Precambrian Rocks of England and Wales

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Chapter 5

Shropshire, Radnor and Llangynog

INTRODUCTION

D. Wilson

The Precambrian rocks between Shropshire and Llangynog form a major network of GCR sites disposed in a series of inliers that are bounded by faults and unconformities associated with the major Church Stretton and Pontesford fault systems (Figure 5.1). The inliers vary in size, the largest being that of the Long Mynd massif, and they generally give rise to the impressive hills and fine scenery exemplified by the Wrekin range and the area around Church Stretton (Figure 5.2). Two major groupings of Precambrian rocks crop out within the region. The older of these is the Uriconian Group, which consists of rhyolitic lavas and associated pyroclastic rocks, with acid and basic intrusions, representing a style of highly explosive, subaerial volcanism. At the Wrekin GCR site, the volcanic rocks are juxtaposed with older schistose and gneissose rocks whose origins are conjectural, but which may in part represent fragments of an earlier Precambrian basement. The Uriconian rocks are overlain by clastic sedimentary sequences of the Longmyndian Supergroup, which represents one of the thickest continuous successions of latest Precambrian sedimentary rocks in southern Britain. These strata record a transition from marine to fluviatile environments and contain, at Ashes Hollow GCR site,

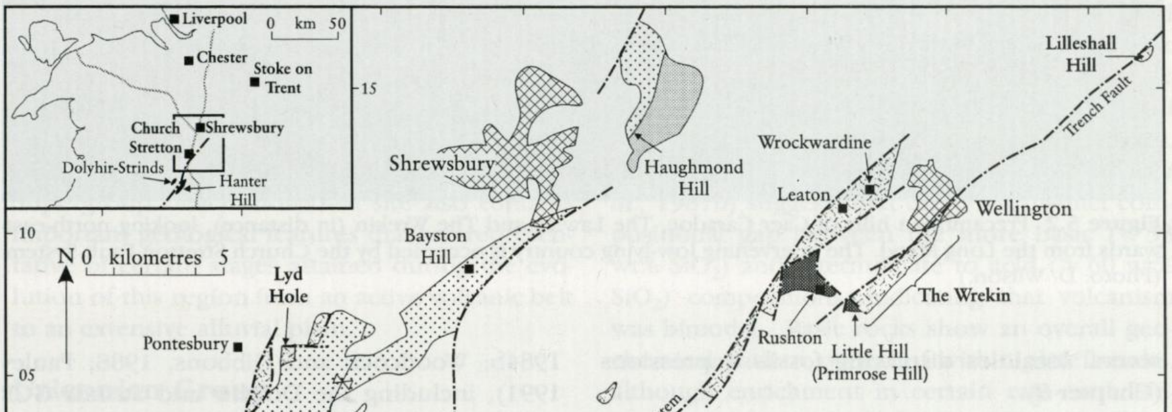

Figure 5.1 Geological map of the Shropshire Precambrian outcrops (modified from Pauley, 1991), with the GCR sites indicated by bold lettering. The Radnor inliers (Dolyhir and Strinds quarries and Hanter Hill sites) are shown by the inset at top left. The location of the Llangynog site is given in Figure 6.1.

Figure 5.2 Precambrian hills of Caer Caradoc, The Lawley and The Wrekin (in distance), looking north-eastwards from the Long Mynd. The intervening low-lying country is occupied by the Church Stretton Fault System. (Photo: D. Wilson.)

several localities displaying fossils impressions (Chapter 8).

The smaller inliers at Old Radnor (Dolyhir and Strinds GCR site) and Llangynog are included here because they have strong lithological affinities with the sedimentary and volcanic sequences of Shropshire. The Stanner–Hanter inlier is unusual, however, in exposing plutonic rock suites. All of these outcrops are representative of rocks comprising the Wrekin Terrane (Figure 1.1), the geological composition of which is discussed in the introductory chapter of this volume.

One of the most important aspects of the well-exposed Shropshire area is its highly complex structural history, in particular the evolution and significance of the major structures that affect Precambrian rocks. These include the Church Stretton and Pontesford fault zones (or 'lineaments') that form part of the Welsh Borderland Fault System (Woodcock and Gibbons, 1988), a long-lived crustal structure that controlled late Precambrian volcanism and sedimentation (Baker, 1973), and may represent a Gondwanan terrane boundary (Figure 1.1). Studies of the Welsh Borderland Fault System (Woodcock 1984b; Woodcock and Gibbons, 1988; Pauley 1991), including the Dolyhir and Strinds GCR site (Woodcock, 1988), have demonstrated the importance of strike-slip displacements periodically during its evolution. Movement along its component faults is thought to have resulted in considerable dismemberment of the Uriconian and Longmyndian sequences (Pauley, 1991). The presence of a locally penetrative cleavage within the Precambrian rocks along the Welsh Borderland Fault System, and its absence from the Cambrian strata, which unconformably overlie them, are generally taken as evidence of a major late Precambrian tectonic event along this important structure. The cleavage that affects the Precambrian rocks strikes NE–NNE, transecting the associated NNE–N fold axes (including the Long Mynd Syncline) in a clockwise sense by angles of up to 22°. This fold-cleavage geometry was considered by Pauley (1991) to have developed during transpressional deformation associated with sinistral displacements along the Welsh Borderland Fault System; he further suggested that the movements were associated with plate tectonic readjustments that terminated Uriconian magmatic activity in late Precambrian to early Cambrian times.

Studies to constrain the age of the cleavage have been carried out on samples collected from various locations, including the Lightspout Hollow GCR site, and are based on the $^{40}Ar/^{39}Ar$ systematics of tectonic micas in deformed Longmyndian mudrocks (BGS, work in progress). The preliminary results suggest that this was probably not a Caledonian event, although it occurred rather later than the end-Avalonian (Cadomian) tectonic dismemberment of terranes that is documented to have occured in the period between 570 and 550 Ma farther west (Gibbons and Horák, 1996). That event included the uplift of blueschists in Anglesey at 560–550 Ma (Dallmeyer and Gibbons, 1987; see also, Chapter 7 of this volume).

The twelve sites of this GCR network, shown in Figure 5.1, are chosen so as to provide an overview of the stratigraphy and structure of these Precambrian rocks, with particular reference to sedimentological evolution within the lower to middle part of the Longmyndian Supergroup. Each individual site also contains important geological features that are representative of certain stages attained during the evolution of this region from an active volcanic belt to an extensive alluvial plain.

Uriconian Group

The term 'Uriconian' was first applied by Callaway (1886) to the acid, intermediate and basic tuffs and lavas, intruded by granophyre and several dolerite dykes, that crop out at the Wrekin and Lilleshall Hill GCR sites. It was at the former site that Callaway (1879) demonstrated the unconformable nature of the Uriconian and Cambrian, on the basis of derived volcanic pebbles in the overlying Cambrian quartzite. However, it was not until Lapworth (1888) proved the existence of an early Cambrian fauna in the quartzites that a Precambrian age for the Uriconian volcanics along the Church Stretton Fault System was accepted. Even so, correlation of these 'Eastern Uriconian' rocks with the so-called 'Western Uriconian' along the Pontesford–Linley Fault System proved controversial (Callaway, 1882; Blake, 1890; Lapworth and Watts, 1910), and was only established when the structure of the Long Mynd was revealed as a major syncline (James, 1952, 1956). The disparate Uriconian outcrops are now generally considered to lie on opposing limbs of the syncline, which is cored by sedimentary rocks of the Longmyndian Supergroup. The term 'Uriconian volcanic complex' was used for these rocks by Pauley (1986, 1991); however, although the sequence is as yet undivided it is well stratified and could, eventually, be amenable to further subdivision. The name 'Uriconian Group' (Pharaoh and Gibbons, 1994) is therefore preferred here.

The Uriconian rocks have been widely studied, as summarized in: Pocock *et al.* (1938); Greig *et al.* (1968); Dunning (1975) and Pharaoh and Gibbons (1994), and feature in a number of isotope investigations and plate-tectonic reconstructions of the British Isles (Baker, 1973; Thorpe, 1974; Patchett *et al.*, 1980; Dewey, 1982; Thorpe *et al.*, 1984; Pharaoh *et al.*, 1987b; Piper and Strange, 1989; Tucker and Pharaoh, 1991). The lavas are largely potassic rhyolites and the intrusions include both basic and acidic types, with an important phase of intrusion represented by the Ercall Granophyre. The geochemistry of Uriconian lavas (Pharaoh *et al.*, 1987b) suggests that there is a distinct compositional gap between the more basic (< 53 wt% SiO_2) and intermediate to acid (> 60 wt% SiO_2) compositions, indicating that volcanism was bimodal. Basic rocks show an overall geochemical pattern typical of within-plate basalts, although enrichment in certain rare-earth elements indicates a subduction-related component such as would be found in a volcanic arc environment. Current opinion (Pharaoh and Gibbons, 1994) is that the Uriconian rocks were erupted in a tectonic setting that was transitional between the two, such as a fault-controlled ensialic marginal basin within the Avalonian volcanic arc (Thorpe *et al.*, 1984; Figure 1.4).

The eruptive age of the Uriconian Group has been subject to some debate (summarized by Pharaoh and Gibbons, 1994). A maximum age of eruption is given by a Rb-Sr metamorphic age of 667 ± 20 Ma from the Rushton Schists (Thorpe *et al.*, 1984), which are in probable faulted contact with largely unmetamorphosed Uriconian volcanic rocks (Coppack, 1974). More reliable data has come from Patchett *et al.* (1980), who published a Rb-Sr whole rock isochron of 558 ± 16 Ma from acid tuffs within the Eastern Uriconian. This age has been generally confirmed by a U-Pb zircon date from a rhyolite lava near Leaton, in the Wrockwardine inlier (Figure 5.1) which gave an age of 566 ± 2 Ma, considered to be the eruptive age (Tucker and

Pharaoh, 1991). That Uriconian magmatism was also diachronous, however, is shown by chemical studies of tuffaceous strata in the Longmyndian Supergroup that constitute the 'Batch Volcanic Beds', summarized in the description of the Long Batch–Jonathan's Hollow GCR site. The upper limit to Uriconian volcanicity is to some extent constrained by the fact that the Lower Cambrian Wrekin Quartzite, which overlies Uriconian intrusive rocks of the Ercall Granophyre at the Wrekin range GCR site, is probably late Tommotian, around 530–534 Ma, in age (Rushton, 1974; Cope and Gibbons, 1987). Radiometric dating of the Ercall Granophyre provides a more accurate estimate for magmatic cessation. The U-Pb zircon age of 560 ± 1 Ma is regarded as that of emplacement (Tucker and Pharaoh, 1991) and it indicates that the Uriconian and much of the Longmyndian arose within a relatively short space of time, between 566 and 560 Ma. The age determination also provides a reliable constraint for deposition of the early Cambrian strata of England.

Longmyndian Supergroup

The Longmyndian Supergroup (Toghill and Schell, 1984) forms a sequence at least 6500 m thick dominated by sedimentary rocks. Its strata are mainly disposed in the Longmynd Syncline (James, 1956), a major eastward-facing isoclinal fold between the Pontesford–Linley and Church Stretton fault zones (Figure 5.1). The Precambrian age of the Longmyndian has long been recognized (Callaway, 1879), and the rocks have received considerable attention for their fine sedimentary sequences and structures, as well as for their enigmatic stratigraphical and structural relationships with the Uriconian Group. Despite this, the broad stratigraphy of the Longmyndian has changed little from the time of Lapworth and Watts (1910) who, based on earlier work by Blake (1890) and Cobbold (1900), divided the sequence into an eastern and western group (named the Stretton and Wentnor 'Series' respectively). These were further subdivided into a number of 'Groups' (the terms 'Series' and 'Group' have now been redesignated as 'Group' and 'Formation' respectively; Dunning, 1975). The succession has been described in considerable detail by Greig *et al.* (1968), who considered that the rocks formed in a range of depositional environments. Modern sedimentological analysis of the Longmyndian Supergroup has been undertaken by Pauley (1986; 1990a,b; 1991), who has recognized basinal, turbiditic, deltaic and fluviatile sequences within the succession, but has retained the existing lithostratigraphical subdivisions of Greig *et al.* (1968).

The relationship between the Uriconian Group and Longmyndian Supergroup is not easily demonstrated owing to a lack of suitable sections, and the faulted nature of the Precambrian outcrop in the Welsh Borderlands. One of the more important sections where a contact is seen is the Lyd Hole GCR site, described below. The abundant volcanic fragments within the Longmyndian have been matched with Uriconian volcanic rocks, indicating that the latter was the source for much of the sediment. Work by James (1952, 1956) firmly established that Uriconian rocks underlie the Longmyndian east and west of the Long Mynd, a relationship that had previously been in dispute (cf. Callaway, 1882; Blake, 1890); more importantly, it confirmed that the main period of volcanism predated Longmyndian sedimentation. The possibility that the waning phase of Uriconian volcanic activity overlapped with Longmyndian sedimentation (Greig *et al.*, 1968) is indicated by the sporadic occurrence of thin volcanic horizons within the Longmyndian succession, the most important of which are the 'Batch Volcanic Beds' (Cobbold, 1900), a group of intermediate felsic tuffs. It suggests that the Longmyndian sediments were derived from erosion of the Uriconian volcanic arc, and that no great time interval separated volcanism and sedimentation.

Perhaps the most important feature of the Longmyndian is its sedimentary evolution, from marine to fluviatile depositional environments, and the implications that follow for reconstructing the tectonic events that may have accompanied the waning phases of Uriconian magmatism. One of the most important sections is at the Ashes Hollow GCR site, where fossil impressions are also well displayed (Chapter 8). It demonstrates a major regressive sequence within the Stretton Group, revealing a transition from deep-water mudstones (Stretton Shale Formation), through progradational turbidites and subaqueous deltaic mudstones and siltstones in the middle and upper parts of the Burway Formation. The top of the Burway Formation is characterized by fluvial environments, represented by the Cardingmill Grit

(Pauley, 1990a). Fluviatile, sandstone-dominated facies characterize the later parts of the Stretton Group, with alluvial floodplain environments commencing in the Synalds Formation. In this formation the occurrence of the Batch Volcanic Beds, best seen at The Pike and Long Batch–Jonathan's Hollow GCR sites, represents the youngest record of volcanic activity contemporary with Longmyndian sedimentation. The sandstone component thickens upwards through the Lightspout Formation, reflecting deposition by multiple sheetflood events, but at the Lightspout Hollow GCR site there is evidence for an upwards transition into strata indicative of braidplain deposits.

A marked change in sedimentation occurred at the junction between the Stretton Group and overlying Wentnor Group, as described at the Hawkham Hollows GCR site. The change from alluvial floodplain sedimentation in the Portway Formation, to the deposition of braidplain conglomerates in the Bayston–Oakswood Formation, basal Wentnor Group, is an abrupt one that has been linked to rejuvenation of the source area by movements along the Church Stretton Fault Zone (Pauley, 1990b).

Coomb Volcanic Formation and Johnston Diorite Complex

The Coomb Volcanic Formation, described at the Llangynog site (see Figure 6.1 for location), is a bimodal suite of basaltic and rhyolitic lavas and associated volcaniclastic rocks (Cope and Bevins, 1993). The volcaniclastic rocks show the features of shallow water to subaerial deposition, and an active continental margin tectonic setting was inferred by Bevins *et al.* (1995) on geochemical grounds.

Although the Coomb Volcanic Formation as yet remains undated, the presence of Ediacaran faunas within associated sedimentary rocks at Coed Cochion (Chapter 8) provides a possible parallel with similar fossils from the Eastern Avalonian, Mistaken Point Formation in Newfoundland. These have a minimum age constraint provided by overlying rhyolitic tuffs dated at 565 ± 3 Ma (Benus, 1988). As a Uriconian rhyolite from Shropshire has yielded a date of 566 ± 2 Ma (see above), and the Coomb Formation and Uriconian are geochemically very similar (Bevins *et al.*, 1995), there seem strong grounds for correlating the two.

Stanner–Hanter Complex

The intrusive rocks of this igneous complex form part of the chain of Precambrian inliers localized along the Church Stretton Fault System (Figure 1.1). At the Hanter Hill GCR site, the gabbros, dolerites and minor granitoids are comparable with lithologies in the Malverns Complex (Chapter 4). The Stanner–Hanter rocks nevertheless appear to be older, with a Rb-Sr isochron age of 702 ± 8 Ma (Patchett *et al.*, 1980).

Another component of the Wrekin Terrane, the Johnston Diorite Complex (Figure 6.1), consists of a heterogeneous, calc-alkaline granodiorite–granite suite comparable to the Malverns Complex and estimated to have been emplaced at about 643 Ma (Patchett and Jocelyn, 1979). Closely associated with it, the Benton Group contains dacitic to rhyolitic tuffs. In a geochemical survey by Bevins *et al.* (1995), the tuffs were shown to plot within or close to the within-plate granite field on Nb-Y vs Zr-TiO_2 diagrams, and are thus comparable to volcanic rocks of the Wrekin Terrane. No GCR sites have been selected for these occurrences.

Metamorphic rocks

These rocks constitute the Rushton Schist and Primrose Hill Gneiss and Schist (Figure 5.1). The latter are poorly exposed at the Wrekin GCR site, where they are in faulted contact with Uriconian rocks and, in relatively strain-free areas, appear to consist of gneissified granophyre and possible acid tuffs. The Rb-Sr metamorphic age of 667 ± 20 Ma determined for the Rushton Schist (Thorpe *et al.*, 1984) indicates that it was formed at least 100 million years before eruption of the Uriconian Group. The model age, of about 1700 Ma, further suggests that the Rushton Schist could represent basement to the Uriconian Group, with the two units subsequently juxtaposed by faulting.

THE WREKIN RANGE (SJ 617 074–647 103)

D. Wilson

Introduction

The Wrekin range is the type area for the late Precambrian Uriconian Group (Pauley, 1986,

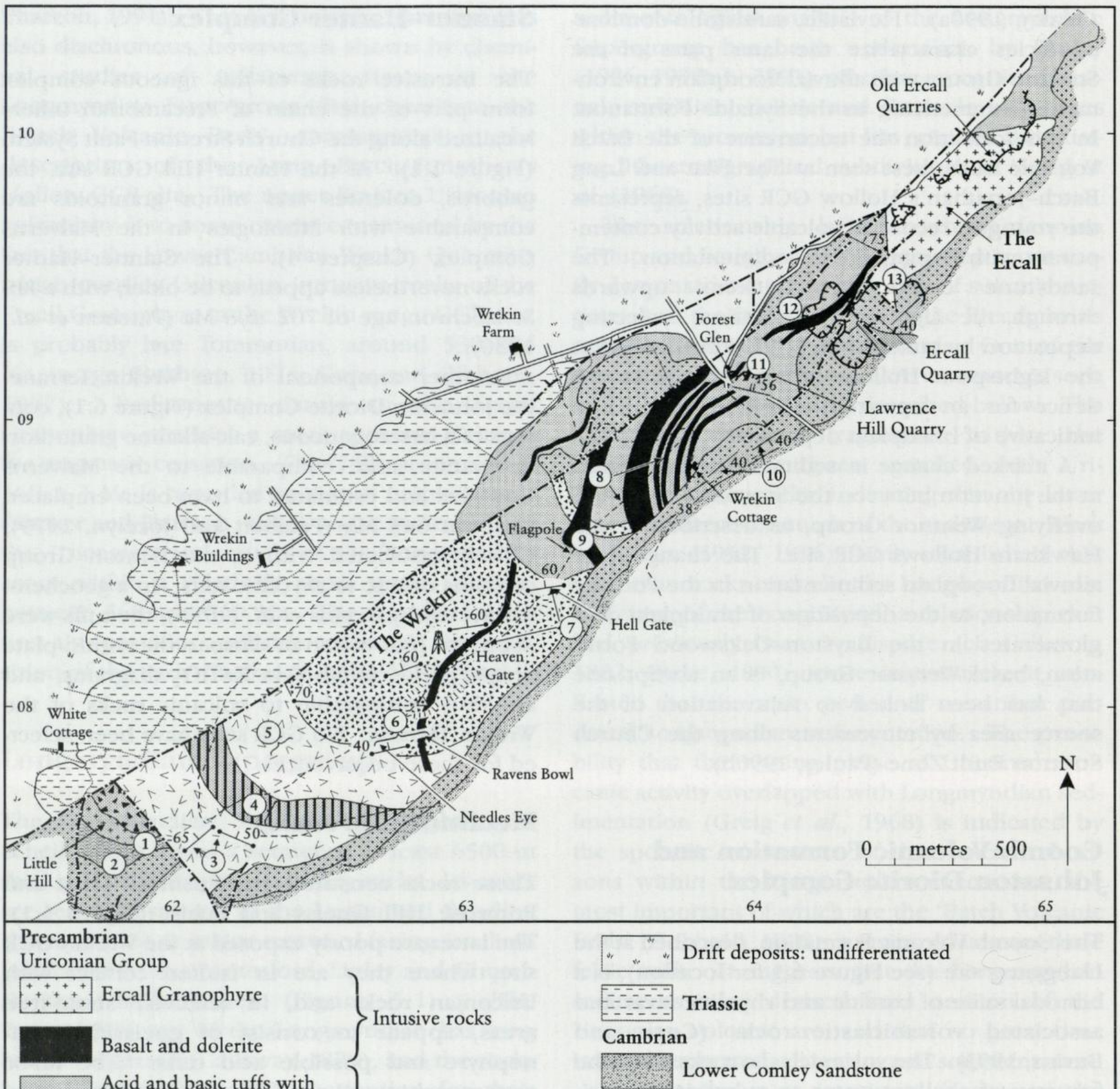

Figure 5.3 Geological map of the Wrekin range.

