Carboniferous and Permian Igneous Rocks of Great Britain

North of the Variscan Front

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Chapter 2

Dinantian volcanic rocks of the Midland Valley of Scotland and adjacent areas

INTRODUCTION

D. Stephenson

In contrast to the Northumberland, Solway and Tweed basins, where local volcanism commenced during Tournaisian time (see Chapter 3), volcanism did not start in the Midland Valley of Scotland until earliest Visean time. However, the resulting eruptions rapidly built up vast lava fields, which are by far the thickest and most extensive in the whole Carboniferous–Permian Igneous Province of northern Britain and probably constitute over 90% by volume of its eruptive products. Tomkeieff (1937) estimated that almost 6000 km^3 of magma was erupted during Dinantian times and some of the preserved volcanic successions are up to 1000 m thick (Paterson *et al.*, 1990). The most extensive lava fields are those of the Clyde Plateau, which dominate the landscape of the western Midland Valley to the north, west and south of Glasgow (Geikie, 1897). Possible outliers of these fields occur just across the Highland Boundary Fault in southern Kintyre and on Ben Bowie, near Helensburgh. In the eastern Midland Valley, the largest lava field is centred upon the Garleton Hills of East Lothian, and small but prominent outcrops occur within the city of Edinburgh. Later in Visean time, volcanic activity became centred upon the Burntisland area of Fife and the Bathgate Hills of West Lothian. The stratigraphical range of each succession of Dinantian volcanic rocks is shown in Figure 2.1 and their distribution in and around the Midland Valley is shown in Figure 2.2.

The early Visean volcanic successions are dominated by lavas, although considerable thicknesses of pyroclastic rocks and volcaniclastic sedimentary rocks occur locally. Uniquely within the Carboniferous–Permian Igneous Province of northern Britain, they are characterized by a wide range of compositions and, although most successions are dominated by transitional to mildly alkaline basalts, hawaiites and mugearites, many include

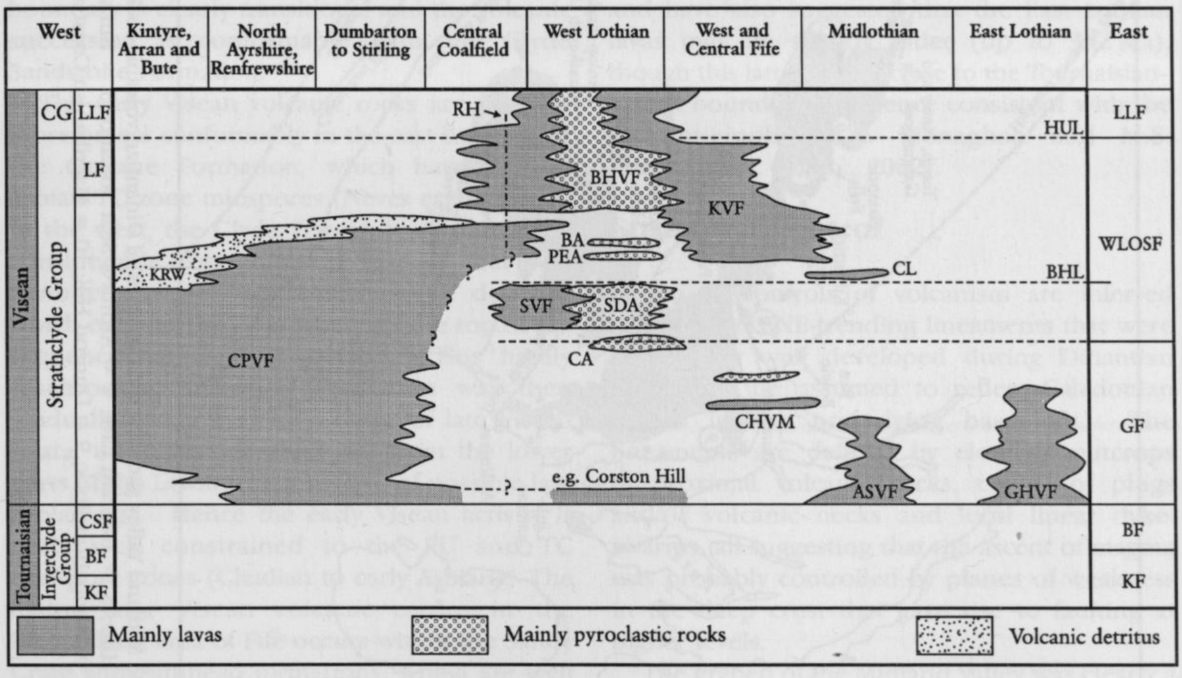


Figure 2.1 Stratigraphical and geographical distribution of Dinantian volcanic rocks in and around the Midland Valley of Scotland. Note that the diagram attempts to show stratigraphical range and not thickness of volcanic successions. (ASVF = Arthur's Seat Volcanic Formation; BA = Barracks Ash; BF = Ballagan Formation; BHL = Burdiehouse Limestone; BHVF = Bathgate Hills Volcanic Formation; CA = Crosswood Ash; CG = Clackmannan Group; CHVM = Charles Hill Volcanic Member; CL = Carlops Lava; CPVF = Clyde Plateau Volcanic Formation; CSF = Clyde Sandstone Formation; GF = Gullane Formation; HUL = Hurlet Limestone; KF = Kinnesswood Formation; KRW = Kirkwood Formation; LF = Lawmuir Formation; LLF = Lower Limestone Formation; PEA = Port Edgar Ash; RH = Rashiehill Borehole; SDA = Seafield–Deans Ash; SVF = Salsburgh Volcanic Formation; WLOSF = West Lothian Oil-shale Formation (equivalent to the Aberlady Formation in East Lothian).) Fife formation names omitted.

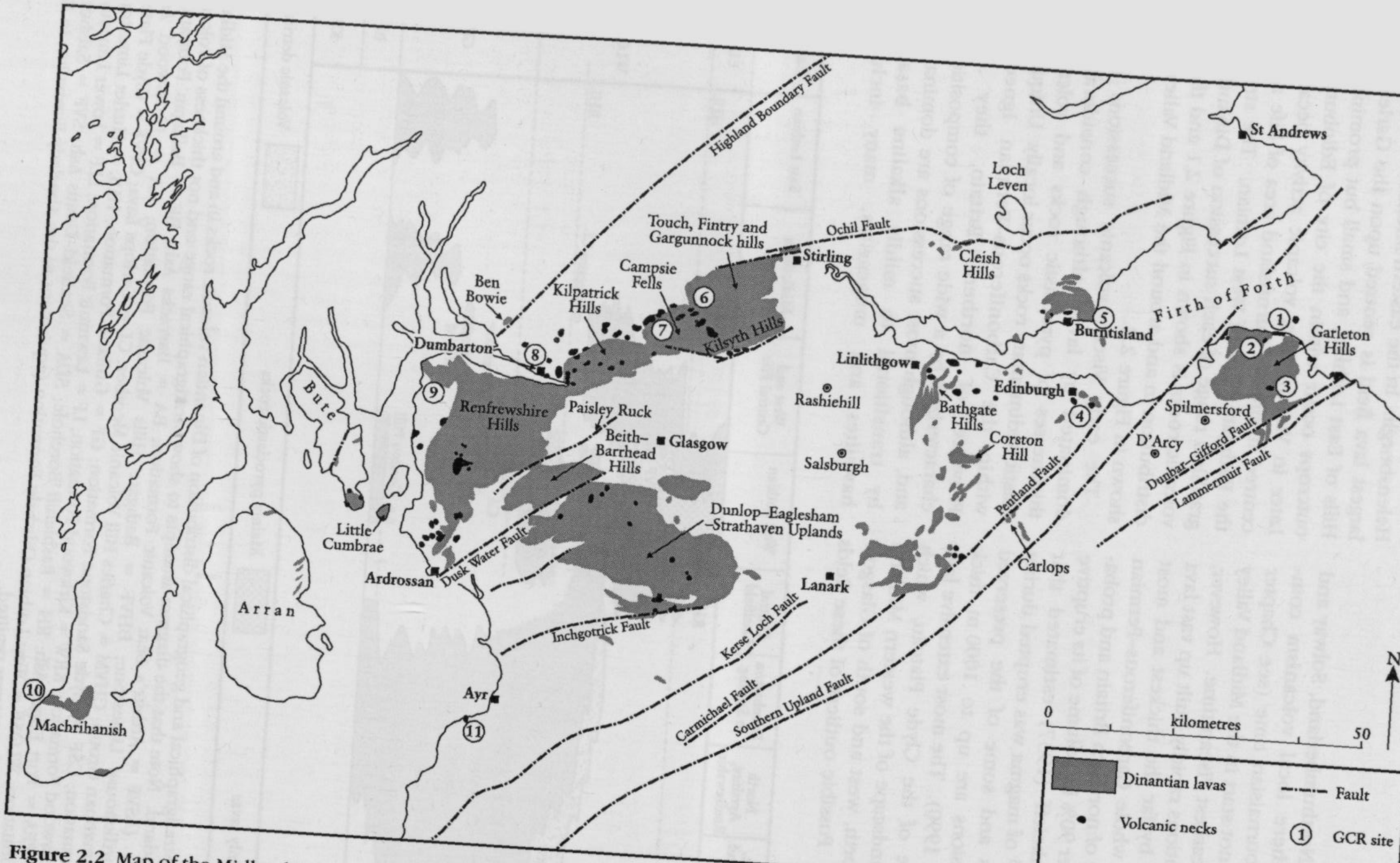


Figure 2.2 Map of the Midland Valley showing the outcrops of Dinantian volcanic rocks and the major structural components. GCR sites: 1 = North Berwick Coast; 2 = Garleton Hills; 3 = Traprain Law; 4 = Arthur's Seat Volcano; 5 = Burntisland to Kinghorn Coast; 6 = Touch, Fintry and Gargunnock Hills; 7 = Campsie Fells; 8 = Dumbarton Rock; 9 = Dunrod Hill; 10 = Macrihanish Coast and South Kintyre; 11 = Heads of Ayr. After Cameron and Stephenson (1985).

trachytic rocks and some have rhyolites. Late Visean volcanism, in contrast, was exclusively basic in character.

The volcanism appears to have started approximately synchronously across much of the Midland Valley, around the Tournaisian–Visean boundary. In the east, the earliest volcanic rocks rest conformably upon the Ballagan Formation, and CM to PU zone miospores and plant remains have been found just below and within them (Davies, 1974; Bateman and Scott, 1990; Scott, 1990). In the west, the Clyde Plateau Volcanic Formation rests with regional unconformity on strata that range from the Stratheden Group (Upper Devonian) to the Clyde Sandstone Formation, which overlies the Ballagan Formation at the top of the Inverclyde Group (Paterson and Hall, 1986). However, in the south-west of the Kilpatrick Hills, thin sandstones occur in the lowest part of the Clyde Plateau Volcanic Formation, and below the northern escarpment of the Touch and Gargunnock hills tuffs occur in the top of the Clyde Sandstone Formation. In these areas, the boundary is clearly transitional and the volcanic succession is conformable with the Clyde Sandstone Formation.

The early Visean volcanic rocks are overlain more-or-less conformably in the east by strata of the Gullane Formation, which have yielded Asbian TC zone miospores (Neves *et al.*, 1973). In the west, the Clyde Plateau lavas built up a considerable topographical feature that was denuded to produce volcaniclastic detritus, which overlies the lavas wherever the top of the sequence has been preserved. This highly diachronous Kirkwood Formation was then gradually overlapped by a range of late Visean strata, the oldest of which are from the lower parts of the Lawmuir Formation, of possible late Asbian age. Hence the early Visean activity is quite well constrained to the PU and TC miospore zones (Chadian to early Asbian). The major later Visean volcanic centre in the Burntisland area of Fife occurs within the Sandy Craig and Pathhead formations, which are well constrained elsewhere by miospore data to the Asbian–Brigantian interval (Brindley and Spinner, 1987, 1989; Browne *et al.*, 1996). In the Bathgate Hills, volcanism commenced in latest Asbian time and continued well into Namurian time.

