}\text{C}$ bulk values have a wider range and values for the specific plant fragments are similar to those in the bulk samples.

	Bulk ‰ $\delta^{13}\text{C}$ org.		Specific ‰ $\delta^{13}\text{C}$ org.	
	from	to	from	to
Coquetdale	-21.12	-21.9	-20.14	-24.84
Coquetdale 'plant bed'	-20.83	-22.69	-20.67	-23.54
Coldstream	-21.94	-24.81	-21.61	-24.42

Table 6.1: Bulk and specific carbon isotope values for Tournaisian sites in this study. Sites in the Northumberland and Tweed basins. $\delta^{13}\text{C}$ values in relation to VPDB.

At Coldstream (Fig. 6.1), the $\delta^{13}\text{C}$ org. values for specific plant fragments track the $\delta^{13}\text{C}$ org. bulk values reasonably closely, and are generally slightly less negative. Of the eleven bulk data points (average -22.85‰), two samples are characterised by significantly more negative values of -24.51 and -24.81‰ and also have low values for their specific equivalents (-22.91‰ and -24.42‰ respectively). These two samples are from beds with particularly abundant fossil material. One of these samples has a particularly high TOC value (8.9% TOC at 36 cm) and includes large plant straps immediately below the sampled bed (Fig. 6.4), whereas the other has low TOC (0.89% at 300 cm) despite abundant fossil material.

The siltstone sample at 300 cm contains abundant fossil fragments including stacked ostracod shells, but terrestrial organic matter was noted to be very sparse during picking for specific pieces for isotope analysis. The organic material picked from this sample appeared to be plant material but *may* have included clumps of amorphous organic matter (AOM), possibly of marine origin.

Legend for logs below

legend			
facies			
f1	siltstone/sandstone with mud clasts and shell fragments	f9	mudstone, significantly disturbed by roots
f2	mudstone, laminated, sometimes fossil-rich	f10	carbonate cemented siltstones with faunal fragments
f3	siltstone, bioturbated	f11	fine to medium sandstone, erosional, sometimes with HCS
f4	siltstone, laminated, sometimes current or wave-rippled	facies assemblages	
f5	mudstone and siltstone, with some roots		bay fill assemblage
f6	very fine to fine sandstone, sometimes cross-bedded, sometimes rooted		vegetated floodplain assemblage
f7	mixed silts and microcrystalline calcite/dolomite (fossil rich)		shallow marginal marine bay assemblage
f8	sandstone, planar-bedded, wave-rippled		shallow marine near-shore assemblage
			pyrite nodule
			wave ripple
			ripple cross-lamination
			horizontal lamination
			mottled fabric
			planar cross bedding
			rootlets
			strap-like leaves
			plant material
			intraclasts
			ostracod
			bivalve
			broken shells
			bioturbation
			bone/scale
			articulated bones
			teeth
			spirorbiform microconchid
			gastropod
			convolute bedding

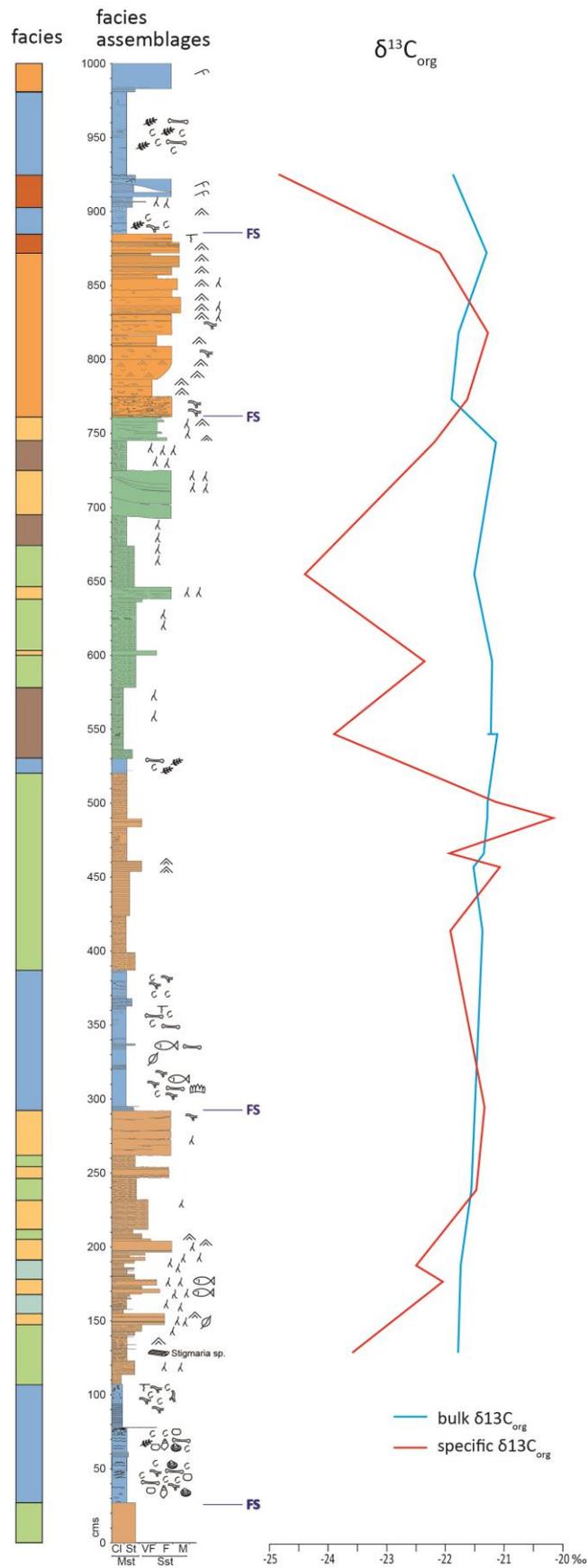


Figure 6.2: Coquetdale log showing facies and facies associations with bulk and specific $\delta^{13}C$ values. Note that the specific values are frequently more negative than the bulk values.

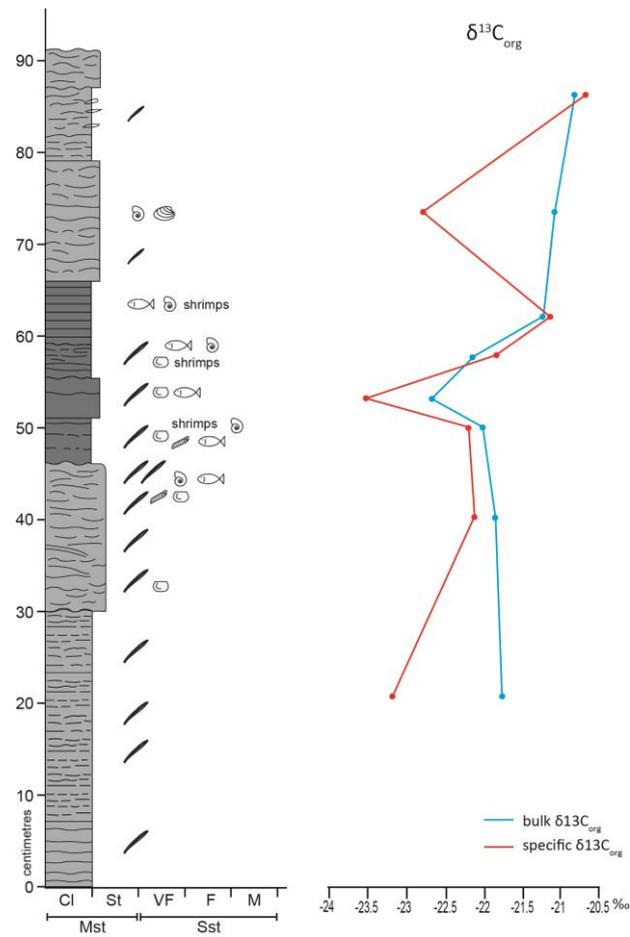


Figure 6.3: Coquetdale 'plant bed' section showing facies, facies associations and carbon isotope curves. Carbon isotope curves illustrate the variable, and generally more negative specific $\delta^{13}C_{org}$ results compared to the equivalent bulk results. These data are comparable to the data for the main section at Coquetdale. Note that this section is only 90 cm long.

At Coquetdale (Fig. 6.2) the $\delta^{13}C$ bulk values are more consistent than those at Coldstream with an average value of $-21.46 \pm 0.25\text{‰}$ whereas the specific values are much more variable (average = $-22.2 \pm 1.24\text{‰}$). Ten out of eighteen data points are also more negative than their equivalent bulk values.

TOC values are very low, with only one point being greater than 1%. The most negative $\delta^{13}C$ values for the specific material occur in the more organic-rich beds. Visual inspection of the organic material indicates dominance of terrestrial debris. The most negative values are in facies 4 (f4) 'siltstones, laminated, sometimes current or wave-rippled' and in facies 9 (f9) – the 'mudstones significantly disturbed by roots', the palaeosols. The lowest value, at 924 cm, is associated with organic material in siltstones overlying storm deposits.

Eight bulk and eight specific samples were also taken from a 90 cm organic-rich section upstream from the main site at Coquetdale, the 'plant bed' section (Fig. 6.3). Here the bulk values are similar to those in the main section with an average $\delta^{13}\text{C}_{\text{org}}$ value of $-21.81 \pm 0.61\text{‰}$ whereas the specific samples are again more variable, with an average of $-22.19 \pm 1.0\text{‰}$. Five out of the eight specific samples are more negative than their equivalent bulk samples, with the most negative values being in organic rich beds or siltstones.

6.4 Bulk carbon isotope data compared to data from the Ballagan Formation mudstones

The bulk carbon isotope data from the Ballagan Formation sampled at Coldstream and Coquetdale were compared to $\delta^{13}\text{C}$ values generated from samples in the Ballagan Formation in the Norham borehole and at Burnmouth (Bennett *et al*, 2016). Fig. 6.4 shows the complete data set of bulk $\delta^{13}\text{C}_{\text{org}}$ and total organic carbon (TOC) values, whereas Fig. 6.5 shows the spread of the majority of values in greater detail, only omitting the few samples with a TOC greater than 4%.

The Norham borehole $\delta^{13}\text{C}_{\text{org}}$ data (bulk values) are predominantly between -20.5 and -22.5‰ , with the black siltstones being mainly towards -22.5‰ . A significant proportion of Burnmouth data, is generally slightly less negative, plotting mainly between -20 and -21.5‰ $\delta^{13}\text{C}_{\text{org}}$, with a greater number of less negative data points.

The Coquetdale data from the main section plot well within the Norham data field, and are grouped closely together reflecting the minimal variation of the bulk values; the data from the 'plant bed' section is a little more variable but is still within the field of data for other Ballagan sites. The Coldstream samples are more negative than the Coquetdale data (Figs. 6.4 & 6.5). The two most negative Coldstream outliers are the two particularly fossil-rich samples.

TOC values at Coldstream and Coquetdale are generally less than 1%, with a few more organic rich samples (3.5 – 8.9% TOC) at Coldstream, and are similar to those from

Burnmouth where a few samples are 1-2 % TOC and occasional samples are 2-3 % TOC. However, TOC values in the 'plant bed' section are all above 1%. Samples with TOC greater than 4% all contain abundant plant material.

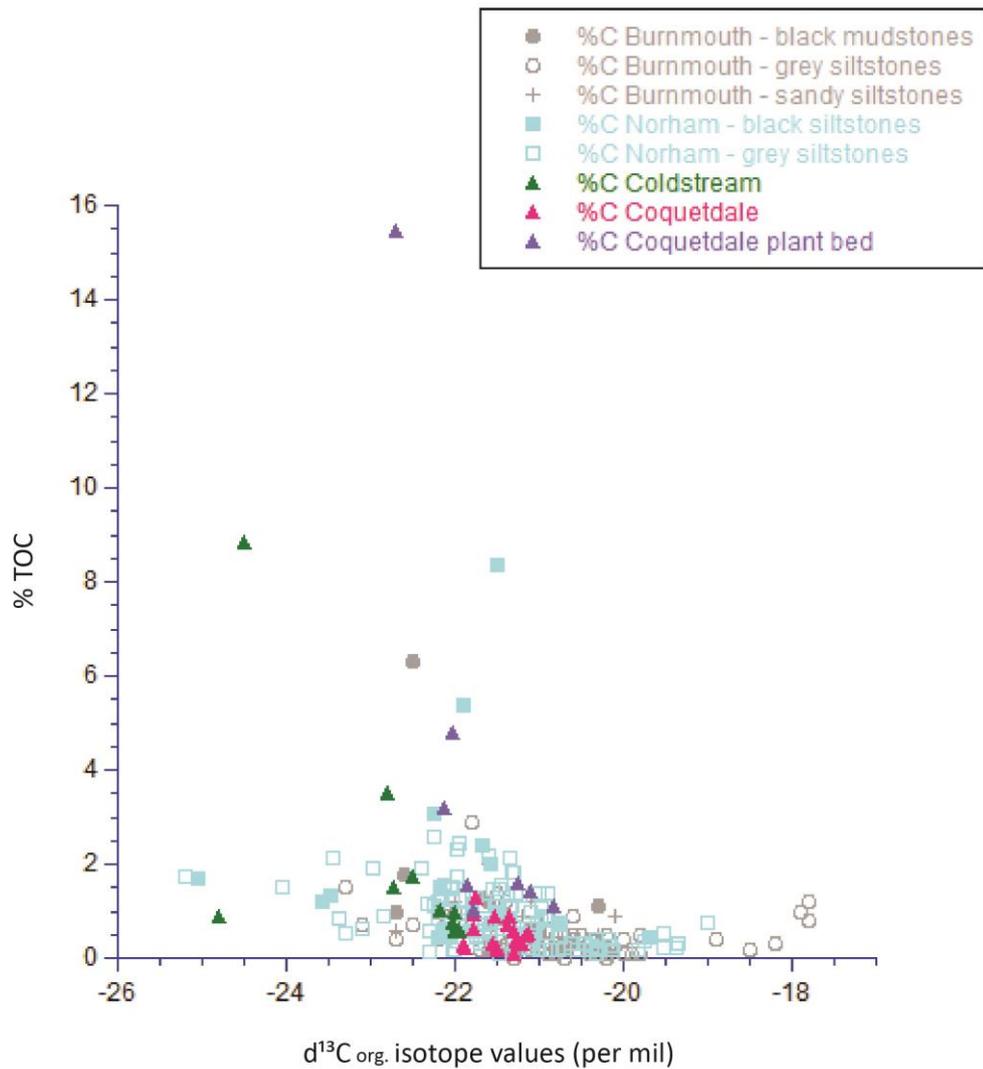


Figure 6.4: Bulk carbon isotope values for the Ballagan Formation sites. Note the low %TOC values for the majority of these samples, characteristic of the Early Carboniferous. Bulk values for Coldstream and both Coquetdale sections are similar to data from Burnmouth and Norham. The isotope values for Coldstream are generally a little more negative than the Coquetdale data, and those from the 'plant bed' section at Coquetdale are a little more variable than those in the main section but are still within the field of data for the Ballagan Formation. TOC is greater than 1% in the 'plant bed' section.

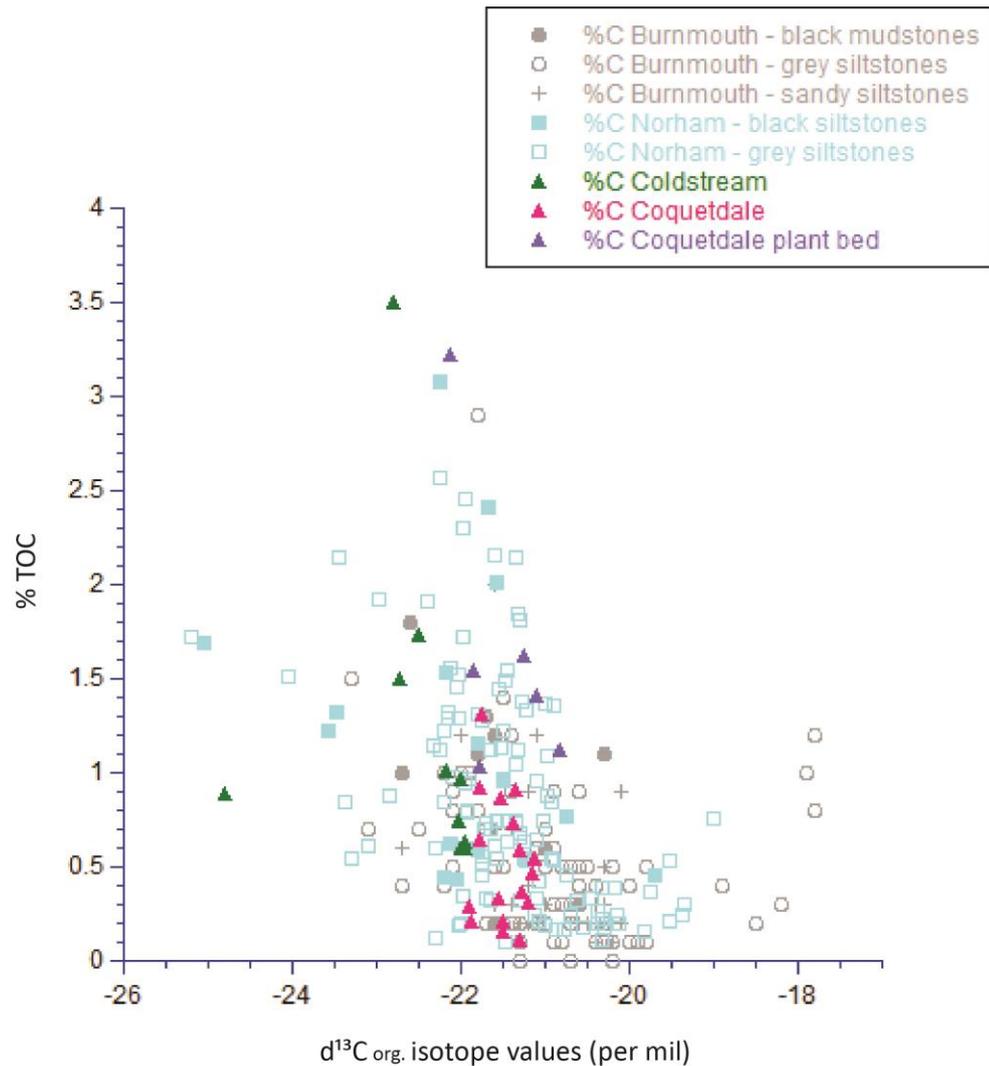


Figure 6.5: Bulk $\delta^{13}\text{C}_{\text{org}}$ values for the Ballagan Formation sites, to 4% TOC. Burnmouth samples are mainly below 1% TOC, whereas Norham values are often richer in TOC, perhaps demonstrating a preservational bias. Bulk $\delta^{13}\text{C}$ values for Coldstream and Coquetdale are similar to Norham, but are generally more negative than those from the coastal alluvial plain environment of Burnmouth (Bennett et al., 2016).