1991), and has been included as a GCR site because it undoubtedly provides the best and most varied exposures of these rocks in England. The intrusive rocks, lavas and pyroclastic beds of the Wrekin (Pocock *et al.*, 1938; Hamblin and Coppack, 1995) form part of the so-called Eastern Uriconian (Callaway, 1886), extending in a series of isolated outcrops along the line of the Church Stretton Fault System. The site (Figure 5.3) covers an area of some 4 km^2 and includes the impressive ridge of ancient rocks that runs south-west from the Ercall, across the Wrekin to Primrose Hill. This area is of great historical importance, for it was here that the volcanic nature of these rocks was first demonstrated (Allport, 1877), and the term 'Uriconian' applied to them by Callaway (1886). It was here also that Callaway (1879) demonstrated the unconformable nature of the Uriconian and Cambrian strata, which flank the Wrekin to the east and south, as discussed in the introduction to this chapter.

Although most of this ridge is composed of Uriconian rocks, there is also a small but stratigraphically significant outcrop of Primrose Hill Gneiss, which has been interpreted as a rare inlier of Malvernian-type basement (see also, Chapter 4).

Description

The oldest Precambrian representatives at the Wrekin site are considered to be the metamorphic and intrusive rocks of Primrose Hill (renamed Little Hill on modern OS maps), a small feature separated from the main mass of the Wrekin by a depression containing the Primrose Hill Fault. The poorly exposed Primrose Hill Gneisses (Pocock *et al.*, 1938; Wright, 1969; Thorpe, 1974) comprise a range of rock types that include granitic and amphibolitic gneisses, altered tuffs, intrusive granite and pegmatite. The best exposures are near the summit of Little Hill (Figure 5.3, Locality 1) and in the steep path to the south (Locality 2). They reveal a group of gneissose rocks consisting of dark grey hornblende-rich layers, which alternate with layers rich in the leucocratic minerals, quartz, plagioclase and alkali feldspar. Granite and granitic lithologies, which crop out at the summit and down the western slope, are albite-microcline-quartz rocks with accessory muscovite and chloritized biotite (Thorpe, 1974). In places they appear gneissose, thin sections suggesting that the fabric may result from streaking-out of original granitic or granophyric material. Specimens obtained by Pocock *et al.* (1938), from the summit, show this group of rocks to contain chloritic and garnetiferous concentrations, probably representing highly altered xenoliths. The uppermost group, which crops out on the northern and north-eastern slopes, is composed of acid igneous material, resembling intensely deformed crystal tuffs with a strong cataclastic texture (Pocock *et al.*, 1938). All the rocks on Little Hill have undergone a variable amount of solid state granulation and recrystallization, and are considerably veined by coarse pegmatitic material.

The main mass of the Wrekin is formed by a northerly dipping succession of up to 700 m of rhyolitic lavas and interbedded pyroclastic rocks. The lower part, which crops out on the steep south-western slopes of the hill, is well displayed in crags (SJ 6214 0760 and SJ 6204 0756) on the forest path (Locality 3). It consists of at least 200 m of fine to coarse, variably welded, basic and subordinate acid crystal-lithic lapilli tuffs, in places with clasts of rhyolite up to 10 mm, interbedded with tuffaceous conglomerates (Pocock *et al.*, 1938). Higher in the sequence, about 80 m of rhyolitic crystal-lithic lapilli tuffs, with andesitic and dacitic fragments, vitric tuffs and devitrified purplish red rhyolites are exposed in crags (SJ 6233 0771) east of the summit path (Locality 4) and on the south-eastern slopes of the hill. The succeeding pyroclastic rocks comprise a thick (200 m) sequence of basic lithic lapilli tuffs, crystal-lithic tuffs and some vitric-crystal tuffs showing variable degrees of welding and alteration. There are thin horizons of largely devitrified flow-banded rhyolite, and beds of lithic lapilli tuff composed of large fragments of rhyolite in crags (SJ 6233 0790) south-west of the summit (Locality 5). On the south-eastern side of the Wrekin, the pyroclastic rocks and rhyolite are intruded by a series of narrow veins of granitic material, composed of fine-grained irregular intergrowths of quartz and alkali feldspar locally with pronounced spherulitic structure (Pocock *et al.*, 1938).

The summit of the Wrekin is composed of rhyolitic rocks (the 'Summit Rhyolite' of Pocock *et al.*, 1938) overlying the pyroclastic succession. They form a series of flows with a few thin pyroclastic horizons, and are well exposed in the impressive crags of the Raven's Bowl (SJ 6283 0799), Bladder Stone (SJ 6275 0795) and

Figure 5.4 Crags in massive, silicified rhyolitic lavas of the Uriconian Group at Raven's Bowl, south-east side of The Wrekin. (Photo: D. Wilson.)

Needle's Eye (SJ 6267 0795) 150 m south of the summit (Figure 5.4; Locality 6), and elsewhere on the flanks of the hill. The rhyolites are generally massive, silicified, and largely devitrified quartz-alkali feldspar rocks with abundant spherulitic feldspar growths, minor chlorite intergrowths and a few albite phenocrysts. They are coloured pink, purplish red, brown and greenish grey, typically weathering brown and buff. A strongly developed pink to brown flow-banding foliation is present in many exposures and is particularly noticeable in small crags (SJ 6303 0832; SJ 6324 0853) along the footpath leading north from the summit (Locality 7). The bands dip generally northward, but are rarely planar, showing varying degrees of puckering, folding and brecciation.

At Hell Gate (SJ 6323 0853), about 270 m north-east of the track to the television transmitter, rhyolitic rocks with well-developed flow-banding are interbedded with strongly welded lithic lapilli tuffs containing 5 mm clasts of dacitic or andesitic rock (Locality 7). From this point northward along the track, pyroclastic material becomes increasingly coarser and more abundant, and the northern end of the Wrekin is composed of at least 400 m of mostly coarse-grained rhyolitic, andesitic and dacitic tuffs. The tuffs are intermittently exposed on the path (Locality 8) that leads from the entrance of the

Country Park to the summit, via Wrekin Cottage (SJ 6371 0891). They are generally unwelded crystal-lithic lapilli tuffs, containing fragments of andesitic, rhyolitic and dacitic rocks in varying proportions, older welded tuff and quartzite, together with broken quartz and plagioclase phenocrysts; lithic clasts may be locally up to 50 mm in diameter. At intervals there are beds of fine-grained vitric-crystal tuffs within the sequence, but these are generally subordinate to the coarse material. West of Wrekin Cottage, one of several N- to NE-trending dolerite dykes intruding the tuffs is exposed in the track leading to the summit near the flagpole marking the firing range (SJ 6330 0865) (Locality 9). East of Wrekin Cottage, quartzose sandstones of the Lower Cambrian Wrekin Quartzite are exposed intermittently on the footpath leading to the park entrance at Forest Glen (Locality 10). The unconformity with the underlying tuffs is not well exposed, but the quartzite locally contains pebbles of highly weathered igneous material.

On the north side of Forest Glen, which separates the Wrekin from Lawrence Hill and the Ercall, three of the previously mentioned NE-trending dolerite dykes are exposed in the disused Lawrence Hill Quarry (SJ 6390 0923). The quarry is important, however, for its section through coarse-grained pyroclastic rocks and lavas that represent the upper part of the sequence exposed on the northern part of the Wrekin (Locality 11). The section reveals a group of coarse purple and green, andesitic and rhyolitic, lithic-lapilli tuffs, with some vitric and crystal tuffs, strongly welded and graded in places. In the quarry face there is a volcanic breccia horizon with blocks of rhyolite and quartz locally up to 0.3 m long (Pocock *et al.*, 1938; Toghill and Beale, 1994). In the upper part of the quarry, a sequence of vitric and lithic tuffs and lapilli tuffs includes vesicular olivine basalt considered to be a lava flow (Pocock *et al.*, 1938). At the top of the quarry is a rhyolite or dacite, which forms part of a thick sequence of flow-banded rhyolites that outcrop over much of the ground between Lawrence Hill and the Ercall.

The Ercall, lying north-east of Lawrence Hill, is largely composed of a granophyre pluton, which in thin section shows coarse micrographic intergrowths of quartz and feldspar enclosing rectangular plagioclase (albite-oligoclase) phenocrysts. Previous names given to the rock include 'granitoidite', 'eurite', 'aplite' and 'ercallite'. The Ercall Granophyre appears to have intruded the Uriconian rocks (cf. Bonney, 1879), although field evidence is now limited, and it has inevitably been compared with the intrusive granitic rocks of the Wrekin and Primrose Hill (Pocock *et al.*, 1938; cf. Thorpe, 1974). There are many exposures of the rock in disused and overgrown quarries (SJ 6455 1015) beside the road from Buckatree Hall to Wellington, but the principal ones are now in the disused Ercall Quarries (SJ 6435 0955), where the granophyre is overlain unconformably by the Wrekin Quartzite (Figure 5.5). The quarries expose a range of lithologies and relationships, described in detail by Wright *et al.* (1993) and Toghill and Beale (1994). In the main western quarry (SJ 6407 0952), flow-banded rhyolites and unwelded crystal-lithic tuffs (Dearnley, 1966) are intruded by, or possibly faulted against, a dolerite dyke that represents a continuation of one of those exposed in Lawrence Hill Quarry (Locality 12). The basal conglomerates of the Wrekin Quartzite, and the quartzite itself, overlie and are possibly faulted against the Uriconian rocks in the south-eastern corner of this quarry (SJ 6410 0945); however, the unconformity is exceptionally well displayed in the large north-eastern quarry (SJ 6436 0962) (Locality 13). The irregular top of the Ercall Granophyre is variably weathered to a depth of 0.5 m, and the overlying conglomerates of the basal Wrekin Quartzite contain pebbles up to 0.1 m, mostly of Uriconian volcanics and, rarely, granophyre (Cope and Gibbons, 1987).

Interpretation

The Precambrian rocks of the Wrekin include the stratigraphically important but poorly understood Primrose Hill Gneisses. Locally these bear a strong resemblance to the Ercall Granophyre, and thus could be the recrystallized and tectonically reworked equivalents of Uriconian intrusive lithologies, although the textures of the metamorphic rocks are generally coarser (Hamblin and Coppack, 1995). It is noteworthy that Callaway (1886) described pebbles of schist and gneiss in nearby Uriconian sedimentary rocks that may be comparable to Primrose Hill metamorphic lithologies, but this has not been confirmed.

The geochemistry of the Uriconian Group (Pharaoh *et al.*, 1987b), summarized in the

Figure 5.5 Exposures in Ercall Quarry showing the Ercall Granophyre (left of picture) unconformably overlain by well-bedded Lower Cambrian Wrekin Quartzite. (Photo: D. Wilson.)

introduction to this chapter, suggests that this is a bimodal magmatic suite, which was generated in an ensialic marginal basin environment within a late Precambrian volcanic arc system (Figure 1.4). Radiometric studies further suggest that eruption occurred between 566 ± 2 and 560 Ma, the latter being the date of the Ercall Granophyre, which is considered to be the terminating episode of Uriconian magmatism. Most of the pyroclastic lithologies on the Wrekin are either acidic (rhyolitic and dacitic) or basic in composition, with subordinate intermediate rocks such as andesite; the lavas are largely potassic rhyolites. The volume and extent of pyroclastic material within this sequence is indicative of a highly explosive style of volcanism, whereas the occurrence of welded tuffs, and a general absence of pillowed basalts, suggests that the eruptions probably occurred subaerially. The abundance of coarsely fragmental lithologies, such as lithic-lapilli tuffs and local volcanic breccias, within the succession indicates proximity to volcanic centres, but no such source has yet been identified within the Uriconian Group.

The 560 Ma age of the Ercall Granophyre is also the maximum possible value for the age of the unconformable Cambrian strata. In order to more closely constrain the latter's age, however, Wright *et al.* (1993) have quoted a Rb-Sr age of 533 ± 13 Ma obtained by Patchett *et al.* (1980) on the granophyre. On the assumption that it reflects a resetting of isotope systems by a major post-Longmyndian tectonic episode, this lower value could be regarded as providing a rather younger, albeit rather imprecise, maximum age for the Cambrian marine transgression. It is in keeping with the late Tommotian age that has been proposed for the Lower Cambrian Wrekin Quartzite by Cope and Gibbons (1987).

Conclusions

The Wrekin range includes excellent examples of a wide variety of pyroclastic lithologies, rhyolite and dacite lavas, subordinate basic intrusions and plutonic granite. The extrusive rocks belong to the late Precambrian Uriconian Group, which was erupted between 566 and 560 Ma, the latter date being that of a large intrusive body, the Ercall Granophyre, considered to be the final product of Uriconian magmatism. The numerous important sections of the extrusive rocks are critical to understanding the nature of late Precambrian volcanism, and the tectonic setting in which it occurred. Most of the lithologies are pyroclastic in origin and of rhyolitic composition, in keeping with the violent, explosive character of Uriconian volcanism. The Primrose Hill Gneisses are of enigmatic origin, but may testify to locally intense shearing of these rocks at a later stage in the tectonic evolution of this region. The Uriconian rocks are overlain unconformably by early Cambrian strata, which are thus no older than 560 Ma, and could be younger than about 533 Ma.

LILLESHALL HILL (SJ 729 156)

POTENTIAL GCR SITE

J. N. Carney

Introduction

Lilleshall Hill has been proposed as a GCR site because it represents the northernmost outcrop of the Uriconian Group, whose main development occurs 10 km to the south-west, at the Wrekin range GCR site. Callaway (1879), in commenting on the general appearance and structure of the predominantly acidic tuffites exposed here, noted their probable Precambrian age and equivalence to the Wrekin volcanic rocks. The sequence was described in considerably greater detail, and studied petrographically, by Whitehead *et al.* (1928), who suggested that it mainly consisted of 'albite rhyolites' along with tuffs, breccias and possible lava flows. The Lilleshall site (Figure 5.6) occupies an important tectonic position since it is bounded by major displacements (Figure 5.1) forming the northerly continuation of the Church Stretton Fault System (e.g. Pauley, 1991). Structures affecting the Precambrian rocks are therefore likely to reflect early movements along the Welsh Borderland Fault System (Woodcock and Gibbons, 1988), a major Precambrian terrane boundary that also defines the western limit of the Midlands Microcraton basement block (Smith, 1987).

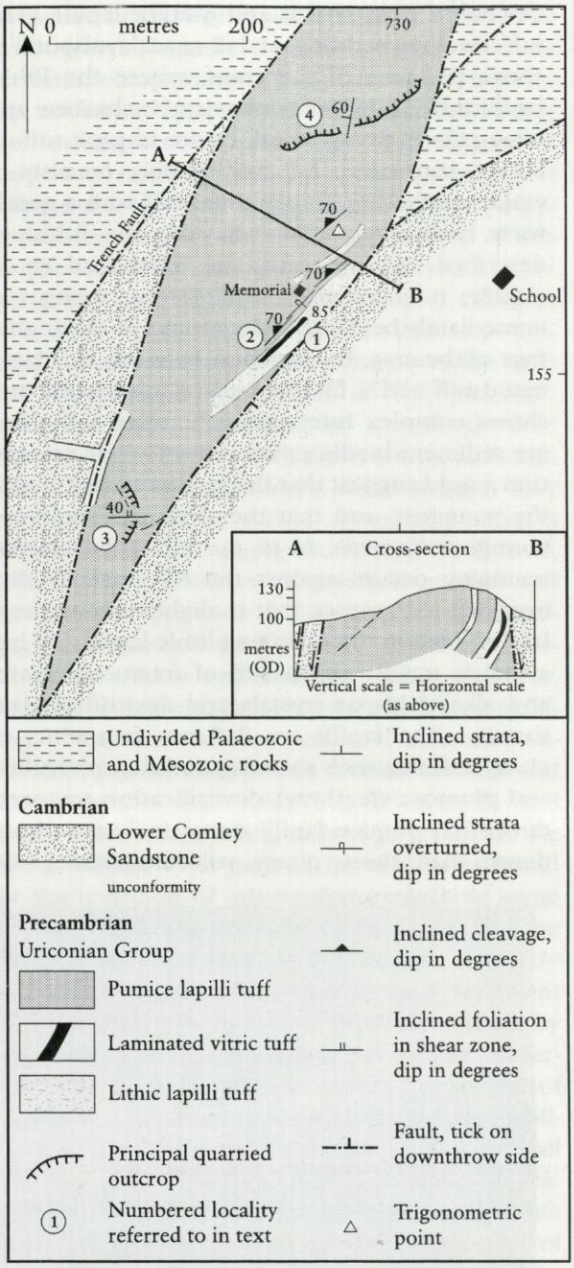

Figure 5.6 Geological sketch map of Lilleshall Hill, with cross section at lower right.

Description

The most accessible exposures are on the crags along the eastern margin of the hill, below and to the south of the Memorial. The stepped appearance of these crags resembles that produced by gently dipping bedding (Figure 5.7). On closer inspection, however, the sequence is sub-vertical to steeply eastward dipping, and the faces shown in this photograph correspond to the bases of beds (see below). The lower crags, at Locality 1 (Figure 5.6), consist of massive, pink to grey-weathering lithic-lapilli tuffs and breccias containing 20–30 per cent of angular to subangular, pale grey or pink glassy volcanic clasts averaging 5 mm across (maximum 30 mm), together with subordinate crystals of grey quartz. The matrix is grey to cream and coarse-grained, one thin section showing abundant devitrified glass shards and pumice lapilli. The tuff develops bedding, at 1–2 m scale, about 4 m above the base of the crags, where the lithic-lapilli tuff locally contains wispy inclusions and large rafts of fine-grained, cream to pink tuff. A highly distinctive, 2.2 m-thick bed, consists of white weathering, fine-grained tuff with a gently wavy lamination. This may be the lithology described as a vitric tuff by Whitehead *et al.* (1928); it is laterally impersistent, dying out immediately below the Memorial. In the higher part of the crag, the junction between this laminated tuff and a further bed of lithic-lapilli tuff shows complex intermixing textures. If these are sediment loading phenomena, their orientation would suggest that the lapilli tuff bed is also the youngest, and that therefore the sequence youngs westwards, in to the hill; this inferred younging occurs against the 85° easterly dip, implying a sequence that is slightly overturned. In thin section the overlying lithic-lapilli tuff has a matrix mainly composed of fractured quartz and alkali-feldspar crystals and devitrified glass shards. The lapilli are 2–5 mm fragments of glassy volcanic rock showing perlitic, spherulitic and plumose (feathery) devitrification textures; devitrified pumice lapilli occur in lesser abundance than these glassy volcanic clasts. An important feature of this tuff is that it fines upwards, in the inferred direction of younging, to a lithology containing diffusely-bounded layers of fine- to medium-grained tuff seen in pavement exposures at the crest of the crag.

Across the summit of the hill, at Locality 2, the sequence is mainly in unbedded, grey to pink tuffs, with a 'silty' surface texture, and with sporadic lithic-lapilli tuff layers up to several centimetres thick. Apart from the latter, these tuffs appear uniformly fine-grained, but polished rock slabs show that they are crammed with pale grey or cream, rounded to subangular lapilli between 2 and 8 mm in size. A thin section of similar rocks exposed at the Memorial shows that most of the shadowy fragments consist of devitrified pumice, interspersed with small quartz and alkali feldspar crystals, classifying the rock as a pumice lapilli tuff. This is the dominant type of lithology in other exposures and crags along the western parts of Lilleshall Hill.

The basic rocks exposed in a former quarry near the south-western extremity of the hill (Locality 3) are those described by Whitehead *et al.* (1928, pp. 10–11). In the south of the quarry they are dark green to grey, chloritic schists, overlying mylonitized siliceous tuffs within a major shear zone dipping 40° to the north.

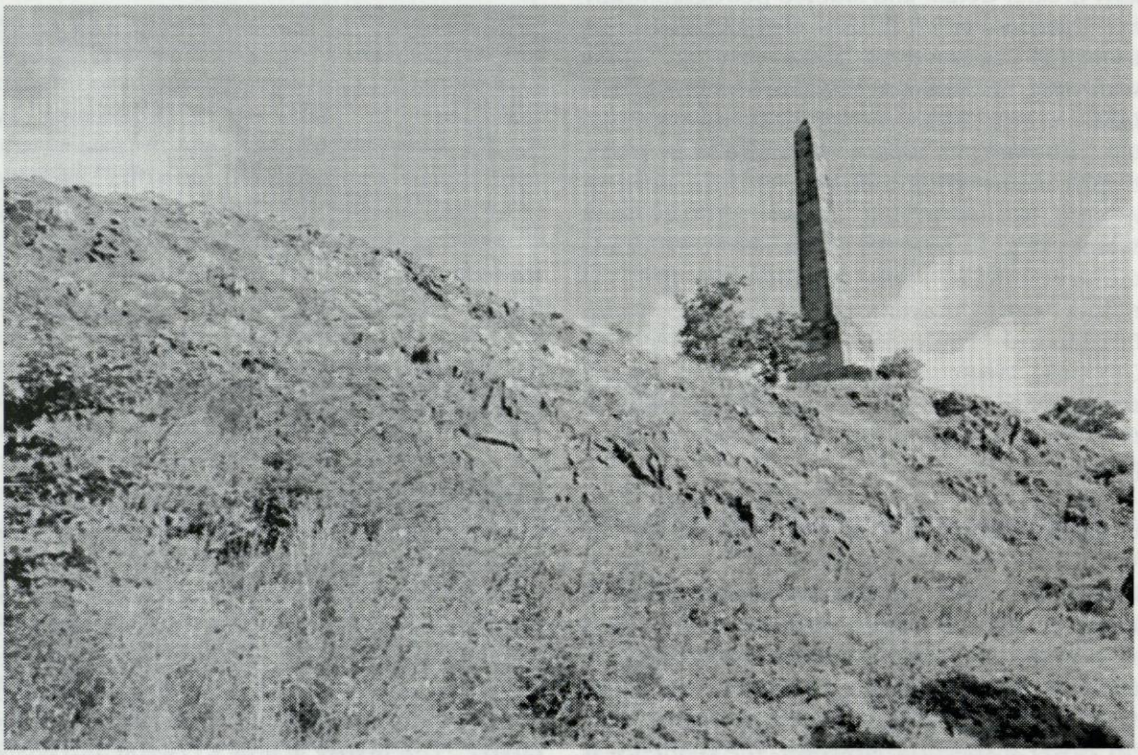

Figure 5.7 East-facing crags just below the Memorial at Lillieshall Hill, showing sub-vertical lithic lapilli tuffs of the Uriconian Group. (Photo: J.N. Carney.)