Radiometric ages obtained from Dinantian volcanic rocks of the Midland Valley are confusing, mainly because of a scarcity of suitable material, for which only K-Ar whole-rock determinations have been published. De Souza (1982), summarizing his earlier work, suggested that the bulk of the Clyde Plateau Volcanic Formation lavas were erupted between 335 Ma and 325 Ma and the De Souza (1979) data, adjusted for new constants, gave an age of *c.* 326 Ma for the Kinghorn lavas and 326–316 Ma for the Bathgate Hills lavas. These dates fit well with the biostratigraphical data and the Gradstein and Ogg (1996) timescale. The East Lothian phonolitic intrusions of Traprain Law and North Berwick Law gave dates of *c.* 328 Ma, suggesting that they were contemporaneous with the Visean activity. However, dates for the Arthur's Seat and East Lothian lavas (Fitch *et al.*, 1970; De Souza, 1974), adjusted for new constants, ranged from 355 Ma to 345 Ma, suggesting an earlier, Tournaisian episode, which is inconsistent with the biostratigraphy. Recently obtained, more precise Ar-Ar dates from separated minerals have confirmed the age of the Clyde Plateau lavas at 335 Ma to 329 Ma and have also suggested that the East Lothian lavas may be slightly older (up to 342 Ma), though this latter date is close to the Tournaisian–Visean boundary and hence consistent with the biostratigraphy (A.A. Monaghan and M.S. Pringle, pers. comm., 2002).

Structural control

Structural controls of volcanism are inferred from NE- to ENE-trending lineaments that were particularly well developed during Dinantian times and are assumed to reflect Caledonian trends in the underlying basement. The lineaments are defined by elongate outcrops of proximal volcanic rocks, chains of plugs and/or volcanic necks and local linear dyke-swarms, all suggesting that the ascent of magma was probably controlled by planes of weakness in the deep crust that gave rise to faulting at higher levels.

The graben of the Midland Valley was clearly a major control, yet there is no evidence of volcanism directly associated with the Highland Boundary Fault. The most north-westerly extrusive centres on this side of the Midland Valley are concentrated on the NNW side of the main volcanic outcrops, within a 2–3 km-wide zone that extends ENE for some 27 km, from Dumbarton towards Stirling (Figure 2.3).

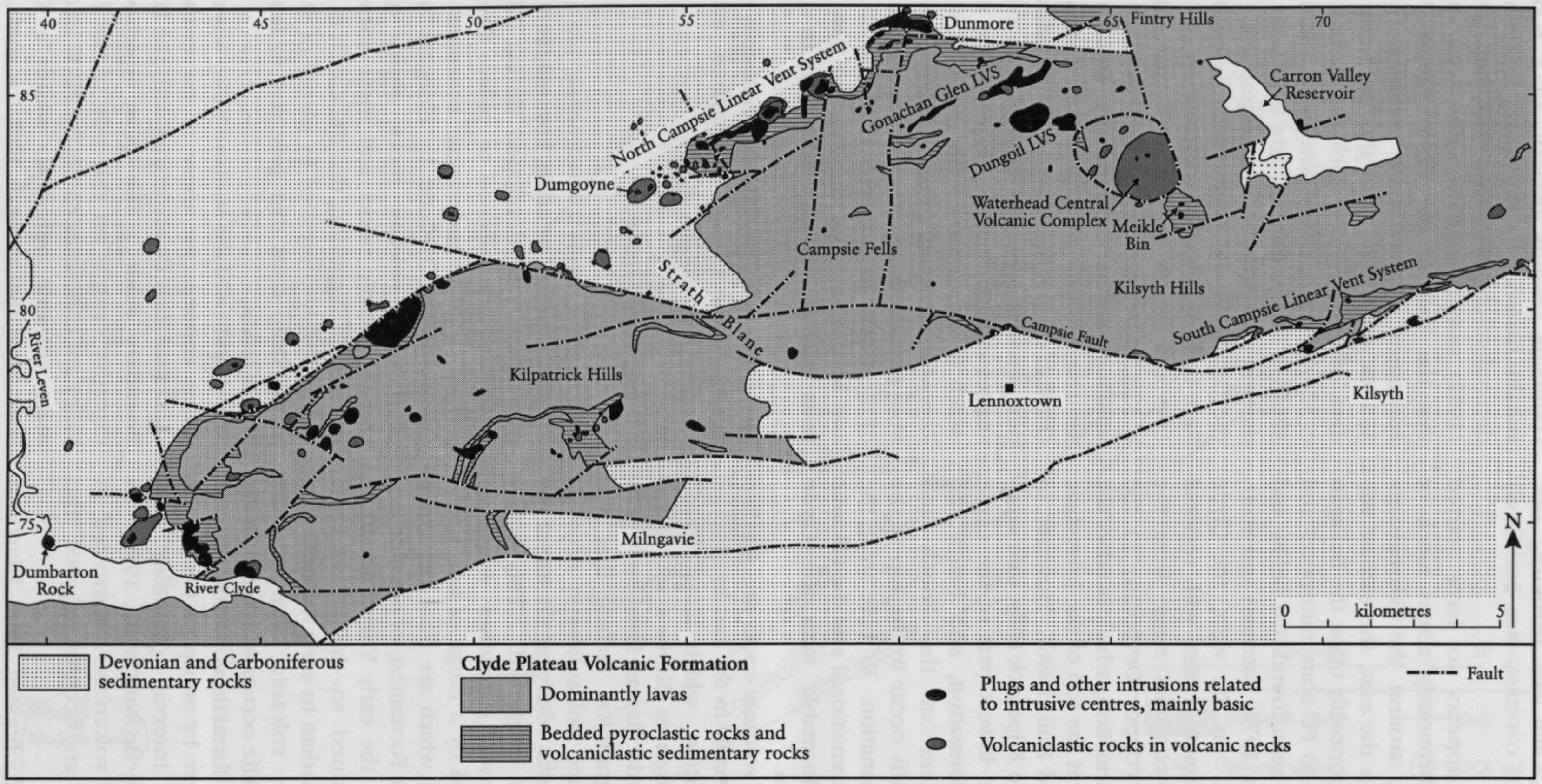


Figure 2.3 Map of the Kilpatrick Hills and Campsie Fells, showing outcrops of the Clyde Plateau Volcanic Formation and volcanotectonic lineaments defined by plugs, necks and proximal volcaniclastic beds. The most prominent lineament, along the north-west edge of the volcanic outcrops, is the Dumbarton–Fintry Line of Whyte and MacDonald (1974). Based on British Geological Survey 1:50 000 sheets 30W, Greenock (1990); 30E, Glasgow (1993); and 31W, Airdrie (1992).

This lineament has become known as the 'Dumbarton–Fintry Line' (Whyte and MacDonald, 1974; Craig and Hall, 1975). It is marked by numerous intrusions (many of them forming prominent landmarks such as Dumbarton Rock, Dumgoyne and Dunmore), pipes occupied by fragmentary materials, and proximal bedded pyroclastic rocks that probably represent degraded tuff-cones. These features are particularly well seen in the **Campsie Fells** GCR site. A concentration of dykes along a similar trend to the south-west suggests that the lineament may continue through the Renfrewshire Hills to south Bute. The north-eastern end of the lineament comprises the North Campsie Linear Vent System, and two slightly younger linear vent systems form separate *en échelon* lines within the Campsie lava block to the south-east (Forsyth *et al.*, 1996). On the south side of the Campsie Block, the South Campsie Linear Vent System forms a 15 km-long lineament close to the Campsie Fault (Craig and Hall, 1975; Forsyth *et al.*, 1996).

Other lines of necks and plugs in the northern Clyde Plateau outcrop are aligned WNW–ESE to north-west–south-east and may reflect less extensive conjugate Caledonian fractures at depth. Notable examples occur at each end of the Kilpatrick Hills, at Dumbarton and Strath Blane, where their coalescing tephra cones may have acted as local barriers between adjacent lava fields during the earlier phases of volcanism (Hall *et al.*, 1998).

Other WSW- to nearly W–E-trending faults throughout the Midland Valley that may have been utilized by rising magmas, also probably formed active escarpments controlling local accumulations of lava. Consequently they are commonly marked by significant changes in thickness of the volcanic piles. These faults include the Paisley Ruck, and the Dusk Water and Inchgotrick faults in Ayrshire (McLean, 1966; Hall, 1974; Rollin in Monro, 1999), as well as the Campsie and Ochil faults (Rippon *et al.*, 1996) (Figure 2.2). The unconformity beneath the lavas in the west has been attributed to a localized mid-Dinantian east–west compressional event (Paterson *et al.*, 1990). However, a close association between the maximum excision of strata and the thickest developments of volcanic rocks led Monro (1982) and Forsyth *et al.* (1996) to suggest that it was due, at least in part, to magmatic updoming in advance of the Visean eruptions.

At the south-eastern limit of the Midland Valley, the north-eastern extrapolation of the main Southern Upland Fault, together with NE-trending splays to the south-east, in particular the Dunbar–Gifford Fault, are thought to pass beneath Carboniferous strata (Max, 1976; Floyd, 1994) and to have controlled the rise of magma throughout much of the Carboniferous Period (Upton, 1982).

Palaeogeography and styles of eruption

In the eastern Midland Valley, the earliest volcanic rocks seem to rest conformably on the mudstones, siltstones and dolomitic limestones of the Ballagan Formation. Therefore, they were probably erupted onto flat low-lying coastal plains and deltas, with semi-marine lagoons and sabkhas. The explosive interaction of magma with surface and ground water resulted initially in phreatomagmatic eruptions, evidence for which is well seen in East Lothian, particularly in the **North Berwick Coast** GCR site. Abundant small vent structures and bedded pyroclastic rocks with base-surge and ash-fall characteristics are interbedded with the sedimentary succession, suggesting the development of shallow tuff-rings, probably less than 1 km in diameter. Lacustrine sedimentary rocks in some of the vents suggest the presence of crater lakes (maars) and some preserve remnants of early terrestrial tropical vegetation.

Farther inland, generally to the west, were semi-arid floodplains, with outwash fans and playa lakes. This transition from coastal plain to the entirely terrestrial, fluvial environment, typified by the Clyde Sandstone Formation and indicating regional uplift, was probably diachronous across the Midland Valley and broadly coincided with the onset of volcanism. The plains may have been divided initially by NE-trending ridges and escarpments formed from pre-Carboniferous rocks but, with a rise in magma productivity, rapidly accumulating lava fields began to form major landscape features. These topographic highs dominated the late Visean palaeogeography and also had a long-lasting influence on subsequent basin development. In such areas, eruptions were almost entirely subaerial and lavas were usually of aa type, though rare pahoehoe features have been reported. Flow surfaces are rarely preserved and thick red-brown boles occur on the top of most flows, resulting from the development of tropical or sub-tropical lateritic soils, indicative of significant interludes of quiescence between eruptions. Basic to mugearitic lavas, typically between 5 m and 30 m thick, were erupted through relatively small shield volcanoes. Together with cinder cones of coarse pyroclastic rocks, these commonly coalesced along NE-trending lineaments, now marked by upstanding plugs and volcanic necks. This style of eruption is represented by the **Campsie Fells** GCR site.

