The Old Red Sandstone of Great Britain

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Chapter 2

The Orcadian Basin

INTRODUCTION

P. Stone

The Devonian strata preserved in the north-east of mainland Scotland and the Orkney and Shetland islands are the remnants of deposits that accumulated in a major, intramontane lake rift-basin, the Orcadian Basin (Figure 2.1). Offshore well records and seismic data show a north-eastwards continuation of the basin beneath the North Sea, with probably equivalent strata also present to the west, in the West Orkney, Outer Isles and North Minches basins. The succession is over 4 km thick in Caithness and possibly twice that in parts of Shetland, where three distinct basins appear to have been juxtaposed by dextral strike-slip faulting (Mykura and Phemister, 1976). A restoration of the Shetland basins in relation to the main Orcadian depocentre is shown in Figure 2.2. Overviews of the Orcadian Basin successions and their regional context were provided by Anderton *et al.* (1979), Dineley (1999a), and

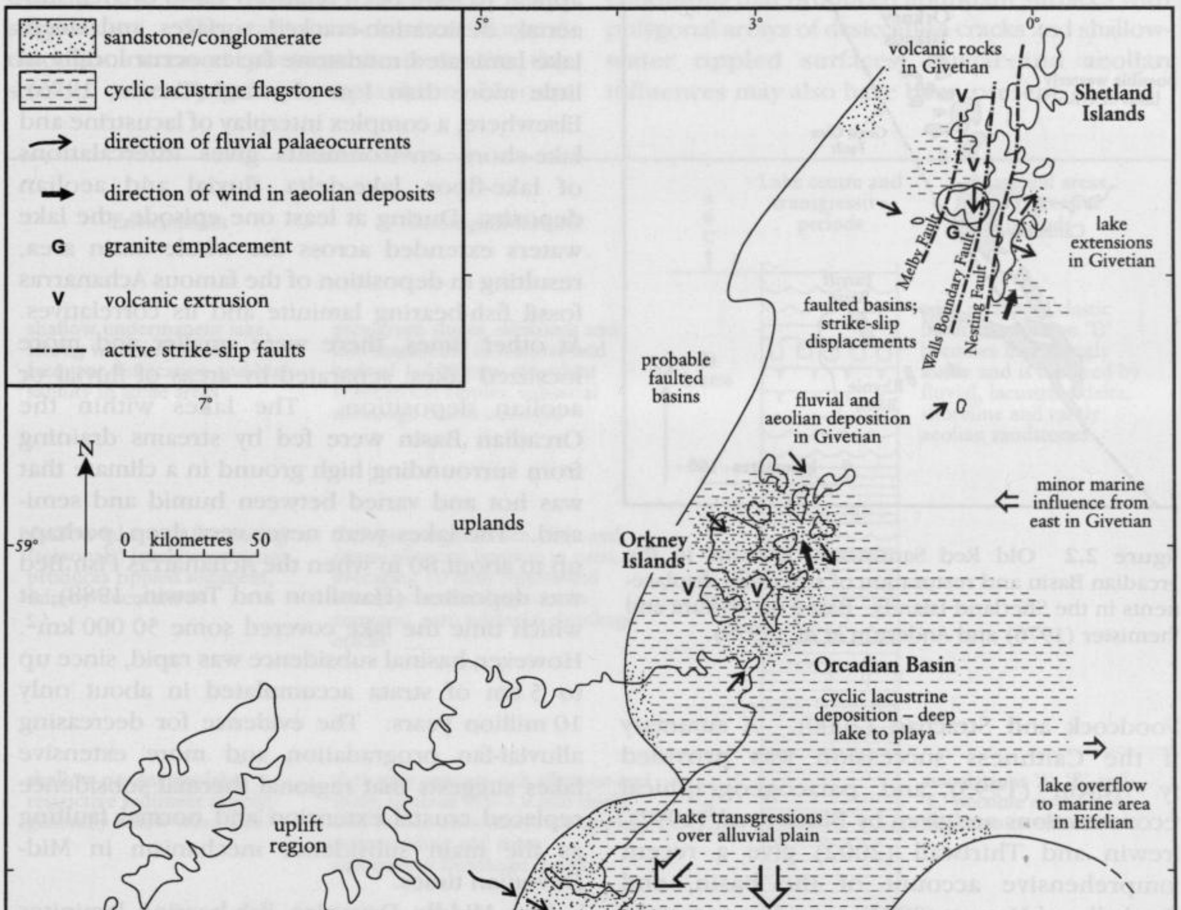


Figure 2.1 Generalized Mid-Devonian palaeogeography based on present geography of northern Scotland. After Trewin and Thirlwall (2002).

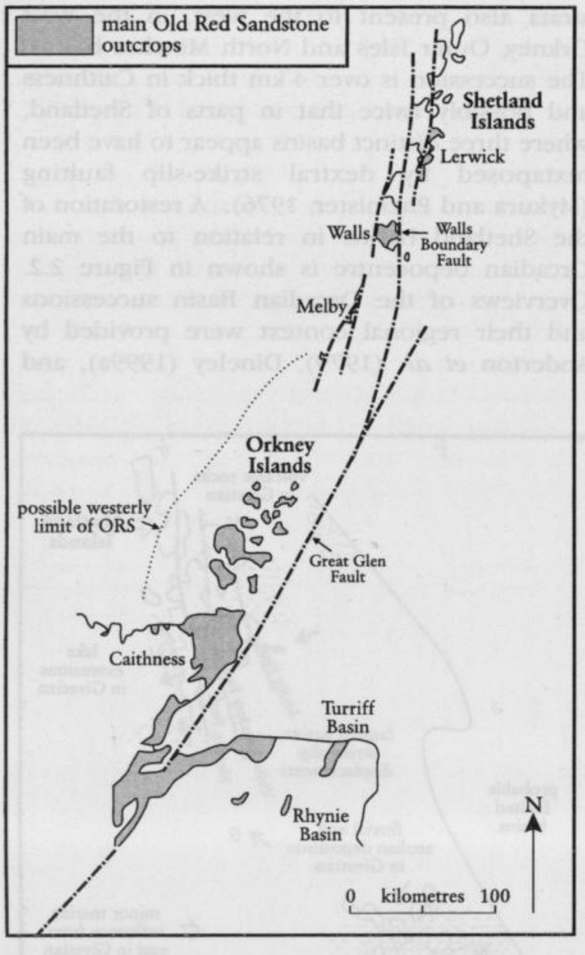


Figure 2.2 Old Red Sandstone outcrops in the Orcadian Basin and restoration of strike-slip displacements in the Shetland Islands. Based on Mykura and Phemister (1976) and Anderton *et al.* (1979).

Woodcock and Strachan (2000). A summary of the Caithness succession was provided by Trewin (1993) and palaeogeographical reconstructions are given by Bluck *et al.* (1992). Trewin and Thirlwall (2002) give a recent comprehensive account of the basin, and Marshall and Hewett (2003) provide a summary. The following introductory account is based on these sources.

Uplift and erosion dominated the northern Scottish region during Early Devonian times. Uplift was largely driven by the intrusion of an extensive suite of granitic plutons and was accompanied by strike-slip movement on major NE–SW-trending faults. Hence, the accumulation of Lower Devonian strata was localized and of variable lithology, lying unconformably on a surface of Moine, Dalradian and intrusive granitic rocks of the Highlands metamorphic complexes that had considerable local relief variations. A similar pattern continued into Mid-Devonian times, with the localized deposition of lake-margin facies, including carbonates, stromatolites and coarse alluvial breccias on the basement unconformity, but passing distally into and transgressively covered by fluvio-deltaic sandstones and lithologies of the developing deep-lake facies.

Lake transgression and regression occurred repeatedly through Mid-Devonian times and appear to have been relatively rapid events; sub-aerial, desiccation-cracked surfaces and deep-lake laminated mudstone facies occur locally in little more than 1 m of strata (Trewin, 1976). Elsewhere, a complex interplay of lacustrine and lake-shore environments gives intercalations of lake-floor, lake-delta, fluvial and aeolian deposits. During at least one episode, the lake waters extended across the whole basin area, resulting in deposition of the famous Achanarras fossil fish-bearing laminite and its correlatives. At other times, there were smaller and more localized lakes, separated by areas of fluvial or aeolian deposition. The lakes within the Orcadian Basin were fed by streams draining from surrounding high ground in a climate that was hot and varied between humid and semi-arid. The lakes were never very deep, perhaps up to about 80 m when the Achanarras Fish Bed was deposited (Hamilton and Trewin, 1988), at which time the lake covered some 50 000 km^2. However, basinal subsidence was rapid, since up to 5 km of strata accumulated in about only 10 million years. The evidence for decreasing alluvial-fan progradation and more extensive lakes suggests that regional thermal subsidence replaced crustal extension and normal faulting as the main subsidence mechanism in Mid-Devonian times.

The Middle Devonian fish-bearing laminites are a highly unusual lithofacies. Their regular, parallel, planar bedding facilitated their use as flagstones, examples of which can be seen in many British cities, and they are commonly known as 'Caithness Flags'. The GCR sites at **Achanarras Quarry** and **Wick Quarries** are disused flagstone quarries. The flagstones were deposited in the large, ephemeral lakes and have a cyclic character. Donovan (1980) recognized a basic sequence of four lithofacies associations

('A', 'B', 'C' and 'D'), representing a range of conditions from deep to shallow lake and exposed playa surface (Figure 2.3; Trewin, 1993; Trewin and Thirlwall, 2002). In practice, a great deal of facies variation arose through local differences in sediment supply, subsidence, carbonate production and water conditions. Donovan's 'A association' – the fish-bearing laminites – were deposited during the deep lake phase as intercalated silt, carbonate and organic laminae. They are interpreted as seasonal varves, controlled by annual climate variations. Clastic laminae represent input from rivers during the wet season, deposited from suspension. The carbonate laminae formed through increased photosynthesis in the dry season when algal phytoplankton blooms flourished in the lake waters, and the organic laminae formed by the subsequent annual decay of the phytoplankton. Shallower lake facies, the 'B & C associations' of Donovan, are thicker laminated and wave-rippled, with probable subaqueous shrinkage (synaeresis) cracks (although this interpretation is not universally accepted; see Astin and Rogers (1991) and subsequent discussions (Astin and Rogers, 1992, 1993; Trewin, 1992; Barclay *et al.*, 1993); see also Plummer and Gostin (1981) for a discussion on shrinkage cracks). Donovan's 'D association' formed in shallow, ephemeral lake conditions that produced abundant surfaces with polygonal arrays of desiccation cracks and shallow-water rippled surfaces; fluvial and aeolian influences may also have been present.

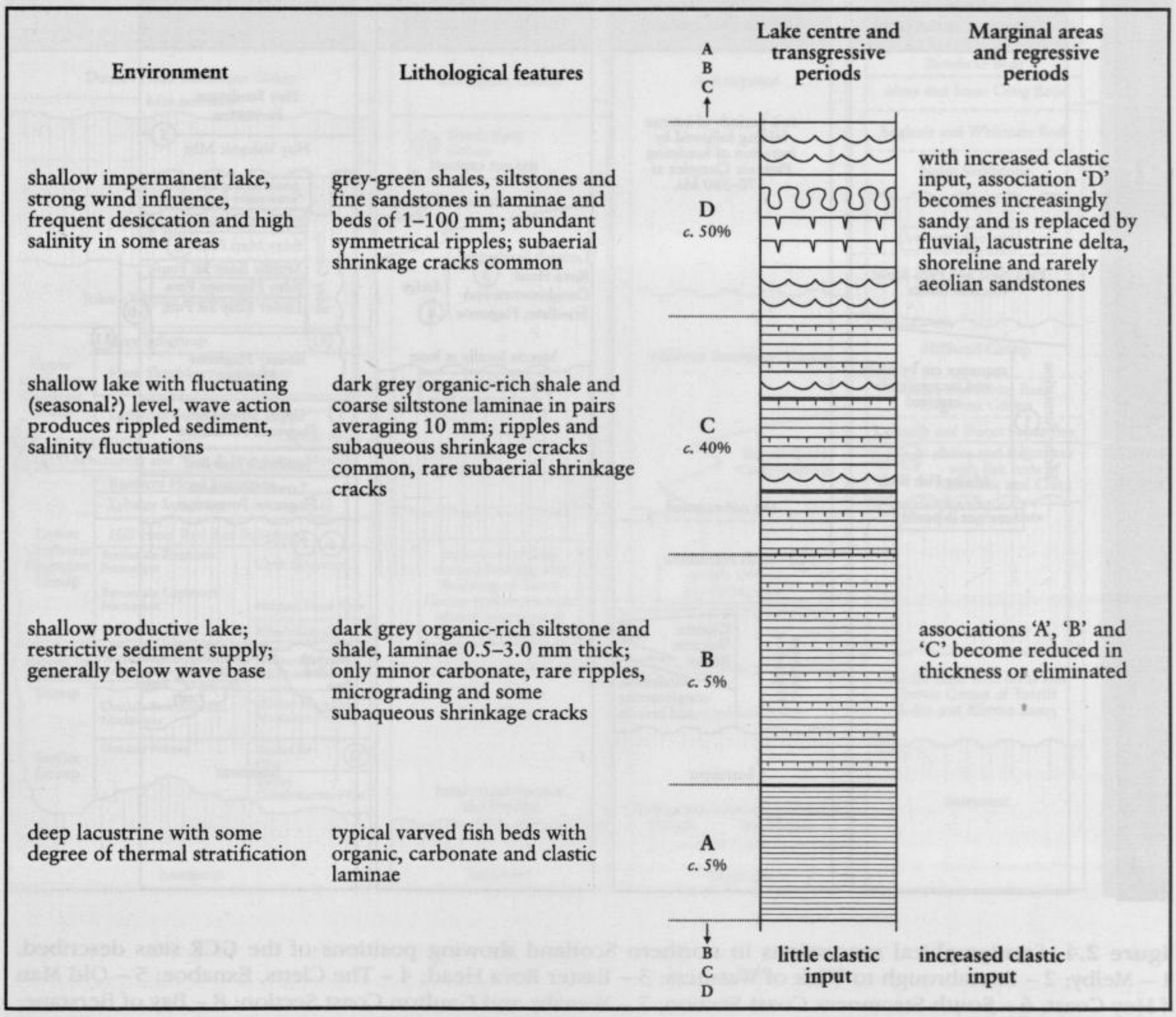


Figure 2.3 Cyclic lacustrine facies in the Caithness Flagstone Group. After Trewin and Thirlwall (2002), from Donovan (1980).

The Orcadian Basin lake(s) probably drained south-eastwards to the sea, the earliest evidence for possible marine influence within the basin being seen late in the Middle Devonian succession of Orkney (Marshall *et al.*, 1996). A marine connection became more important during Late Devonian times. Over most of the basin, the boundary between Middle and Upper Old Red Sandstone is conformable, but there is some evidence for localized tilting of the Middle Devonian strata, interpreted as a phase of limited basin inversion prior to Late Devonian sedimentation. More commonly, the base of the Upper Devonian Series is marked by a change in depositional environment. The Mid-Devonian lakes were replaced in the basin centre by sabkha plains that were periodically inundated by the sea, whilst meandering- and braided-river deposits accumulated on the basin margins.

Most of the preserved succession and the GCR sites in the Orcadian Basin are of Mid-Devonian age. The stratigraphical positions and ranges of the sites are shown in Figure 2.4. The lithostratigraphical scheme for the basin is not wholly satisfactory in modern terms and is currently under review. The version used in this account is largely that used by Dineley (1999a) in the *Fossil Fishes of Great Britain* GCR volume (Dineley and Metcalf, 1999), with small modifications to take account of recent developments.

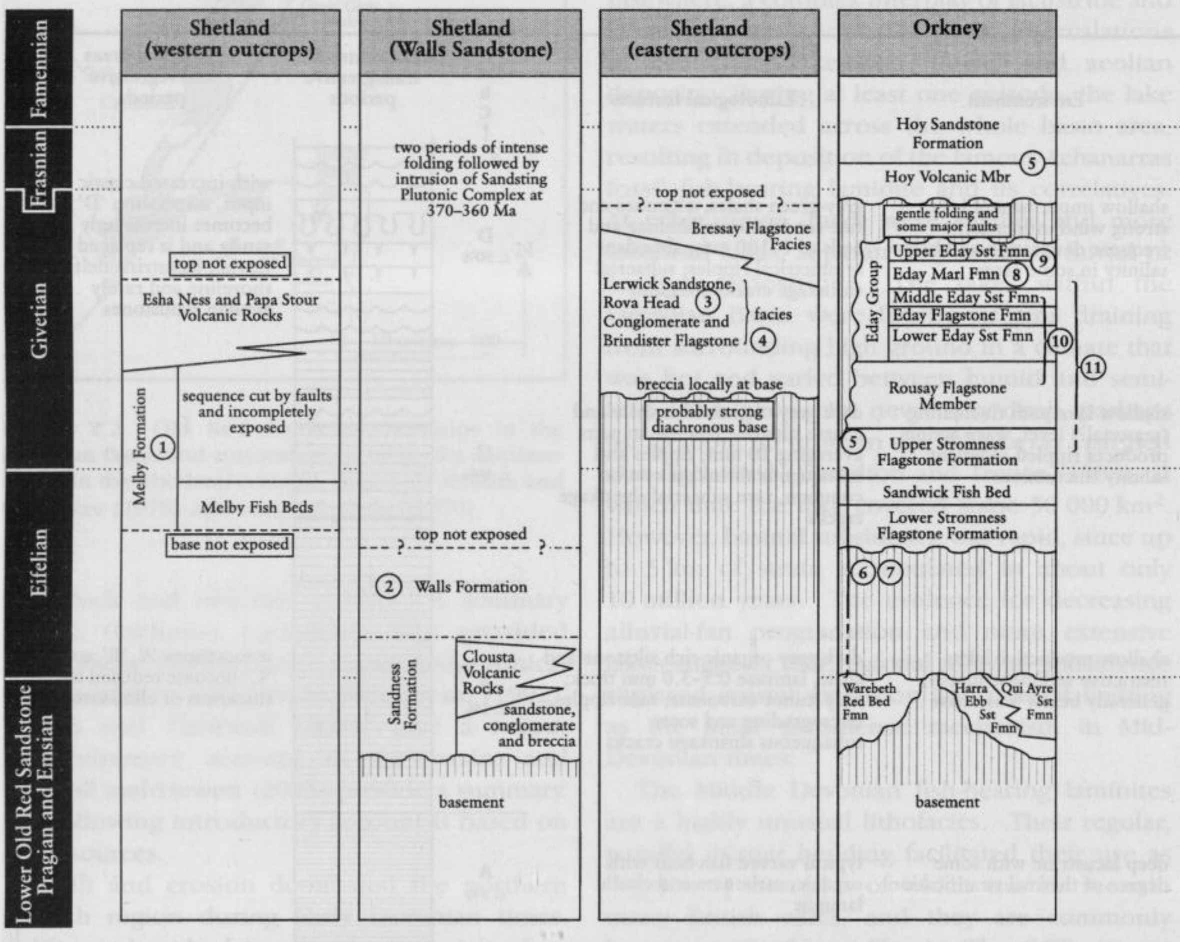


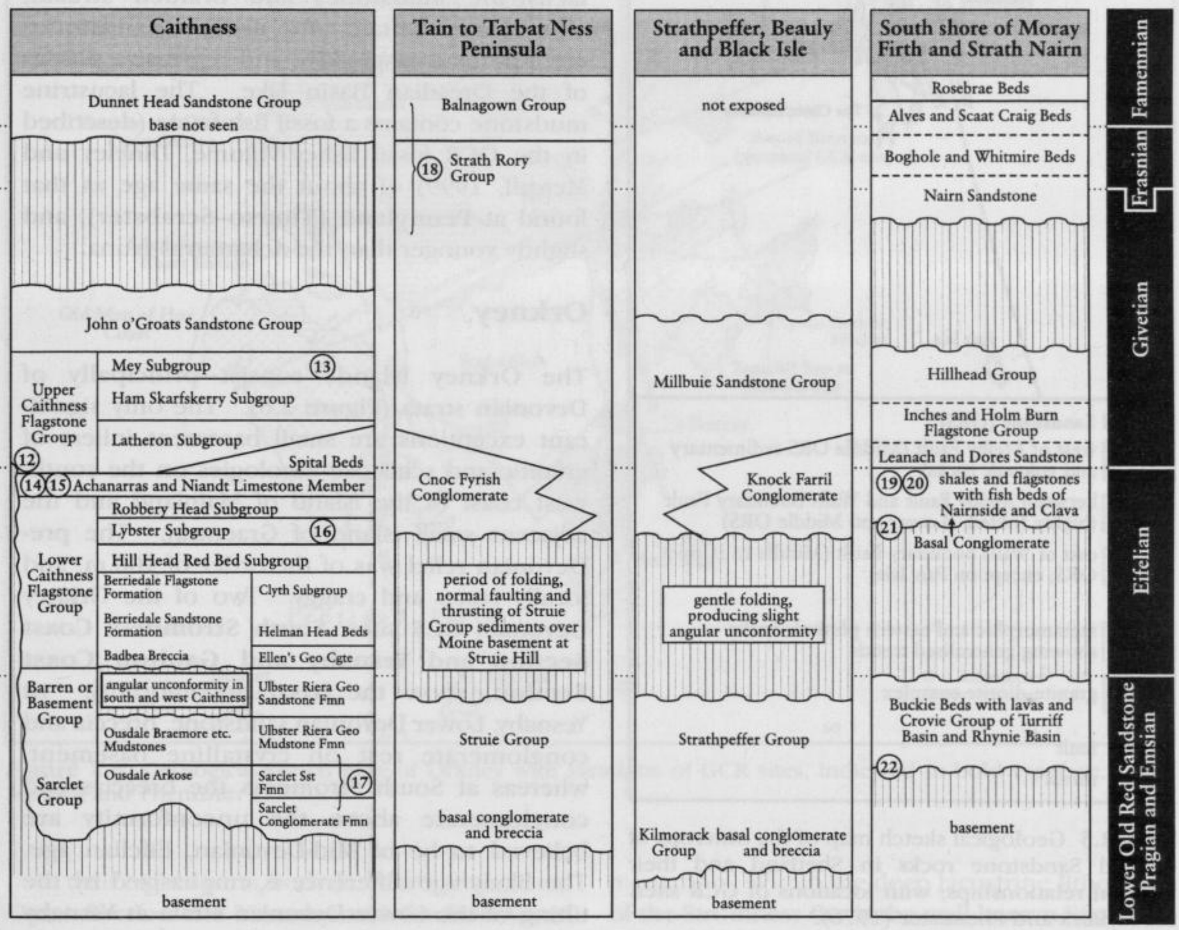
Figure 2.4 Stratigraphical successions in northern Scotland showing positions of the GCR sites described. (1 – Melby; 2 – Footabrough to Wick of Watsness; 3 – Easter Rova Head; 4 – The Cletts, Exnaboe; 5 – Old Man of Hoy Coast; 6 – South Stromness Coast Section; 7 – Yesnaby and Gaulton Coast Section; 8 – Bay of Berstane; 9 – Greenan Nev Coast, Eday; 10 – South Fersness Bay, Eday; 11 – Taracliff Bay to Newark Bay; 12 – Red Point; 13 – Pennyland (Thurso–Scrabster); 14 – Achanarras Quarry; 15 – John o'Groats; 16 – Wick Quarries; 17 – Sarclet; 18 – Tarbat Ness; 19 – Tynet Burn; 20 – Den of Findon; 21 – Dun Chia Hill (Loch Duntelchaig); 22 – Rhynie.)

Shetland

The Devonian strata of Shetland are separated into discrete terranes by a series of major north–south faults (Figure 2.5). There are considerable lithological and structural differences between the sequences in each of the terranes, but all are of Eifelian to Givetian, Mid-Devonian age (Figure 2.4). They have been related to three separate depocentres, juxtaposed by dextral strike-slip fault movements (Mykura and Phemister, 1976). Their possible restoration in relation to the main Orcadian Basin is shown in Figure 2.2.

The westernmost terrane, to the west of the Melby Fault, is the smallest and may have originated closest to the main Orcadian Basin; its Old Red Sandstone lithofacies, of late Eifelian age, is certainly the most similar of the Shetland varieties to those of Caithness and Orkney. The **Melby** GCR site shows two wholly lacustrine intervals alternating with thick fluvial sandstones. The lacustrine beds are thinly laminated, silty mudstones with carbonate nodules that contain an important fossil fish fauna, described in the *Fossil Fishes of Great Britain* GCR volume (Dineley and Metcalf, 1999). This fauna is considered to be slightly older than the Achanarras (Caithness) fauna.

East of the Melby Fault, and bounded to the east by the Walls Boundary Fault, lies a terrane that contains the Devonian succession of the Walls Peninsula. This is represented by probable



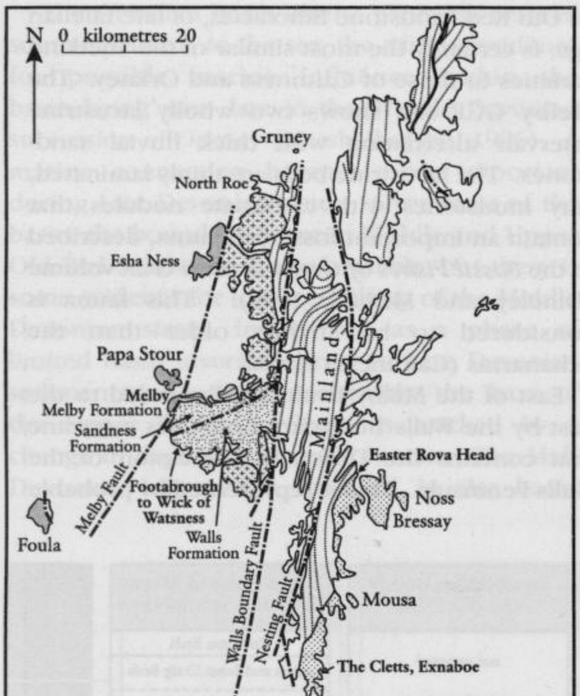


Figure 2.5 Geological sketch map of the outcrops of Old Red Sandstone rocks in Shetland and their structural relationships; with locations of GCR sites. After Mykura and Phemister (1976).

Givetian strata within the **Footabrough to Wick of Watsness** GCR site. The thick alternations of sandstone, siltstone and mudstone may have formed by turbidite deposition in a deep lake environment, but some combination of braided fluvial channel and overbank, littoral and shallow lacustrine lithofacies is perhaps more likely. The Devonian strata of the Walls Peninsula are more severely deformed than any other parts of the Orcadian Basin successions.

The Devonian succession of the eastern terrane, east of the Nesting Fault, is separated from those to the west by a basement terrane devoid of Devonian cover. There are two GCR sites, both with strata of Givetian age, at **Easter Rova Head** near Lerwick and at **The Cletts** near Exnaboe. The Easter Rova Head site shows a coarse, conglomeratic lake-margin alluvial fan comprising sheet-flood and mass-flow deposits. It probably rests unconformably on basement rocks, although the contact is not seen. Farther south, at The Cletts, a sequence of interbedded lacustrine mudstones and braided stream, alluvial-fan, deltaic and aeolian sandstones records the transgressive and regressive phases of the Orcadian Basin lake. The lacustrine mudstone contains a fossil fish fauna (described in the GCR fossil fishes volume, Dineley and Metcalf, 1999) of about the same age as that found at **Pennyland (Thurso–Scrabster)**, and slightly younger than the Achanarras fauna.

Orkney

The Orkney Islands consist principally of Devonian strata (Figure 2.6). The only significant exceptions are small basement inliers of granitic and schistose lithologies on the south-west coast of the island of Mainland and the adjacent small island of Graemsay. The pre-Devonian relief was of the order of 100 m, and locally steep and craggy. Two of the Orkney Devonian GCR sites, **South Stromness Coast Section** and **Yesnaby and Gaulton Coast Section**, show the basal unconformity. At Yesnaby, Lower Devonian sandstone, breccia and conglomerate rest on crystalline basement, whereas at South Stromness the breccias and conglomerate above the unconformity are believed to be of Mid-Devonian, Eifelian age. The likely age difference is emphasized by the tilting of the Lower Devonian strata at Yesnaby so that the basal sedimentary sequence is overlain unconformably by Middle Devonian rocks; at South Stromness the sedimentary relationships above the basal unconformity are all conformable. The Lower Devonian sequence represents alluvial-fan, lacustrine mudflat, aeolian dune and beach deposition.

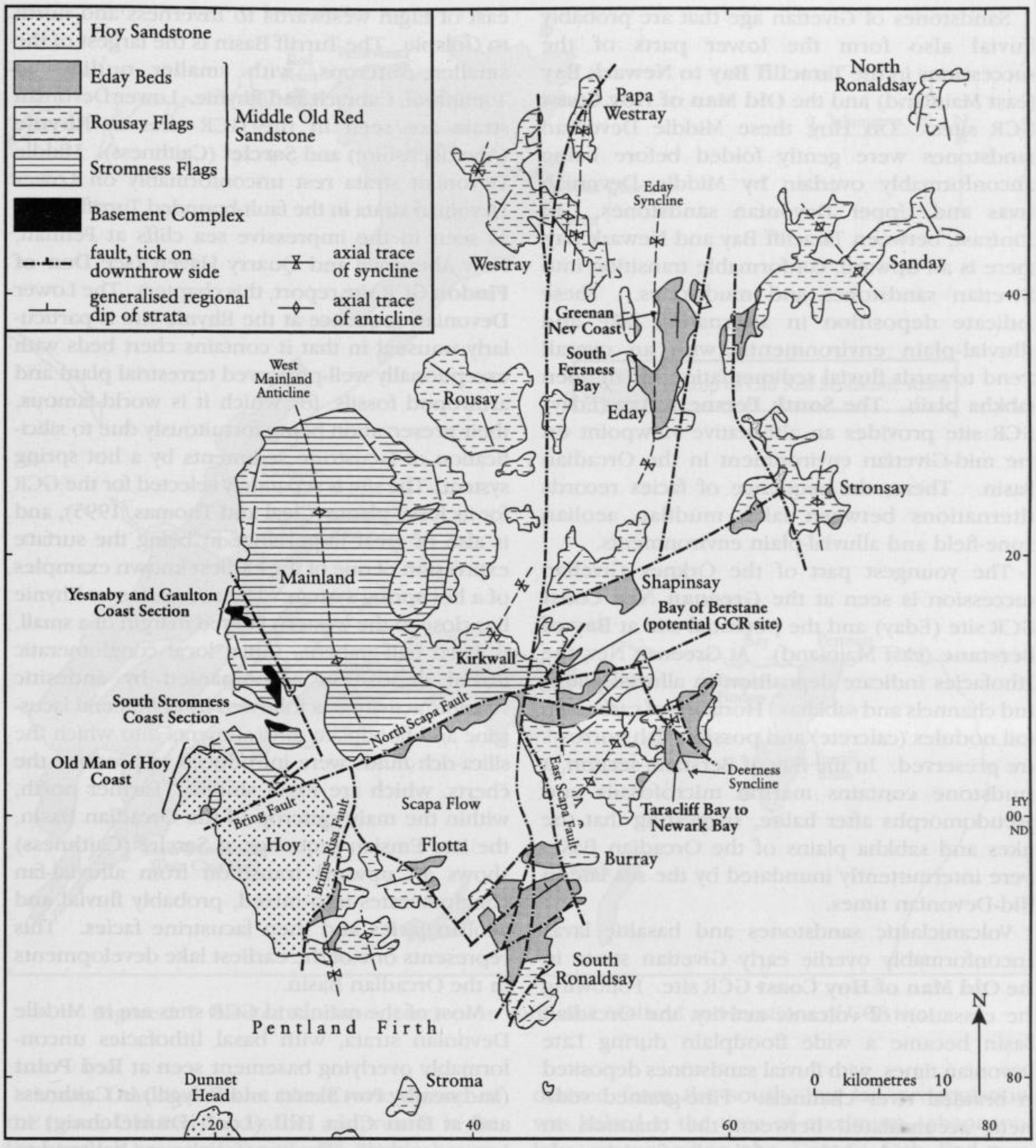


Figure 2.6 Geological sketch map of Orkney with locations of GCR sites, indicated in bold typeface. After Mykura and Phemister (1976).

In their higher parts, both the **South Stromness Coast Section** and **Yesnaby and Gaulton Coast Section** provide splendid sections through the Eifelian to Givetian lacustrine succession of the Orcadian Basin. A series of sedimentary cycles records lake transgression and regression, with evidence for sequential deep-lake, shallow-lake margin with alluvial fans, and subaerial, desiccated lake environments, as well as sporadic fluvial sedimentation. Within the deep lacustrine lithofacies of the Stromness site is the well-known Sandwick Fish Bed, the fossil fauna of which is described in the *Fossil Fishes of Great Britain* GCR volume (Dineley and Metcalf, 1999). The Sandwick Fish Bed is the stratigraphical equivalent of the fish-bearing laminite at Achanarras, thus demonstrating the wide extent of this deep lacustrine facies during a phase of maximum lake development.

Sandstones of Givetian age that are probably fluvial also form the lower parts of the successions in the **Taracliff Bay to Newark Bay** (east Mainland) and the **Old Man of Hoy Coast** GCR sites. On Hoy, these Middle Devonian sandstones were gently folded before being unconformably overlain by Middle Devonian lavas and Upper Devonian sandstones. In contrast, between Taracliff Bay and Newark Bay, there is an upward, conformable transition into Givetian sandstones and mudstones. These indicate deposition in alternating lake and alluvial-plain environments, with an overall trend towards fluvial sedimentation on an open sabkha plain. The **South Fersness Bay** (Eday) GCR site provides an alternative viewpoint on the mid-Givetian environment in the Orcadian Basin. There, the sequence of facies records alternations between lake, mudflat, aeolian dune-field and alluvial-plain environments.

The youngest part of the Orkney Givetian succession is seen at the **Greenan Nev Coast** GCR site (Eday) and the potential site at **Bay of Berstane** (east Mainland). At Greenan Nev, the lithofacies indicate deposition in alluvial plains and channels and sabkhas. Horizons of carbonate soil nodules (calcrete) and possible fish burrows are preserved. In the Bay of Berstane section, a mudstone contains marine microfossils and pseudomorphs after halite, suggesting that the lakes and sabkha plains of the Orcadian Basin were intermittently inundated by the sea late in Mid-Devonian times.

Volcaniclastic sandstones and basaltic lavas unconformably overlie early Givetian strata in the **Old Man of Hoy Coast** GCR site. Following the cessation of volcanic activity, the Orcadian Basin became a wide floodplain during Late Devonian times, with fluvial sandstones deposited in braided river channels. Fine-grained sediment accumulated between the channels in small lakes and sabkhas, whilst the fluvial sands were locally reworked into aeolian dune-fields. This sandstone-dominated succession forms the spectacular, 350 m-high cliffs that form the north-west coast of Hoy, as well as the cliffs of Dunnet Head on the Scottish mainland.

North-eastern Scottish mainland

The outcrops of Devonian strata in north-east Scotland, and the location of the GCR sites are shown in Figure 2.7. The principal outcrops are in eastern Caithness, and in the Moray Firth from east of Elgin westwards to Inverness and north to Golspie. The Turriff Basin is the largest of the smaller outcrops, with smaller outliers at Tomintoul, Cabrach and Rhynie. Lower Devonian strata are seen in the GCR sites at **Rhynie** (Aberdeenshire) and **Sarclet** (Caithness). Middle Devonian strata rest unconformably on Lower Devonian strata in the fault-bounded Turriff Basin, as seen in the impressive sea cliffs at Pennan, New Aberdour and Quarry Haven (see **Den of Findon** GCR site report, this chapter). The Lower Devonian sequence at the Rhynie site is particularly unusual in that it contains chert beds with exceptionally well-preserved terrestrial plant and arthropod fossils, for which it is world-famous, their preservation being fortuitously due to silicification of lacustrine sediments by a hot spring system. The site is separately selected for the GCR for its fossil plants (Cleal and Thomas, 1995), and is also of great importance in being the surface expression of one of the earliest known examples of a hot spring system. The succession at Rhynie lies close to the western faulted margin of a small, isolated half-graben. Initial local conglomeratic fluvial deposition accompanied by andesitic volcanic activity was followed by ephemeral lacustrine and floodplain environments into which the silica-rich fluids were introduced to produce the cherts, which are fossil sinters. Farther north, within the main outcrop of the Orcadian Basin, the late Emsian sequence at Sarclet (Caithness) shows an upward transition from alluvial-fan conglomerates into mixed, probably fluvial and aeolian facies and then lacustrine facies. This represents one of the earliest lake developments in the Orcadian Basin.

Most of the mainland GCR sites are in Middle Devonian strata, with basal lithofacies unconformably overlying basement seen at **Red Point** (and nearby Port Skerra and Balygill) in Caithness and at **Dun Chia Hill (Loch Duntelchaig)** in Inverness-shire. The successions are believed to be Eifelian to Givetian in age and to merge laterally with deeper-water lacustrine deposits farther into the basin. At Red Point, towards the western margin of the Orcadian Basin, limestone and coarse breccia tongues rest unconformably on a high-relief granodiorite basement and merge laterally with fluvial and lacustrine sandstone and mudstone. The Dun Chia Hill site shows the basal relationships at the southern margin of the Orcadian Basin. There, a coarse debris-flow breccia unconformably overlies a steeply undulating basement of Dalradian psammites.

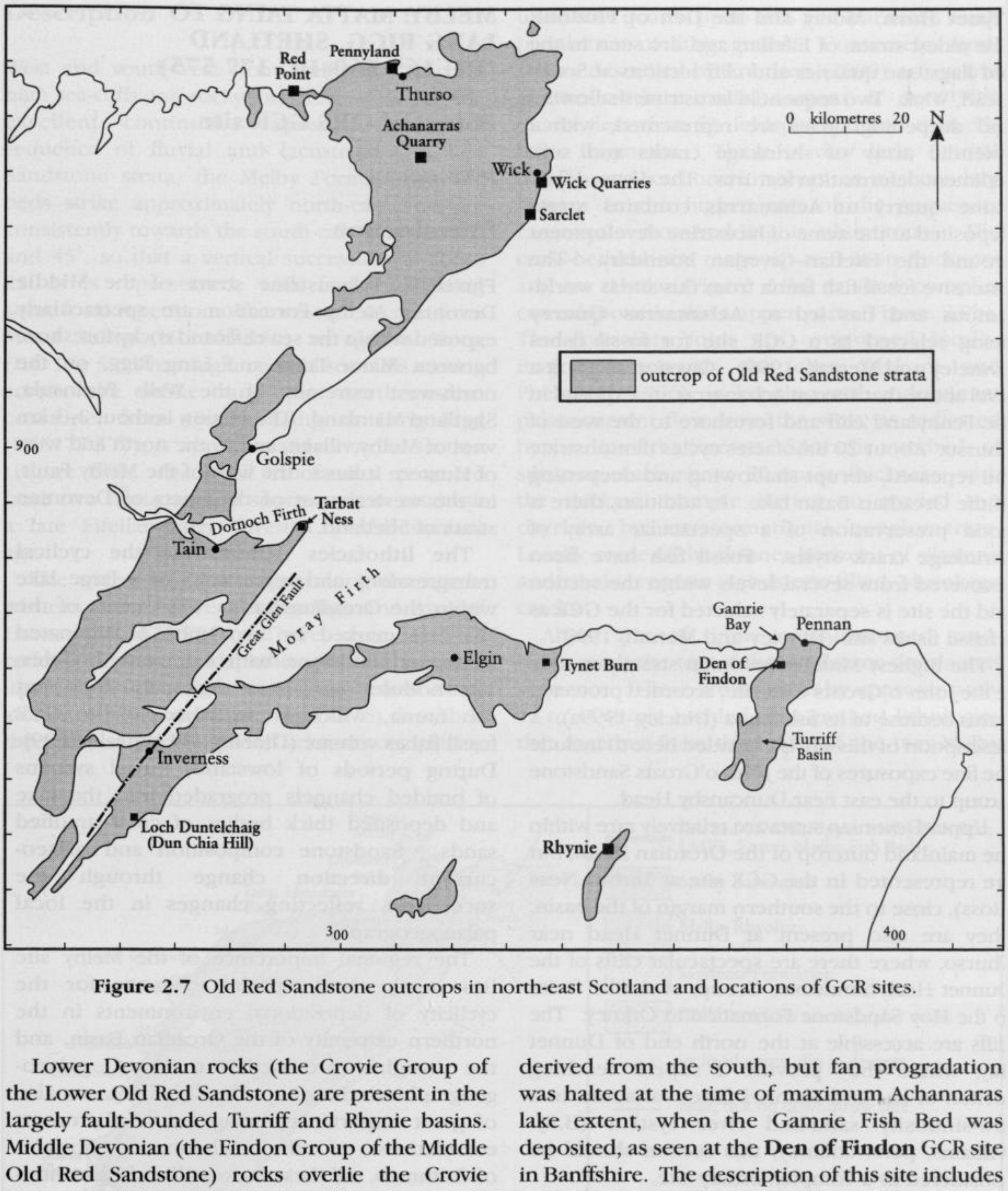


Figure 2.7 Old Red Sandstone outcrops in north-east Scotland and locations of GCR sites.

Lower Devonian rocks (the Crovie Group of the Lower Old Red Sandstone) are present in the largely fault-bounded Turriff and Rhynie basins. Middle Devonian (the Findon Group of the Middle Old Red Sandstone) rocks overlie the Crovie Group unconformably. The Turriff Basin extends about 30 km inland from the superb exposures of its fill on the Moray Firth coast at Gamrie Bay and New Aberdour. Coarse, basal conglomerates of the Crovie Group fill an irregular Dalradian basement topography and are succeeded by a 600 m-thick variable succession of alluvial-fan conglomerates and sandstones and floodplain/playa mudstones. The Findon Group also comprises basin-margin alluvial-fan deposits, derived from the south, but fan progradation was halted at the time of maximum Achannaras lake extent, when the Gamrie Fish Bed was deposited, as seen at the **Den of Findon** GCR site in Banffshire. The description of this site includes a brief account of the magnificent cliff and foreshore sections of the Crovie and Findon groups nearby in Gamrie Bay, New Aberdour and Quarry Haven, which also merit protected status.

The main lacustrine lithofacies of the Orcadian Basin, spanning the Eifelian and Givetian stages, are represented by three GCR sites in Caithness (**Wick Quarries**, **Achanarras Quarry**, and **Pennyland (Thurso–Scrabster)**), in addition to two sites on the southern margin of the basin

(**Tynet Burn**, Moray, and the **Den of Findon**). The oldest strata, of Eifelian age, are seen in the old flagstone quarries and cliff sections at South Head, Wick. Two sequential lacustrine shallowing and deepening cycles are represented, with a splendid array of shrinkage cracks and soft-sediment deformation features. The disused flagstone quarry at Achanarras contains strata deposited at the acme of lacustrine development around the Eifelian–Givetian boundary. The extensive fossil fish fauna from this site is world-famous and has led to **Achanarras Quarry** being selected as a GCR site for fossil fishes (Dineley and Metcalf, 1999). Givetian strata at a level above that seen at Achanarras are exposed in the Pennyland cliff and foreshore to the west of Thurso. About 20 lithofacies cycles demonstrate the repeated, abrupt shallowing and deepening of the Orcadian Basin lake. In addition, there is good preservation of a spectacular array of shrinkage crack styles. Fossil fish have been recovered from several levels within the section and the site is separately selected for the GCR as a fossil fishes site (Dineley and Metcalf, 1999).

The highest Middle Devonian strata are seen at the **John o'Groats** GCR site, accorded protected status because of its fish fauna (Dineley, 1999a). A description of this site is extended here to include the fine exposures of the John o'Groats Sandstone Group to the east near Duncansby Head.

Upper Devonian strata are relatively rare within the mainland outcrop of the Orcadian Basin, but are represented in the GCR site at **Tarbat Ness** (Ross), close to the southern margin of the basin. They are also present at Dunnet Head near Thurso, where there are spectacular cliffs of the Dunnet Head Sandstone Group, of similar facies to the Hoy Sandstone Formation in Orkney. The cliffs are accessible at the north end of Dunnet Bay, where they provide a superb teaching section of bar and channel facies variations in a low-sinuosity sand-bed river system (B.P.J. Williams, pers. comm.); this section should be considered as a complementary site.

In the lower part of the succession exposed at Tarbat Ness, Givetian fluvial sandstones with sporadic lacustrine mudstones lie at the top of the Middle Devonian succession. They are conformably overlain by fluvial, aeolian and sabkha-like deposits, believed to be Frasnian to Famennian in age. The sabkha deposits contain striking examples of desiccation and evaporitic structures.

MELBY: MATTA TAING TO LANG RIGG, SHETLAND (HU 165 560–HU 177 575)

Potential ORS GCR site

P. Stone

Introduction

Fluvial and lacustrine strata of the Middle Devonian Melby Formation are spectacularly exposed within the sea cliffs and rocky foreshore between Matta Taing and Lang Rigg, on the north-west extremity of the Walls Peninsula, Shetland Mainland. The section is about 2–3 km west of Melby village, and to the north and west of Huxter. It lies to the west of the Melby Fault, in the westernmost of the tracts of Devonian strata of Shetland.

The lithofacies demonstrate the cyclical transgressions and regressions of a large lake within the Orcadian Basin. Highstands of the lake are marked by two units of laminated lacustrine mudstone with carbonate nodules. The nodules contain a regionally important fish fauna, which is described in the GCR fossil fishes volume (Dineley and Metcalf, 1999). During periods of lowstand, fluvial systems of braided channels prograded into the lake and deposited thick bodies of cross-stratified sands. Sandstone composition and palaeocurrent direction change through the succession, reflecting changes in the local palaeogeography.

The regional importance of the Melby site lies in the well-exposed evidence for the cyclicity of depositional environments in the northern extremity of the Orcadian Basin, and the insights provided into local palaeogeographical changes. The fish beds are also of great importance, being the northernmost correlatives of the Achannaras horizon of Caithness. All these factors allow better interpretation and characterization of the regional tectonic framework and development of the Orcadian Basin. Overviews of the geology are provided by Mykura (1976, 1991) and a detailed account of the section is given by Mykura and Phemister (1976). Details of the fish beds are given by Hall and Donovan (1978) and Dineley (1999a).

Description

West and south-west from Melby and Huxter, high sea-cliffs and rocky foreshore areas provide excellent, continuous exposure through a sequence of fluvial and lacustrine Old Red Sandstone strata, the Melby Formation. The beds strike approximately north-east and dip consistently towards the south-east between 25° and 45°, so that a vertical succession of about 250 m is represented, with faulting causing some local repetition. The western section from Matta Taing (HU 166 560) to Quilva Taing (HU 171 572) has high, rugged sea-cliffs cutting across the strike, and the north-western coastline, from Quilva Taing to Lang Rigg (HU 177 577), is approximately a strike section (Figure 2.8). The fossil fish faunas recovered from two horizons and spore analysis establish a late Eifelian–early Givetian (Mid-Devonian) age.

The oldest strata are seen along the north-western coast between Lang Rigg and Ayre of Huxter (HU 173 574). There, over 50 m of reddish brown, feldspathic sandstones pass upwards into grey-brown, calcareous sandstones. Mykura and Phemister (1976) described the rocks as poorly graded arkoses, comprising up to 80% quartz, up to 40% feldspar and a small proportion of lithic grains. Conversely, sandstones described by Knudsen (2000) are quartz arenites with a modal content of over 90% quartz, less than 5% feldspar and less than 5% lithic fragments. There is very little detrital matrix, and carbonate occurs as a secondary cement in the higher part of the sequence. Large-scale cross-bedding is ubiquitous. The cross-bedded sets range up to 3 m in thickness and are tabular (Figure 2.9a) or show complex convolutions in their upper parts (Figure 2.9b). The convolute folds are overturned locally, invariably towards the east (Mykura and Phemister, 1976), in keeping with a dominant palaeocurrent flow from the west, as indicated by the dip of the foresets. Sporadic interbeds of siltstone reach about 50 cm in thickness, with the thickest example having an eroded top overlain by intraformational conglomerate. Locally, some of the thinner horizons of siltstone and shale contain sandstone-filled shrinkage cracks.

About 2 m of ripple cross-laminated sandstone mark the top of the sandstone-dominated sequence. These are overlain by approximately 12 m of thinly bedded, fine-grained lithologies that together form the lower of the two Melby

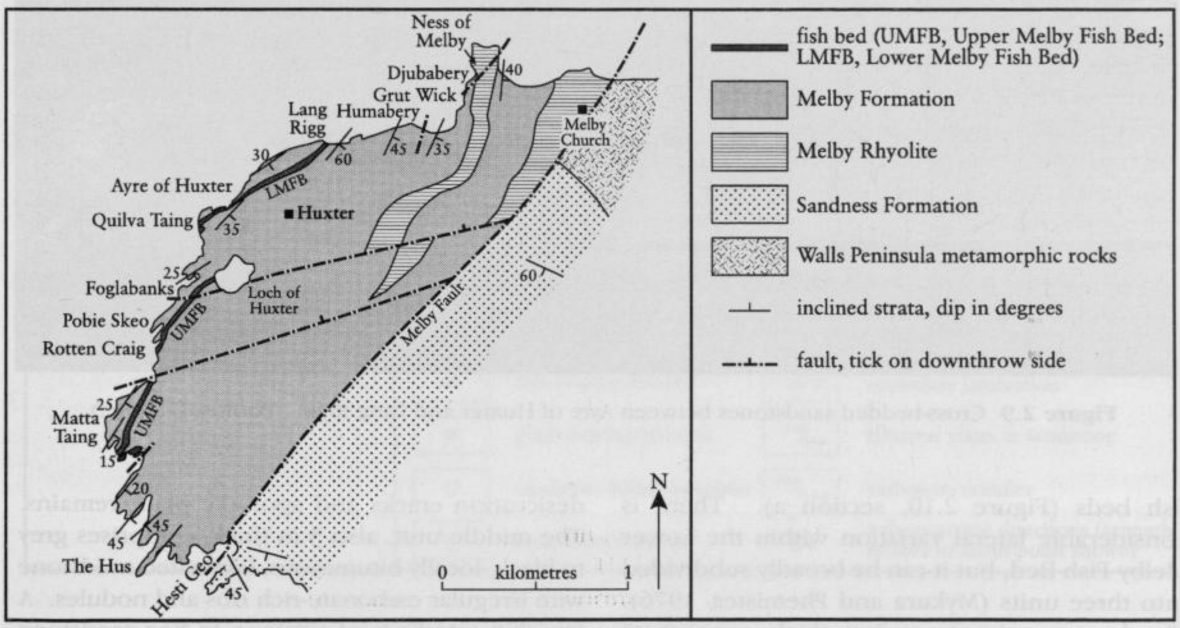


Figure 2.8 Geological sketch map of the Melby Formation on Shetland Mainland. After Mykura and Phemister (1976).