6.5 Discussion

The low total organic carbon content of these earliest Carboniferous samples, typically <1% carbon, is a common characteristic of the Ballagan Formation (Monaghan, 2014). These low values contrast with later in the Carboniferous. The total organic carbon in younger mudstone successions, for example the Asbian Gullane Formation (up to 4.8% TOC) and the Asbian-Brigantian West Lothian Oil-Shale Formation (up to 30% TOC) is typically higher (Monaghan, 2014), although it is recognised that setting also

influences TOC values in these locations. Through the Carboniferous there is a trend for increasing accumulation and burial of organic matter (Saltzman *et al*, 2004).

Limited vegetation on floodplains and low productivity in the water column may contribute to impoverished TOC in the earliest Carboniferous, but it may also reflect poor preservational potential of land plants at the time. Preservation potential of plant material could affect both TOC and bulk $\delta^{13}\text{C}$ org. values. Lignin, found in woody material, and which developed in the Early Devonian, resists biodegradation; it may be that plants found in the marginal marine and floodplain environments of the earliest Carboniferous did not contain a lot of lignin so preservation was poor, but that increasing TOC preserved in sedimentary rocks throughout the Carboniferous reflect an increase in the amount of lignin-containing plant material (Peters-Kottig *et al*, 2006), and a recovery in vegetation following the Hangenberg crisis (Kaiser *et al*, 2015).

A number of studies, e.g. Stephenson *et al* (2008); Könitzer *et al* (2014), show that bulk $\delta^{13}\text{C}$ org. values are generally more negative than the selected woody material analysed because bulk material includes both terrestrial debris and amorphous organic matter. Amorphous organic matter (AOM) has been interpreted as marine algae formed in the water column (Stephenson *et al*, 2008) and is likely synonymous with the early Carboniferous amorphous kerogens of Lewan (1986). The greater the proportion of AOM, typically the more negative the $\delta^{13}\text{C}$ value of the bulk sample.

The bulk $\delta^{13}\text{C}$ values measured in samples from the earliest Carboniferous rocks from the Tweed and Northumberland Basins are generally between -20 to -22‰ (samples from this study average -22.04‰) and are considerably less negative than the average bulk $\delta^{13}\text{C}$ values of -28.0‰ to -29.0‰ of Könitzer *et al* (2014) (late Mississippian), or -28.5‰ determined for 'amorphous kerogens' of Mississippian bulk material by Lewan (1986). It is difficult to compare the Early Carboniferous values from this study directly with the late Mississippian values of Könitzer *et al*. (2014) because the latter are known to contain a mix of terrestrial material and AOM. There is also known to be a shift in specific carbon isotope values between the Tournaisian and later in the Carboniferous (Peters-Kottig *et al*, 2006). Average bulk values from this study

(-22.04‰) are very similar to the equivalent average specific values (-22.2‰) but both are comparable to the published Early Carboniferous (Mississippian) $\delta^{13}\text{C}$ org. isotope value of -23‰ for terrestrial organic matter (Peters-Kottig *et al*, 2006). This suggests that terrestrial organic material dominates the bulk values in all three sites in this study and that there is very little, if any, AOM in the bulk organic matter.

The comparison of bulk and specific $\delta^{13}\text{C}$ org. values from Coldstream (Fig. 6.1) are largely consistent with observations by other authors (e.g. Könitzer *et al*, 2014) in that the specific values are almost always slightly less negative than the bulk values. The bulk and specific values are more negative in samples with abundant fossil material that are interpreted as deposition during or following a marine transgression (see Chapter 3). A more negative bulk value would be expected if AOM was more abundant in these fossiliferous samples, but the $\delta^{13}\text{C}$ org. value of woody material would be expected to have a similar value to the specific values from wood in other samples in this section. Abundant woody plant material was available for picking for the sample at the base of the section (8.9% TOC) and it is unlikely that AOM was selected in error. Although the sample at 300 cm contains only sparse plant material (TOC = 0.89%) again only woody plant material was picked for analysis.

The $\delta^{13}\text{C}$ org. values measured from the Coquetdale samples are more challenging to interpret. The bulk values are remarkably consistent whereas the specific values are much more variable.

The lowest specific $\delta^{13}\text{C}$ org. value in the main section at Coquetdale (-24.8‰, 924 cm) is from a siltstone overlying the storm beds within the uppermost shallow marginal marine bay assemblage (see Chapter 4). The sample at 128 cm (TOC = 0.92%) also has a more negative specific $\delta^{13}\text{C}$ org. value of -23.6‰ but a bulk $\delta^{13}\text{C}$ org. value that is similar to the rest of the section. Other beds with particularly negative values, at 546 and 654 cm, contain roots. Specific $\delta^{13}\text{C}$ org. values are also very variable in the short 'plant bed' section at Coquetdale, the most negative specific values being in an organic-rich mudstone and a fossil-rich siltstone where plant material is abundant.

It is not yet understood why the $\delta^{13}\text{C}$ value of specific material in these samples, with obvious plant material, have the most negative $\delta^{13}\text{C}$ values in the section and are more negative than the bulk. This is counter to observations in other studies where the plant material is generally less negative than bulk material. It is difficult to explain the differences and variation in specific $\delta^{13}\text{C}$ values at Coquetdale but there are several possible contributory factors. The observation that the bulk and specific $\delta^{13}\text{C}$ values track one another at Coldstream may indicate that variations reflect changes in the larger-scale carbon cycle, or that the plants are different, leading to differences in $\delta^{13}\text{C}$ values. More data would be needed to determine which hypothesis is correct.

Variations in plant type or habitat likely influenced the $\delta^{13}\text{C}$ values; both these effects are known to shift $\delta^{13}\text{C}$ values by several per mil (Arens *et al*, 2000; Peters-Kottig *et al*, 2006). Hartman and Danin (2010) also found variations in $\delta^{13}\text{C}$ values in recent plant types and in varying climatic conditions; values related to water stress, with $\delta^{13}\text{C}$ values being generally more negative in wetter conditions. It may be that the very variable specific $\delta^{13}\text{C}$ values at Coquetdale are more influenced by variation in floodplain conditions and seasonality, impacting on the plant cover (Kearsey *et al*, 2016). Today there is also variation in $\delta^{13}\text{C}$ values of different plant tissues, with non-photosynthetic material having less negative $\delta^{13}\text{C}$ values than photosynthetic material (Szpak *et al*, 2013); it is likely that this was also the case during the Carboniferous. This *may* contribute to an explanation for the negative values where plant material (plant straps) has been transported, but not to the more negative values that coincide with palaeosols where roots seem to be the main organic material.

Variation in plant type may also be attributed to seasonal climatic variation. A palaeolatitude position of 4°S of the equator (Scotese and McKerrow, 1990) and interpretation of a tropical climate with seasonal monsoonal rainfall (Falcon-Lang, 1999; Kearsey *et al*, 2016) may have contributed to the frequent variation in the carbon isotope signal.

The palaeogeography of the Northumberland Trough in the Tournaisian, the extent of the floodplains and vegetation, and the reach and depth of marine transgressions from the west, must have defined the particular conditions at each of the sites at the time.

Lagoons and restricted bays seem to have been a feature of the northern edge of the Northumberland Trough at this time, as seen at Coquetdale (Chapter 4) and Whitrope Burn (Chapter 5). It is suggested that the long, shallow nature of the Northumberland Trough (Fig. 1.2) meant that marine incursions would have had a relatively greater effect on the environment at individual sites, small sea level rises having a widespread effect, with shallow lagoons or bays being proportionally more affected. The resulting changing distance from shore would have disproportionately altered the amounts of terrestrial material reaching a particular spot and hence the bulk $\delta^{13}\text{C}$ value for a particular sample. Rapidly fluctuating water levels, whether they be from local sea level variation, storm surges, or monsoonal climatic variability, may also have contributed to variation in plant type at any given time, and thus to the variability seen in the specific values at Coquetdale.

$\delta^{13}\text{C}$ isotope data from the bulk organic material of the coastal alluvial plain sediments of Burnmouth and Norham are more similar in value to the organic matter in lagoonal/marginal marine sediments of Coquetdale than to organic matter in the more marine bay-fill sediments of Coldstream.

6.6 Conclusions

- Ballagan Formation mudstones are generally low in TOC (<1%), and are comparable to other Early Carboniferous mudstones (e.g. Monaghan, 2014). This low %TOC may be due to plant preservation potential (lower proportions of lignin) and variation in plant types, the rapid colonisers perhaps having low preservational potential.
- The average $\delta^{13}\text{C}$ org. values plot within the broad field of Ballagan Formation $\delta^{13}\text{C}$ values and are distinctly less negative than published Mississippian $\delta^{13}\text{C}$ values of specific and bulk material (Peters-Kottig *et al*, 2006; Lewan, 1986).

- At Coldstream, bulk and specific values track one another. The most negative values are associated with fossiliferous mudstones that are interpreted as part of the bay fill and shallow marginal marine bay environments. The low bulk and specific $\delta^{13}\text{C}$ values could reflect different plants incorporated into the sediments.
- At Coquetdale, the bulk organic $\delta^{13}\text{C}$ values are consistent whereas specific values are very variable and often more negative than the bulk values. The explanation requires further investigation.
- The shallow, low-lying nature of the Northumberland Basin in the Tournaisian means that small changes in sea level could intermittently expose large areas for plant colonisation. Variation in plant types adjacent to the shoreline as sea level fluctuated likely contributed to the variable specific carbon isotope values for the Coquetdale section.

7 Discussion and conclusions

The fundamental question asked in this project is what is the sedimentological context of Early Carboniferous vertebrate fossils finds (specifically lungfish and tetrapods) at three key sites; Coldstream, Coquetdale and Whitrope Burn?

The three sites examined are all short sections in comparison to other Tweed project sites (e.g. 500 m thickness examined in the coastal section at Burnmouth and in the core from Norham) but although sections are thin (1.4 – 10 metres) they are important because of the known presence of lungfish and tetrapod material. The detailed sedimentological analysis completed at Burnmouth and Norham (Bennett *et al*, 2016; Kearsy *et al*, 2016) enables the context of these small snapshots to be compared and the differences determined.

Palynological dating has as yet only been completed for Whitrope Burn, the section being latest Tournaisian/early Viséan (John Marshall, pers. comm.). The section at Coquetdale is probably late Tournaisian because it is only about 80 metres below the base of the Fell Sandstone Formation (see Fig. 1.1), and the Coldstream site is likely to be latest Tournaisian since the latest BGS survey determined the nearby rocks at Lennel Braes (<1 km away) as latest Tournaisian.

Eventually the revised palaeostratigraphy undertaken as part of the Tweed project should enable these sites to be correlated with those at Burnmouth and Norham but this will not happen until after completion of this study.

7.1 Comparison of the depositional environments at the study sites

In the Early Carboniferous it is challenging to distinguish between marine and non-marine environments, differences between sites are subtle and have to be based on associations and the weight of evidence. Many fauna inhabited marginal marine environments and were either euryhaline or are interpreted as being brackish to

marginal marine in some studies (e.g. rhizodonts in Carpenter *et al* (2014), but freshwater in others (e.g. rhizodonts in Jeffery (2012)), sometimes because different species were adapted to different environments. Carboniferous fauna are thus ambiguous indicators of environment. In the case of many fauna the issue is exacerbated because their origin and affinity is not well-known, for example the Carboniferous ostracod genera *Cavellina*, *Shemonaella* and *Carbonita* were all extinct by the end-Permian (Moore, 1961; Bennett *et al.*, 2012).

Facies assemblages at each of the three sites have been illustrated and described in Chapters 3-5. Some individual facies within these assemblages occur at only one of the three sites, although the sites also have facies in common (see Table 7.1 for facies descriptions). Facies 1, 2, 3, 7 and 10 are fauna-bearing siltstones. The fauna present, including marginal marine to brackish water ostracods e.g. *Cavellina* and *Shemonaella* (Williams *et al*, 2005, 2006; Bennett *et al*, 2012) and the bivalve *Modiolus* (Wilson, 1965; Ballèvre and Lardeux, 2005), in association with brackish water spirorbids (Williams *et al*, 2005; Taylor and Vinn, 2006; Zaton *et al*, 2012) indicate that these are likely to be marginal marine siltstones. During the Early Carboniferous, spirorbids and ostracods were euryhaline (Bennett, 2008), invading brackish and freshwater environments; here the faunal associations indicate that the spirorbids were a brackish water species, and were particularly abundant in the restricted facies 7, mixed siltstone and microcrystalline calcite/dolomite.

Facies 6, 8 and 11 are sandstones, dominated by wave- and storm-generated structures with energy increasing from facies 6 to 11. The erosional, storm-influenced sandstones of facies 11 are found only at Coquetdale.

Facies 4, 5 and 9 are mudstones and siltstones with evidence of colonisation by vegetation, with roots most abundant in facies 9. Facies 9 is present only at Coquetdale and is interpreted as an immature palaeosol. Facies 9 is absent at Coldstream suggesting that, although plants were established, insufficient time was available to develop palaeosols.

Facies	Facies number	Depositional Processes	Facies location	Facies Assemblage Table 7.2
Siltstone/mudstone with mud clasts and shell fragments	f1	Deposition of transported material - rounded mud clasts and bone/scale/shell fragments; deposition by small mass flows	CS &WB	Bay fill
Mudstone, laminated, sometimes fossil-rich	f2	Intermittent deposition of transported material (sediment and fossils) from brackish water/marginal marine area alternating with settling of fine-grained material. Absence of burrows and presence of framboidal pyrite indicate a reduced oxygen environment	CS only	Bay fill
Siltstone, bioturbated	f3	Remnant lamination generated by flows into still water, includes transport of plant material, bioturbated – better oxygenated water relative to f2, some pyrite in sediment and on organic material	CS only	Bay fill
Siltstone, laminated, sometimes current or wave-rippled	f4	Transport of silt-sized material into standing body of water likely from fluvial input, fluctuating velocities in transporting flow, sufficient to transport plant material, settling-out of fine plant material between flows, occasional wave action rework sands. Some plant growth (occasional roots)	CD & CS	Bay fill, and Vegetated floodplain
Mudstone and siltstone, with some roots	f5	Transport of silt-sized material, occasionally sand grade, into very shallow standing water (rooted, organic-rich area at Coldstream); wave reworking of coarser transported sediment, sub-aerial exposure - roots are common	CD & CS	Bay fill, and Vegetated floodplain
Sandstone, very fine to fine, sometimes cross-bedded, sometimes rooted	f6	Sand-grade sediment deposited in very shallow standing water, wave reworking more prevalent at Coquetdale. Roots are common – very shallow water to sub-aerial exposure. Where cross-bedding is identified larger scale dune formation indicated	CD & CS	Bay fill, and Vegetated floodplain
Mixed siltstone and microcrystalline calcite/dolomite	f7	Silt and mud transported to very shallow standing water; cup-in-cup ostracods indicate gentle reworking of abundant ostracod valves; possible early formation of dolomite in clays/some calcite formation between grains so dolomite may be diagenetic; presence of dolomite interpreted as marine waters	CS only	Shallow marginal marine bay
Sandstone, planar-bedded, wave-rippled	f8	Current transport of sand-grade material, dune formation, 'flaggy', wave-rippled at Coldstream; wave reworking at Coquetdale	CD & CS	Shallow marine near-shore
Mudstone, significantly disturbed by roots	f9	Subaerial deposition of muds and silts, oxidation of sediment, roots – plant growth, almost total destruction of lamination. Immature palaeosol development at Coquetdale only.	CD only	Vegetated floodplain
Carbonate cemented siltstones with fossil fragments	f10	Siliciclastic material and faunal fragments transported and deposited into shallow water. Fossils indicate semi-restricted bay/lagoon; occasional sub-aerial exposure or very shallow water – desiccation cracks or syneresis cracks. Dolomite growth within clays, and early diagenetic alteration of calcite to dolomite – dolomitised limestones. Dolostone packages develop from coarser-grained lower beds to fine-grained laminated or bioturbated upper beds.	CD & WB	Shallow marginal marine bay
Sandstone, fine to medium, erosional, sometimes with HCS	f11	Sand deposition during wave ravinement or from erosive storm-generated flows with the gutter casts and hummocky cross-stratification	CD only	Shallow marine near-shore, and Shallow marginal marine bay

Table 7.1: Facies and depositional processes found at Coldstream, Coquetdale and Whitrope Burn.