Farther north, the quarry exposes fine- to medium-grained, dark grey, quartz microgabbro, possibly the unsheared equivalent of the chloritic schists. At the north-western end of Lilleshall Hill (Locality 4), a cutting into a former quarried slope revealed *c.* 2 m of dark green-grey, coarse-textured chlorite-rich foliated basic rock intercalated with very fine-grained, pale grey, siliceous tuffs in a sequence dipping 60° west.

Bedding at Lilleshall Hill was originally described as being inclined at about 40° to the NNW (Callaway, 1879), a view accepted by Whitehead *et al.* (1928). The structure is more complex than this (Figure 5.6), and the dips previously reported in the vicinity of the Memorial may have been measured on particularly prominent fracture surfaces, which were wrongly identified as bedding planes. The eastern margin of the hill coincides with a zone in which bedding dips at 85° to the SSE. The evidence discussed above, that these beds also young to the west, is tenuous, but if confirmed would indicate that the sequence is overturned. Coinciding with this eastern sub-vertical zone is a highly pervasive cleavage (Figure 5.6), visible at millimetre to sub-millimetre scale, dipping at about 70° to the north-west (strike 054°). Vitric tuffs affected by this cleavage show boudin-like structures on crags immediately below (to the east of) the Memorial, and at a further locality 15 m west of the trigonometric point. On the western slopes of the hill the beds dip westwards and it is assumed that they are the right way up; no cleavage structure was noted here.

The major shear zone, originally noted by Whitehead *et al.* (1928), exposed in the quarry to the south-west (Locality 3), is characterized by an intense, north-dipping mylonitic foliation within siliceous tuffs. The foliation consists of millimetre-size, discontinuous quartz ribbons; it is also present in adjacent basic rocks, defined by chlorite laths. No elongation lineation is developed in either lithology, however. In the siliceous tuffs the foliation locally shows asymmetric drag folds indicative of a (top-down-to-north) normal movement component of the shear zone.

Interpretation

The Uriconian sequence at Lilleshall Hill includes a range of pyroclastic rocks suggestive of a highly explosive style of volcanism. The presence of alkali feldspar crystals in association with quartz suggests that the magmas were of rhyolitic composition. Based on tenuous way-up evidence, the *c.* 120 m-thick succession may be interpreted as part of a density-graded sequence characterized by a concentration of lithic material towards the base. The lithic-lapilli tuffs forming part of the lower *c.* 40 m of this sequence have matrixes rich in vitric constituents, but the lapilli size fraction contains abundant dense material (lithic clasts and crystals). In these compositional respects the tuffs are comparable to certain of the 'heavies'-enriched layers produced by pyroclastic flow activity (e.g. Wilson and Walker, 1982), although the absence of large blocks could suggest they are of a distal facies. A type of normal grading is suggested in the upper lithic-lapilli tuff bed, which contains layers of medium-grained tuff towards the top; such diffuse layering and grading has been described from subaerial pyroclastic flows (Fisher and Heiken, 1982); however, there is no textural evidence for significant welding. The discontinuous nature of laminated vitric tuff intercalations is due to their erosion and incorporation into overlying lithic-lapilli tuff beds, further supporting a flowage origin for the latter. These finer-grained intercalations could be the remnants of intervening pyroclastic surge deposits (e.g. Walker, 1984), but may also be accumulations of fine ash falls during relatively quiescent periods. The associated soft sediment deformation structures are possibly caused by the loading of coarse-grained tuff into an underlying fine-grained tuff bed, and are indicative of deposition on a wet substrate. The pumice lapilli tuffs forming the upper *c.* 80 m of the Lilleshall Hill sequence show apparent uniformity of texture, and absence of bedding or grading, which could suggest deposition in the distal parts of pumice flows (e.g. Perrotta and Scarpati, 1994). In the upper part of the sequence, microgabbro intrusive sheets are present, but there is no evidence for basic extrusive activity.

Lilleshall Hill provides a structural 'window' into early deformation that took place along the northern Welsh Borderlands Fault System. A heterogeneous style of deformation is suggested by bedding attitudes that steepen to become sub-vertical, or locally overturned, in the east of the outcrop, with the cleavage also becoming more intensely developed in that direction. These structures suggest that compression of the sequence was localized along one of the major faults defining the eastern margin of the Precambrian inlier. This event could be as young as Acadian (Siluro–Devonian), but the deformation is also of a similar style to that affecting the Longmyndian and Uriconian rocks of the Church Stretton area (Pauley, 1991), and which has been dated to the latest Precambrian or early Cambrian (see the introduction to this chapter). Cleavage formation could therefore be related to one of the major late Precambrian terrane accretion events along the Welsh Borderlands Fault System (e.g. Gibbons and Horák, 1996). Further complexity is suggested by the discrete shear zone, trending across the southern part of the outcrop, which shows evidence for a phase of normal faulting.

Conclusions

Lilleshall Hill contains many informative sections through about 120 m of massive to thickly bedded Uriconian acidic pyroclastic rocks. Most lithologies are probably pumice lapilli tuffs with rhyolitic compositions, suggesting an origin by highly explosive volcanic eruptions. The succession shows an upward diminution in the content of relatively dense constituents (volcanic rock fragments and crystals), as opposed to pumice, recalling a trend that is commonly seen in deposits formed by pyroclastic flows. This may only be part of the story, however, because the few horizons with sedimentary structures suggest that deposition occurred on a wet substrate. Basic rocks are restricted to minor developments of chloritic tuffs or microgabbro intrusions, locally converted to chlorite schists, in the upper (i.e. westernmost) part of the sequence. Following deposition, the succession was affected by unevenly-distributed deformation involving westwards tilting and the progressive steepening of beds, which became sub-vertical to locally overturned, and penetratively cleaved, on the eastern flank of the inlier. The age of this deformation is not known but is possibly Precambrian, related to compression along the Welsh Borderlands Fault System.

LYD HOLE (SJ 414 055)

D. Wilson

Introduction

The stream section at Lyd Hole is of particular importance as being one of the few places in the Welsh Borderlands where a contact between Uriconian and Longmyndian rocks is exposed. It also includes excellent exposures of volcanic and sedimentary rocks in the river bed and along the steep sides of the gorge, although parts of the section are more accessible during periods of low water. The Uriconian-Longmyndian junction is controversial, having attracted the interest of a succession of researchers who have provided a number of interpretations discussed in the introduction to this chapter.

The earliest descriptions of the section include those of Callaway (1882), who believed the junction between the Uriconian and Longmyndian to be a fault, and Blake (1890) who considered it an intrusive contact. Pocock *et al.* (1938) subsequently decided that the contact showed some evidence of disturbance, but regarded it as essentially conformable. These ideas were partly influenced by regional considerations of the relationships between the various Uriconian outcrops and their association with Longmyndian sediments. The presence of Uriconian volcanic fragments within the Longmyndian demonstrated that the latter were younger, but general acceptance that the disparate outcrops of lithologically similar Uriconian rocks form part of the same volcanic complex came only gradually (cf. Blake, 1890; Lapworth and Watts, 1910). Work by James (1952, 1956) finally established that Uriconian rocks underlie the Longmyndian succession east and west of the Long Mynd, and that the volcanic sequences were everywhere older than the Longmyndian sediments. It follows that the relationship of the Western Uriconian to the Longmyndian, such as that displayed at Lyd Hole, was not an intrusive contact but must be either an unconformity or a fault.

Description

The Uriconian rocks below the waterfalls at Lyd Hole mainly consist of a sequence of ochreous-weathered and silicified acid tuffs and spherulitic rhyolites, separated by a fine-grained, altered vesicular basic rock. The rhyolites are massive, reddish brown, flow-banded lavas, largely devitrified and somewhat ferruginous, with numerous amygdaloidal silica segregations in places, which probably fill vesicles (Figure 5.8, Locality 1). Spherulites of microcrystalline feldspar, concentrated along the foliation,

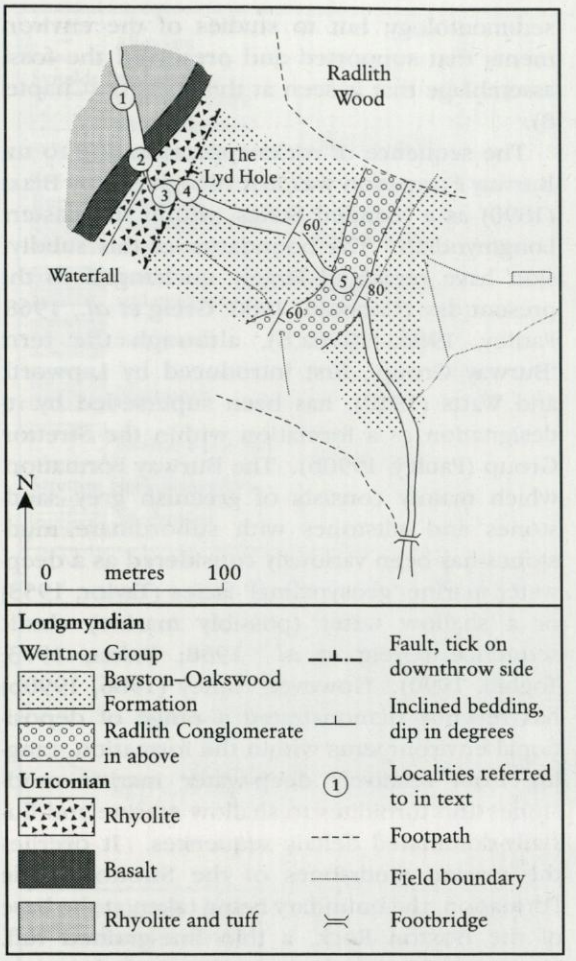

Figure 5.8 Geological map of the Lyd Hole site.

weather out to give the rock a nodular appearance, well demonstrated at the waterfalls (Locality 3). The pyroclastic rocks are vitric-lapilli tuffs which, in thin section, locally reveal a partially welded fabric enclosing rounded and altered igneous lithoclasts. The basalt, below the lower waterfall (Locality 2), appears to form part of the volcanic succession (Pocock *et al.*, 1938) although it is highly altered and its contacts are not immediately obvious. In thin section it consists of a microcrystalline groundmass of albitized oligoclase feldspar and glass, with vesicles mainly containing feldspar, quartz and chlorite aggregates.

Immediately above the waterfalls, Longmyndian sandstones overlie the Uriconian rocks, the junction being straight, sharp and approximately vertical, striking at about 015°. The rocks on either side show local evidence of movement, with a small amount of brecciation and quartz veining resulting in a poorly developed fault gouge; slickensides also occur along the contact and in the adjacent Uriconian rocks (Locality 4). Dean and Dineley (1961) have commented on a small disparity that occurs between the contact and the strike of bedding in the overlying Longmyndian sandstones.

The Longmyndian strata above the contact are representatives of the Bayston–Oakswood Formation (Pocock *et al.*, 1938; Dunning, 1975), the lowermost division of the western Longmyndian Wentnor Group. They comprise a succession of up to 120 m of dull purplish red, medium- to coarse-grained sandstones (lithic arenites), containing a significant proportion of volcanic material, in beds up to 1 m thick, with local thin micaceous partings. About 100 m upstream from the waterfall they are succeeded by a sequence of at least 25 m of quartz-rich, clast-supported granule-pebble conglomerates, in planar beds up to 0.2 m thick, with subordinate coarse, locally cross-bedded, pebbly volcaniclastic sandstones (Locality 5). This division has been named the Radlith Conglomerate (Pocock *et al.*, 1938), the lowest of three major conglomeratic units that punctuate the Bayston–Oakswood Formation in this area. The top of the conglomerate is unexposed, although the overlying sandstones reappear in the river near the footbridge.

Interpretation

The Uriconian rocks of Lyd Hole compare with those in other parts of the Welsh Borderlands and demonstrate the explosive, probably subaerial volcanism that characterized this succession. The basaltic rock was regarded by Pocock *et al.* (1938) as a lava within the eruptive sequence, although it is identical to other basic rocks on nearby Pontesford Hill which were considered as intrusive. Despite its equivocal contacts, it is considered more likely that the basalt represents a later intrusion into the Uriconian rocks, although there is no reason to suppose it is much younger. Indeed Pocock *et al.* (1938, p. 34) show this and other rocks at Lyd Hole to be overlain unconformably by Ordovician strata. It is likely that it forms part of a widely recognized suite of late Precambrian Uriconian basic intrusions (James, 1956; Greig *et al.*, 1968).

The boundary between the Uriconian lavas and Longmyndian strata is clearly a zone of movement. Pocock *et al.* (1938) considered that there must have been some movement along the junction, but that it did not amount to very much. The small obliquity of the contact to the strike of bedding in the basal Longmyndian, noticed by Dean and Dineley (1961), is strongly suggestive of a normal fault although, in the absence of any definitive evidence, it is difficult to say how much movement has occurred on the structure.

The overlying Bayston–Oakswood Formation is representative of the braided fluvial sedimentary facies that extended across the region during the early part of Wentnor Group (Pauley, 1990b). The Radlith Conglomerate is one of a series of conglomerates that probably developed as longitudinal bars in these high-energy environments. The lateral extent of these deposits suggests that they formed on an alluvial braidplain rather than within a restricted fluvial system.

Conclusions

Lyd Hole is an informative site that provides an opportunity to study a rare contact between Uriconian and Longmyndian Wentnor Group rocks. Although the contact appears to be faulted, the magnitude of the displacement is unknown. The site displays important sections through the otherwise poorly exposed and much faulted 'Western Uriconian' volcanic sequence, as well as providing a good reference section through fluvial sedimentary rocks of the western Longmyndian, Bayston–Oakswood Formation.

ASHES HOLLOW–DEVIL'S MOUTH (SO 435 930; SO 440 942)

D. Wilson

Introduction

Ashes Hollow and Devil's Mouth, both of which fall within the existing Long Mynd SSSI, have been selected as GCR sites as they jointly provide a complete section through the sedimentary rocks representative of the lower part of the Longmyndian Stretton Group. The sections (Figure 5.9) record the transition from the marine environments that dominated the early period of Longmyndian sedimentation in the Stretton Shale and Burway formations, to the alluvial sequences that first appear in the overlying Synalds Formation. This progression is of great importance not only to late Precambrian sedimentology, but to studies of the environments that supported and preserved the fossil assemblage that is seen at this locality (Chapter 8).

The sequence of rocks approximating to the Burway Formation was first recognized by Blake (1890) as a 'Banded Series' within the Eastern Longmyndian. The boundaries of this subdivision have remained largely unchanged to the present day (Cobbold, 1900; Greig *et al.*, 1968; Pauley, 1986, 1990a,b), although the term 'Burway Group', first introduced by Lapworth and Watts (1910), has been superseded by its designation as a formation within the Stretton Group (Pauley, 1990b). The Burway Formation, which mainly consists of greenish grey sandstones and siltstones with subordinate mudstones has been variously considered as a deep-water marine 'geosynclinal' facies (Taylor, 1958) or a shallow water (possibly marine) clastic sequence (Greig *et al.*, 1968; Baker, 1973; Toghill, 1990). However, Pauley (1986, 1990b) has recently demonstrated a range of depositional environments within the formation, ranging from relatively deep-water marine mudstones and turbidites to shallow marine and fluvially-dominated deltaic sequences. It overlies the marine mudstones of the Stretton Shale Formation, the boundary being taken at the base of the Buxton Rock, a thin fine-grained tuff. However, the lithological transition does not always coincide with this tuff horizon, and it is common for the thinly laminated siltstone-mudstone facies of the Stretton Shale Formation to persist into the lower part of the Burway Formation. The top of the formation is taken at the top of the Cardingmill Grit, above which there is a generally sharp transition into the purplish red-brown mudstones of the Synalds Formation.

Description

The steep valley sides of Ashes Hollow provide a virtually complete section through the Burway Formation, largely undisturbed by faulting except in the upper part, and estimated here to be about 640 m thick. The beds generally dip steeply westwards at angles between 60° and 90° and also young to the west, as determined by sedimentary structures. The highest beds of the underlying Stretton Shale Formation, consisting of weathered, dark greenish grey, laminated mudstones, crop out by the footbridge next to

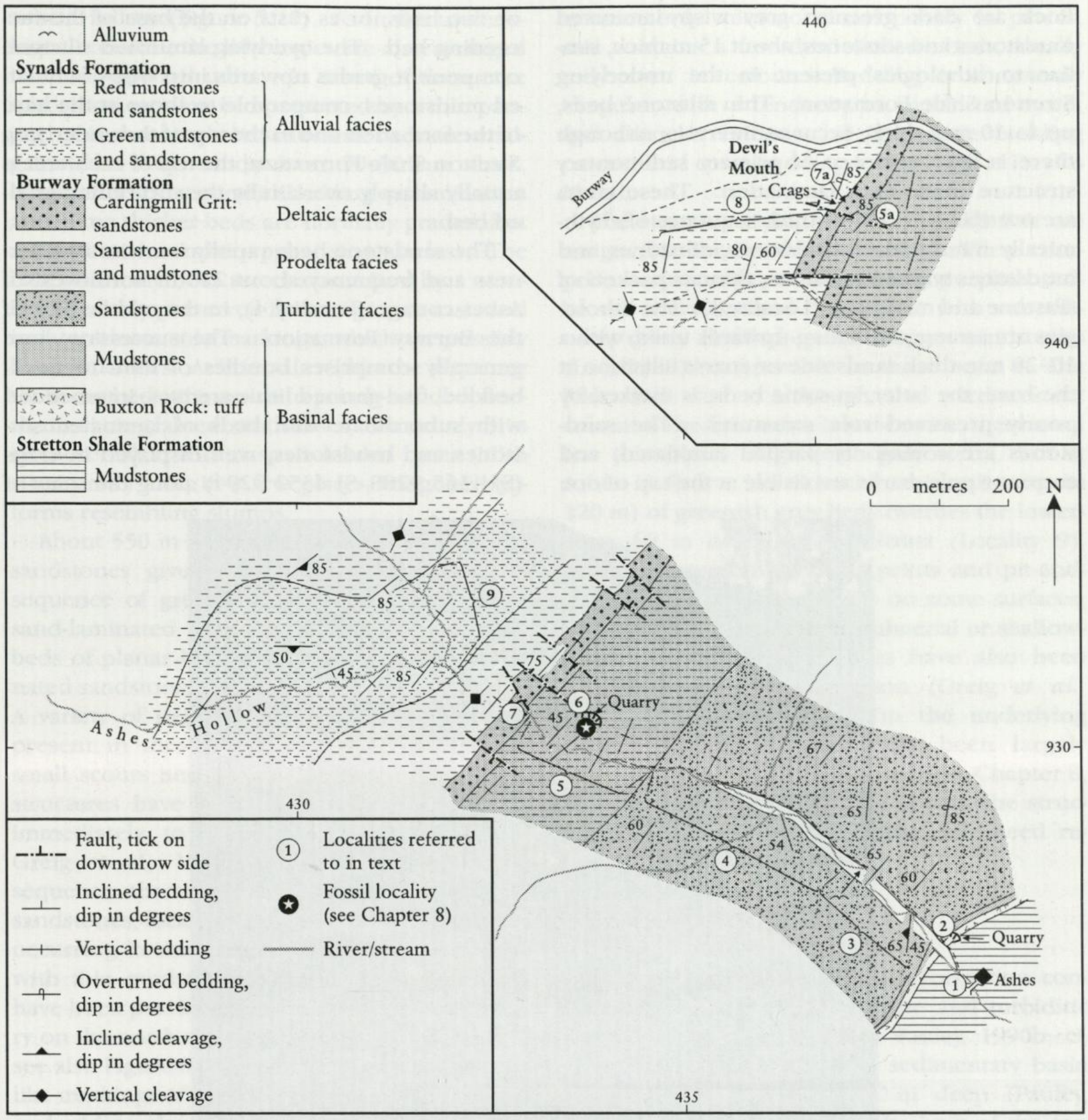

Figure 5.9 Geological map of the Ashes Hollow and Devil's Mouth sites. The range of sedimentary environments in this part of the Stretton Group is indicated on the explanation at top left.

the abandoned Ashes cottage (Figure 5.9, Locality 1).

About 70 m north-west of Ashes cottage, a small quarry (SJ 4392 9267) (Locality 2) exposes the local equivalent of the Buxton Rock, the lowermost member of the Burway Formation. It comprises 4–5 m of a pale greenish grey, highly silicified, fine-grained tuff in beds 20–50 mm thick, but locally up to 0.1 m, with subordinate thin shaly mudstone partings. Thin sections show the tuff to be generally composed of a groundmass of clay minerals, chlorite, quartz and feldspar in which there are a few aggregates of albite feldspar or radially arranged chlorite, both possibly of authigenic origin, together with some small scattered feldspar pyroclasts; geochemical analyses indicate a rhyolitic composition (Greig *et al.*, 1968). The quarry face displays a small thrust duplex that has disrupted the tuff, but its gradational top is exposed in adjacent crags and characterized by alternations of thin (10–20 mm), porcellaneous tuff beds, shaly mudstones and siltstones.

The beds immediately overlying the Buxton

Rock are dark greenish grey wispy-laminated mudstones and siltstones, about 15 m thick, similar to lithologies present in the underlying Stretton Shale Formation. Thin siltstone beds, up to 10 mm thick, occur at intervals, although there is little evidence of primary sedimentary structure other than lamination. These strata are overlain by a 150 m-thick sequence of rhythmically interbedded sandstones, siltstones and mudstones with subordinate, thinner packets of siltstone and mudstone (Locality 3). The lithologies are arranged in fining-upwards units, with a 10–20 mm-thick sandstone or coarse siltstone at the base; the latter, in some beds, is marked by poorly preserved sole structures. The sandstones are commonly parallel laminated, and current ripple marks are visible at the top of one or two beds, or as casts on the base of the succeeding bed. The overlying laminated siltstone component grades upwards into wispy laminated mudstones, comparable to those at the base of the formation and in the top of the underlying Stretton Shale Formation; the top of each unit is usually sharply overlain by the succeeding graded bed.