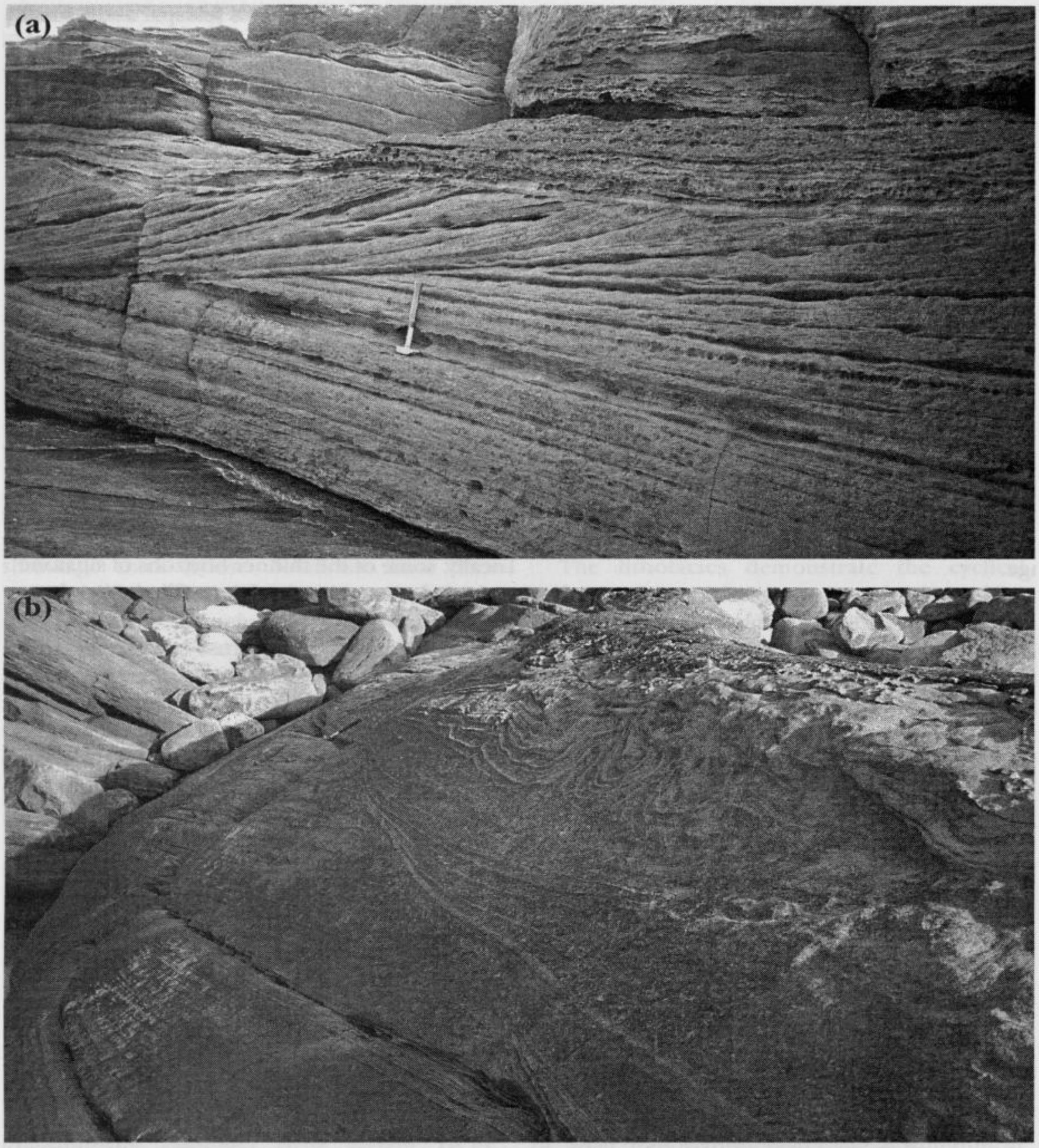


Figure 2.9 Cross-bedded sandstones between Ayre of Huxter and Lang Rigg. (Photos: P. Stone.)

fish beds (Figure 2.10, section a). There is considerable lateral variation within the Lower Melby Fish Bed, but it can be broadly subdivided into three units (Mykura and Phemister, 1976). The lowest unit, about 3 m thick, consists of pale grey, laminated siltstone and mudstone with desiccation cracks and sporadic plant remains. The middle unit, also 3 m thick, comprises grey to black, locally bituminous, laminated mudstone with irregular carbonate-rich ribs and nodules. A few thin interbeds of siltstone to fine sandstone have rippled upper surfaces. Plant remains are

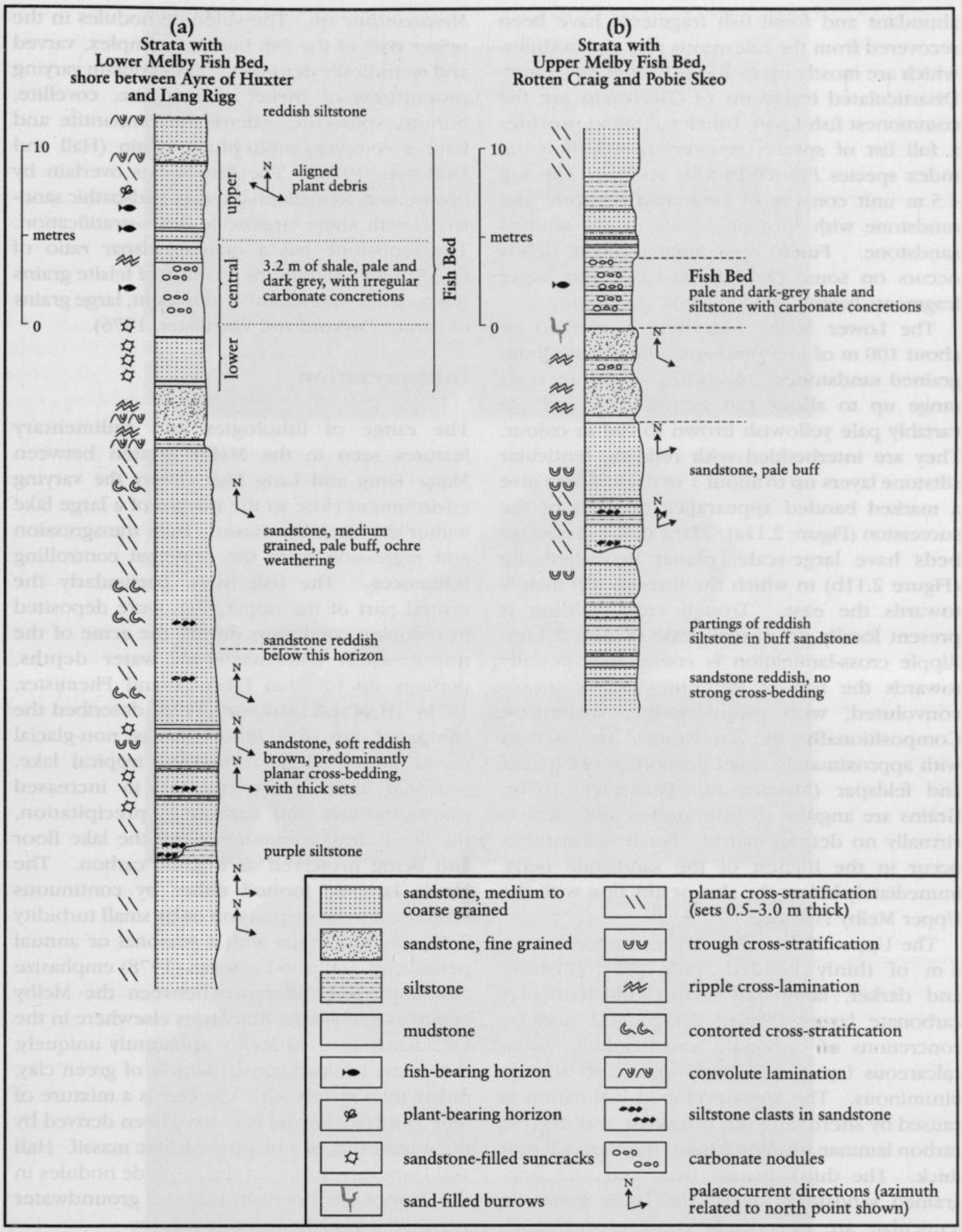


Figure 2.10 Stratal settings of the Melby fish beds. After Mykura and Phemister (1976), reproduced from Dineley (1999a).

abundant and fossil fish fragments have been recovered from the calcareous ribs and nodules, which are mostly up to about 25 mm by 75 mm. Disarticulated fragments of *Coccosteus* are the commonest fish fossil; Dineley (1999a) provides a full list of species recovered, including the index species *Pterichthyodes milleri*. The top 4.5 m unit consists of laminated siltstone and mudstone with sporadic lenses of fine-grained sandstone. Finely disseminated plant debris occurs on some bedding surfaces, the larger fragments having an ESE–WNW alignment.

The Lower Melby Fish Bed is overlain by about 100 m of predominantly fine- to medium-grained sandstones. Individual sandstone beds range up to about 1 m in thickness and are variably pale yellowish brown to red in colour. They are interbedded with reddish, lenticular siltstone layers up to about 1 m thick, which give a marked banded appearance to parts of the succession (Figure 2.11a). Many of the sandstone beds have large-scale, planar cross-bedding (Figure 2.11b) in which the foresets dip mainly towards the east. Trough cross-bedding is present locally on a small-scale (Figure 2.11c). Ripple cross-lamination is common, especially towards the top of the unit, and is locally convoluted, with pseudonodule structures. Compositionally, the sandstones are arkoses with approximately equal proportions of quartz and feldspar (Mykura and Phemister, 1976). Grains are angular to subrounded and there is virtually no detrital matrix. Small worm tubes occur in the highest of the sandstone beds, immediately below the abrupt junction with the Upper Melby Fish Bed.

The Upper Melby Fish Bed comprises about 4 m of thinly bedded, pale grey siltstone and darker, laminated shale with lenticular carbonate layers (Figure 2.12b) and nodular concretions of carbonate and sulphide; some calcareous layers are nearly black and slightly bituminous. The best-developed lamination is caused by alternating silt, carbonate and organic carbon laminae forming triplets 0.3 mm to 2 mm thick. The thin siltstone beds and rare fine-grained sandstone layers, the latter generally lenticular, are pervasively convoluted (Figure 2.12b) with rippled upper bedding surfaces (Figure 2.12c) and contain sporadic plant material. The carbonate layers and nodules contain a well-preserved but fragmentary fossil fish fauna listed by Dineley (1999a) that includes *Coccosteus cuspidatus*, *Homosteus milleri* and *Mesacanthus* sp.. The sulphide nodules in the upper part of the fish bed are complex, varved and tectonically deformed. They contain varying proportions of pyrite, chalcopyrite, covellite, bornite, sphalerite, galena and tennantite and have a complex, multi-phase origin (Hall and Donovan, 1978). The fish bed is overlain by brown, fine- to medium-grained, feldspathic sandstone with some large-scale cross-stratification. The sandstone has a quartz/feldspar ratio of 65:35, with a significant quantity of felsite grains (up to 10%), and relatively abundant, large grains of garnet (Mykura and Phemister, 1976).

Interpretation

The range of lithologies and sedimentary features seen in the Melby section between Matta Taing and Lang Rigg reflect the varying environment close to the margin of a large lake within the Orcadian Basin. Lake transgression and regression were the principal controlling influences. The fish beds, particularly the central part of the upper one, were deposited in reducing conditions during the acme of the transgression and maximum water depths, perhaps up to 50 m (Mykura and Phemister, 1976). Hall and Donovan (1978) described the laminated fish bed lithofacies as non-glacial varves deposited in a stratified tropical lake. Seasonal algal bloom resulted in increased photosynthesis and carbonate precipitation, the dead algae accumulating on the lake floor and being preserved as organic carbon. The clastic laminae formed either by continuous settlement from suspension or by small turbidity flow influx, perhaps with a seasonal or annual periodicity. Hall and Donovan (1978) emphasize one important difference between the Melby laminites and similar lithofacies elsewhere in the Orcadian Basin. At Melby, apparently uniquely, the clastic laminae consist largely of green clay, rather than mainly silt. The clay is a mixture of illite and chlorite and may have been derived by the weathering of a nearby volcanic massif. Hall and Donovan attributed the sulphide nodules in the Upper Melby Fish Bed to groundwater enriched by nearby volcanic activity.

The Lower Melby Fish Bed, and the top and bottom parts of the Upper Melby Fish Bed, are variable and record the irregular progress of lake transgressions and regressions. In these sections lacustrine sediments alternate with low-energy, fluvial sediments of channel, crevasse

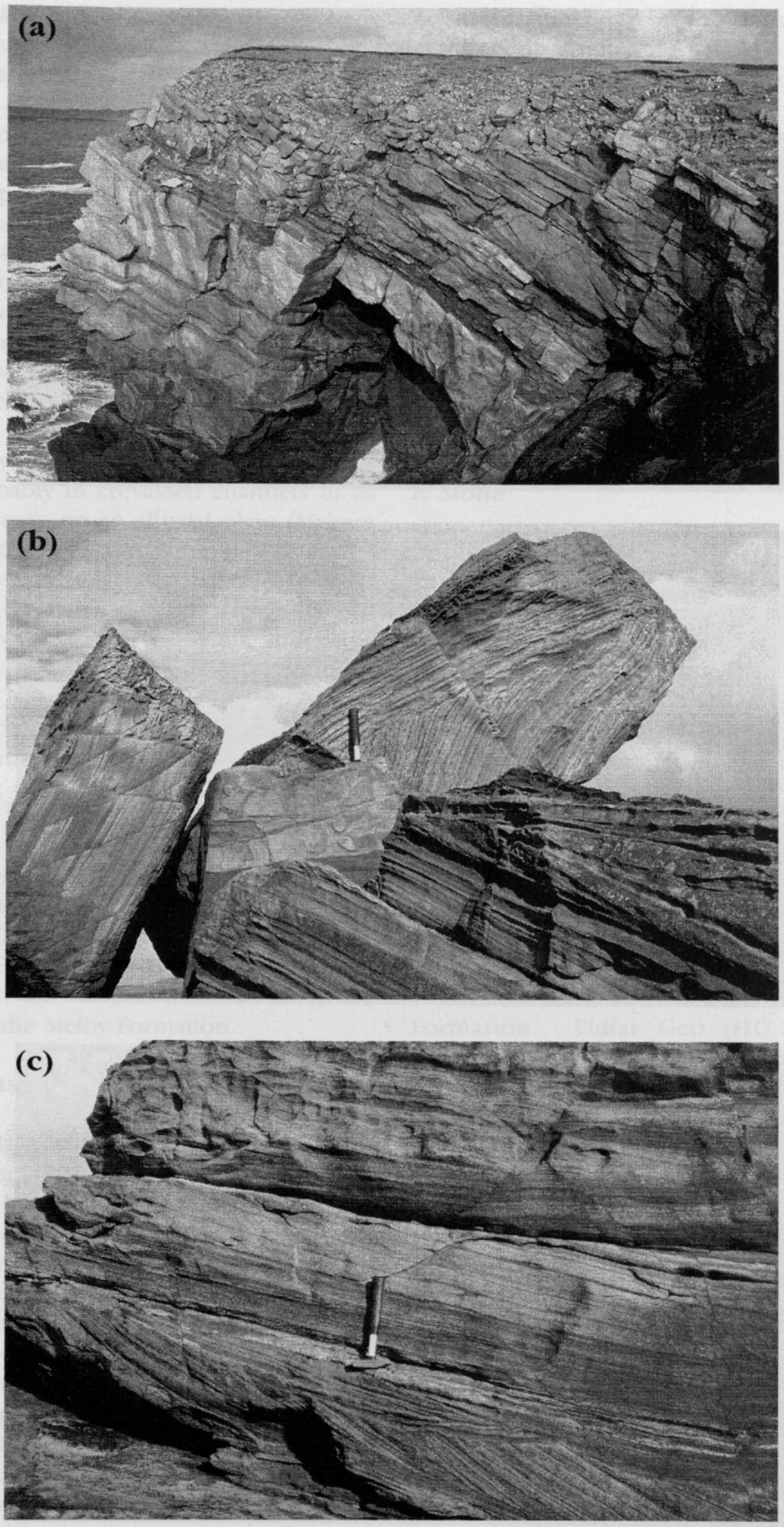


Figure 2.11 Sandstones between the two Melby fish beds. (a) Pobie Skeo; (b) large-scale planar cross-bedding; (c) small-scale trough cross-bedding. (Photos: P. Stone.)

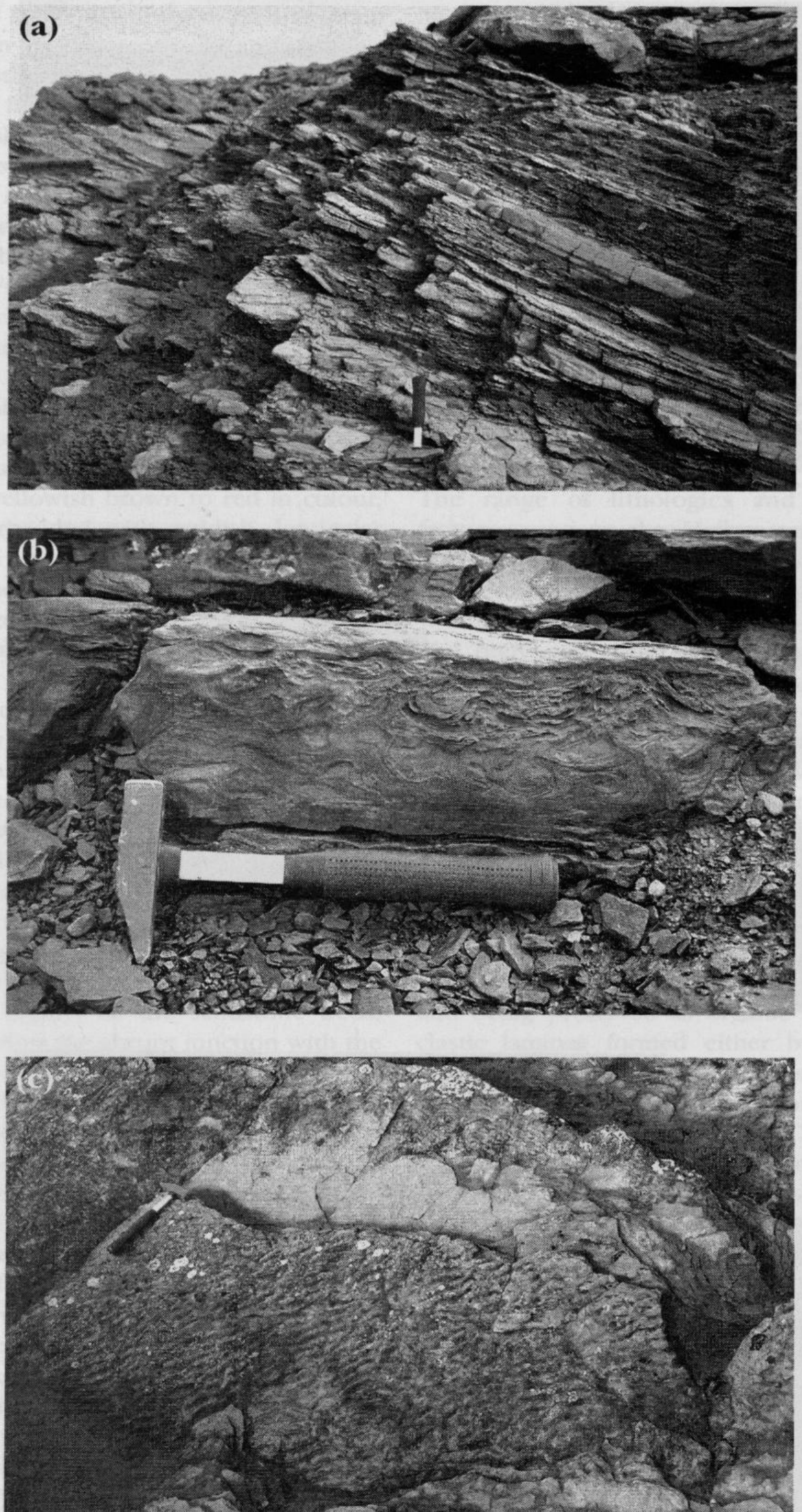


Figure 2.12 The Upper Melby Fish Bed, Rotten Craig. (a) Laminated mudstones with dolomitic layers; (b) convolute lamination in thin sandstone bed; (c) rippled siltstone bedding surface. (Photos: P. Stone.)

and overbank origin. Full lake regression allowed relatively high-energy fluvial systems to prograde into the shallow lake margins, depositing thick, cross-stratified sands. Convoluted tops to many of the cross-bedding sets suggest rapid de-watering of the sediment. Below the Upper Melby Fish Bed, the foresets of most planar cross-bedded sandstones dip eastwards, suggesting a current flow predominantly from the west. This broad uniformity suggests deposition in braided channels, with the sporadic finer-grained beds representing the remains of overbank sediment accumulations. Thinly bedded, ripple cross-laminated sandstones below and between the two fish beds were probably laid down by relatively sluggish currents, probably in crevassed channels or as overbank deposits on an alluvial plain (Mykura and Phemister, 1976).

Above the Upper Melby Fish Bed, palaeocurrent flow direction shows a reversal, with consistent foreset dips to the west, indicating current flow from the east (Mykura and Phemister, 1976). This change coincides with the appearance in the sandstones of significant quantities of mainly felsitic volcanic detritus, but also including some basalt. The volcanic rocks being eroded may have been the precursors to the Esha Ness and Papa Stour volcanic complexes, which are believed to overlie the Melby Formation. The identification by Hall and Donovan (1978) of volcaniclastic clay in the Upper Melby Fish Bed is further evidence for the active erosion of volcanic rocks during the deposition of the Melby Formation.

Conclusions

The cliffs and shoreline containing the Melby GCR site provide a well-exposed, representative section of the Middle Devonian (upper Eifelian–lower Givetian) strata in the westernmost of the Shetland structural tracts. The section demonstrates the varying depositional environments during lake transgression and regression, with lacustrine sediments interbedded within a dominantly fluvial sequence. The lacustrine strata are fine-grained, thinly bedded or laminated lithologies with carbonate (and locally sulphide) nodules. The carbonate nodules contain an important fossil fish fauna, allowing regional correlation with the Achanarras horizon of Caithness. Fluvial strata are mainly thick, cross-stratified, arkosic sandstones. Palaeoflow in the braided channel systems shows a marked change of direction within the succession exposed in the GCR site. The change coincides with a new volcanic source of detritus and is of regional palaeogeographical significance. Overall, the site provides important insights into lake evolution in the northernmost part of the Orcadian Basin and its fish fauna allows regional correlation with the Middle Devonian strata of mainland Scotland.

FOOTABROUGH TO WICK OF WATSNESS, SHETLAND (HU 179 502–HU 201 495)

P. Stone

Introduction

This section was originally named 'Fidlar Geo to Watsness' in the GCR archive. However, Fidlar Geo is in the middle of the section, and the south-east end of the site is better described as 'Voe of Footabrough', or simply as 'Footabrough'. At the north-west end of the site 'Watsness' should be either 'Wats Ness' (a location that is actually beyond the site boundary) or 'Wick of Watsness'.

The sea cliffs between Wick of Watsness and Voe of Footabrough on the west coast of the Walls Peninsula, Shetland Mainland, provide excellent, continuous exposure through folded strata of the Middle Devonian (Givetian) Walls Formation. Fidlar Geo (HU 190 493) is a prominent inlet in the central part of this section. The Walls Formation is part of the sequence lying between the N–S-trending Melby and Walls Boundary faults, and so forms part of the central Devonian structural tract within Shetland. Interpretation of the lithofacies present has proved difficult and controversial. The sequence is dominated by intercalated sandstone and shale that have been interpreted as turbidites deposited in a deep, lacustrine environment. However, a broad consensus from the most recent work (Melvin, 1985) is that a range of alluvial, braided stream, shoreline and shallow lacustrine environments are represented.

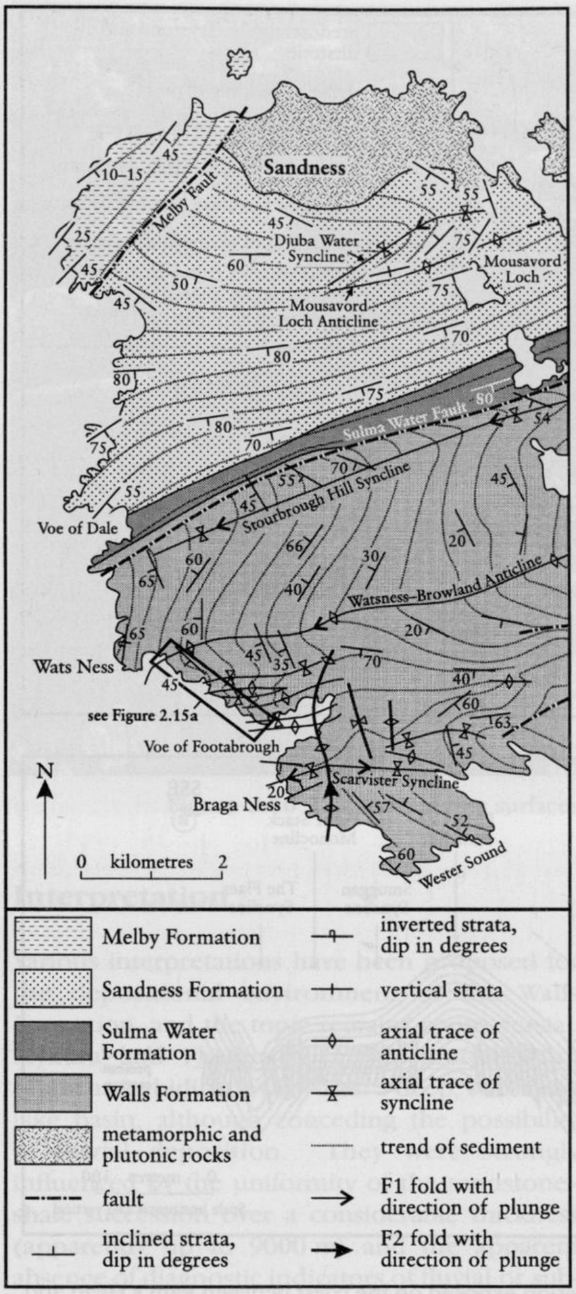
The regional importance of the site lies in the correct interpretation of its enigmatic lithofacies, thus allowing elucidation of depositional geometry within an otherwise poorly understood part of the Orcadian Basin. In this respect the site has an important role in characterizing the central north–south Devonian structural tract in Shetland. Detailed descriptions of the Walls Formation in the GCR site area are provided by Melvin (1976, 1985), Mykura and Phemister (1976) and Astin (1982). An overview of the geology is provided by Mykura (1976, 1991).

Description

Spectacular sea-cliffs along the coastline of the Walls Peninsula form the western extremity of Shetland Mainland. The cliffs expose an extensive section through part of the Walls Formation between Wick of Watsness and Voe of Footabrough (Figure 2.13). The sandstone-dominated formation was formerly thought to exceed 9000 m in thickness (Mykura, 1976), but may be much thinner (as discussed below). Only a part of this is exposed within the GCR site, but faulting and locally intense folding make thickness estimation very difficult. A sparse and fragmentary fish fauna, together with some indeterminate plant detritus, has been recovered from the Walls Formation (including three localities within the GCR site) (Mykura and Phemister, 1976), but is indicative only of a broadly Mid-Devonian age. Highly carbonized palynomorphs collected by Melvin (1985) suggest a Devonian age no older than Emsian, and a more recent study of miospores has confirmed a Givetian age (Marshall, 2000). An approximate upper age limit is provided by the 360 ± 11 Ma date (K–Ar) from the Sandsting Plutonic Complex, which intrudes the Walls Formation in the south-east part of its outcrop (Mykura and Phemister, 1976).

The Walls Formation lies between the Melby and Walls Boundary faults, and so forms part of the central of the north–south structural tracts that contain the Shetland Devonian succession (Figure 2.2). The strata are folded about a series of upright, open to close synclines and anticlines (Figure 2.14) that trend approximately ENE and plunge gently towards the WSW.

Figure 2.13 The Walls Formation at Cotti Geo, Ram's Head. (Photo: P. Stone.)



◂**Figure 2.14** Major stratigraphical and structural features of the Walls Peninsula. After Mykura (1976).

Hence, bedding dips range from horizontal to sub-vertical and strikes are variable (Figure 2.15a,b). A locally strong cleavage is developed broadly axial planar to the folds. At least some of the deformation has been related to the intrusion of the Sandsting Plutonic Complex (Mykura, 1991).

Mykura and Phemister (1976) noted the vertical and lateral lithological uniformity of the Walls Formation. It comprises stacked cycles, each consisting of a basal fine-grained, dark grey sandstone that passes up into a finer-grained, thinly bedded unit consisting, in varying proportions, of shale, siltstone and, rarely, impure limestone. Complete cycles range in thickness from 0.75 m to 20 m, with individual sandstones up to 18 m thick. The ratio of sandstone to shale and siltstone varies between different parts of the formation, as is well illustrated southwards along the coast from Wats Ness, where shale-rich cycles are dominant for up to about 20 m but then alternate with sandstone-rich cycles 25–45 m thick. Melvin (1976, 1985) estimated that various types of sandstone-rich cycles form almost 80% of the succession, with the shale-rich cycles making up the remaining 20%.

The sandstone beds generally have a sharp base, with flute and groove marks developed locally, resting on an erosion surface. Rip-up clasts of shale and siltstone are common, but nowhere abundant, in the lowest few centimetres of the sandstone. Cross-bedding of various types is widespread, ranging from large-scale, planar forms to small-scale, trough cross-lamination. In some places, concentrations of heavy minerals line the cross-bedding foresets. Convolute lamination is also widespread, and some examples are truncated at the base of the overlying sandstone bed. Texturally, most of the sandstone is fine- or medium-grained and fairly well-sorted, with a matrix content up to about 25% (much of which is carbonate). The main detrital components are mono- and poly-crystalline quartz, potassium feldspar (mainly orthoclase) and some plagioclase; accessories include muscovite, biotite, granite and garnetiferous quartz-mica gneiss (Melvin, 1976, 1985; Knudsen, 2000). The approximate overall modal proportions are: 70% quartz, 25% feldspar and 5% lithic fragments, which classify the sandstone as a subarkose. Detrital heavy minerals are abundant in some of the sandstones and may comprise up to 17% of the grains. Zircon, tourmaline and epidote are the most abundant, with less common sphene, rutile, clinozoisite and apatite.

The shale-rich parts of the succession contain many beds with well-developed ripple cross-

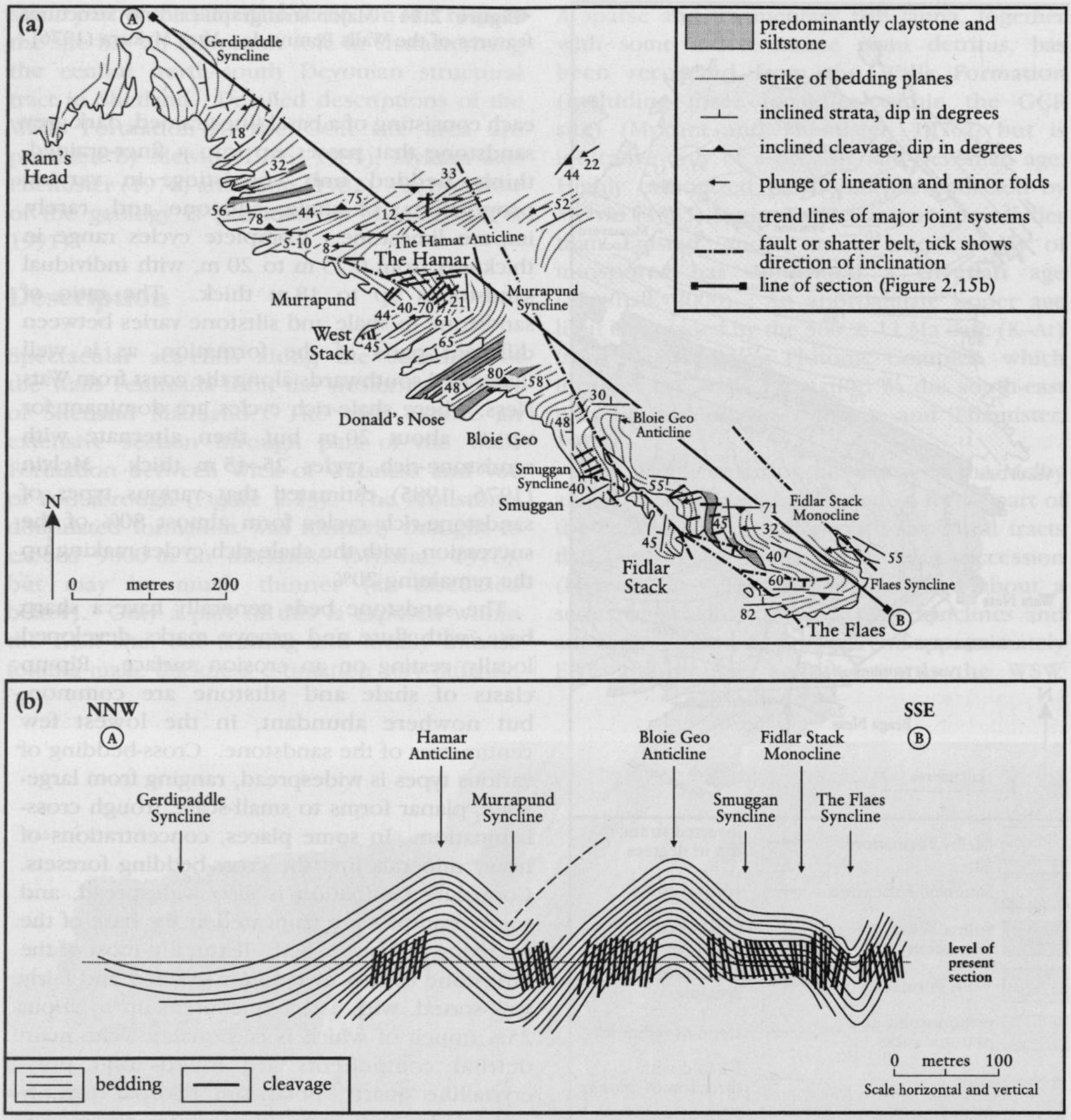


Figure 2.15 (a) Sketch map of the rocks of the Walls Formation exposed on the coast between Ram's Head and The Flaes. (b) Cross-section showing the structural pattern between Ram's Head and The Flaes. After Mykura and Phemister (1976).

lamination (Figure 2.16), which is locally accentuated by the coincidence of ripple crests with the bedding–cleavage intersection lineation. In some cases, successive beds show markedly different ripple orientations. Load casts are widespread and locally form pseudonodule layers. Desiccation polygons have been reported on some bedding surfaces by Melvin (1985), although Mykura and Phemister (1976) commented on their apparent absence from the Walls Formation. A typical section through the shale-rich lithofacies is shown in Figure 2.17a (Melvin, 1985). Thin beds of limestone and calcareous mudstone occur sporadically and are invariably finely laminated, with some included carbonaceous films. Most are disrupted by soft-sediment deformation, with subsequent preferential deformation by tectonic processes.

Figure 2.16 Rippled bedding surfaces in the Walls Formation. (Photo: P. Stone).

Interpretation

Various interpretations have been proposed for the depositional environment of the Walls Formation, and the topic remains controversial. Mykura and Phemister (1976) preferred an origin as turbidity flows within a deep, subsiding lake basin, although conceding the possibility of fluvial deposition. They were strongly influenced by the uniformity of the sandstone–shale succession over a considerable thickness (apparently up to 9000 m) and the apparent absence of diagnostic indicators of fluvial or subaerial environments. Examples of the latter, such as desiccation crack polygons, were subsequently discovered, and later workers (Melvin, 1976; Astin, 1982) preferred a shallow-water to fluvial depositional interpretation. Astin questioned the apparently great thickness of the succession and demonstrated sedimentological continuity northwards across the Sulma Water Fault (Figure 2.14) into the Sandness Formation, where fluvial depositional features are well developed. Further doubt was cast on the thickness estimate of 9000 m when Marshall (2000) showed from miospore evidence that both the Walls and Sandness formations were entirely Givetian in age.

The coast section forming the GCR site was shown by Melvin (1976, 1985) to consist largely of thick, multi-storey sequences of channelized, trough cross-stratified sandstone. Relationships between the successive channel units are complex, with cross-cutting scour surfaces underlying the bases of many sandstones. The assemblage of features suggests that in-channel deposition dominated, probably within a braided stream environment. A few of the sandstone-rich cycles show a particularly marked fining-upward trend and these were interpreted by Melvin (1985) to represent a progression of depositional environments from channel to sandbar, levee and floodplain (Figure 2.17b). The shale-rich cycles were interpreted by Melvin as shallow-water, floodplain deposits. Some thin, rippled sandstone beds within this overbank sequence were probably deposited from intermittent sheet floods, and the sporadic desiccation cracks suggest periods of subaerial exposure. More prolonged lacustrine deposition gave rise to the carbonate-rich beds, reflecting increased phytoplankton abundance.

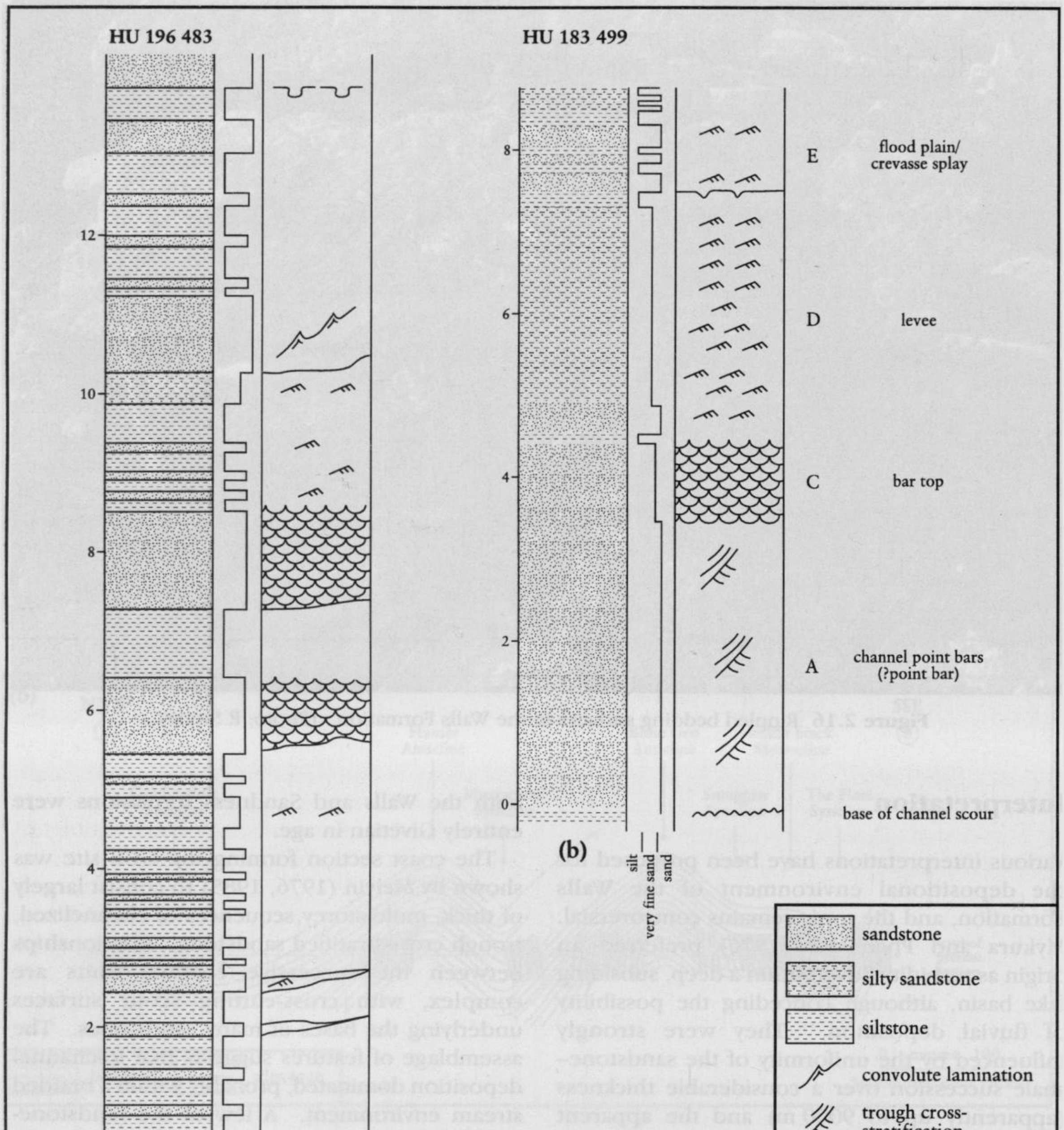


Figure 2.17 Examples of lithofacies associations in the Walls Formation. After Melvin (1985).

An important variation on a fluvial depositional interpretation was proposed by Astin (1982), who recognized a lateral facies transition from north-east to south-west. In his interpretation, the north-east part of the Walls Formation consists of an alluvial fan derived from the metamorphic rocks to the north (equivalent and probably contiguous with the Sandness

Formation). Towards the south-west the alluvial fan merges with more distal, lacustrine deposits and, most controversially, with beach ridges, the latter well represented within the GCR site as parts of the channel-bar and bar-top sandstone assemblages of Melvin (1976, 1985). The correct lithofacies interpretation is important for an accurate assessment of basin geometry and regional tectonics. The current consensus view has moved away from a deep lacustrine, turbidity-flow origin for the Walls Formation and now favours braided alluvial plain to shallow lacustrine and beach environments.

The tectonic deformation of the Walls Formation, involving two separate folding episodes (Mykura and Phemister, 1976), makes it difficult to deduce palaeocurrent flow directions from the sedimentary features. However, both Melvin (1976, 1985) and Astin (1982) deduced a broad flow regime from north to south, with sediment provenance to the north. The suite of detritus, together with its isotopic characteristics (Knudsen, 2000), suggests that the source was a composite granitic and metamorphic terrane with characteristics similar to those of the basement rocks now seen between the Melby and Walls Boundary faults. Thus, Devonian strata now occupying the central of the Shetland structural tracts could have been derived by erosion of the basement rocks from that same structural tract.

Conclusions

The Footabrough to Wick of Watsness GCR site provides a well-exposed section through part of the Walls Formation that is representative of the central of the Devonian structural tracts in Shetland. Sandstone, siltstone and shale combine in varying proportions to give broad alternations, over tens of metres, of sandstone-rich and shale-rich units with strong sedimentary cyclicity. The thick, sandstone-dominated units have a range of sedimentary features that have been interpreted in different ways. Turbidite deposition in a deep lake has been proposed, but the more recent consensus favours a combination of braided fluvial channel, littoral and shallow lacustrine environments. The interpretation of sedimentary depositional environments remains a subject of debate, but is of great importance in assessment of the regional basin geometry and tectonics.

EASTER ROVA HEAD, SHETLAND (HU 475 454)

P. Stone

Introduction

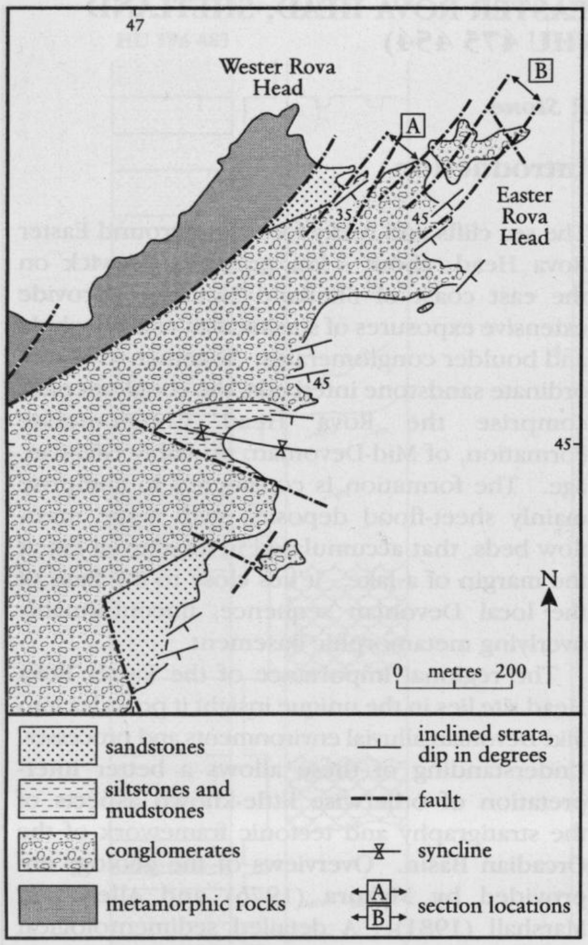
The sea cliffs and rocky foreshore around Easter Rova Head, about 4 km north of Lerwick on the east coast of Shetland Mainland, provide extensive exposures of spectacular, coarse cobble and boulder conglomerates. Together with subordinate sandstone interbeds, the conglomerates comprise the Rova Head Conglomerate Formation, of Mid-Devonian, probably Givetian, age. The formation is considered to comprise mainly sheet-flood deposits, with some mass-flow beds, that accumulated in an alluvial fan at the margin of a lake. It lies close to the base of the local Devonian sequence, unconformably overlying metamorphic basement.

The regional importance of the Easter Rova Head site lies in the unique insight it provides into Mid-Devonian alluvial environments and processes. Understanding of these allows a better interpretation of otherwise little-known aspects of the stratigraphy and tectonic framework of the Orcadian Basin. Overviews of the geology are provided by Mykura (1976) and Allen and Marshall (1981). A detailed sedimentological account of the site is given by Allen (1981a).

Description

At Easter Rova Head, the spectacularly coarse Rova Head Conglomerate Formation is well exposed in low sea-cliffs and across a rocky foreshore. The GCR site covers the island and the contiguous coastal outcrops on the mainland (Figure 2.18). Within the site area, the conglomerates and interbedded sandstones dip moderately to the south-east so that a vertical thickness of about 150 m is represented. The formation lies close to the base of the Devonian succession of eastern Shetland, within the structural tract east of the Nesting Fault (Figure 2.7) and is regarded as probably Givetian (Mid-Devonian) in age.

The clasts of the Rova Head Conglomerate Formation are rounded to well-rounded, generally with a moderate to high sphericity. They range up to about 75 cm in diameter and are poorly sorted. Quartzite is the most abundant clast lithology, but vein quartz and various granitic types are well represented,



◄**Figure 2.18** Outline geology of the Rova Head area. After Allen (1981a).

together with lesser amounts of psammite, felsite, schist and foliated granodiorite. Allen (1981a) recognized three types of conglomerate:

1. Clast- (or framework-) supported conglomerate, inversely graded at the base of the bed, ungraded and only crudely stratified in the centre of the bed, and normally graded at the top of the bed. This is the commonest type in the formation.
2. Matrix-rich or matrix-supported conglomerate, either ungraded or with inverse grading.
3. Clast- (or framework-) supported conglomerate with normal grading.

All three types occur at all levels in the succession, but there appears to be an up-sequence increase in the proportion of clast-supported types. However, many of the conglomerate beds have ambiguous characteristics and are not readily assigned to a specific category.

The lowest conglomerates within the site are seen at its north-west margin, along the shore of the North Bight of Rovahead. There, conglomerate beds range up to about 2 m, separated by thin sandstone beds and lenses up to about 20 cm thick (Figure 2.19). The conglomerates

Figure 2.19 Variably matrix- and clast-supported conglomerates with sandstone interbeds of the Rova Head Conglomerate Formation. (Photo: P. Stone.)

are variably matrix- or clast-supported, with both textures commonly present in the same bed where pockets of pebbles are concentrated in an otherwise matrix-rich background. Some of the beds are normally graded and fine upwards, but others have a well-developed, inversely graded, upward-coarsening trend. There is some sporadic, but seemingly irregular pebble imbrication. Both the matrix and the sandstone interbeds consist of coarse-grained sand, and the interbeds show planar and cross-lamination. There are sporadic thin beds of fine-grained sandstone and red siltstone, the latter containing desiccation cracks (Allen, 1981a). A graphic log of this part of the succession by Allen (1981a) is shown in Figure 2.20A.

Higher in the succession, along strike from the lighthouse and towards the South Bight of Rovahead, there is a tendency for the conglomerates to be clast-supported rather than matrix-rich (Figure 2.21). There is also an over-all up-sequence (SE-directed) trend to thinner beds (although still ranging up to about 2 m maximum thickness), a smaller proportion of conglomerate relative to sandstone (although

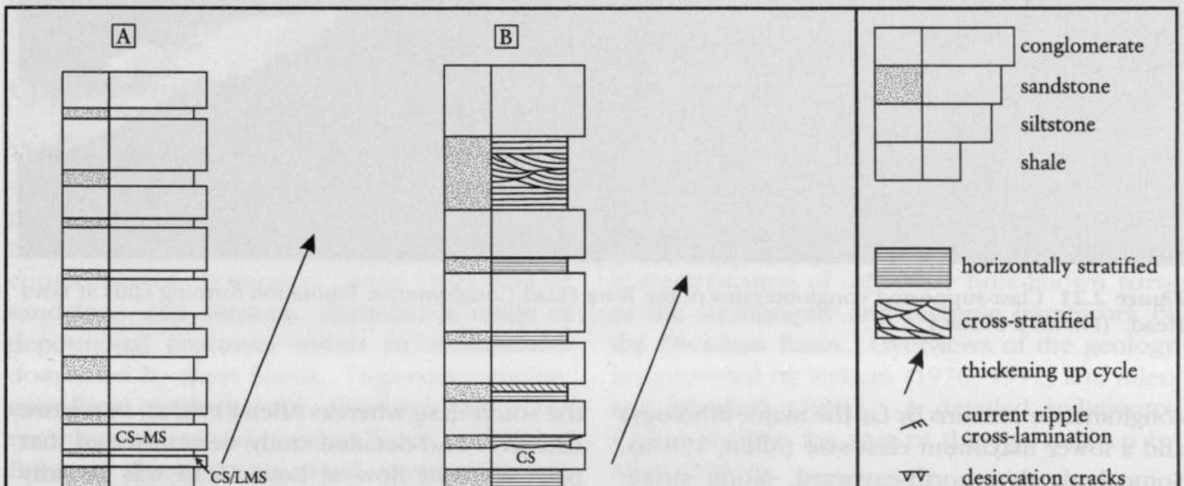


Figure 2.20 Graphic logs of part of the Easter Rova Head succession. See Figure 2.18 for the location of sections A and B. After Allen (1981a).

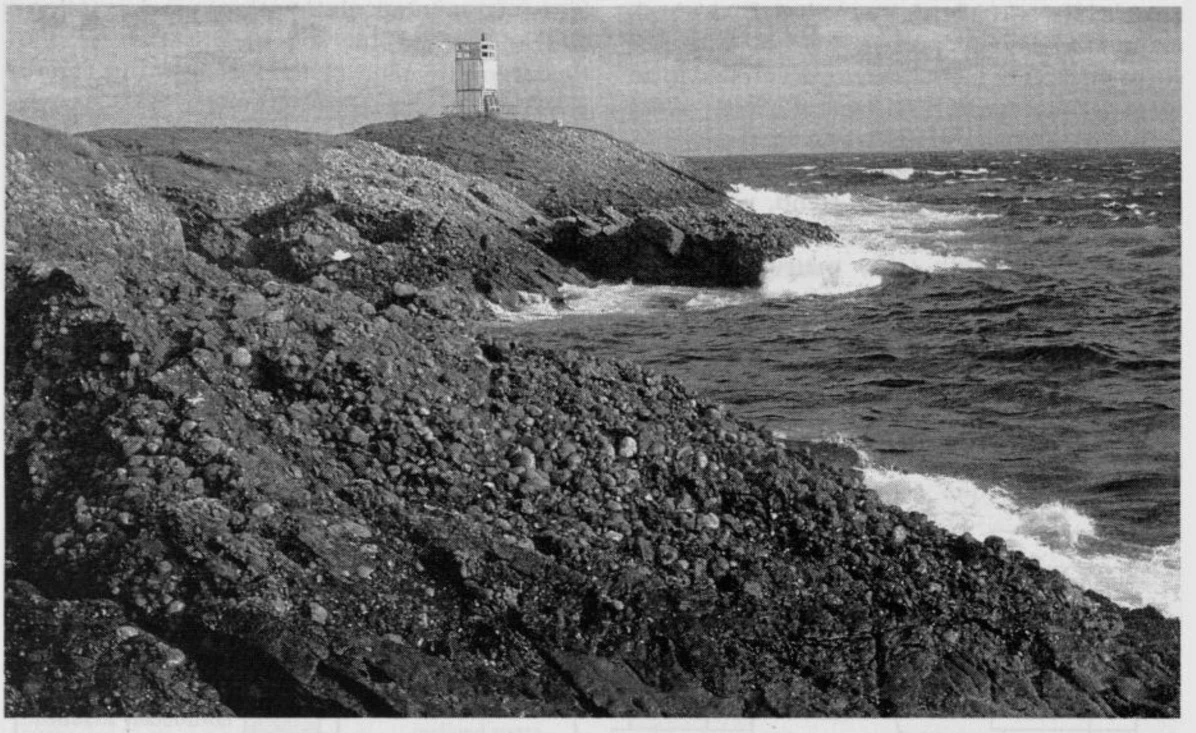


Figure 2.21 Clast-supported conglomerates of the Rova Head Conglomerate Formation forming cliffs at Rova Head. (Photo: P. Stone.)

conglomerate remains by far the major lithology) and a lower maximum clast size (Allen, 1981a). Some beds thin north-eastward, along strike. There are apparently no systematic changes, either with stratigraphical level or laterally, in clast compositional range, shape or modal size. Sandstone interbeds in the higher part of the sequence are rather more extensive than those lower down, but there are no compositional or textural changes, and both planar and cross-stratified beds are present throughout. A graphic log of the higher part of the succession within the GCR site, taken from Allen (1981a), is shown in Figure 2.20B.