Facies assemblage	Coldstream facies	Coquetdale facies	Whitrope Burn facies	Comments
Bay fill	f1, f2, f3, f4, f5, f6	f4, f5, f6	f1	Coldstream: facies 1-6 represent a complete coarsening-upward bay fill facies assemblage, from mudstones with bioturbation to fine-grained sandstones; bioturbation and fauna in lower facies suggest a more marine signature, upper facies, with roots, becoming more marginal marine. At Coquetdale facies 4-6 are current or wave rippled with some roots and indicate shallower water. At Whitrope Burn only f1 is present with a marginal marine/eurytopic fauna.
Vegetated floodplain	f5, f6	f4, f6, f9	Not present	Coquetdale and Coldstream: facies assemblage indicate very shallow water and/or subaerial exposure. Coquetdale has roots and immature palaeosols. Coldstream: roots but no palaeosols identified.
Shallow marginal marine bay	f4, f7, f11	f8, f10, f11	f10	Coquetdale and Coldstream: Wave-ripples or abundant monospecific fossil assemblages and cup-in-cup ostracods in facies, 4, 7 and 8 indicate shallow marginal marine waters, sometimes in a shoreline setting. Coldstream: Facies 11 is interpreted as sands deposited during transgression. Coquetdale: Facies 11 is interpreted as erosional, sometimes storm-generated sandstones. Facies 10, present at Coquetdale and Whitrope Burn, is represented by the dolostone beds characteristic of the Ballagan Formation. Facies 10 is not represented at Coldstream, although dolomite is present as a minor cement.
Shallow marine near-shore	f8	f8, f11	Not present	Coldstream and Coquetdale: Facies 8 is present but expressed differently at each of these locations. Coldstream: facies 8 is represented by wave-rippled 'flaggy' sandstones that overlie the shallow marginal marine assemblage and is interpreted as a mouth bar. Coquetdale: facies 8 contains bioturbation, including escape traces, characteristic of a foreshore to shoreface environment. Facies 8, the base of which is a muddy sandstone with combined flow and wave ripples, overlies the vegetated floodplain assemblage and in this context is interpreted as shoreface sands which are being transgressed as sea level rises. Increasing energy at Coquetdale is illustrated by the presence of facies 11. Context enables the interpretation of facies 8 as differing environments at these locations.

Table 7.2: Facies assemblages found at Coldstream, Coquetdale and Whitrope Burn.

All four facies assemblages (Table 7.2) are present at Coquetdale and Coldstream, but at Whitrope Burn only two assemblages are present in the very short section exposed. Neither the vegetated floodplain assemblage nor the shallow marine near-shore assemblage are represented at Whitrope Burn.

The shallow marginal marine bay assemblage is represented at all three sites. The least marine end member of this facies assemblage is seen at Coldstream where there is direct fluvial input and abundant brackish water fauna, e.g. spirorbids (Fig. 3.6A), and very shallow-water 'cup in cup' ostracods are identified. More marine examples of this facies assemblage are seen at Coquetdale and Whitrope Burn where dolostones hosting marginal marine fauna, calcifying green algae and diverse bioturbation all support a shallow water environment within the photic zone. At Coquetdale and Whitrope Burn the dolostones appear to have mainly originated as shallow marine limestones which have later been dolomitised, with some possible direct precipitation of dolomite. At Coldstream dolomite is scarcer but the typical dolostones of the Ballagan Formation are not present. Dolostones may have developed at Coquetdale and Whitrope Burn because there is a greater marine influence supplying the necessary sulphate, and a degree of restriction to concentrate the marine waters.

The shallow marine near-shore facies assemblage is represented at Coquetdale and Coldstream but is interpreted differently based on subtle differences in character and context. The planar-bedded, wave-rippled sandstones of Coquetdale are interpreted as foreshore deposits in deepening water, and a more open setting because of the abundance of bivalve escape traces, other bioturbation characteristic of a foreshore position and occasional storm beds. As transgression occurred and the planar-bedded sandstones were deposited, it appears that the bay was more open.

At Coldstream the shallow marine near-shore facies assemblage is represented by sharp-based, flaggy, wave-rippled sandstones, with some bioturbation. Here these beds are interpreted as mouth bar deposits developed as the bar prograded out into the bay (e.g. Elliott, 1974), there being direct fluvial input at Coldstream.

As interpreted in Chapter 3, the Coldstream site was initially a shallow, near-shore bay that was adjacent to a coastal alluvial plain similar to that interpreted from the successions at Burnmouth and Norham (also in the Tweed Basin). During periods of high rainfall a range of terrestrial material, including the tetrapod material and large plant straps, was transported by fluvial systems directly into the bay.

Coquetdale and Whitrope Burn are both located along the northern edge of the Northumberland Basin and at these locations lagoons and shallow, marginal marine bays are interpreted from the facies assemblages. Direct fluvial input appears to have been less important here and it is more likely that material was moved around within the bays. At Coquetdale, the presence of roots in fine-grained intervals but without significant soil development are interpreted as short-lived floodplain development in a very marginal location. At Coldstream the floodplain was of even more brief duration and even immature palaeosols did not have time to develop.

7.2 Palaeogeography

During the Tournaisian and early Viséan the Tweed Basin and the Northumberland Basin were separated by the Cheviot Block (Fig. 1.2). The Northumberland Basin is a half-graben running west to east along the southern edge of the Southern Uplands and the Cheviot Block (Stone *et al*, 2010).

The sedimentological evidence, including the rapid alternation between wave-ripples, vegetated surfaces and lagoons with biolaminites and calcifying green algae, particularly at Coquetdale, suggests frequently fluctuating water levels and that the northern edge of the Northumberland Basin was very shallow and flat-lying during the Tournaisian. All sites in this study are interpreted as shallow marine, positioned close to the shoreline. Fig. 7.1 is an illustrative model (not to scale) depicting the likely positions of sites in this study compared to those at Burnmouth and Norham.

A deeper marine environment existed to the west (Fig. 1.2), and a greater marine influence in the west of the Northumberland Basin is consistent with the presence of

dolomitised marine limestones and more typically marine fauna (chondrichthyans) identified at Whitrope Burn. Marine transgressions, probably local rather than eustatic in nature because of their frequency, would have affected large areas in this low-lying trough, sometimes inundating and connecting, sometimes partially separating areas. Shallow lagoons seem to be a feature of this northern edge of the basin, each with a degree of restriction, and promoting conditions which varied from place to place.

The site at Coldstream (Tweed basin; Figs. 1.2; 2.1) is generally more marine than the coastal alluvial plains of Burnmouth and Norham (cf. Bennett *et al*, 2016), with bay fill facies and more marginal marine fauna. The influence of direct fluvial input freshened the waters of the bay and provided habitats for brackish to marginal marine species, it also directed faunal and floral debris into the bay rather than across the floodplain as at Burnmouth. This direct fluvial, and hence siliciclastic, input explains the fossil-rich silty laminae found in the lower part of the section, the fluvial mouth bar sands, and the lack of the 'sandy siltstones' found at Burnmouth since material was moved directly into the bay rather than across a floodplain.

The westerly location of Coquetdale on the shallow, northern edge of the Northumberland Trough, and a succession that suggests a semi-restricted shallow lagoon indicate a shallow coastal setting, with rapid alternation of environment and a less open setting than at Coldstream. Alternation on a decimetre scale of wave-rippled sandstones and desiccated and vegetated horizons at Coquetdale suggest that water depth fluctuation occurred frequently and rapidly. This is typical of very shallow, low-lying marginal sites.

In contrast to Coldstream there does not appear to be a direct fluvial input of siliciclastic material at either Coquetdale or Whitrope Burn. At Coquetdale siliciclastic sediment filled the bays/lagoons, transported into the area by redistribution rather than by direct input from fluvial systems; there is no evidence of fluvial bodies. Rare small flows into the lagoon (e.g. thin fossil-rich dolostones found on top of thicker dolostones) are interpreted as the result of increased inland rainfall. At Whitrope Burn there is no evidence of fluvial bodies and the fossil-rich bed and the bioturbated

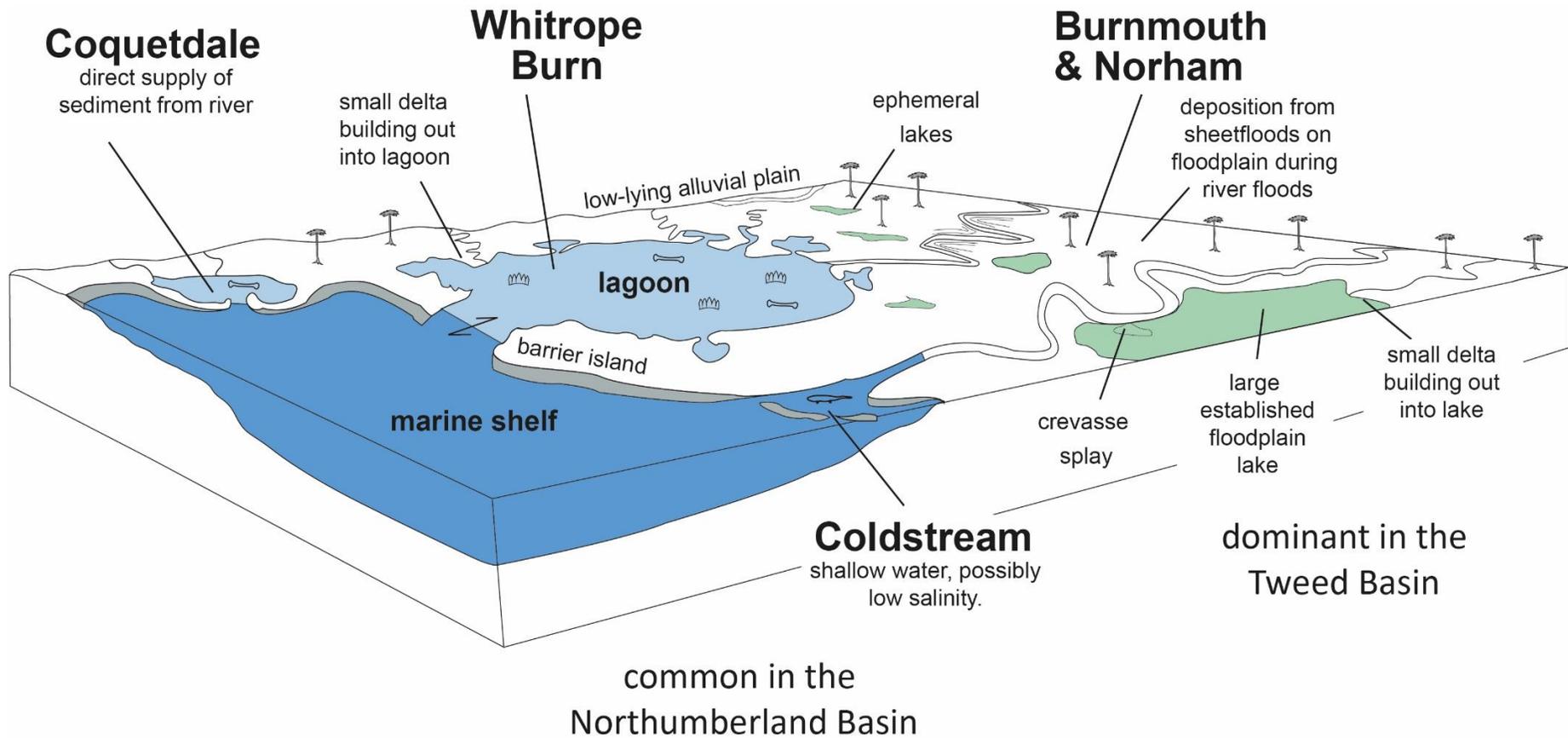


Figure 7.1: Model of likely site locations illustrating their comparative positions and suggested marine/terrestrial influences. Model based on evidence from this study and other localities in the Tweed basin (Bennett et al., 2016). Coldstream is in the Tweed basin, and Coquetdale and Whitrope Burn, with their similar lagoonal/restricted bay environments, are on the northern edge of the Northumberland Trough.

sandstone above it appear to have been event beds deposited into a standing body of water following inland storms, with the bay-fill siltstones again being the result of passive deposition.

7.3 Carbon isotopes

Carbon isotope values ($\delta^{13}\text{C}$) and % total organic carbon (TOC) were measured on material from Coquetdale and Coldstream. TOC values for the bulk samples were generally of less than 1% and are consistent with other Early Carboniferous values. The carbon isotope data plot with data from Burnmouth and Norham, showing consistency of values in the Ballagan Formation, but differ from published values of -23‰ for the Early Carboniferous (e.g. Peters-Kottig *et al.*, 2006). The generally less negative bulk values of the Coquetdale data are influenced by a greater proportion of terrestrial material compared to those at Coldstream and support the interpretation of slightly more marginal conditions. Some of the Burnmouth and Norham data is less negative than that at Coquetdale and Coldstream and indicates the greater influence of plant material from the terrestrial floodplain facies there.

The $\delta^{13}\text{C}$ values are challenging to explain because there are differences in the data compared to the published literature. Published data (Stephenson *et al.*, 2008; Könitzer *et al.*, 2014) have shown that bulk $\delta^{13}\text{C}$ org. values are typically more negative than the values measured on specific fragments of wood. This has been interpreted as the bulk $\delta^{13}\text{C}$ org. values representing a mix of terrestrial organic material (typically less negative) and amorphous organic matter (typically more negative). At Coldstream this relationship is observed although the specific values track the bulk values closely and are similar. These similar values at Coldstream suggest that the bulk $\delta^{13}\text{C}$ values are dominated by terrestrial organic material. This close tracking is unusual and deserves further investigation. It may be part of a larger trend in the carbon isotope curve which is not yet fully understood.

At Coquetdale the bulk values are relatively consistent, as may be expected for a short section, but the specific values vary significantly. There are two possible reasons for this. The rapid variation in the specific carbon isotope values may reflect variations in

vegetation types generated by the far-reaching effects of transgressions and regressions: for example the vegetation growing on newly exposed areas following sea level fall are likely to have been very different to plants on more established areas, as observed in younger Carboniferous successions and reported in Davies and McLean (1996) and Hawkins *et al.* (2013). There may also be a variation in preservation potential, according to how much lignin different plant types possess, with more mature woody plants possessing more lignin than rapid-growing opportunistic plants (Peters-Kottig *et al.*, 2006). The greater the preservation potential the greater the likelihood of the plant material influencing the specific values. Alternatively, variations in climate, and the seasonality of a monsoonal climate, may also contribute to variation in plant type, and hence preservation potential.

These two sites present only a small amount of data, with unusual variations. From the short sections and the data available it is difficult to determine what is happening, and what the primary influences are on the $\delta^{13}\text{C}$ values. The carbon isotope data from these sections could be further investigated in the light of other Ballagan Formation data and Carboniferous trends to further our understanding.

7.4 Preservation of fossil material

Detailed, lamina by lamina studies have been conducted on certain parts of sections where vertebrate, and particularly tetrapod or lungfish material, has been found, or was anticipated. These are detailed snapshots within a larger context. The remaining parts of the succession have been examined carefully but not to the same level of sampling.

Detailed analysis of a block extracted from facies 2 of the bay-fill assemblage at Coldstream (equivalent to where tetrapod material had been found previously, T. Smithson, pers. comm.) revealed abundant vertebrate and invertebrate fossil material and enabled precise recording of where, and in which types of sediment, vertebrate material was preserved.

At Coquetdale particular attention was paid initially to the dolostone beds since fossil material was noted there, and at Whitrope Burn attention was focussed on a dolostone bed with abundant and obvious vertebrate fauna. This study puts these observations into the context of each section.

Where are fossils preserved?

At Coldstream vertebrate and invertebrate fossils are preserved in the bay-fill assemblage. Vertebrate material, including tetrapod and lungfish fossils, is more common towards the base of the section, in facies 1 and 2; invertebrates are present throughout the facies assemblage but decrease in abundance upwards. The vertebrate and invertebrate material is preserved in silt-rich laminae interpreted as frequent silty flows into the basin. Invertebrate fossils are again abundant in the shallow marginal marine bay facies assemblage (facies 7) where water depths are very shallow and shoreline conditions prevailed.