The sandstone beds rapidly increase in thickness and frequency about 250 m north-west of Ashes cottage (Locality 4), in the middle part of the Burway Formation. The succession here generally comprises bundles of thin- to thick-bedded, fine-grained lithic arenites, interbedded with subordinate thin beds of laminated siltstones and mudstones, well displayed in crags (SJ 4365 9288–SJ 4352 9294) along the sides of

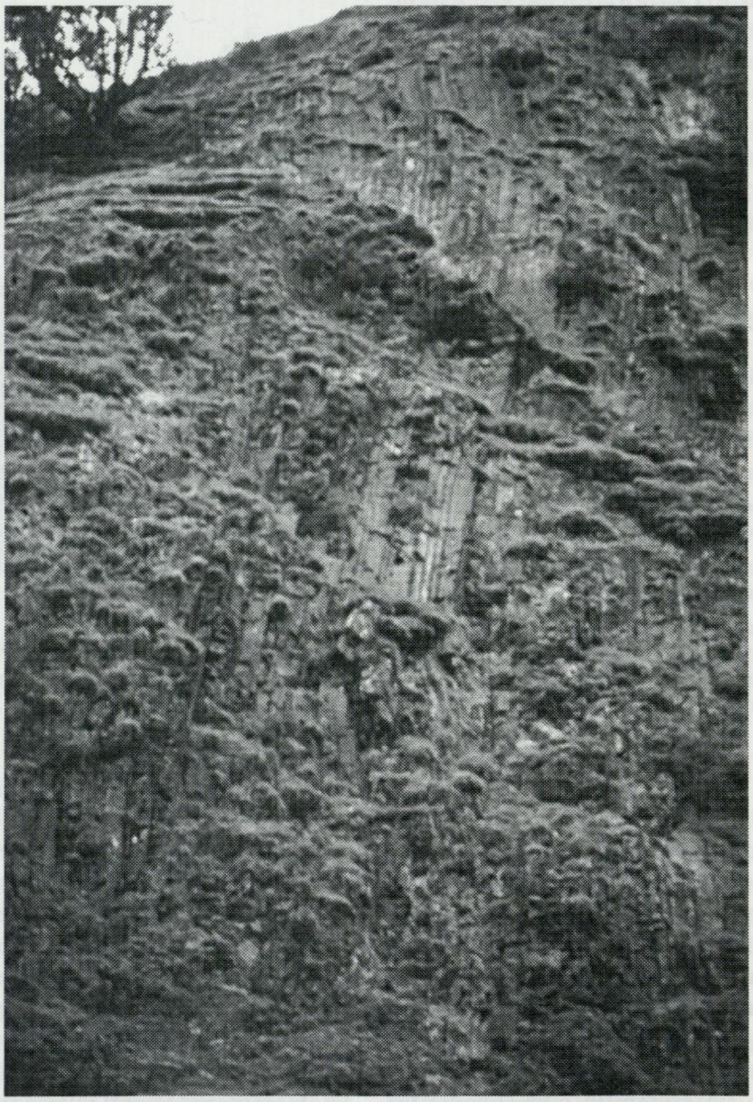

Figure 5.10 Crags along Ashes Hollow showing vertical beds of rhythmically bedded, thick to thin-bedded turbiditic sandstones and siltstones of the Burway Formation. (Photo: D. Wilson.)

Ashes Hollow (Figure 5.10). As in the underlying succession, the lithologies are also arranged in fining-upward units. In some of these, the mudstone capping is represented by a thin parting, or is absent where sandstone beds are amalgamated. The sandstones, 20–30 mm thick, commonly show parallel bedding or lamination and some thicker beds are normally graded, but a substantial number appear structureless. The bases of many beds are characterized by a number of poorly preserved flute or groove casts and indeterminate sole markings, which are possibly load structures. At intervals within the sequence there are beds of massive to poorly bedded, fine- to medium-grained sandstones, commonly 3 m or more thick, with cuspate loaded bases ('dish structures') and, in a few places, irregular bedforms resembling slumps.

About 550 m north-west of Ashes cottage, the sandstones give way rapidly upwards into a sequence of greenish grey colour-banded and sand-laminated siltstones and mudstones with beds of planar laminated and ripple cross-laminated sandstones up to 50 mm thick (Locality 5). A variety of delicate sedimentary structures is present in these rocks, including microfaults, small scours and graded bedding; comparable structures have been recorded from the area immediately to the north-east (Salter, 1857; Greig *et al.*, 1968). At intervals within the sequence there are fine- to medium-grained sandstones, locally parallel- and cross-laminated, occurring in amalgamated beds up to 3 m thick with thin mudstone partings. Four such beds have been previously worked from a small quarry on the north side of Ashes Hollow (Locality 6; see also Figure 8.9a). Here, unusual ring- or pit-like markings of possible organic origin are seen on bedding planes, and are described in Chapter 8 (Cope, this volume).

Beds of fine- to medium-grained, parallel- and cross-laminated sandstone, containing appreciable amounts of mica, appear towards the top of the succession. Up to 0.3 m thick, such beds are particularly noticeable in the crags of Devil's Mouth, north-east of Ashes Hollow (Locality 5a), interbedded with disrupted and possibly slumped siltstones containing small sandstone clasts. They are overlain by the Cardingmill Grit, the highest member of the Burway Formation, which is well exposed although much faulted in Ashes Hollow and at Devil's Mouth (Locality 7, 7a). The grit is a massive to fairly well-bedded, coarse-grained, lithic arenite about 30 m thick; it is commonly micaceous and pebbly in places, and contains a few thin siltstone beds, locally red-brown in colour. Bedding is commonly irregular, with a tendency to pinch and swell, and there is evidence of large-scale bedding cutouts and cross-bedding in the crags at Devil's Mouth.

The strata overlying the Cardingmill Grit are assigned to the Synalds Formation. These comprise a succession of alternating purplish red-brown siltstones and mudstones with subordinate, thinly bedded, purplish grey sandstones (Locality 8), comparable to those seen at The Pike GCR site. Although the predominant colour of the formation is red-brown, there are a few thin layers of greenish grey strata, and at Ashes Hollow a considerable thickness (up to 120 m) of greenish grey beds overlies the lowermost 60 m of red-brown strata (Locality 9). Features resembling rain imprints and pit-and-mound structures are visible on some surfaces, and a range of comparable subaerial or shallow-water sedimentary structures have also been recorded from the formation (Greig *et al.*, 1968). Similar structures in the underlying Burway Formation have now been largely accorded a biogenic origin (see Cope, Chapter 8, this volume) and in the light of this, the structures in the Synalds Formation may need re-evaluation.

Interpretation

The lower part of the Burway Formation contains facies indicative of marine and turbiditic sedimentation (Taylor, 1958; Pauley, 1990b; cf. Greig *et al.*, 1968) within a sedimentary basin estimated to be about 700 m deep (Pauley, 1990a). The mudstones at the base of the formation represent a continuation of the basinal sedimentation that typified the deposition of the Stretton Shale Formation. These wispy-laminated sediments are comparable to hemipelagic sequences (Stow and Piper, 1984; Davies *et al.*, 1997), representing the terrigenous muddy background sediment of slope apron and turbidite systems, deposited on a basin-plain largely beyond the influence of current activity. The locally interbedded thin silt beds represent fine-grained turbidite flows (Stow and Piper, 1984).

The presence of an interbedded subaqueous tuff, the Buxton Rock, at the base of the Burway Formation, is an indicator of the sporadic volcanicity that accompanied sedimentation.

Although this tuff is a significant marker horizon, it is debatable whether it represents an appropriate formational boundary as it does not coincide exactly with any major lithological change (cf. Greig *et al.*, 1968). Undoubtedly, the most important factor in its preservation has been its deposition within a hemipelagic facies, with vertical accretion as the most significant process and an absence of current winnowing. The Buxton Rock and other volcanic horizons within the Longmyndian succession have been used as evidence of continuing Uriconian volcanicity (e.g. Greig *et al.*, 1968; Pauley, 1990b). Baker (1973), however, considered that these discrete volcanic events were linked to fault movements not connected directly to Uriconian activity. Whatever the reason, it is possible that this volcanicity was more extensive than has been previously considered, in view of the low preservation potential of much of the Longmyndian succession.

The sandstones that appear in the Burway Formation a short distance above the Buxton Rock are interpreted as a turbiditic facies, each sandstone–siltstone–mudstone unit representing an individual turbidite flow, which deposited a fining-upward sequence as current velocity waned (Bouma, 1962; Walker, 1967; Pauley, 1990b). The types of sedimentary structure within both the thick- and thin-bedded turbidite units are similar, suggesting that broadly comparable types of flow deposited them; the thicker turbidite units are considered to represent flows of greater magnitude. There is, however, no corresponding increase in the coarseness of the sand fraction within these thicker flows, and the common occurrence of fine-grained sandstones suggests that all the turbidites were derived from a common source depleted in coarse material. The range of sedimentary structures within the turbidite units are typical of the complex sequence of divisions ($T_{a\text{-}e}$) found in classical Bouma-style turbidite units (Bouma, 1962) overlying a basal current-scoured surface with flute and groove casts. The normal grading displayed by a few sandstones represents the typical Bouma (T_a) division, recording relatively rapid deposition from a turbulent flow with an initial high coarse-fraction concentration. The structureless sandstones are somewhat different and may be compared to high-density flows in which sediment concentration and sediment support mechanisms, rather than fluid turbulence, was the more important factor in their transportation (Lowe, 1982). Deposition of these flows may have resulted from rapid 'freezing' of the flow at critical velocities.

Parallel lamination, representing the typical Bouma (T_b) division, is the commonest structure within the sandstones, indicating that the majority of flows were deposited by traction under upper stage, plane-bed, flow conditions (Bouma, 1962; Walker, 1967). Lower flow conditions as indicated by the ripple cross-laminated Bouma (T_c) division is only rarely developed, suggesting that reworking of the top of the flow by the tail of the turbidity current was limited. The overlying siltstone–mudstone units are interpreted as Bouma (T_d) and (T_e) divisions respectively, which have been deposited from suspension. The wispy lamination within the mudstones is comparable to the hemipelagic sedimentation that characterizes the lowest part of the formation. The poor development locally of the T_d and T_e divisions, and the amalgamation of some sandstone beds within the middle part of the formation, suggests that many flows approximate to top-cut-out flow units deposited by high-flow regime turbidity currents (Walker, 1967; Stow, 1986).

The thickly bedded, massive sandstones, found at intervals within the turbiditic sequence have been interpreted as channel fills (Pauley, 1990b). They do not appear to be adequately covered by the Bouma (1962) scheme, but may be comparable to the coarse-grained deposits described by Lowe (1982) and Davies *et al.* (1997). They appear to have been deposited from high-concentration turbidity currents, the dominantly structureless sandstone units recording rapid sediment fall-out from high-density suspension. The resultant loosely packed sediments were probably susceptible to post-depositional liquefaction and movement, as recorded by slumping and water expulsion features such as dish structures.

The thin-bedded sandstone turbidites, which alternate with units of hemipelagic mudstone, are considered to be a relatively distal facies deposited in an outer fan environment. The thick-bedded turbidites in the middle part of the Burway Formation are comparable to facies deposited in a mid-fan setting. The succession from laminated mudstones with thin fine-grained turbidites to an increasingly sandstone turbidite-dominated sequence in the middle part of the Burway Formation is probably the result of turbidite fan progradation (Walker, 1978). However, the lack of coarsening upwards and gradation between the thin- and thick-bedded turbidite sequences, and the absence of thickening-upward cycles within these facies, led Pauley (1990b) to suggest that the succession does not conform to the classical Walker (1978) model of fan progradation. He favoured instead their deposition as turbidite sheets supplied from multiple small sources rather than a single large system; these sources were considered to be channels on a prodelta slope (Pauley, 1990b).

The siltstones and mudstones that succeed the turbidites of the Burway Formation have been interpreted as delta front and delta slope deposits, which were influenced by fluvially-generated bottom currents (Pauley, 1990b). The thicker sandstones that occur at intervals within the sequence may represent periodic delta advance. Progradation of the delta (the 'Cardingmill Delta' of Pauley, 1990b) is recorded by the increasing frequency and thickness of cross-laminated sandstones within the upper part of the formation, culminating in deposition of the Cardingmill Grit. The latter has been interpreted as a shallow marine deposit (Greig *et al.*, 1968), possibly a marine bar, although it was considered more likely to be a fluvial distributary channel fill of the delta by Pauley (1990b), from the absence of any wave-dominated sedimentary structures. Palaeocurrent evidence suggests that the distributary channels flowed in a general north-easterly direction.

The succeeding Synalds Formation, like that observed at The Pike and Long Batch GCR sites, shows a range of structures and facies indicative of the alluvial floodplain environments that prograded across the Cardingmill Delta.

Conclusions

Ashes Hollow provides one of the most complete and well-exposed sections through the Burway Formation, and is highly instructive in showing the evolution of sedimentation within a diverse group of marine beds laid down in basin plain, turbiditic and deltaic environments. It is also one of the few places within the Longmyndian Supergroup that documents the progression from beds accumulated in deep marine environments to those characteristic of fluvial conditions, represented by the Synalds Formation. Not only do these exposures give insights into the early sedimentary history of the Longmyndian basin, but they also provide a palaeoenvironmental framework for the possible organic remains found in these beds (Chapter 8).

THE PIKE (SO 442 950)

D. Wilson

Introduction

The sharp spur that forms The Pike rises between small tributaries on the northern side of the Carding Mill Brook, immediately north-west of Bodbury Hill (SJ 4455 9480). The ridge has been selected as a GCR site (Figure 5.11) because it contains one of the most informative and well-exposed sections through the Synalds Formation, part of a succession of late Precambrian (Longmyndian) fluvial sediments. As such it provides an important reference to other GCR sites within the thick Longmyndian sequence; the site falls within the existing Long Mynd SSSI.

The Synalds Formation forms part of the Eastern Longmyndian Stretton Group. It was first recorded by Blake (1890; his 'Purple Slate' group), although the term 'Synalds Group' for

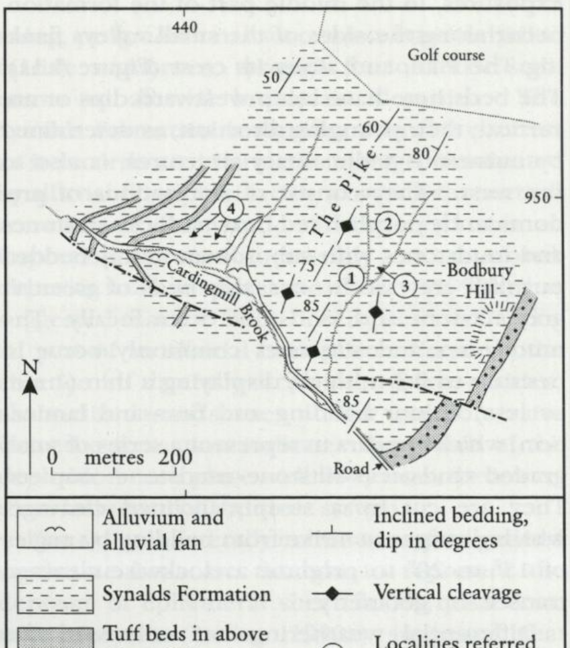

Figure 5.11 Geological map of The Pike site.

rocks approximating to this division was introduced by Lapworth and Watts (1910); the sequence was given formational status by Pauley (1990b). The formation comprises up to 850 m of mudstones and siltstones with beds of sandstone. Its base is taken at the top of the Cardingmill Grit, the highest member of the underlying Burway Formation. A series of tuff horizons, the Batch Volcanic Beds (Cobbold, 1900; Greig *et al.*, 1968), occur in the upper part of the sequence at this site. The strata have been described in considerable detail by Greig *et al.* (1968), who considered that they formed in shallow water depositional environments. Modern sedimentological analysis of the formation has been undertaken by Pauley (1986; 1990a,b; 1991), who has suggested that they represent a sequence of fluviatile sediments within the Longmyndian.

Description

The base of the Synalds Formation crosses Bodbury Hill, a short distance east of The Pike. The north-western slope of the hill approximates to a dip-slope in the lowermost beds, which are, however, poorly exposed. The best exposures, in the middle part of the formation, occur along the sides of the small valleys flanking The Pike, and along its crest (Figure 5.11). The beds here have steep westward dips or are vertical; their younging direction, as determined by numerous sedimentary structures, is also to the west. They consist of alternations of predominantly purplish red-brown blocky siltstones and mudstones with subordinate thinly bedded purplish grey sandstones; thin beds of greenish grey siltstone and sandstone occur locally. The mudstones and siltstones commonly occur in beds up to 0.3 m thick, displaying a thin (1 mm or less) colour banding and fine-sand lamination, which appears to represent a series of small graded sandstone–siltstone–mudstone couplets. They are cut by a steeply inclined cleavage, which diverges in strike from bedding by angles of 15° to 20° to produce a clockwise cleavage transection geometry.

Differential weathering has enhanced the lithological contrasts, so that the sandstones form prominent ribs in most exposures; notable examples are displayed in crags (SJ 4425 9488 to 4435 9506) on the eastern side of The Pike (Figure 5.12). The sandstones generally occur as laterally persistent, sharp-based, tabular or wavy beds, from 10 mm to 0.15 m thick, in places in bundles up to several metres thick. They are mostly fine- to medium-grained lithic arenites composed largely of volcanic material, with an appreciable quartz component and a small amount of mica. Many sandstones are planar laminated and, in crags on the south-eastern side of The Pike (Figure 5.11, Locality 1), examples of small-scale, low-angle ripple cross-lamination are visible in one or two places. A few thick beds, up to 0.75 m, of massive, fine- to medium-grained, somewhat discontinuous sandstone occur at intervals within the succession; a typical example crops out over 30 m across a ridge east of The Pike (Locality 2). These sandstones show little evidence of internal structure, other than a local faint lamination and possible trough cross-bedding. Bedding is usually planar but, in places, the basal surface is erosional, truncating the underlying siltstone and mudstone laminae at a shallow angle.

A conspicuous feature of the Synalds Formation is the local abundance of small (1–5 mm) circular or ellipsoidal pits on the upper surfaces of mudstone and siltstone beds. They are readily observed on bedding surfaces on the eastern slopes of The Pike (Locality 3) and in crags (Locality 4) to the west, in the upper part of the formation. The indentations are probably gravitational load structures, or may be due to the escape of entrapped air, but Greig *et al.* (1968) have recognized other types that may be raindrop imprints or wave-foam bubble impressions. The long axes of the ellipsoidal pits are commonly orientated with cleavage, and may represent formerly circular imprints that have undergone deformation, rather than an oblique impact phenomenon (cf. Salter, 1857; James, 1956). They locally occur with, and are cut by, small branching grooves akin to rill marks. Desiccation cracks have also been reported from other parts of the Synalds Formation crop (Salter, 1857; Greig *et al.*, 1968).

The Batch Volcanic Beds, consisting of two or three horizons of coarse-grained, intermediate, crystal ash and lapilli tuff, crop out on the sides of the Carding Mill Valley, about 350 m west of The Pike. They are described fully in the section on Jonathan's Hollow–Long Batch GCR site.

Interpretation

The common occurrence of rain pits, rill marks, desiccation cracks and allied sedimentary struc-

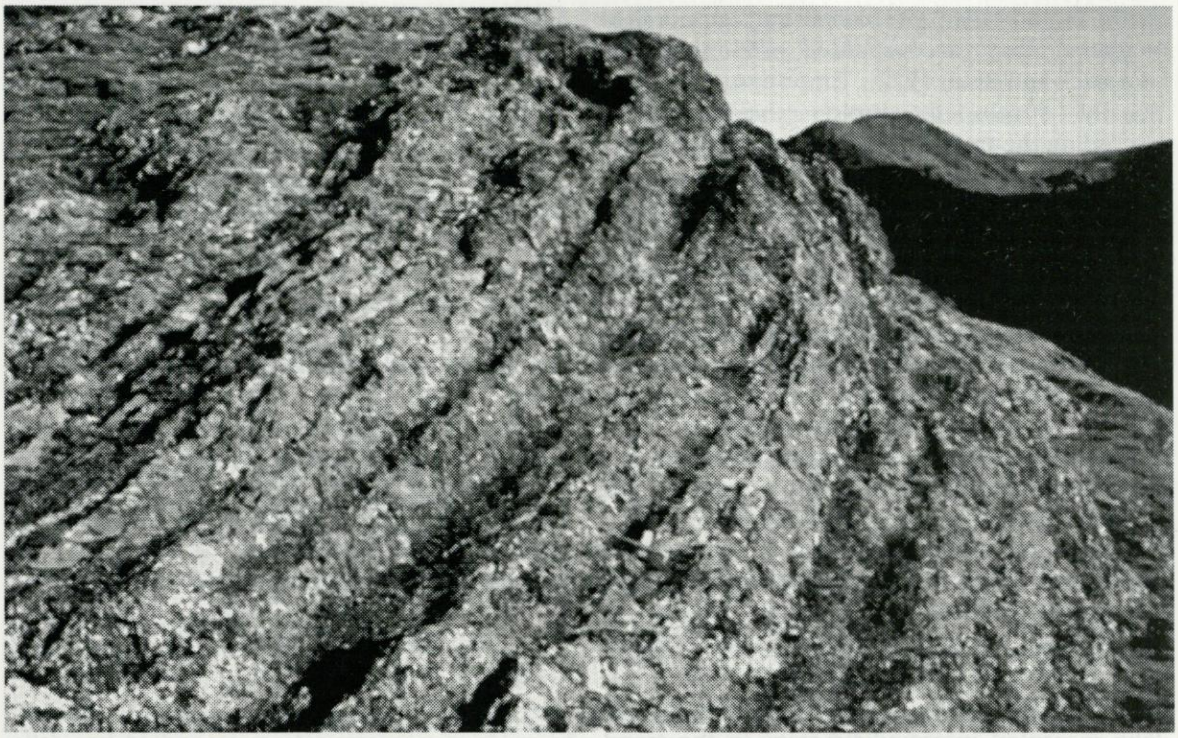

Figure 5.12 Differential erosion in crags on the east side of The Pike, showing alternations of sheeted sandstones, siltstones and mudstones typical of the Synalds Formation. (Photo: D. Wilson.)

tures within the Synalds Formation (Greig *et al.*, 1968) point to a succession that was subject to repeated subaerial exposure. The widespread red coloration of the mudstones and siltstones may be an early diagenetic feature (Walker *et al.*, 1978; Besley and Turner, 1983), indicative of sediments that underwent oxidation within the weathering profile, above a lowered water table. These are features characteristic of sediments deposited in fluvial or paralic (deltaic, estuarine, tidal flat) environments, although they were at one time regarded with the rest of the eastern Longmyndian, as an essentially deep water, turbiditic ('flysch') facies (Taylor, 1958).