Interpretation

The Rova Head Conglomerate Formation lies close to the base of the Devonian succession and interdigitates laterally towards the south-west with a breccia derived from, and believed to unconformably overlie, a basement of metamorphic rocks (Mykura, 1976). Palaeocurrent analysis shows the conglomerates and interbedded sandstones to have been derived from a broadly western source. Mykura (1976) recorded regional palaeocurrent flow towards the south-east, whereas Allen (1981a), in a more extensive and detailed study, demonstrated that palaeocurrent flow at Rova Head was broadly towards the north-east. From petrological and Sm–Nd isotope studies, Knudsen (2000) identified relatively local basement sources for the clast assemblage, all to the east of the Walls Boundary Fault and within the same structural tract as the Devonian strata.

Allen (1981a) concluded that the Rova Head conglomerates accumulated as successive sheet-flood deposits in a low-inclination alluvial fan at the margin of a lake. The inverse to normally graded, mostly clast-supported conglomerates were deposited as thick sheets and bars on a surface with little topographic relief. Finer material was rapidly transported during rising flood conditions and deposited as poorly sorted, unstratified gravels. As the flood developed, increased water flow transported and deposited coarser material, producing inverse grading. As the flood waned, normal grading and horizontal stratification were produced in progressively decreasing concentrations of sediment. The matrix-supported conglomerates, many of which are inversely graded, record high-concentration, mass-flow events.

Sand wedges were deposited over and against the coarse gravel bodies as the current flow waned. Thicker interbeds of horizontally or cross-laminated sandstone represent the migration of sand waves over the gravel bodies and are probably unrelated to the sheet-flood events. However, some of the most extensive sandstone units may have been deposited during small sheet-floods restricted to broad, shallow channels. Flood events may have also produced overbank floodplain deposition, represented by the sporadic red siltstone interbeds. Desiccation cracks in these siltstones show that the floodplain was subjected to periodic subaerial conditions in an arid climate (Allen, 1981a).

Conclusions

The Easter Rova Head GCR site provides a well-exposed and instructive section through part of an alluvial-fan sequence developed at the margins of a Mid-Devonian lake. The conglomerate sequence, with subordinate sandstone and siltstone, illustrates a range of depositional processes within an environment dominated by sheet floods. High-concentration, mass-flow conglomerates alternate with those recording rising and waning flood conditions. Between flood events, sand waves migrated across the conglomerate bodies in channelized fluvial deposition, and fine-grained overbank deposition and subaerial conditions were intermittently established. The conglomerates have a local source, and their presence provides an important control on the geometry of the depositional basin and its geotectonic evolution. The site is therefore of great importance in the regional understanding of the Orcadian Basin.

THE CLETTS, EXNABOE, SHETLAND (HU 409 137–HU 406 122)

P. Stone

Introduction

The sea cliffs between Vaakel Craigs and Millburn Geo, on the south-east coast of Shetland Mainland near Exnaboe provide excellent, continuous exposure through varied lacustrine, fluvial and aeolian strata of the Middle Devonian Brindister Flagstone Formation. The Cletts is a large and prominent bedding surface of pebbly sandstone within this section. The sequence lies to the east of the N–S-trending Nesting Fault, and so forms part of the easternmost of the three Devonian structural tracts of Shetland. The lithofacies present were produced during the cyclical deepening and shallowing of a large lake occupying part of the Orcadian Basin. Highstands of the lake are marked by two units of laminated mudstones, one including thin dolostone layers and lenses. The fossil fish fauna from this locality is of great importance, and is described in the GCR fossil fishes volume (Dineley and Metcalf, 1999). During lowstand periods of shallow-water, alluvial-fan gravels and deltaic sands prograded into the lake, and aeolian dunes prograded across the basin during periods of subaerial emergence.

The regional importance of The Cletts (Exnaboe) site lies in the well-preserved evidence for the cyclicity of depositional environments. This allows interpretation and characterization of otherwise little-known parts of the stratigraphy and tectonic framework of the Orcadian Basin. Overviews of the geology are provided by Mykura (1976, 1991) and Allen and Marshall (1981). A detailed sedimentological analysis for part of the site is given by Allen (1981b).

Description

The incised sea-cliffs forming the south-east coast of Shetland Mainland provide extensive exposure through a sequence of fluvial, lacustrine and aeolian Old Red Sandstone strata. The section spanning The Cletts (Figure 2.22), near the village of Exnaboe, lies within these cliffs, about 5 km north of Sumburgh Head. The strata are informally assigned to the Brindister Flagstone Formation and are of Givetian (Mid-Devonian) age. The beds dip consistently between 15° and 30° towards the south-east, resulting in several hundred metres of strata being exposed within the GCR site, although the continuity is locally disrupted by minor faulting, particularly at the northern end of the section. The section lies in the easternmost of three north–south, fault-bounded tracts of Devonian rocks in Shetland, and is characteristic of the strata in the tract.

The oldest strata present within the site crop out at its northern margin in the vicinity of

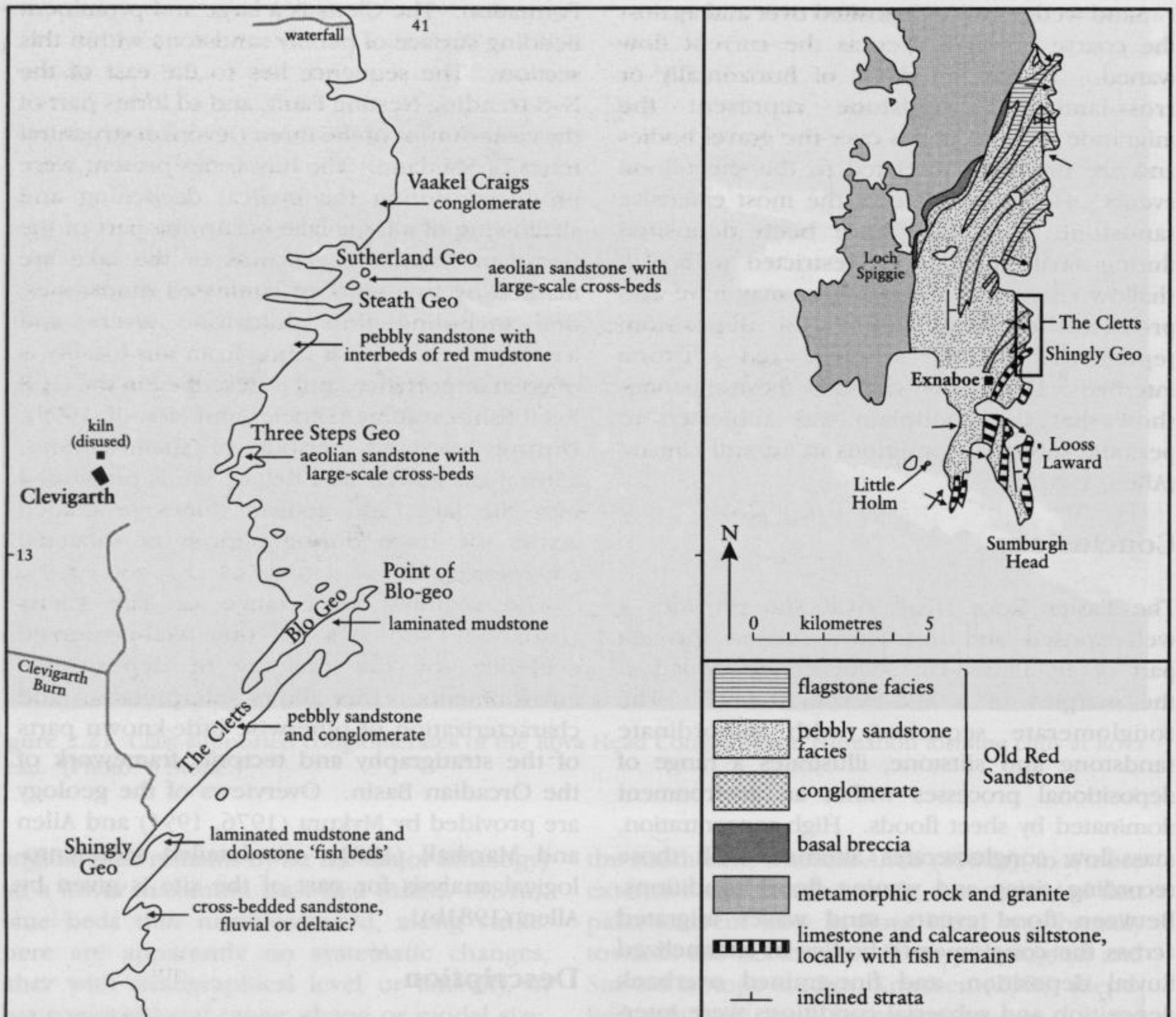


Figure 2.22 Geological map of the Old Red Sandstone of south-east Shetland Mainland and detailed map of the Cletts GCR site. After Mykura (1976).

Vaakel Craigs. There, coarse pebble and cobble conglomerate (Figure 2.23) is interbedded with thin, lenticular bodies of coarse-grained sandstone in a sequence at least 30 m thick. The conglomerate clasts range up to about 30 cm across, but most are 5–10 cm. The conglomerates occur in irregular bodies, either matrix- or clast-supported, in which clast imbrication is only weakly and locally developed. The clasts are predominantly of vein quartz and quartzite (both massive and laminated), but sporadic pebbles of granite (some of which are foliated) and gneiss are also present. Most clasts are subangular to well-rounded, but the degree of sphericity is variable. The interbedded sandstone lenses are commonly cross-bedded.

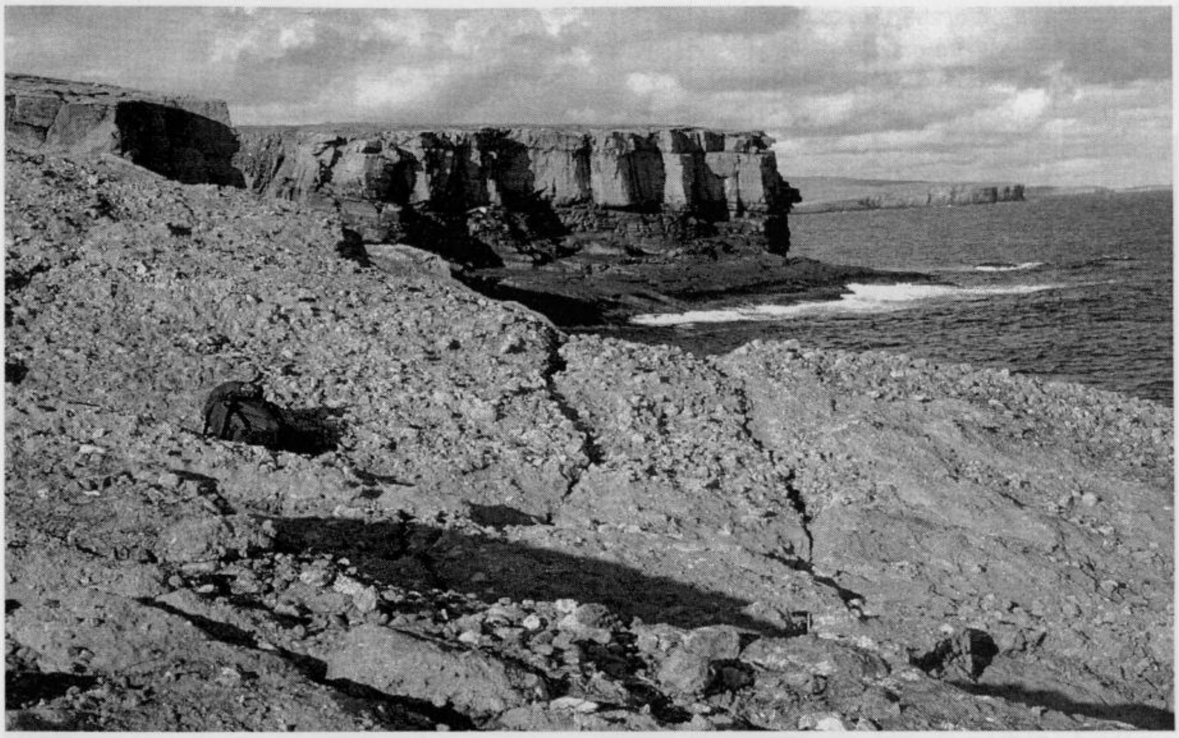


Figure 2.23 Conglomerates at Vaakel Craigs. (Photo: P. Stone.)

Southwards, and up the succession, the coarse conglomerate is succeeded conformably but quite abruptly by cross-bedded sandstone in thick sets up to about 2 m high. About 20 m of this cross-bedded sandstone form the cliffs around Sutherland and Steath Geo. The sandstone becomes more pebbly upwards, but the clasts are generally isolated and supported within the matrix. The clasts are generally less than 10 cm, with a few up to 25 cm, mostly rounded or well-rounded and have a high degree of sphericity. Most of them are of quartzite, commonly with a pinkish hue, but there are also some rare examples of granitic lithologies. Interbedded with the pebbly sandstone are thinly bedded units, up to about 50 cm thick, of red, laminated mudstone and siltstone.

South from Steath Geo, and up-sequence, the pebbles decrease in abundance and thick, cross-bedded sandstone bodies become dominant. These spectacular units are well exposed around Three Steps Geo (Figure 2.24), where individual sets range up to 2 m in thickness; sets up to 5 m thick (Mykura, 1976) and up to 3 m thick (Mykura, 1991) have been reported from elsewhere in this area. The sandstone is pale yellowish brown, variably quartzose or arkosic, and commonly with rounded grains. At least 50 m of the cross-bedded sandstone form the coastal cliffs between Steath Geo and Blo Geo, where they are abruptly overlain by about 4 m of very thinly bedded, laminated mudstone, siltstone and fine-grained, ripple-laminated sandstone, the last showing grading to siltstone over less than 5 mm. Preferential erosion of this unit created the deep inlet of Blo Geo.

Above the Blo Geo unit, there is an abrupt return to fairly massive pebbly sandstone. The concentration of pebbles increases irregularly upwards and most of them consist of quartz or quartzite; the pebbles range from subangular to well-rounded, sphericity is highly variable, and most are matrix-supported. Crude, large-scale cross-stratification is present locally. This lithology forms the prominent headland at Point of Blo-geo, and The Cletts is an extensive bedding surface running south-west from that point. Shingly Geo lies immediately south of The Cletts and is another example of the preferential erosion of a fine-grained, thinly bedded unit (Figure 2.25), this one abruptly overlying The Cletts pebbly sandstone.

About 4 m of laminated mudstone, siltstone and fine-grained sandstone with much small-

Figure 2.24 Planar cross-bedded, probably aeolian, sandstones, Three Steps Geo. (Photo: P. Stone.)

Figure 2.25 The Cletts (bedding plane of pebbly sandstone, left), and gully eroded through the fish bed. (Photo: P. Stone.)

scale cross-bedding and extensive rippled bedding planes are exposed at Shingly Geo. Within the top half of the unit are thin interbeds of dolomitic limestone that become more abundant and thicker upwards, to a maximum of about 5 cm. Fossil fish fragments have been recovered from this part of the sequence, the Exnaboe Fish Bed. Although not abundant, they include *Dipterus* sp., the commonest species, and *Stegotrachelus finlayi*, the second commonest. This is the only species of the genus and is found only in Shetland (Dineley, 1999a). The limestone layers are disrupted by soft-sediment structures, with evidence for both extensional and compressional deformation (Figure 2.26). There is also evidence for tectonic deformation, with an incipient cleavage cutting obliquely across the bedding and crenulating the more finely laminated lithologies. The exposure at Shingly Geo was modified by some limited, long-ceased quarrying, the limestone probably being burnt at the old kiln adjacent to the ruins of Clevigarth, about 500 m to the north.

The Shingly Geo unit is abruptly overlain by fine- to medium-grained, grey, micaceous, cross-laminated sandstone in beds generally 40–60 cm thick, with some up to about 1 m. This lithofacies continues to and beyond the southern margin of the GCR site.

Interpretation

The range of lithologies present within The Cletts GCR site represents deposition in the cyclically varying environments at the margin of a large lacustrine basin. The complexity of the facies relationships was stressed by Allen (1981b), who recognized the influence of both wave and current action in water depths of less than 5 m, and a deeper, offshore environment characterized by a lack of evidence for either wave- or current-action. The deep-water environment, during phases of maximum lake extent is represented by the thinly bedded units at Blo Geo, and, particularly, Shingly Geo. In both cases, mudstone/siltstone laminites were probably deposited from suspension of seasonal influx of clastic sediment, with the fine sandstone to siltstone gradations perhaps having a turbiditic origin. The carbonate laminae, lenses and thin beds may have formed during periods

Figure 2.26 Disrupted dolomitic layers in lacustrine laminites. Exnaboe Fish Bed at The Cletts. (Photo: P. Stone.)

of increased phytoplankton abundance, but some diagenetic re-distribution of carbonate seems likely. The primary bedding was further disrupted by soft-sediment deformation, facilitated by the earlier lithification of the dolostone, allowing it to behave in a brittle fashion while the enclosing clastic laminae were still ductile. The fish fossils at Shingly Geo are further evidence that this unit marks the maximum lake extent and probably the deepest water seen in the section. Water depths may have been up to 80 m (Hamilton and Trewin, 1988).

The pebbly sandstones that make up much of the lower part of the succession, and also separate the two thinly bedded lacustrine units, are probably braided stream and alluvial-fan deposits that prograded into the lake during periods of relatively low water-level. The matrix-supported nature of much of the pebbly sandstone suggests that slumping and mass-flow processes were also active. In their lower part, the pebbly sandstones are sporadically interbedded with thin red mudstone layers, representing overbank facies. The cross-bedded, fine-grained, micaceous sandstones at the top of the succession within the GCR site were probably deposited in prograding deltas during a relative fall in the lake water-level. The regional current flow pattern indicates dispersal of sediment in a general south-easterly direction (Mykura, 1991). The preponderance of quartz and quartzite clasts in the pebbly sandstones and conglomerates suggests that those lithologies were dominant in the sediment provenance area, although the high percentage of well-rounded quartzite pebbles may be due to recycling of older conglomerates. The granite and gneiss clasts are more positively ascribed to basement rocks that were probably exposed in a north–south mountain range that formed the western margin of a large lacustrine basin.

The lowest water-levels within the lake allowed aeolian sand dunes to advance across areas that had once been underwater. The large-scale, cross-bedded sandstones around Sutherland, Steath and Three Steps geos are good examples of such dune sandstones (Mykura, 1976, 1991). They contain rounded grains of quartz and feldspar, occur in sets up to 5 m thick, and appear to have been driven by winds from the south and south-west (Allen and Marshall, 1981).

Conclusions

The Cletts, Exnaboe GCR site provides a well-exposed section representative of the Devonian strata in the easternmost of the Shetland structural tracts. It demonstrates the interplay of depositional environments at the margin of a large lake within the Orcadian Basin. Fluctuation in lake water-level produced a vertical succession of stacked cycles, in which lacustrine, locally fish-bearing, fine-grained laminites alternate with braided stream, alluvial-fan and ?deltaic sandstones. Low lake water-levels allowed the encroachment of aeolian dunes. The succession exposed in this section provides important insights into the evolution of an otherwise little-known part of the Orcadian Basin. The importance of this site is further enhanced by the presence of the Exnaboe Fish Bed, which contains *Stegotrachelus finlayi*, the only species of this genus, which is unique to Shetland. The fish bed is younger than others in Shetland, and may be the only late Givetian site this far north.

SOUTH STROMNESS COAST SECTION, ORKNEY (HY 223 102–HY 254 078)

E.A. Pickett

Introduction

The South Stromness Coast Section GCR site lies on the south-west coast of the Orkney Mainland and extends along the coast from the Skerry of Ness, at the mouth of Stromness harbour, to Billia Croo, about 3 km WNW of Stromness. This stretch of coastline is excellently exposed and accessible, and provides the best section in Orkney through the Lower and Upper Stromness Flagstone formations. These formations make up the Middle Devonian Caithness Flagstone Group (approximately equivalent to the Middle Old Red Sandstone) in Orkney. At the eastern end of the site, the Lower Stromness Flagstone Formation rests unconformably, with a basal conglomerate, on Precambrian metamorphic rocks, exposed here in the largest of a few small inliers of basement in Orkney. Information from the overlying sequence has been critical in the recognition and palaeo-environmental interpretation of the rhythmic units or 'cycles' that characterize the Middle

Devonian fluvio-lacustrine sequences of the Orcadian Basin (Fannin, 1970; Mykura, 1976). Features of particular sedimentary and palaeoenvironmental significance include impressive desiccation cracks, ripples, early diagenetic chert nodules, beds of algal debris, and large stromatolitic mounds.

The section also contains fault-repeated exposures of the Sandwick Fish Bed, the base of which is the junction between the Lower and Upper Stromness Flagstone formations (Astin, 1990). This horizon is important in regional correlation and justly famous for its fossil fishes (see also **Yesnaby and Gaulton Coast Section** GCR site report, this chapter).

Description

At the Skerry of Ness, which lies at the eastern end of the site (Figure 2.27), the Lower Stromness Flagstone Formation overlies a granite-gneiss basement surface that slopes gently to the south. This is one of a few small inliers of pre-Devonian basement which lie along an axis stretching from the island of Graemsay (directly south of Stromness) through Stromness to Yesnaby on the western coast of the Mainland (see also **Yesnaby and Gaulton Coast Section** GCR site report, this chapter). The basement is composed of pink, foliated feldspar-phyric granite with enclaves of foliated metamorphic rocks that resemble rocks of the Lewisian inliers within the outcrop of the Moine Supergroup of the Northern Highlands (Wilson *et al.*, 1935).

Unconformably overlying the basement are 5–20 m of coarse, poorly sorted, clast-supported breccia and conglomerate. These form the basal part of the Lower Stromness Flagstone Formation and consist of subrounded to angular clasts of gneiss, granite, schist and quartzite set in an arkosic, sandy matrix. The clasts are generally cobble-sized (up to 20 cm by 10 cm), although some are up to 50 cm. Fannin (1970) noted that some of the clasts show a poorly developed imbrication, dipping westward off the crystalline mass. Some of the clasts have thin algal coatings of interlaminated dolostone and siltstone. Above the conglomerates, there are further examples of algal coatings, as well as stromatolitic mounds, mats and ridges. The basal facies passes up into finer-grained conglomerates and arkosic, pebbly sandstones. The sequence fines markedly upwards, with pebble content decreasing and with very few pebbles in the beds higher than about 25 m above the base.

The BGS Warebeth Borehole (Mykura, 1976; Clarke, 1990) lies in the site, providing a further point of interest. The borehole proved 61 m of red-purple siltstones, sandstones and breccias (the 'Warebeth Red Bed Formation'), which are absent from the succession exposed on the coast. They lie above the metamorphic basement and below the basal conglomerate of the

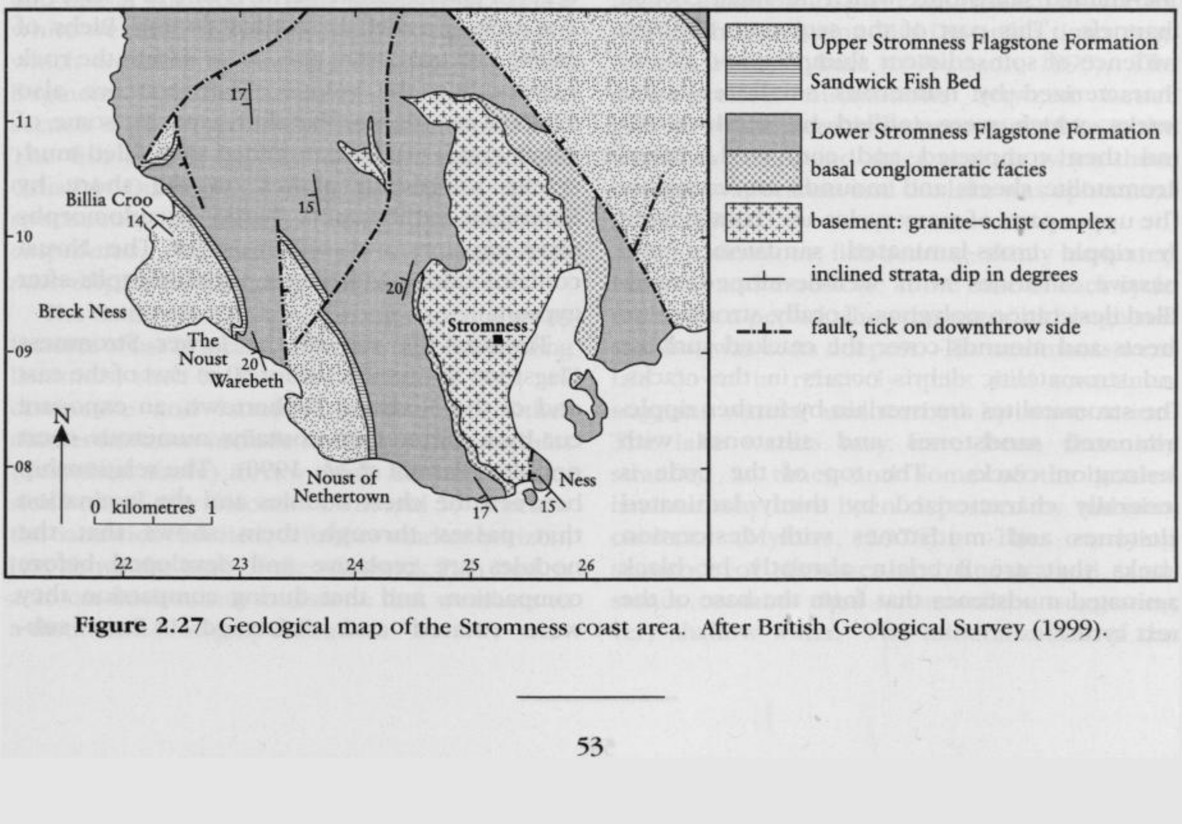


Figure 2.27 Geological map of the Stromness coast area. After British Geological Survey (1999).

Lower Stromness Flagstone Formation and are tentatively assigned to the Lower Old Red Sandstone (Lower Devonian).

The Lower Stromness Flagstone Formation is about 250 m thick and the Upper Stromness Flagstone Formation at least 500 m thick. Both formations, with the exception of the basal sequence of the Lower Stromness Flagstone Formation, comprise a series of cyclic units, each unit or 'cycle' ranging from 2 m to 15 m thick. Fannin (1970) studied the sequence in detail and subdivided the units into the Lower Stromness Flags (24 cycles), the 'Sandwick Fish Bed Cycle' (1 cycle), the 'Hoy Cycles' (4 cycles) and the Upper Stromness Flags (20+ cycles). The classification was subsequently simplified so that the informally defined 'Sandwick Fish Bed Cycle' and 'Hoy Cycles' of Fannin (1970) now form part of the Upper Stromness Flagstone Formation (Astin, 1990). The facies of a typical cycle were described by Mykura (1976) and are summarized below.

The base of each cycle is taken at the appearance of dark grey to black, calcareous, silty mudstone interlaminated with siltstone. This facies generally contains graded laminae that may have either a high bitumen and pyrite content or a high carbonate (mostly ferroan dolomite) content. Fossil fish remains may also be present. The dark mudstones and siltstones pass up into thinly interbedded bituminous siltstones and fine-grained sandstones, together with some discrete beds of massive siltstone and fine-grained sandstone, which fill small eroded channels. This part of the sequence contains evidence of soft-sediment slumping and is also characterized by numerous small lenticular cracks, which were infilled by sand or silt and then compacted and contorted. Algal stromatolite sheets and mounds are common. The upper parts of many cycles are characterized by ripple cross-laminated sandstones and massive siltstones with well-developed sand-filled desiccation polygons. Locally, stromatolite sheets and mounds cover the cracked surfaces and stromatolitic debris occurs in the cracks. The stromatolites are overlain by further ripple-laminated sandstones and siltstones with desiccation cracks. The top of the cycle is generally characterized by thinly laminated siltstones and mudstones with desiccation cracks that are overlain abruptly by black laminated mudstones that form the base of the next cycle.

Much of the sequence is rich in ferroan dolomite, causing the beds to weather to distinctive yellow, orange and buff colours. Fresh surfaces are commonly dark grey owing to their organic content. One of the most impressive features of the section is the presence of extensive, gently dipping bedding planes (Figure 2.28) covered with wave- and current-ripple marks or infilled desiccation cracks.

The algal stromatolites are an interesting and important feature of this site. About 800 m west of the Point of Ness are large (metre-wide) stromatolite-covered mounds (HY 2510 0765), described in detail by Fannin (1969). The mounds were built around coarse sand cores on an irregular erosion surface. The sandy cores are coated with linked hemispheroidal stromatolites and flanked by breccias of stromatolitic debris. The stromatolite-bearing sequence is locally silicified, with chert layers occurring at the top of some mounds. The succession above and below the mounds is rich in stromatolitic debris and smaller stromatolite developments are common in the beds above the mounds. Most of the stromatolitic rocks are mineralized by galena and sphalerite.

Small shows of mineralization, particularly galena-sphalerite-barite, are common elsewhere throughout the section. Barite veins also cut the metamorphic basement. Lead was mined at Warebeth Beach around 1775 and an infilled adit is still visible. Below the adit on the beach, a brecciated zone of normal faulting that contains veins of galena-calcite-barite is one of a series of 2 m-wide, parallel, brecciated zones. Blebs of galena and sphalerite also occur within the rock surrounding the veins. The sequence also exhibits pseudomorphs after gypsum, some of which appear to have modified sand-filled mud-cracks. Bedding planes on the shore by Stennigor exhibit pink barite pseudomorphs after gypsum and one west of The Noust contains calcite and quartz pseudomorphs after gypsum.

Towards the top of the Lower Stromness Flagstone Formation, about 50 m east of the east end of the Noust of Nethertown, an exposure cut by a minor fault contains numerous chert nodules (Parnell *et al.*, 1990). The relationship between the chert nodules and the lamination that passes through them shows that the nodules are replacive and developed before compaction, and that during compaction they were rotated and deformed. Rare sub-

Figure 2.28 Upper Stromness Flagstone Formation at Warebeth. View south-west towards Hoy. (Photo: E.A. Pickett.)

millimetre sized sphalerite crystals in the chert are thought to be syndiagenetic (Parnell, 1987).

The Sandwick Fish Bed, the base of which marks the junction between the Lower and Upper Stromness Flagstone formations, crops out at three localities in the site owing to fault repetition – two localities at the Noust of Nethertown, and one at The Noust (near Breck Ness). It is a white-weathering carbonate laminite, above which are black lenses of chert and fossil fish fragments. At the Noust of Nethertown the section is deformed by N–S-trending box folds. The fish bed is rich in fossil fish specimens, Cruaday Quarry GCR site yielding 14 or 15 specimens, including common *Coccosteus cuspidatus*, *Mesacanthus* sp. and *Osteolepis macrolepidotus* (Dineley, 1999a).

The Caithness Flagstone Group contains many features of use in local correlation. The fish-bearing laminites, such as the Sandwick Fish Bed, are distinctive and laterally persistent. Other marker horizons include the large chert nodules (described above) in two thinly laminated iron-rich beds of silty dolomite 14 m and 59 m below the Sandwick Fish Bed. A further marker horizon, just above the base of the Sandwick Fish Bed, is a 25 cm-thick calcite mudstone that weathers to a distinctive blue-grey colour.

Interpretation

This site has the best section in Orkney of the rhythmic units or 'cycles' that form the Caithness Flagstone Group. These cycles are interpreted as reflecting fluctuations in water-level and sediment input in a major lacustrine rift basin, the Orcadian Basin, that extended across Orkney and Caithness during Mid-Devonian times (e.g. Fannin, 1970; Mykura, 1976; Astin, 1990). The cycles reflect fluctuations between wetter periods when there was a permanent lake in the Orcadian Basin, and drier periods when there was net evaporation and the lake was ephemeral (Astin, 1990).

The calcareous, fish-bearing, thinly laminated siltstone and mudstone at the base of each cycle represent a period when the water, although still shallow, was at its deepest. The sediments were laid down in relatively quiet and sometimes stagnant waters, undisturbed by wave action. The lake waters may have been thermally stratified at times and some of the graded laminae may have been deposited by turbidity currents (Mykura, 1976). The overlying sequence of siltstones and fine-grained sandstones contains algal stromatolites, suggesting very shallow water. The mineralization of the stromatolites by galena and sphalerite may also have environmental implications, Muir and Ridgway (1975) suggesting that the degradation of algal material produced organo-metallic complexes that precipitated sulphides in the presence of sulphide-reducing bacteria. Numerous small cracks, originally interpreted as subaqueous synaeresis cracks (e.g. Fannin, 1970; Donovan and Foster, 1972) have been re-interpreted as pseudomorphs of gypsum (Astin and Rogers, 1991; Rogers and Astin, 1991), but this remains controversial (Astin and Rogers, 1992, 1993; Trewin, 1992; Barclay *et al.*, 1993; Trewin and Thirlwall, 2002).

Progressive shallowing of the lake produced mudflats that periodically dried to produce polygonal arrays of desiccation cracks. The presence of barite, quartz and calcite pseudomorphs after evaporite minerals such as gypsum suggest a playa-lake setting. The lake muds were covered by fine-grained, rippled sands and silts of alluvial fans that prograded over the shallow or dry lake floor. These fluvial and deltaic sediments were succeeded by thin-bedded silts and sands with mud-filled desiccation cracks, abruptly overlain by deeper-water fish-bed facies at the base of the next cycle.

The cyclicity of the sequences, together with the presence of wave-reworked surfaces, pseudomorphs after gypsum and mudcracks, reflects the frequent oscillation between shallow temporary lake and desiccated lake basin, the former estimated to have lasted in the range of 100–200 years to 5000–10 000 years (Astin, 1990). Palaeocurrent analysis of the sequence indicates that there was a dominant current flow towards the south. Fannin (1970) suggested that the Orcadian lake had a roughly E–W-trending shoreline and was fed by rivers entering from the north. Astin (1990) measured palaeocurrent directions from current ripples in sheet-flood sandstones in the Upper Stromness Flagstone Formation on the northern coast of the Mainland (north of this GCR site) and found the predominant sheet-flood drainage direction to be towards the SSW.

In a detailed study of the cyclicity in the Upper Stromness Flagstone Formation across Orkney, Astin (1990) recognized about 45 'first order' lake cycles between the Sandwick Fish Bed and the Lower Eday Sandstone Formation (which lies above the Upper Stromness Flagstone Formation). He interpreted these cycles as Milankovitch-type cycles controlled predominantly by orbital precession, causing long-term fluctuation in rainfall over about 1 million years. He also suggested that there is evidence for eccentric orbital cycles of longer timescale.

Conclusions

This site is of national importance as it is the best section in Orkney (and arguably the best anywhere) through the Caithness Flagstone Group, and its underlying metamorphic basement. The site provides a beautifully exposed section through a series of sedimentary 'cycles' that reflect the periodic oscillation between a shallow temporary lake and a desiccated lake basin. These cycles, together with the wide range of lithologies, sedimentary structures and biogenic structures that they contain, have been critical for the interpretation of sedimentary environments of the Orcadian Basin during Mid-Devonian times. The site also contains exposures of the well-known Sandwick Fish Bed, which separates the Lower and Upper Stromness Flagstone formations and is an important marker horizon in regional correlation.

TARACLIFF BAY TO NEWARK BAY, ORKNEY (HY 553 035–HY 568 043)

E.A. Pickett

Introduction

This site extends from Taracliff Bay to Newark Bay on the south-east coast of the Deerness Peninsula, in the south-eastern part of the Mainland. It exposes a long section through Middle Devonian (Middle Old Red Sandstone) strata comprising the uppermost part of the Caithness Flagstone Group and the overlying Eday Group (including the Lower Eday Sandstone, Eday Flagstone and Eday Marl formations). The section of the Eday Flagstone Formation is the thickest and best exposed in Orkney, and the site is important in containing evidence of contemporaneous volcanism. It is also particularly important for the interpretation of the depositional environments of the Eday Group and the palaeogeography of Orkney during late Mid-Devonian times, providing a comparison with the more proximal strata of the group in the Fersness Bay section on Eday (see **South Fersness Bay** GCR site report, this chapter).

Description

At the western end of the section, interbedded grey-green, calcareous siltstones and yellowish buff sandstones of the Caithness Flagstone Group are exposed in the vertical cliffs and on a wide shore platform (Figure 2.29). The sandstones are fine- to medium-grained, thin- to thick-bedded, planar and cross-laminated, and locally cariously weathered. The finer-grained, silty interbeds are rich in grey mudstone rip-up clasts and also display carious weathering. The Caithness Flagstone Group is overlain, with a transitional boundary, by the arenaceous Lower Eday Sandstone Formation. One of the best

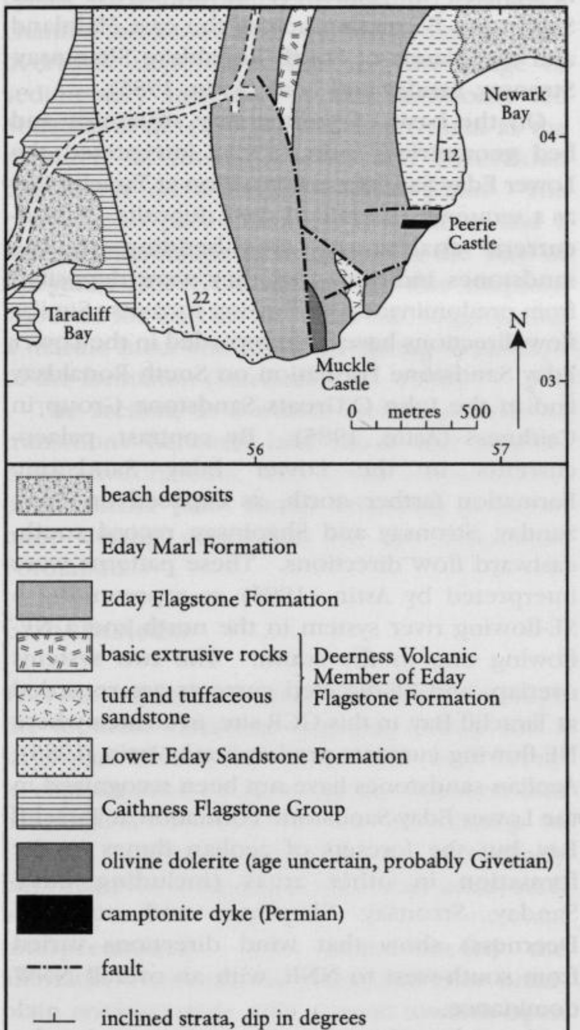


Figure 2.29 Geological map of the Taracliff Bay to Newark Bay area. Based on Kellock (1969) and British Geological Survey (1999).

sections of the Lower Eday Sandstone Formation in Orkney lies west of the sea stack known as 'Muckle Castle', where over 175 m of quartzitic, well-bedded, cross-bedded and channelized sandstones with rare, scattered pebbles can be observed. A distinctive feature of these sandstones on Deerness and elsewhere is the presence of convolute or slumped bedding. The boundary between the Lower Eday Sandstone Formation and the Eday Flagstone Formation occurs just west of Muckle Castle. Muckle Castle itself (HY 562 032) is formed almost entirely of an intrusive plug of olivine dolerite that is fault-bounded on its western side (Kellock, 1969).

The thickest development and best exposures of the Eday Flagstone Formation in Orkney are found in and around the Deerness Peninsula (included in this site), where the formation is about 150 m thick. It thins northwards to 100 m in Shapinsay, 50 m in Stronsay and 10 m in Sanday. The formation comprises up to 12 cyclic sequences, which range in thickness from 2 m to several tens of metres. Most of these cycles comprise an upward-fining phase of buff, yellow, or more rarely, red sandstone and silty sandstone, and a phase of grey, black and locally purple, flaggy siltstones and mudstones (Mykura, 1976). Fish remains are common in the muddy facies at the base of some cycles, and two beds of flaggy siltstone near Muckle Camay have yielded remains of *Tristichopterus alatus* and *Pentlandia macroptera*. Desiccation cracks and small lenticular cracks, possibly pseudomorphs after gypsum, are common in the siltstones and fine-grained sandstones. Thin sandstones within the flaggy siltstones commonly have slump structures and load-casts at their bases.

Less than 200 m east of Muckle Castle, the Eday Flagstone Formation contains about 5 m of interbedded greenish tuffs and tuffaceous sandstones with sandstone and lava pebbles (HY 563 033). These form part of the Deerness Volcanic Member, which also includes thin basalt lava flows and basaltic breccias elsewhere on the Deerness Peninsula. Analcime, natrolite (a zeolite) and alkali feldspar occur in these rocks, but Thirlwall (1979) showed that the analcime and natrolite are secondary and that the basalts are calc-alkaline.

The Eday Flagstone Formation has a faulted contact with the Eday Marl Formation near Peerie Castle, a small sea stack on the west side of Newark Bay. The Eday Marl Formation forms cliffs on the western side of Newark Bay and comprises a series of fining-upward cycles, each made up of a sandstone unit 50 cm to 2 m thick overlain by reddish calcareous mudstones and siltstones with thin beds of convoluted sandstone (Figure 2.30). Grey-green, calcareous mudstones and siltstones are also common in the formation, some carious weathering in the mudstones testifying to the presence of carbonate. The sandstones at the base of the cycles are sharp-based, resting on erosion surfaces, and are commonly medium-grained, grading normally from coarse-grained bases. Other features include rip-up clasts of red siltstone, cross-bedding and desiccation cracks. These fining-upward sequences appear to grade upwards into more massive sandstones in the sequence exposed at Newark Bay.

Interpretation

The oldest part of the section is the Caithness Flagstone Group at the western end of the site. Although divided into the Lower and Upper Stromness Flagstone formations in the west Mainland, the group is undivided in east

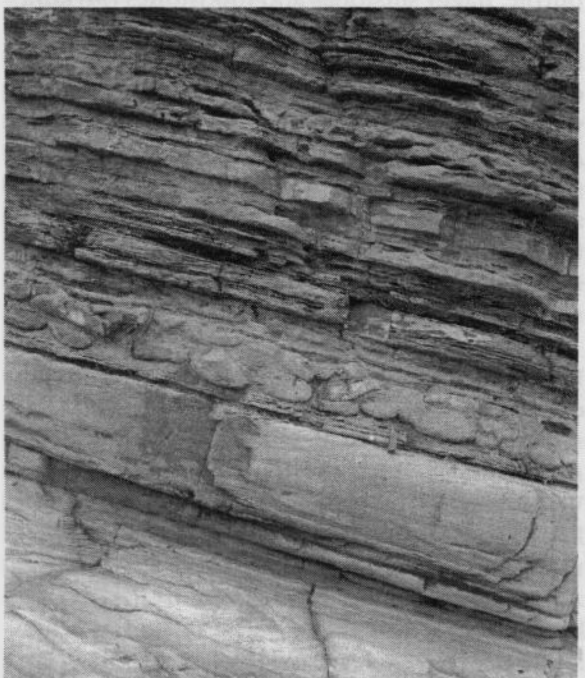


Figure 2.30 Load structures and convolute bedding in the Eday Flagstone Formation near Muckle Castle, south-west of Newark Bay. (Photo: BGS No. D1522, reproduced with the permission of the Director, British Geological Survey, © NERC.)

Mainland on the current published geological map (British Geological Survey, 1999). The Caithness Flagstone Group comprises a series of rhythmic units that record successive repeated changes in a lacustrine environment, from deep-water lake to ephemeral shallow lake with accompanying lake-beach and prograding alluvial-fan deposits. This cyclicity may have been driven by climate change (Astin, 1990).

The overlying Eday Group records the progradation of alluvial fans across the lake basin to form a regional braidplain, perhaps as a result of an episode of active extensional faulting along the basin margins (Astin, 1985). Braided river, aeolian-dune and lake-beach deposits are recognized in sequences of the Lower Eday Sandstone Formation across the east Mainland and the islands of South Ronaldsay, Shapinsay, Stronsay, Sanday and Eday (Astin, 1985).

On the basis of sedimentary structures and bed geometries, Astin (1985) interpreted the Lower Eday Sandstone Formation at Taracliff Bay as a sequence of braided river deposits. Palaeocurrent analysis of cross-bedding in the sandstones indicates that they were deposited from predominantly NE-flowing currents. Similar flow directions have been recorded in the Lower Eday Sandstone Formation on South Ronaldsay and in the John O'Groats Sandstone Group in Caithness (Astin, 1985). By contrast, palaeocurrents in the Lower Eday Sandstone Formation farther north, as exposed on Eday, Sanday, Stronsay and Shapinsay, record south-eastward flow directions. These patterns were interpreted by Astin (1985) as representing a SE-flowing river system in the north and a NE-flowing one in the south. The two systems overlap, and SE-directed currents are recorded at Taracliff Bay, in this GCR site, in an area where NE-flowing currents predominate (Astin, 1985). Aeolian sandstones have not been recognized in the Lower Eday Sandstone Formation at Taracliff Bay, but the foresets of aeolian dunes in the formation in other areas (including Eday, Sanday, Stronsay, Shapinsay and northern Deerness) show that wind directions varied from south-west to NNE, with an overall NNW dominance.

The Eday Flagstone Formation marks the re-establishment of cyclic, lacustrine sedimentation. However, in contrast to the cycles of the Caithness Flagstone Group, many of the cycles contain a buff or yellow, locally pebbly, channelized sandstone phase. The formation exhibits great lateral changes in thickness and lithology, thinning northward and south-westward away from Deerness. The thinning is due to changes in facies within the individual cycles, with the sandstone phases of some cycles becoming thicker and coarser-grained away from Deerness, whereas the lacustrine phases become thinner and eventually wedge out. Contemporaneous, sporadic volcanic activity is represented by the olivine dolerite intrusion and tuffaceous sandstones at the site, which are typical of the volcanic rocks of Mid-Devonian age found in Orkney.

The Eday Marl Formation, at the eastern end of the section, comprises fluvial, fining-upward cycles, each interpreted by Mykura (1976) as a channel sandstone unit overlain by fine-grained overbank deposits. However, the range of sedimentary structures in the sandstones led Marshall *et al.* (1996) to interpret them as the products of fluvial channel and sheet-flood deposition on a muddy sabkha plain. The presence of pseudomorphs after halite and a marine microfauna at a locality in the **Bay of Berstane** (see GCR site report, this chapter), about 12 km north-west of this site, suggests that a marine incursion occurred during deposition of the formation (Marshall *et al.*, 1996).

The section, as a whole, records a series of transitions between lake basin and alluvial-plain environments, with fluvial sedimentation on a sabkha plain finally becoming predominant during deposition of the Eday Marl Formation.

Conclusions

This site displays a well-exposed section through the topmost Caithness Flagstone Group and a large part of the Eday Group. The Eday Group is characterized by great lateral thickness and lithological variations across its outcrop in Orkney. This succession is particularly important for comparison with other areas, such as Eday, and for regional palaeoenvironmental interpretations. The strata record the alternation between lake-basin and river braid-plain environments, with a trend towards open sabkha environments. An intrusive igneous plug and related tuffaceous sandstones within the Eday Flagstone Formation are also important features, indicating contemporaneous volcanism in late Mid-Devonian times.

GREENAN NEV COAST, EDAY, ORKNEY (HY 549 367)

E.A. Pickett

Introduction

This site extends for about 700 m around the point of Greenan Nev on the west coast of Eday and provides one of the best sections through the Eday Marl Formation in Orkney. It is the only accessible section of the formation in Eday, exposing the entire thickness, and it is the type section of the formation. The site lies on the eastern limb of the Eday Syncline, the axial trace of which passes just offshore from Greenan Nev. The site is especially important for palaeo-environmental interpretations of late Mid- to early Late Devonian times as it preserves well-exposed floodplain sandstones, calcareous mudstones and siltstones with fossil carbonate soil (calcrete) horizons, a facies association not observed elsewhere in the Orcadian Basin. The presence of abundant trace fossils including *Beaconites* and *Cornulatichnus* also adds to the importance of the site by providing further information on the depositional environments, and the habitats and types of animals that lived in the Orcadian Basin during Mid- to Late Devonian times.

Description

At Greenan Nev (Figure 2.31), the Eday Marl Formation comprises several fining-upward sequences consisting of buff, clean, fine- to medium-grained sandstone beds up to 2 m thick, interbedded with up to 15 m of bright red, calcareous mudstones and sandy siltstones (Figure 2.32). The sandstone beds are finely laminated and cross-bedded, with foresets picked out by red coloration. They commonly contain reddish wisps of calcareous mudstone, fine mudstone partings and red mudstone rip-up clasts. Some sandstone beds have undulose bases with erosion of, and slight loading into, the underlying calcareous mudstones. Sandstones containing convolute bedding and trough and tabular cross-bedding occur in an outcrop of Eday Marl Formation at the **Bay of Berstane** on the Mainland (Marshall *et al.*, 1996).

The red mudstones and siltstones are micaceous and contain ripple marks, desiccation cracks, small scours, hard calcareous layers and

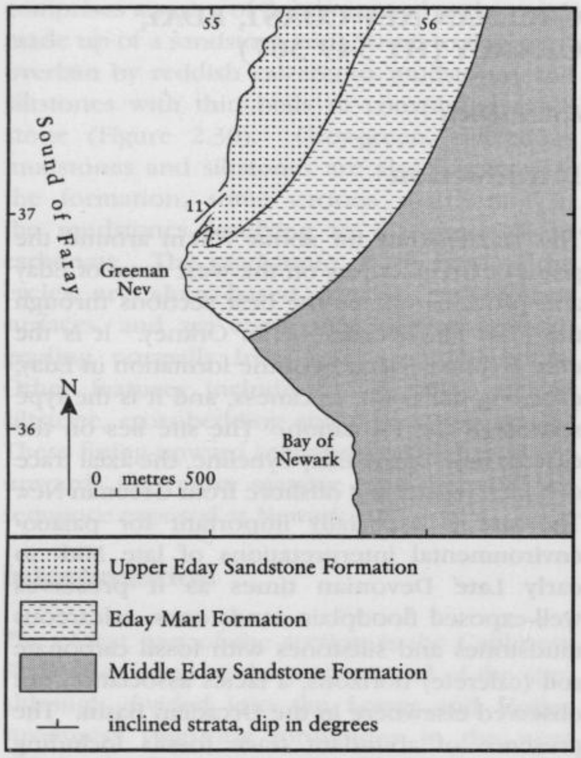


Figure 2.31 Geological sketch map of the Greenan Nev area. After British Geological Survey (1999).

abundant pale calcareous (calcrete) concretions. Bleached white reduction zones are especially common along joints and cracks. The mudstones and siltstones were bioturbated by burrowing organisms, bioturbation being most abundant in the more massive mudstones, as is seen particularly well on the gently dipping bedding planes at the base of the sea cliffs. Carroll (1991) recorded a diverse ichnofauna of 11 species within the Eday Marl Formation and the overlying Upper Eday Sandstone Formation. Carroll and Trewin (1995) described a newly recognized trace fossil *Cornulatichnus edayensis* in the Eday Marl Formation on Eday and the east Mainland, examples of which were observed at this site. The trace consists of a sub-vertical, downward-tapering conical burrow up to 40 cm long with a wide sub-circular cross-section up to 9 cm wide. Carroll and Trewin (1995) also recorded large back-filled burrows identified as *Beaconites* along with *C. edayensis* in the Eday Marl Formation.

The ratio of sandstone to mudstone increases northwards and up-sequence and there is a gradual passage into the overlying Upper Eday Sandstone Formation approximately at the northern boundary of the site.

Figure 2.32 Interbedded red marl and sandstone of the Eday Marl Formation at Greenan Nev. View towards the north-east. (Photo: E.A. Pickett.)

Interpretation

The Eday Marl Formation exposed at this site contains fining-upward cycles of sandstone overlain by calcareous mudstone and siltstone. The fining-upward cycles and the structures within the sandstones, including flat lamination and cross-lamination, as well as erosion surfaces at their bases, suggest that they represent fluvial-channel and sheet-flood (possibly channel overspill and crevasse-splay) deposition (Ridgway, 1974; Carroll, 1991; Marshall *et al.*, 1996). The cross-bedding suggests south-directed currents, similar to those determined by Astin (1985) for the Lower Eday Sandstone Formation.