At both Coquetdale and Whitrope Burn vertebrate fossils are preserved in the dolostones of the shallow marginal marine bay assemblage; both sites are interpreted as lagoonal environments. At Coquetdale vertebrate (but not tetrapods or lungfish) and invertebrate material is found in the dolostones, particularly on the top surfaces, or in thin dolostone beds overlying thicker dolostones. Vertebrate and invertebrate material has been transported in repeated flows into or around the basin and then been deposited with the silty material, and also settled out of the water column to be preserved later on the top of dolostones beds. At Coquetdale occasional vertebrate fossils have also been found in the sandier facies of the bay-fill assemblage, although a systematic analysis of fossil content has not been undertaken. Lungfish material has been found nearby in mudstones/siltstones.

At Whitrope Burn vertebrate and invertebrate fossils, including tetrapod and lungfish fragments, have been transported in a single density flow, as an event bed. Here the diverse chondrichthyan fauna and the tetrapods and lungfish have been transported from their original site on the edge of, or within, the lagoon into deeper water further

out in the lagoon, perhaps as a result of excessive inland rainfall and consistent with the abundance of large plant straps.

How are the fossils preserved?

In the bay-fill assemblage at Coldstream many of the fossil fragments in the lower part of the section are preserved in pyrite, with minimal crushing even of plant material. Framboidal pyrite indicates a marine setting with abundant sulphur available for early pyritisation and preservation of fossil fragments (e.g. Raiswell, 1982; see also Chapter 3). Material preserved in the shallow, marginal marine bay assemblage at each of the sites was likely subject to early calcite cementation and/or dolomitisation of some of the sediments. This early cementation also contributed to the largely uncrushed nature of many fossil fragments and to their overall preservation.

At all three sites both vertebrate and invertebrate material has been preserved in facies showing evidence for transport and possible concentration, sometimes in thin (millimetre-scale) silty flows and sometimes as thicker density flows. Rarely, invertebrate material has been preserved either in situ or very close to its site of origin, as in the shallow marginal marine bay assemblage at Coldstream. The palaeogeography at the time, the sedimentology and the marginal marine nature of the environments imply that fossil fragments, including tetrapod material, were preserved in lagoons and shallow marine bays.

Tetrapods likely inhabited the fresher water parts of the marginal marine environment (Kearsey *et al.*, 2016; Smithson *et al.*, 2012) and were intermittently transported into, and preserved in, shallow water facies (perhaps upon death). The association of abundant fossil fragments and plant straps infers that transportation took place during times of heightened run-off from increased rainfall events inland, thus transporting terrestrial plant material and fossil fragments, including tetrapods and lungfish, into the marginal marine realm. Fossil fragments are generally not badly abraded, and at Coldstream some bivalve shells are preserved in the 'butterfly' position; it is thus likely that fossils were deposited relatively close to their location of origin and in some cases may not have moved any distance at all.

There are similarities between the mode of preservation at the sites in this study and those in other key sections of the TW:eed project (Burnmouth and Norham), but the environments of deposition are different. The Burnmouth and Norham sections are interpreted as a coastal alluvial floodplain environment (Anderton, 1985; Andrews and Nabi, 1994, 1998; Andrews *et al.*, 1991; Bennett *et al.*, 2016; Kearsley *et al.*, 2016; Scott, 1986; Stephenson *et al.*, 2002; 2004). At Burnmouth tetrapod fossils are found in fine-grained beds interpreted as cohesive debris flow deposits, and defined as 'sandy siltstones' (Bennett *et al.*, 2016), that originated in overbank settings through overflow from rivers, floodplain lakes, and erosion of land surfaces at times of high rainfall. Tetrapod fossils have been concentrated by local erosion and reworking and deposited into temporary lakes on the floodplain (Bennett *et al.*, 2016). Similar processes were active at the sites examined in this study but the silt-rich flows transporting fossil material here differ from those at Burnmouth in that they transported material directly offshore, by fluvial channels, or into and around the lagoons of these marginal marine sites.

It appears that tetrapod material found at these sites, and others in the TW:eed project, has been preserved following transport, the sites of deposition being related to their palaeogeographic position. The localities in which tetrapods were living at the time remain an enigma.

7.5 Conclusions

- All three sites in this study are interpreted as being shallow marginal marine environments. Framboidal pyrite found at all three sites, in association with marginal marine species, rapid fluctuations in water level, repeated deposition of dolostones during times of transgression, and wave-rippled foreshore or mouth bar sandstones confirm these as deposits from a marginal marine setting with local fluctuations in sea level.

- Vertebrate fossils, including lungfish, have been found at all three sites, but tetrapods have so far only been found at Coldstream and Whitrope Burn.
- All vertebrate, and most invertebrate material has been transported in silty flows from its place of origin, with material occasionally settling out of the water column. In certain conditions invertebrates have thrived, become abundant and been preserved in situ.
- The shelfal area of the Tweed Basin was evidently relatively open, with direct fluvial input at Coldstream and consequent deposition into a relatively open bay. The position of the Coquetdale and Whitrope Burn sites on the northern edge of the shallow, low-lying Northumberland Trough means that vertebrate material is preserved in lagoons.
- The Coquetdale and Whitrope Burn sites were temporally separated, however differences in fauna between the two locations, with marginal marine (actinopterygians and rhizodonts) and lagoonal species (dasyclads) at Coquetdale and a diverse chondrichthyan fauna at Whitrope Burn, suggest that the bays and lagoons that existed along the northern margin of the Northumberland basin were isolated from one another. This spatial separation enabled the degree of endemism or niche separation observed (see also (Richards *et al*, in review).
- Brackish water fauna (e.g. spirorbids and ostracods) predominate at Coldstream reflecting the direct fluvial freshwater input into a shallow marine bay.

- Transport and deposition of vertebrate fossils into the shallow marine environments of these sites contrasts with other sites in the TW:eed project; at Burnmouth and Norham vertebrate fossils were transported across a floodplain and are preserved in 'sandy siltstones'.
- Palaeogeographic location also determines the presence or absence of the typical dolostone beds of the Ballagan Formation. At the sites in this study dolostones have only formed in the semi-restricted environments where there was limited exchange of marine waters, and insignificant freshwater input (generating a concentration of Mg ions). At the semi-restricted lagoonal sites of Coquetdale and Whitrope Burn dolostone beds are an important part of the sequences, but in the more open position of Coldstream, with the freshwater input, there was limited production of dolomite and the typical dolostone beds are absent.
- Dolostones occur in packages which record reducing siliciclastic supply within each package. The coarser lower beds, sometimes with rounded clasts, indicate transgressions and ravinement; microbial lamination and bioturbation in the upper parts of packages indicate reducing energy. Dolostone packages are thus interpreted as having been deposited at times of sea level rise. Bed thicknesses and the proportion of siliciclastic content suggest that carbonate production must have been high at these times. Fossil content indicates a shallow marginal marine environment, fossils commonly occurring in thin dolostone beds representing late surges of material into the basin. Marginal marine conodonts recorded in the dolostones at Coquetdale several years ago confirm marine conditions (M. Purnell, pers. comm.).
- Almost all dolomite found at these sites is interpreted as diagenetically altered calcite. In almost all cases dolomite is found in conjunction with calcite, and is

usually seen to be replacing it. The dolostone beds of Coquetdale and Whitrope Burn are interpreted as dolomitised limestones. At Coldstream the generally more open nature of the location has restricted the amount of dolomite formation and the typical dolostone beds are absent. Some dolomite is however, found at Coldstream, again as a replacement to calcite, in the thin (20 cm) marginal marine bay assemblage, and occasionally without an obvious calcite precursor as individual rhombs in the shallow marine nearshore assemblage. In comparison, other restricted sites in the Tweed Basin (Burnmouth and Norham) have dolostone beds composed of authigenic dolomite and are interpreted as lacustrine (Bennett *et al*, 2016).

- TOC values are consistent with other Early Carboniferous data, but the $\delta^{13}\text{C}$ org. values are generally less negative than published data. Values from both sites vary from expected patterns. At Coldstream specific values track the bulk values very closely inferring that terrestrial organic material dominates the bulk values. At Coquetdale the specific values are very variable and include many data points more negative than the bulk value equivalents. Specific $\delta^{13}\text{C}$ org. values may have been affected by plant type and preservation potential, or something, as yet unresolved, affecting the carbon cycle. This deserves further investigation.
- Since all recovered tetrapod material seems to have been transported, even if only short distances, further searches for tetrapod fossils may be best pursued in rocks interpreted to have been deposited as fine-grained (silt - fine sand grade) marginal marine or brackish water bay fill assemblages or in lagoons, depositional locations close to shore. Coarser floodplain deposits are also another possible source.

Appendix A - Sites images

Coldstream



Figure A.1: The site at Coldstream, with location of extracted block. View northeast, left bank of River Tweed. Hammer for scale.

Coquetdale

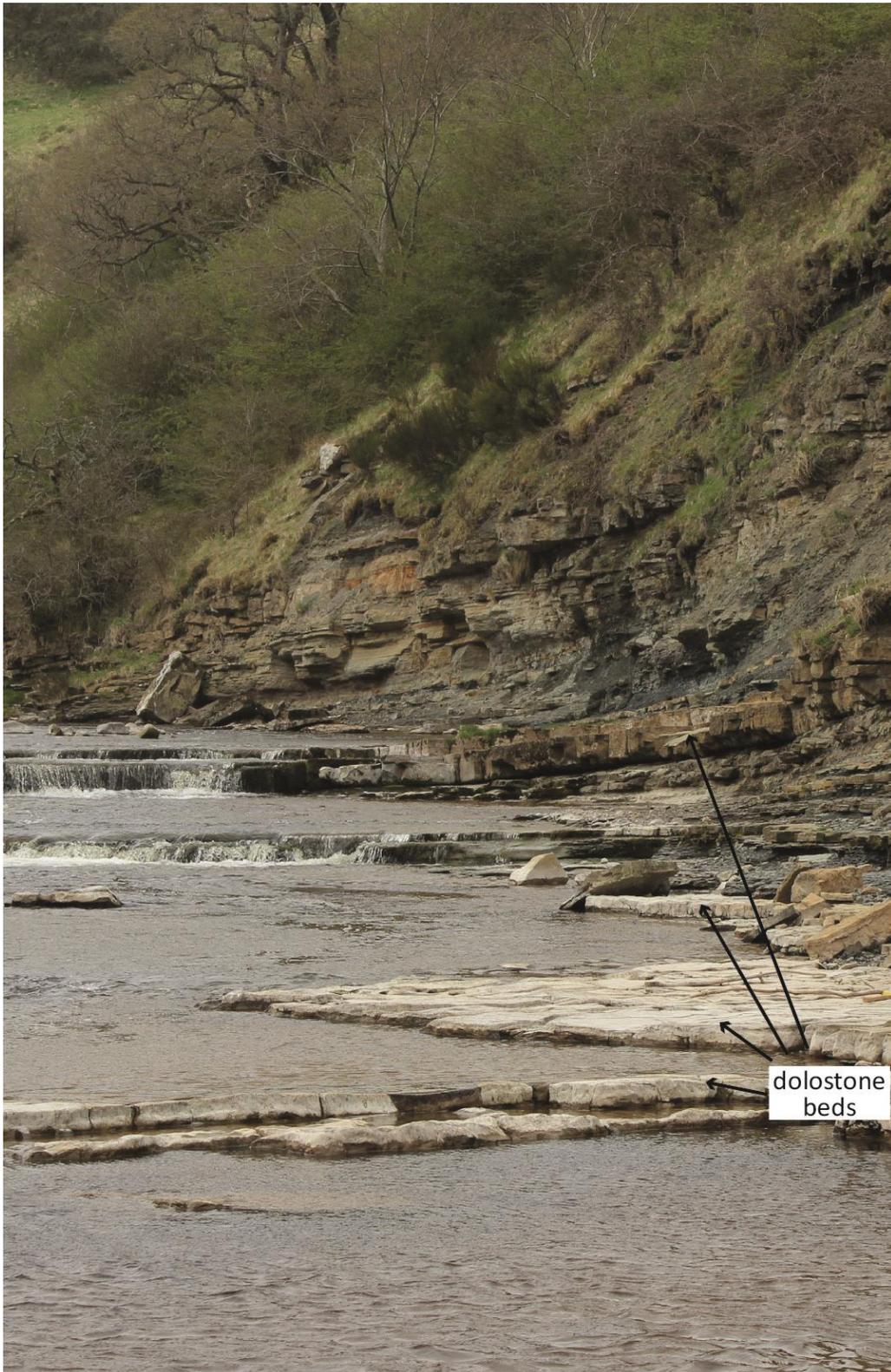


Figure A.2: The site at Coquetdale. View upstream, to west. Exposed bedding planes in the foreground are dolostone beds; bay fill assemblage between, and vegetated floodplain assemblage above. Upper beds are shallow marine near-shore assemblage overlain by dolostones and a return to the shallow marginal marine bay assemblage.

Whitrope Burn



Figure A.3: The site at Whitrope Burn. View downstream, to southwest. Dolostone beds in the foreground. Siltstone beds are exposed slightly downstream in the opposite bank. Image courtesy of T. Smithson.

Appendix B - XRD results

Coldstream

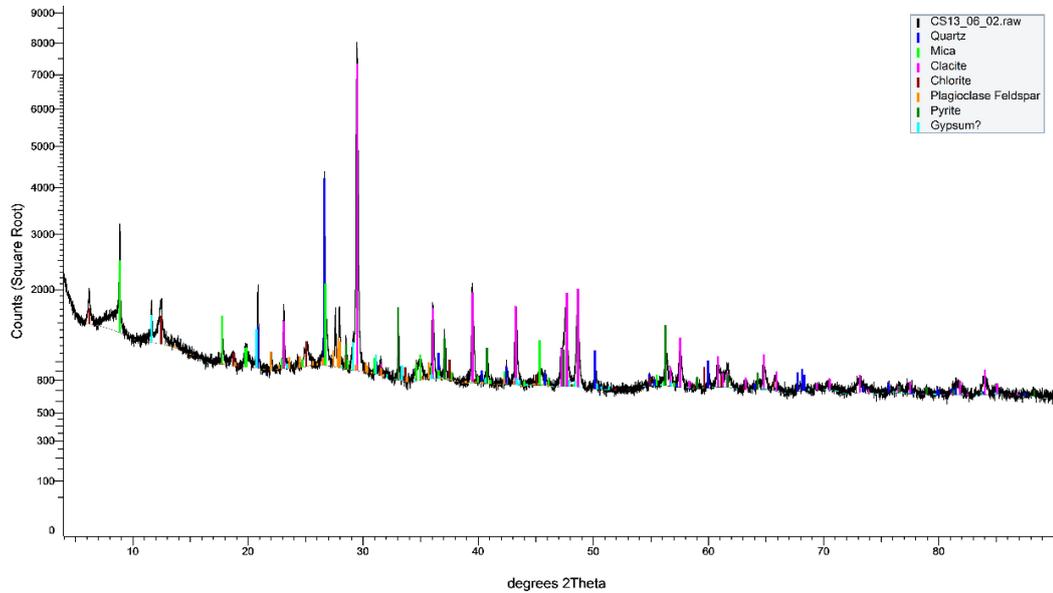


Figure B.1: Coldstream: XRD trace at 36 cm on log. The colours illustrate different minerals present in the sample. Note: the size of peaks does not correlate with relative amounts of individual minerals.

Coquetdale

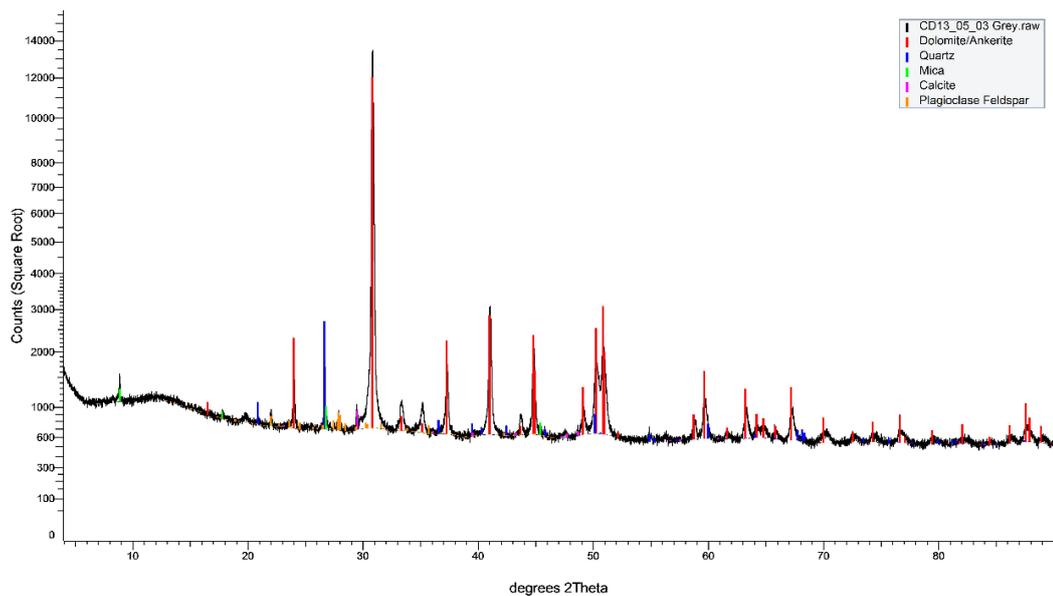


Figure B.2: Coquetdale: XRD trace at 40 cm on log.