The absence of volcanic necks and proximal pyroclastic rocks from some areas suggests eruption from fissure volcanoes. The **Touch, Fintry and Gargunnock Hills** GCR site represents one such area, and the lateral continuity of some flows for over 6 km in the escarpment of the Gargunnock Hills has been cited as further evidence of fissure eruptions (Read in Francis *et al.*, 1970). Regional dyke-swarms that may have acted as feeders to the fissure eruptions are not obvious in most lava successions. However, there is a marked concentration of ENE- to NE-trending dykes up to 12 m wide along a south-western continuation of the Dumbarton–Fintry volcanotectonic line, which can be traced beneath the thickest part of the Renfrewshire Hills succession (Paterson *et al.*, 1990), across Great Cumbrae (Tyrrell, 1917a) and into southern Bute (Smellie, 1916). Outwith the major lava fields, individual volcanoes such as Arthur's Seat and the Heads of Ayr were possibly up to 5 km in diameter and rose to heights of about 1000 m above the plain (Whyte, 1963b; Black, 1966).

More evolved lava compositions are common locally in the Dinantian lava fields, and the abundance of trachytic extrusive rocks in the southern crop of the Clyde Plateau Volcanic Formation between Greenock and Strathaven implies that higher stratovolcanoes may have developed in this region (MacPherson *et al.*, 2001). The best-documented example is the 8 km-wide Misty Law Trachytic Centre in the Renfrewshire Hills, which comprises trachytic pyroclastic rocks, massive lavas of trachyte and rhyolite, and trachytic plugs and necks (Johnstone, 1965; Stephenson in Paterson *et al.*, 1990). Trachyte lavas are also abundant in the upper part of the Garleton Hills Volcanic Formation and the **Garleton Hills** GCR site has been selected to represent this style of volcanism. They are also present in the upper part of the Machrihanish succession (see **Machrihanish Coast and South Kintyre** GCR site report). Rhyolites occur locally in the upper part of the succession in the western Campsie Fells and near the base of the sequence in the Cleish Hills (Geikie, 1900). Flow banding in many of these evolved lavas indicates viscous flow; they probably never extended more than a few kilometres from their source and may even have formed steep-sided lava domes (e.g. the trachyte of Skerry Fell Fad, near Machrihanish and the rhyolite at Swinlees in the southern Renfrewshire Hills).

Calderas may have developed over some of the principal salic centres and the thick trachyte lavas forming the Garleton Hills of East Lothian may have been ponded in a caldera (B.G.J. Upton, pers. comm., 2001). However, the best-documented evidence occurs in the Waterhead Volcanic Complex of the Campsie Fells (Craig, 1980; Forsyth *et al.*, 1996). Here, a large multiple neck and several smaller necks, plugged by a wide variety of rock-types, occur within an oval ring-fault 2 km by 2.5 km. The complex is underlain by a positive gravity anomaly (Cotton, 1968) and the enclosed basic lavas show intense brecciation and hydrothermal alteration and are intruded by a variety of dykes (MacDonald, 1973). Some of the dykes are felsic, and trachytic pyroclastic rocks in the adjacent tephra cone of Meikle Bin have been attributed to the centre, although there are no felsic lavas preserved.

Despite the abundance of felsic volcanic rocks and the inferred presence of calderas in some areas, there is little evidence for pyroclastic flows, which are typical of such activity elsewhere. Well-bedded, carbonated and haematitized trachytic tuffs near the Weak Law Vent in East Lothian that were originally interpreted as welded ash-flow tuffs (Upton in Sabine and Sutherland, 1982), are now considered to be ash-fall material (see **North Berwick Coast** GCR site report). Welded trachytic lapilli-tuffs near Eaglesham have also been interpreted as ash fall (MacPherson and Phillips, 1998).

In the West Lothian oil-shale field, the land surface remained close to sea level during Dinantian times and similar conditions prevailed during most of Silesian time throughout the eastern Midland Valley. Relatively small basaltic volcanoes erupted onto coastal plains with lagoons and into shallow seas, locally building volcanic islands, fringed by reefs that were periodically eroded and submerged. Initial eruptions were explosive (phreatomagmatic), leading to widespread pyroclastic deposits, but later eruptions in any one area were dominantly of lavas. Pillow lavas and hyaloclastites at Kinghorn testify to local subaqueous eruptions, but most of the lavas were probably subaerial.

The eastern Midland Valley (early Visean)

Outcrops of the Garleton Hills Volcanic Formation of East Lothian lie entirely between the projected north-easterly continuation of the Southern Upland Fault at depth and NE-trending splays to the south-east, such as the Dunbar–Gifford Fault (McAdam and Tulloch, 1985; Davies *et al.*, 1986). They therefore overlie Lower Palaeozoic rocks of the Southern Uplands terrane at no great depth. Superb coastal exposures around North Berwick show the relationships of the basal basaltic pyroclastic rocks to associated necks and sedimentary country rocks (see **North Berwick Coast** GCR site report), and overlying basaltic to trachytic lavas form the **Garleton Hills** (see GCR site report). The sequence is up to 520 m thick. Thinner successions have been encountered to the south-west in the Spilmersford and D'Arcy boreholes and in a small outcrop near Borthwick, still within the same fault-bound block.

Some flows of analcime trachybasalt in East Lothian appear to have contained leucite originally (Bennett, 1945). Apart from one other flow in the Campsie Fells, which is phonolitic, these are the only known silica-undersaturated evolved lavas in the Dinantian lava successions. However, the East Lothian lava field is unusual because of its apparent association, backed by limited K-Ar whole-rock dates (De Souza, 1974, 1979), with several large high-level intrusions (plugs and laccoliths) of silica-undersaturated phonolitic rocks. The latter form the prominent landmarks of Traprain Law, North Berwick Law and the Bass Rock (e.g. see Figure 2.6 – **North Berwick Coast** GCR site report) as well as a sill at Hairy Craig (Bailey in Clough *et al.*, 1910; MacGregor and Ennos, 1922; Campbell and Stenhouse, 1933; McAdam and Tulloch, 1985). They are represented by the **Traprain Law** GCR site.

In Edinburgh, the lavas around Arthur's Seat and Calton Hill, together with associated intrusions such as the Lion's Head and Lion's Haunch vents and the basalt plug of the Castle Rock, dominate the city landscape, and the **Arthur's Seat Volcano** GCR site is one of the most widely appreciated geological localities in Britain. At least 13 lavas, ranging from olivine-clinopyroxene-phyric basalts to hawaiite and mugearite, form a succession 400–500 m thick (Clark, 1956; Black, 1966). A more restricted, 90 m-thick succession of tuffs and olivine-clinopyroxene-phyric basalts forms Craiglockhart Hill, 6 km to the south-west. Together these volcanic sequences comprise the Arthur's Seat Volcanic Formation.

Burntisland and Bathgate Hills (early Visean to Namurian)

In the Burntisland area of Fife, two volcanic developments within the Anstruther Formation are probably younger than the Arthur's Seat and Garleton Hills volcanic formations of the Lothians (Figure 2.1). At the base of the formation, the Charles Hill Volcanic Member consists of tuffs and olivine-microphyric basalts that crop out on the limbs of a shallow anticline centred upon the island of Inchcolm. The higher unit, of coarse tuffs and agglomerates, is known from boreholes and poor exposures onshore to the north. Major outcrops of late Visean volcanic rocks occur around the Burntisland Anticline, where up to 485 m of olivine-microphyric basalt lavas ('Dalmeny' and 'Hillhouse' types) with subordinate pyroclastic rocks and volcaniclastic sedimentary rocks constitute the Kinghorn Volcanic Formation. This formation, which is represented onshore by the **Burntisland to Kinghorn Coast** GCR site, is also well developed offshore to the east, as was seen underground in Seafield Colliery, and on the island of Inchkeith in the Firth of Forth. The succession is dominantly subaerial, but with periodic submergence beneath freshwater lakes or marine incursions during which some pillow lavas and hyaloclastites were formed.

Within and around the West Lothian oil-shale field, volcanic activity was of a distinctly different nature to the generally earlier Visean activity elsewhere in the Midland Valley. In contrast to the wide compositional variety within the Garleton Hills and most Clyde Plateau successions, this volcanism was entirely basaltic. Initial activity may have been contemporaneous with later phases of the essentially subaerial Clyde Plateau Volcanic Formation, but subsequently the terrestrial lava pile was overlapped and the volcanism continued to develop in a coastal-plain–lagoonal–shallow-marine environment that was a precursor to Silesian volcanic settings. The earlier phases of this activity are poorly exposed and much information has come from boreholes and underground workings (Mitchell and Mykura, 1962). Although later lavas are well exposed in parts of the Bathgate Hills, no suitable GCR site has been identified.

The Crosswood Ash, known from exposures and boreholes around Crosswood Reservoir, occurs at the base of the West Lothian Oil-shale Formation, and the 100 m-thick Seafield–Deans Ash of the West Calder area underlies the freshwater Burdiehouse Limestone that marks the base of the Hopetoun Member slightly higher in that formation. Other thin but widespread volcaniclastic beds occur within the Hopetoun Member (e.g. the Port Edgar Ash and Barracks Ash) and a basalt lava occurs at this general stratigraphical level near Carlops, in the Midlothian Basin. Farther west, in an oil-well at Salsburgh in the Central Coalfield, supposed Lower Devonian volcanic rocks are overlain by 100 m of basaltic tuffs and lavas with interbeds of limestone and mudstone that have been termed the 'Salsburgh Volcanic Formation' (Cameron *et al.*, 1998). These are succeeded directly by the Burdiehouse Limestone and hence are probably contemporaneous with the Seafield–Deans Ash.

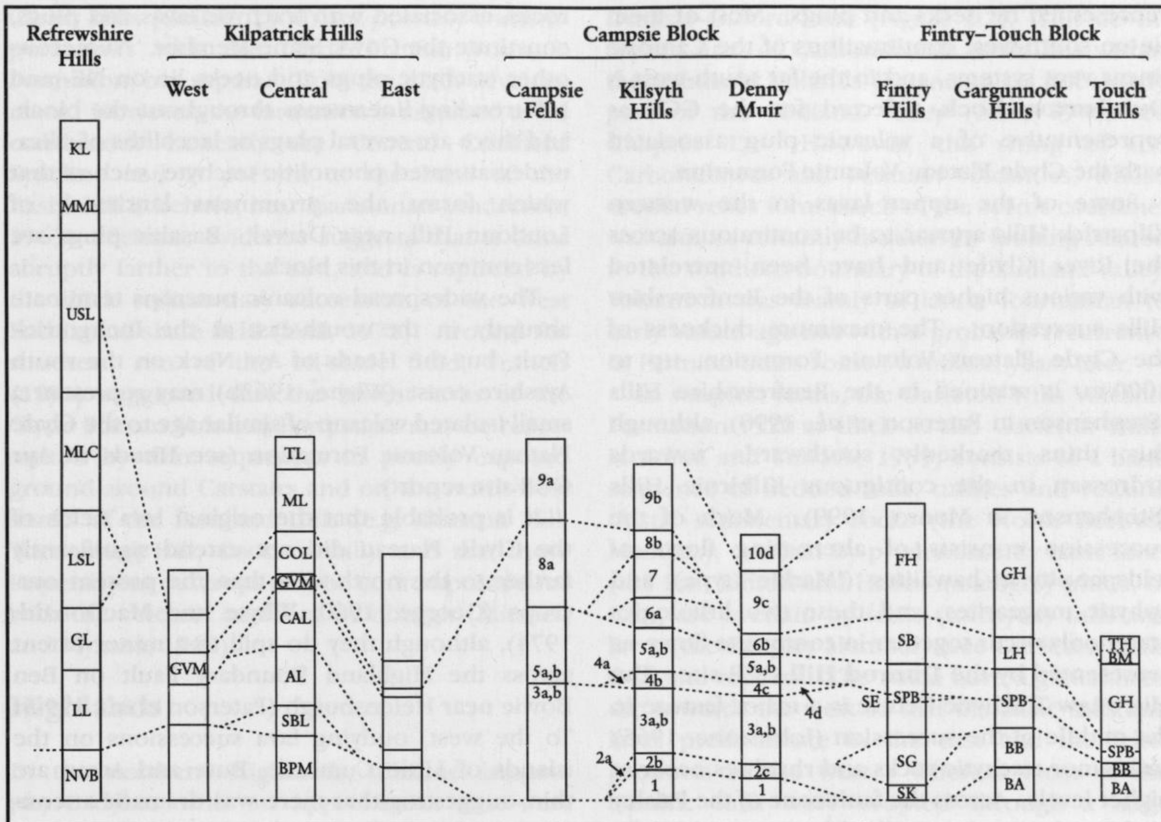
Higher still in the Hopetoun Member, above the Houston Marls, thick and widespread pyroclastic rocks mark the base of the Bathgate Hills Volcanic Formation. Beneath the Central Coalfield to the west, in the Rashiehill Borehole, this major volcanic formation rests directly upon the Clyde Plateau Volcanic Formation. In the Bathgate Hills area it interdigitates with the Visean sedimentary succession and extends well into the Namurian, accumulating a total thickness of about 600 m of volcanic rocks (Cadell, 1925; Smith *et al.*, 1994; Stephenson in Cameron *et al.*, 1998). Rather than split the description between chapters, the formation is described in its entirety in the 'Introduction' to Chapter 4.