The calcareous mudstones and siltstones were interpreted as floodplain overbank deposits by Mykura (1976) and as floodplain lake deposits by Ridgway (1974) and Carroll (1991). The trace fossils within the Eday Marl Formation were interpreted by Carroll and Trewin (1995) as the subaqueous shelter burrows of eel-like fish. Periodic and prolonged drying out of the floodplain is indicated by desiccation cracks and calcrete horizons. Sparite-filled vugs in the Eday Marl Formation on the Mainland have been interpreted as the moulds of evaporite nodules that formed in a sabkha-type environment (Marshall *et al.*, 1996). The marine microfauna, and pseudomorphs after halite, described by Marshall *et al.* (1996) in the Eday Marl Formation at the **Bay of Berstane** on the Mainland (see GCR site report, this chapter) suggest periodic marine inundation of the sabkhas and floodplains in mid- to late Givetian times.

Conclusions

This site is important as the type section of the Eday Marl Formation. It contains a well-exposed range of the rock types and sedimentary and biogenic structures that characterize this formation. The rocks are interpreted as floodplain sediments with associated river channel deposits, evaporitic sabkha deposits, pedogenic carbonate (calcrete) horizons and evidence of burrowing by eel-like fish. This represents an environment which is not found elsewhere in the rock record of the Orcadian Basin and is therefore of great significance in palaeoenvironmental reconstruction and the overall interpretation of the Mid- to Late Devonian evolution of the Orcadian Basin.

SOUTH FERSNESS BAY, EDAY, ORKNEY (HY 531 346–HY 543 332)

E.A. Pickett

Introduction

The South Fersness Bay GCR site extends for about 2 km along the south-west side of Fersness Bay on the west coast of Eday. It lies on the western limb of the Eday Syncline and provides a complete section through the Lower Eday Sandstone, Eday Flagstone and most of the Middle Eday Sandstone formations of the Middle Devonian Eday Group. The oldest part of the sequence is represented in the western part of the site by the Rousay Flagstone Member (forming the upper part of the Upper Stromness Flagstone Formation). The site is important because it contains an uninterrupted, well-exposed and accessible section through much of the lower part of the Eday Group. The rocks are not only representative of the stratigraphy of this part of the Orcadian Basin, but also contain a range of sedimentary structures that suggests the interaction of several depositional environments, including lake-beach, lake, braided river and aeolian dune-field. The site is therefore very important in palaeoenvironmental interpretations and for comparison with sites elsewhere in the Orcadian Basin.

Description

The rocks at the site young from west to east and form the western limb of the Eday Syncline, the axis of which runs approximately north–south through the centre of Fersness Bay. At the western end of the site, Fers Ness is composed of about 170 m of the Rousay Flagstone Member (formerly the Rousay Flags) at the top of the Upper Stromness Flagstone Formation (Figure 2.33). These comprise grey, flaggy rhythmic sequences of laminated, fish-bearing mudstone, siltstone and fine-grained sandstone. A distinctive pebbly sandstone at the top of the Rousay Flagstone Member (the Sacquoy Sandstone Member) is about 4 m thick and contains pebbles of quartzite, psammite and dolomitic limestone, with lesser amounts of quartz, granite and chert (Astin, 1990).

The Sacquoy Sandstone Member is overlain by the Lower Eday Sandstone Formation, which

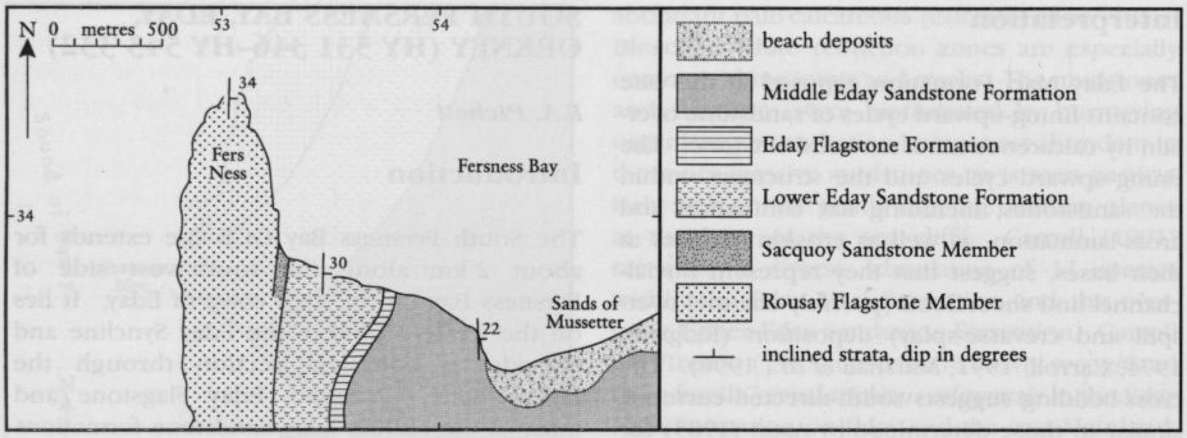


Figure 2.33 Geological sketch map of the south-west side of Fersness Bay. Based on Mykura (1976) and British Geological Survey (1999).

is about 200 m thick and consists of two main facies types (Mykura, 1976; Astin, 1985). The lower facies comprises reddish purple, medium- to coarse-grained, trough cross-bedded sandstone. Pebbly lenses and conglomeratic beds lie about 30 m above the base. The pebbles in the conglomerate are up to 7 cm across and include pink granite and pegmatite, granitic gneiss, quartzite, chert and vein quartz, with lesser amounts of sandstone (Mykura, 1976). The upper facies is a predominantly yellow, medium-grained, planar and trough cross-bedded sandstone with fewer pebbles.

The overlying Eday Flagstone Formation is relatively thin (less than 20 m) at this locality and consists of two cycles with fish beds. The cycles comprise a basal finely laminated fish-bearing facies, and overlying flaggy siltstone and mudstone with subordinate sandstone which pass up into buff, yellow or red sandstone and sandy siltstone.

A higher fish-bearing, calcareous siltstone (HY 530 335) lies about 110 m above the base of the overlying Middle Eday Sandstone Formation. The Middle Eday Sandstone Formation is an orange-yellow, pebbly and locally conglomeratic, trough cross-bedded sandstone containing scour structures, pebbly lags, mudstone rip-up clasts, convolute bedding and ripple-drift lamination (Figure 2.34). Some parts of the section contain abundant pebbles, up to 6 cm across, of quartz, porphyritic and spherulitic rhyolite, scoriaceous basic lava and granite/gneiss. The rocks locally display carious 'honeycomb' weathering and diagenetic red and yellow mottling.

Interpretation

The rhythmic units or 'cycles' that form the Rousay Flagstone Member of the Upper Stromness Flagstone Formation are interpreted as reflecting fluctuations in water-level and sediment input into a large, shallow lake that extended across Orkney and Caithness in Mid-Devonian times (e.g. Fannin, 1970; Mykura, 1976; Astin, 1990). The development of the Orcadian Basin was probably tectonically controlled, with the rhythmic sedimentation being caused by the interplay of tectonic and climatic changes. The dark, calcareous, fish-bearing siltstone/mudstone laminites at the base of each cycle represent deposition from suspension when the lake was at its deepest. Progressive shallowing culminated in mudflats that periodically dried up. Alluvial fans prograded across the shallow or dried-up lake floor producing fine-grained, rippled sandstones and siltstones. These fluvial sediments were deposited as sheet-flood and channel sands and show sediment transport to the south and south-west (Astin, 1990). The base of the following cycle is marked by an abrupt reversion to the deeper-water fish-bearing facies.

The Rousay Flagstone Member passes up into the Sacquoy Sandstone Member, which has distinctly different palaeocurrent directions, having been deposited from eastward-flowing rivers. This marks the first incursion of alluvial-fan sediments prograding from the north-westerly basin margin, probably sourced from metamorphic rocks or older Devonian conglomerates lying to the west and exposed as

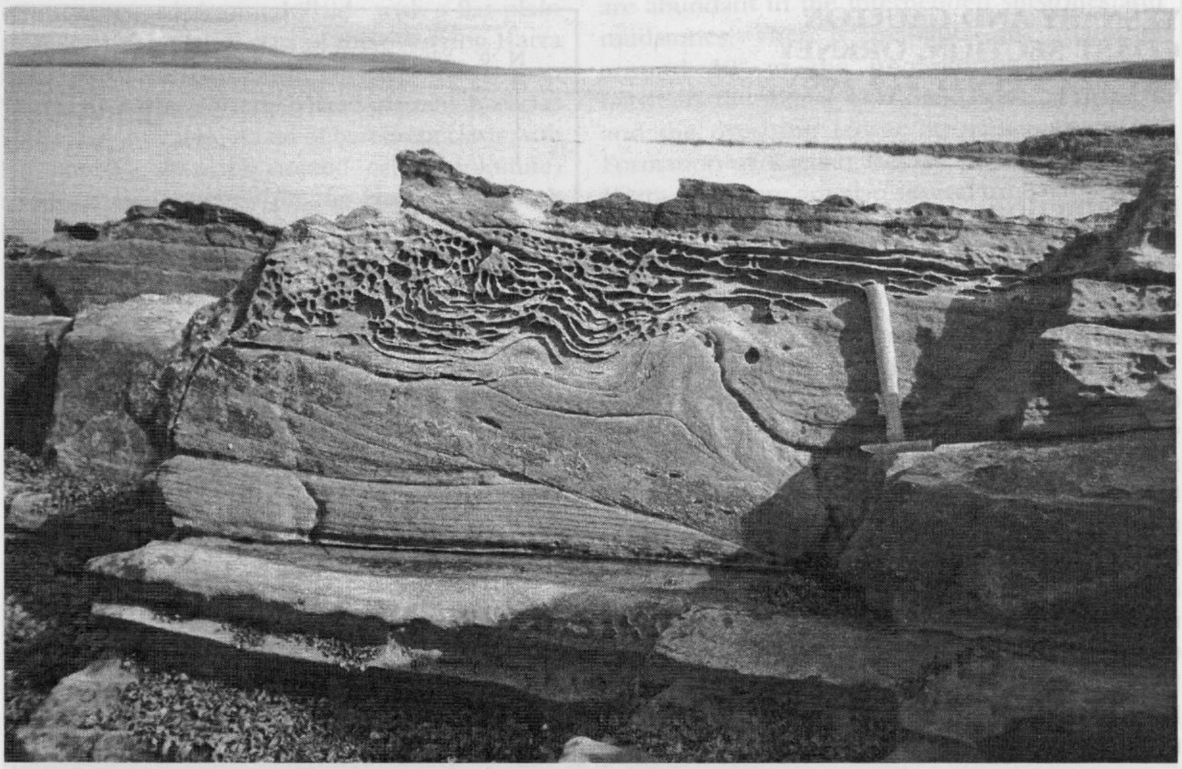


Figure 2.34 Middle Eday Sandstone Formation at Fersness Bay showing cross-bedding, convolute bedding and calcareous weathering. View towards the ENE. (Photo: E.A. Pickett.)

a result of footwall uplift on a basin-margin fault (Astin, 1990). The lower facies of the overlying Lower Eday Sandstone Formation was probably deposited in the channels of large, fast-flowing rivers in a fan system prograding south-eastwards. The yellow sandstones of the upper facies include large-scale cross-bedded dune sets up to 3.5 m thick and are interpreted as aeolian sands (Astin, 1985). In western Eday (e.g. HY 535 336) these aeolian sandstones form part of a distinct unit up to 70 m thick, interbedded with fluvial sandstones.

The Eday Flagstone Formation represents a break from the predominantly fluvial regimes of the Lower and Middle Eday Sandstone formations and a return to cyclic, lacustrine sedimentation. However, in contrast to the Rousay Flagstone Member, the cycles contain thick channel-fill sequences of yellow and red sandstones. As with the Lower Eday Sandstone Formation, palaeocurrents indicate flow towards the south-east. The thick channel-fill sandstones indicate that the lake was fed by larger and higher-energy rivers than those present during deposition of the Rousay Flagstone Member. The overlying Middle Eday Sandstone Formation is entirely fluvial, and this, together with SE-directed palaeocurrents, shows that a fan system similar to that of the Lower Eday Sandstone Formation prograded over the whole area.

Conclusions

The site at Fersness Bay is very important in the reconstruction of environments existing in the Orcadian Basin during Mid-Devonian times. The site provides a complete section through the Lower Eday Sandstone, Eday Flagstone and most of the Middle Eday Sandstone formations. The sequences record changes in the environment of deposition within the Orcadian Basin from lakes to mudflats to aeolian dune-fields and alluvial fans. Against this background of overall shallowing, smaller-scale fluctuations and periodic returns to deeper water are recorded in the cycles of the Rousay Flagstone Member and the Eday Flagstone Formation. These probably reflect the interplay between climatic changes, basin subsidence and sedimentation rates and tectonic activity along basin-bounding faults.

YESNABY AND GAULTON COAST SECTION, ORKNEY (HY 222 144–HY 224 166)

E.A. Pickett

Introduction

This impressive and extensive site extends from south of Neban Point to Row Head, comprising over 5 km of the western coast of the Mainland. The site contains superb cliff sections through the Lower Devonian Yesnaby Sandstone Group, which does not occur anywhere else in Orkney, and the Middle Devonian Lower Stromness Flagstone Formation. The Yesnaby Sandstone Group comprises two formations, the older Harra Ebb Sandstone Formation and the Qui Ayre Sandstone Formation (formerly known as the Yesnaby Sandstone Formation; Fannin, 1970; Clarke, 1990; Trewin and Thirlwall, 2002). These units are separated by the Garthna Geo Fault and both overlie crystalline basement. Importantly, the Qui Ayre Sandstone Formation includes aeolian sandstones (Fannin, 1970; Trewin and Thirlwall, 2002), a facies not found anywhere else in the Lower Devonian succession of the Orcadian Basin. Other aeolian sediments in the Lower Old Red Sandstone are known only in the Smerwick Group in the Dingle Peninsula, south-west Ireland (Richmond and Williams, 2000). The Yesnaby Sandstone Group is unconformably overlain by the Lower Stromness Flagstone Formation. This comprises a series of rhythmic fluvio-lacustrine units, representing a wide range of lake sediments, and includes the best stromatolite beds in the Orcadian Basin. Also exposed within the site is the Sandwick Fish Bed, which lies at the base of the overlying Upper Stromness Flagstone Formation and is an important marker horizon in regional correlation.

Description

The shore on the south side of the prominent inlet of Garthna Geo, which has been eroded along a fault, exposes extensive outcrops of the metamorphic basement (Figure 2.35). The basement includes large areas of siliceous schist, and, just north of Harra Ebb, some hornblende- and biotite-schist. The schist and gneiss are intruded by veins and sheets of pre-Devonian granite. The pre-Devonian land surface was a

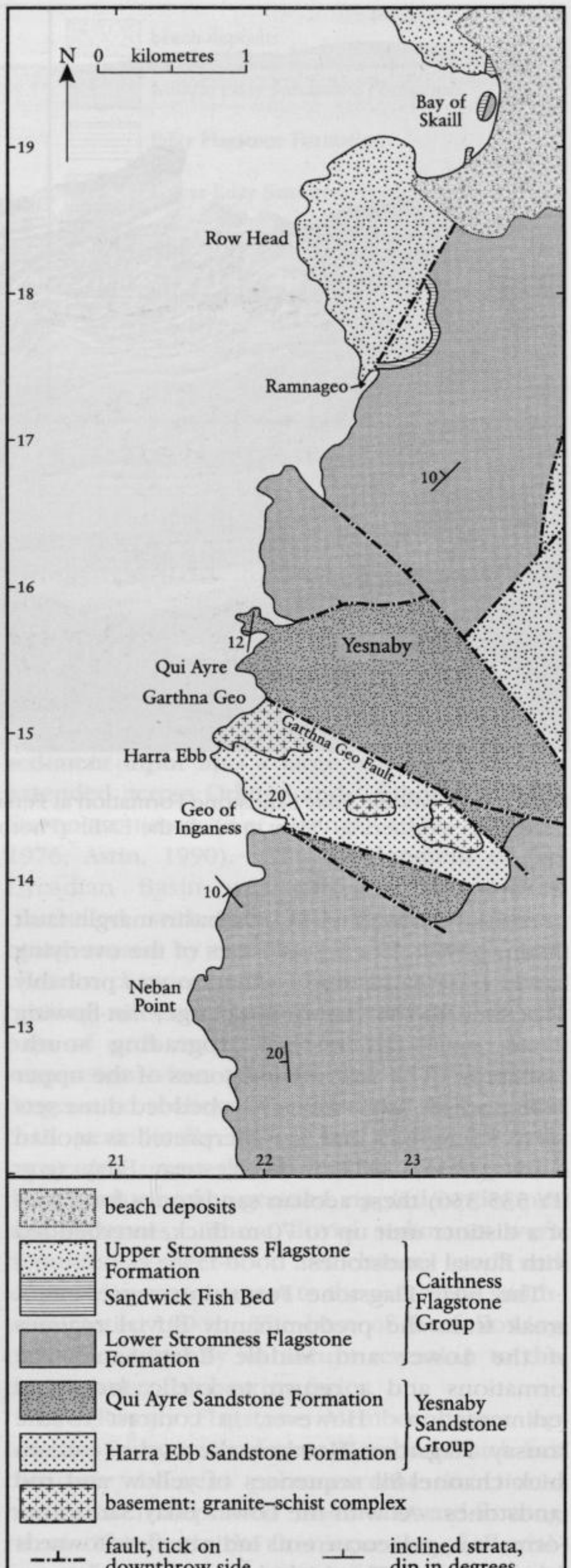


Figure 2.35 Geological map of the Yesnaby coast, west Mainland. Based on Fannin (1970), Mykura (1976), Clarke (1990) and British Geological Survey (1999).

steep westward-sloping hillside with a flat plain at its foot. The basal part of the overlying Harra Ebb Sandstone Formation consists of less than 50 m of poorly sorted, coarse-grained breccias and conglomerates. A bed of basement clasts with laminated carbonate (mixed calcite-dolomite) coatings rests directly on the basement, which has fractures up to 50 cm wide and 6 m deep, filled with similar laminated carbonate. Cavities within the carbonate fracture-fillings contain quartz, barite and fluid hydrocarbon. The breccias and conglomerates are overlain by up to 100 m of interbedded sandstones, dolomitic mudstones and red and green siltstones, with a few thin tabular beds of conglomerate. Trough cross-bedding and ripple cross-lamination are common in the sandstones; desiccation cracks are abundant in the interbedded siltstones and mudstones. There is an angular discordance of around 10° at the unconformable contact between the Harra Ebb Sandstone Formation and the overlying Lower Stromness Flagstone Formation at Kaellan Hellier (HY 219 145). At Harra Ebb there are also several unusual vertical breccia masses, one of which forms a spectacular stack composed of angular to subrounded sandstone boulders in a matrix of comminuted carbonate-rich sand.

On the north side of the Garthna Geo Fault, the rocks of the Yesnaby Sandstone Group, formerly named the 'Yesnaby Sandstone Formation' by Fannin (1970), Mykura (1976) and Clarke (1990), are now known as the 'Qui Ayre Sandstone Formation' (Figure 2.36).

Figure 2.36 Cliffs of large-scale cross-bedded sandstones of the Qui Ayre Sandstone Formation at Yesnaby. In the background is Garthna Geo with exposed basement. View towards the south. (Photo: BGS No. D1545, reproduced with the permission of the Director, British Geological Survey, © NERC.)

Although the Garthna Geo Fault separates the major outcrops of the Harra Ebb and Qui Ayre Sandstone formations, Clarke (1990) noted that the Qui Ayre Sandstone Formation rests directly and conformably on the Harra Ebb Sandstone Formation, with channels filled by the former incised into the latter (HY 2179 1482; HY 2181 1488). The Qui Ayre Sandstone Formation can be subdivided into two members, the lower of which comprises at least 30 m of rusty-weathering, grey, fine- to medium-grained, well-sorted sandstones with well-rounded grains. The sequence displays large-scale cross-stratification in sets up to 3 m, with steep foresets and planar truncation surfaces. The truncation surfaces are spaced about 5–10 m apart.

The well-sorted, cross-bedded sandstones pass up into the upper member, which is best exposed in Old Millstone Quarry (HY 217 155) at the point of Qui Ayre. It consists of at least 25 m of massive, planar-bedded and ripple-bedded sandstones, which are locally trough cross-bedded and ripple cross-laminated with pebble trains. Siltstone interbeds have desiccation cracks. The sandstone is locally impregnated with bitumen, and black, bituminous sandstone can be observed on the cliff top near the sea stack known as the 'Castle of Qui Ayre'. At Old Millstone Quarry, the Qui Ayre Sandstone Formation is unconformably overlain by the Lower Stromness Flagstone Formation with an angular discordance of 10°. The unconformity surface is draped by a thin, matrix-supported conglomerate in which the clasts are subangular to subrounded and generally up to 10 cm by 6 cm. They include metaquartzite, siliceous mica schist, sparry dolomite and vein quartz, all set in a matrix of sandstone cemented by ferroan dolomite. Rarer clast types include banded gneiss, pegmatite and chert. The conglomerate is overlain by 1–2 m of ripple cross-laminated sandstone, followed by laminated, calcareous beds.

The Yesnaby Sandstone Group may contain lateral equivalents of the 61 m-thick Warebeth Red Bed Formation, which were encountered between basement and the Lower Stromness Flagstone Formation in a borehole at Warebeth, near Stromness (see **South Stromness Coast Section** GCR site report, this chapter).

The Lower Stromness Flagstone Formation is well exposed along the cliff tops northwards towards the Hill of Borwick and Ramnageo. It comprises a blue-grey, flaggy, rhythmic sequence in which laminated mudstones pass up into siltstones and fine-grained sandstones, forming a series of coarsening-upward sequences 1–8 m thick. The section includes stromatolites known locally as 'Horse Tooth Stone' (HY 220 161), of which there are particularly fine examples on the cliff top at Yesnaby, just to the north of the road end. A prominent bedding plane is covered with stromatolites, which locally form elliptical mounds with the heads inclined inwards. It is intruded by a camptonite dyke, and there are minor traces of galena mineralization at the contact. On the cliff top between Borwick and Ramnageo, 0.5 m of interbedded calcitic and dolomitic mudstones contain dark, laminated chert nodules that are concentrated in the calcitic layers and form polygonal arrays.

The top of the Lower Stromness Flagstone Formation is placed at the base of the Sandwick Fish Bed. This laminated mudstone is exposed in a small quarry at the head of Ramnageo, a spectacular, long inlet with parallel vertical walls, one of which is a fault plane. Shiny, black fragments of fossil fish are abundant on the exposed bedding planes. The quarry yielded a great quantity of fossil fish remains in the 1840s and 1850s. One of the best-known localities for the Sandwick Fish Bed is Cruaday Quarry, about 4 km north-east of the north end of the GCR site. The quarry has been a source of fossil fish specimens since the 19th century and is described in the *Fossil Fishes of Great Britain* GCR volume (Dineley and Metcalf, 1999). The fauna includes *Coccosteus cuspidatus* and *Osteolepis macrolepidotus* (both common as whole specimens) and *Mesacanthus* sp. (Dineley, 1999a). Detailed studies at this site have allowed a close correlation with the Niandt Limestone Member at Achanarras, Caithness (Trewin, 1976). An exposure of the Sandwick Fish Bed at the Bay of Skaill, about 2 km north-east of the north end of this GCR site, has yielded some of the best examples of Mid-Devonian plant fossils in Britain and is described in the GCR volume on *Palaeozoic Palaeobotany of Great Britain* (Cleal and Thomas, 1995). The flora includes the holotype of the oldest known progymnosperm *Protopteridium thomsonii*.

The overlying Upper Stromness Flagstone Formation comprises a sequence of rhythmic units similar to that of the Lower Stromness Flagstone Formation.

Interpretation

The sequence at Yesnaby is interpreted as recording a changing sequence of fluvial, lacustrine and aeolian environments which characterized the Orcadian Basin in Early to Mid-Devonian times. The aeolian facies is very important in being one of only two occurrences recognized in the Lower Old Red Sandstone of Britain and Ireland. Trewin and Thirlwall (2002) suggest that a drier climate prevailed during its deposition. The basal breccias of the Harra Ebb Sandstone Formation may represent talus deposits which accumulated on the slopes of the basement palaeotopography. Fannin (1969) interpreted the coated basement clasts as algal oncolites, although there seems to be a gradation from these structures to the fracture-fillings in the basement and the pebble coatings may therefore be travertine drapes. The overlying sandstones, siltstones, mudstones and dolomicrites of the Harra Ebb Sandstone Formation are interpreted as representing alluvial-fan and sandflat deposits that prograded over a playa-lake system and its associated mudflats (Fannin, 1970; Clarke, 1990). Deposition was predominantly by sheet-flood, with some channelized flow, palaeocurrent directions suggesting south-westerly progradation of the alluvial-fan system (Fannin, 1970). The laminated carbonates probably had a biogenic origin as algal or microbial mats in the playa lake. The vertical breccia bodies have been interpreted as breccia pipes (cryptovents) produced by gas fluxion associated with Permian igneous activity (Mykura, 1976).

Clarke (1990) interpreted the basal part of the Qui Ayre Sandstone Formation as the deposits of a sandy braided river system that flowed south-westwards over the playa mudflats of the Harra Ebb Sandstone Formation. This was followed by a period of emergence and aeolian dune-field migration, when the predominant wind direction was towards the east or ESE (Fannin, 1970). Clarke (1990) interpreted the overlying facies as the result of renewed fluvial deposition by ephemeral streams and overbank sheet-floods, followed by development of sandy, low-sinuosity braided rivers flowing to the south-west and south-east. A different interpretation was presented by Fannin (1970) who tentatively suggested that the facies overlying the aeolian dunes represent a northward-encroaching, shallow, clastic shoreline that inundated and reworked the upper part of the dune deposits. He based this interpretation on analogy with modern marine shoreline environments, but pointed out that there is no evidence to show whether the advancing water body was fresh or marine.

Fannin (1970) distinguished three sources for the pebbles lying on the unconformity surface at the top of the Qui Ayre Sandstone Formation – the adjacent crystalline basement, the underlying Yesnaby Sandstone Group and an exotic source that lay outside the area and supplied a range of metamorphic and metasedimentary clasts.

The Caithness Flagstone Group, comprising the Lower and Upper Stromness Flagstone formations, marks a change to rhythmic, coarsening-upward lacustrine and alluvial cycles. Each cycle records a successive change in environment from deep-water lake to ephemeral shallow lake, lake-beach and alluvial fan. This cycle was repeated many times and may have been driven by a combination of cyclic climate change and tectonic activity (see **South Stromness Coast Section** GCR site report, this chapter, for a more detailed interpretation of the 'cycles' within the Caithness Flagstone Group).

Conclusions

This site is important in exposing a thick succession of Old Red Sandstone strata, including the Harra Ebb and Qui Ayre Sandstone formations of probable Early Devonian age, the oldest sedimentary rocks exposed in the Orcadian Basin. These represent a combination of alluvial-fan, aeolian dune and lake-beach deposits and are of great importance in the interpretation of sedimentary environments in the early development of the Orcadian Basin. The aeolian sandstones are particularly important in the uniqueness of their occurrence and its implications for early Devonian climate and the palaeogeography of Britain and Ireland. The well-exposed basement complex and its mantle of breccias and conglomerates also allow reconstruction of part of the pre-Devonian landscape. The Caithness Flagstone Group records the establishment of a lake environment and a cyclic pattern of sedimentation in which the lake deepened rapidly and shallowed slowly, with deep lake muds at the base of each cycle passing up into coarser-grained sediments containing evidence of shallowing and emergence. Rivers formed alluvial fans that built out over the dry lake-bed before becoming inundated as the cycle started again.

OLD MAN OF HOY COAST, ORKNEY (HY 211 052–HY 174 991)

E.A. Pickett

Introduction

This site includes about 8 km of the spectacular coastline of north-western Hoy. It extends from Rora Head to Hamar Hellia and includes the 137 m-high Old Man of Hoy sea stack and the 350 m-high cliffs at St John's Head (Figure 2.37). The cliff sections, which are largely inaccessible from inland, show the basal volcanic part (Hoy Volcanic Member) of the mainly Upper Devonian Hoy Sandstone Formation resting unconformably on the Middle Devonian Upper Stromness Flagstone Formation. The Hoy Volcanic Member has been dated as 379 ± 10 Ma (Halliday *et al.*, 1982) which places it close to the Mid–Late Devonian boundary and of probable Givetian age. The site is important because the Hoy Volcanic Member, seen only on Hoy, allows comparison with other Middle Devonian volcanic rocks of the Orcadian Basin. It is described in the GCR volume on the Caledonian igneous rocks (Stephenson *et al.*, 1999). Above the lavas lies a thick succession of sandstones of the Hoy Sandstone Formation, forming the impressive red sea-cliffs that dominate this site. These cliffs, where accessible, show excellent examples of the braided stream and aeolian facies that are typical of the Late Devonian of the Orcadian Basin. Sedimentary structures within the aeolian sandstones suggest wind directions different from those that prevailed in Early Devonian times (on the evidence of the unique aeolian sandstones at the **Yesnaby and Gaulton Coast Section** (see GCR site report, this chapter) and Mid-Devonian times (see **South Fersness Bay** GCR site report, this chapter). This site is therefore of great importance in the interpretation of the palaeoenvironments and palaeogeography of the Orcadian Basin.

Description

The oldest part of the exposed sequence is the cross-bedded, faulted and gently folded Upper Stromness Flagstone Formation, which is exposed at the base of the Old Man of Hoy sea stack (HY 176 008) and from the Geo of Hellia (HY 190 042) northwards (Figure 2.37). A particularly good section of the rhythmites of the formation is at the Bay of the Tongue (HY 207 047), near the northern end of the site. The cliffs at Kame of Hoy and the Bay of the

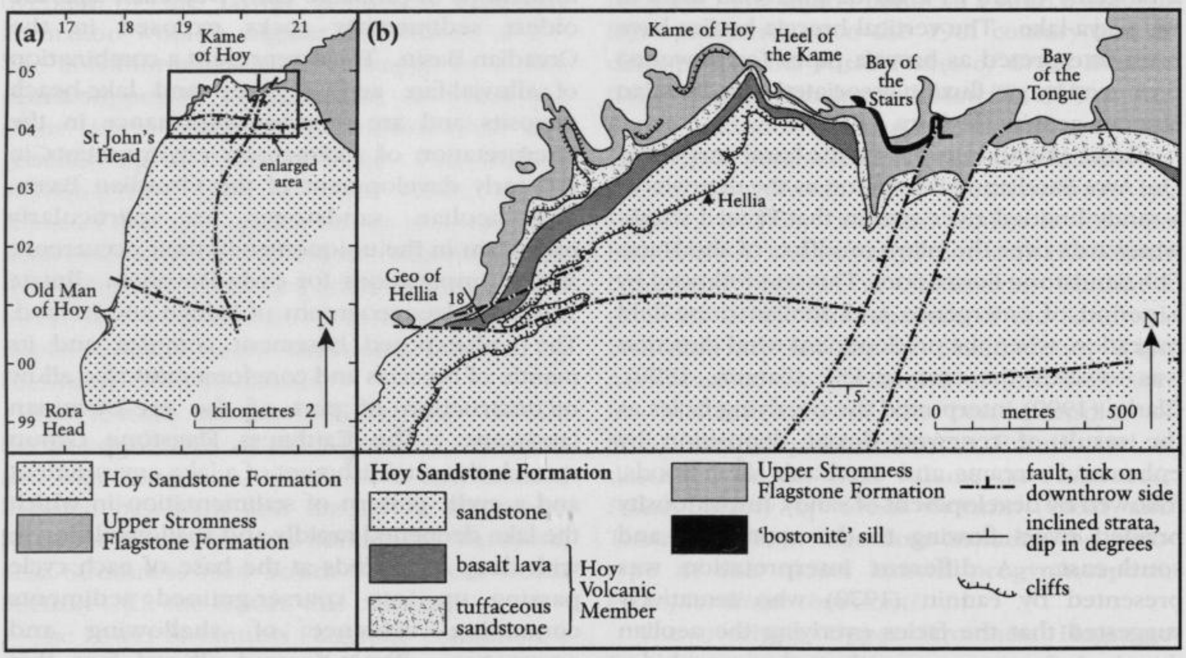


Figure 2.37 (a) Simplified geological map of north-west Hoy (after British Geological Survey, 1999). (b) Detailed geological map of the northern part of the GCR site (after Mykura, 1976).

Stairs consist mainly of cyclic sequences of mudstones, siltstones and sandstones of the formation. At Yelting Geo, about 300 m north-east of the eastern end of the site, Fannin (1970) noted thin (up to 4 cm) beds of tuffaceous siltstone in the formation.

The Upper Stromness Flagstone Formation is truncated by an undulating surface and unconformably overlain by the Hoy Volcanic Member of the Hoy Sandstone Formation, the unconformity being well exposed at the base of the Old Man of Hoy. The basal part of the Hoy Volcanic Member consists of tuffaceous sandstones and breccio-conglomerates. Between the Bay of the Tongue and the Geo of Hellia the tuffaceous sandstones are up to 15 m thick, brownish red, locally cross-bedded and contain angular blocks and pebbles of basalt and finely comminuted basaltic material (Mykura, 1976). At the Geo of Hellia and the Old Man of Hoy, these basal tuffaceous rocks are absent and the Upper Stromness Flagstone Formation is directly over-lain by lavas.

The lavas of the Hoy Volcanic Member form five disconnected outcrops in the north-west of Hoy and one on the south coast of the island. McAlpine (1978) recognized five separate lava flows, although Mykura (1976) thought it impossible to determine whether the outcrops represent one or several flows. The lavas are alkali olivine basalts with porphyritic crystals of olivine and feldspar in a groundmass of iron oxides, augite and plagioclase (Wilson *et al.*, 1935). They are nepheline-normative (Thirlwall, 1979, 1981) and have been dated at 379 ± 10 Ma (Halliday *et al.*, 1982). A maximum exposed thickness of about 90 m is seen at Hellia, where the lavas form part of the impressive cliffs. They comprise a grey-weathering, vesicular lower part, a massive 60 m-thick columnar-jointed central portion and an upper, 15 m-thick, purplish slaggy zone (Mykura, 1976). At the Old Man of Hoy, the lavas are 3 m to over 7 m thick, and the lateral variations in the lavas can be seen particularly well at their base. Their upper surface at the Old Man of Hoy is a fairly even plane that appears to have been eroded before deposition of the overlying sandstones (Wilson *et al.*, 1935).

Above the Hoy Volcanic Member, the Hoy Sandstone Formation comprises over 1000 m of medium- to thick-bedded, medium- to coarse-grained, red and yellow sandstones with thin interbedded siltstone partings (Figure 2.38). The sandstones display planar and trough cross-bedding with individual cross-sets up to 1.2 m in height. Slumping and convolute bedding are locally common. Many cross-sets contain clasts of red and purple siltstone, and intraformational conglomerates at the bases of troughs are common. There are also rare small lenses of extraformational conglomerate containing pebbles of quartz, schist and gneiss.

McAlpine (1978) subdivided the sandstones into four members – the Lang Geo Sandstone, Lyre Geo Sandstone, Haist Pebbly Sandstone and Trowie Glen Sandstone. The Hoy Sandstone Formation in the northern part of the site is undivided, whereas the Lang Geo Sandstone Member and the overlying Lyre Geo Sandstone Member have been recognized in the southern part (British Geological Survey, 1999). The Lang Geo Sandstone Member, which forms the cliffs at the Old Man of Hoy, is distinguished from the Lyre Geo Sandstone Member mainly by its thinner-bedded and better-cemented sand-stones. Both members contain fining-upward sandstone units and subordinate siltstones and mudstones.

A further feature of interest at the site is a microsyenitic ('bostonite') sill of uncertain age which cuts the Upper Stromness Flagstone Formation at the base of the cliff at Bay of the Stairs (HY 203 047). The sill consists predomi-nantly of feldspar (mainly orthoclase) and some chlorite and carbonate (Mykura, 1976).

McAlpine (1978) correlated in detail the Hoy Sandstone Formation with the Dunnet Head Sandstone Group at Dunnet Head on the Caithness mainland, which can be seen from Hoy. The cliffs at Dunnet Head provide specta-cular exposure of laterally continuous fluvial (braided river) sandstones (Trewin, 1993).

Interpretation

The coarsening-upward rhythmic cycles in the Upper Stromness Flagstone Formation at the base of the succession record changes in environment from relatively deep-water to ephemeral, shallow lake and prograding alluvial-fan. This cycle was repeated many times and may have been driven by a combination of cyclic climate change and tectonic activity (see **South Stromness Coast Section** GCR site report, this chapter). Thin bands of tuffaceous siltstone within the Upper Stromness Flagstone Formation near the eastern end of the site may represent

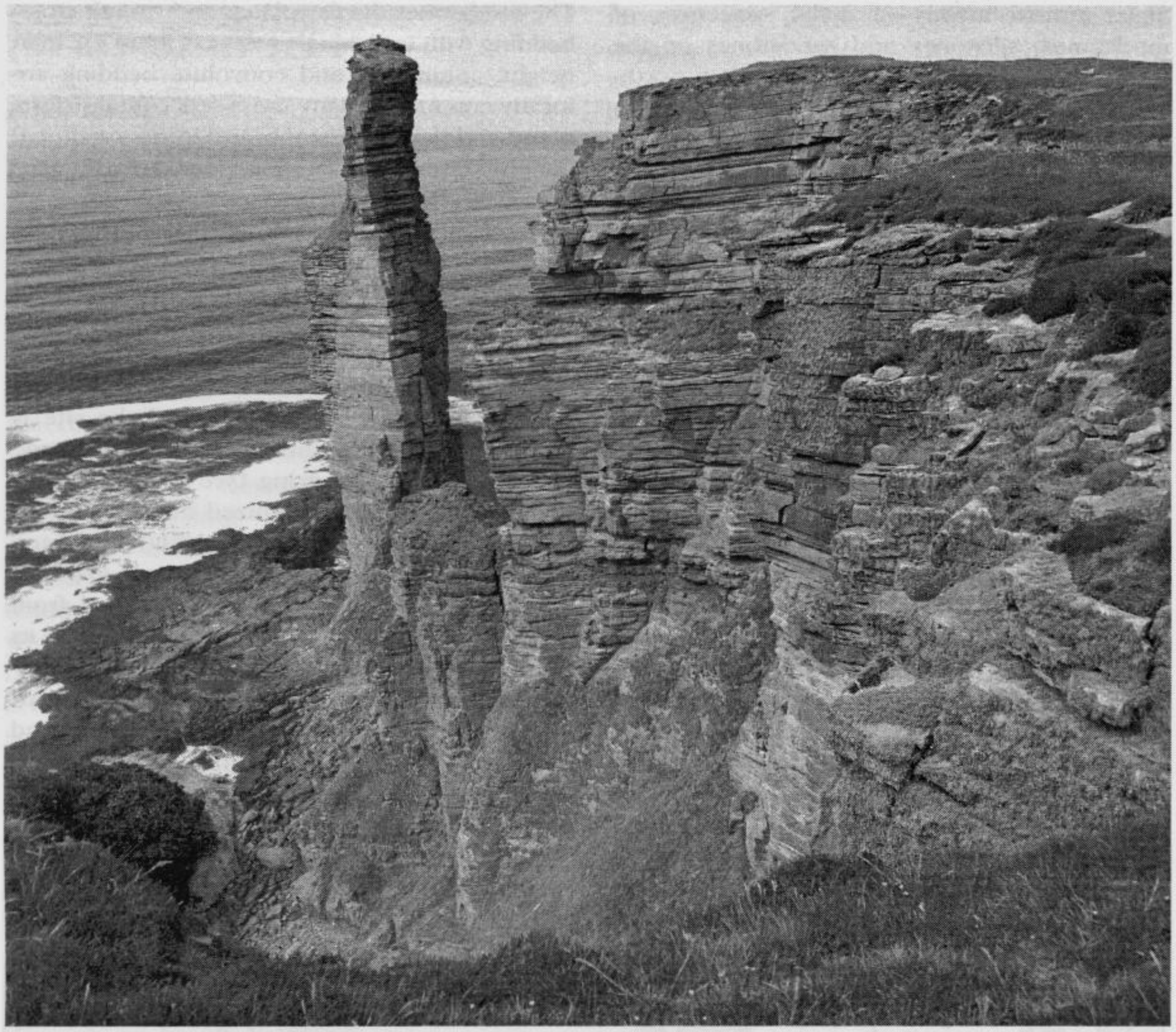


Figure 2.38 The Old Man of Hoy, a sea stack of Hoy Sandstone Formation lying just off the spectacular cliffs of north-west Hoy. The basal member, the Hoy Volcanic Member, lies at the base of the stack where it overlies the Upper Stromness Flagstone Formation. (Photo: BGS No. D1539, reproduced with the permission of the Director, British Geological Survey, © NERC.)

the start of Mid-Devonian volcanic activity that culminated in the extrusion of the volcanic rocks at Hoy and those in the Eday Flagstone Formation at Deerness on the Mainland (Fannin, 1970). The tuffaceous layers thin and wedge out northwards, suggesting that they emanated as small ash eruptions from a volcanic centre that lay farther south (Fannin, 1970).

The basal tuffaceous sandstones and conglomerates of the Hoy Volcanic Member were deposited on an undulating erosion surface. Mykura (1976) suggested that most of the volcaniclastic rocks were deposited as subaerial ash falls, since their bedding planes are parallel to the slopes of the underlying hummocky basement surface. However, McAlpine (1978) interpreted the tuffaceous beds as shallow-water deposits. Extrusion of basaltic lavas followed, probably as several flows. The planar upper surface of the lavas suggests that some erosion occurred before deposition of the overlying sandstones.

The Hoy Sandstone Formation records a period when over 1000 m of sands and subordinate silts were deposited in a predominantly fluvial, probably braided stream environment (Mykura, 1976). The great thickness of the formation has important implications for the subsidence history of the Orcadian Basin in Late Devonian times. The common fining-upward fluvial sequences are interbedded with sabkha and playa-lake siltstones and poorly cemented aeolian sandstones. McAlpine (1978) suggested a system of active wadi fans in which sediments were deposited as wadi-flood deposits. Beyond the distal margins of the wadi fans, sand accumulated as small barchan dunes while sand and silt accumulated in playa lakes and sabkhas in the interdune areas. Palaeocurrent analysis of the fluvial sandstones indicates north-east flow, and the aeolian sandstones appear to have formed by reworking of the fluvial sands by winds that blew predominantly towards the south-west (McAlpine, 1978).

Conclusions

The island of Hoy is the only locality in the Orcadian Basin where the Hoy Sandstone Formation and its basal member, the Hoy Volcanic Member, are exposed. For this reason Hoy, and more specifically this GCR site, are of great significance in the regional interpretation of the Devonian Orcadian Basin. Along a coastline of spectacular and commonly inaccessible cliffs, this large site exposes a series of important stratigraphical contacts and fine sections through the succession. At the base of the Old Man of Hoy, fluvial sandstones of the Upper Stromness Flagstone Formation are unconformably overlain by tuffaceous sandstones and basaltic lava of the Hoy Volcanic Member. Although a significant period of erosion probably occurred between deposition of the sandstones and eruption of the lavas, volcanic activity had already started during deposition of the Upper Stromness Flagstone Formation, as shown by the presence of thin tuffaceous sandstones. Volcanic activity ceased by Late Devonian times, when the Orcadian Basin was a wide plain with braided rivers, wadis, lakes, sabkhas and small dune-fields. The sediments laid down in these environments are preserved in the cliffs at this site. Sedimentary structures show that rivers flowed towards the north-east and dune-forming winds blew towards the south-west. These directions contrast with data from Lower Devonian and Middle Devonian rocks of the Orcadian Basin, thus making this site important for interpreting the palaeogeographical and palaeoenvironmental evolution of the basin through the Devonian Period.

BAY OF BERSTANE, ORKNEY (HY 476 111)

Potential GCR site

E.A. Pickett

Introduction

This potential GCR site lies in the north-west corner of the Bay of Berstane near Kirkwall on the Orkney Mainland. A short section of the Eday Marl Formation, including a mudstone containing a marine microfauna and pseudomorphs after halite, is exposed in a small, fault-bounded block. The mudstone is interpreted as representing a short marine incursion into the Orcadian Basin during mid–late Givetian (Mid-Devonian) times (Marshall *et al.*, 1996). If this interpretation is correct, this is the only record of a Devonian marine incursion onshore in Scotland, and the site is therefore of crucial importance for palaeoenvironmental and palaeogeographical interpretations of the Orcadian Basin. The site also provides the first onshore evidence to confirm the results of a BP/Chevron offshore well, drilled in the East Orkney Basin, which proved three intervals with marine microfossils in upper Givetian to lower Frasnian rocks (Marshall *et al.*, 1996).

Description

The following account is based wholly on a description by Marshall *et al.* (1996). The site (Figure 2.39) exposes about 7 m of red and green, calcareous mudstones and argillaceous siltstones and sandstones, punctuated by cleaner red and buff sandstones up to about 50 cm thick. The most important feature is a 1 m-thick bed of grey-green, parallel-laminated, mudstone and siltstone, with red mudstone layers spaced at about 10 cm intervals (Figure 2.40), which contains evidence suggesting a brief marine incursion into the Orcadian Basin. Many of the siltstone laminae are trains of starved wave ripples, the troughs of which contain common hopper-crystal pseudomorphs after halite up to 2 cm across. Only the corners of many of the larger pseudomorphs are present, or they are truncated below the overlying siltstone. The smaller pseudomorphs are generally complete. The lamination in the red mudstone layers is disrupted and compressed into tepee structures.

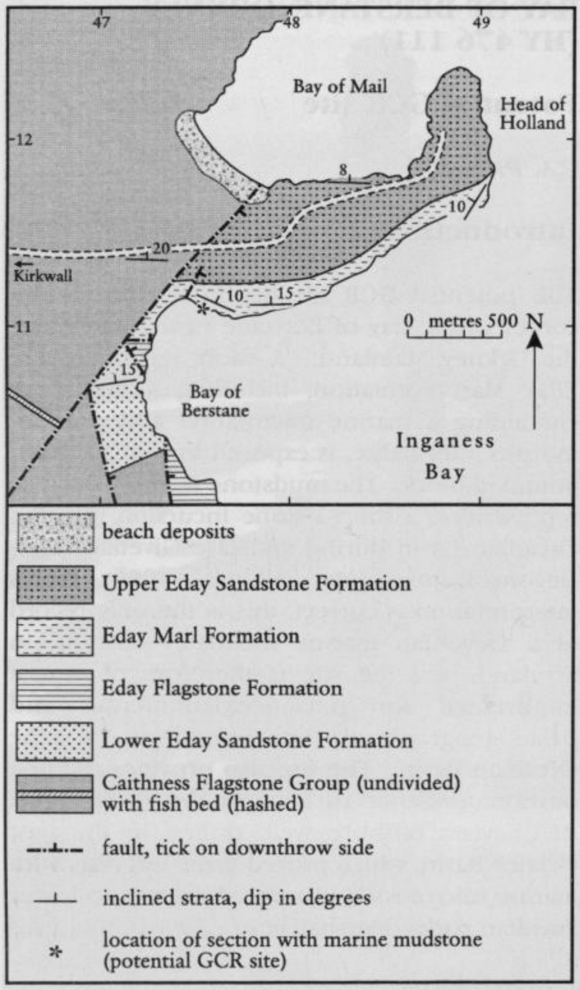


Figure 2.39 Geological map of the Bay of Berstane area. After British Geological Survey (1999).

The red mudstones above and below the grey-green unit are mostly apparently structureless and disrupted by successive polygonal arrays of desiccation cracks, or have a 'turbate', locally brecciated structure defined by contorted lenses and laminae of cleaner or coarser-grained sand. A diamict texture is present locally. Other less abundant structures include wave-rippled surfaces, deflation lags of coarse sand or granules (some with ripple form), gutter casts and water-escape structures. Traces of cross-lamination and flat lamination are preserved locally within the 'turbate' layers, and there are a few burrow-mottled horizons with tube-like, U-shaped, *Thalassinoides* and *?Beaconites* burrows. Common calcite nodules are either small calcrete concretions or more commonly sparite-filled vugs, giving the mudstones a vesicular appearance. Some are filled with dog-tooth spar and others have geopetal fillings.

Samples of the grey-green mudstone contain a low-diversity palynomorph assemblage dominated by *Geminospora lemurata*, regarded as mid- to late Givetian in age. However, the main importance of this site lies in the discovery by Marshall *et al.* (1996) of two scolecodonts in the palynomorph assemblage. These are the chitinous jaw parts of polychaete worms, which are today restricted to marine environments, although all the sedimentological evidence at the site points to a non-marine origin.

This site is also important in providing an onshore link with offshore data obtained from the BP/Chevron offshore well 14/6-1, drilled in 1985 at the south-east margin of the East Orkney Basin (about 100 km ESE of Orkney). The well penetrated Devonian sedimentary rocks that are dated palynologically as late Givetian to early Frasnian in age (Marshall *et al.*, 1996), and three grey-green mudstone intervals, yielded marine microfossils, including an acritarch in the upper one and scolecodonts in the lower two.

Interpretation

The red beds of the Eday Marl Formation are interpreted by Marshall *et al.* (1996) as the deposits of a muddy sabkha plain, the 'turbate' structures being caused by growth of evaporite crusts and nodules, deposition and erosion around them by wind and water, repeated desiccation, remobilization of saturated sediment and bioturbation by burrowing organisms. The sparite-filled vugs are probably the moulds of dissolved evaporite nodules.

The presence of scolecodonts and pseudomorphs after halite in the grey-green mudstone in the Eday Marl Formation at the Bay of Berstane has been attributed to marine inundation of the sabkha plain (Marshall *et al.*, 1996). All living polychaete worms are found in marine environments and are accepted as indicators of similar environments in the geological past. Acritarchs would be expected to predominate over scolecodonts in a normal marine setting, and the dominance of scolecodonts in offshore well 14/6-1, as well as their restriction to simple forms such as *Micrhystridium* and *Veryhachium*, may be due to a 'stressed', marginal marine environment (Marshall *et al.*, 1996). This is supported by the fact that polychaete worms are common in the modern

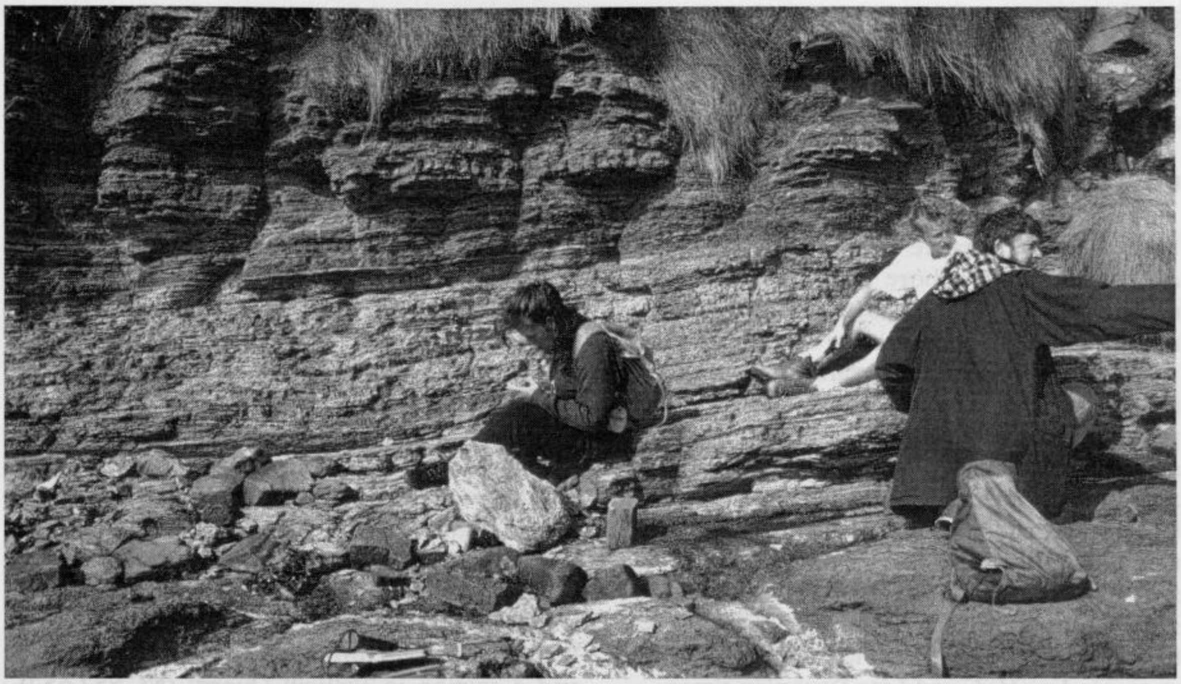


Figure 2.40 Grey-green mudstone unit within the Eday Marl Formation at the Bay of Berstane. (Photo: J.E.A. Marshall.)

nearshore environment, where they can tolerate wide salinity variations.