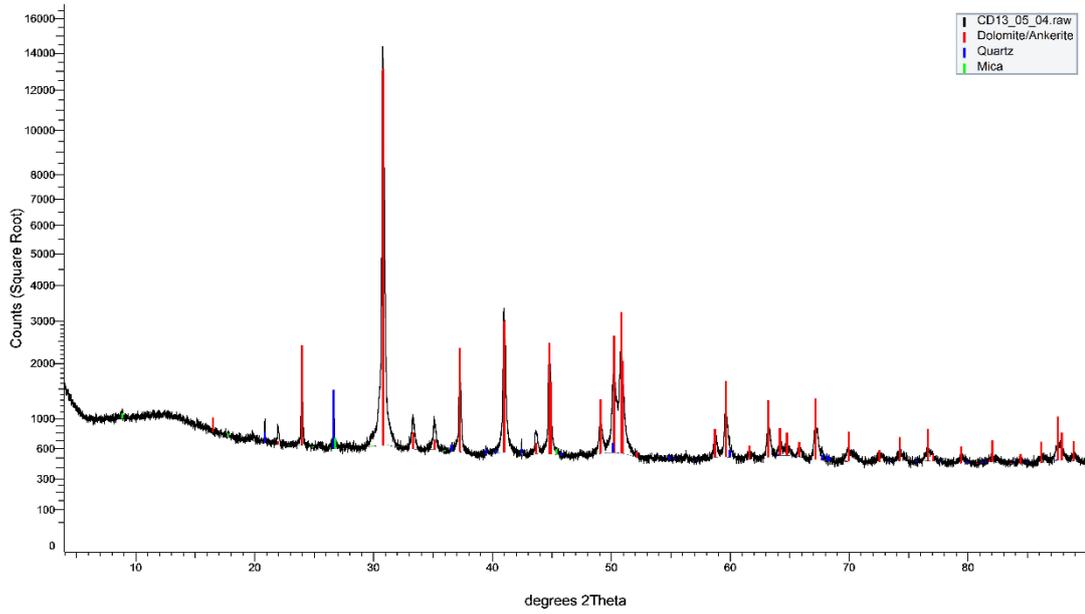


Figure B.3: Coquetdale: XRD trace at 55 cm on log.

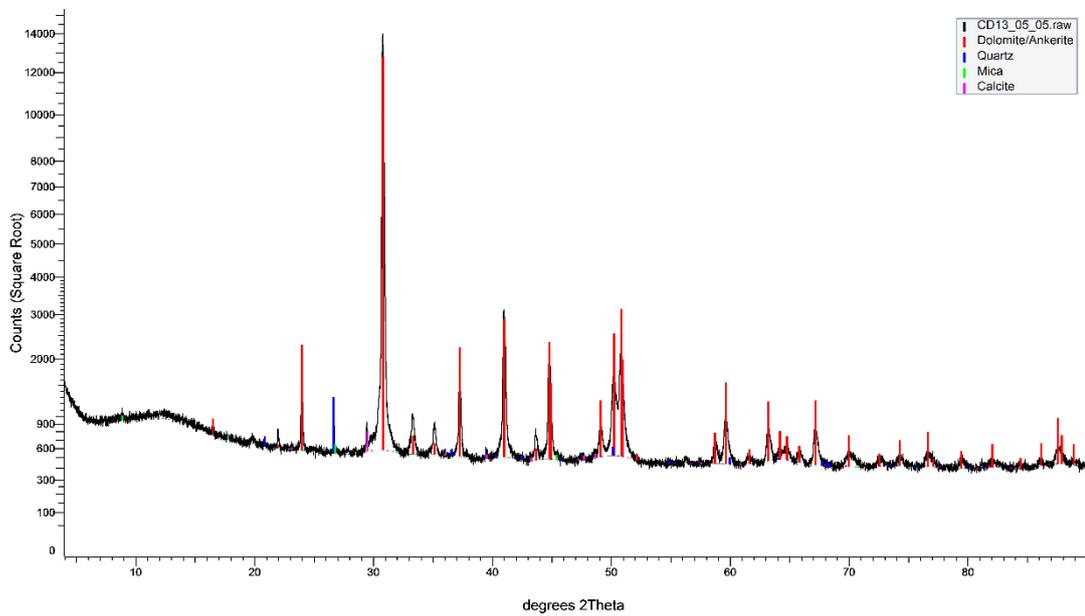


Figure B.4: Coquetdale: XRD trace at 70 cm on log.

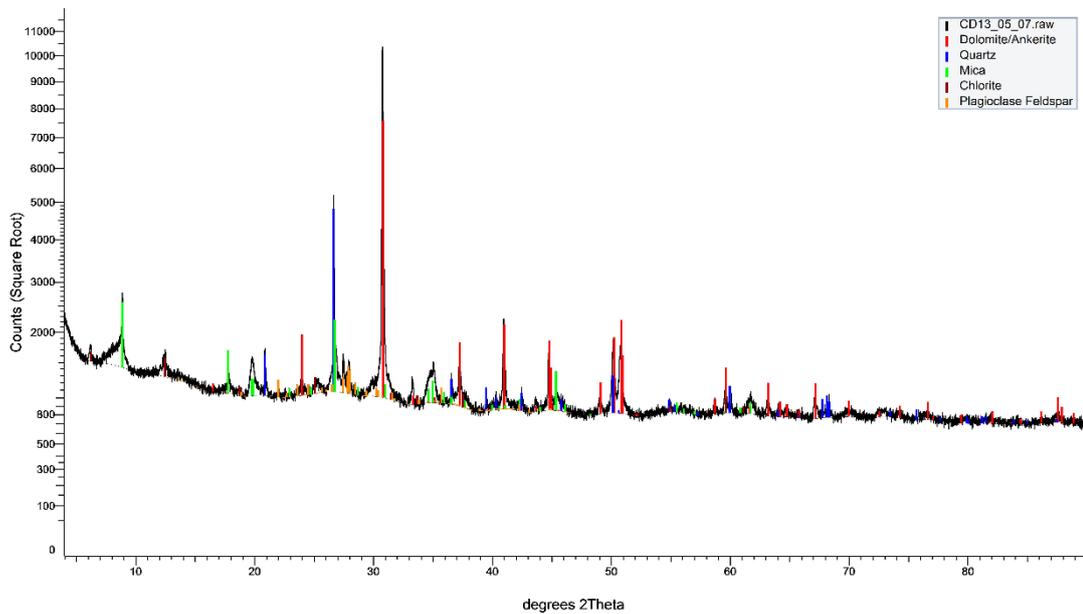


Figure B.5: Coquetdale: XRD trace at 80 cm on log.

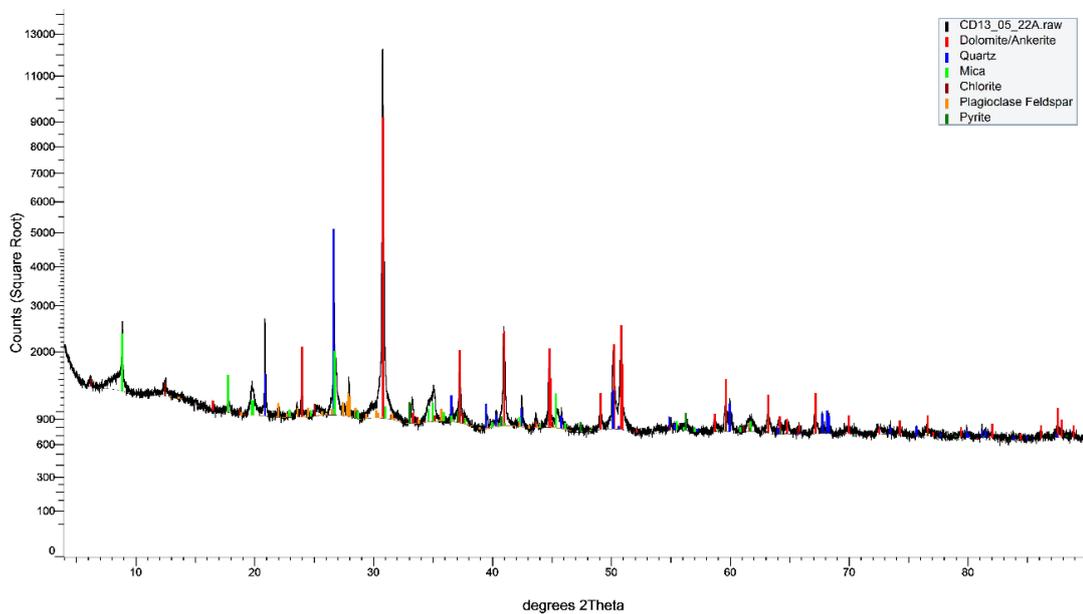


Figure B.6: Coquetdale: XRD trace at 365 cm on log.

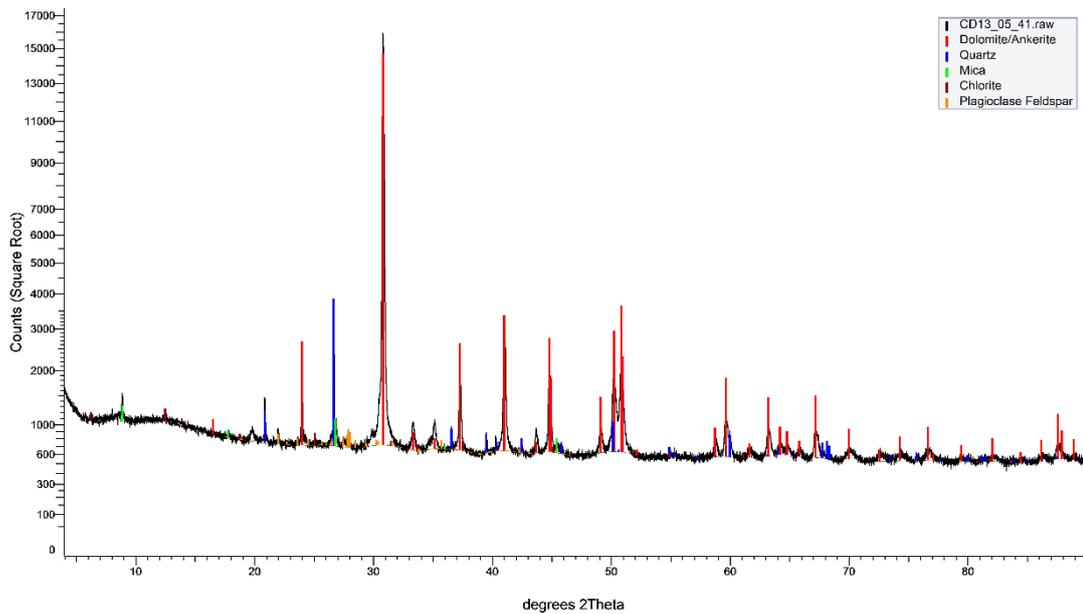


Figure B.7: Coquetdale: XRD trace at 890 cm on log.

Whitrope Burn

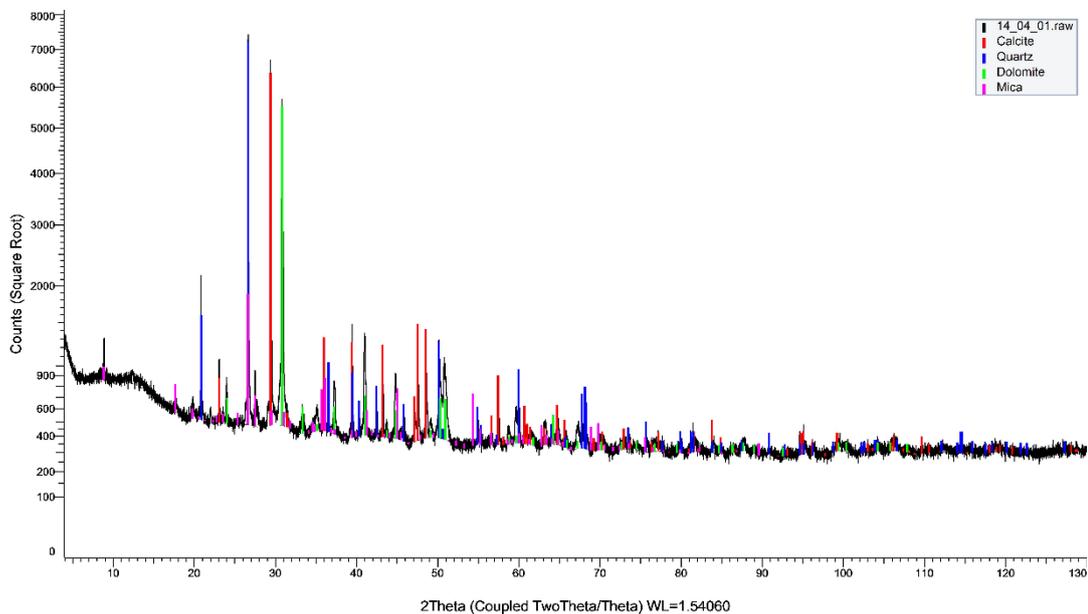


Figure B.8: Whitrope Burn: XRD trace at 3 cm on log.

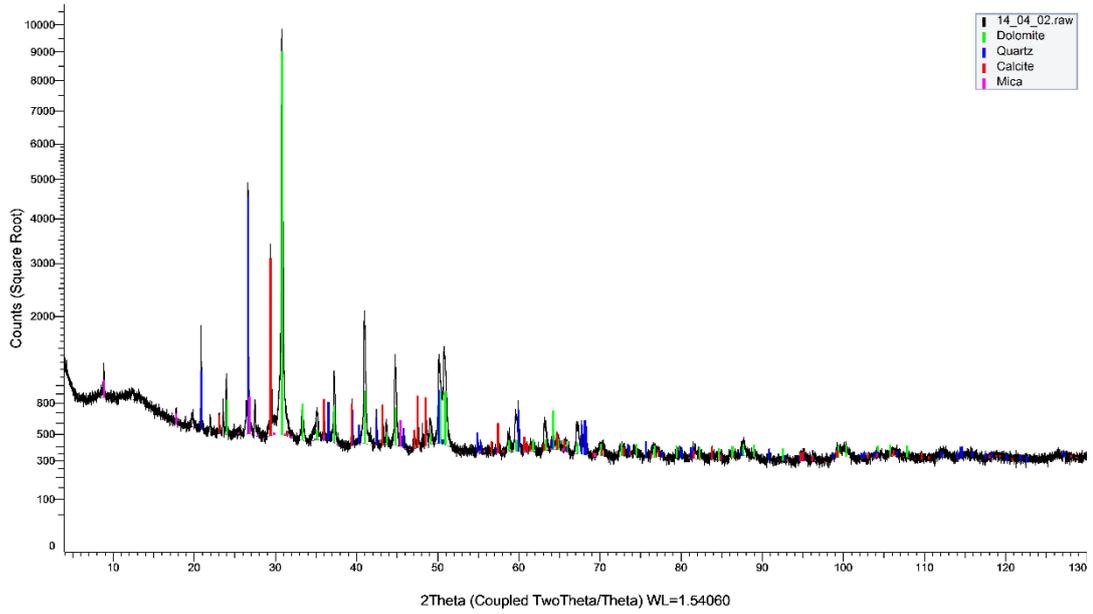


Figure B.9: Whitrope Burn: XRD trace at 20 cm on log.

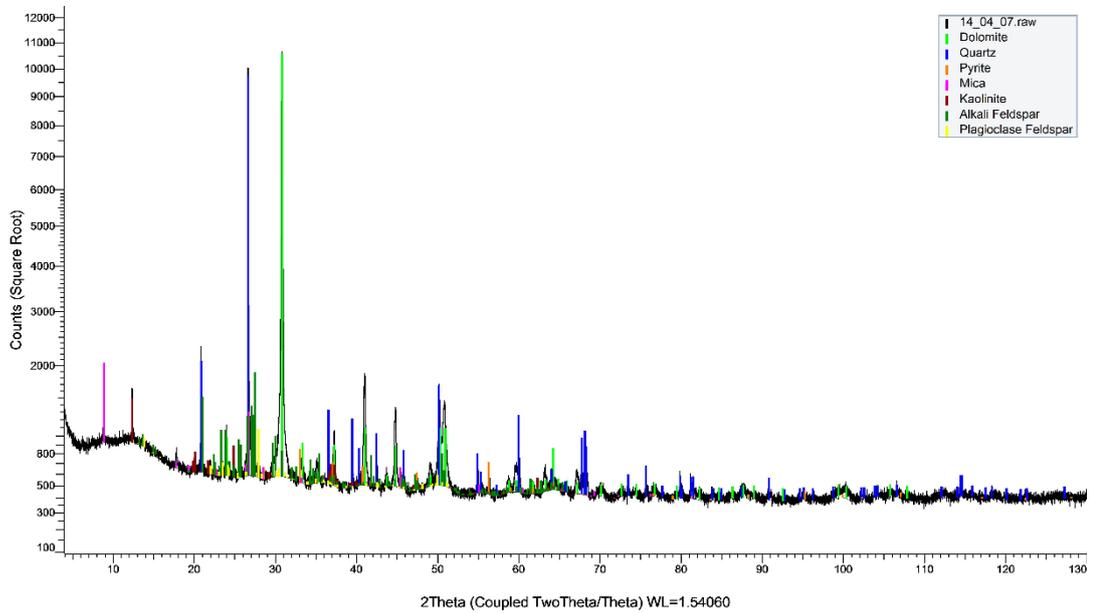


Figure B.10: Whitrope Burn: XRD trace at 60 cm on log.