Greig *et al.* (1968) were the first to propose that the majority of Longmyndian sediments were deposited in shallow water, probably in a tidal estuary or delta; they assumed, as had Cobbold and Whittard (1935), that deposition immediately followed, or was partly coeval with, the eruption of the Uriconian volcanic sequence. Most subsequent authors have followed this view (e.g. Baker, 1973; Toghill and Schell, 1984). However, recent work (Pauley, 1990a,b; 1991) has demonstrated that Longmyndian stratigraphy is characterized by a broadly upwards-coarsening, progradational succession from turbiditic marine sequences, through fluvio-deltaic sediments to alluvial braidplain deposits. The Synalds Formation is significant in this context, as it records an intermediate stage, dominated by alluvial floodplain facies, in which the vertical accretion of fine sand, silt and mud was an important process, possibly as overbank deposits from suspension. Pauley (1990b) has suggested that the formation represents the distal part of an alluvial braidplain. Certainly, the absence of laterally accreted, coarse-grained, channelized sandstones and/or conglomerates from the formation indicates that much of the deposition took place well away from the immediate influence of the river system.

The thinner, laterally continuous, fine-grained sandstone beds are thought to represent the deposits of ephemeral sheet floods; palaeocurrent analyses (Pauley, 1990a) suggested that these flowed in a general westerly direction. Evidence that the sandstones were deposited rapidly, each probably by a single flood event, is suggested by the local abundance of load structures caused by dewatering. The parallel lamination that characterizes many of the sandstone beds is indicative of high-energy upper-flow regimes that accompanied sedimentation (Frostick and Reid, 1977; Tunbridge, 1981), the locally developed ripple cross-lamination recording the tractional re-working of the upper parts of the beds as flows waned. The thicker, discontinuous sandstones have been interpreted as the deposits of shallow channels (Pauley, 1990b). It is possible that these developed by channelization of flow during an initial erosive stage, by floods of unusual severity.

Conclusions

The Pike contains some of the finest exposures through the Synalds Formation, a sequence of Longmyndian sedimentary strata laid down on a late Precambrian alluvial floodplain. The various types of sandstone bed, and the diversity of well-preserved, small-scale sedimentary structures at this locality, are important pointers to this depositional environment. By reference to the other nearby GCR sites, the strata represented at The Pike record an intermediate stage in the trend that ultimately led to deposition of coarse-grained beds in the Stretton Group of the eastern Longmyndian sequence.

LONG BATCH–JONATHAN'S HOLLOW (SO 445 961)

D. Wilson

Introduction

The steep spur that lies between Jonathan's Hollow and Long Batch is the type section of the Batch Volcanic Beds, a series of tuff horizons that lie within the upper part of the Synalds Formation. This well-exposed section (Figure 5.13) has been selected as a GCR site as it reveals an almost complete sequence through the upper part of the formation and the lower part of the overlying Lightspout Formation, the tuffs providing an important series of correlateable marker horizons within the eastern Longmyndian. The volcanic origin of the tuffs was realized by Murchison (1867), but it was Cobbold (1900) who first named and described these beds. Although many authors have subsequently commented on the tuff horizons (James, 1956; Taylor, 1958; Toghill and Schell, 1984), they have never been formally designated, and the term 'Batch Volcanic Beds' remains in informal usage.

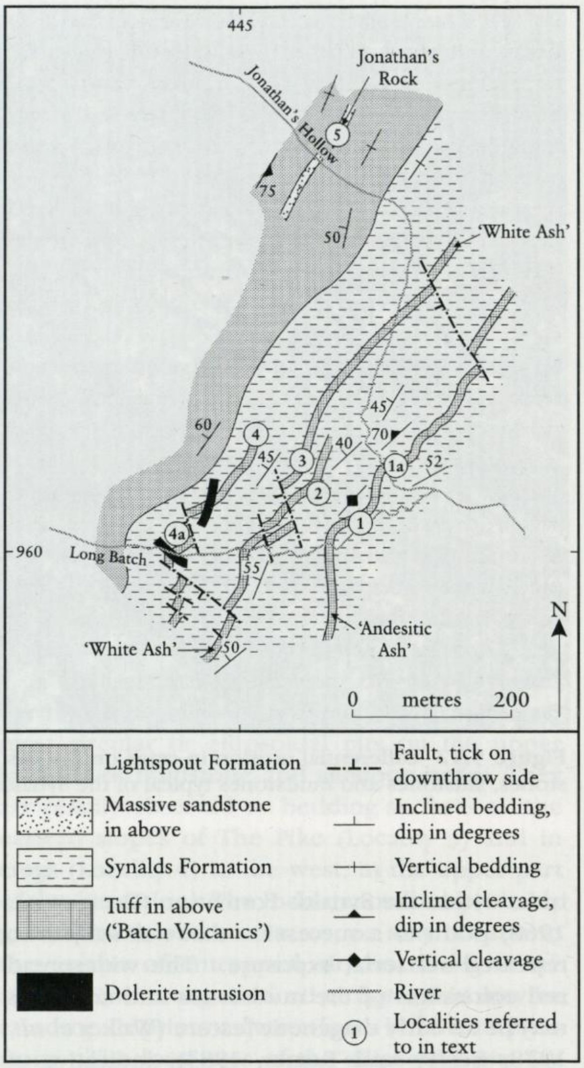

Figure 5.13 Geological map of the Long Batch–Jonathan's Hollow site.

It was Greig *et al.* (1968) who gave their first detailed petrographical descriptions and suggested that they represent a final phase of Uriconian volcanic activity. Until recently, however, little was known of the detailed geochemistry of the tuffs, or their exact relationship to the earlier Uriconian volcanism.

Description

The section at Long Batch exposes steeply westward dipping strata, comprising about 150 m of the upper part of the Synalds Formation and the lower 100 m of the overlying Lightspout Formation. The Synalds Formation is typical of that exposed elsewhere on the Long Mynd, namely a succession of alternating purplish red-brown siltstones and mudstones with subordinate thinly bedded purplish grey sandstones. The sandstones are generally lithic arenites with a significant volcanic component, and some appear to be largely composed of tuffaceous material. Up to four horizons of acid to intermediate tuff are interbedded with the strata in the upper part of the formation in this area, but only two have been traced for any distance laterally (Greig *et al.*, 1968).

The lowermost tuff bed has been informally termed the 'Andesitic Ash' (Cobbold, 1900; Greig *et al.*, 1968). It is a generally massive, poorly sorted, coarse-grained, crystal lithic tuff, about 5 m thick with a sharp base and top, passing in places into a lithic-lapilli tuff (classification of Le Maitre *et al.*, 1989). It is well exposed on the lower part of the spur dividing Long Batch and Jonathan's Hollow (Figure 5.14) and on the valley slopes to the north-east (Figure 5.13, Locality 1a). The tuff is characteristically greenish grey in its lower part, grading to pale purplish red higher up, and contains pyroclasts of large feldspar phenocrysts together with many conspicuous dark green epidotic aggregates, visible in hand specimen and locally up to 5 mm diameter. The latter display alteration rims of chlorite and white mica, commonly aligned within the rough cleavage that transects the tuff. The lower part of the tuff is noticeably coarser grained in places and, in addition, contains clasts up to 10 mm diameter of intermediate or basic volcanic rocks, reddened siltstone and mudstone.

About 45 m of mainly well-cleaved mudstones with thin sandstones separate the 'Andesitic Ash' from the next higher volcanic bed (Locality 2), an unnamed greenish grey, fine-grained tuff, up to 5 m thick, which crops out sporadically over the Long Mynd (Greig *et al.*, 1968). Up to 12 m of mudstone separate this horizon from the overlying 'White Ash' (Cobbold, 1900; Greig *et al.*, 1968). This is a distinctive, sharply defined bed of pale grey, white-weathering, coarsely feldspathic, crystal-lithic ash or lapilli tuff of intermediate composition, which contains many small elongate epidotic aggregates and conspicuous feldspar phenocrysts. There is, in places, a crude alignment of the feldspar laths and some of the epidote aggregates, suggestive of layering or grading. The tuff, which is 2–3 m thick in this

Figure 5.14 View looking west at the junction of Long Batch (left) and Jonathan's Hollow (from right of picture). Crags at base of spur are formed by the Andesitic Ash crossing from right to left across the spur. (Photo: A9425, reproduced by kind permission of the Director, British Geological Survey, © NERC.)

area, crops out in the higher crags of the spur between Long Batch and Jonathan's Hollow (Locality 3), and can be traced into the adjoining valleys. It is overlain by up to 40 m of purplish red mudstones and thin sandstones before the highest representative of the Batch Volcanic Beds supervenes. This is an unnamed pale grey and purplish red, fine-grained tuff, about 2 m thick, outcropping on the north-eastern slopes of Long Batch (Locality 4, 4a).

In thin section, there is little petrological difference between the various tuff horizons of the Batch Volcanic Beds. The matrix of the tuffs is composed mainly of micaceous aggregates, epidote and chlorite, which define a rough cleavage within the groundmass of the rock. The most abundant pyroclasts are albite or sodic oligoclase feldspar phenocrysts, occurring as stubby subhedral prisms or irregular fragments up to 1 mm across, in part altered to white mica, chlorite and epidote. Quartz also occurs, but is less abundant than feldspar and the fragments are usually smaller. Subordinate pyroclastic constituents include altered greenish brown biotite and, very rarely, prisms of apatite. Finely crystalline aggregates of chlorite, quartz and feldspar, some of which enclose small prisms of plagioclase, probably represent the alteration products of rhyolitic or andesitic fragments. The larger sericite-chlorite-epidote aggregates possibly represent reworked tuffaceous material (Greig *et al.*, 1968). Sedimentary lithoclasts include reddened siltstone and mudstone, presumably from the Synalds Formation, as well as small sandstone pebbles; metamorphic quartzite has also been reported (Greig *et al.*, 1968).

The base of the Lightspout Formation occurs about 20 m above the highest tuff bed, on the spur between Jonathan's Hollow and Long Batch, the intervening strata of purplish red mudstone and siltstone representing the highest beds of the Synalds Formation. The Lightspout Formation is comparable to strata in its type area at the Lightspout Hollow GCR site. It typically comprises alternations of greenish grey siltstones with subordinate red beds and packets of thinly to thickly bedded sandstone. One of these packets, vertically disposed on the slope below Jonathan's Rock (Locality 5), consists of alternating bundles of thick (0.3 m) and thin (20–30 mm), fine- to medium-grained sandstone and subordinate laminated siltstone beds. A massive 4 m-thick bed of medium- to coarse-grained sandstone within this packet forms Jonathan's Rock. It possibly correlates with the Haddon Hill Grit (Whittard *et al.*, 1953; James, 1956) or a similar sandstone in the lower part of the formation in the Carding Mill Valley, as seen at the Lightspout Hollow GCR site.

Interpretation

The strata of the Synalds and Lightspout formations are comparable to those that crop out around the Carding Mill Valley. They illustrate the fluvial processes that took place on the extensive alluvial floodplain that developed during the early Longmyndian. Their deposition followed a period of marine and fluvio-deltaic sedimentation, reflected in the underlying formations of the Stretton Group (Pauley, 1990a,b, 1991), and was part of a gradual shallowing, upwards-coarsening and progradation of facies that occurred throughout the eastern Longmyndian sequence. The composition of the large quantities of volcanic detritus within these sediments suggests that they were derived from sources largely within the Uriconian Group (Lapworth and Watts, 1910; Greig *et al.*, 1968; Pauley, 1990b).

It has been suggested that the Longmyndian sediments were deposited in a near-shore setting, within a fore-arc basin to a Uriconian volcanic arc with an axis situated to the south or south-east (Thorpe *et al.*, 1984; Pauley, 1990a). The long-standing idea that sedimentation was partly coeval with Uriconian igneous activity (Cobbold and Whittard, 1935; Greig *et al.*, 1968; Pauley, 1990b) has been largely re-inforced by the presence of pyroclastic lithologies within the Longmyndian succession, including the Batch Volcanic Beds seen here. Baker (1973), however, did not consider the need for such a direct link, preferring a later, extensional fault-controlled period of mild volcanicity to explain the presence of the tuff beds.

The Batch Volcanic Beds show no textural evidence for welding. In some places the lowest tuff (the 'Andesitic Ash') reveals a concentration of coarser lithic material towards the base, which could be interpreted as density grading within a pyroclastic flow unit. The alignment of elongate pyroclasts within the tuffs is a possible indicator of laminar flow (Williams and McBirney, 1979), but there has undoubtedly been overprinting of this fabric by subsequent deformation. The tuff beds may each record an individual ash-fall event, although from their poor sorting, coarse grain size and bedding characteristics they may be alternatively regarded as a series of single-event, subaerial pyroclastic ash-flows possibly deposited at some distance from the seat of the eruption. All of the Batch Volcanics were erupted across the alluvial floodplain sediments of the Synalds Formation, a depositional environment that has undoubtedly contributed towards the preservation of the tuff beds.

Conclusions

Long Batch and Jonathan's Hollow contain many informative sections through the upper part of the Longmyndian Synalds Formation and lower part of the Lightspout Formation. In particular, they provide the most complete sequence through the Batch Volcanic Beds, a group of intermediate ash and lapilli tuffs, which have been considered as late stage eruptions from the Uriconian Group. Evidence suggests that the tuffs were deposited subaerially, either as ash-fall deposits or distal pyroclastic flows in which welding fabrics were not developed. Their preservation may have occurred in areas that fortuitously escaped reworking in the alluvial floodplain environments within which they were deposited, and they may therefore represent the remains of a once more extensive period of Longmyndian volcanism.

LIGHTSPOUT HOLLOW (SO 435 952)

D. Wilson

Introduction

This site includes a small tributary on the western side of the Carding Mill Brook and the upper reaches of the Carding Mill Valley (Figure 5.15). It falls within the existing Longmynd SSSI and was selected for the GCR because of its exceptional exposures that are designated as the stratotype for the Lightspout Formation. This unit comprises about 500 to 800 m of thinly to thickly interbedded siltstones and sandstones with one or more thin tuff beds in the upper part. Its base is taken at the change from the predominantly purplish red-brown colouration of the underlying Synalds Formation to the greenish grey that typifies the Lightspout Formation, a somewhat arbitrary division, which nevertheless generally occurs at a comparable distance above the Batch Volcanic Beds throughout the region (James, 1956). The colour change is commonly accompanied by a lithological transition, with sandstones noticeably more abundant than in the Synalds Formation. Purple-red colouration re-appears in the upper part of the Lightspout Formation and becomes more persistent towards the top. The top of the formation is taken at the base of the Huckster Conglomerate, a well-defined marker bed that can be traced across most of the Long Mynd (James, 1956; Greig *et al.*, 1968).

The Lightspout Formation is a sequence of fluvial sandstones and siltstones within the middle part of the eastern Longmyndian Stretton Group, and is an important reference to the changing depositional environments of the Longmyndian, when studied in conjunction with other sites in the general area. Although the succession had been described in considerable detail by Murchison (1839), Salter (1857), Blake (1890) and others, the term 'Lightspout Group' was first given by Lapworth and Watts (1910), and the division has been generally accepted by subsequent workers. Significant contributions to the description and depositional environment of the formation have been provided by James (1956), Greig *et al.* (1968) and Pauley (1986, 1990a,b). Dunning (1975) modified the definitions used by Greig, replacing the term 'group' for the more formal designation of Lightspout Formation.

Description

The boundary between the Synalds and Lightspout formations can be readily identified in the numerous crags on the north and south sides of the Carding Mill Valley, where the beds dip steeply westward at 60° to 80° (Figure 5.15, Locality 1). The accompanying colour change occurs over several metres within the finer-grained lithologies, and the transition is marked by the appearance of flaggy siltstones and sandstones within the Lightspout Formation. The bulk of the formation along the Carding Mill Valley and Lightspout Hollow consists of alternations of silty mudstone and siltstone, with packets of sandstone and sporadic beds of purplish red mudstone and siltstone. The siltstones and silty mudstones are thinly colour banded from pale to dark greenish grey, and commonly contain sandstone laminae and beds up to 20 mm thick. Worn and smoothed surfaces on the paths

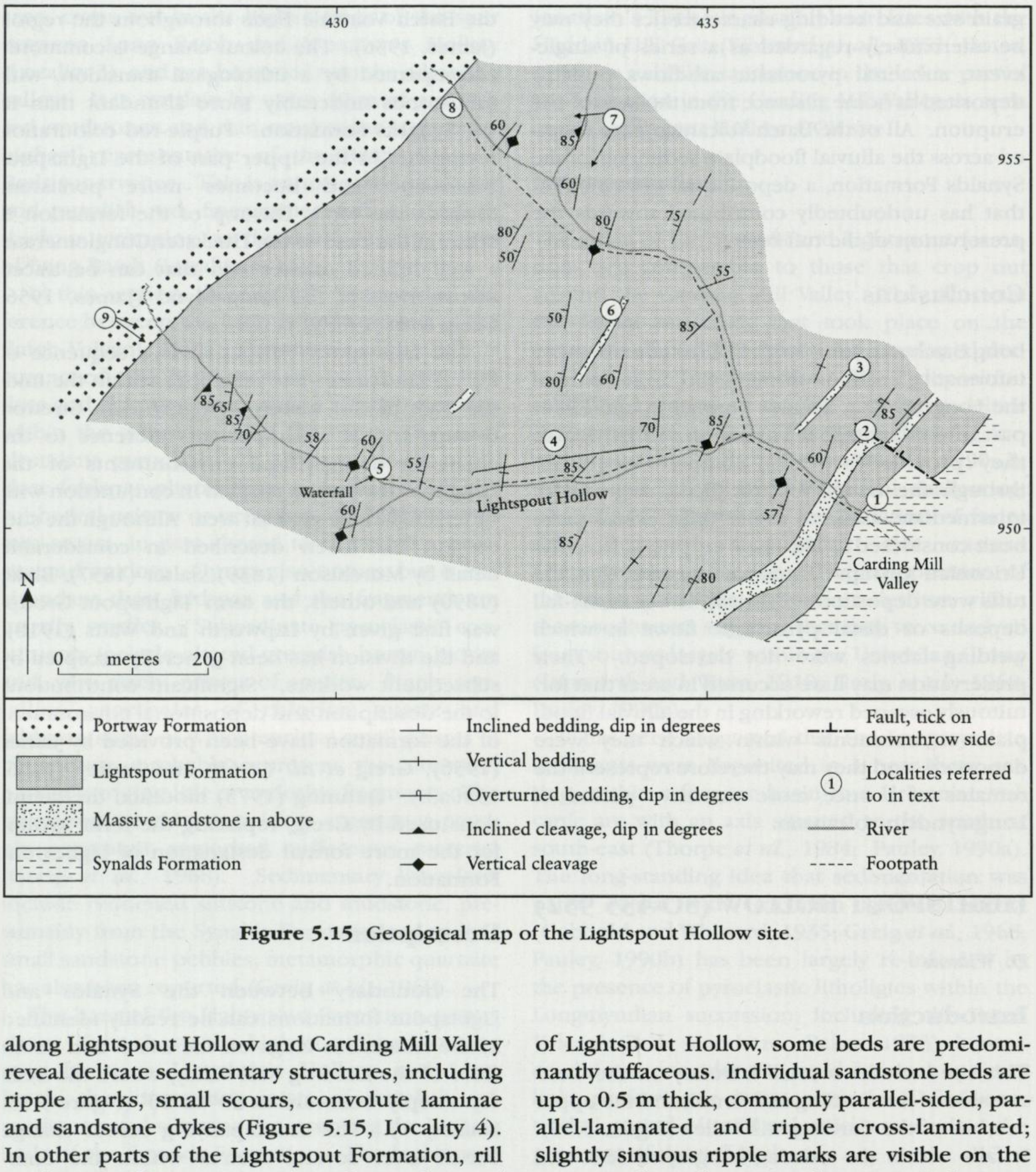

Figure 5.15 Geological map of the Lightspout Hollow site.

along Lightspout Hollow and Carding Mill Valley reveal delicate sedimentary structures, including ripple marks, small scours, convolute laminae and sandstone dykes (Figure 5.15, Locality 4). In other parts of the Lightspout Formation, rill markings and groove casts have been observed (Greig *et al.*, 1968). It was from similar facies that Peat (1984a) obtained a variety of algal microfilaments.

Sandstones occur throughout the Lightspout Formation in this area, in packets up to 10 m thick, interbedded with subordinate thin mudstones and siltstones. They are generally thinly to thickly bedded, fine- to medium-grained, lithic arenites, with a significant Uriconian volcanic detrital component, subordinate quartz and some mica (Greig *et al.*, 1968); in the upper part of Lightspout Hollow, some beds are predominantly tuffaceous. Individual sandstone beds are up to 0.5 m thick, commonly parallel-sided, parallel-laminated and ripple cross-laminated; slightly sinuous ripple marks are visible on the upper surface of some sandstones at Lightspout waterfall (Locality 5) (Greig *et al.*, 1968) and in several loose blocks. Poorly developed trough cross-bedding is discernible on some surfaces (Figure 5.16). Massive or irregularly bedded, medium-grained sandstone beds up to 4 m thick occur in places, first appearing about 30 m above the base of the formation and then at intervals throughout the succession (Locality 2, 3 and 6). One such bed occurs on the northern side of the Carding Mill Valley (Locality 2) and has been named the Haddon Hill Grit (James,

Figure 5.16 Exposures in the Lightspout Formation by the path along Lightspout Hollow, showing thinly laminated siltstones and mudstones low-angle ripple cross-lamination and small scours or convolute laminae (centre of picture). Overlain by cross-laminated sandstone. (Photo: D. Wilson.)