The western Midland Valley (early Visean)

The Clyde Plateau Volcanic Formation comprises the major part of the Strathclyde Group in the western Midland Valley. Its extensive main outcrop encircles Glasgow on three sides, forming the Touch, Fintry and Gargunnock hills, the Campsie Fells and the Kilpatrick Hills to the north, and the Renfrewshire Hills, the Beith–Barrhead Hills and the Dunlop–Eaglesham–Strathaven Hills to the south (Figure 2.2). The outcrop is divided by major faults into several discrete 'blocks', each with its own succession. Most have been described in some detail in Geological Survey memoirs and some attempt has been made to correlate parts of successions between blocks, although in some cases this is extremely tentative (Figure 2.4).

In the north-east of the outcrop the Gargunnock, Touch and Fintry hills form a coherent block (the Fintry–Touch Block, see Figure 2.4) with a volcanic sequence that is 300–400 m thick. The lavas are mainly of feldspar-phyric basalts and hawaiites ('Markle' and 'Jedburgh' types), with subordinate trachybasalts and mugearites (Read in Francis *et al.*, 1970). Volcanic necks and proximal volcaniclastic rocks are rare, so the **Touch, Fintry and Gargunnock Hills** GCR site represents part of the Clyde Plateau Volcanic Formation that may have originated mainly from fissure eruptions. The Campsie Fells, Kilsyth Hills and Denny Muir, forming the next block to the south-west (the Campsie Block), have a sequence in excess of 500 m thick. A wide range of lava compositions from olivine-clinopyroxene-phyric basalts to trachyte and a rare phonolitic trachyte are represented, though here too feldspar-phyric basalts and hawaiites are dominant and parts of the succession can be traced into the Touch Hills (Craig, 1980; Hall in Forsyth *et al.*, 1996; Hall *et al.*, 1998). Numerous volcanic necks are concentrated along four NE-trending 'linear vent systems' and the **Campsie Fells** GCR site has been selected to represent this multiple vent volcanism. To the east of the GCR site is the major Waterhead Central Volcanic Complex, which dominated the later extrusive phases and may have developed a caldera.

The sequence in the Kilpatrick Hills Block is separated from that of the Campsie Fells by the E–W-trending Campsie Fault, and correlations between these two blocks are only tentative.



Renfrewshire Hills

KL	Kilbarchan Lava Member
MML	Marshall Moor Lava Member
USL	upper Strathgryfe Lava Member
MLC	Misty Law Trachytic Member
LSL	lower Strathgryfe Lava Member
GL	Greeto Lava Member
LL	Largs Lava Member
NVB	Noddsdale Volcaniclastic Member

Kilpatrick Hills

TL	Tambowie Lava Member
ML	Mugdock Lava Member
COL	Cochno Lava Member
GVM	Greenside Volcaniclastic Member
CAL	Carbeth Lava Member
AL	Auchineden Lava Member
SBL	Saughen Braes Lava Member
BPM	Burncrooks Volcaniclastic Member

Campsie Block

10d	Garvald Lava Member
9c	Denny Muir Lava Member
9b	Kilsyth Hills Lava Member
9a	Holehead Lava Member
8b	Upper Lecket Hill Lava Member
8a	Fin Glen Lava Member
7	Boyd's Burn Lava Member
6b	Langhill Lava Member
6a	Lower Leckett Hill Lava Member
5b	Upper South Campsie Lava Member
5a	Upper North Campsie Lava Member
4d	Overton Lava Member
4c	Laird's Hill Lava Member
4b	Loup of Fintry Lava Member
4a	Craigentimpin Lava Member
3b	Lower South Campsie Lava Member
3a	Lower North Campsie Lava Member
2c	Carron Bridge Lava Member
2b	Laird's Loup Lava Member
2a	Burnhouse Lava Member
1	Drumnessie Lava Member

Fintry–Touch Block

TH	Touch House Member
BM	Black Mount Member
GH	Gargunnock Hills Member
FH	Fintry Hills Member
SB	Shelloch Burn Member
LH	Lees Hill Member
SPB	Spout of Ballochleam Member
SE	Stronend Interbasaltic Member
SG	Slackgun Interbasaltic Member
BB	Baston Burn Member
SK	Skiddaw Member
BA	Basal Member

Figure 2.4 Correlation of composite sections in the Clyde Plateau Volcanic Formation. Based on information in Forsyth *et al.* (1996); Hall *et al.* (1998); and Paterson *et al.* (1990). N.B. formal designation of these units as members is currently in progress.

The 400 m-thick Kilpatrick sequence includes many olivine- and olivine-clinopyroxene-phyric basalts ('Dalmeny', 'Dunsapie' and 'Craiglockhart' types) and is generally more mafic than that of the Campsie Block. Many lavas thin eastwards towards Strath Blane, where Hall *et al.* (1998) suggested that high ground, possibly formed by early tephra cones, formed a barrier. Later lavas, possibly emanating from the Waterhead Central Volcanic Complex, can be correlated across Strath Blane, suggesting that the barrier had become ineffective by this time. Many of the lavas in the Kilpatrick Hills originated from small central volcanoes, now represented by necks and plugs. Most of them lie on south-west continuations of the Campsie linear vent systems, and in the far south-west is **Dumbarton Rock**, selected for the GCR as representative of a volcanic plug associated with the Clyde Plateau Volcanic Formation.

Some of the upper lavas in the western Kilpatrick Hills appear to be continuous across the River Clyde and have been correlated with various higher parts of the Renfrewshire Hills succession. The maximum thickness of the Clyde Plateau Volcanic Formation, up to 1000 m, is attained in the Renfrewshire Hills (Stephenson in Paterson *et al.*, 1990), although this thins markedly southwards towards Ardrossan in the contiguous Kilbirnie Hills (Stephenson in Monro, 1999). Much of the succession consists of alternating flows of feldspar-phyric hawaiites ('Markle' type) and aphyric mugearites, and these two lithologies commonly occur together in composite flows, as represented by the **Dunrod Hill** GCR site. The Misty Law Trachytic Centre is a major feature in the middle of the succession (Johnstone, 1965) and minor trachytic rocks and rhyolites occur at higher levels. Across the fault zone of the Paisley Ruck, to the south-east, the lava sequence in the Beith–Barrhead Hills is probably less than 300 m thick, being arranged in a gentle SW-plunging anticline. The varied but essentially basaltic lavas may be equivalent to only the upper, more mafic part of the adjoining Renfrewshire Hills succession (Stephenson in Paterson *et al.*, 1990; in Hall *et al.*, 1998; and in Monro, 1999). This contrasts with the view of De Souza (1979), based on K-Ar age determinations (337 ± 7 Ma, or *c.* 344 Ma using new constants) and with the Ar-Ar date of 335 ± 2 Ma (A.A. Monaghan and M.S. Pringle, pers. comm., 2002), which both suggest that the Beith Hills include some of the oldest lavas of the Clyde Plateau Volcanic Formation. Farther south-east, across the Dusk Water Fault, lies a vast outcrop of poorly exposed varied volcanic rocks that form undulating high moorland between Dunlop, Eaglesham and Strathaven. Their stratigraphy and structure are difficult to determine but thickness estimates range from 500 m to 900 m and a full range of lithologies is present, from mafic basalts to trachyandesites, trachytes and rhyolites (Richey *et al.*, 1930; Paterson *et al.*, 1998; MacPherson *et al.*, 2001). Towards the east of this block, widespread proximal pyroclastic rocks and volcaniclastic sedimentary rocks, associated with trachytic lavas and plugs, constitute the Gowk Stane Member. Numerous other trachytic plugs and necks lie on NE- and NW-trending lineaments throughout the block, and there are several plugs or laccoliths of silica-undersaturated phonolitic trachyte, such as that which forms the prominent landmark of Loudoun Hill, near Darvel. Basaltic plugs are less common in this block.

The widespread volcanic outcrops terminate abruptly in the south-east at the Inchgotrick Fault, but the Heads of Ayr Neck on the south Ayrshire coast (Whyte, 1963b) may represent a small isolated volcano of similar age to the Clyde Plateau Volcanic Formation (see **Heads of Ayr** GCR site report).

It is probable that the original lava fields of the Clyde Plateau did not extend significantly farther to the north-west than the present outcrops (George, 1960; Whyte and MacDonald, 1974), although they do spill to a minor extent across the Highland Boundary Fault on Ben Bowie near Helensburgh (Paterson *et al.*, 1990). To the west, outlying lava successions on the islands of Little Cumbrae, Bute and Arran are thin, suggesting that there was dramatic attenuation westwards from the main outcrops. Significantly, on Bute and Arran, up to 16 m of fluvio-deltaic mudstones with sandstones and thin coals occur beneath the volcanic rocks, but with a sharp erosional base above the Clyde Sandstone Formation, suggesting that the onset of volcanism was a little later here than in the main outcrop. However, in south Kintyre a highly varied volcanic succession, ranging from mafic basalts to trachyte and up to 400 m thick, rests upon the Kinnesswood Formation, Lower Old Red Sandstone and Dalradian rocks, north-west of the Highland Boundary Fault (McCallien, 1927). The Kintyre succession is overlain unconformably by volcaniclastic detritus akin to the Kirkwood Formation and by Lower Limestone Formation strata and hence has been assigned to the Clyde Plateau Volcanic Formation, though it is likely that it constitutes an entirely separate lava field (see **Machrihanish Coast and South Kintyre** GCR site report). In fact it is closer to the Visean volcanic rocks at Ballycastle in Northern Ireland (Wilson and Robbie, 1966) than it is to the main outcrops of the Clyde Plateau.