Appendix C - Coldstream block for detailed study.

Block taken from 33-40 cm on log.

This block was extracted for detailed examination because it is from a horizon which contains lungfish material.

This mudstone block was very fragile so was wrapped in foil to reduce the effects of drying out and splitting. Once back in the labs it was photographed to provide a reference, and then cut with a band saw to reveal the un-weathered internal structure. One cut surface was then consolidated with 'Paraloid B' prior to cutting for blocks for thin sections. Thin section blocks were selected so that they covered all the beds. Six ultrathin (20 μm) thin sections were made.



Figure C.1: Coldstream block for detailed study, vertical section, from 33-40 cm on log. Letters indicate described beds. Block is about 30 cm across.

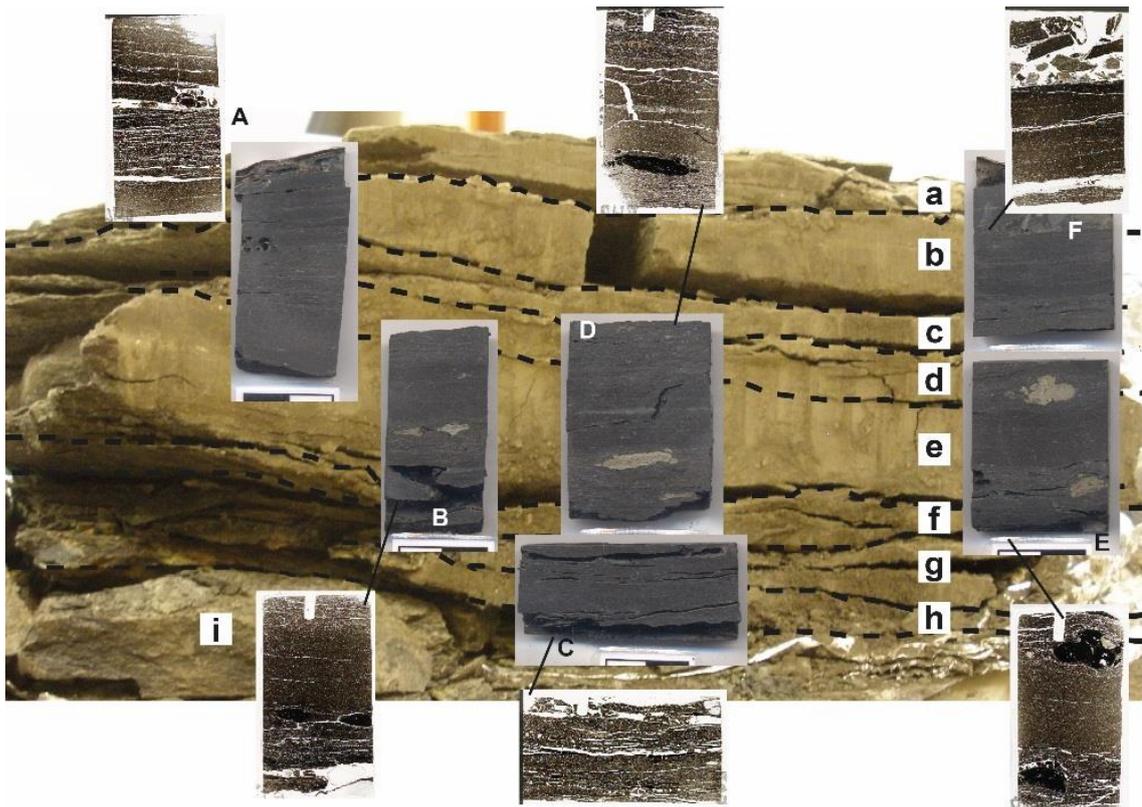


Figure C.2: Coldstream block, indicating beds (a-h) and location and nature of thin sections and polished blocks. Scale: thin section A (top left) is 2 cm wide.

Description and interpretation of block.

The block naturally divided into nine individual beds which are described below. Fossil finds from each bed are detailed in the table in the main text (Table 3.1). Beds were stripped off from the top of the block, one by one, and examined under a binocular microscope. Since it was visually obvious that ostracods, plant, bone and pyrite nodules were unequally distributed throughout the aim was to record the lithology and type of fossil fragments within each bed in the expectation that this would suggest reasons for the presence of bone material at this horizon. The following is an assessment of each bed, starting from the base.

Bed i, the lowest bed, appears to be rather different from the overlying beds and probably represents the bed below this main, plant-rich block. This idea is supported by the fact that bed **i** contains only one plant fragment, whereas the main part of the block is rich in plant material and is underlain by large plant straps.

Ostracods are abundant but are mainly small (0.5 – 1 mm) and are now moulds only; slightly larger ostracods (up to 2 mm) are present and these retain parts of thin, fractured carapaces. This suggests that the smallest ostracods had very thin carapaces and that these have dissolved post deposition. Towards the base of this bed ostracods are abundant and have retained their carapaces, perhaps suggesting that (acidic?) diagenetic fluids from overlying, plant-rich, beds have not infiltrated as far as the base of the bed. Spirorbiform microconchids ('spirorbids') are common, and bivalves of the genus *Modiolus* are also present, each of these retains at least parts of their shells, probably because they are thicker.

At the time of deposition the bay must have been sufficiently oxygenated to support the abundance of ostracods but not perhaps their development, hence the presence of predominantly small ostracods. Alternatively, the preponderance of small ostracods may be a result of winnowing during deposition. There is little in the way of other faunal material. Plant material within the bed is scarce but is represented by fragments of strap-like leaves.

The whole block is underlain by large pieces of strap-like leaves, these were likely washed into position by the increased fluvial input of a river in flood (see the interpretation of the bay-fill assemblage), as represented by the slightly coarser facies 1 on the log. The strap-like leaves may have been brought into the basin and then settled down after the previous flow as they would likely have floated for longer than the siliclastic input. These would then underlie the sedimentation of this particularly fossiliferous block.

Bed h contains only plant material, both strap-like and pieces with nodes, these are similar to the nodes on *Stigmaria* but are more subtle and less regular. There are also many shiny imprints of plant material. The lack of any fauna, including ostracods, renders this bed unique within the block and suggests that conditions at the time of deposition were somewhat inhospitable; probably due to microbial degradation of the plant material present and the resulting reduced oxygen concentration in the water. There is no lamination in this bed and the base of the sample undulates, suggesting that the deposit rapidly infilled an uneven surface.

Beds g - c contain abundant plant material, particularly in the lowest part of bed **e** where the base is composed almost entirely of compressed plant material. All beds contain ostracods and spirorbids; beds **f**, **d** and **c** also contain bone fragments, the scales of rhizodonts and actinopterygians, and bed **c** contains some lungfish material. Ostracods are particularly abundant in beds **f**, **d**, and **c**, and the base of bed **e**, and spores are common in bed **f**, in association with the plant material.

The main part of bed **e** is interesting in that it is darker and very fine-grained, a micaceous mudstone, (Figs. 3.2C; C1). Ostracods are still common but are generally very small (occasional larger ones are up to 2 mm in size). Other fossil fauna are absent from this bed apart from occasional 'spirorbids'. However, this central part of the bed does contain occasional pale grey nodules, these appear to be slightly siltier lenses which have been compressed and cemented, with one nodule measuring 8.5 cm from its centre to the tapered edge. Similar, but smaller, grey nodules are present in bed **d**.

Both the upper and lower edges of the fine-grained centre of bed **e** are gradational, with the lower edge fining up and the upper edge coarsening up; the slightly coarser edges of the bed contain a lot of fossil material (Figs. 3.2C; C2). The gradational nature of the contact between some beds makes it difficult to split them apart, it may be that what has been described as the base of bed **e** should really be ascribed to the top of bed **f**.

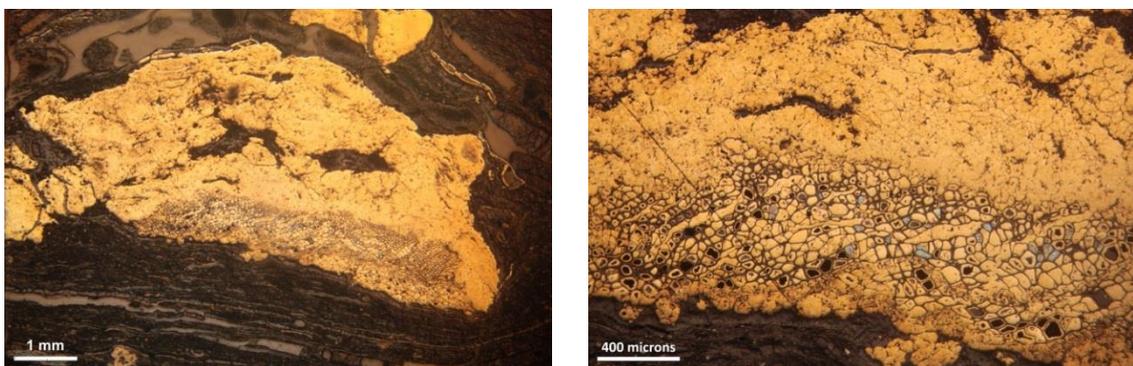


Figure C.3: Pyrite nodules in the coarser edges either side of bed **e**; it is suggested that this is pyritised plant material.

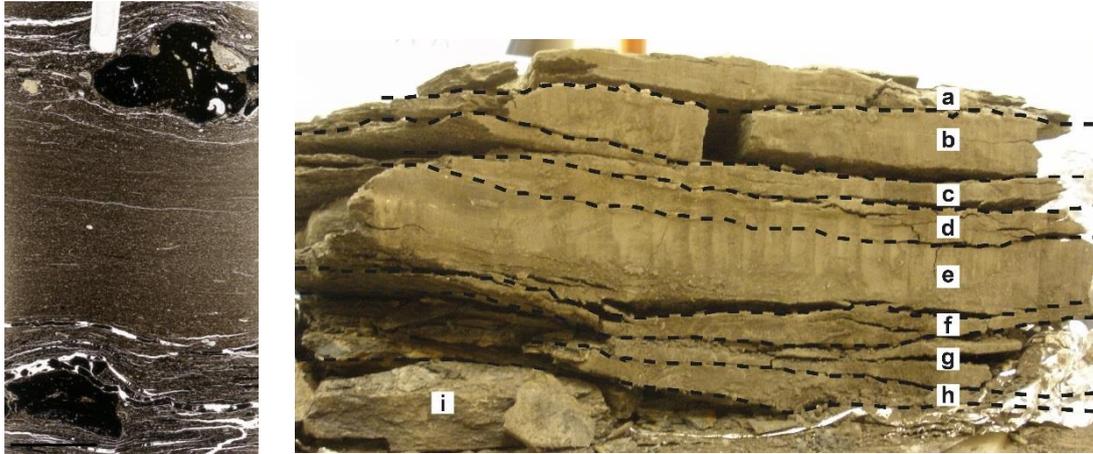


Figure C.4: Thin section E (1.8cm wide) of fine-grained central bed e with coarser laminae either side, containing faunal fragments and pyrite nodules.

Beds **d**, and the lower part of bed **e**, contain a lot of plant and faunal fossil material (Fig. C2); these beds also contain large (up to 15 mm) pyritised nodules, as seen in thin sections. The pyrite nodules contain uncrushed plant material and fossil fragments. Beds **f** and **g** also contain some pyritised ostracods and 'spirorbids'. The presence of pyrite coincides with the greatest abundance of plant material and fossil fragments. At its simplest, it is suggested that the pyrite formed diagenetically under reducing conditions caused by microbial sulphate reduction of the plant material (Deer et al., 2013). Pyrite formation occurs in marine environments and usually in shallow water; here pyrite formation must have occurred soon after deposition because the pyrite nodules and the pyritised ostracods and 'spirorbids' are uncrushed. It is suggested that pyritisation occurred early and easily as reducing fluids circulated easily through the loosely compacted sediment, held open by the abundance of fossil fragments. Towards the top of this block, beds **c** to **a**, bivalves of the genus *Modiolus* are present. *Modiolus* is regarded as being tolerant of a wide range of environmental conditions but here, in association with 'spirorbids', it is interpreted as being quasi-marine/brackish. It is in these beds (**c – a**) that lungfish material and an *Ageleodus* tooth (bed **b**) were found; these species also suggest a marginal marine/brackish/non-marine environment.

These upper beds contain abundant ostracods, 'spirorbids' are common, and both rhizodont and actinopterygian scales are present. Rhizodonts are found in fresh to brackish water environments (Carpenter et al., 2014), and actinopterygians are found in facies ranging from freshwater to marine (Falcon-Lang et al. 2006; Štamberg 2006; Carpenter et al. 2011). The tetrapod skull bones and vertebrae, *Gyracanthus* spines and scorpion fragments found in these beds (Smithson et al., 2012) are also deemed to be of brackish to marginal marine origin. Tetrapods were likely to have lived in fresh to brackish water and were just becoming terrestrial at this time (Clack, 2002); *Gyracanthus* species inhabited brackish to marginal marine environments (Carpenter et al., 2014).

The ostracods found in this block include *Shemonaella* sp., *Cavellina* (possibly *C. incurvescens* and *C. coela*) and *Carbonita* (C. Bennett pers. comm.). The first two are marginal marine to brackish water species and *Carbonita* is a predominantly freshwater species tolerant of slightly brackish water (Gray, 1988; Williams et al., 2005; Bennett et al., 2012). A butterfly mould of an ostracod in bed **f** suggests that the abundant ostracods present on and between the plant material had not been transported far. The abundance of plant material and spores suggests a location close to shore, and the presence of pyrite, common in shallow marine and marginal environments, supports this.

All these factors suggest a marginal marine, shallow water environment close to shore, possibly a protected bay environment, which was generally home to a rich variety of species - rhizodonts, actinopterygians, lungfish, eurypterids, 'spirorbids', bivalves and ostracods. Initially (bed **i**) sedimentation was fine-grained, with only small ostracods, spirorbids and some bivalves. An increasing input of plant material and spores, a piece of eurypterid carapace (Hibbertopteroidea, A. Ross, pers. comm.) which had likely been transported with the plant material, ostracods and 'spirorbids' (beds **g** – **f**) suggest greater flow into the basin (bay) either due to storm input (more plant material), or progradation of the hinterland. The proximity to the shore, or the increased siliciclastic input and the oxidation of the water, provided an environment conducive to colonisation by a rich variety of species, and sometimes their remobilisation and concentration into certain beds.

Intermittently sedimentation tailed off and finer-grained sediment (bed **e**) was deposited, again restricting the size of the ostracods either by sorting or limiting conditions conducive to their development. This was followed by increased sedimentation again and more fossil material was introduced into the basin (bed **d**). The presence of pyrite below (and above) the fine-grained central bed (**e**) is interpreted as being due to the abundance of plant material and its rapid microbial reduction.

Beds **c** – **a** indicate increasingly marginal/brackish conditions as a freshwater source, a fluvial system, prograded and the location became more proximal.

This sample is envisaged as being part of a bay fill assemblage. The nature of the fauna present suggests a restricted, marginal marine environment.

Appendix D - Carbon isotope values

Bulk and specific carbon isotope values ($\delta^{13}\text{C}$) in ‰, and % TOC are detailed in the table below for the main Coquetdale section, the short nearby 'plant bed' section at Coquetdale, and the section at Coldstream.