1956), although it is probably not traceable for any great distance (Greig *et al.*, 1968). Another, outcropping at the waterfall (Locality 5), has been named the Lightspout Grit (Taylor, 1958). A particular feature of many sandstone beds is quartz veining accompanying a strong pressure-solution cleavage along which the quartz is concentrated.

In the upper part of the formation, at the head of Carding Mill Valley, the strata revert to the dull purplish red-brown colour more typical of the underlying Synalds Formation (Locality 7). The colour change is gradual, and initially affects only the finer-grained lithologies. The highest beds are entirely purplish red, consisting of alternations of siltstone and coarse-grained sandstone and a coarse purple tuff, which is seen as debris along the path beside the Carding Mill Brook (Locality 8) (Greig *et al.*, 1968). The Huxter Conglomerate, defining the top of the formation, is only locally present in this area, being represented by small exposures of a massive, purplish red coarse-grained sandstone with scattered pebbles of quartz and red siltstone, about 350 m above the waterfall in Lightspout Hollow (Locality 9).

Interpretation

The Lightspout Formation records a continuation of the fluvial sedimentation that began with the deposition of the Synalds Formation. A similar range of subaerial and shallow water sedimentary structures is displayed in the siltstones and mudstones of both formations, although the Lightspout Formation contains the additional component of thicker, laterally extensive and more abundant sandstone units, reflecting the overall coarsening-upwards trend that occurs throughout the Longmyndian. These multistorey sandstone packets are thought to have been deposited predominantly by successive major unconfined sheet floods under high flow regimes (Tunbridge, 1981; Pauley, 1990a,b); ripple marks record waning flow conditions and reworking of the tops of the sandstone beds. Bed thickness variations between successive bundles of sandstones may reflect their relative proximity to the point of discharge, rather than the magnitude of individual flows.

The laminated siltstones and mudstones were probably deposited from suspension, during periods of low flow towards the end of each flood event. The small groove casts and scours sometimes seen in these deposits are presumably caused by erosion of the cohesive sediment by the succeeding flow. Palaeocurrent analyses of these and other structures by Pauley (1990a) indicated sheet floods that discharged in a west to WNW direction. Rapid sediment loading by successive flows was probably responsible for the small convolutions and injection structures seen in some of the laminated siltstone–mudstone sequences.

The massive, thick sandstones that occur sporadically throughout the Lightspout Formation, and some trough cross-bedded units, probably represent shallow channels (Pauley, 1990b), although some may represent flood deposits of unusual magnitude; there is no evidence to suggest that they are laterally accreted channel fills. However, the thick, coarse-grained, trough cross-bedded sandstones that appear in places in the upper part of the Lightspout Formation have been interpreted as braidplain deposits, representing incursions of this facies onto the alluvial floodplain as the fluvial system prograded north-westwards (Pauley, 1990a,b). A major advance of the braidplain occurred with the deposition of the widely developed Huxter Conglomerate, immediately succeeding the Lightspout Formation.

Conclusions

The excellent exposures along Lightspout Hollow and Carding Mill Valley provide the stratotype for the Lightspout Formation, a unit representing the younger part of the alluvial plain sedimentary facies that characterized Longmyndian Supergroup deposition during the late Precambrian. The site is highly instructive for sedimentologists since it demonstrates an upward evolution, from strata deposited as a result of sheet-flood activity to sequences indicative of alluvial braidplain environments. In conjunction with other GCR sites nearby, the site emphasizes the range of fluvial sedimentary environments that existed during deposition of the Eastern Longmyndian of Shropshire.

HAWKHAM HOLLOW (SO 432 975)

D. Wilson

Introduction

Hawkham Hollow, a small valley at the northern end of the Long Mynd SSSI, has been designated as a GCR site because it contains exposures that span the boundary between the eastern Longmyndian Stretton Group and the western Longmyndian Wentnor Group. The site (Figure 5.17) is important as the only place where the supposed unconformity between these groups

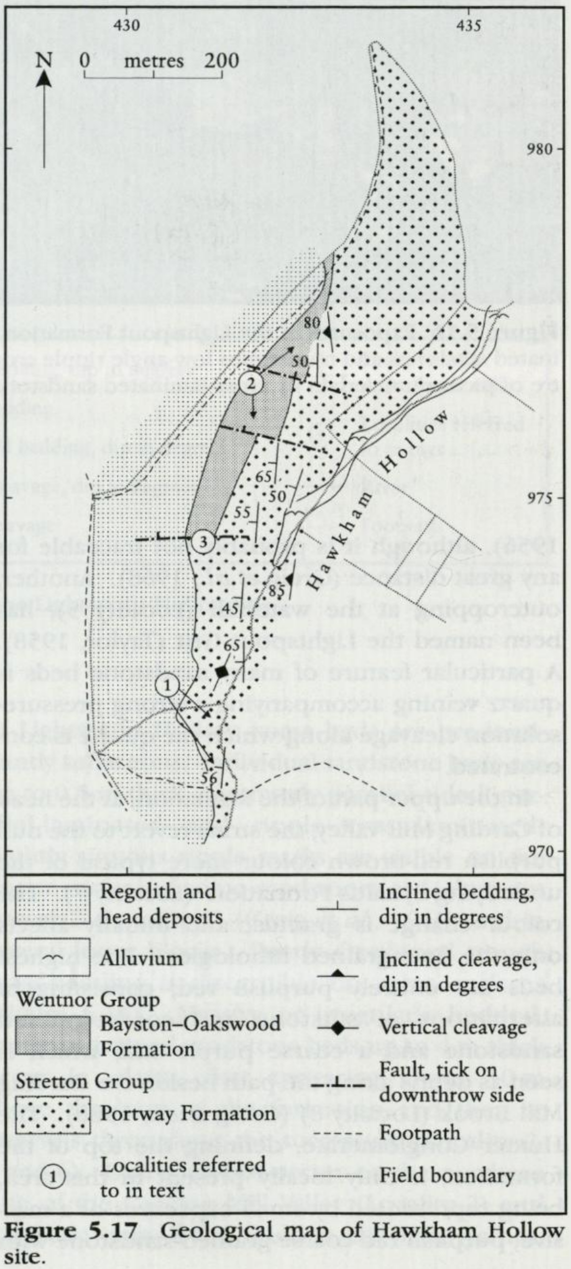

Figure 5.17 Geological map of Hawkham Hollow site.

can be studied, as well as providing a section through the generally poorly exposed Portway Formation, at the top of the Stretton Group.

Murchison (1839, 1867), Callaway (1891) and other early workers generally regarded the Wentnor and Stretton groups as parts of a conformable succession. It was Blake (1890) who first considered the possibility of an unconformity to explain stratigraphical problems between the eastern and western Longmyndian, although he regarded the latter as Cambrian in age. Although Blake's unconformity locally coincided with the base of the Wentnor Group, it was Cobbold (1925), and subsequently Cobbold and Whittard (1935), who proposed this as a major unconformable surface, the evidence cited being the greater geographical distribution of the overlying strata, implying widespread overstep of the Stretton Group. Until recently, this unconformity has been accepted by most authors (James, 1956; Dean, 1964; Toghill and Schell, 1984), support being provided by the apparent overstep of the Portway Formation and earlier divisions of the Stretton Group (Greig *et al.*, 1968), and a regional divergence in strike between the Wentnor and Stretton groups (James, 1956). However, Pauley (1991) has re-interpreted these overstep relationships and has suggested that the Wentnor Group does not unconformably overlie the Stretton Group; he considered the apparent unconformity to be largely the result of post-Longmyndian tectonism.

In a further development, James (1956) suggested that an unconformity was present at the base of the Portway Formation. He regarded the formation as equal to the Stretton and Wentnor groups and elevated its status accordingly (to his 'Mintonian' series). However, this idea has not been widely accepted, and has been dismissed on stratigraphical and structural grounds by Greig *et al.* (1968).

Description

The principal exposures through the upper part of the Portway Formation (Stretton Group) and the lowest beds of the overlying Bayston–Oakswood Formation (Wentnor Group) are to be found on the western slopes of Hawkham Hollow. The Portway Formation, intermittently exposed in crags in the upper part of the valley (Figure 5.17, Locality 1), comprises a sequence of dull purplish brown and subordinate greenish grey, planar laminated mudstones, siltstones and sandstones. The sandstones, which are usually bundled, are fine- to medium-grained lithic arenites occurring in beds up to 0.25 m thick; locally pebbly, they contain intraformational red mudstone clasts in the basal few centimetres. A few rather massive sandstone beds, up to 0.4 m thick, locally displaying low-angled erosional bases, occur at intervals within the sequence, and cross-lamination is visible in a few beds.

A series of massive, parallel-bedded and locally trough cross-bedded, fine- to medium-grained purplish red sandstones (lithic arenites), in beds up to 3 m or more thick, locally with erosional contacts, define the base of the overlying Bayston–Oakswood Formation in the higher crags above Hawkham Hollow (Locality 2). They contain scattered pebbles, thin pebbly layers and local thin conglomeratic beds, with pebbles up to 10 mm commonly visible on weathered surfaces. Although the incoming of these coarse sandstones is well defined in the crags, the junction between the Bayston–Oakswood and Portway formations is not marked by any angular discordance, bedding in both divisions being inclined westwards at angles between 50° and 60°. The formational boundary is cut by a number of small east-trending faults, which feature as 'slacks' on the steep hillside (Locality 3)

Interpretation

The Portway Formation represents a continuation of the Stretton Group alluvial floodplain facies that began with deposition of the Synalds and Lightspout formations (Pauley, 1990b). The lithologies within the Portway Formation are interpreted in a similar way to these earlier divisions, the siltstones mainly representing overbank deposits, and the sandstones recording a series of sheet floods or crevasse splay deposits; the thicker sandstones are probably the deposits of shallow channels. The common occurrence of laminated and locally cross-laminated sandstones within the succession indicates a greater proximity to the fluvial source than with the earlier formations. Conglomeratic beds within the Portway Formation have been interpreted as the first incursions of coarse braidplain deposits onto the alluvial floodplain (Pauley, 1990a,b). Their petrographical similarity to the trough cross-bedded sandstones and conglomerates of the overlying Bayston–Oakswood Formation, and the lack of any angular discordance at the formational boundary, led Pauley (1991) to suggest that the latter represented the culmination of braidplain progradation that began within the upper part of the Stretton Group. Palaeocurrent analyses (Pauley, 1990a) indicated that the braided fluvial system flowed in a general NNW direction. The relatively abrupt appearance of this facies indicates a rapid influx of coarse detritus, without any transitional facies, and has been linked to rejuvenation of the source areas, possibly by contemporaneous faulting (Pauley, 1990a). The supposed major unconformity between the Stretton and Wentnor groups could therefore be regarded as local disconformity, related to uplift and facies progradation.

The apparent regional overstep of the Portway Formation by the Bayston–Oakswood Formation cannot be demonstrated at Hawkham Hollow. However, there is a regional divergence in strike between the Wentnor and Stretton groups over this site, which is mirrored at other localities, and has been cited by James (1956) as evidence of major unconformity. In detail, this is evident as a gradual anticlockwise rotation of cleavage and bedding within the Portway Formation, as the boundary with the Bayston–Oakswood Formation is approached. Pauley (1990b, 1991) has suggested that the rotation is tectonic in nature as it affects structural elements (cleavage and fold axes) as well as bedding. He interpreted it as an incremental rotation within the Portway Formation, either through shear along cleavage and fault planes (Pauley, 1990b), or by rigid block rotation of the sandier Wentnor Group when the more argillaceous Stretton Group was undergoing ductile deformation (Pauley, 1991). He considered that the rotation is due to strike-slip movements, which occurred during or after folding of the Stretton and Wentnor groups. The movements were considered to have been taken up along major boundary faults within the Church Stretton and Pontesford lineaments, which have a documented history of strike-slip movement (Lynas, 1988; Woodcock, 1984b, 1988; Woodcock and Gibbons, 1988). Pauley's model does not, however, explain the relationship whereby the Bayston–Oakswood Formation (Wentnor Group) locally rests on successively older formations within the Stretton Group (Greig *et al.*, 1968), unless this can be attributed to facies change, faulting at the Portway to Bayston–Oakswood junction, or a combination of both.

Conclusions

Hawkham Hollow is an interesting and informative site, as it provides an opportunity to study the sedimentological and structural relationships between the Wentnor and Stretton groups. Although an unconformity between these groups remains to be proved, the site enables comparisons to be made of the evidence that has supported the previous sedimentological and structural models relevant to the evolution of this part of the Longmyndian Supergroup. It particularly demonstrates an abrupt sedimentary change, from proximal alluvial floodplain environments in the Portway Formation to the proximal braidplain conglomerate facies of the overlying Bayston–Oakswood Formation.

DOLYHIR AND STRINDS QUARRIES (SO 243 580)

N. H. Woodcock

Introduction

The Old Radnor Inlier offers the last westward glimpse of the Precambrian basement to the Midlands Microcraton, before it re-appears near Carmarthen, some hundred kilometres to the south-west (Pharaoh and Gibbons, 1994). Working quarries in the inlier have, for at least the last 100 years, provided fine sections through sedimentary rocks now assigned to the Longmyndian Supergroup, together with their unconformable cover of Silurian limestone and mudstone. This GCR site (Figure 5.18) therefore displays the best exposures of Longmyndian strata outside the type area of the Long Mynd, itself represented by the preceding sites of this chapter.

Early observers (Murchison, 1854) saw the rocks of the inlier as the conformable lower part of the local Silurian sequence. The recognition by Callaway (1900) of the unconformable relationship of the Silurian on older rocks that resemble the Longmyndian of Shropshire established the true significance of the Old Radnor Inlier.

The Dolyhir and Strinds quarries expose rocks providing an important window into the basement geology of southern Britain, framed by clear stratigraphical relationships with younger

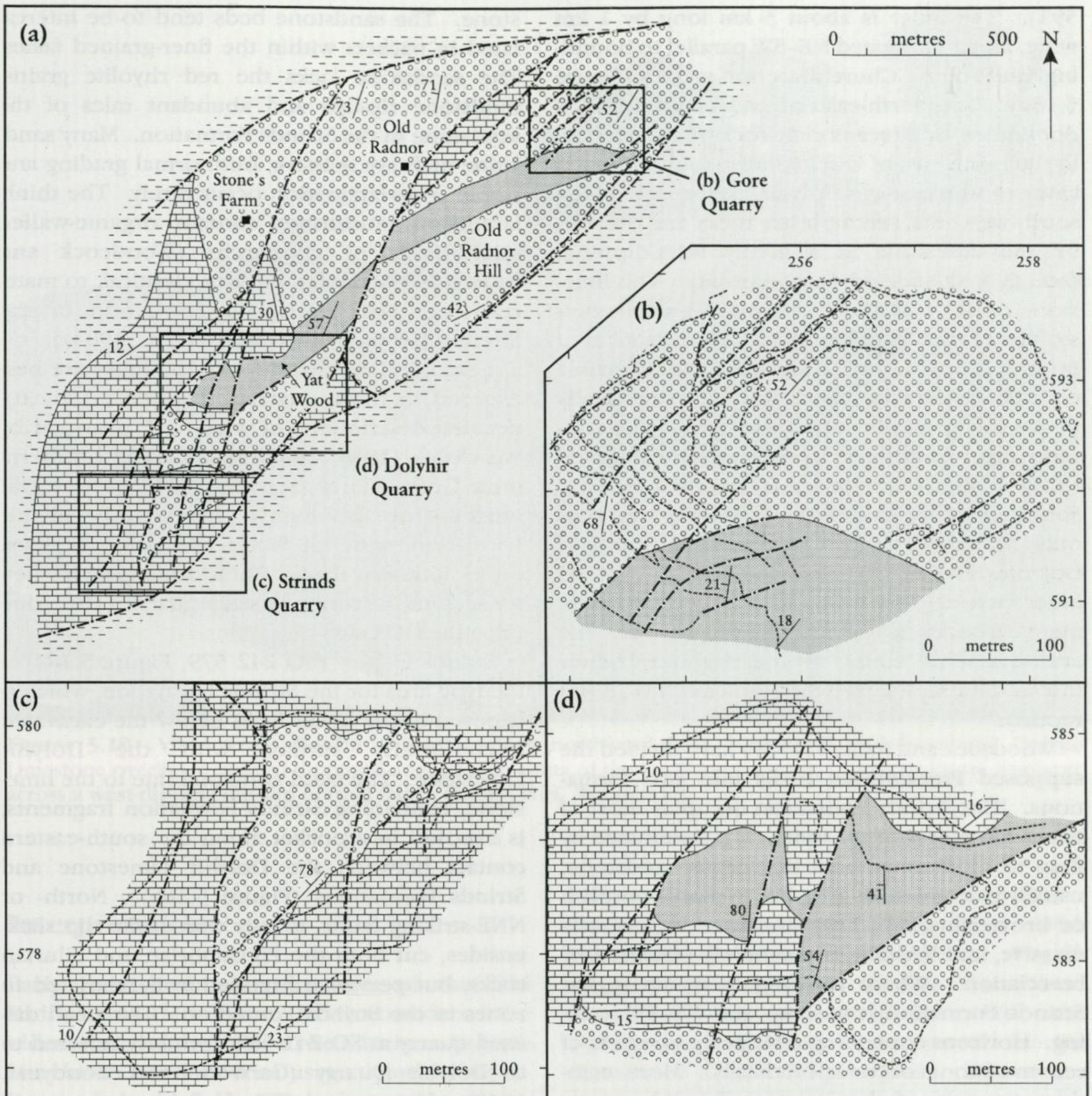

Figure 5.18 Geological maps of (a) the Old Radnor Inlier and (b, c, d) the three most extensive quarries within it.

overlying Silurian rocks. Further quarrying activity guarantees fresh exposure for scientific work, and a changing transect through complex three-dimensional lithological and structural relationships.

Description

The Old Radnor Inlier lies 5 km north-west of the town of Kington (e.g. Figure 1.1, 5.1), and is centred on the village of Old Radnor (SO 250

591). The inlier is about 3 km long by 1 km wide, and is elongated NE–SW parallel to bounding faults of the Church Stretton system (Figure 5.18a). The north-eastern end of the inlier is dominated by Precambrian rocks but includes, to the north-east of Old Radnor, a fault-bounded sliver of Wenlock-age Dolyhir Limestone. In the south-west, the Precambrian rocks are overlain by the limestone at a gently west-dipping, though fault-affected, unconformity. The limestone passes westward again into Wenlock mudstones of the Coalbrookdale Formation at a poorly exposed and supposedly conformable contact. These mudstones crop out sporadically in the low ground around the inlier. The contacts, apart from those in the west, are assumed to be faulted (Garwood and Goodyear, 1918), although direct evidence of faulting can now only be seen in the Weythel Brook south of Dolyhir (SO 243 577). The internal faults of the inlier record predominantly strike-slip movement (Woodcock, 1988). To the east lies the fault-bounded sliver of the Stanner–Hanter Intrusive Complex (see the Hanter Hill GCR site report).

Woodcock and Pauley (1989) have divided the supposed Precambrian rocks into two formations. The Strinds Formation occupies most of the outcrop area of the inlier. It is dominated by fine- to medium-grained micaceous sandstone, usually pale greenish grey, but in places purplish or brownish grey. The sandstone is generally massive, and bedding is commonly obscured by brecciation. Where bedding is visible in the Strinds Formation it is steep and NE–SW striking. Horizons containing tabular rip-up clasts of red mudstone occur sporadically. More common are units of clast-supported conglomerate of two compositions. The 'grey conglomerates' of Garwood and Goodyear (1918) comprise well-rounded clasts of vein-quartz with subordinate purple rhyolite and mica-schist. Their 'red conglomerates' contain a high proportion of purple rhyolite and reddened quartzite. Holgate and Hallowes (1941) matched clasts of granite and quartz-porphyry with similar lithologies in the Stanner–Hanter Complex to the east.

The Yat Wood Formation crops out in a NE–SW strip, the observed contacts of which are faults against Strinds Formation rocks (Figure 5.18a). The Yat Wood Formation contains three facies. Laminated or massive pale green siltstone, thinly laminated grey and green-grey mudstone, and green-grey fine-grained sandstone. The sandstone beds tend to be intercalated in packets within the finer-grained facies. The sandstone lacks the red rhyolite grains, mudstone rip-ups and abundant mica of the sandstone in the Strinds Formation. Many sandstone beds are massive, but normal grading and ripple cross-lamination occur rarely. The thinly laminated mudstones contain organic-walled microfossils or cryptarchs (Woodcock and Pauley, 1989) similar, but not identical, to material from the Lightspout Formation of the Stretton Group in Shropshire (Peat, 1984a).

Geological relationships in the inlier are best exposed in three quarries (Figures 5.18b,c,d), detailed descriptions of which are provided by Woodcock (1988, 1992, 1993). The northernmost Gore Quarry (SO 256 592, Figure 5.18b) mostly reveals highly brecciated Strinds Formation, with Yat Wood Formation exposed on its southern flank. Both formations are better seen in the south-western quarries that comprise the GCR site.

Strinds Quarry (SO 242 579, Figure 5.18c) is the type area for the Strinds Formation, which is exposed in the lower faces below the excavated unconformity surface with the Dolyhir Limestone. The basal conglomerate to the limestone, containing Strinds Formation fragments, is exposed in a pocket along the south-eastern contact between the Dolyhir Limestone and Strinds Formations (Figure 5.18a). North- or NNE-striking faults, mostly with strike-slip slickensides, cut both the Precambrian and Silurian rocks, but pervasive brecciation is restricted to zones in the Strinds Formation. The small disused quarry at SO 241 581, formerly referred to as Dolyhir Quarry (Garwood and Goodyear, 1918), also exposes Strinds Formation unconformably beneath Dolyhir Limestone.