East of Stirling, thin sequences of tuffs, basalts and felsic lavas close to the Ochil Fault near Dollar (Browne and Thirlwall, 1981) and in the

Cleish Hills (Geikie, 1900) are possibly contemporaneous with the Clyde Plateau Volcanic Formation, but separate from the main development. However, the formation is assumed to be continuous beneath the Central Coalfield Syncline and is present at the base of the Rashiehill Borehole, near Slamannan (Anderson, 1963). Seismic evidence suggests that it thins abruptly farther to the east, and is replaced by the thick sedimentary succession of the West Lothian oil-shale field (Hall, 1971). Around the southern rim of the oil-shale field, Francis (1991) suggested that the lowest lavas of the Clyde Plateau Volcanic Formation may be represented by thin sequences in poorly exposed ground around Carstairs and on the north-west flanks of the Pentland Hills (e.g. Corston Hill, Torweaving Hill, Cockburnhill), and that these impersistent outcrops may be contemporaneous with the volcanic rocks of Edinburgh (Mitchell and Mykura, 1962).

Highlands

In the northern Highlands, alkaline lamprophyre dykes with an approximate east–west trend (see 'Introduction' to Chapter 5) include some that have been assigned an age of *c.* 326 Ma (Baxter and Mitchell, 1984; Esang and Piper, 1984), making them contemporaneous with some of the Visean activity farther south. It is therefore possible that some of the diatremes in the northern Highlands that appear to be associated with these dykes are also of Visean age (see 'Introduction' to Chapter 4).

NORTH BERWICK COAST, EAST LOTHIAN (NT 496 858–NT 624 829)

B.G.J. Upton

Introduction

The North Berwick Coast GCR site, extending for some 17 km between Fidra and Dunbar along the coast of East Lothian, exposes a succession from sedimentary rocks of the Ballagan Formation up through the dominantly volcanic Garleton Hills Volcanic Formation to the unconformity with the overlying sedimentary Gullane Formation. The volcanic rocks form part of the East Lothian volcanic field that crops out between the extrapolation at depth of the main splay of the Southern Upland Fault and the Dunbar–Gifford Fault (Figure 2.2). This down-faulted area of Carboniferous strata lies within the Southern Uplands Terrane and is not strictly part of the Midland Valley (Max, 1976; see Chapter 1). However, the siting of the Carboniferous and Permian volcanoes, whose eroded relics form much of the scenic coastline, was almost certainly dictated by faulting related to the southern boundary of the Midland Valley. Volcanism was mainly of latest Tournaisian to early Visean age but with a probable recurrence in Permian times some 50 million years later.

In simplest terms, the Garleton Hills Volcanic Formation (520 m thick in the Garleton Hills; McAdam and Tulloch, 1985) consists of a basal sequence of bedded tuffs, tuffites and volcaniclastic sedimentary rocks (the North Berwick Member), overlain by predominantly basic lavas (the East Linton and Hailes members) which, in turn, are overlain by felsic (trachytic) tuffs and lavas (the Bangley Member) (see **Garleton Hills** GCR site report). The strata have a generalized dip towards the west so that the basic tuffs and lavas predominate in the east and the felsic products in the west. Whereas most of the basaltic tuffs were erupted through a large number of small volcanoes whose eroded relics are now seen as vents, larger volcanoes, from which the lavas and trachytic tuffs were erupted, developed at a later stage. Some 14 volcanic vents have been recognized along this coast and it may be supposed that many more lie both out to sea and inland beneath drift deposits.

Inland exposures, other than of trachytes in the Garleton Hills, are generally poor whereas the largely unspoiled and rocky coast affords excellent sections. Although some of the outcrops are above high-water mark, many lie in the intertidal zone. Four prominent islands lie off this coast and outwith the North Berwick Coast GCR site; these are, from west to east, Fidra, The Lamb, Craigleith and Bass Rock. Of these, the first three consist of alkali basalt or dolerite sills whilst the Bass Rock is made of phonolitic trachyte, probably forming a sub-cylindrical stock.

Studies of the East Lothian volcanic rocks have not made the same impact on the history of geology as those closer to Edinburgh (see **Arthur's Seat Volcano** GCR site report) or on the opposite side of the Firth of Forth (see **East Fife Coast** GCR site report). However, they are of first-order importance in presenting (1) a superb set of shallowly eroded Late Palaeozoic tuff-rings, (2) some remarkable welded ash-fall tuffs, and (3) inclusions of middle- to lower-crustal and upper-mantle material, allowing insight into the rock-types present at depth that have been sampled by rising magmas. The North Berwick Coast GCR site also includes two localities, Oxroads Bay and Weaklaw, where plant remains of international importance are preserved in the volcaniclastic rocks (Cleal and Thomas, 1995). The earliest descriptions of the volcanic rocks were in the Geological Survey sheet memoir (Howell *et al.*, 1866), but considerably more detail was given in the second edition (Clough *et al.*, 1910). The section was subsequently investigated by T.C. Day, who was responsible for the identification of many of the volcanic vents and published a series of detailed papers between 1916 and 1936. The most recent Geological Survey revision of the area led to an overall appraisal of the volcanism by Martin (1955), which synthesized the work of Day and formed the basis for the current maps and memoirs (McAdam and Tulloch, 1985; Davies *et al.*, 1986). More recent work has concentrated upon the inclusions of crustal and upper-mantle material that are common in many of the vents and intrusions (Upton *et al.*, 1976, 1984, 1999; Graham and Upton, 1978; Halliday *et al.*, 1984; Hunter *et al.*, 1984; Aspen *et al.*, 1990). The section is a popular venue for field excursions and is described in various field guides (McAdam *et al.* in Upton, 1969; McAdam in McAdam and Clarkson, 1986).

Description

Since there is a generalized younging of strata from east to west, the sites are described in this sequence (Figure 2.5).

Scoughall Rocks, at the south-east extremity of the GCR site (NT 623 829), expose the '*Pilmour Volcano*', a 900 m-broad vent containing blocks of bedded pyroclastic rock, tuffite, marl and sandstone, up to 100 m across, together with several small basaltic intrusions (McAdam and Clarkson, 1986; Davies *et al.*, 1986). Between Scoughall Rocks and The Car peninsula (NT 610 850) are marls and sandstones with intercalations of fine-grained volcaniclastic sedimentary rocks and tuffites. The concentration of clasts of tuffite and lava increases and boulders of lava with big augite phenocrysts become abundant towards The Car. Igneous clasts, present in abundance along particular horizons in the red marls outside the vents, give them the appearance of agglomerates but they were thought by Clough *et al.* (1910) to have been deposited from 'sheet-floods' of great violence. The bedded rocks are transected by two small vents known as the Scoughall and Seacliff Tower vents. The eastern part of the *Scoughall Vent* contains red unbedded tuffite, blocks of sandstone and basaltic bombs. The exposed eastern part of the *Seacliff Tower Vent* is occupied by red pyroclastic breccia. Large tuffite blocks are prominent in this vent, together with basaltic bombs and fragments of sandstone and dolostone ('cementstone').

Day (1930b) described two vents at *The Car*, which cut sandstones, tuffites and ostracode-bearing marls (Clough *et al.*, 1910). The older of the two vents forms the end of the tidal peninsula whilst the cross-cutting younger vent forms the eminence known as 'Great Car'. The older vent contains stratified tuffite and volcaniclastic breccias dipping uniformly towards the north-west at *c.* 50°. The younger vent is likewise occupied by stratified pyroclastic materials that dip inwards to the north or north-west at 30°–70°. The clasts in these vents are highly scoriaceous with conspicuous augite phenocrysts. Described by Day (1930b) as limburgitic, they were re-defined as leucite basanites by Balsillie (1936). The same rock-type also occurs as irregular masses intruding the northern vent and one of these forms the islet of St Baldred's Boat to the east of The Car. A third vent on the south-east side of the peninsula, partially exposed at very low tide, was recorded by Day (1930b).

Some 500 m south-west of The Car, further exposures of pyroclastic rocks form The Gegan (NT 603 848) and the headland at Seacliff Harbour. At the latter, red tuffites show large-scale cross-bedding. Immediately inland, on the steep coastal slopes, are two basaltic intrusions. Whereas the Primrose Bank body is an olivine-augite-phyric basaltic sill, it is unclear whether the aphyric basalt around the old quarry at Auldhame is part of a plug or a sill. The intrusion contains celestine-bearing veins.

Between The Gegan and Oxroad Bay is a down-faulted mass of mainly greenish tuffites with intercalated dolostones ('cementstones'), ripple-bedded sandstone and impure limestone that correlates with strata cropping out farther west near North Berwick. Bedding is highly disturbed by faulting, with dips of up to 60° in varying directions. The green colouration is due

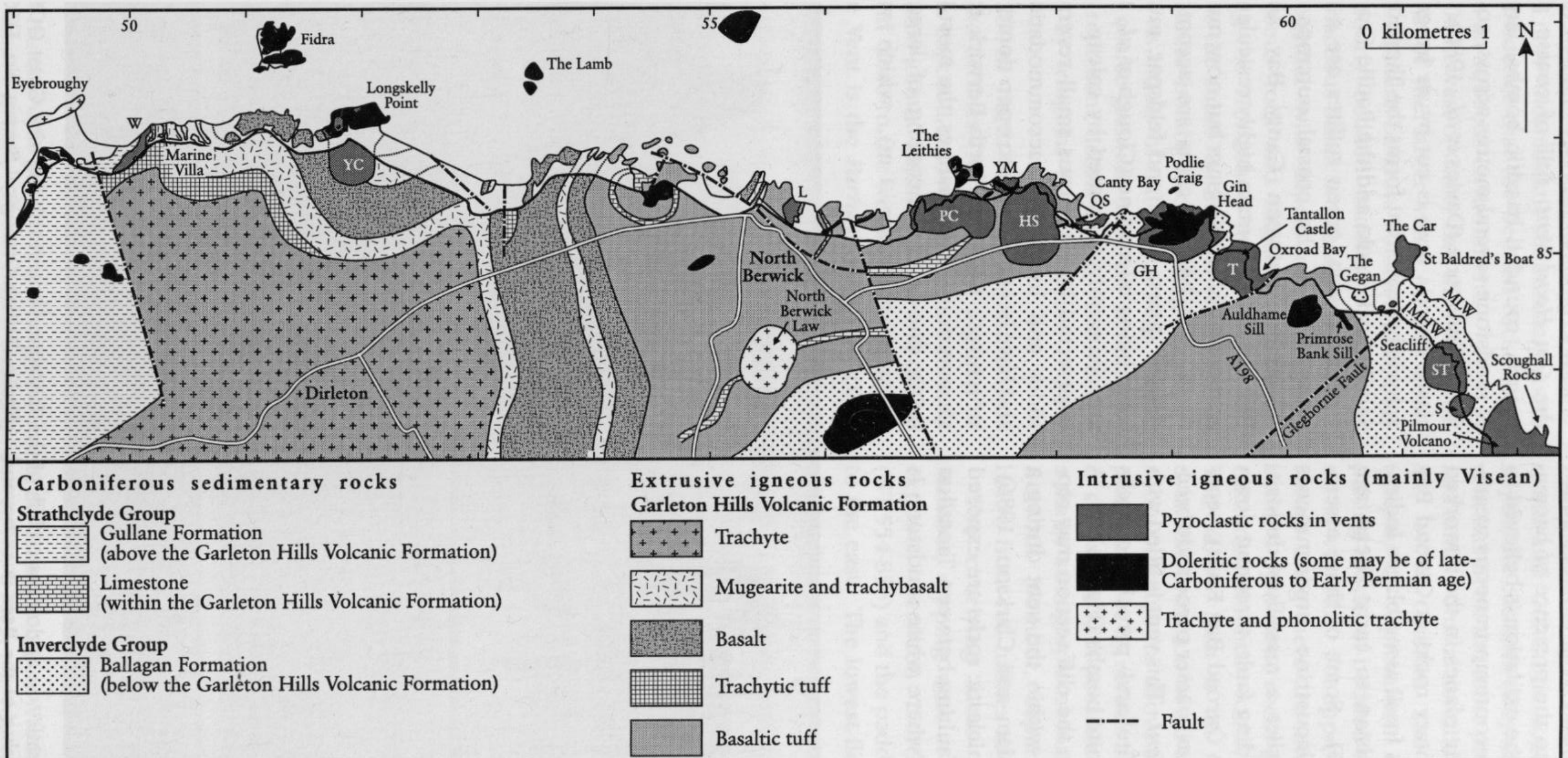


Figure 2.5 Map of the area around the North Berwick Coast GCR site. (GH = Gin Head Vent; HS = Horseshoe Vent; L = The Lecks Vent; PC = Partan Craig Vent; QS = Quarrel Sands Vent; S = Scoughall Vent; ST = Seacliff Tower Vent; T = Tantallon Vent; W = Weaklaw Vent; YC = Yellow Craig Plantation Vent; YM = Yellow Man Vent.) After McAdam (in McAdam and Clarkson, 1986); and British Geological Survey 1:50 000 sheets 33W, Haddington (1985); and 33E, Dunbar (1986).