The pseudomorphs after halite are also cited by Marshall *et al.* (1996) as important evidence for a marine incursion, although they also occur commonly in playa-lake deposits. Although evaporite pseudomorphs are widespread in the Orcadian Basin (e.g. Astin and Rogers, 1991) and the evaporitic crusts have been interpreted as causing the disrupted bedding and tepee structures in the Bay of Berstane section, the pseudomorphs dominantly replace gypsum, indicating the presence of sulphate-rich brines. The pseudomorphs after halite are thought to indicate a short episode when NaCl-rich brines flooded the basin, and their large size and high concentration suggest that marine waters were the source of the brines. The absence of desiccation cracks suggests that any periods of emergence were short. Marshall *et al.* (1996) envisaged an extensive, low-relief sabkha plain that was episodically flooded by marine waters, analogous to the Ranns of Kutch today, which is flooded during the monsoon season and is a dry, evaporitic sabka plain for the remainder of the year.

The evidence of the marine incursions in BP/Chevron Well 14/6-1 and in the Bay of Berstane section has led to a major revision of late Mid- and Late Devonian palaeogeography of the Orcadian Basin and North Sea. The discovery of marine Devonian rocks in the Argyll Field, farther south in the North Sea (Pennington, 1975) necessitated revision of the palaeogeographical model, incorporating a connection with the marine Devonian rocks of southern Britain and northern Europe along a narrow, probably fault-bounded zone. The discovery by Marshall *et al.* (1996) of the miospore *Archaeoperisaccus* in the marine beds in well 14/6-1, a genus restricted to the northern part of Laurasia to the north, indicates that the marine waters did not enter the Orcadian Basin through a southern connection. Marshall *et al.* (1996) suggest that the connection was from the east, along the Tornquist Zone, around the southern margin of the Fenno–Scandian High. When sea levels dropped, the connection was severed and continental sedimentation resumed in the Orcadian Basin.

The marine incursions represented at the Bay of Berstane and in well 14/6-1 were tentatively attributed by Marshall *et al.* (1996) and Marshall and Hewett (2003) to global sea-level rises, the Berstane occurrence being correlated with the Givetian Taghanic highstand of the 'standard' Devonian sea-level curve, as defined in New York

State (House and Kirchgasser, 1993). Marshall *et al.* (1996) further speculated that the sea level was controlled by two orders of orbital cyclicity, at periodicities of 39.5 ky and 413.9 ky.

Conclusions

This site's importance lies in providing what may be the first onshore record of a Devonian marine flooding event in Scotland. The marine interpretation is contentious, but if proved to be correct, there are important implications for the study of the palaeoenvironments and palaeogeography of the Orcadian Basin. A short section of the Eday Marl Formation contains a grey-green mudstone that has yielded marine microfossils and pseudomorphs of halite crystals. These, together with evidence from an offshore well drilled in the East Orkney Basin, may provide evidence that the sabkha plains of the Orcadian Basin were periodically inundated by marine waters in late Mid- and Late Devonian times.

RED POINT, CAITHNESS (NC 932 659)

P. Stone and W.J. Barclay

Introduction

The cliff sections around Red Point provide remarkable and unique examples of Middle Devonian lake margin deposits unconformably resting on granodiorite and gneiss basement. The lithofacies relationships are important in establishing fluctuations in playa-lake conditions and their interaction with the variable topography of the basin margin. Particularly noteworthy are the lake margin carbonate and coarse breccia deposits, and their lateral transition into braid-plain sandstones and deeper-water, lacustrine, laminated mudstones. A comprehensive description of the site is given by Donovan (1975), summarized by Trewin and Thirlwall (2002). Field guides are provided by Donovan (1978), Parnell *et al.* (1990) and Trewin (1993).

The regional importance of the Red Point site lies in the exceptionally well-preserved examples of marginal lacustrine deposits. The lithofacies are unique in their diversity and the clarity of their relationships with the basement rocks. The site, along with similar exposures to the west at Port Skerra and Baligill, allows interpretation and characterization of an otherwise unrepresented part of the stratigraphy of the Orcadian Basin, and apart from its intrinsic sedimentological value, it is important for the broader interpretation of the tectonics and palaeogeography of the basin.

The relationships between irregular basement topography and lowermost Old Red Sandstone conglomerates and sandstones are also well displayed at Port Skerra (HC 878 663) (Trewin, 1993). Here, cliff exposures show locally intense fracturing in the basement that was exploited and opened by Devonian weathering. In the bay, sheet-flood conglomerates and hard silica-cemented sandstones fill hollows, drape knolls and infill a 3 m-deep fissure in the irregular gneiss surface. Sedimentary structures include excellent convolute lamination in a sandstone exposed near low-tide mark and sand volcanoes (B.P.J. Williams, pers. comm.). West of Port Skerra, outcrops at Balligill (NC 855 659) and a section in Balligill Quarry (NC 852 657) provide excellent exposures of deeper lacustrine and lake-margin facies, the latter including a limestone mantling a basement hill (NC 857 661) (Trewin, 1993). The lacustrine laminites are fish-bearing at two horizons in the quarry. Fine examples of load structures are seen in the bases of fine-grained turbidite sandstones, as well as 'flow-roll' de-watering structures, diagnostic of lacustrine deposition. Trewin (1993) provided details of these excellent localities.

The dangers of the Red Point site should be stressed. The sea cliffs around Red Point are high and precipitous with much loose rock, and most of the critical exposures are perched at the top of the cliffs in precarious positions. Visitors must therefore take appropriate precautions to ensure their safety.

Description

Red Point lies about 18 km west of Thurso on the north coast of Caithness. It provides a remarkable section through marginal lithofacies of an Orcadian Basin lake, which rest unconformably on rusty brown, granodioritic and locally gneissose basement rocks. The basement relief is at least 30 m (Donovan, 1975), with the overlying sedimentary strata dipping variably away from the elevated basement areas. Local minor faulting complicates the unconformable relationship. The outline geology of the site and a sketch cross-section illustrating the relief are shown in Figures 2.41

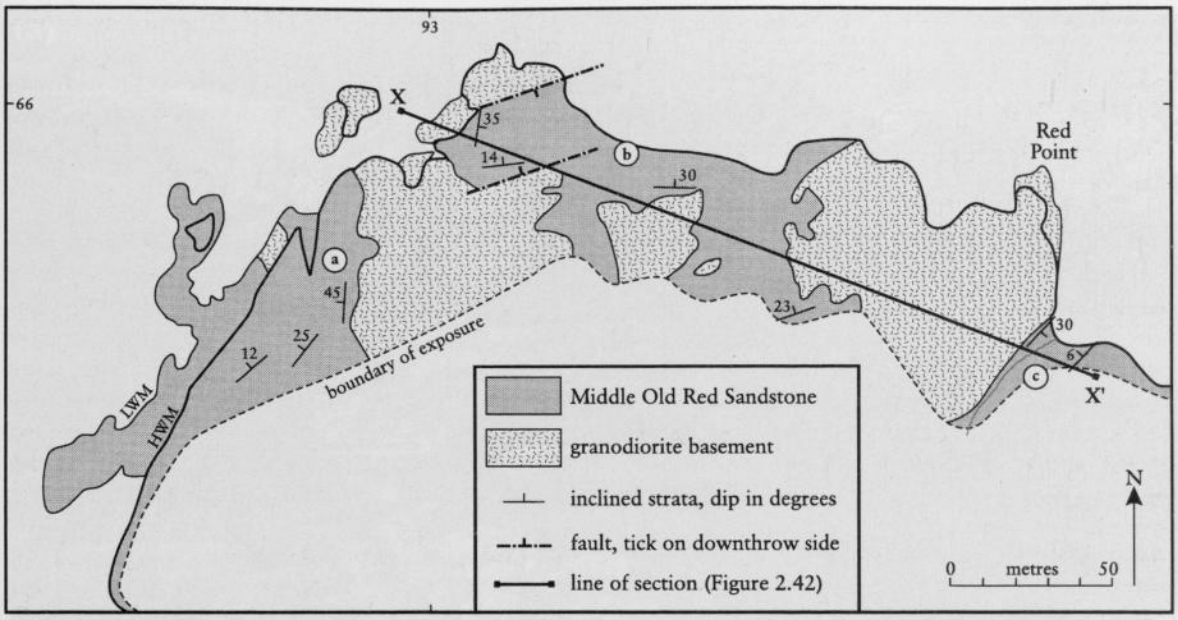


Figure 2.41 Locality map of the Red Point area. Based on Donovan (1975) and Trewin (1993).

and 2.42 respectively. There is no direct evidence for the age of the basin-margin strata, but they probably belong to the lower part of the Latheron Subgroup, the lowest division of the Upper Caithness Flagstone Group, and are therefore likely to be early Givetian (Mid-Devonian) in age. The deposition of the Red Point sediments was probably coincident with the maximum expansion of the Orcadian Basin lake.

Along the western side of the basement inlier ('a' on Figure 2.41) a remarkable limestone facies rests directly on the basement. The lowest limestone is a thin, calcite-cemented breccia of basement clasts, which is particularly concentrated in hollows in the erosion surface. Above this, up to 2 m of grey, fairly massive, limestone shows evidence for slumping, is irregularly silicified locally, and appears to merge laterally down-slope into laminated mudstone. Janaway and Parnell (1989) noted lower dolomitic and upper calcitic units and small laminated domes of possible algal origin. A coarse breccia of limestone and basement clasts in a sandy matrix overlies the limestone (Figure 2.43). Clasts are angular to subrounded and range up to about 40 cm across.

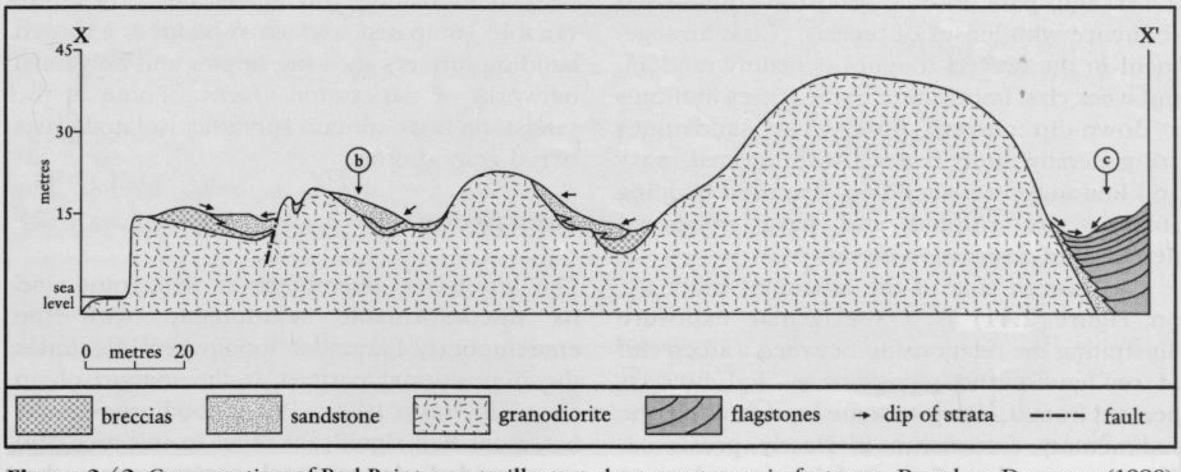


Figure 2.42 Cross-section of Red Point area to illustrate basement margin features. Based on Donovan (1980) and Trewin (1993).

Figure 2.43 Coarse breccia of basement and limestone clasts in a sandy matrix, overlying massive limestone, Red Point. (Photo: P. Stone.)

Across the central part of the basement inlier ('b' on Figure 2.41), breccias lie directly on the unconformity, although minor faulting affects the outcrop pattern. The breccias are cemented by calcite locally, and some beds are draped by very thin, carbonate-rich mudstone laminae. Northwards, and dipping away from the elevated basement area, the breccias are increasingly interbedded with coarse sandstone. Breccia tongues thin fairly abruptly down-dip (Figure 2.44) and pass into a sandstone-dominated sequence with lenses of breccia. Clast arrangement in the breccia tongues is mainly random, but weak clast imbrication in the lenses indicates a down-dip current flow. The sandstones are generally thinly bedded with internal, wavy and low-angle cross-bedding. Exposed bedding surfaces are rippled and have polygonal networks of desiccation cracks.

The eastern side of the basement inlier ('c' on Figure 2.41) is a spectacular exposure illustrating the relationship between a steep cliff of red gneiss and granodiorite, and breccias derived from it. Despite some local faulting, the sedimentary association is clearly preserved. The coarse breccia contains angular to subrounded clasts, mainly of the adjacent granodiorite, but there are also some clasts of limestone similar to that on the western side of the basement inlier. Locally, the breccia fills fissures in the steep basement cliff, and overall reaches a maximum thickness of about 10 m. The beds dip steeply away from the basement cliff, towards the east, but the dip decreases markedly in that direction and is sub-horizontal 50 m away, the decrease coincident with the breccias merging laterally with thinly interbedded sandstone and mudstone. These are variably laminated and cross-bedded, exposed bedding surfaces showing ripples and polygonal networks of desiccation cracks. Some of the sandstone beds contain sporadic, isolated clasts of red granodiorite.

Interpretation

The lithofacies assemblage at Red Point, and its unconformable relationship with the contemporary basement topography, illustrates the depositional pattern at the margin of an Orcadian Basin lake. The granodiorite–gneiss basement had significant relief in this area and so provided depositional environments that varied in response to fluctuating lake levels. The

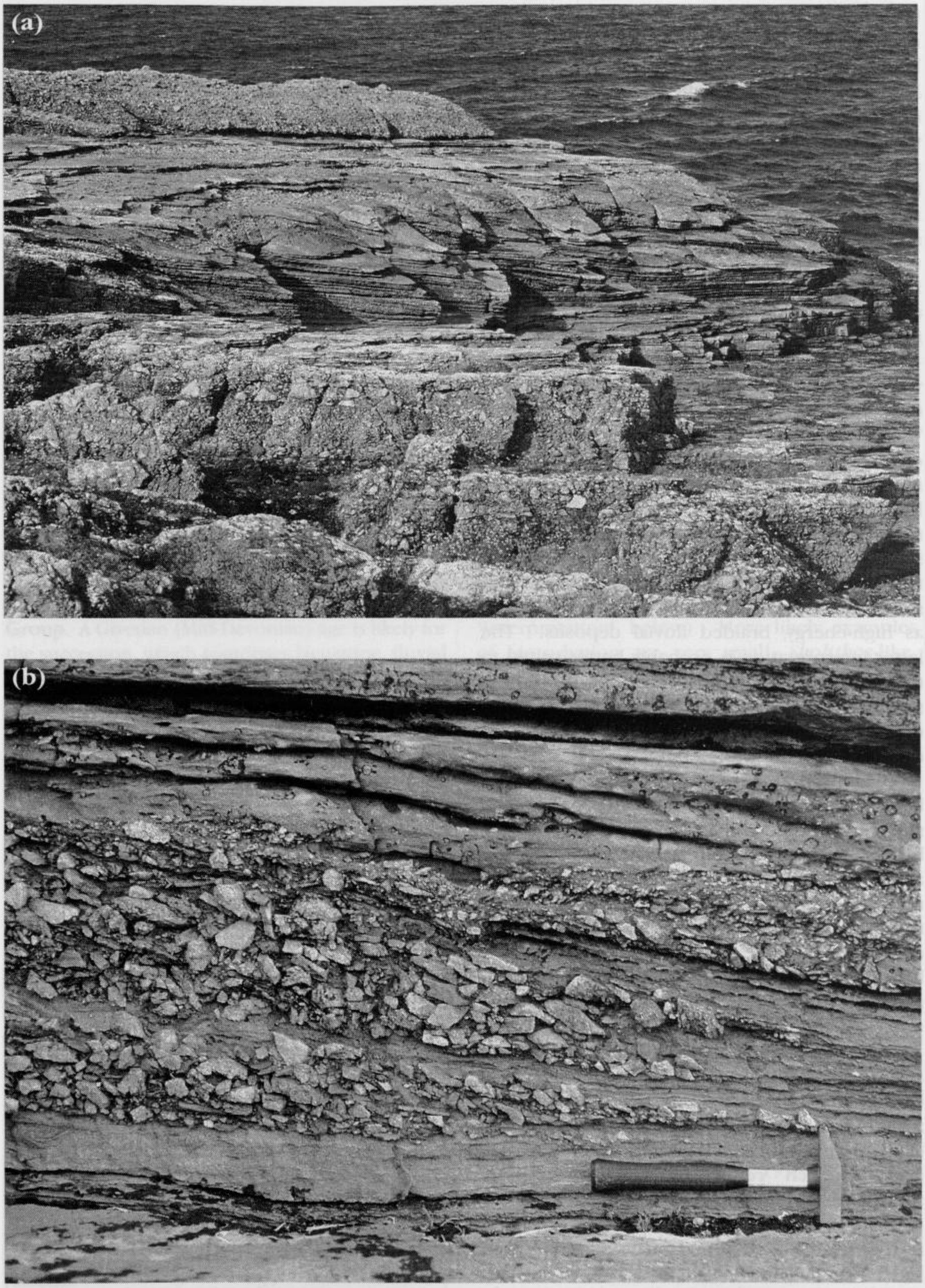


Figure 2.44 Breccia tongues interfingering with sandstone. (a) General view; (b) detail. (Photos: P. Stone.)

limestone to the west of the basement inlier represents lacustrine deposition at maximum lake water-level (a coincident lake margin of Donovan, 1975), when the water-level was high enough to transgress the margin of the basin. Carbonate deposition was controlled by photosynthesizing algae, with the upward transition from dolomitic to calcitic limestone probably reflecting reduced water salinity (Janaway and Parnell, 1989). Down-slope slumping of the limestone occurred prior to its complete lithification; in the same direction, it grades into the deeper-water mudstone–sandstone laminite facies.

The coarse breccias overlying the limestone were lake-margin beach deposits that prograded into the lake in response to a drop in water-level. Their angular limestone pebbles were clearly derived from fully lithified limestone. Again, there is a lateral transition, away from the elevated basement topography, as breccia tongues interfinger with thinly bedded sandstones. The sandstones may be a slightly deeper-water, nearshore lithofacies, but their sedimentary structures are equally compatible with an origin as high-energy, braided fluvial deposits. The latter interpretation is probably better suited to the current-imbricated breccia lenses within the sandstones. Wave-rippled bedding planes cut by desiccation polygons point to intermittent emergence of the lake floor.

The most dramatic example of basement relief is seen on the eastern side of the inlier. The steepness of the basement cliff against which the breccias are banked, and the abrupt transition away from the cliff into laminated sandstone and mudstone, suggests a scree-like accumulation at the lake margin. The laminated sequence contains many examples of wave-rippled bedding surfaces and polygonal desiccation crack networks, indicating alternating shallow-water and emergent conditions.

The range of marginal lacustrine lithofacies around Red Point is remarkable, but it is very hard to link the different exposures into a single, comprehensive model. Instead, it seems most likely that a range of stratigraphical levels is present, each reflecting slightly different conditions at different times in the evolution of the lake margin. The lacustrine transgressions and regressions recorded may possibly have had a climatic control. The local, minor faulting that cuts the site may also be significant in juxtaposing different parts of the sequence.

Conclusions

The Red Point GCR site exposes a unique assemblage of marginal lacustrine deposits and shows their relationships. It preserves a fragment of an Orcadian Basin lake margin, probably at the time of deposition of the earliest strata of the Upper Caithness Flagstone Group farther out in the lake and at the time of maximum lake extent in the basin. The lacustrine strata rest unconformably on a basement of red granodiorite with a visible topographic relief exceeding 30 m. Exceptional features of the site are the shallow-water limestones and coarse breccia tongues of possible beach origin. These pass laterally into high-energy, probably fluvial sandstones with clast-imbricated breccia lenses, and lacustrine mudstone–sandstone laminites. The latter display wave-generated ripples and polygonal desiccation crack networks, a combination that is strongly suggestive of a playa-lake environment. Overall, the Red Point site is critically important to an understanding of the palaeogeography of the Orcadian Basin and provides a unique insight into its lake margin environments.

PENNYLAND (THURSO–SCRABSTER), CAITHNESS (ND 102 696–ND 115 685)

P. Stone

Introduction

The cliff and foreshore section between Thurso and Scrabster provides excellent exposures of part of the Middle Devonian Upper Caithness Flagstone Group. The lithofacies present were produced during the cyclical deepening and shallowing of a large lake occupying part of the Orcadian Basin; about 20 cycles are represented, with abrupt transitions from laminated mudstones deposited in a deep lake, to cross-bedded sandstones deposited in an ephemeral, playa-lake environment. A remarkable range of shrinkage cracks have either a subaqueous or subaerial origin, and their correct interpretation is crucial for an understanding of lake history within the Orcadian Basin. Fossil fish faunas from the laminated mudstones are of great importance and have led to the site being separately selected for the GCR for its fossil fishes (Dineley and Metcalf, 1999). Of the 20 laminite horizons, 16 yield fish remains, most of them small fragments, but 7 or 8 laminites contain complete plates or complete fishes. Field guides to the site are provided by Donovan (1978), Parnell *et al.* (1990) and Trewin (1993).

The regional importance of the Pennyland (Thurso–Scrabster) site lies in the exceptionally well-preserved evidence for lacustrine depositional cyclicity, allowing interpretation and characterization of an otherwise poorly exposed part of the Orcadian Basin succession. The site is also important in the broader interpretation of the tectonics, lithostratigraphy and palaeogeography of the Orcadian Basin. The conservation value and importance of the site also rests in the good quality of its fossil fish remains.

Description

Along the south-west side of Thurso Bay, between Thurso and Scrabster, an extensive coastal section affords excellent exposure through part of the Mey Subgroup of the Upper Caithness Flagstone Group. The highest beds seen are probably transitional into the overlying John o'Groats Sandstone Group. A Givetian (Mid-Devonian) age is likely for the succession, which comprises lacustrine, fluvial and aeolian facies. Gentle, consistent north-west dips result in almost 200 m of strata being exposed along about 1.5 km of shoreline (Figure 2.45), but continuity is repeatedly disrupted by minor faulting. Exposure is in the cliffs and on a wide intertidal rock platform.

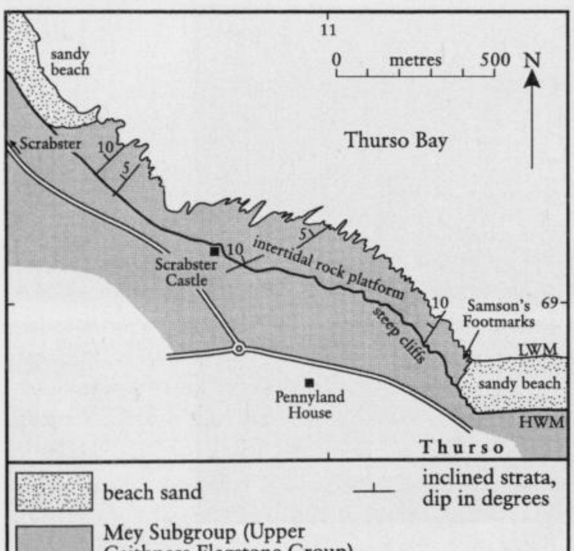


Figure 2.45 Locality map of Pennyland GCR site.

The oldest strata within the GCR site, at the eastern end of the section, are thinly bedded, laminated grey-green shales and paler, fine-grained, locally lenticular sandstones. They are affected by a range of styles of shrinkage cracks, all well exposed on bedding planes on the intertidal platform. Polygonal arrays (Figure 2.46a) are particularly well-developed (probably giving rise to the local name for this area – 'Samson's Footmarks') and grade into more orthogonal patterns of more widely spaced cracks (Figure 2.46b). Lenticular cracks tend to be smaller and to form complex, interlocking patterns (Figure 2.46c). Rippled surfaces commonly alternate with those affected by shrinkage cracks and some of them are also cracked. The sandstone layers commonly extend downwards into the small shrinkage cracks that penetrate the underlying shale or mudstone to create the so-called 'fang structures' of Donovan (1980). Differential compaction strongly deformed many of the crack infills so that in cross-section they are bulbous or sinuous (Figure 2.47). Originally mistaken for burrow-fills (Crampton and Carruthers, 1914), their origin remains a subject of controversy (see 'Interpretation', below). More likely examples of bioturbation are very small, *Skolithos*-like, circular structures less than 1 mm across on some bedding surfaces.

Finely interlaminated organic-rich mudstone and carbonate beds, some containing scattered, disarticulated fish remains, occur sporadically within the sandstone–mudstone laminites. There are 20 such units within the Thurso–Scrabster section, 16 of which contain fish. However, most of these units are disrupted by shrinkage cracks and contain only small fish fragments, and only seven or eight thicker calcareous laminites with no cracks contain complete plates or complete fishes (Dineley, 1999a). Complete specimens of *Dipterus* are fairly common and *Millerosteus minor* is very well-preserved as disarticulated and semi-articulated plates in the lower part of the section (ND 113 688–ND 110 692). Orange- or brown-weathering stromatolites also occur in the organic, fish-bearing strata, mainly as small, millimetre-scale hemispheroids (Parnell *et al.*, 1990) and as sheets (Trewin, 1993). The fish-bearing laminites occur on average at about 7 m intervals through the lower (eastern) part of the section (Parnell *et al.*, 1990). This cyclicity is superimposed on a general, but irregular, trend of upward-increasing sandstone proportion. Many of the sandstone beds have soft-sediment

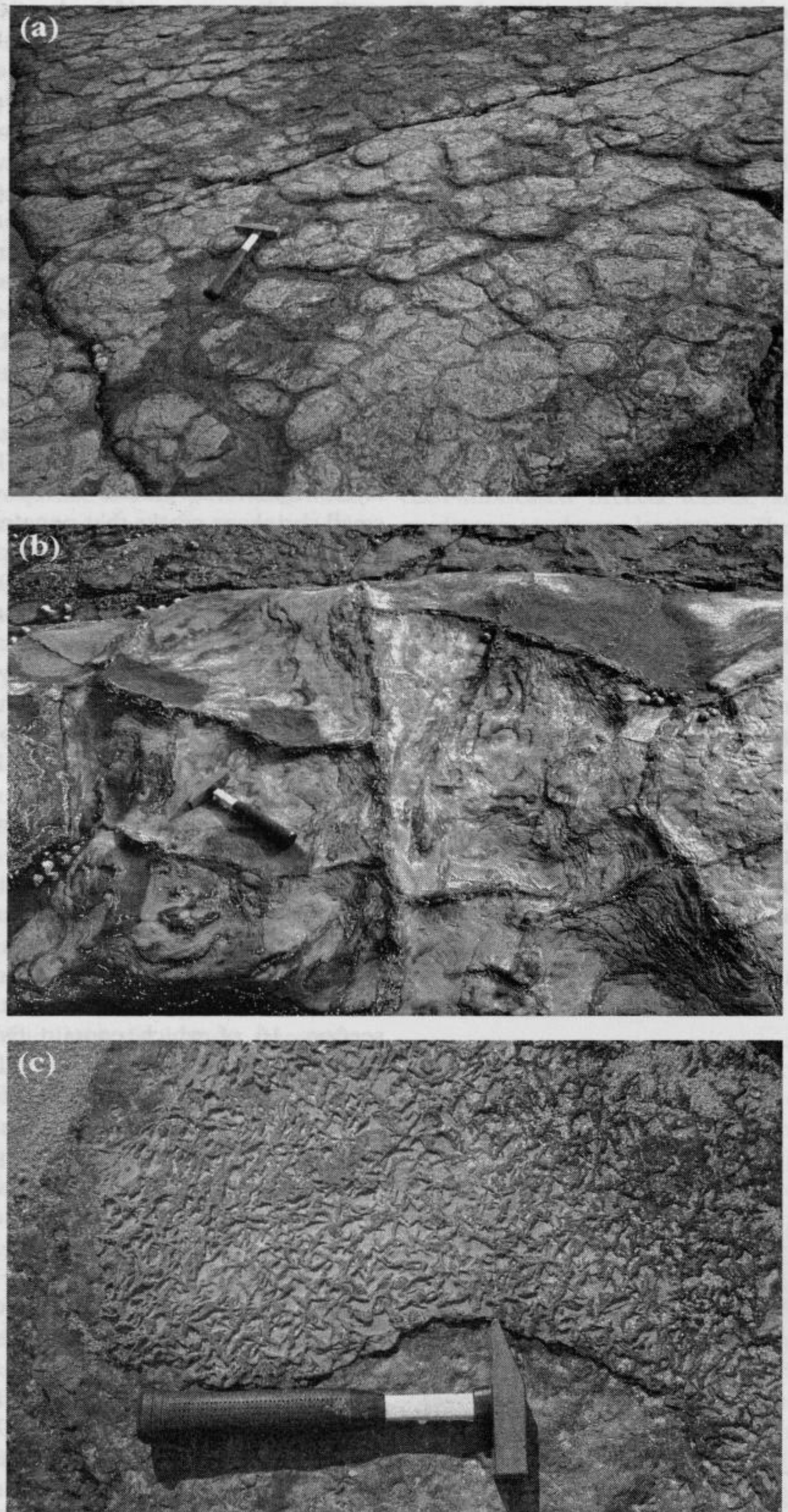


Figure 2.46 Styles of shrinkage cracks affecting thinly bedded mudstones and fine-grained sandstones. (a) Polygonal array of desiccation cracks; (b) more widely spaced orthogonal array of desiccation cracks; (c) lenticular shrinkage cracks forming complex interlocking pattern. (Photos: P. Stone.)

Figure 2.47 Compressed, deformed shrinkage cracks in cross-section. (Photo: BGS No. P547102, reproduced with the permission of the Director, British Geological Survey, © NERC.)

deformation structures, with widespread convolute lamination, load-induced 'pseudonodule' layers (Figure 2.48) and sandstone dykes, which cut up to 3 m of strata. Cross-bedded sandstone bodies up to 4 m thick occur towards the top (the western end) of the section. Some have trough cross-bedding, with locally abundant rip-up mudstone clasts, suggesting a fluvial origin. Others have low-angle cross-bedding, with lag accumulations of coarse, well-rounded quartz grains, and are probably aeolian. Small pyrite nodules are locally common in some of the more massive sandstone beds. Stratigraphically, this part of the Thurso section forms a link with the overlying John o'Groats Sandstone Group, in which high-energy, fluvial sandstones are

Figure 2.48 Load-induced 'pseudonodule' layer. (Photo: P. Stone.)

dominant. These can be examined near the type locality of **John o'Groats** south of Duncansby Head (ND 405 735; see GCR site report, this chapter). There, trough and planar cross-bedding, low-angle planar bedding and primary current lamination are seen in broadly cyclic, upward-fining channelized units, and mudstone clasts are common in the channel bases (Trewin, 1993).

Interpretation

The strata exposed in the coast section between Thurso and Scrabster record the interaction of lacustrine, fluvial and aeolian processes in and around an Orcadian Basin lake of fluctuating water-level. The lithofacies can be correlated with the facies associations recognized by Donovan (1980) (Figure 2.3), and record the repeated, abrupt variations in lake level from the deepest-water, fish-bearing laminites ('association A') to intermittently emergent playa-lake deposits ('association D'). The shallower lake deposits ('associations B and C') are poorly represented in the Thurso section.

The fish-bearing laminites ('association A') represent the slow accumulation from suspension of fine-grained sediment in the deeper part of a large lake over a period of hundreds, possibly thousands, of years. The sub-millimetre scale alternations of fine clastic siltstone, carbonate and organic laminae have been interpreted by Donovan (1980) as varved sediments deposited under annual, seasonal, climatic control. Trewin (1993) interpreted the clastic layers as representing input from rivers in the rainy season, the carbonate laminae as the deposits of the dry, warm summer season when the photosynthetic activity of phytoplankton was at a maximum, and the organic laminae as the product of autumnal, annual decay of the phytoplankton. Water depths may have been as great as 80 m (Hamilton and Trewin, 1988), but the presence of stromatolitic layers within the laminites necessitates shallower water in the photic zone, perhaps up to about 50 m, for at least part of the depositional cycle.

The laminites pass gradationally up through thinly interbedded fine-grained sandstones and mudstones into interbedded internally cross-laminated siltstones and sandstones. The latter lithofacies forms most of the succession between the fish-bearing laminites. It indicates a depositional environment in which the lake floor was periodically emergent and sand was transported by shallow streams and reworked by wave action (Trewin, 1993). The abundant polygonal desiccation cracks are evidence of the intermittent emergence and drying out of the lake-bed sediments, but the origin of the equally abundant, lenticular cracks has been the subject of debate. Donovan and Foster (1972), Trewin (1992) and Barclay *et al.* (1993) considered that the lenticular cracks were subaqueous in origin and developed by a synaeresis-like process perhaps linked with salinity changes in the lake. Astin and Rogers (1991, 1992, 1993) expressed a contrary view, suggesting that the formation of gypsum (and possibly halite) crystals during the drying-out of the lake (with commensurate increase in salinity) was an important precursor to the formation of the lenticular cracks. These were then initiated by the crystal pattern in a subaerial environment with the type of crack pattern developed being controlled by the extent of desiccation and the thickness of the sediment layer involved. During dry periods, wind-blown sand filled the open cracks and was deposited as thin lenses and laminae on the exposed playa-lake floor. The interpretation of Astin and Rogers requires many more and longer periods of subaerial conditions than the Donovan and Foster model.

Superimposed on the cycles of lake sedimentation is the overall upward increase in both the thickness and frequency of sandstone beds. These show a combination of sedimentary features that suggests aeolian reworking of fluvial sands into small dunes and rippled sheets. The environment was probably marginal to a receding lake, with fluvial sedimentation in broad, shallow channels and intermittent influxes of wind-blown sand. The thicker sandstones are commonly convoluted and disrupted by a combination of loading and water-escape structures.

Conclusions

The Pennyland (Thurso–Scrabster) GCR site provides an exceptionally well-exposed representative section for part of the Middle Devonian Upper Caithness Flagstone Group and its transition into the overlying John o'Groats Sandstone Group. The cyclical arrangement of the strata represents variations in the depth of the Orcadian Basin lake. The site is therefore of great importance in regional interpretation of the tectonics and palaeogeography of the Orcadian Basin. In addition, there is good preservation of a spectacular array of shrinkage cracks, the origins of which remain a matter of debate. The site is also of great importance because of its well-preserved late Givetian fossil fish fauna (Dineley, 1999a).

JOHN O'GROATS, CAITHNESS (ND 380 735–ND 407 735)

Potential ORS GCR site

W.J. Barclay

Introduction

The John o'Groats GCR site is already accorded protected status on account of its Mid-Devonian fish fauna, preserved in the John o'Groats Fish Bed. This is the best site for the highest Mid-Devonian fauna in Caithness and is described in the companion GCR volume on the *Fossil Fishes of Great Britain* (Dineley and Metcalf, 1999). The following account summarizes the account of Dineley (in Dineley and Metcalf, 1999) and incorporates important Old Red Sandstone sections to the east at Ness of Duncansby and Duncansby Head (Figure 2.49). The area provides the type locality of the John o'Groats Sandstone Group, the highest Middle Devonian unit of Caithness. The John o'Groats Fish Bed crops out on the foreshore 365 m north-east of the John o'Groats Hotel. Since the first reported discovery of *Microbrachius dicki* by Peach (1868), the bed has yielded complete fish remains, although in less rich concentrations than the earlier fish beds in the Caithness Flagstone Group (Westoll, 1948). Trewin's (1993) field guide to the section and to the outcrops at Ness of Duncanby and Duncansby Head provides the basis of this account. Donovan (1978) gave a summary.

Description

Red sandstones of the Last House Formation (Foster, 1972; Donovan *et al.*, 1974) dominate the lowest outcrops, seen immediately east of the harbour (ND 380 735). They occur in beds up to 50 cm thick, in which trough and planar cross-bedding, parallel lamination with primary current lineation and ripple cross-lamination are seen. Soft-sediment deformation de-watering structures are common (Astin, 1985). Green to grey, thin-bedded sediments 100 m east of the harbour contain arrays of polygonal desiccation cracks, as well as wave- and current-rippled beds and lenticular cracks of subaqueous origin. Dineley (1999a) gives a detailed section. Three thin, dark grey, calcareous, varved siltstone laminites with pale carbonate concretions contain scattered fish fragments. The lowest of these is the 0.25 m-thick John o'Groats Fish Bed, which has yielded *Tristichopterus alatus*, *Pentlandia macroptera*, *Microbrachius dicki*, *Watsonosteus fletti* and *Dipterus* sp.. The fauna of this fish bed is typical of the Eday Group of Orkney, and although fish fragments occur elsewhere, this is the only known bed to yield whole specimens in the 610 m-thick John o'Groats Sandstone Group. The beds are displaced by a small NE-trending fault in a notch on the foreshore, and are only visible at low tide. Red sandstones and grey-green laminites with fish scales crop out to the east.

Permian volcanic rocks intrude the John o'Groats Sandstone Group in the axial area of a syncline at Ness of Duncansby (ND 390 739). There are two exposures, but it is not clear if they represent a single vent or two separate ones. Nepheline basalt dykes cutting the vent agglomerate have been K–Ar whole-rock dated at around 270 Ma (Macintyre *et al.*, 1981; Stephenson *et al.*, 2003).

To the east of Ness of Duncansby, sandstones and laminites lie on the eastern limb of the syncline, dips increasing to 40° in the Bay of Sannick. A dark grey laminite visible at low tide here may be the John o'Groats Fish Bed. To the east, the John o'Groats Sandstone Group is faulted against the Mey Subgroup (Upper Caithness Flagstone Group), the latter forming Duncansby Head (ND 405 735). The fault is marked by a gully, and, on the foreshore, a zone of fractured rocks. The John o'Groats Sandstone Group here comprises high-energy, channelized fluvial sandstones with trough and planar cross-bedding and low-angle planar bedding, and lacks the lacustrine facies seen to the west. Fining-upward cycles commence with flat-bedded, fine- to medium-grained sandstones with large-scale planar and trough cross-bedding. Intraformational mudstone clasts line the bases of units, which overlie erosion surfaces. The cycles end with thinner-bedded parallel- and ripple-laminated sandstones passing up into red and pink mudstones. The trough cross-bedding indicates consistent north-east palaeocurrents. Individual sandbodies are wedge-shaped and up to 1 m thick and 10 m

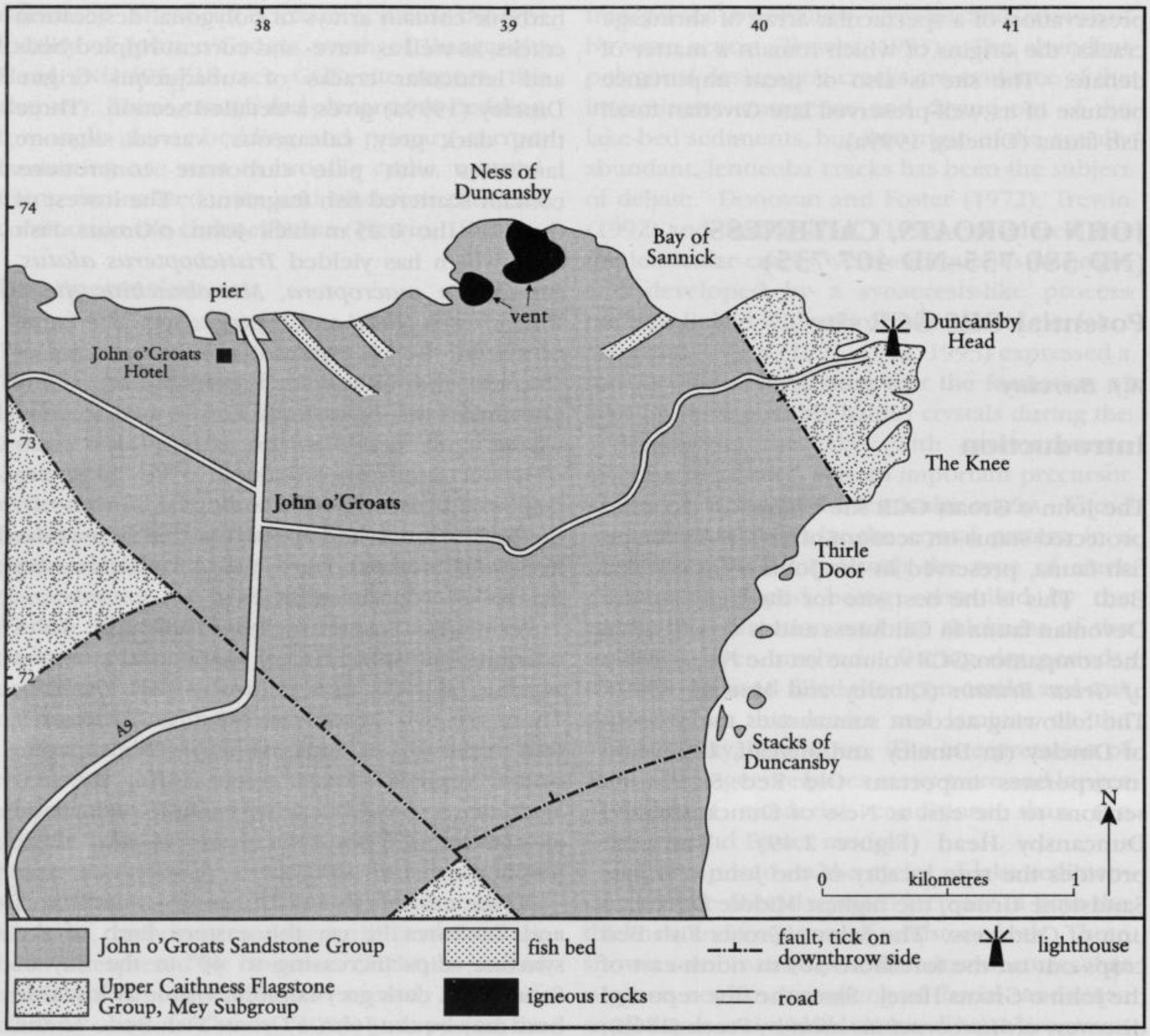


Figure 2.49 Map of the John o'Groats–Duncansby Head area. Based on British Geological Survey 1:50 000 Sheet 116 (Scotland), Wick (1985) and Trewin (1993).

wide. The Stacks of Duncansby offshore preserve more continuous sheet-like bodies up to 1 m thick. The Mey Subgroup comprises the typical grey-green, thinly bedded flagstones.

Interpretation

The John o'Groats Sandstone Group is interpreted as the deposits of shallow, braided or low-sinuosity streams on a broad, low-angle alluvial fan (Foster, 1972) that was subject to periodic inundation by lake waters. The John o'Groats Fish Bed has yielded fewer species than the earlier fish beds of the Orcadian Basin, suggesting to Dineley (1999a) that the fluvial-dominated environments were less suitable for some fish groups and that a long period of climatic stability may have given way to more aridity and only punctuated lacustrine lake development and fish habitation during cooler and/or wetter periods. The dark grey, fish-bearing laminites record periods of more permanent, deeper lacustrine conditions.

There appears to be a substantial amount of strata cut out by the fault separating the Mey Subgroup from the John o'Groats Sandstone Group east of the Bay of Sannick. The John o'Groats Sandstone Group is correlated with the Eday Group of Orkney (Astin, 1985), and on the basis of their respective fish faunas, it appears that all of the *Asterolepis orcadensis* Zone (Watson, 1935) is absent in Caithness (Dineley, 1999a).

Conclusions

The John o'Groats site is protected because of the presence of the John o'Groats Fish Bed, the best occurrence of a latest Mid-Devonian fish fauna in Caithness. It allows comparison with the richer faunas of Orkney and shows that lake development here was sporadic, with alluvial deposition being predominant. The outcrops around and to the east of the site are also important in providing the type locality of the John o'Groats Sandstone Group and the most easterly outcrops of the Mey Subgroup on the Scottish mainland.

WICK QUARRIES, CAITHNESS (ND 377 498–ND 374 491)

P. Stone

Introduction

At South Head, Wick, good exposure through part of the Middle Devonian Lower Caithness Flagstone Group is provided by a series of disused flagstone quarries cut into the sea cliffs. The quarries were last worked almost 100 years ago, since when differential weathering of the cliff faces has revealed an extraordinary array of sedimentary features. These and the lithologies present comprise a series of lithofacies produced during the sequential shallowing and deepening of a large lake occupying part of the Orcadian Basin. Of particular note are the laminated lacustrine deposits, including a fish-bearing bed; the remarkable range of shrinkage cracks, which have been variously interpreted as subaqueous and/or subaerial in origin; and the soft-sediment deformation structures showing both compressional and extensional bed-parallel movement. Field guides to the site are provided by Donovan (1978), Parnell *et al.* (1990) and Trewin (1993).

The importance of the Wick Quarries site lies in its exceptionally well-displayed sedimentary features. This allows interpretation and characterization of an otherwise poorly exposed part of the stratigraphy of the Orcadian Basin. The site is also important for the broader tectonic, lithostratigraphical and palaeogeographical interpretation of the Orcadian Basin, as well as providing exposure of rocks that have a high total organic carbon content (TOC) and are a source for oil.

Description

The South Head promontory forms the south side of Wick Bay. Between the coastguard station and the Castle of Old Wick (Figure 2.50), a sea cliff, modified by extensive quarrying, provides an outstanding section through lacustrine strata of the Lybster Subgroup of the Lower Caithness Flagstone Group. The strata are probably of Eifelian (Mid-Devonian) age. The beds dip consistently northwards by about 10°, so that just over 100 m of strata are present within the GCR site, although minor faulting at the northern end of the section disrupts continuity.

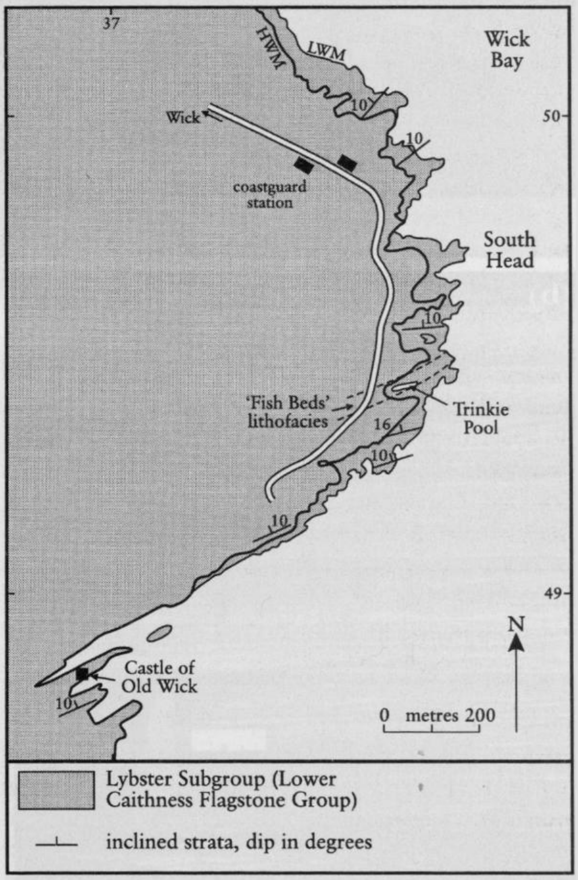
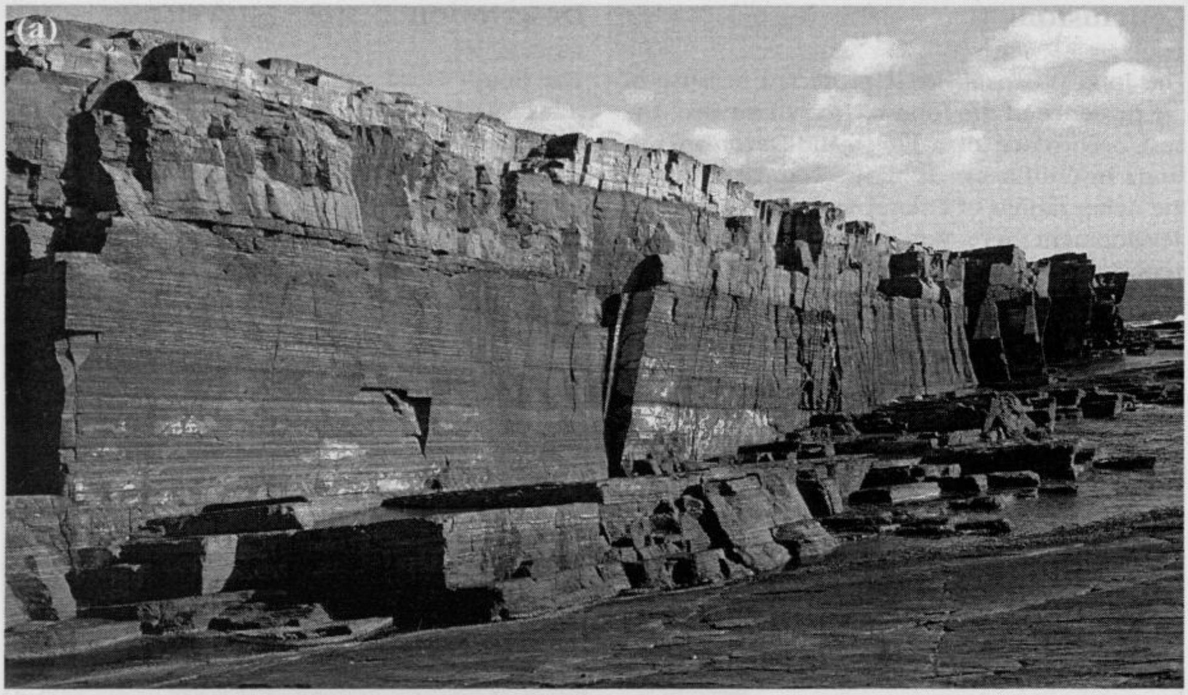


Figure 2.50 Map of the Wick Quarries GCR site.

The southernmost of the old flagstone quarries within the site area, adjacent to the road-end (Figure 2.50), affords a fine section through thinly laminated mudstones and fine-grained sandstones (Figures 2.51a,b). This transitionally overlies a sequence dominated by thicker sandstone beds, ranging up to about



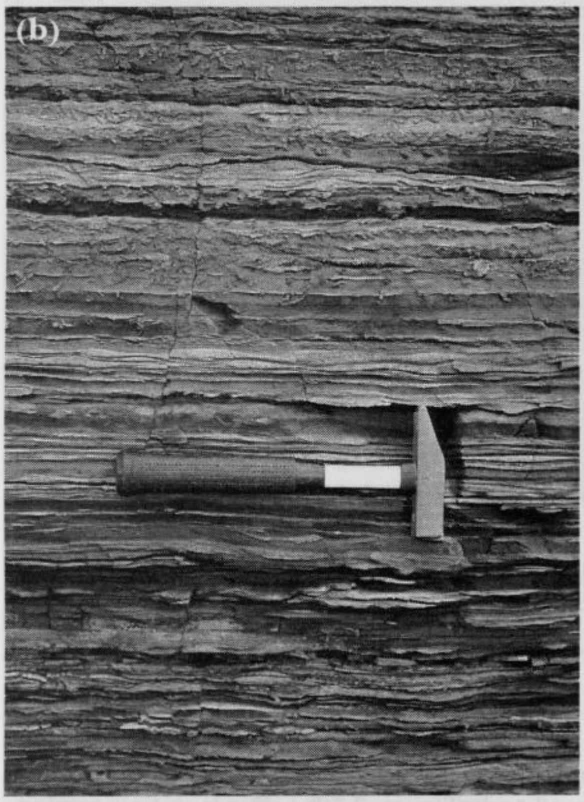


Figure 2.51 (a) Laminated mudstones and thin sandstones of the Lower Caithness Flagstone Group, South Head. (b) Detail of laminites, Lower Caithness Flagstone Group, South Head. (Photos: P. Stone.)