The present Dolyhir Quarry (SO 244 584, Figure 5.18d), a northward extension of the former Yat Wood Quarries of Garwood and Goodyear (1918), contains Yat Wood Formation, faulted against the Strinds Formation in its south-east corner. Rapid working of this quarry changes the detailed exposure from year to year. The Wenlock unconformity is well exposed (Figure 5.19), with pockets of the basal conglomerate lying along it. The Yat Wood Formation dips moderately WNW, with good sections usually exposed near SO 245 584. Woodcock and Pauley (1989) provide a lithologic log of a typical section. NNE-striking faults cut both the Wenlock and underlying rocks in this quarry.

Figure 5.19 View across Dolyhir Quarry, looking northward from SO 244 583. Sub-horizontal Dolyhir Limestone overlies westward-dipping Longmyndian strata of the Yat Wood Formation. Both units are displaced across a west-dipping dip-slip normal fault. (Photo: N.H. Woodcock.)

Interpretation

Early published interpretations of the Old Radnor area (Murchison, 1854) considered the oldest rocks of the inlier to be Silurian – in both Murchison's and modern usage – and to underlie conformably the Wenlock-age limestone. Callaway (1900) demonstrated an unconformity beneath the limestone, and a lithological similarity of the older rocks to the better-known Longmyndian of Shropshire. Garwood and Goodyear (1918) supported this correlation, and specifically a match with the sandstone and conglomerate of their 'Bayston Group'.

The subdivision of the Longmyndian within the inlier (Woodcock and Pauley, 1989) has prompted a more detailed correlation with the type area. The predominant Strinds Formation is still correlated with the Bayston–Oakswood Formation of the Wentnor Group, interpreted by Pauley (1986, 1990a,b) as the deposits of a braided alluvial system (see also, the Hawkham Hollow GCR site report). The Yat Wood Formation is more tentatively correlated with parts of the underlying Stretton Group. The filamentous microfossils in the Yat Wood Formation urge a correlation with the Lightspout Formation, interpreted as the deposits of an alluvial floodplain. However, a match with the alluvial facies of the Synalds Formation or with the pro-deltaic facies of the Burway Formation is also considered possible (Woodcock and Pauley, 1989). The sandstones in both the Strinds and Yat Wood formations are rich in lithic fragments, like those of the type Longmyndian, suggesting derivation from an undissected magmatic arc.

The detailed correlations of the Strinds and Yat Wood formations with the type Longmyndian are important in establishing their possible Precambrian age. Local stratigraphical relationships only constrain the age of these formations to before the earliest Wenlock, the age of the Dolyhir Limestone. A smaller inlier of supposed Longmyndian rocks at Pedwardine, 15 km north-east, is associated with Tremadoc rocks, but any original unconformity between these two units has been replaced by a thrust (Boynton and Holland, 1997). The stratigraphical relationships of the Shropshire Longmyndian are also complex, as discussed in the introduction to this chapter, but it is probable (Pauley, 1990a,b, 1991) that at least one of its fault-detached components – the Willstone Hill Conglomerate – is unconformably overlain by Lower Cambrian strata.

The structural setting of the Old Radnor Longmyndian is also relevant to discussion of its probable age. Mapping by Kirk (1947, 1951, 1952) clarified Garwood and Goodyear's (1918) view of the inlier as a fault-bounded basement block along the Church Stretton Fault System. This is the same setting as that demonstrated by Holgate and Hallowes (1941) for the adjacent Stanner–Hanter Inlier, which contains plutonic rocks yielding an Rb-Sr isochron age of 702 ± 8 Ma (Patchett *et al.*, 1980). Holgate and Hallowes' (1941) identification in the Old Radnor Inlier of derived clasts from the Stanner–Hanter Complex suggests that the local Longmyndian is younger than this age. Woodcock (1988) showed that the internal faults, and by inference the bounding faults, of the Old Radnor Inlier record dominantly strike-slip displacement, mostly of post-Wenlock, probably Acadian, age. A gravity survey by Coster *et al.* (1997) suggests that the faults that bound the inliers of high density rock, particularly of the Stanner–Hanter ridge, cannot be vertical, but must converge at depth to isolate relatively small masses of Precambrian rock at a high structural level.

Conclusions

The Dolyhir and Strinds quarries provide the most instructive exposures in the Old Radnor Inlier, a lozenge of old rock caught up along the major Church Stretton Fault System. The unconformity with the overlying Wenlock (middle Silurian) rocks is magnificently displayed in former and currently active quarries, showing that the older rocks had been tilted and deformed before the Silurian strata were laid down. The older rocks can be subdivided into two formations, both of which can be matched with rocks of the late Precambrian Longmyndian Supergroup of Shropshire. The Old Radnor Inlier provides a rare view of the old basement rocks of the Welsh Borderland, and is a crucial link between the larger Precambrian outcrops of Shropshire and south-west Wales.

HANTER HILL (SO 252 572)
POTENTIAL GCR SITE

K. A. Jones

Introduction

The choice of Hanter Hill as a GCR site is justified by its excellent exposures of the Stanner–Hanter Intrusive Complex. The importance of these rocks lies in the fact that they represent what is possibly the oldest and, compared to the Malverns Complex, the least deformed, intrusive igneous association to be found in the English and Welsh Precambrian terranes. The earliest accounts of the Stanner–Hanter Complex are those of Murchison (1839, 1867), who described it as a 'hypersthene trap' intrusive into Ludlow (late Silurian) sedimentary strata. Callaway (1879) envisaged the complex to form a continuation of a ridge of Precambrian rocks stretching southwards from Lilleshall Hill and the Wrekin. Cole (1886) who classified the igneous rocks of Stanner but did not discuss their age provided the first petrological descriptions of the complex.

In past years there has been much debate over the age and affinities of these rocks. Callaway (1900) expounded upon his earlier ideas when he compared them with certain lithologies of the Church Stretton area, and suggested that the Stanner–Hanter Complex was of Uriconian type and probably 'Archaean' in age. On the other hand, Raw (1904) believed the exposed mass of Hanter Hill to be a Carboniferous laccolithic body intrusive into the Ludlow strata. Watts (1906) suggested that the complex was Tertiary in age on the basis of supposed petrographical similarities with the igneous complexes of Skye. Pocock and Whitehead (1935) followed Raw's ideas in suggesting that the complex was intrusive into Silurian shales and represented the continuation of an ESE–WSW trending belt of Carboniferous basalts and dolerites, which included the Clee Hills and Wyre Forest. Detailed petrographical and petrological studies carried out by Holgate and Hallowes (1941) resulted in a re-classification of the main rock types and elucidation of the order of intrusion. They further concluded that the complex had the form of a fault-bounded inlier located along the southerly extension of the Church Stretton Fault System. This is now known to be a Precambrian terrane boundary (Figure 1.1), and is in reality a plethora of structures with complicated movement histories, as discussed by Woodcock (1988). Holgate and Hallowes favoured a probable Precambrian age for the Stanner–Hanter rocks, partly because of their petrographical similarities to intrusions associated with the Uriconian Group of the Wrekin. They also found, however, that closely similar igneous rock types occur as pebbles in 'Longmyndian' conglomerates, described earlier in this chapter at the adjacent Dolyhir and Strinds GCR site. The latter observation, if verified, would indicate not only that the complex is Precambrian in age, but also that it was rather older than the Uriconian Group, having been exhumed and exposed to erosion prior to accumulation of the Longmyndian Supergroup.

The controversy over the age of the Stanner–Hanter rocks has been partly resolved by an Rb-Sr isochron date of 702 ± 8 Ma, obtained by Patchett *et al.* (1980), on a granophyre (an 'Acid Type' according to Holgate and Hallowes) from the summit of the Stanner ridge, north-east of Hanter Hill. Given the evidence discussed below, of substantial magma mixing within these rocks, this value may represent an emplacement age for the entire complex.

Description

The three en-echelon ridges that make up the Stanner–Hanter Inlier comprise, from NNE to SSW: Stanner, Worsell Wood and Hanter Hill. The outcrops of Stanner and Worsell Wood are poorly exposed owing to forestation and consequently only the better exposures, limited to Hanter Hill (Figure 5.20), are described here. No formal redefinition of these rocks has been made, and the terminology adopted follows the rather dated scheme of Holgate and Hallowes (1941).

According to Holgate and Hallowes the oldest component of the complex is the 'Fine Dolerites', which occupy the majority of north facing slopes on Hanter Hill. The best exposures are on crags (SO 2525 5754) above Lower Hanter Farm (Figure 5.21, Locality 1) and along the footpath (SO 2553 5740) leading to Red Vallet Wood. At outcrop the dolerites (which include microgabbros) are fine-grained, variable in colour, and show textures ranging from ophitic to porphyritic. Primary structures are not visible owing to subsequent deformation and weathering. In thin section the main varieties comprise olivine-free, pyroxene (augite)

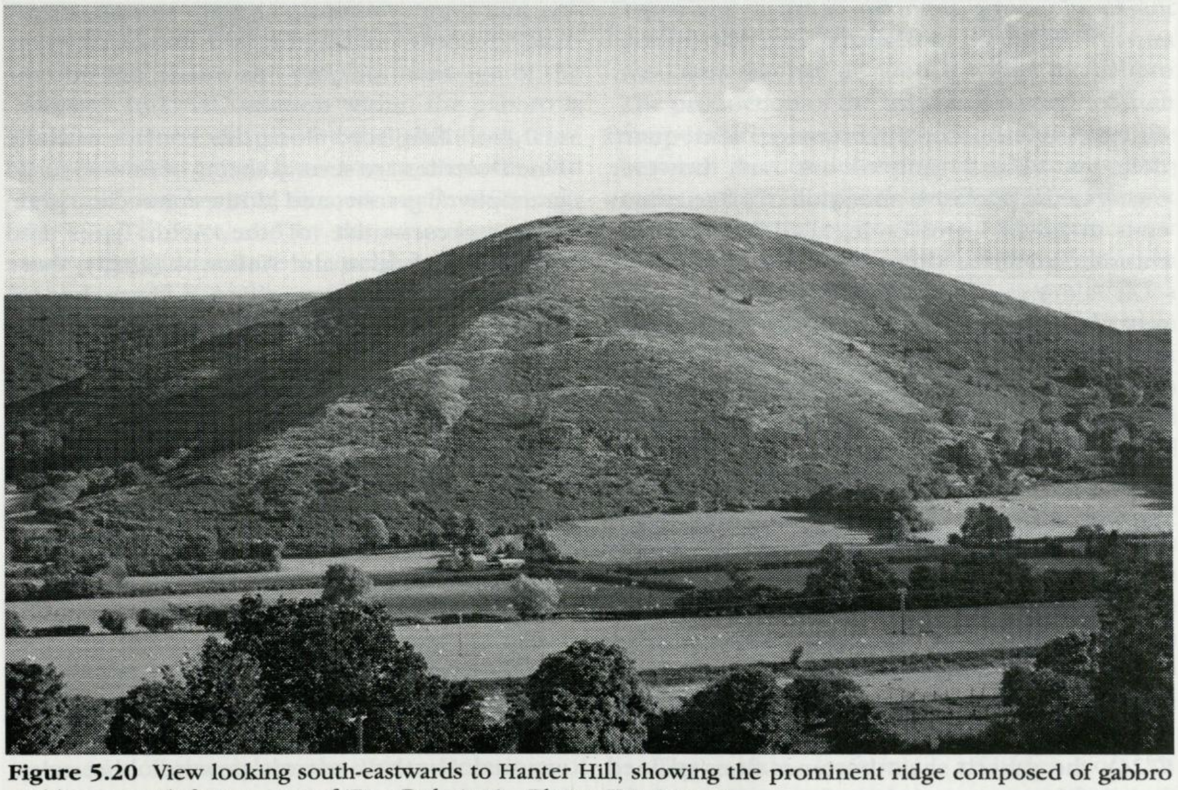

Figure 5.20 View looking south-eastwards to Hanter Hill, showing the prominent ridge composed of gabbro and lower north facing crags of 'Fine Dolerites'. (Photo: K.A. Jones.)

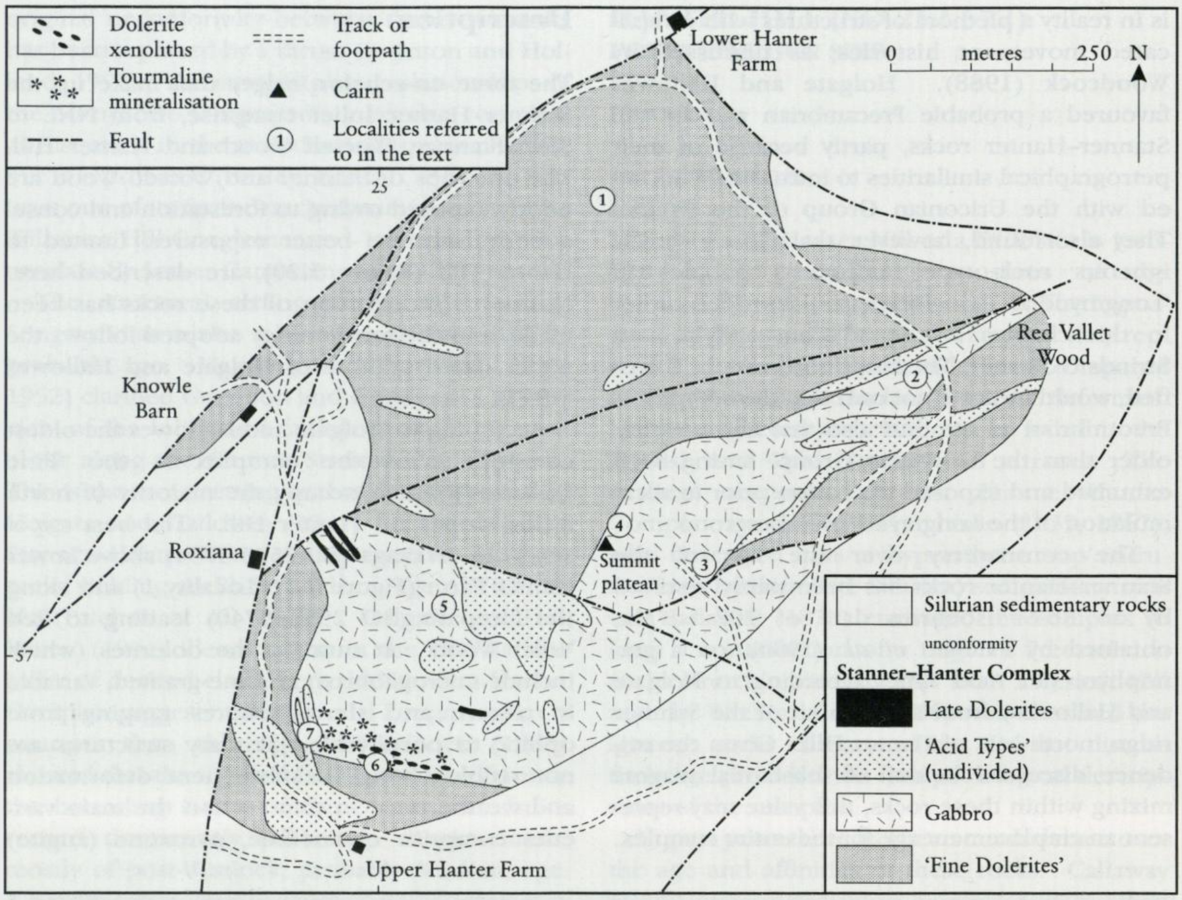

Figure 5.21 Geological sketch map of Hanter Hill (modified after Holgate and Hallowes, 1941).

dolerites with interstitial quartz, and quartz dolerites. Fresh samples are rare, however, owing to the pervasive alteration of primary minerals to green hornblende, actinolite, epidote, chlorite and albite.

At Red Vallet Wood (Locality 2) a tongue of very coarse-grained gabbro crosses the footpath (SO 2565 5734) and forms a prominent NE–SW trending ridge that leads to the summit plateau. This ridge and its crags are composed of the quartz-free gabbros of Holgate and Hallowes (1941). In thin section fresh samples of this variety are composed entirely of augite and plagioclase (labradorite); however, the present writer collected one sample containing the assemblage: orthopyroxene, clinopyroxene, plagioclase and quartz. In these gabbros the primary assemblage is commonly altered to green-hornblende, actinolite, epidote and chlorite.

On the lower crags of the ridge (SO 2554 5724), the gabbro sheet becomes finer grained and develops a porphyritic texture towards its margins. Emplaced along this contact with the Fine Dolerites are several sheets of fine-textured granophyric granite, and biotite-microcline granites, representative of the 'Acid Types and hybrids' of Holgate and Hallowes (1941); these minor granitic sheets are thinner where they are exposed on the summit. On the lower, south-facing slopes leading to the summit (Locality 3) a chilled margin between individual gabbro sheets is exposed (SO 2537 5710). This contact contains relict screens of the Fine Dolerites, and thin 1 m-wide sheets of biotite granite have subsequently invaded it. The granite sheets enclose abundant dolerite xenoliths, many of the latter displaying 'corrugated' rims, the significance of which is discussed below.

On the summit plateau (Locality 4) the contact between the coarse, quartz-free gabbro and a plagioclase phenocryst-bearing, coarser-grained facies of the Fine Dolerites is well exposed south-westwards towards the cairn. The gabbro sheets and the dolerites are there truncated by the major NW–SE trending fault, which dissects the complex. To the south of this fault (SO 2517 5710) is exposed the largest granitic body of the complex, the granophyric quartz porphyry of Holgate and Hallowes (1941), although it is poorly exposed and intensely fractured by the fault.

Most of the south- and south-west-facing slopes leading from the summit are in quartz-free gabbros. Near the summit these gabbros are strongly leucocratic but farther south, towards Upper Hanter Farm, they become melanocratic. On the western slopes (SO 2517 5705), overlooking Knowle Barn (Locality 5), there are scattered exposures of spectacular gabbro pegmatite featuring plagioclase laths and green-hornblende crystals, the latter up to 5 cm in length and replacing primary augite. These exposures show that the primary mineral assemblage of the gabbros displays various degrees of alteration, to green hornblende, actinolite, epidote and chlorite. The lower contact of the gabbro against the Fine Dolerites is seen on the craggy slopes (SO 2500 5687) above Upper Hanter Farm (Locality 6). It contours around the hill slopes, suggesting the body has the geometry of a shallow-dipping sheet, and is located in the vicinity of a line of small hawthorn trees. Inspection of these crags shows important primary structures that provide key evidence bearing on the nature and emplacement age of the gabbro. Internal variation within the gabbro is indicated, for example, by the abundant 'Fine Dolerites' xenoliths, up to 1 m in diameter, and relict primary layering in more mafic units near the basal contact, and by the more leucocratic and pegmatitic nature of its higher part. The gabbro itself is melanocratic and many of the crags exhibit primary compositional layering. Furthermore, the gabbro contains abundant tourmaline mineralization, the development of which is linked to the intrusion of fine-grained granite sheets (Holgate, 1977). In a small quarry to the west, located above the footpath (SO 2495 5688), there are several sheets of fine-grained biotite granite interleaved along the lower contact of the gabbro and also intruded into the underlying Fine Dolerites.

A further small quarry (SO 2492 5690) above Roxiana (Locality 7) shows in its backwall excellent exposures of the youngest components of the complex. These consist of several fine-grained dolerite dykes, the 'Later Dolerites' of Holgate and Hallowes (1941), cutting the gabbro. In thin section the dolerites are unaltered, with ophitic to sub-ophitic textures and olivine phenocrysts.

Interpretation

The Hanter Hill inlier is an excellent example of a relatively undeformed late Precambrian bimodal (gabbro-granophyre) magmatic complex. Four intrusive stages are recognized, which from oldest to youngest are: 'Fine Dolerites', Gabbro, 'Acid Types and hybrids' and 'Later Dolerites' in the terminology of Holgate and Hallowes (1941). The earliest stage, represented by the Fine Dolerites, probably resulted from the multiple injection of basic magmas as a series of sheets; however, their limited outcrop, fine-grain size and absence of observable primary structures renders them difficult to interpret. The Fine Dolerites were in turn intruded by coarse-grained, quartz-free gabbros, which in the north of the complex form thick near-vertical sheets with internal chills delineating the margins of individual components. On the southern slopes a larger and more homogeneous gabbro, with a shallow-dipping lower contact, is interpreted as a fragment of a small magma chamber.

The subsequent emplacement of the 'Acid Types and hybrids' (granites, granophyres and porphyries), was favoured by zones of structural weakness that had developed within the gabbro. The production of the hybrids is a result of high-temperature physical and chemical interactions between the hot doleritic and gabbroic country rocks and invading granitic sheets, according to Holgate and Hallowes (1941). Furthermore, in the abundant dolerite xenoliths the presence of cuspate or 'corrugated' rims is an unusual feature, which in other parts of the world (Vernon, 1983) is most commonly reported where acid and basic magmas are suspected to have 'mingled' together when still in a molten or semi-solidified state. The Later Dolerite dykes mark the final phase of the intrusive sequence.

The dolerites and gabbros were subsequently altered to mineral assemblages akin to those of the low-amphibolite and greenschist facies of regional metamorphism. Holgate and Hallowes (1941) speculated, however, that this alteration, and the tourmalinization of the gabbros, are phenomena associated with hydrothermal activity linked to the intrusion of the 'Acid Types'.

The plate tectonic affinities of the Stanner–Hanter Complex must await the results of further geochemical study. It probably represents a remnant of a magmatic 'feeder' zone located within the roots of a volcanic arc, although the radiometric evidence suggests that this was somewhat older than the volcanism that gave rise to the Uriconian Group of Shropshire and Llangynog (Figure 1.4).

Conclusions

The exposures of Hanter Hill give a unique insight into processes accompanying the emplacement of the Stanner–Hanter Complex, which is an example of a bimodal (gabbro-granophyre) intrusive association forming part of the igneous basement to the western Wrekin Terrane (Figure 1.1). The majority of the lithologies are basic in composition, but a significant acidic component is also seen as sheet-like granitic intrusions. These were evidently emplaced close to the time of the basic magmatism, resulting in excellent examples of some very rarely reported features that are attributed to the 'mingling' together of magmas with contrasting compositions. The *c.* 702 Ma Rb-Sr radiometric age obtained on the granitic rocks suggests that the Stanner–Hanter Complex may be one of the oldest igneous assemblages within the basement of southern mainland Britain (Figure 1.2).