Coquetdale	height (cm)	Bulk $\delta^{13}\text{C}$	Specific $\delta^{13}\text{C}$	%C
CD13/05/10	128	-21.78	-23.59	0.92
CD13/05/11	176		-22.04	
CD13/05/12	187	-21.74	-22.50	1.31
CD13/05/15A	238	-21.56	-21.47	0.33
CD13/05/18A	294	-21.49	-21.33	0.21
CD13/05/24	413	-21.37	-21.92	0.73
CD13/05/25	456	-21.52	-21.07	0.87
CD13/05/26	465	-21.35	-21.96	0.91
CD13/05/27	489	-21.29	-20.14	0.11
CD13/05/28	500	-21.28	-21.14	0.37
CD13/05/30 A	546	-21.12		0.55
CD13/05/30 B	546	-21.28	-23.90	0.30
CD13/05/30 C	546	-21.23		0.40
CD13/05/31	595	-21.20	-22.35	0.31
CD13/05/33	654	-21.51	-24.40	0.16
CD13/05/35	743	-21.14	-22.19	0.47
CD13/05/37	772	-21.90	-21.63	0.29
CD13/05/38	817	-21.78	-21.28	0.64
CD13/05/39	871	-21.30	-22.10	0.59
CD13/05/43	924	-21.87	-24.84	0.21
Coquetdale plant bed	height (cm)	Bulk $\delta^{13}\text{C}$	Specific $\delta^{13}\text{C}$	%C
CD-PB 14/04/02	21	-21.77	-23.20	1.03
CD-PB 14/04/03	42	-21.86	-22.13	1.55
CD-PB 14/04/13	50	-22.03	-22.21	4.81
CD-PB 14/04/14	52	-22.69	-23.54	15.49
CD-PB 14/04/15	57	-22.13	-21.86	3.22
CD-PB 14/04/16	62	-21.24	-21.15	1.62
CD-PB 14/04/17	73	-21.09	-22.80	1.41
CD-PB 14/04/20	86	-20.83	-20.67	1.12
Coldstream	height (cm)	Bulk $\delta^{13}\text{C}$	Specific $\delta^{13}\text{C}$	%C
CS13/06/01	10	-22.17	-21.80	1.01
CS13/06/02	36	-24.51	-22.91	8.9
CS13/06/08	79	-22.50	-22.24	1.73
CS13/06/03	84	-22.72	-22.06	1.50
CS13/06/09	110	-21.94	-21.84	0.63
CS13/06/10	125	-22.02		0.75
CS13/06/11	126	-22.01	-22.40	0.97
CS13/06/12	129	-21.96		0.60
CS14/06/03	185	-22.82	-22.40	3.50
CS14/06/06	277	-21.99	-21.61	0.60
CS13/06/13	300	-24.81	-24.42	0.89

Table D.1: Details of bulk and specific carbon isotope values (per mil) and % TOC for Coquetdale, Coquetdale plant bed and Coldstream sections. $\delta^{13}\text{C}$ values in relation to VPDB.

Bibliography

- Al-Awadi, M., Clark, W.J., Moore, W.R., Herron, M., Zhang, T., Zhao, W., Hurley, N., Kho, D., Montaron, B. and Sadooni, F. (2009) 'Dolomite: Perspectives on a Perplexing Mineral', *Oilfield Review*: 21, no. 3, pp. 32-45.
- Amorosi, A., Colalongo, M., Pasini, G. and Preti, D. (1999) 'Sedimentary response to Late Quaternary sea-level changes in the Romagna coastal plain (northern Italy)', *Sedimentology*, 46(1), pp. 99-121.
- Anderton, R. (1985) 'Sedimentology of the Dinantian of Foulden, Berwickshire, Scotland', *Transactions of the Royal Society of Edinburgh: Earth Sciences*, 76, pp. 7-12.
- Andrews, J.E., Hamilton, P.J. and Fallick, A.E. (1987) 'The geochemistry of early diagenetic dolostones from a low-salinity Jurassic lagoon', *Journal of the Geological Society*, 144(5), pp. 687-698.
- Andrews, J.E. and Nabi, G. (1998) 'Palaeoclimatic significance of calcretes in the Dinantian of the Cockburnspath Outlier (East Lothian-North Berwickshire)', *Scottish Journal of Geology*, 34, Part 2, pp. 153-164.
- Andrews, J.E. and Nabi, G. (1994) 'Lithostratigraphy of the Dinantian Inverclyde and Strathclyde Groups, Cockburnspath Outlier, East Lothian North Berwickshire', *Scottish Journal of Geology*, 30, pp. 105-119.
- Andrews, J.E., Turner, M.S., Nabi, G. and Spiro, B. (1991) 'The Anatomy of an Early Dinantian Terraced Floodplain - Paleoenvironment and Early Diagenesis', *Sedimentology*, 38(2), pp. 271-287.
- Arens, N.C., Jahren, A.H. and Amundson, R. (2000) 'Can C3 plants faithfully record the carbon isotopic composition of atmospheric carbon dioxide?', *Paleobiology* 26(1), pp. 137-164.
- Bak, K., Wolska, A., Zielinska, M. and Bak, M. (2015) 'Coal-bearing submarine slump sediments from Oligocene-Miocene transition of the Eastern Outer Carpathians (Bieszczady Mountains, SE Poland)', *Geological Quarterly*, 59(2), pp. 300-315.
- Baker, P.A. and Kastner, M. (1981) 'Constraints on the formation of sedimentary dolomite', *Science* 213, pp. 214-216.
- Ballèvre, M. and Lardeux, H. (2005) 'Signification paléocéologique et Paléogéographique des bivalves du Carbonifère inférieur du bassin d'Ancenis (Massif Américain)', *Comptes Rendus Palevol*, 4, pp. 109-121.

- Bann, K.L. and Fielding, C.R. (2004) 'An integrated ichnological and sedimentological comparison of non-deltaic shoreface and subaqueous delta deposits in Permian reservoir units in Australia.', in McIlroy, D. (ed.) *The Application of Ichnology to Palaeoenvironmental and Straigraphic Analysis*. Special Publication, 228 edn. London: Geological Society, pp. 273-310.
- Belt, E., Freshney, E. and Read, W. (1967) 'Sedimentology of Carboniferous cementstone facies, British Isles and eastern Canada', *Journal of Geology*, 75(6), pp. 711-711-721.
- Bennett, C.E. (2008) 'A review of the Carboniferous colonisation of non-marine environments by ostracods', *Senckenbergiana lethaea*, 8(1), pp. 37-46.
- Bennett, C.E., Siveter, D.J., Davies, S.J., Williams, M., Wilkinson, I.P., Browne, M. and Miller, C.G. (2012) 'Ostracods from freshwater and brackish environments of the Carboniferous of the Midland Valley of Scotland: the early colonization of terrestrial water bodies', *Geological Magazine*, 149(3), pp. 366-396.
- Bennett, C.E., Kearsley, T.I., Davies, S.J., Millward, D., Clack, J.A., Smithson, T.R. and Marshall, J.E.A. (2016) 'Early Mississippian sandy siltstones preserve rare vertebrate fossils in seasonal flooding episodes', *Sedimentology*, 63, pp. 1677-1700.
- Braddy, S.J. (2001) 'Eurypterid palaeoecology: palaeobiological, ichnological and comparative evidence for a 'mass-moult-mate' hypothesis', *Palaeogeography, Palaeoclimatology, Palaeoecology*, 172(1-2), pp. 115-132.
- Brandon, A., Riley, N.J., Wilson, A.A. and Ellison, R.A. (1995) 'Three new early Namurian (E1c-E2a) marine bands in central and northern England, UK, and their bearing on correlations with the Askrigg Block', *Proceedings of the Yorkshire Geological and Polytechnic Society*, 50(4), pp. 333-355.
- Brenchley, P.J. and Harper, D.A.T. (1998) *Palaeoecology: Ecosystems, environments and evolution*. London: Chapman & Hall.
- Bridge, J., Collier, R. and Alexander, J. (1998) 'Large-scale structure of Calamus River deposits (Nebraska, USA) revealed using ground-penetrating radar', *Sedimentology*, 45(6), pp. 977-986.
- Bridge, J.S., Smith, N.D., Trent, F., Gabel, S.L. and Bernstein, P. (1986) 'Sedimentology and Morphology of a Low-Sinuosity River - Calamus River, Nebraska Sand Hills', *Sedimentology*, 33(6), pp. 851-870.
- Briggs, D.E.G., Clarkson, E.N.K. (1987) 'An enigmatic chordate from the Lower Carboniferous Granton 'shrimp-bed' of the Edinburgh district, Scotland', *Lethaia*, 20(2), pp.107-115.

- Bristow, C.S. (1988) 'Controls on the sedimentation of the Rough Rock Group (Namurian) from the Pennine Basin of northern England.', in Besly, B.M. and Kelling, G. (eds.) *Sedimentation in a Synorogenic Basin Complex: the Upper Carboniferous of Northwest Europe*. Glasgow: Blackie, pp. 114-131.
- Bromley, R.G. (1996) *Trace fossils: biology, taphonomy and applications*. 2nd edn. London: Chapman and Hall.
- Browne, M.A.E. 1980 *The Upper Devonian and Lower Carboniferous (Dinantian) of the Firth of Tay, Scotland.*, Institute of Geological Sciences, London.
- Browne, M.A.E., Dean, M.T., Hall, I.H.S., McAdam, A.D., Monroe, S.K. & Chisholm, J.I. 1999 *A lithostratigraphical framework for the Carboniferous rocks of the Midland Valley of Scotland.*, NERC, British Geological Survey, Keyworth.
- Carpenter, D., Falcon-Lang, H.J., Benton, M.J. and Nelson, W.J. (2011) 'Fishes and tetrapods in the upper Pennsylvanian (Kasimovian) Cohn Coal Member of the Mattoon Formation of Illinois, United States: systematics, paleoecology, and paleoenvironments', *Palaios*, 26(10), pp. 639-657.
- Carpenter, D.K., Falcon-Lang, H.J., Benton, M.J. and Henderson, E. (2014) 'Carboniferous (Tournaisian) Fish Assemblages from the Isle of Bute, Scotland: Systematics and Palaeoecology', *Palaeontology*, 57(6), pp. 1215-1240.
- Clack, J. (2002) 'An early tetrapod from 'Romer's Gap'', *Nature*, 418(6893), pp. 72-76.
- Clack, J. and Finney, S. (2005) 'Pederpes finneyae, an articulated tetrapod from the Tournaisian of western Scotland', *Journal of Systematic Palaeontology*, 2(4), pp. 311-346.
- Davies, S.J. and McLean, D. (1996) 'Spectral gamma-ray and palynological characterization of kinderscoutian marine bands in the namurian of the pennine basin', *Proceedings of the Yorkshire Geological Society*, 51(2), pp. 103-114.
- Edwards, W. and Stubblefield, C.J. (1947) 'Marine bands and other faunal marker-horizons in relation to the sedimentary cycles of the Middle Coal Measures of Nottinghamshire and Derbyshire.', *Quarterly Journal of the Geological Society*, no. 412, pp. 209-260.
- Elliott, T. (1974) 'Interdistributary bay sequences and their genesis', *Sedimentology*, 21, pp. 611-622.
- Eugster, H.P. and Hardie, L.A. (1975) 'Sedimentation in an Ancient Playa-Lake Complex: The Wilkins Peak Member of the Green River Formation of Wyoming', *Geological Society of America Bulletin*, 86(3), pp. 319-334.

- Falcon-Lang, H. (1999) 'Fire ecology of a Late Carboniferous floodplain, Joggins, Nova Scotia', *Journal of the Geological Society of London*, 156, pp. 137-137-148.
- Falcon-Lang, H.J., Benton, M.J., Braddy, S.J. and Davies, S.J. (2006) 'The Pennsylvanian tropical biome reconstructed from the Joggins Formation of Nova Scotia, Canada', *Journal of the Geological Society*, 163, pp. 561-576.
- Falcon-Lang, H. (1998) 'The impact of wildfire on an Early Carboniferous coastal environment, North Mayo, Ireland', *Palaeogeography, Palaeoclimatology, Palaeoecology*, 139(3-4), pp. 121-138.
- Farquhar, G.D., Ehleringer, J.R. and Hubick, K.T. (1989) 'Carbon Isotope Discrimination and Photosynthesis', *Annual Review of Plant Physiology and Plant Molecular Biology*, 40, pp. 503-537.
- Ferreira, T.O., Nobrega, G.N., Albuquerque, A.G.B.M., Sartor, L.R., Gomes, I.S., Artur, A.G. and Otero, X.L. (2015) 'Pyrite as a proxy for the identification of former coastal lagoons in semiarid NE Brazil', *Geo-Marine Letters*, 35(5), pp. 355-366.
- Flügel, E. (2010) *Microfacies of Carbonate Rocks: Analysis, Interpretation and Application*. second edn. Heidelberg: Springer.
- Folk, R.L. (1959) 'Practical petrographic classification of limestones', *AAPG Bulletin*, 43(1), pp. 1-38.
- Friedman, M. and Sallan, L.C. (2012) 'Five hundred million years of extinction and recovery: a phanerozoic survey of large-scale diversity patterns in fishes', *Palaeontology*, 55, pp. 707-742.
- Gregg, J.M. and Sibley, D.F. (1984) 'Epigenetic Dolomitization and the Origin of Xenotopic Dolomite Texture', *Journal of Sedimentary Petrology*, 54(3), pp. 908-931.
- Guernet, C. and Lethiers, F. (1989) 'Ostracodes et recherche des milieux anciens; possibilites et limites', *Bulletin de la Societe Geologique de France*, V(3), pp. 577-588.
- Guion, P.D., Gutteridge, P. and Davies, S.J. (2000) 'Carboniferous sedimentation and volcanism on the Laurussian margin', in Woodcock, N.H. and Strachan, R.A. (eds.) *Geological History of Britain and Ireland*. 1st edn. Oxford, UK: Blackwell Science Ltd, pp. 227-270.
- Hampson, G.J., Procter, E.J. and Kelly, C. (2008) 'Controls on Isolated Shallow-Marine Sandstone Deposition and Shelf Construction: Late Cretaceous Western Interior Seaway, Northern Utah and Colorado, Usa', *Recent Advances in Models of Siliciclastic Shallow-Marine Stratigraphy*, 90, pp. 355-389.

- Hartman, G. and Danin, A. (2010) 'Isotopic values of plants in relation to water availability in the Eastern Mediterranean region', *Oecologia*, 162(4), pp. 837-852.
- Hawkins, K., Davies, S.J., Mullins, G.L. and Macquaker, J.H.S. (2013) 'Miospore distribution and sedimentological facies distribution as an insight to changing terrestrial palaeoequatorial floral communities during a Pennsylvanian glacio-eustatic sea level cycle', *Review of palaeobotany and palynology*, 197(0), pp. 166-178.
- Hudson, J.D. (1963) 'The recognition of salinity-controlled mollusc assemblages in the Great Estuarine Series (Middle Jurassic) of the Inner Hebrides', *Palaeontology*, 6(2), pp. 318-326.
- Hudson, J.D., Clements, R.G., Riding, J.B., Wakefield, M.I. and Walton, W. (1995) 'Jurassic Paleosalinities and Brackish-Water Communities - a Case-Study', *Palaios*, 10(5), pp. 392-407.
- Humphrey, J.D. (2000) 'New geochemical support for mixing-zone dolomitization at Golden Grove, Barbados', *Journal of Sedimentary Research*, 70(5), pp. 1160-1170.
- Illing, L.V., Wells, A.J. and Taylor, J.C.M. (1965) 'Penecontemporary dolomite in the Persian Gulf.', in Pray, L.C. and Murray, R.C. (eds.) *Dolomitization and Limestone Diagenesis*. SEPM Special Publications, pp. 89-111.
- Jeffery, J.E. (2012) 'Cranial morphology of the Carboniferous rhizodontid *Screbinodus ornatus* (Osteichthyes: Sarcopterygii)', *Journal of Systematic Palaeontology*, 10(3), pp. 475-519.
- Kaiser, S.I., Aretz, M. and Becker, R.T. (2015) 'The global Hangenberg Crisis (Devonian-Carboniferous transition): review of a first-order mass extinction.', in Becker, R.T., Konigshof, P. and Brett, C.E. (eds.) *Devonian Climate, Sea Level and Evolutionary Events*. Special Publications, 423, London: The Geological Society of London.
- Kane, I.A. (2010) 'Turbulence, displacement, death and worms: a day in the life of a fluvial Carboniferous bivalve', *Lethaia*, 43(3), pp. 381-395.
- Kearsey, T.I., Bennett, C.E., Millward, D., Davies, S.J., Gowing, C.J.B., Kemp, S.J., Leng, M.J., Marshall, J.E.A. and Browne, M.A.E. (2016) 'The terrestrial landscapes of tetrapod evolution in earliest Carboniferous seasonal wetlands of SE Scotland', *Palaeogeography, Palaeoclimatology, Palaeoecology*, 457, pp. 52-69.
- Kieft, R.L., Hampson, G.J., Jackson, C.A.-. and Larsen, E. (2011) 'Stratigraphic Architecture of a Net-Transgressive Marginal- to Shallow-Marine Succession: Upper Almond Formation, Rock Springs Uplift, Wyoming, Usa', *Journal of Sedimentary Research*, 81(7-8), pp. 513-533.
- Könitzer, S.F., Davies, S.J., Stephenson, M.H. and Leng, M.J. (2014) 'Depositional Controls on Mudstone Lithofacies in a Basinal Setting: Implications for the