10 cm, which are commonly disrupted by extensive desiccation polygons (Figure 2.52). The quarry face exposes hundreds of couplets of mudstone–fine sandstone, each between a few millimetres and several centimetres thick; it is an excellent example of 'B to C association' flagstones as defined by Donovan (1980) (see 'Introduction', Figure 2.3 for further details). The sandstone layers commonly extend downwards into small shrinkage cracks that penetrate the underlying mudstone, the so-called 'fang structures' of Donovan (1980). Differential compaction strongly deformed many of the crack infills so that they are bulbous or sinuous in cross-section. It is understandable that they were originally interpreted as bioturbation (Crampton and Carruthers, 1914) and their origin remains controversial (see 'Interpretation' below).

Interbedded with the mudstone–sandstone couplets are abundant, thin dolomitic layers up to about 2 cm thick that weather to a distinctive orange-brown colour. During diagenesis, the dolostones were lithified earlier than the adjacent mudstone–fine sandstone couplets and so behaved in a more competent fashion during early deformation. As a result, the dolostones preserve a remarkable array of small-scale compressional and extensional, brittle and ductile

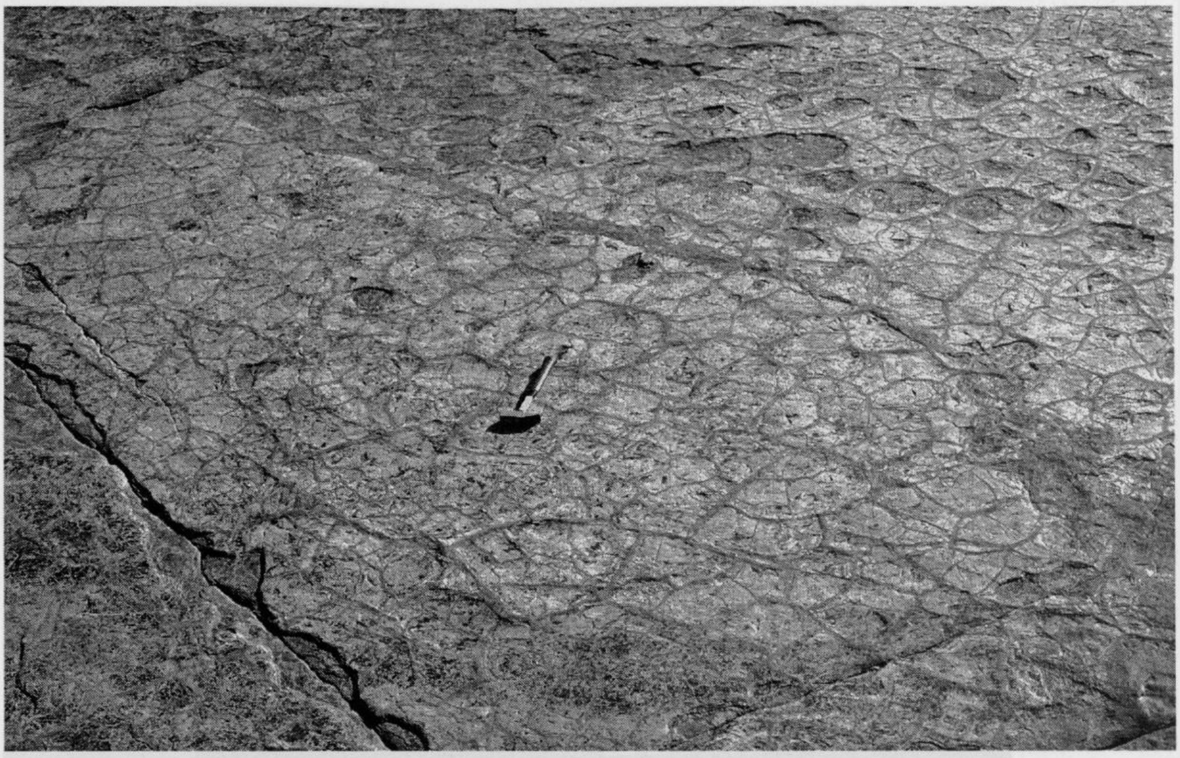


Figure 2.52 Desiccation polygons, Wick Quarries, South Head, Wick. (Photo: P. Stone.)

structures (Figure 2.53). Recumbent folds and thrusts demonstrate considerable local shortening (Figure 2.53a,b), but other beds in close proximity are boudinaged or offset by small arrays of en-echelon normal faults (Figure 2.53c).

The main face of the quarry exposes a sandstone body above the laminites. About 2 m thick, in cross-bedded sets up to about 30 cm, the sandstone is the 'D association' of Donovan (1980). North from here, the central part of the GCR site spans a quarried area around Trinkie Pool, which is slightly higher in the sequence than that exposed in the southernmost quarry. The distinctive mudstone–fine sandstone laminites again dominate, but sporadic thicker (up to about 10 cm) sandstone beds are also present. There is one thicker development of laminated mudstone ('A association' of Donovan, 1980) from which fish remains have been recovered and which is reported to be cut by hydrocarbon-bearing veins (Trewin, 1993). To the north, and higher in the succession, the laminites are interbedded with increasing amounts of thin sandstone beds, which are generally internally cross-laminated and, in some cases, extensively and chaotically convoluted. Thin sandstone dykes cutting obliquely across 1 m or more of strata are further evidence of wet-sediment mobility. Exposed bedding surfaces of the sandstones are extensively rippled (Figure 2.54) and carry a range of cracks. Some are polygonal, but most are commonly lenticular and either randomly arranged or aligned. Locally, the crack patterns interfere to give complex, mixed arrays that are probably polygenetic.

Interpretation

The strata exposed in the series of old quarries at South Head, Wick were originally laid down in a large lake. The lithologies reflect changes in lake conditions, primarily water depth, and these have been grouped together into four general lithofacies associations by Donovan (1980) (see 'Introduction', and Figure 2.3 for further details). The South Head sequence shows two major deepening–shallowing cycles, on which bed-by-bed variations are superimposed.

The southernmost, oldest strata are sandstones, which were probably shallow-water deposits, with some polygonal desiccation cracks indicating sporadic emergence as the lake

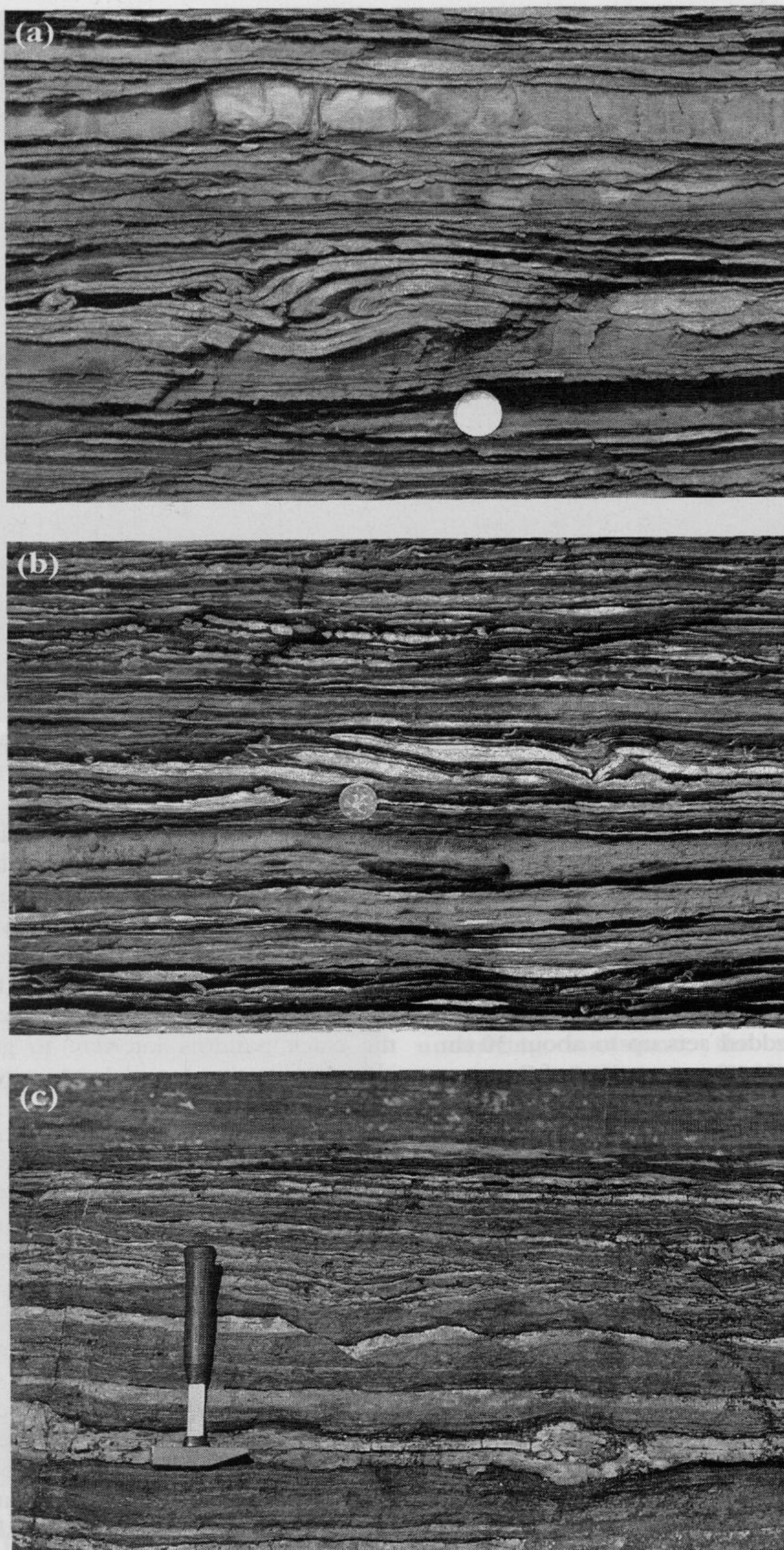


Figure 2.53 Laminites of mudstone, sandstone and dolostone, Wick Quarries. (a,b) Small-scale recumbent shortening structures; (c) small-scale brittle extensional faults. (Photos: P. Stone.)

Figure 2.54 Rippled sandstone surface, Wick Quarries. (Photo: P. Stone.)

dried up. Stratigraphically above these sandstones, the laminated mudstone–fine-grained sandstone couplets ('B to C association') seen in the southernmost quarry are the deposits of a shallow lake with a seasonally fluctuating water-level. Sporadic, thicker sandstone beds are probably turbidite units. Higher in the sequence, the significantly thicker and more massive sandstone body (the 'D association') is evidence for a return to very shallow water depths. It represents an increased clastic input to what had previously been a relatively distal lacustrine area, the shallower water allowing wave or fluvial reworking of the sediment in cross-bedded sets. Parnell *et al.* (1990) go further and suggest a partial aeolian origin for the cross-bedding, implying temporary emergence.

Subsequent deepening of the lake is shown by the re-appearance of the 'B to C association' laminites, a trend that reaches its fullest development with the fish-bearing mudstone laminite close to Trinkie Pool. In detail, this bed comprises sub-0.5 mm, alternating layers of fine, clastic sediment, carbonate and organic carbon probably derived from seasonal algal blooms. The whole assemblage is interpreted as 'a non-glacial varve recording sequential deposition in a tropical, eutrophic lake whose waters were subject to some degree of thermal stratification' (Donovan, 1980). The 'A association' unit is evidence for a deep lake environment remaining relatively stable for at least hundreds, and possibly thousands, of years. Thereafter, renewed shallowing of the lake is shown by the up-sequence (northward) return to 'B to C association' laminites and the increasing proportion of sandstone beds with polygonal desiccation cracks and rippled surfaces.

Polygonal crack patterns on bedding surfaces have been cited above as evidence for shallowing of the lake, with periodic emergence and desiccation of the sediments. However, most of the shrinkage cracks seen in the South Head section are isolated, curved and lenticular, and their interpretation is more controversial. Donovan and Foster (1972) considered them to be subaqueous in origin, and formed by a synaeresis-like process, perhaps linked to salinity changes in the lake. This interpretation has been supported by Trewin (1992, 1993), Barclay *et al.* (1993) and Trewin and Thirlwall (2002), and precludes subaerial conditions. Rogers and Astin (1991) and Astin and Rogers (1992, 1993) expressed a contrary view, suggesting that the formation of gypsum, and possibly halite, crystals during the drying out of the lake (with commensurate increase in salinity) was an important precursor to the formation of the lenticular cracks. The crack patterns and geometry were controlled by that of the crystals, the extent of desiccation and the thickness of sediment layer. During dry periods, wind-blown sand filled the open cracks and was deposited as thin lenses and laminae on the exposed playa-lake floor. The Astin and Rogers interpretation requires longer periods of subaerial conditions than the interpretation of Donovan and Trewin.

One of the most striking features of the 'B to C association' laminites is the widespread presence of small-scale, but intense compressional and extensional structures (Figure 2.43). The structures are picked out by the dolomitic layers that appear to have lithified before the enclosing mudstone, and so responded differently to early deformation. Compressional structures are more common than extensional features, but both are extensively developed and may occur in close proximity to each other, both vertically and horizontally. A complex, polyphase history of bed-parallel movement seems likely. This may have partly occurred during burial and loading, but may also be a result of adjustment during extensional faulting and subsidence of the Orcadian Basin.

Conclusions

The Wick Quarries GCR site provides an exceptionally well-exposed representative section of part of the Middle Devonian Lower Caithness Flagstone Group. Two lithofacies cycles are seen, representing the sequential shallowing and deepening of the Orcadian Basin lake. The site is of great importance in regional interpretation of the tectonics and palaeogeography of the Orcadian Basin. In addition, the excellent preservation of a spectacular array of shrinkage-crack styles provides pertinent evidence to current scientific debate concerning their origin. Polygonal crack patterns have traditionally been associated with desiccation, whereas lenticular cracks have been regarded as subaqueous. A recent re-interpretation of the lenticular cracks in the Wick Quarries section as the result of evaporite mineral formation and subsequent dissolution, subaerial desiccation and infilling by aeolian sand presents a radically different interpretation of lake chemistry and sedimentation in this part of the Orcadian Basin. A further aspect of the importance of the site lies in it being one of the few localities where potential hydrocarbon source rocks can be examined.

ACHANARRAS QUARRY, CAITHNESS (ND 150 544)

Potential ORS GCR site

P. Stone

Introduction

The disused quarry on Achanarras Hill provides a rare exposure of the Middle Devonian Achanarras Limestone Member. This distinctive unit is a lithostratigraphical marker bed separating the Upper and Lower Caithness Flagstone groups and allowing correlation of Orcadian Basin sequences from Shetland and Orkney south to the Moray Firth (Trewin and Thirlwall, 2002; Marshall and Hewett, 2003). The richest Old Red Sandstone fish site in Great Britain, the site is already a palaeontological GCR site, of international importance and renowned for its abundant and varied, well-preserved fossil fish fauna, including many whole specimens. The following account supplements that in the fossil fishes GCR volume (Dineley, 1999a). The seminal modern work on the Achanarras Limestone Member and its remarkable fish fauna is by Trewin (1986), who has also provided a field guide to the site (Trewin, 1993).

Achanarras Quarry was intermittently worked for flagstone and roofing slate from about 1870 to 1961. Since 1980, the quarry has been managed by the Nature Conservancy Council (and subsequently by Scottish Natural Heritage), and strict access and fossil collecting conditions apply. The limestone member represents the fullest development of a deep-water, lacustrine lithofacies seen in the Orcadian Basin. In addition to its unique sedimentological features and fish fauna, the site has important implications for an overview of basin palaeogeography and tectonics. The broad geological setting is described by Johnstone and Mykura (1989) and Mykura (1991).

Description

The GCR site is centred on the disused quarry excavated on the north side of Achanarras Hill, about 2 km west of the village of Spittal (Figure 2.55), where (at the time of writing) flagstones are being quarried from a stratigraphical level slightly above that seen at Achanarras. At Achanarras Quarry, the worked rock faces provide exposure through the Achanarras Limestone Member, a distinctive, 3.6 m-thick unit of fish-bearing, carbonate-rich laminites taken to mark the top of the Lower Caithness Flagstone Group (Donovan *et al.*, 1974; Trewin, 1986). The fish bed is at or slightly below the Eifelian–Givetian boundary (Paton, 1981). In the exposed quarry section (Figure 2.56), the laminites strike approximately north–south and dip a few degrees towards the east.

The section exposed is summarized in Figure 2.57 (after Trewin, 1986, 1993). At the base, thinly bedded, grey siltstone contains paler, silty laminae and small, isolated, rippled lenses of

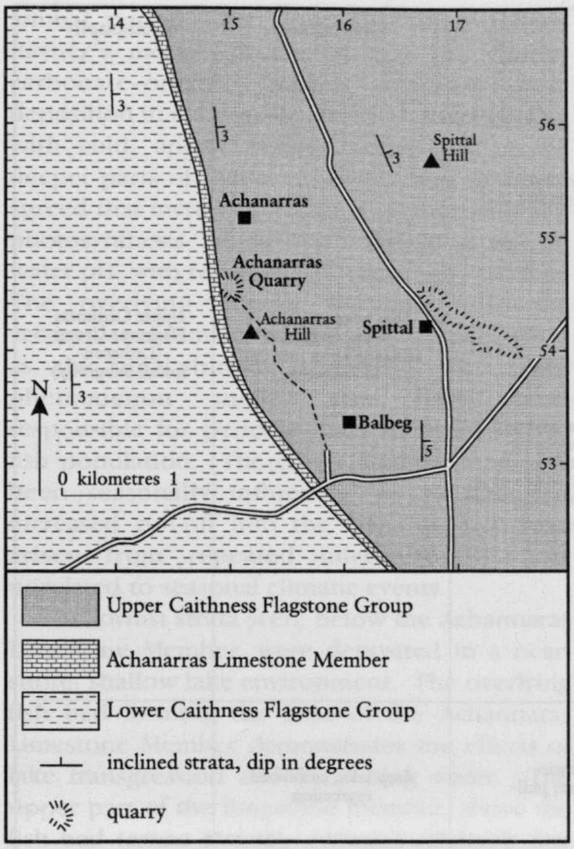


Figure 2.55 Geology of the Achanarras area and location of the Achanarras Quarry GCR site.

fine-grained sandstone; plant detritus is fairly common. These beds are the topmost of the Robbery Head Subgroup of the Lower Caithness Flagstone Group. Above, there is an abrupt transition to the dark grey, finely laminated fish bed, which forms the lowermost 1.95 m of the Achanarras Limestone Member (Trewin, 1986). The lamination is caused by a fine alternation of clastic, organic and carbonate (calcitic or dolomitic) laminae in varying relative proportions. The laminae are sub-millimetre in thickness, with an average clastic–carbonate pair about 0.7 mm thick; the organic laminae are very thin, and mostly less than 0.1 mm (Trewin, 1986). Trewin found six horizons of different fish diversity and relative abundance of species, and full details of the fauna are given by Dineley (1999a). The site is the type locality of ***Rhamphodopsis threiplandi*** Watson and *Palaeospondylus gunni* Traquair. Trewin (1993) noted that *Gyroptychius* is absent at the site, although common in the stratigraphically equivalent Sandwick Fish Bed of Orkney.

The upper part of the Achanarras Limestone Member, from which few fish fossils have been recovered, consists largely of clastic and dolomitic laminae. The fish bed and part of the overlying laminite unit are below water-level in the flooded quarry, but the topmost 50 cm of the

Figure 2.56 Part of the disused Achanarras Quarry, now flooded, showing the upper 50 cm of the Achanarras Limestone Member overlain by the basal beds of the Upper Caithness Flagstone Group. (Photo: P. Stone.)

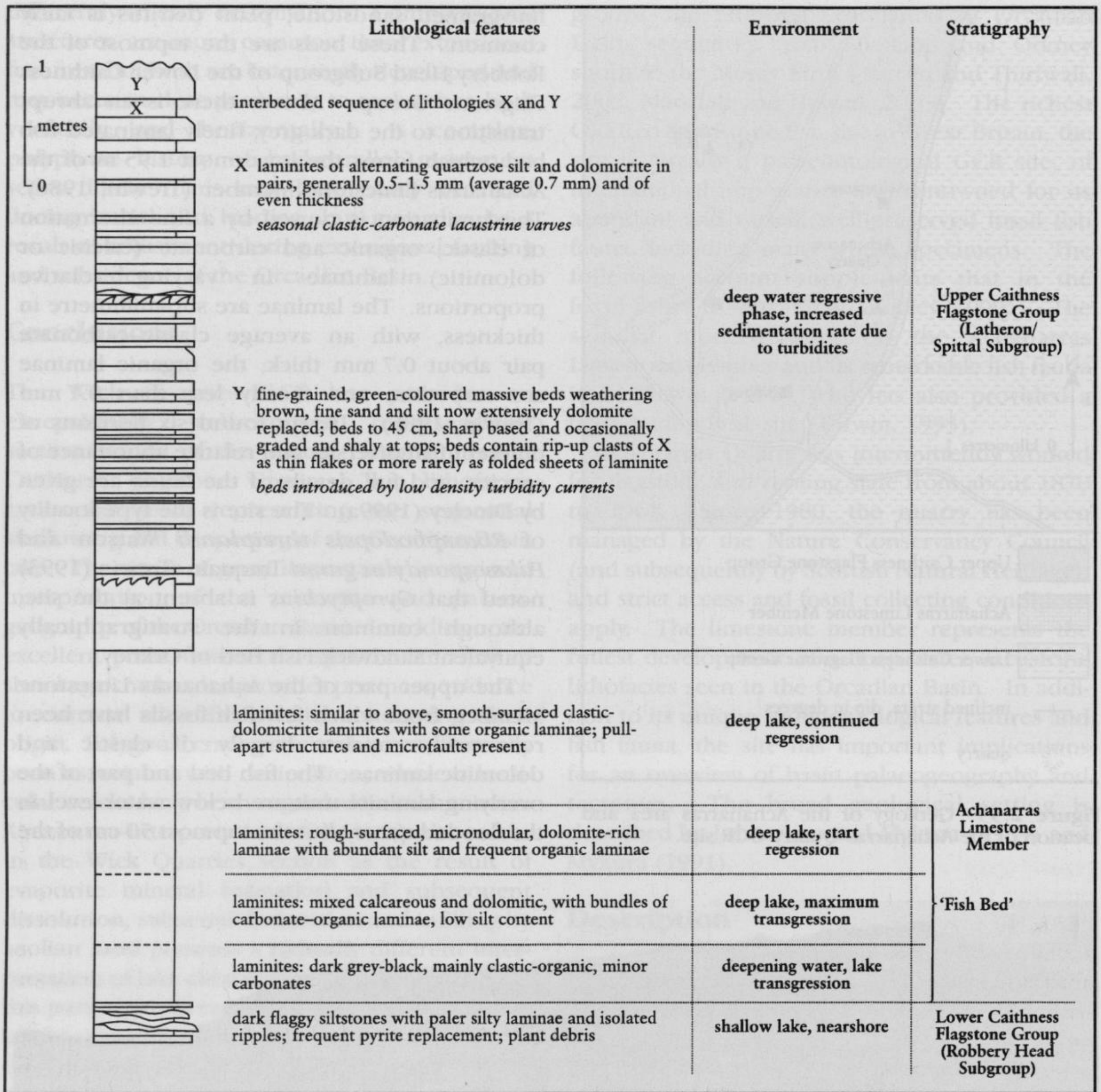


Figure 2.57 Section at Achanarras Quarry. After Trewin (1986, 1993).

limestone member form part of the rock face immediately above water-level (Figure 2.56). Most of the strata seen are the lowest beds of the Upper Caithness Flagstone Group (the Latheron Subgroup, locally known as the 'Spittal Beds'). They consist of thin units of clastic–dolomitic laminites, lithologically similar to those in the upper part of the Achanarras Limestone Member, alternating with beds of fine-grained sandstone ranging up to 45 cm thick. The sandstone beds are fairly massive and greenish grey where fresh, but weather brown owing to a high proportion of secondary dolomite. The bed bases are mostly sharp and some grading is common, sporadically fining upwards to a thin mudstone at the top of the bed. Rip-up clasts of clastic–dolomitic laminite are fairly common in the lower parts of the sandstone beds.

Interpretation

The varying lithofacies in the Achanarras section were ascribed by Trewin (1986) to the effects of lake transgression and regression. The fine laminations of the fish bed were interpreted by Rayner (1963) as lacustrine varves with an annual periodicity. Donovan (1980) refined Rayner's model, proposing that the clastic–carbonate–organic triplets resulted from deposition in a thermally stratified, tropical lake, with cold, anoxic bottom conditions in the deeper parts of the lake. Rayner and Donovan agreed that increased seasonal algal growth and photosynthesis would have caused a rise in water pH, with resulting carbonate precipitation. The overlying organic laminae were the accumulated remains of the dead phytoplankton as the algal bloom decayed. The latter phenomenon might also have been responsible for periodic mass mortality in the fish population. The clastic laminae may have been seasonally influenced by rainfall and increased run-off into the lake, or may have arisen from repeated microturbidite flow unrelated to seasonal climatic events.

The lowest strata seen, below the Achannaras Limestone Member, were deposited in a near-shore, shallow lake environment. The overlying fish bed forming the base of the Achannaras Limestone Member demonstrates the effects of lake transgression and deepening water. The upper part of the limestone member, above the fish bed (*sensu stricto*), contains relatively few organic laminae, reflecting a decline in plankton productivity, and a higher proportion of clastic input. This marks the beginning of lake regression, but the fully developed, deep-water regressive phase is represented by the abrupt incoming of substantial, low-density turbidite sands at the base of the Upper Caithness Flagstone Group. Between turbidity flows, deposition of the clastic–carbonate laminites continued. Assuming a broad seasonal control on laminite deposition, the Achanarras Limestone Member is estimated to represent accumulation over about 4000 years (Rayner, 1963; Trewin, 1986).

Conclusions

The GCR site at Achanarras Quarry provides a rare section through the Achanarras Limestone Member and its contacts with the overlying and underlying strata of the Upper and Lower Caithness Flagstone groups. The limestone member is a valuable stratigraphical marker bed for correlation within the Orcadian Basin, and contains a remarkable fossil fish fauna of international importance. The lithofacies record the transgression and regression of a lacustrine environment, possibly recording the maximum transgression by the Orcadian Basin lake. Lacustrine deposition of the fine laminites was controlled by seasonal increased algal productivity and the resulting changes in lake water chemistry. The organic laminae are the product of decay of seasonal algal blooms, the carbonate laminae were deposited during periods when increased photosynthesis raised the pH of the lake waters. The fine-grained clastic laminae were deposited from suspension of material introduced to the lake, probably mainly during periods of seasonal rainfall and increased run-off.

SARCLET, CAITHNESS (ND 353 431–ND 354 435)

Potential GCR site

P. Stone

Introduction

The cliff sections on either side of The Haven, Sarclet provide good exposure of conglomerates and sandstones of the Lower Devonian Sarclet Group. Fluvial and probable aeolian lithofacies are overlain to the north by mudstone marking a major lacustrine transgression. A spectacular aspect of the section is the presence of deformed sheets overlying bed-parallel basal detachments, with other dislocation planes cutting bedding at a low angle. The deformation is likely to be of tectonic origin, but its age is uncertain. Field guides to the site are provided by Armstrong *et al.* (1978a) and Trewin (1993), and much detail is contained in a PhD thesis by Donovan (1970).

Good exposure of Lower Devonian strata is rare in the northern part of the Orcadian Basin. The Sarclet site is therefore of regional importance as an example of an otherwise poorly represented part of the stratigraphy, providing data necessary for an overview of the tectonics and palaeogeography of the basin. The broad geological setting of the site is given by Johnstone and Mykura (1989) and Mykura (1991).

Description

From the village of Sarclet a steep track leads down into The Haven, a long-abandoned base for herring fishing. Sarclet Head is a rugged promontory on the south side of The Haven. The cliffs around the headland and those extending north from The Haven provide extensive exposure through the Sarclet Conglomerate and Sarclet Sandstone formations of the Sarclet Group. These occupy the broad hinge zone of a large, open anticline that plunges gently towards the north-west and which is broken up locally by intersecting N–S- and ENE–WSW-trending fault sets (Figure 2.58). To the north and west, the Sarclet Sandstone Formation is overlain by mudstone and sandstone of the Ulbster/Riera Geo and Ulbster/Ires Geo formations. A spore assemblage from the latter was assigned a late Emsian (Early Devonian) age (Collins and Donovan, 1977). The strata in the Sarclet section lie on the northern limb of the regional anticline and dip consistently north-west between 20° and 30°, so that 150 m of beds are present.

The lowest strata seen are polymictic conglomerate lenses and pebbly sandstones of the Sarclet Conglomerate Formation that form Sarclet Head, the southern margin of the site. The conglomerates are poorly sorted and contain clasts of granite, schist, quartzite and basalt, the last generally abundant and dominant in some lenses, together with sandstone intraclasts. The clasts are mostly subrounded pebbles, but range up to 30 cm boulders. The junction with the overlying Sarclet Sandstone Formation is transitional, with the proportion of pebbly beds diminishing over a few tens of metres.

The cliffs surrounding The Haven and extending to the north are of reddish brown, medium-grained sandstone in parallel-sided beds up to 50 cm thick. Many of the parallel beds are internally cross-laminated, others contain wispy mud laminae or parallel lamination defined by size variation in the rounded sand grains (Trewin, 1993). The sandstones are quartz-cemented and extensively fractured in some parts. They are also affected by dislocation planes cutting across the sequence at low or moderate angles, and locally by chaotically deformed units overlying basal, bed-parallel slide planes (Figure 2.59). In the latter examples, the degree of deformation decreases upwards from the basal detachment, immediately above which the sandstone is highly fractured with abundant small quartz veins.

The coastal outcrop of the Sarclet Sandstone Formation is terminated in the north by faulting at Riera Geo (ND 354 439), where thinly bedded, greenish mudstones and grey, calcareous siltstones of the Ulbster/Riera Geo Mudstone Formation are thrown down to the north.

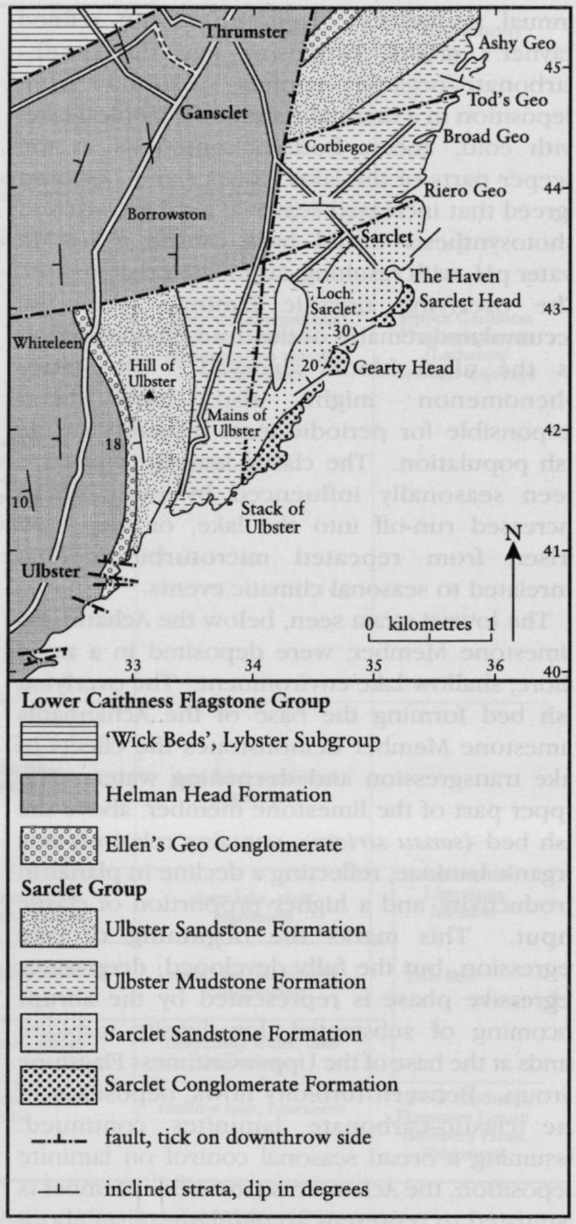


Figure 2.58 Geological map of the Sarclet area. After British Geological Survey 1:50 000 Sheet 110 (Scotland), Latheron (1985).

Interpretation

The lowest strata seen in the site area, the conglomerates and pebbly sandstones of the Sarclet Conglomerate Formation, are alluvial-fan deposits laid down by braided streams. The abundance of basalt detritus is of particular significance, suggesting a nearby area of contemporaneous volcanism (Trewin, 1993), which, from the evidence of the palaeocurrent indicators (Johnstone

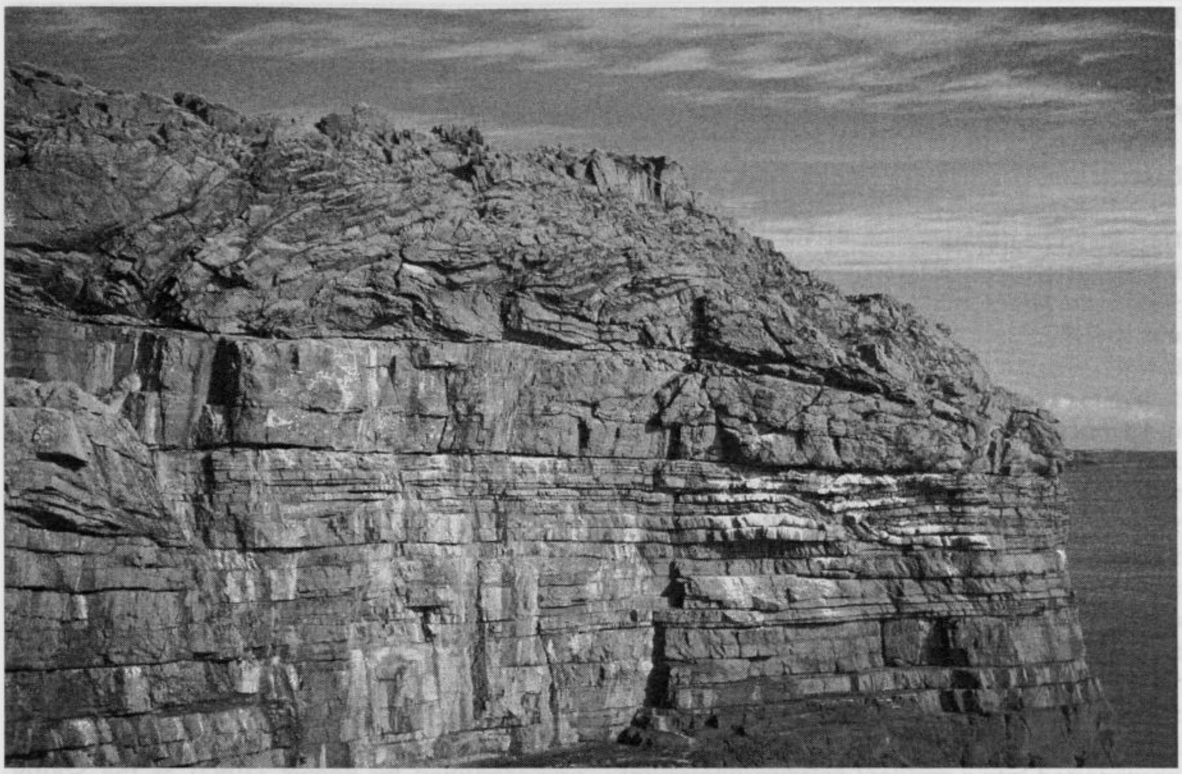


Figure 2.59 Sarclet Sandstone Formation. Deformed sandstones rest on a bedding-parallel detachment; a low-angle dislocation cuts the regularly bedded sandstone below the detachment. (Photo: P. Stone.)

and Mykura, 1989; Mykura, 1991), lay to the south-east. Some of the overlying Sarclet Sandstone Formation may also have a fluvial origin, Mykura (1991) noting evidence for increased meandering of the rivers with time. However, Trewin (1993) noted that the wispy lamination defined by thin, irregular mud laminae resembles aeolian adhesion ripples. The lamination produced by size sorting of rounded sand grains provides further evidence for deposition by aeolian processes. Higher in the succession, the thinly bedded mudstones and calcareous siltstones of the Ulbster/Riera Geo Formation record deposition in a shallow lacustrine environment, and represent one of the earliest, significant lake transgressions in the Orcadian Basin.

Of particular interest at this site is the localized deformation of beds in the Sarclet Sandstone Formation above bed-parallel slide surfaces (Figure 2.59). The overall appearance is suggestive of synsedimentary sheet slumping, but the most deformed sandstone is fractured and quartz-veined, and was clearly lithified when deformed (Trewin, 1993). The age of the deformation at Sarclet therefore remains uncertain.

Conclusions

The Sarclet site provides good exposure through a section representative of the Lower Devonian Sarclet Group. The site is of regional importance as outcrop of Lower Devonian strata is rare within Caithness and the northern part of the Orcadian Basin. Alluvial-fan conglomerates form the base of the exposed succession, with an upwards transition to sandstones that may be either fluvial or aeolian in origin, or formed by a combination of both processes. The Sarclet Sandstone Formation is cut by numerous low-angle dislocation planes and contains chaotically deformed sheets resting on bed-parallel detachment planes. The origin of this deformation is unclear, with both synsedimentary and post-lithification tectonic indicators, and its timing also remains uncertain. Overlying the Sarclet Group, mudstones and calcareous siltstones of the Ulbster/Riera Geo Mudstone Formation represent a major lacustrine transgression, one of the earliest significant lake developments in the Orcadian Basin.

TARBAT NESS, ROSS AND CROMARTY (NH 929 873–NH 939 851)

P. Stone

Introduction

The cliff and foreshore sections on either side of Tarbat Ness afford excellent exposure through parts of the Strath Rory Group (Middle Devonian) and the Balnagown Group (Upper Devonian). The site also provides evidence pertinent to the debate over the relationship between the Middle and Upper Devonian successions. A regional unconformity seems likely, but there appears to be a conformable succession at this site. A range of lithofacies is present, including fluvial pebble conglomerates and pebbly sandstones, aeolian sandstones, and a sabkha-type facies of sandstone and mudstone containing evidence of evaporitic deposition, desiccation, intermittent rainfall and subaerial animal activity. The most comprehensive description and interpretation of the site are by Rogers (1987), whose results were incorporated into a regional palaeogeographical analysis by Marshall *et al.* (1996).

The importance of the Tarbat Ness site is two-fold. Firstly, it provides a complete section through the Upper Devonian succession, allowing interpretation and characterization of an otherwise poorly represented part of the stratigraphy of the Orcadian Basin. It also provides one of the few apparently conformable transitions between the Middle and Upper Devonian strata. The site thus has sedimentological value, and is important to the broader interpretation of the palaeogeography, development and tectonics of the Orcadian Basin.

Description

The Tarbat Ness GCR site consists of the cliff and foreshore exposures to the west and south of the lighthouse. It forms the extreme north-eastern promontory of Easter Ross between the Moray and Dornoch firths. The exposed strata span the Middle Devonian Strath Rory Group and Upper Devonian Balnagown Group, the contact between them being variously interpreted as faulted or conformable. The dip is fairly uniform towards the north-west, steepening in that direction from 10° to nearly 40°, within the south-east limb of the Black Isle Syncline. About 800 m of beds are seen within the site, although continuity is disrupted by faulting towards the top and the bottom of the section.

The oldest strata are preserved on the east coast of the promontory, as a strike section along the shore southwards from a point about 1.5 km south of Tarbat Ness (Figure 2.60). They are thickly bedded, yellow and red, fluvial sandstones with subordinate (but fairly common) thin interbeds of calcareous mudstone. The beds were assigned by Armstrong (1977) to the Strath Rory Group. The sandstones are generally cross-bedded on a fairly large scale and internal lamination is highly convoluted locally. The mudstone interbeds contain sporadic carbonate concretions, some of which have yielded fish fragments that support an early Givetian (Mid-Devonian) age. Elsewhere, the Strath Rory Group ranges down into the Eifelian Stage (Dineley and Metcalf, 1999).

The top of the Strath Rory Group was placed by Armstrong (1977; see also Institute of Geological Sciences, 1973) at a fault on the south side of Port Tarsuinn, probably the same one identified by Rogers (1987) as the South Wilkhaven Fault (Figure 2.60). This interpretation allows for an unconformable relationship between the Strath Rory Group and the overlying strata, the unconformity itself being farther inland and unexposed. By contrast, Rogers (1987) proposed a complete and conformable succession, compiled in various fault blocks around Port Tarsuinn. He noted a thick sandstone body, probably of aeolian origin (the Port Tarsuinn Member of the Rockfield Formation in his informal stratigraphy), at the top of the Strath Rory Group conformably overlain by coarse, pebbly sandstones of the Balnagown Group.

Rocks of the Balnagown Group (Armstrong, 1977) form the headland of Tarbat Ness and are exposed from there along the north coast of the promontory. Red, pebbly sandstones with conglomeratic lenses and interbeds predominate at the base of the group, and are exposed northwards from Port Tarsuinn and around the headland itself. Rogers (1987) informally named these beds the 'Tarbat Ness Formation'. The pebbles are invariably well-rounded and mainly quartzose. They form conglomeratic beds up to about 20 cm thick, but more commonly occur as lenticular bodies (Figure 2.61a) or as isolated clasts strewn along bedding surfaces (Figure 2.61b). The

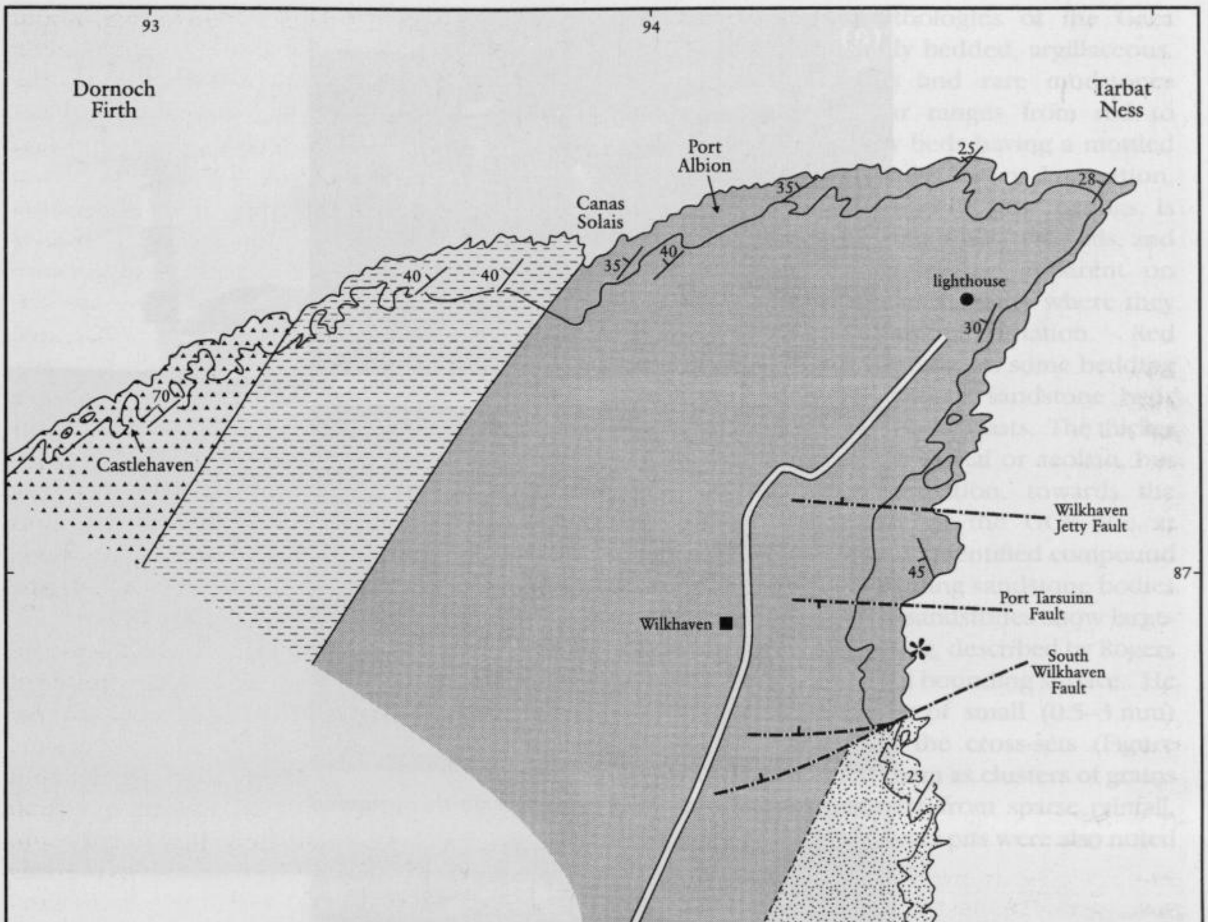


Figure 2.60 Geological map of the Tarbat Ness area.

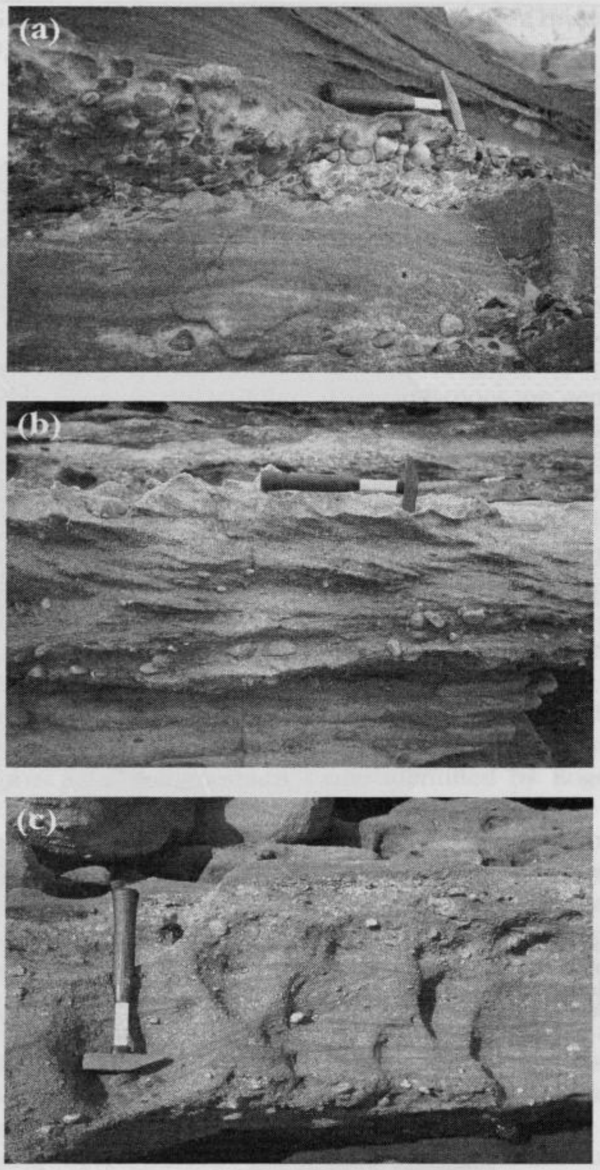
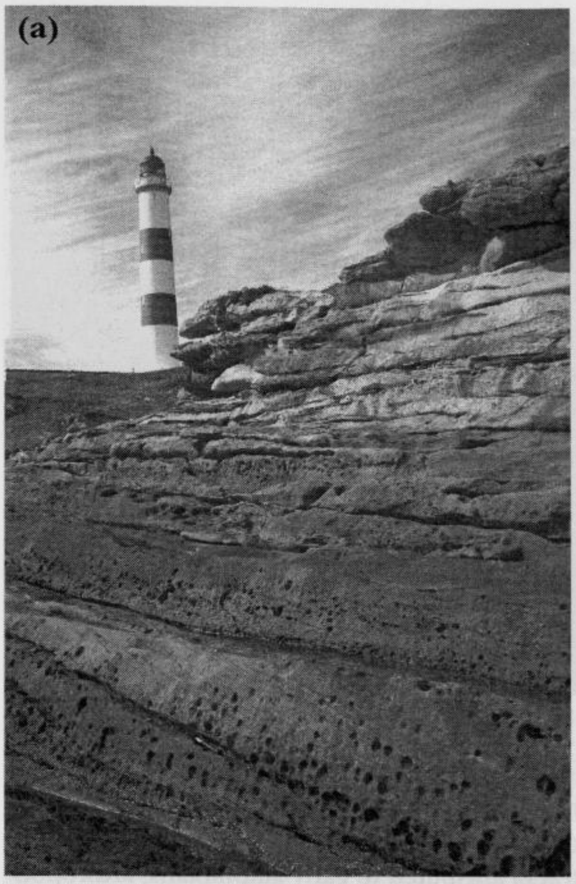


Figure 2.61 Tarbat Ness Formation. (a) Conglomerate lenses; (b,c) pebbly layers and isolated pebbles in cross-bedded sandstones. (Photos: P. Stone.)

formation is cross-bedded on a large scale, with conglomeratic layers forming individual cross-sets and isolated pebbles lining foresets (Figure 2.61c). The concentration of pebbles decreases up-sequence towards Tarbat Ness, but thereafter appears to increase again towards the top of the Tarbat Ness Formation. Around the lighthouse and headland, many of the thick, red sandstone beds contain only sporadic, isolated pebbles and many are pebble-free. The sandstones are cross-bedded on a large scale (Figure 2.62a), and there are many examples of convolute bedding and post-depositional deformation, probably caused by slumping and de-watering, the latter phenomenon having produced sand volcanoes in places (Figure 2.62b).

The top of the Tarbat Ness Formation, to the west of the headland, shows an irregular increase in the proportion of pebbles in the thick, cross-bedded red sandstones towards Port Albion and Canas Solais. There, pebble content and bed thickness abruptly decrease, above which the higher part of the Balnagown Group is dominated by thinly interbedded sandstone and mudstone. Rogers (1987) named these beds the Gaza Formation.



The characteristic lithologies of the Gaza Formation are very thinly bedded, argillaceous, fine-grained sandstones and rare mudstones (Figure 2.63a). Colour ranges from red to yellow-green, with many beds having a mottled appearance. An irregular, wavy lamination, defined by colour variation or silty partings, is fairly ubiquitous in the thin sandstone beds, and very low-amplitude ripples are apparent on some bedding surfaces, particularly where they are picked out by a colour variation. Red mudstone flakes are common on some bedding planes and sporadic, thicker, sandstone beds also contain isolated mudstone clasts. The thicker sandstones may be either fluvial or aeolian, but higher in the Gaza Formation, towards the north-western boundary of the GCR site at Castlehaven, Rogers (1987) identified compound aeolian dunes (draas) forming sandstone bodies up to 14 m thick. These sandstones show large-scale, internal cross-bedding, described by Rogers in terms of three orders of bounding surface. He also noted the presence of small (0.5–3 mm) 'adhesion warts' within the cross-sets (Figure 2.63b) and interpreted them as clusters of grains held together by moisture from sparse rainfall. Rare examples of rain impact pits were also noted elsewhere in the sequence.

Figure 2.62 Tarbat Ness Formation. (a) Tabular cross-bedding; (b) sand volcano. (Photos: P. Stone.)