to the presence of chlorite; more oxidized equivalents are red owing to the presence of haematite. At least some of the oxidation is believed to have been due to penecontemporaneous weathering. Several bedding planes in the reworked volcaniclastic sedimentary rocks at Oxroad Bay reveal important plant fossil assemblages indicative of a late Tournaisian to early Visean age (Bateman *et al.*, 1995). Some of these assemblages formed in lacustrine environments whereas another overlies a mass-flow deposit. An important SW-trending fault, throwing down to the south-east (the Oxroad Bay Fault), separates green tuffites from coarser pyroclastic breccias of the *Tantallon Vent*. This vent is filled with unbedded green tuffite and pyroclastic rocks containing small basanite bombs (Figure 2.6). A sandstone lens high in the cliff section may represent sedimentation within the vent during a dormant phase (McAdam and Clarkson, 1986). Beds below the volcaniclastic rocks are exposed locally as a result of faulting between Tantallon Castle and Gin Head where white sandstone is exposed.

To the west, cutting red and green tuffites, lies the *Gin Head Vent*, full of coarse pyroclastic breccia, rounded basaltic bombs, large masses of sandstone several metres across, and a fossiliferous limestone (Davies *et al.*, 1986). The pyroclastic breccia here is cut by an irregular sill of aphyric basanite that forms the Tapped Rock and Saddle Rock stacks and the Podlie Craig islet.

Farther west, green tuffites are a dominant component of the coastal outcrops for some kilometres between Canty Bay and North Berwick. They contain highly vesicular particles, originally of basaltic glass but now much chloritized. The basaltic clasts are accompanied by detrital quartz, mica and feldspar, much of the latter being microcline (Clough *et al.*, 1910). All are commonly cemented by calcite. They are intersected by numerous small reversed faults, slides and low-angled accommodation planes and have been folded into open dome and basin structures. Close to North Berwick the tuffites show various facies changes, the most important being due to the incoming of larger basaltic clasts (some over 1 m), which vary from

Figure 2.6 Tantallon Castle, on agglomerate cliffs of the Tantallon Vent, North Berwick Coast GCR site, with the phonolitic trachyte plug of the Bass Rock beyond. (Photo: British Geological Survey, No. D3665, reproduced with the permission of the Director, British Geological Survey, © NERC.)

rounded to angular. The interbedding of vocaniclastic and non-volcaniclastic sedimentary rocks is further evidence that the bedded rocks accumulated in an aqueous environment.

Some 500 m west of Canty Bay, the *Horseshoe Vent* is one of the larger vents on this coast section. It contains basanite bombs and coarse-grained, poorly bedded material. Blocks of highly scoriaceous basalt are characteristic (Day, 1928a), together with blocks of sedimentary and tuffaceous rock and fragments of wood. Intersecting the Horseshoe Vent on its north-west side is the small *Yellow Man Vent*, notable for its basanite bombs and the large size (up to 3 m) of its blocks of tuff (Day, 1925). Some layers of coarse-grained material within the vent may represent debris flows. The vent is transected by irregular basanite dykes (Figure 2.7).

Some 200 m north-west of the Horseshoe Vent lie The Leithies (NT 573 858), a group of small islets exposed at low tide that are formed by a columnar-jointed basanite sill. South of The Leithies and close to the western margin of the Horseshoe Vent is the *Partan Craig Vent* (Day, 1925), containing blocks of bedded tuffite, siltstone and dolostone together with bombs of nepheline basanite (Day and Bailey, 1928). This vent has been the most prolific source of lower- and middle-crustal xenoliths in Britain (Upton *et al*., 1976; Graham and Upton, 1978) (see Chapter 1). A prominent debris-flow layer comprising blocks of bedded tuffite is well exposed on its west side. For approximately 1 km west of Partan Craig there are intermittent outcrops of bedded pyroclastic rocks (Figure 2.8) and intercalated sedimentary rocks, cut by the *Lecks Vent*. The Yellow Man is an irregular NE-trending dyke-like intrusion of olivine basalt that forms a rocky prominence above the beach. A wave-cut platform close by reveals a *c*. 3 m-thick sequence of dolostone with possible algal growths. This is a major stratigraphical marker within the North Berwick Member, separating green tuffites below from red tuffites above (Martin, 1955).

The overlying lavas of the East Linton Member, dipping west at *c*. 20°, are very well exposed between North Berwick harbour (NT 554 856) and the paddling pool some 100 m to the east. The lowest flow is *c*. 4 m thick and

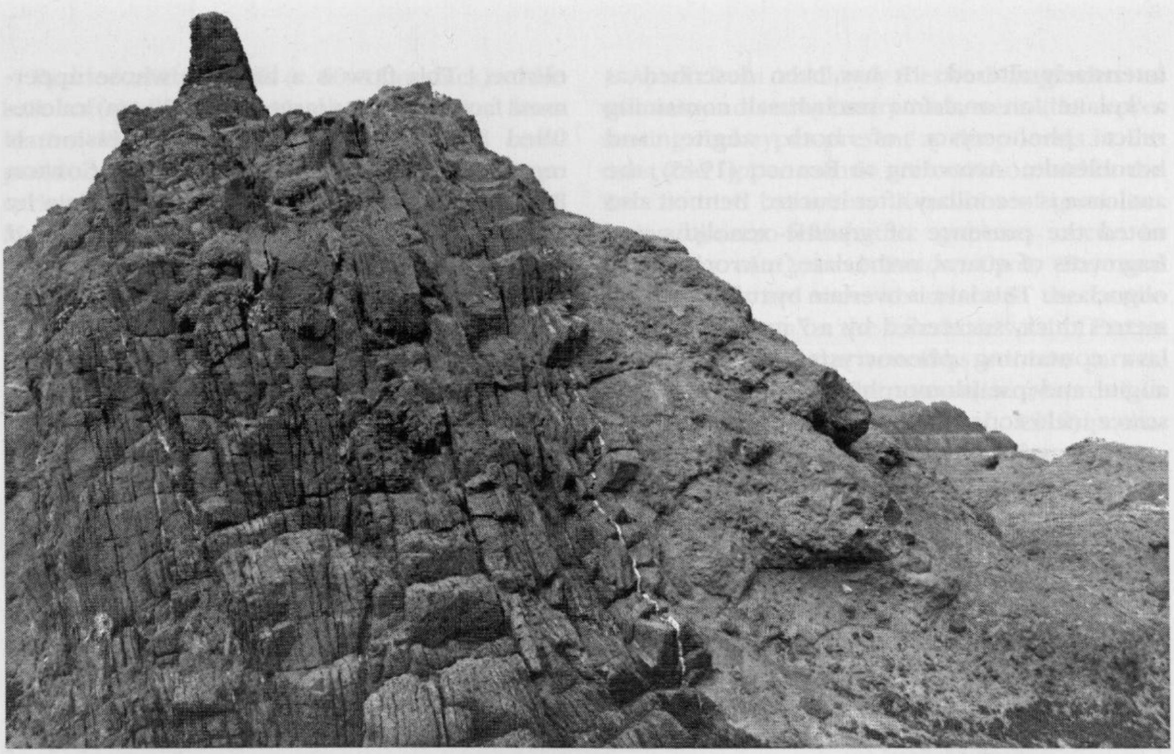


Figure 2.7 Basanite dyke (left) cutting vent agglomerate of the Yellow Man Vent, North Berwick Coast GCR site. The cliff is about 6 m high. (Photo: British Geological Survey, No. D1113, reproduced with the permission of the Director, British Geological Survey, © NERC.)

Figure 2.8 Basaltic bombs (rounded) and blocks (angular) in red, bedded basaltic tuffs at The Lecks, North Berwick Coast GCR site. The hammer head is about 15 cm long. (Photo: British Geological Survey, No. D3044, reproduced with the permission of the Director, British Geological Survey, © NERC.)

intensively altered. It has been described as a 'kulaite', an analcime trachybasalt containing relict phenocrysts of both augite and hornblende. According to Bennett (1945), the analcime is secondary after leucite. Bennett also noted the presence of granitic xenoliths and fragments of quartz, orthoclase, microcline and oligoclase. This lava is overlain by tuffite, several metres thick, succeeded by a 7 m-thick basaltic lava containing phenocrysts of plagioclase, augite and pseudomorphed olivine as well as scarce inclusions (?autoliths) of gabbro.

The basaltic lava is directly overlain by a mugearite lava some 10 m thick with a well-preserved autobrecciated upper surface that marks the base of the Hailes Member (Figure 2.9). The vesicles (now amygdales) in this flow top have been stretched out into elongate, tubular forms, and individual lava blocks have been rotated during flow. The original surface clearly experienced minimal weathering before it was over-ridden by the next lava flow. The latter is *c.* 17 m thick and contains an abundance of plagioclase phenocrysts in association with smaller, scarcer phenocrysts of iddingsitized olivine. This flow is a hawaiite whose uppermost facies contains large (up to 20 cm) calcite-filled amygdales. This lava succession is repeated by faulting to the west at Cowton Rocks, where a porphyritic hawaiite overlain by a mugearite flow whose fissile nature and diagenetic ovoid structures, produced by concentric bands of haematite, are similar to those seen in the mugearite at North Berwick. The mugearite lava can be followed west as far as Marine Villa (NT 503 859) where the original rough and jagged flow top is perfectly preserved beneath bedded trachytic tuffs.

These lavas are cut by the *Yellow Craig Plantation Vent*, composed of poorly stratified tuffite with associated basanite intrusions, the more prominent of which form Longskelly Point (NT 522 863) and the hillock of Yellow Craig (Day, 1932c). Fidra island, lying north-west of the Yellow Craig Plantation Vent, consists of a thick, columnar-jointed basanite sill. The cobble beach on the mainland south-west of the island is largely composed of basanite pebbles, which are almost certainly derived from Fidra or its offshore extensions.