- Delivery of Sedimentary Organic Matter', *Journal of Sedimentary Research*, 84(3-4), pp. 198-214.
- Lamsdell, J.C. and Braddy, S.J. (2010) 'Cope's Rule and Romer's theory: patterns of diversity and gigantism in eurypterids and Palaeozoic vertebrates', *Biology Letters*, 6(2), pp. 265-269.
- Leeder, M. (1974) 'Lower Border Group (Tournaisian) fluvio-deltaic sedimentation and palaeogeography of the Northumberland Basin', *Proceedings of the Yorkshire Geological Society*, 40, part 2, no. 9, pp. 129-180.
- Lewan, M.D. (1986) 'Stable Carbon Isotopes of Amorphous Kerogens from Phanerozoic Sedimentary-Rocks', *Geochimica et Cosmochimica Acta*, 50(8), pp. 1583-1591.
- MacEachern, J.A., Pemberton, S.G., Gingras, M.K. and Bann, K.L. (2007) 'The Ichnofacies Paradigm: A Fifty-Year Retrospective.', in Miller, W.C. (ed.) *Trace Fossils. Concepts, Problems, Prospects*. first edn. Oxford: Elsevier, pp. 52-77.
- Mack, G.H., James, W.C. and Monger, H.C. (1993) 'Classification of paleosols', *Geological Society of America Bulletin*, 105(2), pp. 129-136.
- Marszalek, D.S. (1975) 'Calcsphere ultrastructure and skeletal aragonite from the alga *Acetabularia antillana*', *Journal of Sedimentary Research*, 45(1), pp. 266-271.
- McKenzie, J.A. (1981) 'Holocene Dolomitization of Calcium Carbonate Sediments from the Coastal Sabkhas of Abu Dhabi, U.A.E.: A Stable Isotope Study', *The Journal of geology*, 89(2), pp. 185-198.
- Monaghan, A.A., 2014 *The Carboniferous shales of the Midland Valley of Scotland: geology and resource estimation*, British Geological Survey for the Department of Energy and Climate Change, London, UK.
- Moore, R.C. (1961) *Treatise on Invertebrate Paleontology, Part Q, Arthropoda 3*. Kansas, USA: Geological Society of America and University of Kansas.
- Nelson, R.A. (1979) 'Natural Fracture Systems - Description and Classification', *Aapg Bulletin-American Association of Petroleum Geologists*, 63(12), pp. 2214-2221.
- Nichols, G. and Jones, T. (1992) 'Fusain in Carboniferous Shallow Marine-Sediments, Donegal, Ireland - the Sedimentological Effects of Wildfire', *Sedimentology*, 39(3), pp. 487-502.
- North, C.P. and Davidson, S.K. (2012) 'Unconfined alluvial flow processes: Recognition and interpretation of their deposits, and the significance for palaeogeographic reconstruction', *Earth-Science Reviews*, 111(1-2), pp. 199-223.
- Pearson, N.J., Mangano, M.G., Buatois, L.A., Casadio, S. and Raising, M.R. (2013) 'Environmental variability of *Macaronichnus* ichnofabrics in Eocene tidal-

- embayment deposits of southern Patagonia, Argentina', *Lethaia*, 46(3), pp. 341-354.
- Pemberton, S.G., MacEachern, J.A., Gingras, M.K. and Saunders, T.D.A. (2008) 'Biogenic chaos; cryptobioturbation and the work of sedimentologically friendly organisms', *Palaeogeography, Palaeoclimatology, Palaeoecology*, 270(3-4), pp. 273-279.
- Pemberton, S.G., Spila, M., Pulham, A.J., Saunders, T., MacEachern, J.A., Robbins, D. and Sinclair, I.K. (2001) *Ichonology and Sedimentogy of Shallow to Marginal Marine Systems: Ben Nevis & Avalon reservoirs, Jeanne D'arc Basin*. St John's, Newfoundland: Geological Association of Canada.
- Pemberton, S.G. and Gingras, M.K. (2005) 'Classification and characterizations of biogenically enhanced permeability', *AAPG Bulletin*, 89(11), pp. 1493-1517.
- Peters-Kottig, W., Strauss, H. and Kerp, H. (2006) 'The land plant $\delta^{13}C$ record and plant evolution in the Late Palaeozoic', *Palaeogeography Palaeoclimatology Palaeoecology*, 240(1-2), pp. 237-252.
- Plink-Bjorklund, P. (2008) 'Wave-to-tide facies change in a Campanian shoreline complex, Chinmey Rock Tongue, Wyoming-Utah, U.S.A.', in Hampson, G.J., Steel, R.J., Burgess, P.M. and Dalrymple, R.W. (eds.) *Recent Advances in Models of Siliciclastic Shallow-Marine Stratigraphy*. First edn. Tulsa, U.S.A.: SEPM, pp. 265-292.
- Plint, A.G., Macquaker, J.H.S. and Varban, B.L. (2012) 'Bedload Transport of Mud Across a Wide, Storm Influenced Ramp: Cenomanian-Turonian Kaskapau Formation, Western Canada Foreland Basin', *Journal of Sedimentary Research*, 82(11-12), pp. 801-822.
- Plummer, P.S. and Gostin, V.A. (1981) 'Shrinkage Cracks - Desiccation Or Synaeresis', *Journal of Sedimentary Petrology*, 51(4), pp. 1147-1156.
- Raiswell, R. (1982) 'Pyrite texture, isotopic composition and the availability of iron', *American Journal of Science*, 282(8), pp. 1244-1263.
- Raychaudhuri, I., Brekke, H.G., Pemberton, S.G. and MacEachern, J.A. (1992) Depositional Facies and Trace Fossils of a Low Wave Energy Shoreface Succession, Albian Viking Formation, Chigwell Field, Alberta, Canada, in Pemberton S.G. (ed), *Applications of Ichonolgy to Petroleum Exploration: A Core Workshop*. Tulsa, U.S.A.: SEPM, 17, pp. 319-337.
- Reineck, H.-E. and Singh, I.B. (1980) *Depositional Sedimentary Environments*. Second edn. New York: Springer-Verlag.
- Retallack, G.J. (2001) *Soils of the Past: An Introduction to Paleopedology*. second edn. Oxford: Blackwell Science Ltd.

- Retallack, G.J. (1988) 'Field Recognition of Paleosols', *Geological Society of America Special Papers*, 216, pp. 1-20.
- Rex, G.M. and Scott, A.C. (1987) 'The sedimentology, palaeoecology and preservation of the Lower Carboniferous plant deposits at Pettycur, Fife, Scotland', *Geological Magazine*, 124(1), pp. 43-66.
- Richards, K.R., Sherwin, J.E., Smithson, T.R., Bennion, R.F., Davies, S.J., Marshall, J.E.A. and Clack, J.A. 'Diverse and durophagous: early Carboniferous chondrichthyans from the Scottish Borders', in review.
- Romer, A.S. (1956) 'The Early Evolution of Land Vertebrates', *Proceedings of the American Philosophical Society*, 100(3), pp. 157-167.
- Rygel, M.C. and Gibling, M.R. (2006) 'Natural Geomorphic Variability Recorded in a High-Accommodation Setting: Fluvial Architecture of the Pennsylvanian Joggins Formation of Atlantic Canada', *Journal of Sedimentary Research*, 76(11), pp. 1230-1251.
- Saltzman, M.R., Groessens, E. and Zhuravlev, A.V. (2004) 'Carbon cycle models based on extreme changes in $\delta^{13}\text{C}$: an example from the lower Mississippian', *Palaeogeography, Palaeoclimatology, Palaeoecology*, 213(3-4), pp. 359-377.
- Sambrook Smith, G.H., Ashworth, P.J., Best, J.L., Woodward, J. and Simpson, C.J. (2006) 'The sedimentology and alluvial architecture of the sandy braided South Saskatchewan River, Canada', *Sedimentology*, 53(2), pp. 413-434.
- Schieber, J., Southard, J. and Thaisen, K. (2007) 'Accretion of mudstone beds from migrating floccule ripples', *Science*, 318(5857), pp. 1760-1763.
- Schieber, J. (1998) 'Possible indicators of microbial mat deposits in shales and sandstones: examples from the Mid-Proterozoic Belt Supergroup, Montana, U.S.A.', *Sedimentary Geology*, 120(1-4), pp. 105-124.
- Schultze, H. (2009) 'Interpretation of marine and freshwater paleoenvironments in Permo-Carboniferous deposits', *Palaeogeography Palaeoclimatology Palaeoecology*, 281(1-2), pp. 126-136.
- Scotese, C.R. and McKerrow, W.S. (1990) 'Revised World maps and introduction', *Geological Society, London, Memoirs*, 12(1), pp. 1-21.
- Scott, A.C. (2010) 'Charcoal recognition, taphonomy and uses in palaeoenvironmental analysis', *Palaeogeography, Palaeoclimatology, Palaeoecology*, 291(1-2), pp. 11-39.
- Scott, W. (1986) 'Nodular Carbonates in the Lower Carboniferous, Cementstone Group of the Tweed Embayment, Berwickshire - Evidence for a Former Sulfate Evaporite Facies', *Scottish Journal of Geology*, 22, pp. 325-345.

- Searl, A. (1990) 'Dolomitisation of the Ardross Limestones (Dinantian), East Fife, Scotland', *Sedimentary Geology*, 69(1-2), pp. 77-94.
- Sequeira, S.E.K. and Coates, M.I. (2000) 'Reassessment of 'Cladodus' neilsoni Traquair: a primitive shark from the Lower Carboniferous of East Kilbride, Scotland', *Palaeontology*, 43(1), pp. 153-172.
- Sheldon, N.D. and Tabor, N.J. (2009) 'Quantitative paleoenvironmental and paleoclimatic reconstruction using paleosols', *Earth-Science Reviews*, 95(1-2), pp. 1-52.
- Shiers, M.N., Mountney, N.P., Hodgson, D.M. and Cobain, S. (2014) 'Depositional controls on tidally influenced fluvial successions, Neslen Formation, Utah, USA', *Sedimentary Geology*, 311, pp. 1-16.
- Slowakiewicz, M. and Mikolajewski, Z. (2011) 'Upper Permian Main Dolomite microbial carbonates as potential source rocks for hydrocarbons (W Poland)', *Marine and Petroleum Geology*, 28(8), pp. 1572-1591.
- Smithson, T.R., Richards, K.R. and Clack, J.A. (2016) 'Lungfish Diversity in Romer's Gap: Reaction to the End-Devonian Extinction', *Palaeontology*, 59(1), pp. 29-44.
- Smithson, T.R., Wood, S.P., Marshall, J.E.A. and Clack, J.A. (2012) 'Earliest Carboniferous tetrapod and arthropod faunas from Scotland populate Romer's Gap', *Proceedings of the National Academy of Sciences of the United States of America*, 109(12), pp. 4532-4537.
- Somerville, I.D., Strogon, P., Mitchell, W.I., Somerville, H.E.A. and Higgs, K.T. (2001) 'Stratigraphy of Dinantian Rocks in WB3 Borehole, Co. Armagh', *Irish Journal of Earth Sciences*, 19, pp. 51-78.
- Stamberg, S. (2006) 'Carboniferous-Permian actinopterygian fishes of the continental basin of the Bohemian Massif, Czech Republic: an overview.', in Lucas, S.G., Cassinis, G. and Schneider, J.W. (eds.) *Non-Marine Permian Biostratigraphy nad Biochronology*. Special Publication, 265 edn. The Geological Society of London, pp. 217-230.
- Stephenson, M., Williams, M., Monaghan, A., Arkley, S. and Smith, R. (2002) 'Biostratigraphy and palaeoenvironments of the Ballagan Formation (lower Carboniferous) in Ayrshire', *Scottish Journal of Geology*, 38, pp. 93-111.
- Stephenson, M.H., Millward, D., Leng, M.J. and Vane, C.H. (2008) 'Palaeoecological and possible evolutionary effects of early Namurian (Serpukhovian, Carboniferous) glacioeustatic cyclicity', *Journal of the Geological Society*, 165, pp. 993-1005.
- Stephenson, M., Williams, M., Leng, M. and Monaghan, A. (2004) 'Aquatic plant microfossils of probable non-vascular origin from the Ballagan formation (Lower

- carboniferous), Midland Valley, Scotland', *Proceedings of the Yorkshire Geological Society*, 55, pp. 145-158.
- Stone, P. (2012) *British Regional Geology: South of Scotland*. 4th edn. Nottingham, UK: British Geological Survey.
- Stone, P., Milward, D., Young, B., Merritt, J.W., Clarke, S.M., McCormack, M. and Lawrence, D.J.D. (2010) *British Regional Geology: Northern England*. fifth edn. Keyworth, Nottingham: British Geological Survey.
- Szpak, P., White, C.D., Longstaffe, F.J., Millaire, J. and Vasquez Sanchez, V.F. (2013) 'Carbon and Nitrogen Isotopic Survey of Northern Peruvian Plants: Baselines for Paleodietary and Paleoecological Studies', *PLOS ONE*, 8, pp. e53763.
- Taylor, K.G. and Macquaker, J.H.S. (2000) 'Early diagenetic pyrite morphology in a mudstone-dominated succession: the Lower Jurassic Cleveland Ironstone Formation, eastern England', *Sedimentary Geology*, 131(1-2), pp. 77-86.
- Taylor, P.D. and Vinn, O. (2006) 'Convergent morphology in small spiral worm tubes ('Spirorbis') and its palaeoenvironmental implications', *Journal of the Geological Society*, 163(2), pp. 225-228.
- Tucker, M.E. and Wright, V.P. (1990) *Carbonate Sedimentology*. 1st edn. Oxford, UK: Blackwell Publishing.
- Turner, S., Burrow, C. and Warren, A. (2005) 'Gyracanthides hawkinsi sp nov (Acanthodii, Gyracanthidae) from the Lower Carboniferous of Queensland, Australia, with a review of gyracanthid taxa', *Palaeontology*, 48, pp. 963-1006.
- Vasconcelos, C. and McKenzie, J.A. (1997) 'Microbial mediation of modern dolomite precipitation and diagenesis under anoxic conditions (Lagoa Vermelha, Rio de Janeiro, Brazil)', *Journal of Sedimentary Research*, 67(3), pp. 378-390.
- Vasconcelos, C., Warthmann, R., McKenzie, J.A., Visscher, P.T., Bittermann, A.G. and van Lith, Y. (2006) 'Lithifying microbial mats in Lagoa Vermelha, Brazil: Modern Precambrian relics?', *Sedimentary Geology*, 185(3-4), pp. 175-183.
- Wakefield, M.I. (1995) 'Ostracod biostratigraphy at lagoonal shorelines: examples from the Great Estuarine Group, Middle Jurassic, Scotland', *Proceedings of the Geologists' Association*, 106(3), pp. 211-218.
- Waters, C.N., Somerville, I.D., Jones, N.S., Cleal, C.J., Collinson, J.D., Waters, R.A., Besly, B.M., Dean, M.T., Stephenson, M.H., Davies, J.R., Freshney, E.C., Jackson, D.I., Mitchell, W.I., Powell, J.H., Barclay, W.J., Browne, M.A.E., Leveridge, B.E., Long, S.L. and McLean, D. (eds.) (2011) *A Revised Correlation of Carboniferous Rocks in the British Isles*. Geological Society of London.

- Wignall, P.B. and Newton, R. (1998) 'Pyrite framboid diameter as a measure of oxygen deficiency in ancient mudrocks', *American Journal of Science*, 298(7), pp. 537-552.
- Williams, M., Leng, M.J., Stephenson, M.H., Andrews, J.E., Wilkinson, I.P., Siveter, D.J., Horne, D.J. and Vannier, J.M.C. (2006) 'Evidence that Early Carboniferous ostracods colonised coastal flood plain brackish water environments', *Palaeogeography Palaeoclimatology Palaeoecology*, 230(3-4), pp. 299-318.
- Williams, M., Stephenson, M., Wilkinson, I.P., Leng, M.J. and Miller, C.G. (2005) 'Early Carboniferous (Late Tournaisian-Early Visean) ostracods from the Ballagan Formation, central Scotland, UK', *Journal of Micropalaeontology*, 24, pp. 77-94.
- Wilson, R.B. (1965) 'A revision of some Carboniferous species of *Modiola*', *Bulletin of the Geological Society of Great Britain*, 23, pp. 9-15.
- Wilson, R.B. (1989) 'A study of the Dinantian marine macrofossils of central Scotland', *Earth and Environmental Science Transactions of the Royal Society of Edinburgh*, 80(02), pp. 91-126.
- Worsley, D. and Mork, A. (2001) 'The environmental significance of the trace fossil *Rhizocorallium jenense* in the Lower Triassic of western Spitsbergen', *Polar Research*, 20(1), pp. 37-48.
- Young, J. (1896) 'Notes on the Ostracoda found in the lacustrine or freshwater strata of the Campsie district which alternate with the lower beds of the Carboniferous marine limestone series', *Transactions of the Geological Society of Glasgow*, 10, pp. 334-336.
- Zaton, M., Vinn, O. and Tomescu, A.M.F. (2012) 'Invasion of freshwater and variable marginal marine habitats by microconchid tubeworms - an evolutionary perspective', *Geobios*, 45(6), pp. 603-610.
- Zonneveld, J. and Gingras, M.K. (2013) 'The Ichnotaxonomy of Vertically Oriented, Bivalve-Generated *Equilibrichnia*', *Journal of Paleontology*, 87(2), pp. 243-253.