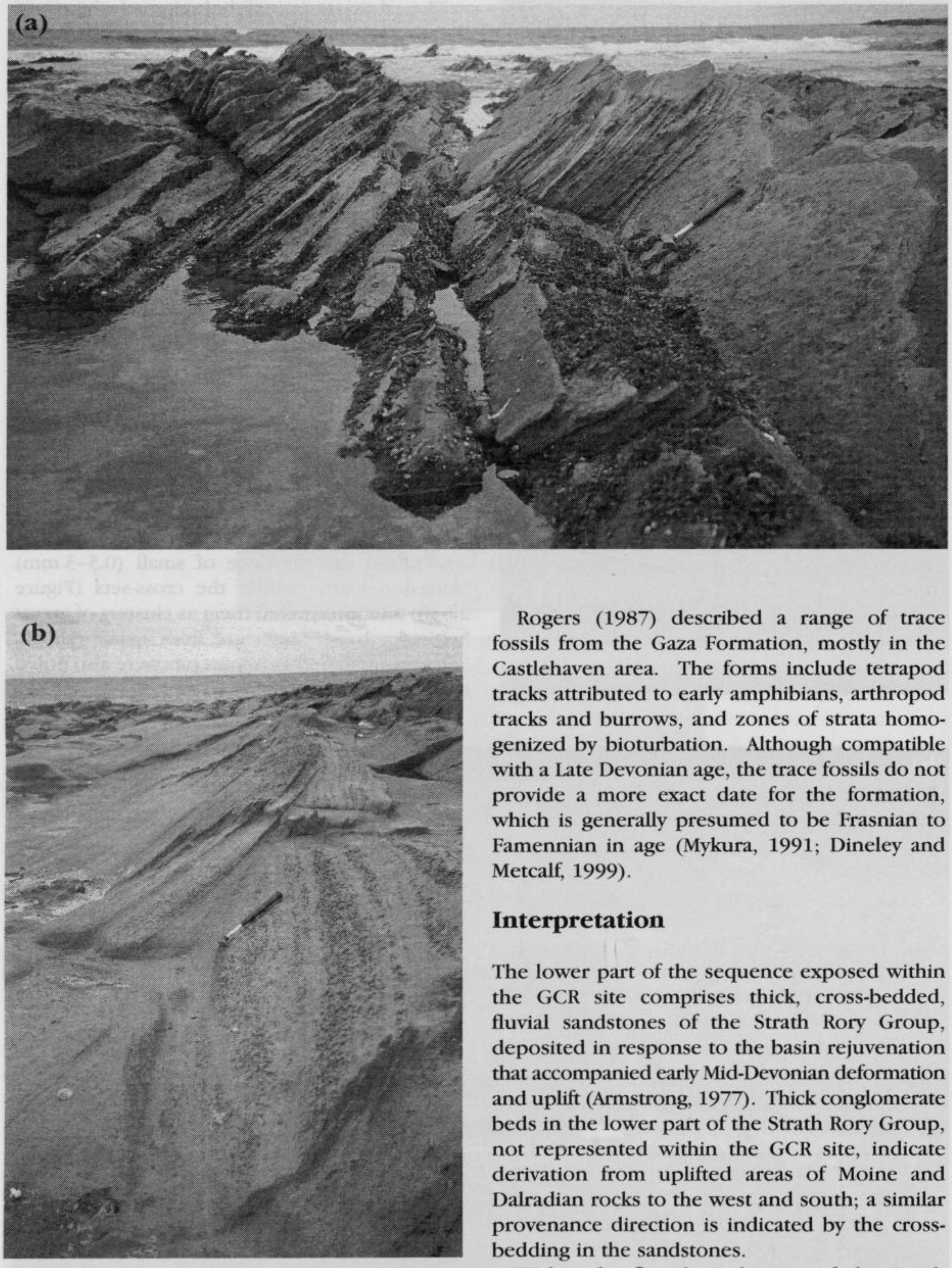


Figure 2.63 (a) Thinly bedded sandstones of the Gaza Formation west of Canas Solais. (b) 'Adhesion warts' in the Gaza Formation near Castlehaven. (Photos: P. Stone.)

Rogers (1987) described a range of trace fossils from the Gaza Formation, mostly in the Castlehaven area. The forms include tetrapod tracks attributed to early amphibians, arthropod tracks and burrows, and zones of strata homogenized by bioturbation. Although compatible with a Late Devonian age, the trace fossils do not provide a more exact date for the formation, which is generally presumed to be Frasnian to Famennian in age (Mykura, 1991; Dineley and Metcalf, 1999).

Interpretation

The lower part of the sequence exposed within the GCR site comprises thick, cross-bedded, fluvial sandstones of the Strath Rory Group, deposited in response to the basin rejuvenation that accompanied early Mid-Devonian deformation and uplift (Armstrong, 1977). Thick conglomerate beds in the lower part of the Strath Rory Group, not represented within the GCR site, indicate derivation from uplifted areas of Moine and Dalradian rocks to the west and south; a similar provenance direction is indicated by the cross-bedding in the sandstones.

Within the fluvial sandstones of the Strath Rory Group, interbeds of red mudstone with calcareous laminae and concretions probably represent lacustrine episodes and indicate the intermittent extension, into the Easter Ross region, of the main Orcadian Basin lake that was centred on Caithness. The thick aeolian sandstones (Port Tarsuinn Member) identified by Rogers (1987) at the top of the Strath Rory Group, provide evidence of temporary subaerial conditions prior to the re-establishment of a fluvial regime at the base of the Balnagown Group. This is marked by the appearance of conglomeratic lenses, pebbly sandstones and thick, cross-bedded fluvial sandstones, which, together with the absence of red mudstone interbeds, are characteristic of the Tarbat Ness Formation. The conglomerate pebbles are invariably well-rounded (Figure 2.61), suggesting some degree of reworking of the underlying Strath Rory Group. The fluvial sandstones show widespread internal convolute lamination and liquefaction phenomena such as sand volcanoes (Figure 2.62b), indicating de-watering during rapid burial of water-saturated sand, probably in a large, braided river system.

The top of the Tarbat Ness Formation marks a major environmental change. The overlying non-pebbly, fluvial sandstones and thinly interbedded sandstones and mudstones (the Gaza Formation) were interpreted by Rogers (1987) as having formed in a sabkha-like environment. The presence of the fragmented remains of desiccated mud layers as mudstone flakes in the sandstones, and 'adhesion warts' produced by differential wetting of a dry sand surface support this interpretation. Rippled surfaces may have formed either by adhesion of blown sand to an irregularly damp surface or by sheet flooding, but Rogers (1987) speculates that, in either case, wind scour may have accentuated the features. Roughly polygonal to irregularly sinuous ridges rising up to 1 cm above some bedding surfaces caused localized disruption of the underlying sandstone laminae. Rogers (1987) interpreted them as a result of surface buckling during expansion due to evaporite formation. This interpretation supports a sabkha origin. Rain pitting, and the tetrapod and arthropod tracks add further evidence of predominantly subaerial environments during deposition of the Gaza Formation.

At the north-western extremity of the GCR site, the Gaza Formation is overlain by a thick, compound aeolian dune sequence identified as a draa by Rogers (1987). From measurements of the internal set and co-set relationships, Rogers calculated dune heights up to about 6 m, with interdune distances of up to 80 m. Most of the animal tracks were recorded in the dune sandstones, on surfaces strewn with 'adhesion warts'.

A palaeoenvironmental model for the Upper Old Red Sandstone of the Tarbat Ness area, after Marshall *et al.* (1996) is shown in Figure 2.64. This envisages fluvially dominated marginal facies passing distally to sandy sabkha with migrating dunes.

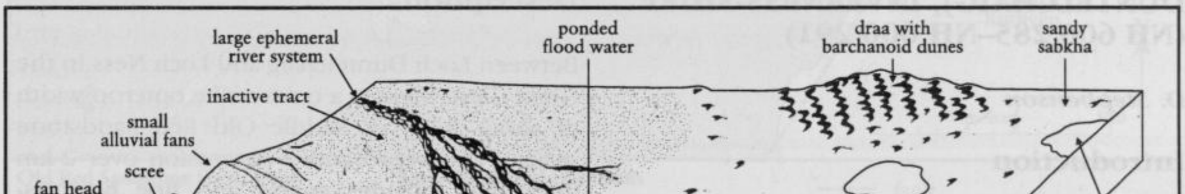


Figure 2.64 Palaeoenvironmental model for the Upper Old Red Sandstone of Tarbat Ness. After Marshall *et al.* (1996).

Conclusions

The Tarbat Ness GCR site provides an exceptionally well-exposed representative section of part of the Middle Devonian Strath Rory Group and its boundary with the overlying Upper Devonian Balnagown Group. This is one of the few localities where the Middle–Upper Devonian boundary can be examined, and here it appears to be conformable, in contradiction to the generally accepted model of a regional unconformity. The range of lithofacies exposed is unusually large. Fluvial sandstones with sporadic lacustrine mudstone interbeds are typical of the upper part of the Strath Rory Group. Two markedly different facies associations are seen in the Balnagown Group – a fluvial association of pebble conglomerate and cross-bedded sandstone forms the Tarbat Ness Formation at the base of the group; above this, a fluvial–aeolian–sabkha association forms the Gaza Formation. The latter contains striking examples of desiccation and evaporitic structures, features associated with sporadic rainfall in an arid environment, and evidence of subaerial animal activity. Overall, Tarbat Ness is a very important site, providing a rare insight into Mid- and Late Devonian environments and habitats at the margin of the Orcadian Basin.

DUN CHIA HILL (LOCH DUNTELCHAIG), INVERNESS-SHIRE (NH 600 285–NH 600 291)

D. Stephenson

Introduction

At the south-west limit of the Orcadian Basin, the outcrop of Middle Old Red Sandstone (MORS) rocks extends down the south-east side of the Great Glen as far as Foyers (Stephenson, 1972, 1977; Mykura, 1982). Within this outcrop 'extension', the basal beds are characterized by coarse breccias, breccio-conglomerates and conglomerates that rest on a highly irregular palaeo-landsurface of Grampian Group (Dalradian) metasedimentary rocks and late Caledonian granitic rocks. The unconformity is well exposed in several places, the most dramatic and instructive being on the south-east side of Dun Chia hill, between Loch Ruthven and Loch Duntelchaig. There, the unconformable junction is exposed continuously for a distance of some 1300 m and the overlying breccio-conglomerates form cliffs from Tom Mor (NH 591 280) in the south-west, through Craig na h-Iolaire (NH 597 282), to the spectacular, 50 m-high Creag nan Clag (Crag of the Bells; NH 601 289) in the north-east (Figure 2.65). The effects of pre-MORS weathering and erosion on the underlying Grampian Group rocks are well seen and the breccio-conglomerates yield much information about sedimentary processes and palaeogeography of the region in Mid-Devonian time. Similar basal beds are also well exposed in cliffs between Creag Dhearg (NH 618 294) and Carn Mor (NH 627 303), 2–3 km to the ENE of Creag nan Clag, but the unconformity is less well-exposed.

The Old Red Sandstone rocks of the area were first described by Wallace (1880), who described the unconformity and the overlying beds at Dun Chia and Creag Dhearg in some detail. Further descriptions and interpretations of these localities were given by Mould (1946) in her account of the nearby Foyers granitic pluton. The most detailed investigation was that of Mykura (1982), which was used in the compilation of the 1:50 000 geological map, Sheet 73E (Foyers) (British Geological Survey, 1996) and is the basis for much of the following account.

Description

Between Loch Duntelchaig and Loch Ness in the Great Glen, there is a cross-strike outcrop width of about 5 km of Middle Old Red Sandstone strata. This represents a succession over 2 km thick that includes coarse and fine breccias, conglomerates, gritty, pebbly feldspathic and fine-grained sandstones, with subordinate mudstones, siltstones and calcareous beds. In general, the lower parts of the succession (up to 900 m) are of breccio-conglomerate and coarse bimodal conglomerate overlain by planar-bedded, medium- to fine-grained sandstones. The succession interdigitates along strike to the north-east with that of the Inverness and Nairnside districts, where fragmentary fish remains and plant spores of Mid-Devonian age have been found (Horne and Hinxman, 1914; Horne, 1923; Fletcher *et al.*, 1996). There is a regional dip of between 10° and 50° to the north-west.

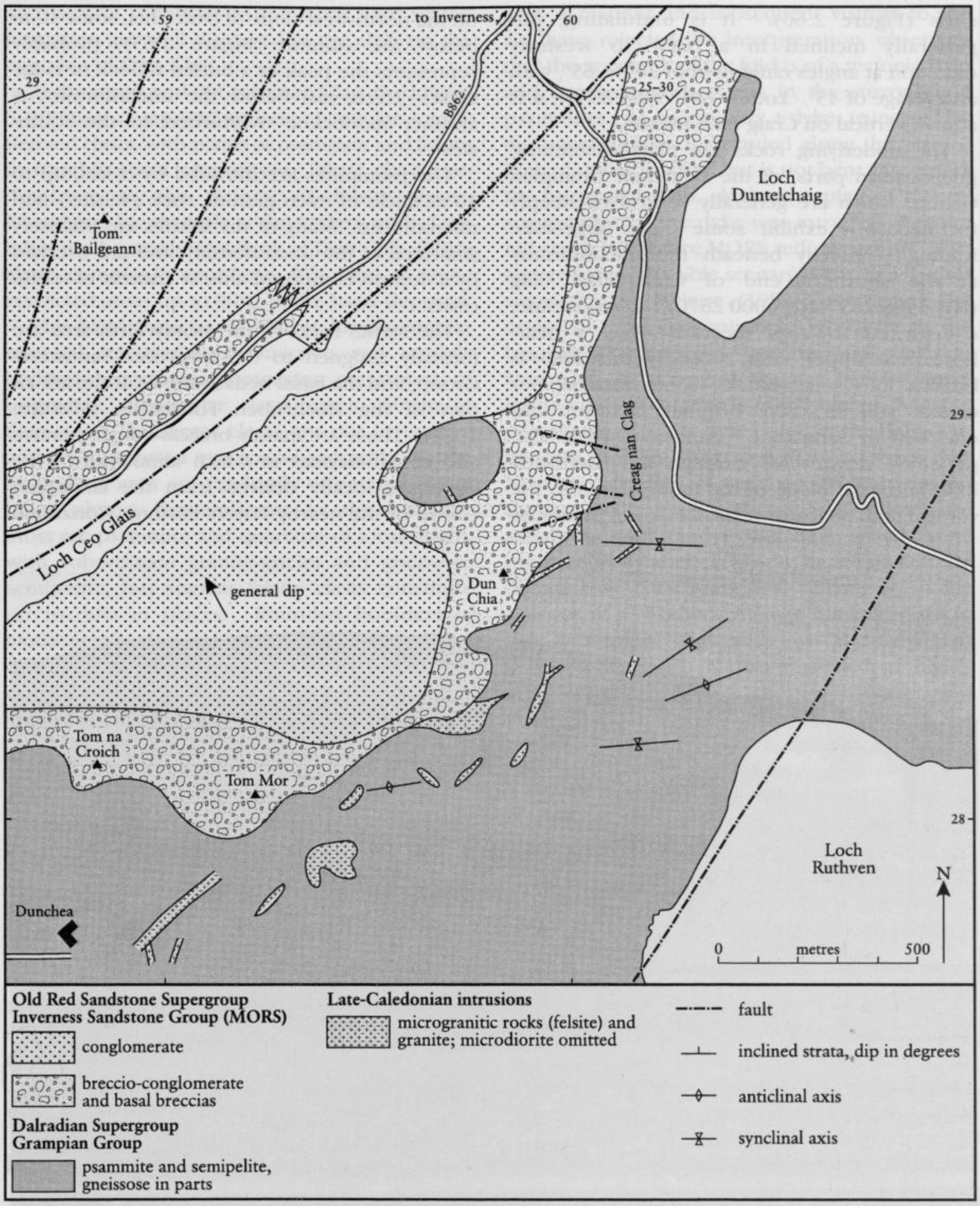


Figure 2.65 Map of the area around Dun Chia hill, at the south-west end of Loch Duntelchaig. After British Geological Survey 1:50 000 Sheet 73W (Scotland), Foyers (1996).

Throughout the area in general, the pre-MORS topography was undulating, with some steep slopes locally that probably reflected active fault scarps (Stephenson, 1972; Mykura, 1982). Within the GCR site, between Tom Mor (NH 591 279) and Creag nan Clag (NH 600 287), the basal unconformity is exposed almost continuously for 1300 m at the foot of steep cliffs (Figure 2.66). It is undulating, but generally inclined in a NNW to westerly direction at angles ranging from 20° to 65°, with an average of 45°. Locally (NH 5985 2835), it is almost vertical on Craig na h-Iolaire.

The underlying rocks are platy psammites, gneissose in parts, of the Dalradian Grampian Group, which are generally vertical or steeply inclined, but exhibit some quite large-scale folding. Directly beneath the unconformity at the southern end of Creag nan Clag (NH 5988 2857–NH 6000 2870), a monocline, overturned towards the west, has a near-horizontal upper limb. Locally this limb is broken into small blocks, some rotated, but others still in their original position and enclosed by sandstone. Sandstone veins also penetrate downwards through the horizontal limb and axial plane of the monocline into the vertical limb beneath. A small inlier of psammite on the north-west slope of Dun Chia, some 40 m below the summit (Figure 2.65), probably represents the peak of a buried hillock or ridge crest. Here, just below the unconformity, is another monocline overturned to the south-west.

The psammites are intruded by dykes, up to 10 m thick, of micrograntic rock ('felsite') and microdiorite. Many of the felsites are intensely shattered at the unconformity and are overlain by a deposit composed almost entirely of felsite clasts.

The MORS rocks of the Moray Firth area are formally assigned to the Inverness Sandstone Group, and the basal beds at the GCR site are all part of the Bochruben Formation (Mykura, 1982). These are mainly breccio-conglomerates and conglomerates in which almost all of the clasts are locally derived from the Grampian Group or the nearby Foyers Pluton. Horizontal

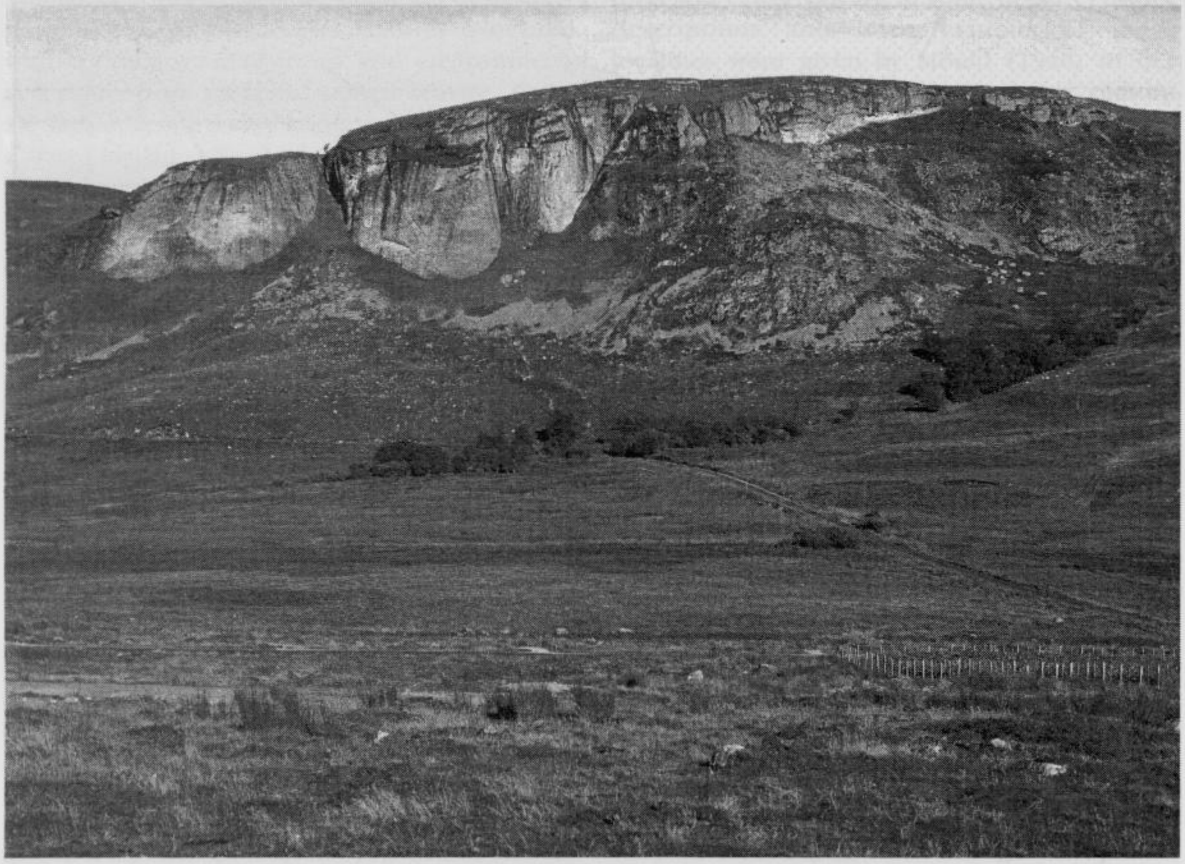


Figure 2.66 The cliffs of Creag nan Clag, on the eastern flank of Dun Chia hill. The undulating unconformity between Middle Old Red Sandstone massive breccio-conglomerates and underlying flaggy psammites of the Grampian Group (Dalradian) can be traced for 1300 m along the base of the steep upper cliffs. (Photo: BGS No. D1813, reproduced with the permission of the Director, British Geological Survey, © NERC.)

and vertical variations in the basal beds of the formation are well displayed in the cliffs on the east side of Dun Chia.

At the south end of Creag nan Clag, the lowest 5–10 m are composed entirely of angular, locally derived clasts, with plates of psammite and felsite in roughly equal proportions; felsite clasts predominate in the vicinity of felsite dykes. Elsewhere, for example at the eastern end of the inlier on the north-west flank of Dun Chia, felsite clasts commonly constitute 40%, and in some restricted areas up to 90%, of the basal deposits.

The basal breccias of almost in-situ angular material are overlain by massive, very crudely bedded breccio-conglomerates composed of ungraded, subangular clasts up to 60 cm in diameter. The larger clasts are not in contact and are supported in a sparse red gritty matrix. Their long axes are randomly orientated, some being almost perpendicular to the bedding. Most are psammite and quartzite of local origin, with subordinate granite-gneiss and gneissose semipelite, but angular felsite clasts constitute up to 30% in places. Rounded boulders of biotite granite also occur sporadically. The breccio-conglomerates vary in thickness, being thickest in the cliffs of Creag nan Clag and thinning to the north-east and south-west, although they still form the greater part of all of the cliff sections.

The north-west flanks of Dun Chia and parts of the summit ridge are composed of bimodal conglomerates with a high proportion of leucogranite clasts. On Craig na h-Iolaire, coarse conglomerates with rounded granite boulders up to 60 cm in diameter overlie Dalradian 'basement' almost directly in places, although residual patches of local breccia only a few metres thick, and some breccio-conglomerates are also present. Sandstone lenses are intercalated locally with the conglomerates, such as those dipping at 15° to the WNW, between 6 m and 10 m above the unconformity at the south-west end of Creag nan Clag.

Interpretation

The prominent monocline in the Dalradian 'basement' at Creag nan Clag has an axial plane almost parallel to the unconformity and was interpreted by Mykura (1982) as flexuring due to hill creep prior to deposition of the MORS sediments. However, this coherent fold is on a remarkably large scale to have been formed in such a manner and subsequent visitors to the site have rejected the interpretation, observing that the geometry of the fold is of a tectonic style and conforms with others in the surrounding psammites. At one locality a dyke, truncated by the unconformity, is intruded along the vertical limb of the monocline but is not bent as it enters the horizontal limb. Mykura explained this by suggesting that the dyke was intruded after the hill creep but before MORS sedimentation. This somewhat implausible scenario does fit a further suggestion by Mykura (1982), concerning the origin of local concentrations of felsite in the basal MORS. Observing that, in several places on Dun Chia and Creag Dearg, the percentage of felsite clasts (40–90%) is much higher than the percentage of felsite dykes seen in the Dalradian beneath, Mykura suggested that the dykes may have been feeders to small extrusive domes on the pre-MORS land surface. Such domes are usually unstable and would have crumbled to breccia shortly after extrusion; hence no definite domes have been identified. However, in the absence of any other evidence for volcanism in the Grampian Highlands in Mid-Devonian time, this scenario is also regarded as highly tenuous.

The local origin of most of the clasts in the basal MORS and the low degree of sorting indicate that many of the deposits were formed quite near to the source of their constituents. Much of the angular material in the basal breccias is virtually in-situ and is in effect a regolith, or, if on a steep original slope, a 'fossil scree'. The lack of movement is particularly noticeable above some felsite dykes where it is difficult to separate the in-situ felsite from 'felsite scree'. This effect is even more characteristic of basal MORS deposits to the south-west of the GCR site, where 'granitic screes' are indistinguishable in places from underlying in-situ granitic rocks (Stephenson, 1972; Mykura, 1982). The freshness of many of the clasts indicates rapid accumulation in a climate in which chemical weathering was minimal. However, oxidation is commonly intense, imparting a red or purple hue to the matrix of the rocks, so climatic conditions during deposition were probably arid or semi-arid.

The breccio-conglomerates that dominate the GCR site merge with the basal breccias and have no marked basal erosion surfaces. They are unstratified and poorly sorted, with mainly subangular clasts of local material 'floating' in a matrix, usually of mud or silt grade. These are the characteristics of debris flows that moved as dense, viscous masses in which waterlogged matrix was able to support and transport clasts up to boulder size. Such deposits form typically in areas of high relief and rapid erosion, in which deposition on steeply sloping alluvial-fans results from occasional sheet floods during sudden violent rainstorms in an otherwise dry area. The thickest development of breccio-conglomerate, in the 50 m-high cliffs of Creag nan Clag, has no horizontal breaks and it is possible that the entire section represents a single debris flow, possibly filling a palaeovalley. Lenses of conglomerate and sandstone within the sequence probably originated as channel-fill deposits within the alluvial fans and some of these have cut down to rest directly on basement in places.

Higher in the sequence, reworking by water becomes apparent as the breccias become slightly better sorted, some bedding is evident and clasts become more rounded. Debris of a less local nature appears, including leucogranite clasts that must have been transported for some distance. Bimodal clast-supported conglomerates, which occur locally as channel-fills within the breccio-conglomerates, become dominant higher in the succession, from the summit of Dun Chia north-westwards. These higher conglomerates are of fluvial origin and probably originated as braided river deposits on the distal parts of fans and on the piedmont plain. Beyond the GCR site, still higher parts of the succession record the eventual burial of the fans by finer-grained sediments that accumulated on a broad floodplain (Mykura, 1982).

By the time of deposition of the MORS sediments, the Grampian Highlands had already experienced considerable rapid uplift at the end of the Caledonian Orogeny. An estimate of the scale of uplift in the northern Grampians can be obtained from petrological evidence in the aureole of the Foyers Pluton, which Marston (1971) suggested had a possible cover of 7 km at its time of emplacement in Early Devonian times. The cover was removed in a maximum of about 20 Ma to expose the pluton in Mid-Devonian time. This represents an erosion rate of at least 3.5 cm per 100 years, comparable with rates in present-day newly formed mountain areas.

The Great Glen Fault was already in existence in Mid-Devonian times and may have been the locus of a major valley that drained north-east and was bounded to the south-east by high mountains (Stephenson, 1972, 1977). The MORS basal sediments were deposited predominantly in coalescing alluvial-fans along the foot of these mountains and in the valley bottom, which widened out north-eastwards into a piedmont plain on the edge of the main Orcadian Basin. The plain probably extended across the Great Glen Fault into the area now occupied by the Black Isle and Tarbat Ness (Mykura, 1982, fig. 11). Close to the Great Glen Fault, in the Foyers–Inverfarigaig area, the highlands to the south-east were probably bounded by major active NE–SW fault scarps, related to the Great Glen Fault and with a component of downthrow to the north-west. However, around Loch Duntelchaig, the edge of the hills probably swung to the ENE; here major fault scarps were probably absent and the boundary zone between plain and mountain may have been a belt of foothills with several small active local fault scarps.

Conclusions

The spectacular cliffs on the south-east flank of Dun Chia hill, at the south-west end of Loch Duntelchaig, provide a representative section of the basal beds of the Middle Old Red Sandstone at the southern limit of the Orcadian Basin. The unconformable junction beneath these beds is very well-exposed almost continuously over a distance of 1300 m and is probably one of the best preserved 'buried landscape' unconformities in Great Britain. It is certainly of national importance and visitors cannot fail to be impressed by the scale of the exposure and by the many exceptional features that are exhibited. Beneath the unconformity, folded Neoproterozoic flaggy metasandstones, cut by felsitic dykes, are disaggregated and shattered close to an undulating former land surface. The overlying angular debris can be matched precisely to the immediately underlying rock in many places and probably accumulated as scree. Poorly sorted mixtures of angular fragments and rounded boulders (breccio-conglomerates) comprise the majority of the section and probably originated as debris flows or sheet floods associated with alluvial fans that debouched from mountainous terrane to the south-east and spread out over a broad piedmont plain to the north on the edge of the alluvial plain of the main Orcadian Basin.

TYNET BURN, MORAY (NJ 383 618)

Potential ORS GCR site

W.J. Barclay and N.H. Trewin

Introduction

Tynet Burn, 5 km north-east of Fochabers, Moray is a classic Middle Old Red Sandstone fossil fish site, discovered in 1838. Since then, it has yielded about 14 species of fish, most of which are found in the upper of two beds of calcareous nodules within laminated shales and limestones deposited in a lake. The fish fauna is correlated with the Middle Devonian (Eifelian) Achanarras Fish Bed in Caithness and other Orcadian Basin equivalents in Orkney and Shetland. The fish-bearing nodule beds alternate with conglomerates and sandstones deposited by rivers on an alluvial plain at the southern margin of the Orcadian Basin (Trewin and Thirlwall, 2002). Periodic transgression of the lake over the alluvial plain resulted in lacustrine sediments being deposited at times of high lake level; the contractions and expansions of the lake were controlled by climatic changes with Milankovitch periodicities. The palaeogeographical and stratigraphical history of the Orcadian Basin has been summarized by Trewin and Thirlwall (2002). Trewin and Davidson (1999) provided details of the sedimentary environments and palaeoecology at Tynet on the southern shore of the Orcadian lake about 387 million years ago.

Description

The geology of the site was described by Peacock *et al.* (1968). Gillen (1987) provided a brief field guide to the locality, and Wood and Norman (1991) described an excavation in the fish bed at Tynet in 1989–1990. Dineley (1999a) and Trewin and Davidson (1999) gave historical details of research at the site, and the latter gave an account of excavations made at the site in 1996, providing the main source for this account. The attractive pink, red and purple colours of the fossil fish, in contrast to the green or beige matrix of the nodules in which they occur, made the fossils very popular, and specimens from Tynet are widely dispersed in museums, university departments and private collections throughout the world.

The site (Figure 2.67) lies in the Tynet Burn below Lower Mills of Tynet (NJ 388 618), where the stream is incised into Old Red Sandstone through a cover of glacial till. The beds belong to the Middle Old Red Sandstone, which comprises about 200 m of red sandstones, conglomerates and mudstones in this area. The fish-bearing succession in the Tynet Burn consists of 28 m of conglomerates, sandstones, mudstones and limestones (Figure 2.68), fault-bounded at the top and bottom. The lowest bed (NJ 3837 6196) is a 0.9 m-thick conglomerate with numerous angular pebbles. The highest beds are faulted against a fining-upward succession of about 120 m of gently dipping beds that are exposed downstream. These comprise a basal red boulder conglomerate 39 m thick passing up through interbedded red sandstones and conglomerates into cross-bedded sandstones with subordinate conglomerate. The Lower Nodule Bed is exposed in the river (NJ 3837 197) just downstream of the 0.9 m-thick conglomerate.

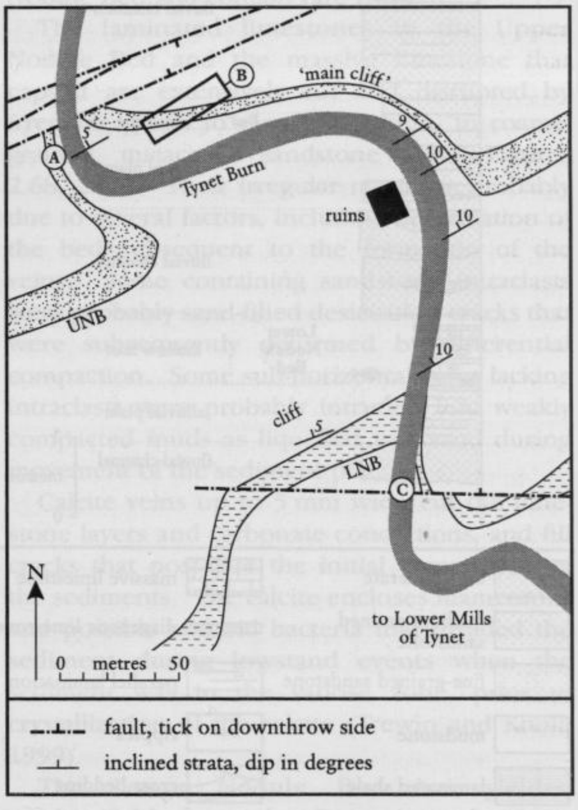


Figure 2.67 Sketch map of Tynet Burn: (A) section cleared in 1996 (Trewin and Davidson, 1999); (B) section excavated in 1989–1990 (Wood and Norman, 1991). Based on Peacock *et al.* (1968) and Davidson and Trewin (1999).

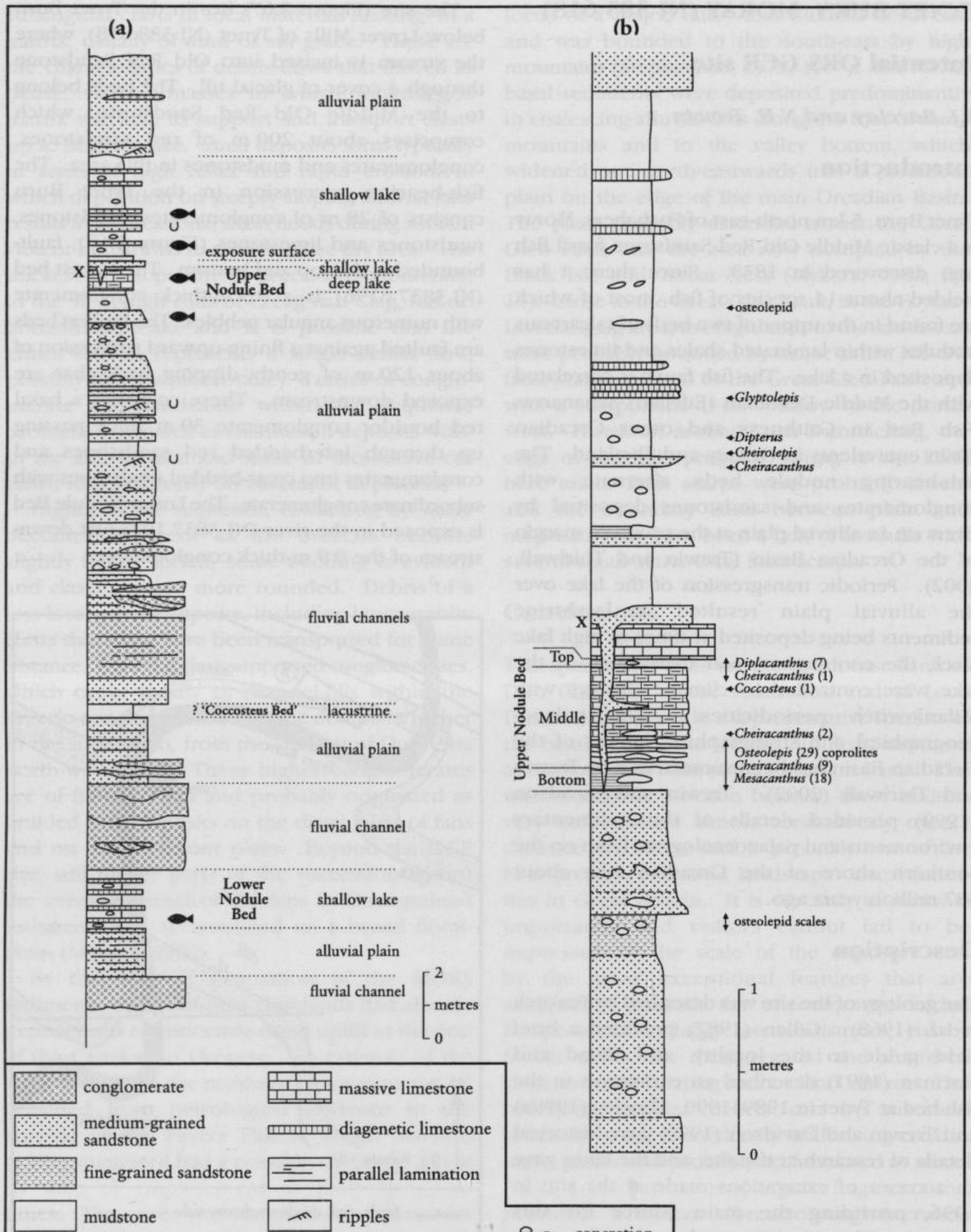


Figure 2.68 (a) Composite log of succession in Tynet Burn between points (A) and (C) on Figure 2.67. The position of the Coccosteus Bed is inferred. (b) Summary log of section excavated at Point (A). After Trewin and Davidson (1999).

The best section was formerly in a 12 m-high cliff (the 'Main Cliff') on the east bank of the stream (NJ 384 620). The higher nodule bed (the Upper Nodule Bed) crops out along the top of this cliff, but the disused quarry is covered by talus and the exposure is obscured. The bed is exposed and more accessible farther downstream in a meander scar on the west bank (NJ 3823 6203) close to the fault that forms the northern limit of the block. Excavations (NJ 3828 6205) near the 'Main Cliff' in 1989 and 1990 (Wood and Norman, 1991) and clearance of the exposure on the west bank (NJ 3823 6203) in 1996 (Trewin and Davidson, 1999) provided detailed sections of the strata here, although soft-sediment extensional and compressional structures, sandstone intrusions and minor tectonization of the beds in proximity to the main fault to the north disrupt the continuity of the beds.

The succession contains two main fish-bearing horizons (the Upper Nodule Bed and Lower Nodule Bed) and a third, intermittent horizon between them (the Coccosteus Bed), the exact position of which is uncertain (Figure 2.68). These horizons represent lake deposition within the fining-upward fluvial to lacustrine cycles. The cycles are truncated by erosion surfaces and commence with conglomerates that contain rounded to subrounded pebbles up to 15 cm in diameter, predominantly of quartzite, but with some of metamorphic and igneous lithologies. The pebbles are set in a red, medium- to coarse-grained sandstone matrix and show sorting and imbrication. Lenses of laminated and cross-bedded sandstones also occur, and there are some cross-bedded conglomerate bodies up to 0.75 m thick.

The sandstones are red with minor patchy reduction zones. They are mainly fine- to medium-grained and micaceous, and occur in beds ranging from 5 cm to 40 cm thick. Parallel lamination is the dominant structure, although most beds are only weakly laminated. A few beds are trough cross-bedded and the tops of some are linguoid current-rippled. The sandstones pass up into brick-red, massive mudstones, with small (up to 10 cm) irregular to subspherical unfossiliferous diagenetic carbonate concretions (or nodules). The mudstones containing the Upper and Lower Nodule beds are red to chocolate-brown and drab grey-green. They are mainly massive to weakly laminated on a millimetre to centimetre scale, but the fish-bearing horizons are well laminated on a sub-millimetre scale. The carbonate concretions occur in several forms within the laminites and are commonly fish-bearing.

There are three types of limestone in the section. A massive white to pale pink limestone with red, hematite-stained limestone intraclasts and small spar-filled cavities forms a prominent bed up to 0.15 m thick immediately above the Upper Nodule Bed in the main exposure (Figures 2.68, 2.69). Secondly, white to dark red, irregularly stained, laminated, platy limestones form the middle part of the Upper Nodule Bed and are interlaminated with siliciclastic laminites, particularly at the top and base of the Upper Nodule Bed. These are typical carbonate, fish-bearing laminites, the fish being preserved in flat concretions as well as in the non-concretionary carbonate laminae. Thirdly, concretions and thin, laterally persistent non-laminated beds of green limestone occur in the mudstone-dominated beds above the Upper Nodule Bed and contain rare fish.

The laminated limestones in the Upper Nodule Bed and the massive limestone that caps it are extensively cut and disrupted by irregular, green to purple, medium- to coarse-grained, micaceous sandstone veins (Figures 2.68, 2.69). Their irregular nature is probably due to several factors, including deformation of the bed subsequent to the formation of the veins. Some containing sandstone intraclasts were probably sand-filled desiccation cracks that were subsequently deformed by differential compaction. Some sub-horizontal veins lacking intraclasts were probably intruded into weakly compacted muds as liquefied wet sand during movement of the sediment pile.

Calcite veins up to 5 mm wide cut the limestone layers and carbonate concretions, and fill cracks that post-date the initial compaction of the sediments. The calcite encloses filamentous and possible coccoid bacteria that invaded the sediment during lowstand events when the sediment was in the vadose zone, prior to crystallization of the calcite (Trewin and Knoll, 1999).

The Lower Nodule Bed has yielded unidentifiable fish scales (Peacock *et al.*, 1968), an articulated osteolepid, scattered osteolepid scales and a *Cheiracanthus* fin spine (Trewin and Davidson, 1999). The Coccosteus Bed has not been traced recently, is probably lenticular,

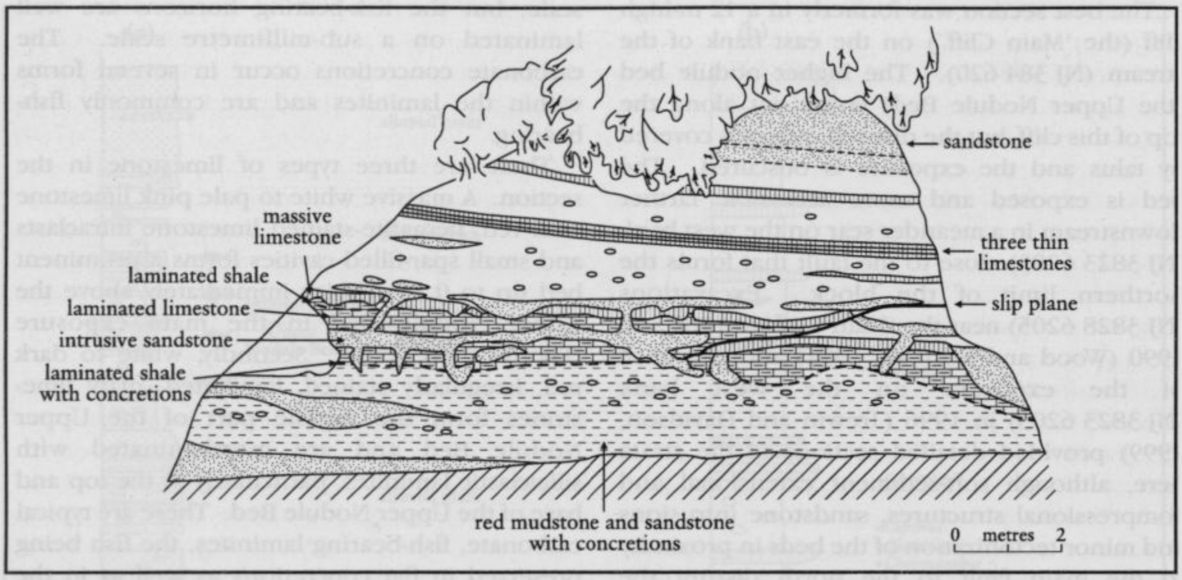


Figure 2.69 Field sketch of exposure at Point A in Tynet Burn showing the disrupted nature of the Upper Nodule Bed. After Trewin and Davidson (1999). This exposure is now largely obscured by talus and vegetation.

and apparently yielded only *Coccosteus* and some scales. *Dipterus* (Malcolmson, 1859) and *Osteolepis* scales (Trewin and Davidson, 1999) have been recorded from the beds below the Upper Nodule Bed, both in sandstones, the latter in small nodules 1 m below the base of the nodule bed.

The Upper Nodule Bed (Figure 2.68b) is the main fish-bearing horizon. Many of the carbonate nodules are shaped like the fish they contain. Detailed sections from excavations were recorded by Wood and Norman (1991) and Trewin and Davidson (1999). Although only 50 m apart, the sections show a marked variation in thickness of the bed. Wood and Norman recorded a thickness of over 2 m, compared to 0.75 m recorded by Trewin and Davidson. The difference may be reconciled by invoking deposition of the bed on a channelled surface of the underlying sands, the irregular topography probably contributing to instability, soft-sediment slumping and disruption of the bed (Figure 2.69).

In detail, the Upper Nodule Bed comprises three horizons (Top, Middle and Bottom). The top and bottom units are chocolate brown, clastic laminites with sporadic fish-bearing concretions. The lower unit ranges from 0.05 m to 0.9 m in thickness and has yielded a fauna dominated by osteolepids but including the acanthodians *Cheiracanthus* and *Mesacanthus*. Acanthodians (*Diplacanthus* and *Cheiracanthus*) dominate in the upper unit, which shows a similar thickness range. The intervening middle laminated limestone unit is 0.5 m thick and has yielded mainly acanthodians.

The complete faunal assemblage from the Upper Nodule Bed is listed by Dineley (1999a) and Trewin and Davidson (1999) as:

Placodermi:
Coccosteus cuspidatus Miller (1841)
Rhamphodopsis trispinatus Watson 1938
Pterichthyodes milleri (Miller 1841)
Acanthodii:
Cheiracanthus murchisoni Agassiz
Cheiracanthus latus Egerton 1861
Diplacanthus crassisimus Duff 1842 (= *D. striatus* Agassiz 1835)
Mesacanthus pusillus Agassiz 1844 (?= *M. peachi* Egerton 1861)
Rhadinacanthus longispinus (Agassiz 1844)
Osteichthyes:
Cheirolepis trailli Agassiz 1844
Glyptolepis leptopterus Agassiz 1844
Glyptolepis paucidens Agassiz 1844 ? (not seen in any collection by R.G. Davidson)
Dipterus valenciennesi Sedgwick & Murchison 1828
Osteolepis macrolepidotus Agassiz 1835
Gyroptychius spp.

The assemblage includes type specimens of *Diplacanthus striatus* and *Rhamphodopsis trispinatus* Watson 1938, and possibly that of *Cheiracanthus latus* Egerton 1861. The material is sufficiently well-preserved to distinguish the scale morphology of several acanthodians (Young, 1995). Wood (in Trewin and Davidson, 1999) carried out a statistical analysis of the assemblage of the Upper Nodule Bed, calculating that acanthodians comprise 82% of the assemblage of the Top Bed and sarcopterygians (mainly *Osteolepis*) 72% of the Bottom Bed. *Osteolepis*?, *Glyptolepis*, *Cheirolepis* and *Cheiracanthus* were collected between 1 m and 2 m above the Upper Nodule Bed by R.G. Davidson.

The preservation of the fish ranges from complete carcasses with full articulation to isolated scales and bones. The best material comes from the nodule beds and limestone laminites of the Upper Nodule Bed, in which the osteolepids at the base and the acanthodians in the Middle and Top beds are commonly articulated or suffered only minor disarticulation before burial. Isolated scales and coprolites also occur in the concretions alongside the whole fishes.

The striking pink, red and purple colour of the fish is the result of oxidation, probably by chemotrophic bacteria (Trewin and Davidson, 1999; Trewin and Knoll, 1999). Zones of iron oxide surround the fish, commonly permeating the scale and bone structure and infilling original canals in the bone. The phosphatic scales are commonly dissolved, leaving calcite-filled cavities. Relics of internal organs and eyes have been recognized in some specimens, represented by dark red to black stains (Davidson and Trewin, 1999, 2002). Gut contents are preserved in one specimen of *Coccosteus*.

Interpretation

Hamilton and Trewin (1988) and Trewin and Davidson (1999) provided a detailed analysis of the sedimentary environments and of the diagenetic processes by which the fish were preserved. Figure 2.68 shows a summary of the range of sedimentary environments in which the succession was deposited.

The conglomerates at the bases of the cycles were the deposits of high-energy, gravel-bed (?braided) stream channels. The overlying mainly parallel-bedded laminated sandstones were probably deposited rapidly as unconfined sheet floods. Some of the finer, thinly bedded sandstones contain *Skolithos* and *Diplocraterion* burrows and may have been deposited in shallow floodplain lakes. Red mudstones with carbonate nodules record formation of incipient caliche soil profiles on the floodplain. On three, or perhaps four occasions, the alluvial plains were rapidly inundated from the north by lake waters, resulting in the deposition of the fish-bearing beds. These lake transgressions record two major lake highstand events, when the Orcadian lake reached its greatest extent and was deep enough for lacustrine laminites to form at Tynet during the deposition of the Upper Nodule Bed.

The massive limestone at the top of the Upper Nodule Bed is interpreted as a shallow-water lake-margin deposit, similar to that seen at the **Red Point** GCR site. Exposure followed its deposition, sand-filled desiccation cracks cutting the limestone.

The Lower Nodule Bed may represent a shallower water and/or shorter-lived lacustrine transgressive event, as may the problematic Coccosteus Bed (discussion in Trewin and Davidson, 1999). Lacustrine conditions were established several times at Tynet for long enough to allow carbonate deposition in low-energy environments in which articulated fish were preserved. The presence of channelized sandstones above some horizons may explain their lateral discontinuity, with fluvial incision removing all or part of the lacustrine interval.

The correlation of the Tynet Burn succession with the Achanarras horizon of Caithness and the Sandwick horizon on Orkney is based on the presence of at least seven species common to both, minor differences in the faunas being explained by ecological factors in their distribution or a slight age difference (Trewin, 1986). If the whole of the Tynet Burn fish-bearing succession is equivalent to the Achanarras Limestone Member, the three cycles here point to a situation more complex than at Achanarras, which was situated in a deeper area of the lake. The marginal position of the Tynet Burn locality, combined with greater clastic influence, made it more sensitive to lake level fluctuation.

The preservation of the fish is summarized by Trewin and Davidson (1999, fig. 12). The differentiation of their post-mortal and pre-burial disarticulation is made difficult because of soft-sediment deformation, sand mobilization, synsedimentary faulting and calcite veining. Also, the carbonate nodules were not entirely nucleated on the fish, and where they are only partially enclosed, post-burial differential compaction and movement resulted in the in-situ breakage of the fish fossils.

Covering of fish carcasses by lake-bed sediment and subsequent compaction was followed by wet sediment deformation and disruption, probably aided by its position overlying an irregular lake bottom. A drop in lake level exposed the lake bed, and desiccation resulted in cracking of the semi-consolidated sediment and infill of the cracks by sand. Further compaction and formation of the carbonate concretions took place in the vadose zone, followed by colonization of fractures in the nodules by chemotrophic bacteria. The selective deposition of iron oxide by the bacteria at haemoglobin-rich organ sites may be the cause of preservation of these soft parts as dark-stained traces.

Conclusions

Tynet Burn is a classic Middle Old Red Sandstone fossil fish site, having yielded a rich fauna of acanthodians, placoderms and bony fishes since its discovery in 1838. The importance of the site lies in the remarkable preservation of the fish in carbonate nodules. The pink, red and purple colours of the fish, in contrast to the green or beige matrix of the nodules, have made the fossils very popular and specimens are widely dispersed throughout the world.

The fish beds were deposited as part of a cyclic fluvial and lacustrine succession on the southern margins of the Mid-Devonian Orcadian lake. The cyclicity was controlled by climatic fluctuations that influenced expansion and contraction of the lake. At times of lake expansion the alluvial plain on the shore of the lake was rapidly inundated by lake waters, which deposited laminated muds and carbonates. Bacterial oxidization in the subsurface preserved the fish in fine detail, including in some cases, traces of soft tissue. The material collected continues to provide new insights into the morphology and diet of Devonian freshwater fish and their lake habitat.

DEN OF FINDON, GAMRIE BAY AND NEW ABERDOUR, ABERDEENSHIRE (NJ 796 635–NJ 882 650)

Potential ORS GCR site

W.J. Barclay

Introduction

The Den of Findon near Gamrie, Banffshire has been independently selected for the GCR for fossil fishes (Dineley and Metcalf, 1999). It is the easternmost occurrence of the Mid-Devonian Achanarras fauna and has yielded 12 species of fossil fishes. The fish bed (the Gamrie Fish Bed) lies within the upper part of the Findon Group of the Middle Old Red Sandstone and the site is important historically as the first prolific Scottish Old Red Sandstone fish site to be exploited. The section is now rather inaccessible and poorly exposed in a steep, overgrown ravine and no collection or excavation is permitted, but the opportunity is taken to describe briefly the important Old Red Sandstone coast sections nearby at Gamrie, and farther east at New Aberdour (Figure 2.70). The following account summarizes that of Dineley (in Dineley and Metcalf, 1999), Sweet (1985), Trewin and Kneller (1987a–c) and Trewin (1987a).