LLANGYNOG (SN 330 145)

POTENTIAL GCR SITE

R. E. Bevins

Introduction

This structural inlier, in the vicinity of Llangynog, Carmarthenshire, comprises one of three important Precambrian sites in Wales lying to the south-east of the Menai Strait Fault System (Figure 6.1). Together with the Llyn Padarn and the St David's Peninsula GCR sites, described in Chapter 6, it provides critical evidence for the nature of the Precambrian basement beneath the Welsh Basin region. The site is of further significance, however, in representing one of the most south-westerly outcrops of the Wrekin Terrane (Figure 1.1). As will be discussed below, its rocks have affinities with the Uriconian Group of Shropshire and are dissimilar to the Pebidian Supergroup of south-west Wales, even though the latter are in closer proximity. A terrane boundary is inferred to separate the two, as shown in Figure 6.1.

The sequence exposed (Figure 5.22) is 1100 m thick and consists of rhyolitic and basaltic lavas, rhyolitic pumiceous and shardic ash-flow tuffs, basaltic autobreccias and uncommon hyaloclastites. These rocks are associated with intrusions of basic and, rarely, intermediate composition.

The earliest geological descriptions of the Llangynog area were by Murchison (1839), based on the rhyolitic rocks of the Capel Bethesda and Castell Cogan areas. The area was first investigated in any detail by the Geological Survey during their geological mapping of the district around Carmarthen. Cantrill and Thomas (1906) presented a full account of the Llangynog area, which was the basis for the subsequent memoir description compiled by Strahan *et al.* (1909). In these accounts the volcanic rocks were assigned an Ordovician age.

Identification of Tremadoc-age strata in the Llangynog area (Cope *et al.*, 1978) led to a renewed interest in the geology of this region, and brought into question the supposedly Arenig age of the volcanic rocks (Arenig being younger than Tremadoc), which actually underlie the Tremadocian strata. In particular, this study led to the conclusion that the rhyolites of the Castell Cogan area are of pre-Tremadoc age. Most significant of all, however, was the discovery of an Ediacaran fauna in associated strata (see the Coed Cochion GCR site report, Chapter 8), which confirmed a Precambrian age for at least some of the rocks of the Llangynog Inlier (Cope, 1977). The discovery has also established this site as one of international geological importance.

Later studies by Cope and Bevins (1993) confirmed that the Ediacaran fauna was preserved in volcaniclastic strata, and, as discussed in the introduction to this chapter, the rhyolites, basalts and related rocks are by association assumed to be late Precambrian. Cope and Bevins (1993) reviewed in detail the volcanic rocks of the Llangynog Inlier and established a new stratigraphical succession. This revised stratigraphical nomenclature is used in the description below.

Bevins *et al.* (1995) presented geochemical data for the basic, (rare) intermediate and silicic rocks from the Coomb Volcanic Formation. They concluded that the various rock types are petrogenetically related, most probably by low-pressure fractional crystallization dominated by

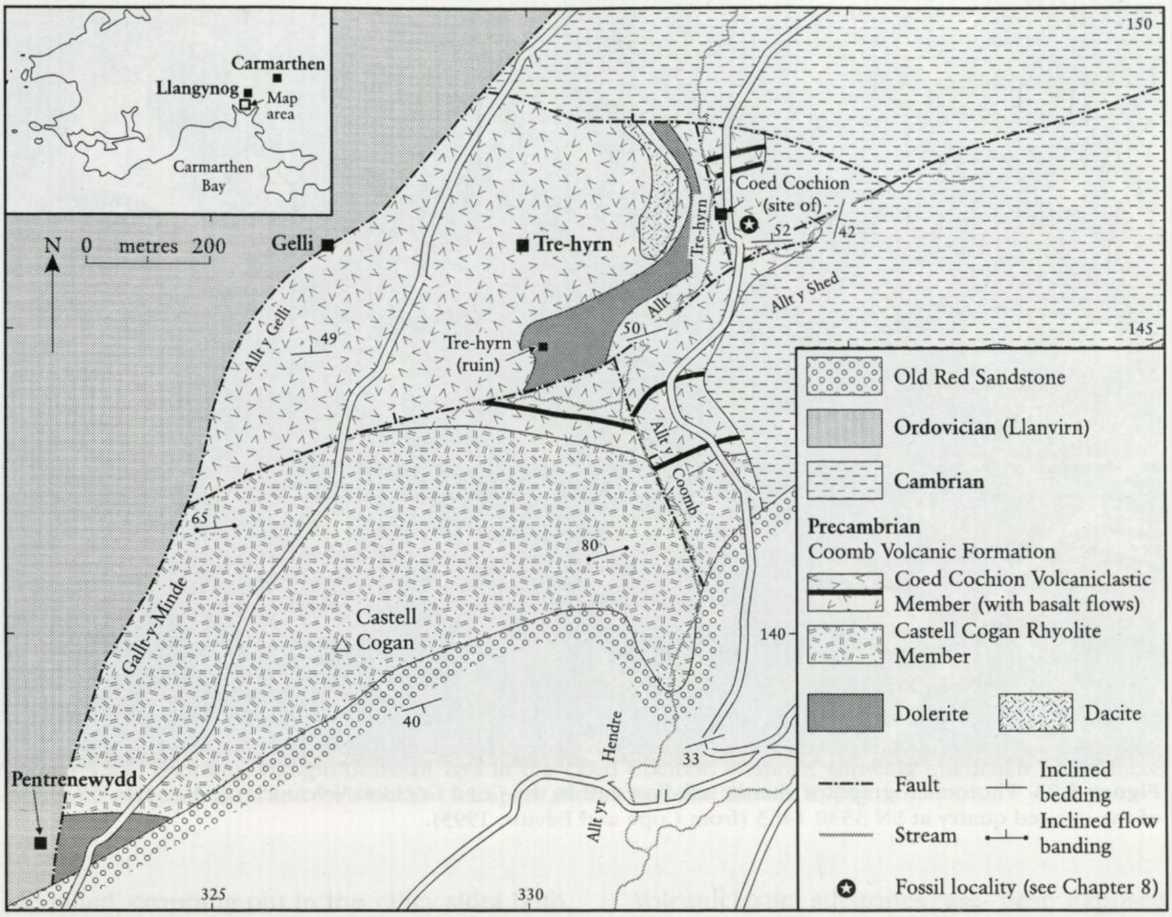

Figure 5.22 Geological map of the Llangynog inlier, including the Coed Cochion palaeontological site. This diagram is reproduced with the permission of the Geological Magazine, Cambridge University Press (modified from Cope and Bevins, 1983).

plagioclase feldspar and possibly amphibole. In particular, the Precambrian volcanic rocks of the Llangynog Inlier constitute a bimodal magmatic suite with marked petrological and geochemical similarities to the Uriconian Group, exposed farther north in Shropshire. On multi-element normalized diagrams, the basalts exhibit slight depletions of Ta and Nb and slight enrichments of Th and Ce indicative of active continental margin tectonic settings, but the Nb/Y diagram also indicates a strong within-plate chemical component. There are, however, certain geochemical differences to the Pebidian volcanic rocks exposed in the St David's Peninsula GCR site, which in fact are more similar to the volcanic rocks of presumed Precambrian age recovered from the Bryn-teg borehole in the Harlech Dome region of North Wales (see Chapter 6). The chemistry therefore supports the model discussed above, featuring a major intervening crustal boundary separating the Cymru and Wrekin terranes (Figure 1.1).

The western boundary of the Precambrian outcrop is fault-defined, with Llanvirn rocks lying to the west (Strahan *et al*., 1909). To the south and east the Precambrian rocks are overlain unconformably either by early Cambrian rocks (Cope and Rushton, 1992) or by Old Red Sandstone (Cope, 1980).

Description

The various Precambrian volcanic and volcaniclastic rocks of the Llangynog Inlier are exposed chiefly in the unnamed river valley and adjacent crags to the south of Llangynog village (Figure 5.22). Cope and Bevins (1993) coined the term Coomb Volcanic Formation for these rocks, recognizing two members, a lower Castell Cogan Rhyolite Member and an upper Coed Cochion Volcaniclastic Member.

The Castell Cogan Rhyolite Member, up to

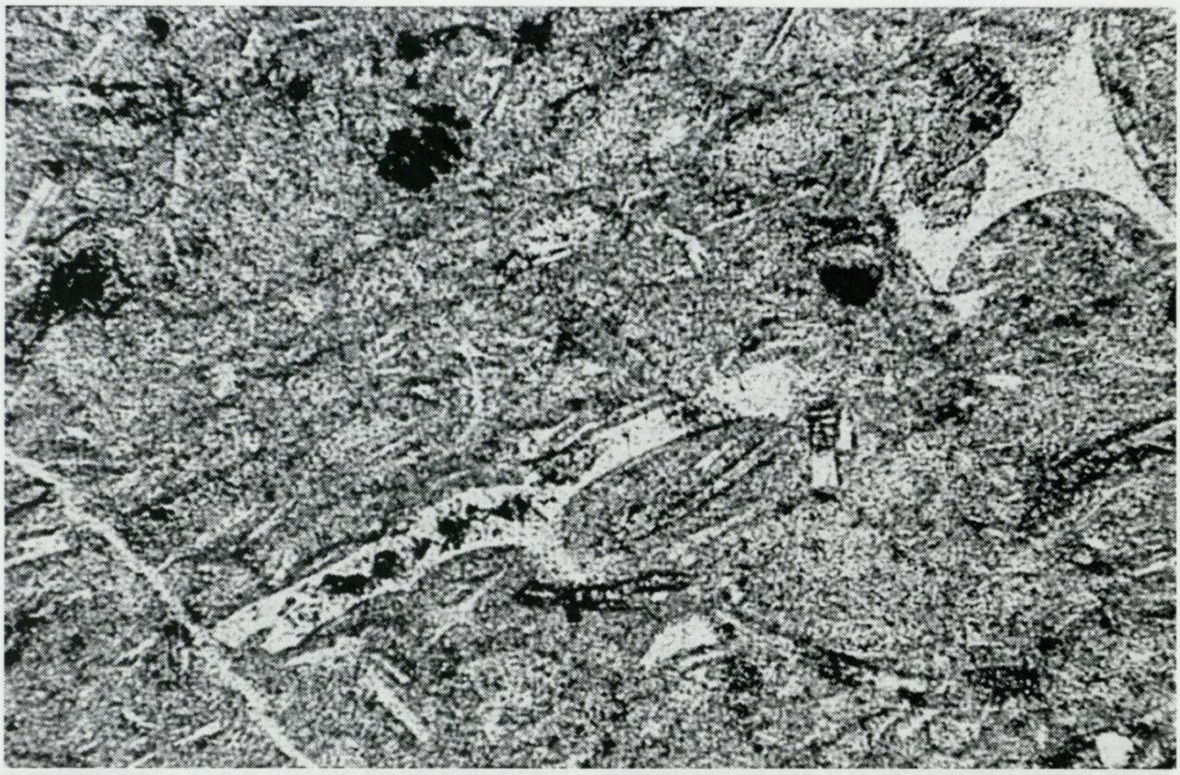

Figure 5.23 Photomicrograph of shardic ash-flow tuff in the Coed Cochion Volcaniclastic Member, 50 m east of the disused quarry at SN 3338 1463 (from Cope and Bevins, 1993).

400 m thick and younging to the north, chiefly comprises rhyolitic lavas, seen for example on the high ground of Castell Cogan (SN 3277 1394) and also on the crags of Allt yr Hendre (SN 3317 1417). More remotely, rhyolitic rocks are also exposed in the vicinity of Capel Bethesda (around SN 363 157) and to the east of Waun-das (SN 3553 1509), respectively lying to the north-east and south of Figure 5.22. The rhyolites show extensive silicification, dominated by mosaics of coarse-grained quartz, and are commonly cut by thin quartz veinlets. They do, however, preserve evidence of primary flow-banding, seen for example in forestry track exposures on Gallt y Minde, about 300 m west of Castell Cogan. The rhyolites are typically sparsely porphyritic, with feldspar phenocrysts up to 2 mm in length, although these are commonly replaced by white mica. The only other primary phase present is zircon.

Conglomerates, which crop out in the vicinity of Capel Bethesda, considered by Cope and Bevins (1993) to belong to the Castell Cogan Member, contain rounded cobbles and boulders of rhyolite up to 0.6 m in diameter. Other conglomerates, exposed farther south on the slopes of Allt y Coomb, to the south-east of old Tre-hyrn Farm (Figure 5.22), are correlated with the Coed Cochion Member.

The fossil-bearing Coed Cochion Member consists predominantly of volcaniclastic rocks, associated with rare basaltic lavas, rhyolitic ash-flow tuffs and an horizon of rhyolitic lava and associated conglomerates. The volcaniclastic rocks, which dip to the north at around 50°, are well exposed around the derelict house at SN 3332 1467 and at the disused quarry at Coed Cochion (SN 3335 1465). Sedimentary structures present include cross-lamination, climbing ripple lamination and sporadic flaser bedding; finely laminated and normally graded beds also occur. It is within these volcaniclastic beds that the Ediacaran fauna (Cope, 1977) occurs (see the Coed Cochion GCR site report, Chapter 8). The ash-flow tuffs display classic shardic textures (Figure 5.23) and contain clasts of welded ash-flow tuff with streaked out pumices (parataxitic texture) similar to inclusions in tuffs from the St David's Peninsula GCR site. Within the succession at least five thin basalt lava sheets have been

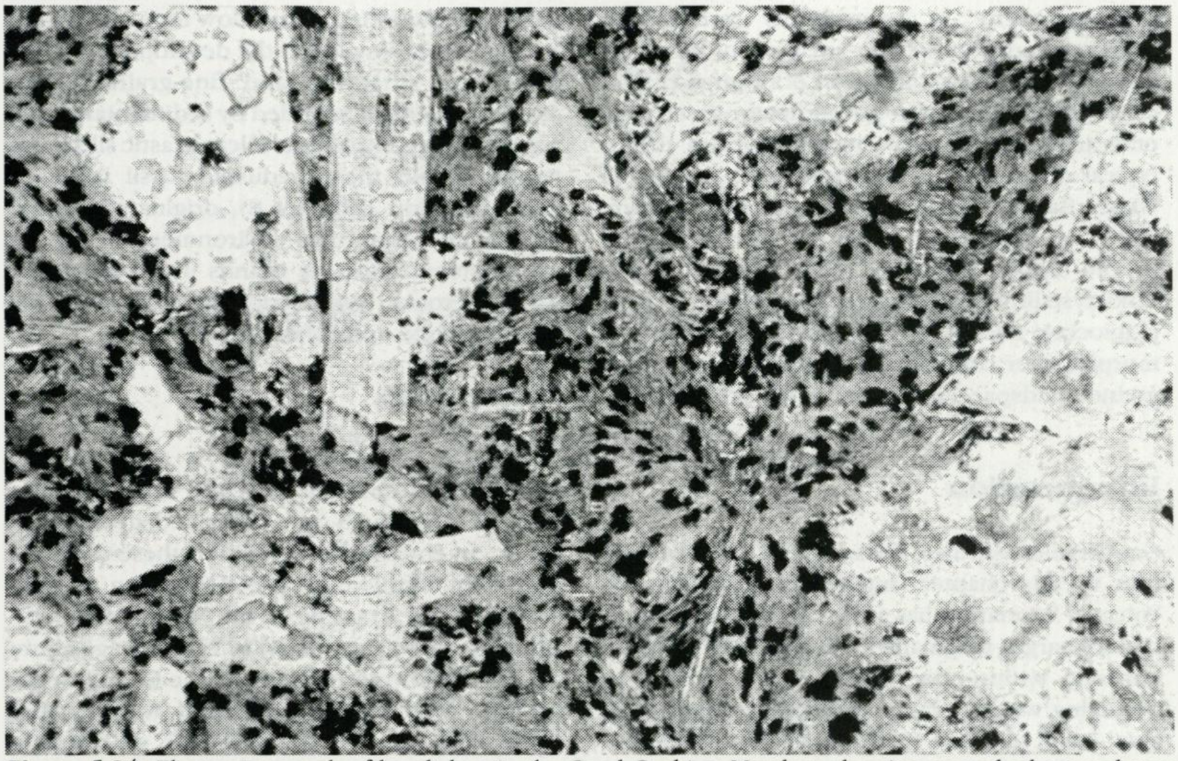

Figure 5.24 Photomicrograph of basalt lava in the Coed Cochion Member, showing quenched groundmass with skeletal, spherulitic and 'belt-buckle' feldspars (from Cope and Bevins, 1993).

identified, cropping out in the valley sides both to the north and south of Coed Cochion.

Basaltic lavas are sparsely porphyritic, with plagioclase phenocrysts up to 4 mm in length. These are set in a groundmass of recrystallized glass (now chlorite), accompanied by chloritic pseudomorphs after primary mafic minerals (most probably olivine and clinopyroxene), and Fe-Ti oxide minerals. The lavas show evidence of quench textures, including skeletal crystal forms, in particular 'swallow-tail' and 'belt-buckle' textures (Figure 5.24). A basalt flow exposed some 140 m to the north of Coed Cochion shows a brecciated margin in contact with adjacent sedimentary rocks. The breccia is composed entirely of former glassy basalt fragments and was originally a hyaloclastite.

The rhyolitic tuffs comprise crystals, chiefly angular quartz and plagioclase, and lithic clasts up to 3 mm in diameter. Other fragments typically include rhyolitic lava and rarer basalt, together with pumice fragments (again up to 3 mm) ubiquitously replaced by sericite and/or chlorite and showing no signs of flattening. Also present are glass shards, now replaced by fine-grained aggregates of quartz and feldspar.

Volcaniclastic siltstones are best exposed along the northern section of Allt y Coomb and on the lower slopes of Allt Tre-hyrn (Figure 5.22). Coarser-grained siltstone beds contain chiefly angular, fractured plagioclase crystals, angular quartz chips, chloritic pseudomorphs after mafic minerals, rarer white micas and scattered aggregates of celadonite. In places, rare basaltic lithic lapilli are present, and one bed has a preponderance of recrystallized glass shards.

Heterolithic conglomerates, exposed to the south-east of old Tre-hyrn Farm, contain a variety of angular to sub-rounded clasts, up to 3 cm in diameter, chiefly rhyolitic lava in character, but also including volcaniclastic sandstone, basaltic lava, and various types of rhyolitic tuff.

A number of basic intrusions are present in the site area, the largest occurring on the east-facing lower slopes of Allt Tre-hyrn and best exposed in a small quarry in the vicinity of old Tre-hyrn Farm (at SN 3312 1445). Primary textures are typically ophitic to sub-ophitic and well preserved, although primary minerals are almost entirely altered. Original Ca-plagioclase feldspar is ubiquitously replaced by albite, and primary mafic minerals are mostly replaced by chlorite (clinopyroxene is sporadically preserved). Prehnite and pumpellyite are present in minor amounts, reflecting alteration of original plagioclase feldspar and also groundmass; these minerals indicate low-grade metamorphism that is typical of the Lower Palaeozoic rocks across the Welsh Basin region (Bevins and Rowbotham, 1983). A single dacite intrusive sheet is present in the upper slopes of Allt Tre-hyrn. It is holocrystalline, with microporphyritic albitized plagioclase feldspar crystals and calcite pseudomorphs after amphibole, contained in a fine-grained feldspar-quartz dominated groundmass.

Interpretation

Geochemical studies, discussed in the introduction to this chapter, suggest that the Coomb Volcanic Formation has attributes of active continental margin magmatism, though with a superimposed within-plate component. Eruptions occurring within a fault-controlled ensialic marginal basin in a late Precambrian Avalonian volcanic arc (Figure 1.4) would probably have produced such a sequence.

Early volcanic activity in the Llangynog area was dominated by silicic magmatism, now represented by flow-banded rhyolites of the Castell Cogan Rhyolite Member that pass upwards into coarse, monolithic breccias. Such a progression suggests that the rhyolites were of extrusive origin, and most probably were subaqueously emplaced in the form of thick flows or domes. Reworking of the brecciated carapaces of these flows or domes probably provided the debris for the heterolithic conglomerates. Sporadic explosive eruptions are represented by the thinner pumiceous and shardic ash-flow tuffs.

Later activity was characterized by both acid and basic magmatism in the Coed Cochion Member. Basaltic lavas show quench textures, suggestive of subaqueous emplacement, as does the presence of hyaloclastite deposits, although no pillow structures have been recorded. Breccias composed almost entirely of basalt lava fragments are thought to have been derived from the reworking of lava flow tops, which with further reworking led to the development of volcaniclastic siltstones and fine-grained sandstones. Evidence provided by sedimentary structures suggests that these volcaniclastic siltstones, together with their medusoid fossil faunas at Coed Cochion (Chapter 8), originated in shallow water to intertidal environments. A model of ephemeral volcanic islands undergoing continuous erosion and reworking has been put forward by Cope and Bevins (1993).

Conclusions

Within the Llangynog Inlier it is possible to examine Precambrian volcanic rocks ranging from acid through rare intermediate to basic types. In chemical composition these rocks are markedly similar to the Uriconian Group, described for example at the Wrekin GCR site, which are thus considered to be close age-equivalents. Some volcanic rocks of the Castell Cogan Rhyolite Member are associated with unusual types of breccias, in keeping with an origin as rhyolitic or basaltic lava flows or domes emplaced in shallow waters. Explosive eruptions also took place, generating the ash-flow tuffs, and it is probable that all of this activity occurred in a continental margin setting, along the axis of a volcanic arc that was undergoing rifting. Erosion of these lavas produced abundant debris, which accumulated to form the thick volcaniclastic deposits of the Coed Cochion Member. Volcanism had not yet ceased, however, as the intercalated basalt flows testify. Shallow-water environments persisted throughout deposition of the volcaniclastic strata, and at times the medusoid Ediacaran faunas of Coed Cochion flourished and were preserved as fossils within the sedimentary layers.

Along with evidence from geological sites in the St David's and Llyn Padarn areas of western and northern Wales respectively, reviewed in Chapter 6, the Llangynog sequences provide an insight into the likely nature of the low-grade metamorphic basement underlying much of the Welsh Basin region.