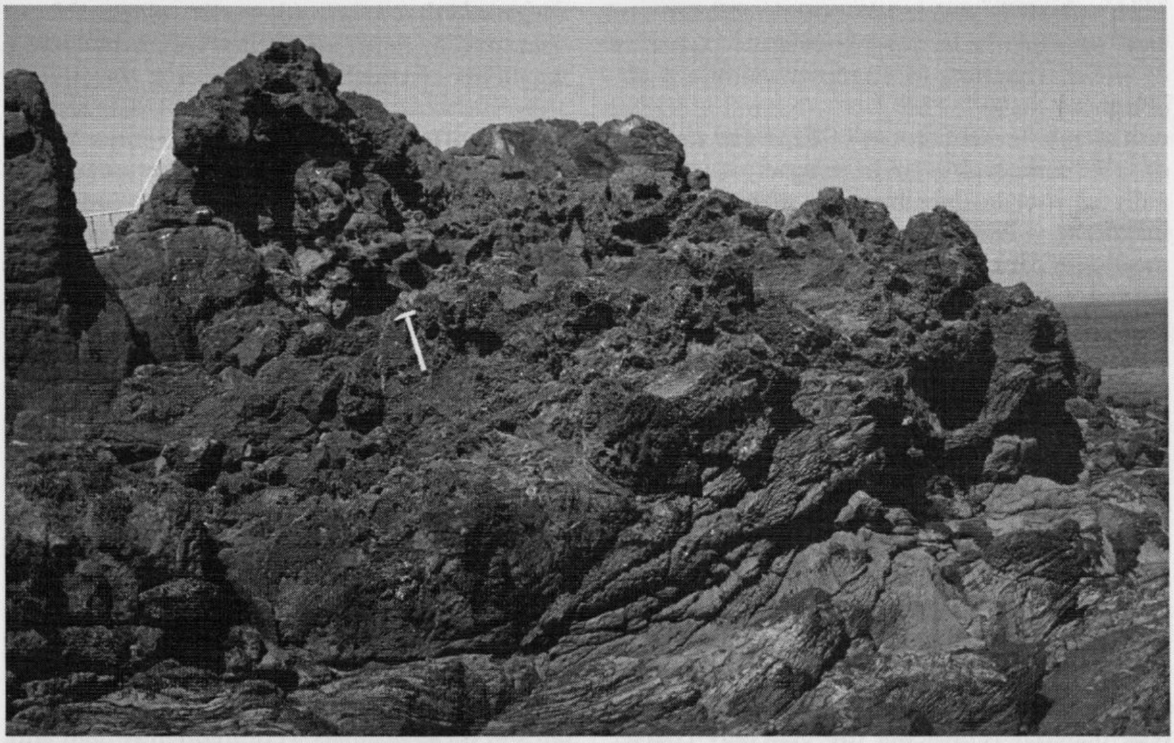


Figure 2.9 Mugearite lava at North Berwick, showing flow lamination in the main body of the flow and a slaggy, amygdaloidal flow top. The hammer shaft is about 35 cm long. (Photo: British Geological Survey, No. D3041, reproduced with the permission of the Director, British Geological Survey, © NERC.)

Bedded trachyte tuffs (*c.* 10–15 m thick) at the base of the Bangley Member crop out from Marine Villa westwards. Their original compositions have been largely obscured by later carbonation and oxidation to haematite. Mugearitic and trachytic clasts in these tuffs are mainly angular but some layers are rich in lenticular particles flattened parallel to the bedding. These lenticles were probably lapilli of very fluid pumice that were deformed beneath accumulating tuff layers above, i.e. they are welded ash-fall tuffs.

The *Weaklaw Vent*, which cuts the bedded trachytic tuffs, contains poorly bedded sandy tuffaceous breccia. Basaltic rocks in the vent have been intensely carbonated. Plant fossils are particularly well preserved (Bateman *et al.*, 1995) and Gordon (1935) considered that the plants grew on the flanks of an active volcano and were killed by ash flows. Immediately west of the vent a 5 m-thick flow of vesicular porphyritic trachyte overlies the bedded trachytic tuffs and is itself overlain by sedimentary rocks that mark the start of a prolonged period of magmatic inactivity.

Apart from shallowly derived fragments of volcanic and sedimentary rock, xenoliths representing rock-types present at deep levels in the crust and upper mantle are common within the East Lothian basanites. Xenoliths of granulite-facies quartzo-feldspathic gneisses occur in intrusions at Quarrel Sands, Canty Bay and Weaklaw but are most abundant in the coarse pyroclastic strata on the foreshore west of Partan Craig. Some of the xenoliths contain garnet, kyanite and rutile. The microcline recorded in the bedded tuffites (Clough *et al.*, 1910) probably came from disaggregation of these gneisses. Xenoliths of other, more basic granulite-facies gneisses, comprising plagioclase together with one or two types of pyroxene and subordinate magnetite, are also present, especially in the Fidra basanite.

Peridotite inclusions occur at Weaklaw and Fidra (Figure 1.8, Chapter 1). Those in the Weaklaw Vent are foliated, highly altered and up to 15 cm in diameter. Spinel is the only primary constituent in these that has not been altered by low-temperature processes. In contrast, the Fidra basanite contains abundant xenoliths of fresh spinel lherzolite, commonly foliated and with orthopyroxene porphyroclasts. Xenoliths of wehrlite grading to clinopyroxenite are also common here. The Fidra basanite is also remarkable for its content of large (up to 3 cm) discrete anhedral crystals (megacrysts), principally of anorthoclase but also of sanidine and magnetite. Rare xenoliths of related apatite-magnetite rock also occur. Peridotite xenoliths have not been found along the coast east of Fidra but xenoliths of biotite-rich ultramafic rock are known from Partan Craig and Beggar's Cap.

Interpretation

The chief distinction between the mapped vents and the surrounding tuffites and volcaniclastic sedimentary rocks is that the latter represent widely distributed (distal facies) fragmental material on and around the volcanic cones, which were subject to water-sorting and admixture with fluviatile detritus, whereas the vent material represents the generally coarser and more chaotic (proximal facies) material collecting as fall-out and talus on the steeper inward-facing slopes. Listric faulting and mass-flow (lahar) processes would have contributed to the complexity. The extreme alteration (dolomitization) of the pyroclastic rocks at Weaklaw may have resulted from extended exhalation of carbonated fluids from the post-eruptive vent.

In Early Carboniferous time the region lay within equatorial latitudes, with the evidence indicating a low-relief lagoonal landscape lying little above sea level, and supplied by fine clastic sediment from slow-flowing rivers. The vents probably represent shallowly dissected tuff-rings created by phreatomagmatic activity where rising magmas encountered wet sediment or standing water. The diameter of these tuff-rings, which would have been little wider than the vents, rarely exceeded 1 km and was commonly significantly less. Eruptions would have been short-lived and violent, yielding pyroclastic products composed mainly of broken fragments of near-surface rocks together with juvenile tephra, including at times lava bombs of substantial size. The identification of lacustrine sediment within some vents suggests the probability that lakes ('maars') formed within the craters. The presence of plant fossils suggests that the emergent slopes were colonized by ferns, equisitales and club-mosses, fragments of which are common in the tuffites and which are well preserved in places, such as at Oxroad Bay and Weaklaw. Dolostone lenses (free of organic remains) within the green tuffs have been regarded as evaporites produced under more arid conditions. Martin (1955) divided the vents into a younger Green Group and an older Red Group, showing that the Red Group vents were active before the eruption of the lavas whereas those of the Green Group cut higher strata and contain fragments of the hawaiite and mugearite lavas. The Quarrel Sands Vent is representative of the older Red Group whereas the Yellow Craig Plantation, Partan Craig, Yellow Man and Horseshoe vents are among the younger Green Group.

The early magmas, represented by tuff fragments and 'bombs', all appear to have been strongly silica-undersaturated basaltic varieties. The compositions of many were primitive, with olivine as the sole phenocryst species. Ascent rates of these magmas were frequently high and uninterrupted, as is evidenced by the common occurrence of ultramafic xenoliths of probable mantle origin. However, some of the magmas contained both olivine and augite phenocrysts, as large and conspicuous crystals, indicative of stagnation and slower cooling (probably at deep crustal levels) during their ascent. It may be significant that the lowest lava in the succession, the analcime trachybasalt at North Berwick, is also silica-undersaturated. This, and a similar lava at the base of the lava succession farther south, are almost unique in the Dinantian lava successions of the Midland Valley in being relatively evolved silica-undersaturated rocks. They may be related to the nearby phonolitic intrusions of North Berwick Law, the Bass Rock and Traprain Law (for discussion, see **Traprain Law** GCR site report).

The overlying lavas at North Berwick and west to Marine Villa signify a situation different from that which produced the foregoing phreatomagmatic materials. They indicate larger volume eruptions under subaerial conditions and their less silica-undersaturated compositions point to greater degrees of melting at shallower mantle depths. The compositional variation (olivine basalt, hawaiite, mugearite and subsequent trachytes) implies extended crystal fractionation in magma chambers, probably in the deep crust. Since the greatest thickness of volcanic rocks in East Lothian is around Haddington (McAdam and Tulloch, 1985), it was probably in this area, some 20 km distant, that the principal volcanoes were sited. The thick trachyte lavas of the **Garleton Hills** (see GCR site report) may well be caldera-filling flows, and the thick trachytic tuff sequence at Marine Villa and Weaklaw could well signify a major pyroclastic eruption from a high-level magma chamber prior to caldera collapse. The disconformity above the trachytes heralded long-term volcanic quiescence in the region.

Basanite sills such as those at The Leithies and Fidra mark a later rejuvenation of activity. At Brigs of Fidra a small basanite sheet, believed to be correlative with that on Fidra Island, cuts hawaiite lava and may be assumed to post-date the trachyte episode. A recent K-Ar date on the Fidra basanite gave its age as 264 ± 10 Ma (Downes *et al.*, 2001). Other recent K-Ar dates include a dyke in the Gin Head Vent at 267 ± 5 Ma and the Yellow Man Dyke at 293 ± 7 Ma (Wallis, 1989). These all suggest early to mid-Permian ages. The younger 'Green Group' of vents (Martin, 1955) may also post-date the early phase of volcanism. These younger magmas may have been produced by depressurization melting of garnet lherzolite mantle sources in response to Variscan earth movements.

The xenoliths and megacrysts provide information concerning the rocks beneath the cover of sedimentary and volcanic rocks. Together with suites of inclusions from elsewhere in Scotland, they enable a detailed interpretation to be made of the composition, structure and history of the lower crust and upper mantle, which probably has not changed significantly from Late Palaeozoic times to the present day. A full account of this interpretation is given in Chapter 1.

Much work remains to be done and the physical volcanology of the shallowly dissected vents in particular awaits a modern study. Although many of the igneous rocks have been seriously affected by low-temperature alteration, they nonetheless offer wide scope for further petrological investigation. Additionally, many of the rocks, in particular the fresher basanites, could now be more satisfactorily dated by Ar-Ar techniques. Deep crustal xenoliths containing zircon invite more precise dating using the U-Pb method. Lastly, the petrology of the xenoliths within the volcanic rocks is being actively studied to learn more of the nature of the rocks at depth below this critical area, close to the junction of the Midland Valley and Southern Upland terranes.

Conclusions

The North Berwick Coast GCR site, covering the coast to the east and west of North Berwick is representative of the Visean Garleton Hills Volcanic Formation, but shows particularly the fragmental and intrusive products of numerous small basaltic volcanoes that characterize the earliest local volcanic activity, before the eruption of lavas and tuffs from larger volcanoes sited in the region of the **Garleton Hills** GCR site. Radiometric dates on later intrusions suggest that small-scale magmatism resumed in the Early Permian, possibly more than 50 million years later.

The rocks at this site provide an insight into an ancient tropical environment of sluggish rivers, lakes and lagoons subjected to short-lived, violent eruptions from small volcanoes that stood above an otherwise flat landscape. The wide variety of early land-plants that flourished on this landscape confers international palaeobotanical importance to the site.