Description

Dineley (1999a and references therein) gave a detailed account of the discovery and history of fish discoveries at the Den of Findon. The fish bed lies up to about 17 m above the base of the Findon Group. The basal unit of the group is a conglomerate comprising rounded clasts of quartzite, felsite and local Dalradian lithologies. Within 0.15 m, the conglomerate fines up into red clay, above which are 1.25 m of grey, laminated mudstones containing fish-bearing calcareous concretions. The parts of the fish within the nodules are preserved complete or with only slight disturbance. The species recovered here include the acanthodians *Diplacanthus striatus*, *D. tenuistriatus*, *D. (Rhadinacanthus) longispinus*, *Cheiracanthus murchisoni*, and *C. latus*; the placoderms *Pterichthyodes milleri* and *Coccosteus cuspidatus*; the actinopterygian *Cheirolepis trailli*; and the osteolepids *Glyptolepis leptopterus*, *Osteolepis*

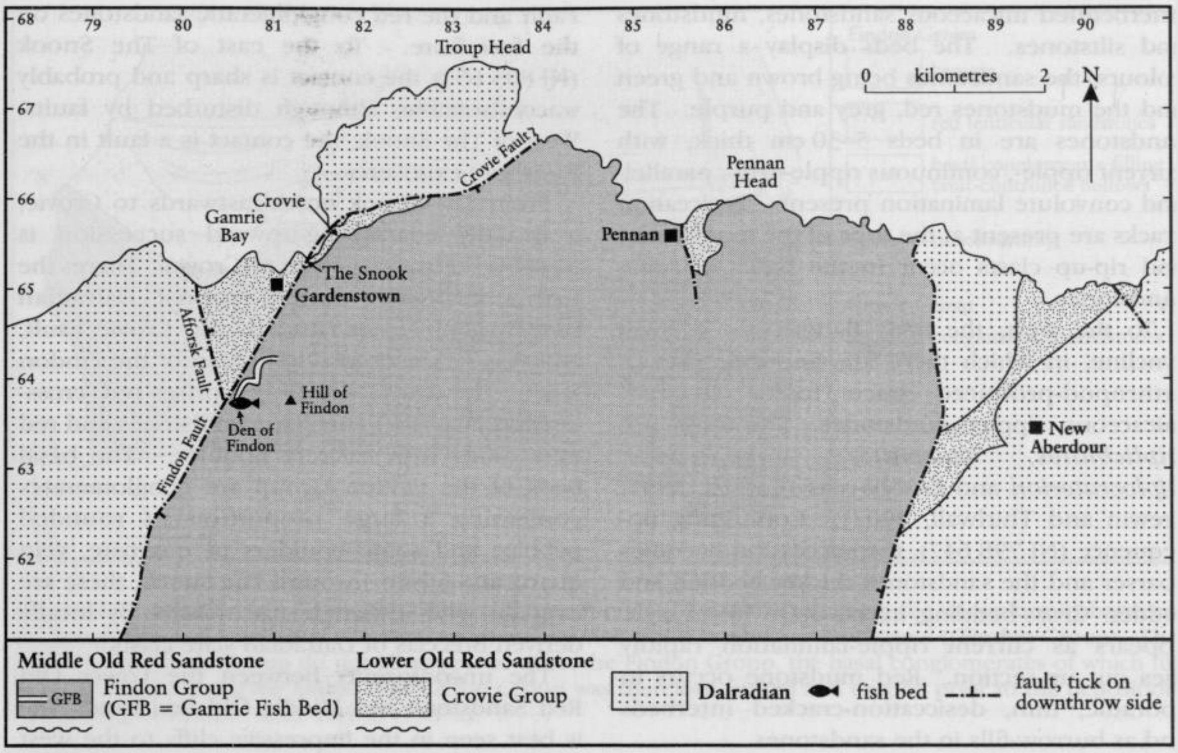


Figure 2.70 Geological map of the Gamrie–New Aberdour area. Based on Institute of Geological Sciences 1:50 000 Sheet 96 (Scotland), Banff (1955) and British Geological Survey 1:50 000 Sheet 97 (Scotland), Fraserburgh (1987).

macrolepidotus and *Gyroptychius* n. spp.. Red clays and shales (0.6 m thick) overlying the laminites are truncated by an erosion surface at the base of a breccia/conglomerate of local Dalradian clasts.

The site lies at the northern end of the Turriff Basin, the fill of which comprises Lower Old Red Sandstone strata (the Crovie Group) resting unconformably on Dalradian basement and unconformably overlain by Middle Old Red Sandstone (the Findon Group). In Gamrie Bay to the north of the Den of Findon, the Afforsk Fault forms the western margin of the basin (Figure 2.70). Conglomerates in the Crovie Group to the east of the fault (NJ 7922 6448) contain locally derived clasts of Dalradian Southern Highland Group lithologies, but also further-travelled clasts of hornblende schist and cleaved greywackes.

A succession of sandstones and conglomerates of the Crovie Group is exposed along the foreshore and in the sea cliffs of Gardenstown between the Afforsk Fault and the Findon Fault to the east. Cut by numerous small faults, the succession is a fine example of a coarsening-upward fluvial sequence, with mudstones and thin sandstones at the base and conglomeratic sandstones at the top (Trewin and Kneller, 1987a). The sandstones are markedly porous and permeable; their red colour is due to the breakdown of large quantities of detrital biotite (Archer, 1978; Donovan *et al.*, 1978). The lowest part of the succession (NJ 802 651), seen on the wave-cut platform between the Findon Fault and the breakwater at the east side of the harbour, comprises thinly bedded grey-green sandstones and mudstones. The sandstones are generally less than 5 cm thick, with parallel lamination and ripple-lamination, as well as small load structures and disrupted beds with detached sandstone balls. Calcrete nodules occur in the mudstones and polygonal desiccation cracks are common at the lowest level exposed, and also near the harbour wall, where the sandstones are thicker and contain more current ripples and ripple-drift lamination.

Immediately west of the harbour wall (NJ 799 648), the succession continues with interbedded micaceous sandstones, mudstones and siltstones. The beds display a range of colours, the sandstones being brown and green and the mudstones red, grey and purple. The sandstones are in beds 5–30 cm thick, with current ripple-, continuous ripple-drift-, parallel- and convolute lamination present. Desiccation cracks are present at the tops of the mudstones, and rip-up clasts occur in the bases of some sandstones.

To the west, the beds flatten into a small syncline, in which there are fine examples of arthropod-produced trace fossils in red, micaceous, rippled sandstones. The traces are *Diplichnites*, *Beaconites*, *Isopodichnus*, *Diplocraterion* and *Rusophycus* (Carroll, 1991; Trewin and Thirlwall, 2002). Continuing up-sequence (NJ 798 647), the succession becomes coarser and the sandstones thicker bedded and redder. Cross-bedding, in sets up to 50 cm high, appears as current ripple-lamination rapidly dies out up-section. Red mudstone occurs in sporadic, thin, desiccation-cracked interbeds and as burrow-fills in the sandstones.

Where small pebbles appear first, they lie mainly in the bases of small channelized, medium-grained sandstone bodies (NJ 796 646), with mudstone interbeds virtually absent. The succession continues to coarsen upwards from the last house on the harbour wall to the cliffs (NJ 7940 6445), with conglomerate lenses lining the bottoms of channels up to 2 m deep. Cross-bedded and parallel-bedded sandstones completing the channel-fills are variously bright red and green as a result of a complex diagenetic history of oxidation and reduction. Dark brown to black zones in some paler sandstones are rich in vanadium. A small fault in the cliff east of the burn mouth (NJ 7925 6445) brings down a conglomeratic sandstone containing angular blocks of red sandstone.

East of the Findon Fault, the Crovie Group outcrops on the foreshore to Crovie and is unconformably overlain by basal conglomerates of the Middle Old Red Sandstone Findon Group. The outcrop of the Crovie Group is much faulted, and Trewin and Kneller (1987a) made no attempt at a correlation of the succession with that to the west of the Findon Fault, as had been proposed by Read (1923) and Donovan *et al.* (1978) (Trewin and Thirlwall, 2002). There is a complex relationship between the conglomerates of the Findon Group east of the Findon Fault and the red conglomeratic sandstones on the foreshore. To the east of The Snook (NJ 803 651) the contact is sharp and probably unconformable, although disturbed by faults. West of The Snook, the contact is a fault in the foreshore exposures.

From The Snook north-eastwards to Crovie, a broadly coarsening-upward succession is traversed down-sequence to Crovie. There, the Crovie Group is faulted against Dalradian basement of Troup Head by the Crovie Fault, which is probably a continuation of the Findon Fault. The Crovie Group comprises red, cross-bedded, mainly channelized sandstones and red mudstones with calcrete nodules. The basal beds of the Findon Group are conglomerates containing a large proportion of rounded pebbles and some boulders of quartzite, vein quartz and felsite. Around The Snook, these are overlain with low-angle discordance by locally derived breccias of Dalradian slate clasts.

The unconformity between the Lower Old Red Sandstone and Middle Old Red Sandstone is best seen in the impressive cliffs to the west of Pennan (NJ 846 655) (Trewin, 1987a). Deposition of the Crovie Group appears to have been controlled by synsedimentary faulting. Re-activation of these faults and erosion before deposition of the Findon Group produced a complex unconformity, the basal conglomerate of the group draping and filling hollows in the underlying eroded surface (Figure 2.71).

At New Aberdour, the Crovie Group is represented by a westward-dipping succession of conglomerates, sandstone and siltstone over 400 m thick and divisible into a lower sandstone–conglomerate unit and an upper mudstone-dominated unit with sandstone lenses in its upper part (Sweet, 1985; Trewin and Kneller, 1987b). The unconformity at the base of the succession is magnificently exposed at the east end of the section (NJ 898 652), where coarse breccia at the base of the Crovie Group rests on an eroded surface of Dalradian andalusite schists and psammites. The unconformity surface dips from 45° to near-vertical, the breccia mantling a highly irregular topography in which resistant greywacke ridges protruded. The breccia is up to about 10 m thick and consists mainly of locally derived psammitic clasts along with some vein quartz and felsite. Westwards, the breccia is absent (NJ 897 651)

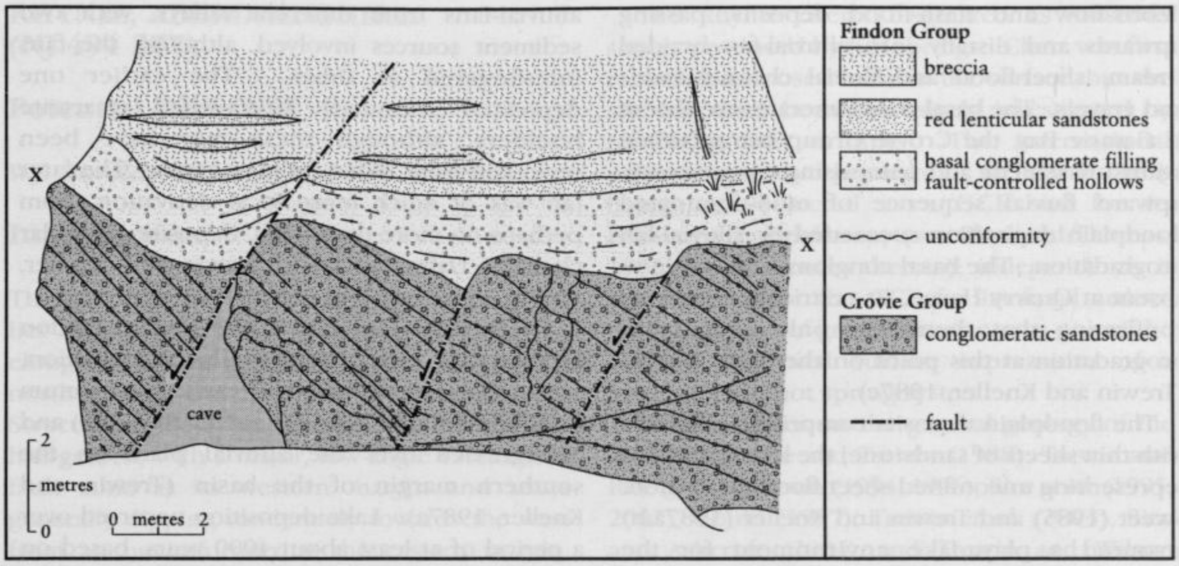


Figure 2.71 Sketch of the lower part of the cliff near Pennan (NJ 842 658) showing the unconformity between the Lower Old Red Sandstone (Crovie Group) and overlying Middle Old Red Sandstone ((Findon Group). The faults in the Crovie Group do not all appear to affect the Findon Group, the basal conglomerates of which fill hollows eroded along the faults. The Crovie Group was therefore faulted and eroded prior to the deposition of the conglomerates. After Trewin (1987a).

and sandstones with conglomerate lenses and granitic debris of angular quartz and feldspar grains overlie the unconformity. Conglomeratic red sandstones rich in granitic debris dominate the succession up-sequence until a marked, but punctuated change into fine-grained sandstones (NJ 894 649). These fine upwards and pass up into red and green mudstones with calcrete nodules, which locally coalesce to form thin, continuous beds (Stage III calcrete) with pseudo-anticlinal structures. Polygonal arrays of desiccation cracks appear to have controlled the development of the calcrete nodules. Thin sheets of laminated and ripple-laminated sandstone interrupt the mud-prone succession.

The fine-grained succession is seen to a point on the wave-cut platform (NJ 839 649) where thin, coarse-grained sandstones appear again and persist upwards. A remarkable feature is the presence of pink felsite pebbles and boulders up to 60 cm in a sandstone or mudstone matrix. They were clearly derived from nearby outcrop and may have slid gently onto the mud- or sand-flat. Another striking feature of this part of the succession is the abundance of *Beaconites* burrows. The succession is terminated by the Dundarg Fault, which throws down to the east, and a similar succession of conglomeratic sandstones (NJ 889 647) and mudstone-dominated beds (NJ 886 648–NJ 882 650) is seen to its west.

Close to the east of New Aberdour, exposures on the east side of Quarry Haven (NJ 9081 6578) reveal a small outlier of Old Red Sandstone rocks, faulted against Dalradian schists and psammites on their western margin and unconformably overlying them to the east (Trewin and Kneller, 1987c). In contrast to New Aberdour, there is no thick development of basal conglomerate facies, the mudstone facies lying much closer to the unconformity.

Interpretation

The Turriff Basin developed in Early Devonian times, its Lower Old Red Sandstone alluvial fill (the Crovie Group) comprising conglomerates and sandstones. Basal breccias mantled the irregular land surface of Dalradian meta-sedimentary basement, infilling valleys and hollows and forming accumulations at the faulted basin margins. Sweet (1985) gave a detailed interpretation of the facies at New Aberdour. The basal conglomerates there filled two SSW-draining palaeovalleys. They represent debris-flow and flash-flood deposits, passing upwards and distally into alluvial-fan braided stream, sheet-flood and fluvial channel sands and gravels. The basal conglomerates are absent at Gamrie Bay, the Crovie Group being faulted against basement and comprising a coarsening-upward fluvial sequence of mud-dominated floodplain deposition succeeded by alluvial-fan progradation. The basal conglomerate is thin to absent at Quarry Haven, its restricted clast suite confirming that there was only limited fan progradation at this point on the basin margin (Trewin and Kneller, 1987c).

The floodplain deposits comprise mudstones with thin sheets of sandstone, the latter probably representing unconfined sheet-flood deposition. Sweet (1985) and Trewin and Kneller (1987a,b) favoured a playa-lake environment for the mudstone deposition, impermanent, shallow lakes forming after flood events. Periods of exposure and low rates of aggradation produced desiccation cracking of the mudflats, with prolonged carbonate soil formation producing calcrete nodules, and in some cases, mature calcrete horizons. A semi-arid, hot climate with seasonal rainfall is inferred. The impressive range of burrows and trails at Gamrie Bay were probably made mainly by arthropods. The mudflats were succeeded by sandy alluvial plains which were combed by rapidly migrating stream channels.

Spores from the Crovie Group at Gamrie Bay are of late Early Devonian (Pragian–Emsian) age (Westoll, 1977). Correlation of the sequences on both sides of the Findon–Crovie Fault in Gamrie Bay was attempted by Read (1923) and Archer (in Donovan *et al.*, 1978). Similarly Sweet (1985) correlated the New Aberdour and Quarry Haven sections. Trewin and Kneller (1987a) adopted a more cautious approach, pointing out that rapid facies variations in the locally derived sediments in the basal part of the succession (probably controlled by synsedimentary faulting), and numerous faults between The Snook and Crovie at New Aberdour make correlation tentative.

Inversion and erosion of the basin-fill followed deposition of the Lower Old Red Sandstone (Crovie Group), represented by a major unconformity overlain by the Middle Old Red Sandstone (Findon Group), spectacularly seen at Pennan. The basal conglomerates and breccias seen at The Snook in Gamrie Bay represent the progradation of two discrete alluvial-fans from different valleys, with two sediment sources involved, although the fans interfingered at times. The earlier one deposited rounded, far-travelled quartzite boulders, although these may have been reworked from older conglomerates. The later fan was of much more local derivation, from perhaps no more than 2 km, depositing angular clasts of Dalradian slate (Trewin and Kneller, 1987a).

The Gamrie Fish Bed at the Den of Findon represents an interruption to alluvial deposition, when the Orcadian lake reached maximum development (at the end of the Eifelian Age) and transgressed over the alluvial plain on the southern margin of the basin (Trewin and Kneller, 1987a). Lake deposition occurred over a period of at least about 4000 years, based on the interpretation of the fine laminations as annual, seasonally controlled varves. The bed is correlated on its fish fauna with the Tynet Burn fish bed (see **Tynet Burn** GCR site report this chapter), the Achanarras Fish Bed of Caithness and the Sandwick Fish Bed of Orkney. There are differences, however, between Gamrie and the other sites, with, for example, *Dipterus*, which is common elsewhere, being absent. Subsequent uplift and erosion took place prior to deposition of the locally derived fan breccia seen at the Den of Findon. The fish bed was probably removed by erosion in Gamrie Bay and Pennan, where the basal Findon Group breccia rests with angular discordance on the conglomerates at the top of the Crovie Group (Dineley, 1999a).

Conclusions

The Gamrie Fish Bed is historically important as one of the first prolific Scottish Old Red Sandstone fossil fish sites. The Den of Findon GCR site is overgrown, but magnificent cliff and foreshore sections nearby in Gamrie Bay, and to the east at Pennan and New Aberdour provide the best sections available of the Turriff Basin. Lower Old Red Sandstone and Middle Old Red Sandstone successions, and the intervening unconformity, are completely exposed, allowing detailed analysis of the sedimentary rocks and the environments in which they were deposited. Of particular importance are the trace fossil assemblages, the variations in facies and the bounding faults with the Dalradian, all spectacularly displayed in these sections.

RHYNIE, ABERDEENSHIRE (NJ 494 277)

Potential ORS GCR site

W.J. Barclay, P. Stone and N.H. Trewin

Introduction

The Rhynie outlier of Old Red Sandstone strata lies about 50 km WNW of Aberdeen. It has an elongate outcrop measuring about 21 km from north to south and up to 3 km from east to west. Structurally, the inlier has been interpreted as a half-graben, with a major, low-angle extensional fault zone at its western margin and strata generally dipping moderately towards the west from an unconformity at the eastern margin. However, recent re-mapping of the northern part of the basin by Rice and Ashcroft (2004) has demonstrated that faulting and folding can be related to basin formation in an Early Devonian regional strike-slip system. The site lies close to the north of the village of Rhynie (Figure 2.72) on the western margin of the basin. Natural exposure within the site is confined to some tuffaceous sandstone, but the site is an established palaeontological GCR site for the Rhynie cherts and its biota, which comprises the best-preserved and most diverse early terrestrial/freshwater ecosystem in the world. The diverse biota was silicified and exceptionally well-preserved by hot spring activity. It includes plants, together with algae, fungi and cyanobacteria, as well as a number of species of terrestrial and freshwater arthropods. Many of the plants and arthropods are unique to this site. It is a world-renowned lagerstätte that has been of crucial importance in providing insights into plant and arthropod evolution. The plants include *Rhynia*, the type-genus of the Rhyniophytina and widely regarded as the archetypal early land plant, and *Asteroxylon*, the earliest well-documented lycopsid. The arthropods include terrestrial opilionid (harvestman) spiders, trigonotarbids, mites, hexapods, euthycarcinoids, centipedes, freshwater crustacea, and several other arthropods of uncertain affinities. A recent re-examination of *Rhyniognatha hirsti*, a fragmentary fossil from Rhynie, suggests that it may have been not only the earliest true insect, but also the first winged species (Engel and Grimaldi, 2004). *Leverhulmia* has also been re-interpreted as an insect (Fayers and Trewin, in press).

The Rhynie cherts originated as siliceous sinters produced by hot springs. Chert was first recorded as loose blocks at the surface, but the site has since been extensively investigated by trenching and drilling. The international palaeobotanical importance of the site led to it being also selected as a GCR site for its Palaeozoic palaeobotany (see Cleal and Thomas, 1995 and references therein); independently it was selected for the GCR for its fossil arthropods. It is also of great importance as one of the earliest preserved surface expressions of a hydrothermal hot spring system.

Recent studies on the general geology of the area are by Rice and Trewin (1988), Trewin and Rice (1992), Trewin (1994, 1996), Rice *et al.* (1995, 2002), Gould (1997), Trewin *et al.* (2003), Rice and Ashcroft (2004) and Trewin and Wilson (2004). A general account of the Old Red Sandstone is provided by Trewin and Thirlwall (2002). Freshwater and terrestrial arthropod faunas recovered from the nearby Windyfield Chert are described by Anderson and Trewin (2003) and Fayers and Trewin (2004). Recent palaeobotanical discoveries include the recognition of gametophytes of some plants (Remy *et al.*, 1993; Kerp *et al.*, 2004), the Zosterophyll plants *Trichopherophyton* (Lyon and Edwards, 1991) and *Ventarura* (Powell *et al.*, 2000a), a parasitic relationship between fungi and green alga (Taylor *et al.*, 1992a,b), ascomycete fungi (Taylor *et al.*, 1999), and the earliest known lichen (Taylor *et al.*, 1997).

The fossiliferous horizons are placed into a context of the evolving geological environments at the margins of a small, active half-graben. The importance of the site lies in the in-situ biota that provides a unique 'snapshot in time' of an Early Devonian terrestrial ecosystem.

Description

The GCR site spans an area of land adjacent to the road from Rhynie to Cabrach (Figure 2.72). Several patches of the distinctive chert have been recorded as float, but there are no natural chert exposures, and the stratigraphical evidence for the position and relationships of the cherts has been obtained from geophysical studies, trenching and drilling. The drilled core and trench samples are housed at the University of Aberdeen. The chert occurs in the highest part of the succession (the Rhynie Chert Member of Gould, 1997; Rhynie Cherts Unit of Rice *et al.*,

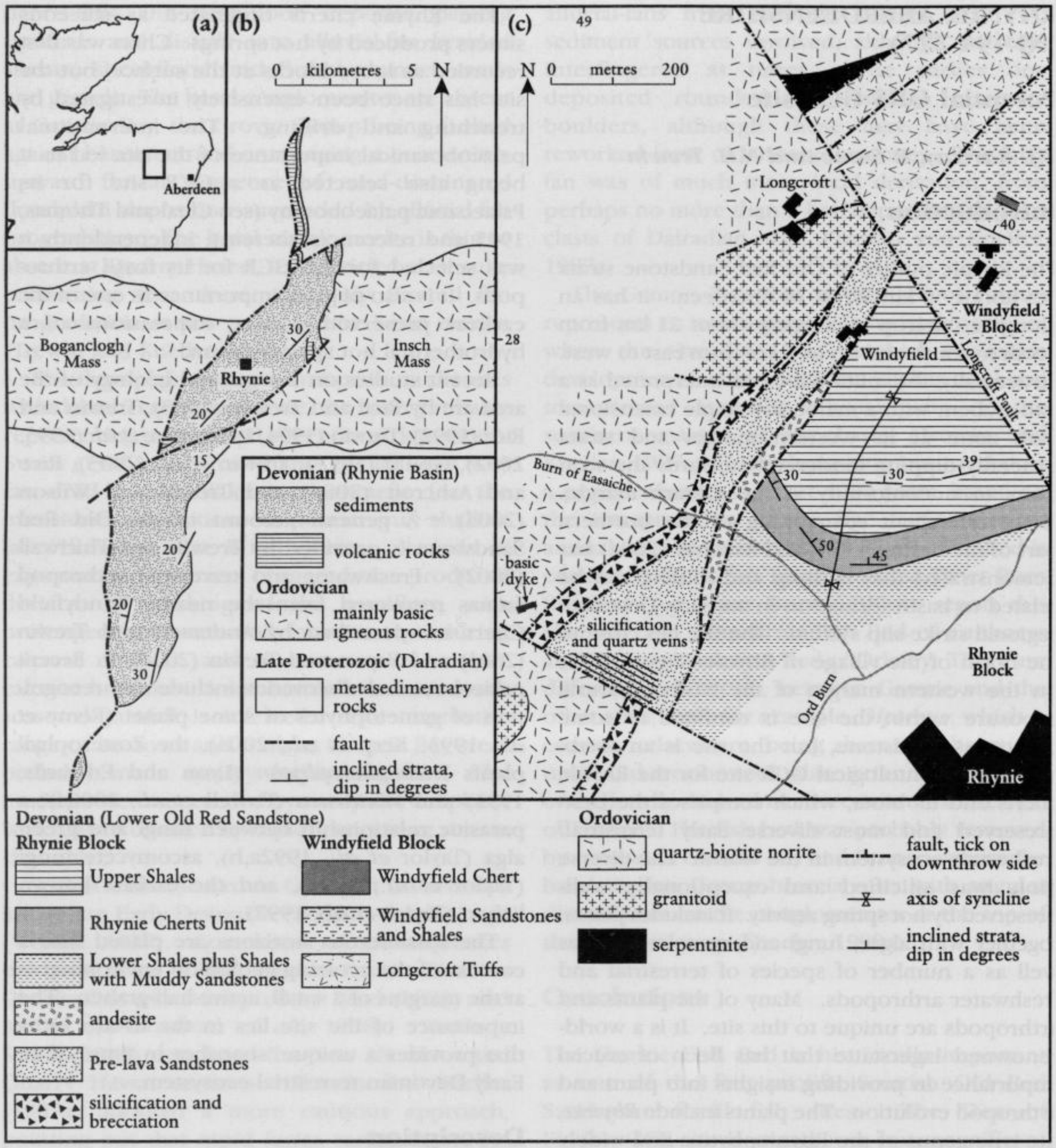


Figure 2.72 Geological map of the area of the Rhynie GCR site. Inset maps show location of the figure. After Rice *et al.* (2002). See Rice and Ashcroft (2004) for a new structural interpretation of the northern part of the basin.

2002) in the Rhynie outlier, in a sequence of shales and tuffaceous sandstones comprising the Dryden Flags Formation (Gould, 1997; Rice *et al.*, 2002). The GCR site lies in the Rhynie Block, the southern of two chert-bearing blocks that are separated by the Longcroft Fault (Figure 2.72). The northern (Windyfield) Block lies beyond the boundary of the palaeobotanical GCR site. In this GCR site however, a basin-margin faulted succession older than the cherts comprises about 20 m of lava and 30 m of sandstone and conglomerate. Most of the Old Red Sandstone succession seen elsewhere in the outlier is absent in the palaeobotanical GCR site area. Recently discovered altered lapilli tuffs east of the Longcroft Fault (the Longcroft Tuffs) are unique to the Rhynie Basin and other Devonian basins in north-east Scotland (Rice *et al.*, 2002). Palynological evidence (Richardson, 1967) suggested a Pragian age for the cherts; Wellman (2004), using borehole samples, has placed the Rhynie succession within the *polygonalis–emsiensis* Spore Assemblage Biozone, giving an early (but not earliest) Pragian to earliest Emsian age. Radiometric (Ar–Ar) dating has given an age of 396 ± 12 Ma (Rice *et al.*, 1995); the Milton of Noth andesite (Figure 2.73) has yielded an ID-TIMS U–Pb zircon-titanite age of 409.6 ± 1.1 Ma that is probably close to the age of the Rhynie Cherts Unit (Parry, 2004), and consistent with the Pragian age based on palynology.

In the Rhynie area the basement to the Old Red Sandstone succession consists mainly of basic igneous rocks (quartz-biotite norite and minor serpentinite) and some granitoids belonging to the Ordovician Boganclogh intrusion. These are separated by an extensional, basin-margin fault zone from the overlying Old Red Sandstone sedimentary and volcanic rocks. The basin-margin fault zones in the Rhynie and Windyfield blocks comprise faulted slices of the basin-fill, as yet uncorrelated between blocks, although Rice *et al.* (2002) estimate that the Longcroft Fault throws down to the north-east by about 160 m, with chert outcrops in the Windyfield Block lying at a higher level than the Rhynie cherts.

The basal part of the succession in the basin-margin fault zone in the Rhynie Block comprises 30 m of mostly lithic sandstone, pebbly sandstone and conglomerate (Figure 2.73), named the 'Pre-lava Sandstones' by Trewin and Rice (1992) and correlated by Gould (1997) and Rice *et al.* (2002) with the Tillybrachty Sandstone Formation of the southern and eastern parts of the Rhynie Basin. The succession fines upwards, and the sandstones are mostly massive, with some parallel lamination and cross-bedding. The abundant pebbles are locally derived from the underlying igneous rocks, some being up to 10 cm in diameter. A calcareous cement is widespread, some sandstones are silicified, and rare calcrete nodules are present (Trewin and Rice, 1992).

A faulted sliver of purple vesicular, andesitic lava separates the Pre-lava Sandstones from the overlying Dryden Flags Formation. This formation is subdivided informally by Rice *et al.* (2002) into five units (Figure 2.73). It is at least 200 m thick and consists mainly of shales and sandstones with minor cherts. Over 40 m of clean, white, parallel- and ripple-laminated sandstones with some cross-bedding (White Sandstones unit) lie at the base of the formation. Beds have sharp bases resting on erosion surfaces, contain large mudstone rip-up clasts and some pebble beds. Mudstone interbeds become commoner upwards and the unit passes into a thick succession of argillaceous sandstones and

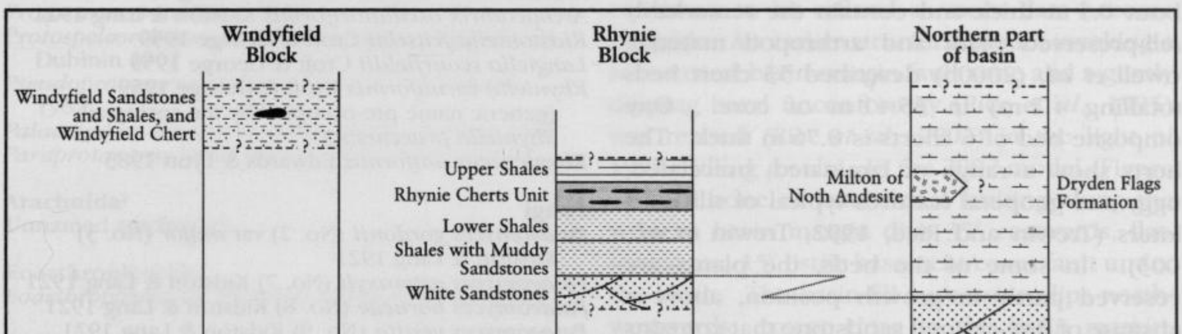


Figure 2.73 Generalized stratigraphy and basin correlations of the Rhynie Basin. After Rice *et al.* (2002).

mudstones (Shales with Muddy Sandstones unit). The argillaceous sandstones are massive, dark, up to 1 m thick and contain numerous small rip-up clasts. There are also some clean, pale sandstones with sharp bases, parallel- and ripple-lamination, and some pebble beds. Interbeds of laminated mudstone are commonly disrupted and locally burrowed. This unit is overlain by laminated mudstones with thin sandstones, subdivided into lower (Lower Shales) and upper (Upper Shales) units that are separated by the Rhynie Cherts Unit. The Lower Shales unit comprises green to blue-black, laminated, locally burrowed mudstone and fine-grained, pale sandstones. The latter occur in beds up to 0.15 m thick that have basal erosion surfaces and are parallel- and ripple-laminated. They contain rip-up clasts, and there are also some beds of pebbles of granitic basement and volcanic rocks. The Upper Shales unit comprises graded green mudstone and siltstone laminae in couplets up to 10 mm, interbedded with sandstones up to 0.1 m thick. The sandstones are similar to those in the Lower Shales, but in addition contain a few desiccation cracks, carbonaceous plant debris and patchy calcite cement.

The Rhynie Cherts Unit (Rhynie Chert Member of British Geological Survey, 1993 and Gould, 1997) is 35 m thick and contains the Rhynie cherts in a succession of interbedded mudstones, carbonaceous sandstones and minor tuffs. Individual chert beds range up to about 0.5 m thick and contain the remarkably well-preserved plant and arthropod material. Powell *et al.* (2000b) described 53 chert beds (totalling 4.2 m) in 35.41 m of core. One composite bed of 6 cherts is 0.76 m thick. The cherts show a range of laminated, brecciated, vuggy and geopetal textures typical of siliceous sinters (Trewin and Rice, 1992; Trewin *et al.*, 2003). In some of the beds, the plants are preserved partly in growth position, above a substrate of fine-grained sandstone that contains plant rhizomes. Trewin and Wilson (2004) have shown that whereas the chert-bearing unit can be correlated over 45 m to 65 m between three boreholes, individual chert beds are laterally impersistent and there is strong lateral variation in the flora. The published fauna and flora of the site are listed below. Several recent finds, including a nematode worm, a large trigonotarbid, an eoarthropleurid and possible spider remains, await publication (H. Kerp, S.F. Fayers, pers. comm.).

PLANTS

Trachyophytes (Sporophytes)
Rhynia gwynne-vaughanii Kidston & Lang 1917, 1920a; Edwards 1986
Horneophyton lignieri Kidston & Lang 1920a; El-Saadawy & Lacey 1979a
Aglaophyton major (Kidston & Lang 1920a); Edwards 1986
Nothia aphylla Lyon 1964; El Saadawy & Lacey 1979b; Kerp *et al.* 2001
Asteroxylon mackiei Kidston & Lang 1920b; Lyon 1964
Trichopherophyton teuchansii Lyon & Edwards 1991
Ventarura lyonii Powell *et al.* 2000a

Trachyophytes (Gametophytes)
Remyophyton delicatum (♀ ♂) (of *Rhynia*) Kerp *et al.* (2004)
Langiophyton mackiei (♀ ♂) (of *Horneophyton*) Remy & Hass 1991a; Kerp *et al.* (2004)
Lyonophyton rhyniensis (♀ ♂) (of *Aglaophyton*) Remy and Remy 1980; Remy & Hass 1991b; Kerp *et al.* (2004)
Kidstonophyton discoides (♀) (of Nothia) Remy & Hass 1991c

Nematophytes
Nematophyton taiti Kidston & Lang 1921
Nematoplexus rhyniensis Lyon 1962

Algae *sensu lato*
Mackiella rotunda Edwards & Lyon 1983
Rhynchertia punctata Edwards & Lyon 1983
Palaeonitella cranii Kidston & Lang 1921; Edwards & Lyon 1983; Kelman *et al.* 2004

Lichen
Winfrenatia reticulata Taylor *et al.* 1997

Cyanobacteria
Archaeothrix contexta Kidston & Lang 1921
Archaeothrix oscillatoriformis Kidston & Lang 1921
Kidstoniella fritschii Croft & George 1959
Langiella scourfieldii Croft & George 1959
Rhyniella vermiformis Croft & George 1959 (generic name pre-occupied by the springtail *Rhyniella praecursor*)
Rhynicoccus uniformis Edwards & Lyon 1983

Fungi
Palaeomyces gordonii (No. 2) var *major* (No. 3) Kidston & Lang 1921
Palaeomyces asteroxyli (No. 7) Kidston & Lang 1921
Palaeomyces horneae (No. 8) Kidston & Lang 1921
Palaeomyces vestita (No. 9) Kidston & Lang 1921
Palaeomyces agglomerata (No. 10) Kidston & Lang 1921
Palaeomyces simpsonii (No. 13) Kidston & Lang 1921
(Kidston & Lang described 15 types of fungi though only 7 were named, 2 being variants of the same species, the others were only given numbers and are not included here)
Glomites rhyniensis Taylor *et al.* 1995
Palaeoblastocladia milleri Remy *et al.* 1994
Milleromyces rhyniensis Taylor *et al.* 1992a
Lyonomyces pyriformis Taylor *et al.* 1992a
Krispiromyces discoides Taylor *et al.* 1992a
Ascomycetes Taylor *et al.* 1999

ANIMALS

'Worms'
Nematoda
Nematode (undescribed) noted in Dunlop *et al.* (2004)
Annelida
Polychaete (undescribed)

Crustaceans
Lepidocaris rhyniensis Scourfield 1926, 1940
Castracollis wilsonae Fayers & Trewin, 2003
Ebullitiocaris oviformis Anderson *et al.* 2004
Nauplii (of *Lepidocaris*?) (Fayers *et al.* in prep.)
Noted in Fayers & Trewin 2004

Euthycarcinoid
Heterocrania rhyniensis Hirst & Maulik, 1926a,b; Anderson & Trewin, 2003

Trigonotarbids
Palaeocharinus rhyniensis Hirst 1923; Shear *et al.* 1987
Palaeocharinus hornei (Hirst 1923); Shear *et al.* 1987
Palaeocharinus tuberculatus Fayers *et al.* (in press)
(*Palaeocharinus hornei* (previously *Palaecharinoides hornei*) and *Palaeocharinus rhyniensis* are probably the only valid species of the five originally described by Hirst)
Unnamed large trigonotarbid Noted in Fayers & Trewin 2004

Araneae (Spiders)
*Palaeocteniza crassipes** Hirst 1923 (*this specimen is now regarded as a juvenile trigonotarbid (Selden *et al.* 1991))

Opilionids
Harvestman spider Dunlop *et al.* 2003
= *Eophalangium sheari* Dunlop *et al.* 2004

Arcari (Mites)
Protacarus crani Hirst 1923
Protospeleorchestes pseudoprotacarus (Hirst 1923); Dubinin 1962
Pseudoprotacarus scoticus (Hirst 1923); Dubinin 1962
Palaeotydeus devonicus (Hirst 1923); Dubinin 1962
Paraprotacarus hirsti (Hirst 1923); Dubinin 1962

Arachnida?
Unnamed arachnid(?)

Eoarthropleurids
Eoarthropleura sp. Noted in Fayers & Trewin 2004

Chilopods
Crussolum sp. Shear *et al.* 1998; Anderson & Trewin 2003
Unnamed scutigeromorph Fayers & Trewin 2004
Unnamed centipede Fayers & Trewin 2004

Hexapods
Rhyniella praecursor Hirst & Maulik 1926
Rhyniognatha hirsti Hirst & Maulik 1926; Engel & Grimaldi 2004
Leverhulmia mariae Anderson & Trewin 2003; Fayers & Trewin (in press)

Arthropoda *incertae sedis*
Rhynimonstrum dunlopi Anderson & Trewin 2003

In the basin-margin fault zone of the Windyfield Block, the Longcroft Tuffs comprises at least 140 m of intensely altered andesitic lapilli tuffs with subordinate sandstones and minor andesite. The tuffs are up to 1 m thick in normally graded beds with lithic, vesicular and non-vesicular clasts. The sandstones are laminated, in fining-upward beds, with convolute lamination and possible bioturbation. The Windyfield Sandstones and Shales comprise at least 65 m of green sandstone and shale. The sandstone beds have sharp basal erosion surfaces and are parallel- and ripple-laminated. There are a few coarse beds of granitic debris, andesite, sandstone and metamorphic rock. The Windyfield Chert (Fayers and Trewin, 2004) comprises chert lenses up to 1 m across and 0.3 m thick within blue claystones with sandstone and pebble beds. The chert contains freshwater and terrestrial plants and arthropods (Anderson and Trewin, 2003; Fayers and Trewin, 2004) and vent sinter occurs as float (Trewin, 1994).

There is widespread hydrothermal alteration of the Devonian rocks, particularly in the vicinity of the basin-margin faults (Rice *et al.*, 1995, 2002). The altered rocks and the chert sinters contain high (but non-commercial) concentrations of gold and arsenic.

Interpretation

The interpretative model of the Rhynie inlier continues to be refined as more data become available. An earlier structural model invoking a sub-vertical basin-margin fault zone and a gently dipping basal unconformity (Rice *et al.*, 1995) was revised (Rice *et al.*, 2002) in the light of 1997 drilling results. In the 2002 model (Figure 2.74), a zone of extensional faulting along the western basin margin dips 35° eastwards, flattening to a 15° listric basement/cover fault under the basin. The basin-fill succession dips northwestwards, except close to the basin margin, where the strata are folded into a syncline. Further mapping supported by excavation and geophysical (magnetic) survey (Rice and Ashcroft, 2004) has revealed more extensive faulting and folding consistent with a dextral strike-slip origin for the northern part of the basin.

In the Rhynie Block, the Pre-lava Sandstone is interpreted as a small basin-margin alluvial-fan produced by a combination of flash-flood and channel deposition (Trewin and Rice, 1992; Rice

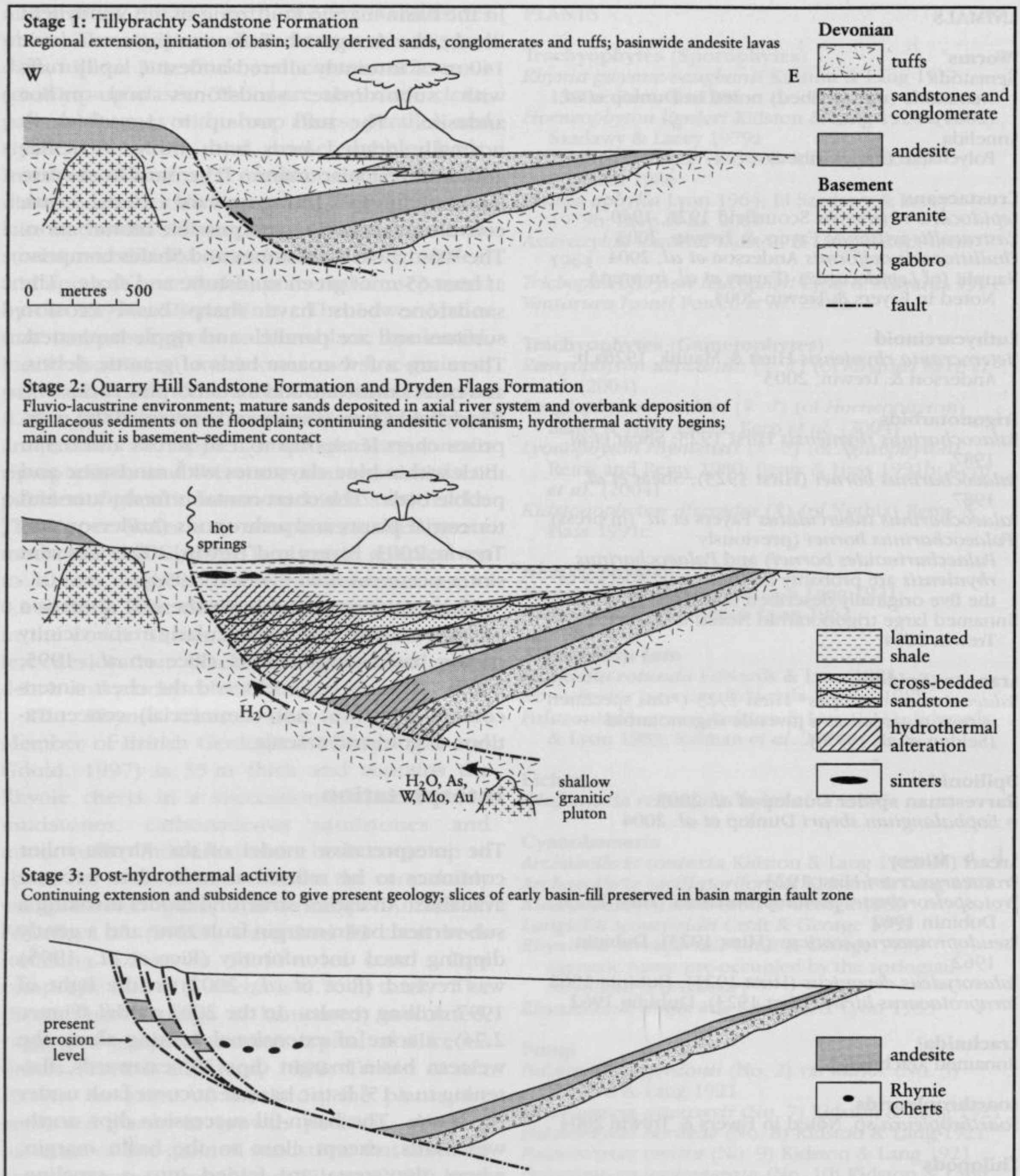


Figure 2.74 Evolution of the Rhynie Basin and hot spring development. After Rice *et al.* (2002).

et al., 2002). The rare calcrete nodules indicate soil formation in a subaerial, seasonally wet environment and semi-arid climate. The clastic rocks are derived from the local igneous basement and there is no evidence for volcanicity prior to the eruption of the overlying andesitic lava flow. The White Sandstones are interpreted as the deposits of shallow, fluvial channels and sheet-flood events. The Shales with Muddy Sandstones, Lower Shales and Upper Shales are predominantly lacustrine and floodplain deposits. Trewin and Rice (1992) and Rice *et al.* (1995,

2002) interpreted the depositional environment as ephemeral, shallow lakes on an alluvial plain that periodically dried to desiccated mudflats. The fine lamination of the mudstones is attributed to minor seasonal changes in sediment supply, the thin, rippled sandstones representing clastic input to the lakes during flooding events. The tuffaceous sandstones at the base of the Shales with Muddy Sandstones (the Tuffaceous Sandstones of Trewin and Rice, 1992) are largely fine grained and quartzose, but also contain angular, vesicular andesitic clasts. Some of the more tuffaceous beds may be volcanic ashfall deposits, implying contemporaneous volcanicity, but most of the volcanic material was transported by water from a local source. Trewin and Rice (1992) placed these beds stratigraphically immediately above the andesitic lava, but the contact between the units is now interpreted as a fault (Rice *et al.*, 2002). Archer (1978) and Trewin and Rice (1992) interpreted the beds as alluvial-plain deposits, with regional axial current flow from the south (Archer, 1978). Trewin and Rice suggested that a nearby tuff cone undergoing erosion provided the volcanic debris, and that the laminated shales were overbank deposits of small ephemeral pools. The rare calcrete nodules are evidence of carbonate soil formation in a climate that continued to be semi-arid and seasonally wet.

The Rhynie Cherts Unit has received much attention on account of its plant-bearing cherts. For a historical summary see Cleal and Thomas (1995). Interpretations are given by Trewin and Rice (1992), Trewin (1994), Rice *et al.* (1995, 2002) and Trewin and Wilson (2004). The revised model (Rice *et al.*, 2002) invokes silica deposition from mineral-rich hydrothermal fluids that emanated from hot (about 100°C) springs along basin-margin faults during the waning of local volcanism. Hot spring activity may have occurred at intervals along at least 2 km of the fault zone as pulses of fluid were released by subsidence movement along the fault zone. The silica-rich fluids permeated an alluvial plain with scattered small ponds, depositing surface sinter. Trewin and Wilson (2004) conclude that sinter deposition took place on a low-angle outwash apron from a hot spring, and that overbank flooding from a river system deposited the sand and mud interbeds (Figure 2.75). The silicification of standing plants and plant debris resulted from flooding of the land surface, but the aquatic arthropods and algae preserved in the chert probably inhabited cooler pools within areas of sinter. The wide range of preservation of the plants is probably due to silicification at different times in the cycle of plant growth and decay, combined with the variable efficiency of the silicification process. An outline model for the chert formation proposed by Rice *et al.* (2002) is shown in Figure 2.74. The sinter textures in the chert compare with those in modern subaerial and

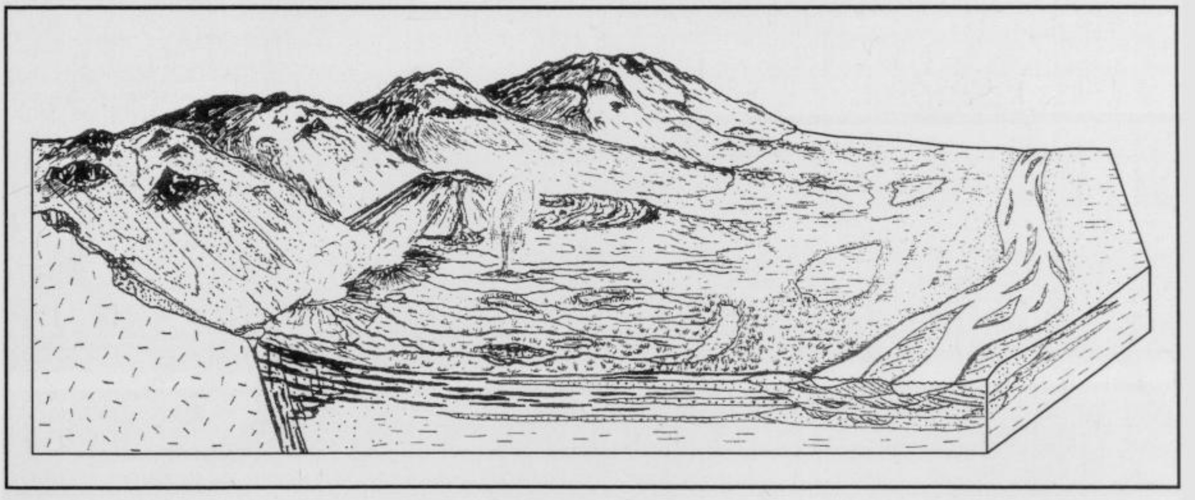


Figure 2.75 Cartoon view looking north to illustrate the Rhynie hot spring system and the low-angle sinter apron crossed by streams emanating from hot spring vents. Plants colonized the apron along stream banks and ponds, and on alluvial floodplain areas. The sinter apron was periodically flooded by waters from the axial river system. Some coarse detritus and reworked volcanic ejecta were sourced from the west. After Trewin and Wilson (2004).

subaqueous sinters in the cooler outwash areas of hot springs in Yellowstone National Park USA, and indicate a depositional temperature below 30°C for the majority of the plant-bearing cherts (Trewin *et al.*, 2003).

In the Windyfield Block, the Longcroft Tuffs are interpreted by Rice *et al.* (2002) as the product of volcanic activity from a nearby centre, producing lava and airfall tuffs, as well as the product of the erosion of the volcanic source. This introduced coarse debris during flood events into an otherwise low-energy fluvio-lacustrine floodplain. The Windyfield Chert was the product of hydrothermal silicification of the contents of freshwater pools in the vicinity of a hot spring vent.

Rice *et al.* (2002) suggest correlation of the strata in the Rhynie and Windyfield blocks with the succession recognized in the northern part of the Rhynie Basin (Figure 2.73). The Pre-lava Sandstones are correlated with the Tillybrachty Sandstone Formation (British Geological Survey, 1993; Gould, 1997) and the White Sandstones may correlate with the topmost part of the Quarry Hill Sandstone Formation. The succeeding beds of both the Rhynie and Windyfield blocks, including the Rhynie and Windyfield cherts, are correlated as part of the Dryden Flags Formation.

Conclusions

The GCR site at Rhynie is arguably the most important Old Red Sandstone site in Great Britain and one of the most important in the world. In addition to its status as a unique Early Devonian faunal and floral lagerstätte, it is also one of the earliest-known occurrences of the surface expression of a hydrothermal spring system. The strata include cherts that contain an exceptionally well-preserved, silicified, internationally important early land-plant and arthropod assemblage, the whole in-situ biota providing an insight into plant and arthropod evolution and the development of terrestrial ecosystems. The fossiliferous horizons are placed into a context of evolving geological environments at the margins of a small, subsiding basin. Initial alluvial-fan deposition was followed by the eruption of andesitic lavas, marking the local onset of volcanicity. Subsequent sedimentation was on an alluvial plain with ephemeral lakes and ponds. The climate throughout was semi-arid and seasonally wet. As volcanicity waned, the basin-margin faults acted as conduits for silica-rich, mineralized fluids. These emerged on the alluvial plain as hot springs and preserved the biota in siliceous sinters, now represented by the Rhynie and Windyfield cherts.