The dissected volcanoes can be interpreted as small 'tuff-rings' of fragmental ejecta, formed as a result of the highly explosive interaction of magma with surface water, with groundwater and with water-saturated sediments. As such they complement the similar, but later volcanoes of the **East Fife Coast** GCR site and are possibly some of the best-preserved examples in Britain. As the chances of survival of such fragile structures are rare in the older geological record, there is potential for further study that could increase their international importance.

The intrusions and volcanic rocks of this section are renowned for the abundance and wide variety of exotic rock fragments that have been brought up from great depths by the magmas. These constitute a unique method of sampling the deeper levels of the Earth's crust and provide information of international value on the nature of the lower crust and the underlying upper mantle.

GARLETON HILLS, EAST LOTHIAN (NT 449 764–NT 520 763)

I.T. Williamson

Introduction

The upper part of the Garleton Hills Volcanic Formation in East Lothian, represented by the Garleton Hills GCR site, is the erosional remnant of a lava field, built up of evolved trachytic flows and associated pyroclastic rocks (Figure 2.10). The lower part of the formation, dominated by basaltic rocks, is represented by the **North Berwick Coast** GCR site.

Early Carboniferous volcanism in the Midland Valley of Scotland was dominated by the construction of basaltic and hawaiitic lava fields. However, mugearite, benmoreite and trachyte flows interbedded with the more basic rock-types also feature in many areas (see for example **Machrihanish Coast and South Kintyre**, **Campsie Fells**, and **Touch, Fintry and Gargunnock Hills** GCR site reports). Discrete trachyte-rhyolite centres were also formed in some areas, most notably the Misty Law Trachytic Centre in the Renfrewshire Hills (Johnstone, 1965; Paterson *et al.*, 1990). Studies of these more evolved members of the suite, such as those exposed in the Garleton Hills GCR site, are therefore important in modelling magma genesis and volcanic processes on a province-wide scale.

The Garleton Hills Volcanic Formation is the basal unit of the Visean Strathclyde Group; it is conformable with both the underlying Ballagan Formation (Inverclyde Group) and the overlying Gullane Formation. The underlying and overlying sedimentary rocks have yielded good biostratigraphical evidence of age. In the Spilmersford borehole (NT 4570 6902), situated 7.25 km south-west of the Garleton Hills, sedimentary rocks from beneath the volcanic rocks have yielded (early Visean) PU zone miospores

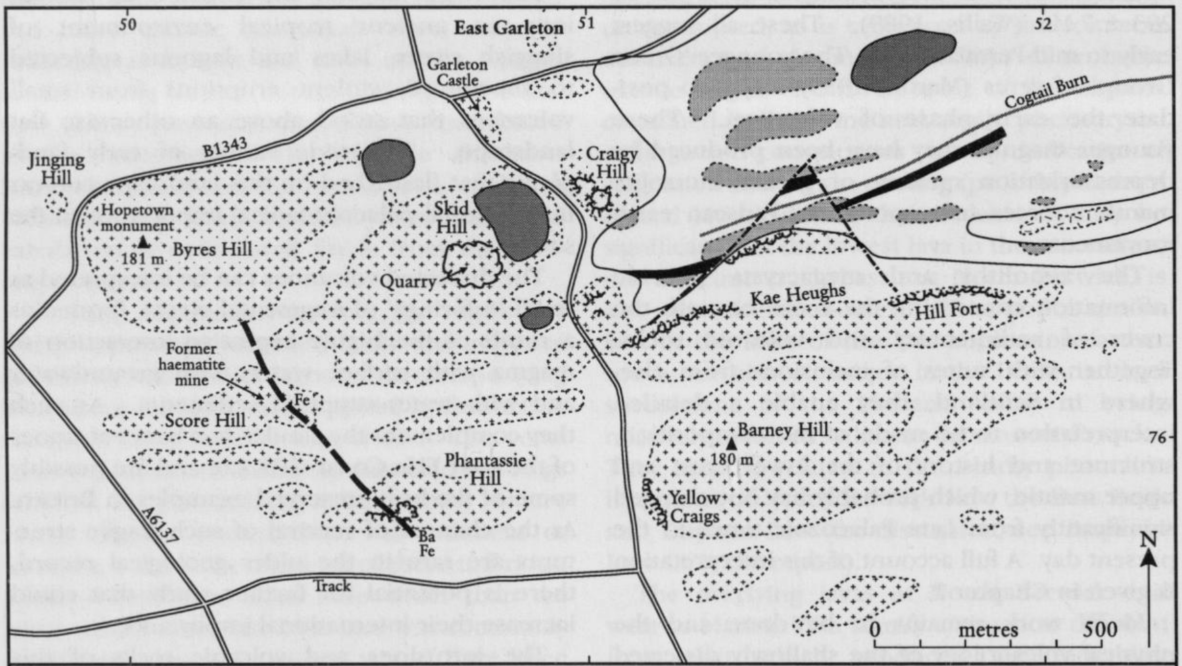


Figure 2.10 Map of the area around the Garleton Hills GCR site. Based on Geological Survey 1:10 560 mapping by M.F. Howells (1961) and A.D. McAdam (1964).

(Davies, 1974). However, Scott (1990) and Bateman and Scott (1990) reported late-Tournaisian (Courceyan–CM zone) plant assemblages from volcaniclastic rocks within the North Berwick Member at the base of the Garleton Hills Volcanic Formation. Sedimentary rocks above the volcanic succession have yielded Asbian–TC zone miospores (Neves *et al.*, 1973) and hence the volcanism can be fairly confidently assigned to early Dinantian time. Radiometric ages obtained by K-Ar whole-rock determinations from lavas at Skid Hill (349 ± 7 Ma) and Phantassie Hill (342 ± 5 Ma) are compatible with an early Dinantian age (De Souza, 1974). However these dates recalculate to *c.* 355 Ma and 348 Ma using new constants, suggesting an early Tournaisian age which is incompatible with the biostratigraphical evidence. A more precise Ar-Ar date of 342.1 ± 1.3 Ma on sanidine separated from a trachyte at Phantassie Hill is, however, compatible with the biostratigraphy (A.A. Monaghan and M.S. Pringle, pers. comm., 2002).

Early descriptions of these rocks are by Howell *et al.* (1866), Geikie (1880, 1897), Hatch (1892) and Clough *et al.* (1910). The geology of the area has been described more recently by McAdam and Tulloch (1985); accounts of the successions in the Spilmersford and East Linton boreholes are by Davies (1974) and Davies *et al.* (1986). Petrographical aspects of the rocks were described by McAdam (1974). Localities within the GCR site are used frequently for educational purposes and are included in excursion guides for the district (McAdam *et al.* in Upton, 1969; Upton and Macdonald in McAdam and Clarkson, 1986).

Description

The Garleton Hills form an area of low hills and escarpments up to 180 m above sea level, about 2.5 km north of Haddington, and in the south-west part of the outcrop of the Garleton Hills Volcanic Formation. The geomorphology is controlled strongly by the effects of both the strike of the rocks and glacial erosion (Figure 2.11). Former glacial drainage channels are a feature of the site with inferred water flow from west to east, parallel to the direction of ice movement.

Figure 2.11 Trap featuring in trachyte lavas, dipping to the right (south), modified by ice action and glacial drainage, at Kae Heughs, Garleton Hills. (Photo: British Geological Survey, No. D3262, reproduced with the permission of the Director, British Geological Survey, © NERC.)

The stratigraphy of the Garleton Hills Volcanic Formation is shown below.

	Thickness (m)
Bangley Member	
Trachyte, quartz-trachyte and augite-phyric quartz-bearing trachyandesite (formerly 'quartz-banakite') lavas, trachytic tuffs	0–160
Hailes Member	
Feldspar-phyric basalts ('Markle' type) and mugearites	25–70
East Linton Member	
Mostly plagioclase-olivine-clinopyroxene-phyric basalts ('Dunsapie' type) and olivine-clinopyroxene-phyric basalts ('Craiglockhart' type), mugearites and analcime-bearing hornblende-phyric trachybasalts (formerly 'kulaites')	10–90
North Berwick Member	
Red basaltic tuffs and agglomerates, green basaltic tuffs and agglomerates, beds of freshwater limestone and dolostone	50–150

The Garleton Hills Volcanic Formation is divided into four laterally persistent members (see above). The Garleton Hills GCR site is situated mainly within the outcrop of the Bangley Member, but some of the upper units of the Hailes Member are also present. A few volcanic necks and minor basic intrusions cut the lava sequence. The thickness of individual lavas is difficult to assess because interflow junctions are not easily identified and any interbedded volcaniclastic rocks are not exposed. In addition, most of the flows have much the same characteristics, making correlation problematical even within this relatively small area. However, many of these lavas are probably more than 20 m thick (Upton, 1982). It is not known if, as seems quite likely, some of the units are shallow intrusions into the lava pile.

The oldest lavas within the area crop out in the relatively poorly exposed ground to the north-east of Kae Heughs (Figure 2.10). Scattered exposures there are of plagioclase ± olivine-phyric basalt belonging to the Hailes Member. Near the top of this unit is a thin unit of mugearite. The basalts are intruded by an ENE-trending quartz-dolerite dyke, interpreted by McAdam and Tulloch (1985) as part of the regionally persistent Prestonpans–Seton Dyke of the Stephanian Midland Valley tholeiitic swarm (see Chapter 6). The dyke is cut by later NW- and NE-trending faults (Figure 2.10).

At Craigy Hill (NT 511 765) the lowest trachyte lava of the Bangley Member seen in the GCR site rests directly upon a basaltic flow of the Hailes Member. Successive trachyte lavas are well exposed on the elongate ridges, which display well-developed stepped or 'trap' features on their northern faces. A particularly fine example of this is exhibited by Kae Heughs (NT 512 762) (Figure 2.11), an E–W-trending ridge composed of two trachyte lavas that are less porphyritic than those higher in the sequence.

The disused quarry cut into the southern flank of Skid Hill (NT 508 763) appears to have been excavated through a single, 20 m-thick lava of massive, but well-jointed, quartz-trachyte. This unit is plagioclase-alkali feldspar-augite-apatite-phyric. A trachyte lava above that at Skid Hill forms Byres Hill (NT 500 764) (180.7 m) and has probably been used in the construction of the Hopetoun Monument, a prominent landmark on the summit. The escarpment at Phantassie Hill (NT 507 758) is composed of at least one thick flow of plagioclase-alkali feldspar-clinopyroxene-apatite-phyric trachyte. Internal structures include patchily developed vesicular facies and zones of alteration and reddening. Yellow Craigs (NT 5115 7585) and Barney Hill (NT 514 760) (179.8 m) expose similar rocks. All of these lavas have developed prominent dip-slopes to the south. Further trachytic lavas and a unit of trachytic tuff in the upper part of the formation crop out to the west of the GCR site. The highest lava is exposed in Bangley Quarry (NT 487 752), which is a GCR site in the Mineralogy of Scotland GCR Block. There, a lava of quartz-trachyte (formerly 'quartz-banakite') is cut by a dyke of trachybasalt that contains phenocrysts or xenocrysts of clear sanidine up to 5 cm long (Day, 1930e).

Near Skid Hill, three small areas (NT 5055 7660, NT 5080 7645, NT 5085 7625), up to 100 m by 200 m, of pyroclastic breccia and tuff, may represent volcanic necks cutting the trachytic lavas.

The volcanic rocks are extracted for road metal and were formerly used for building stone. Haematite and baryte veins that cut the lavas were once exploited commercially and traces of the old haematite workings can be seen north-west of Phantassie Hill, where working ceased in 1876 (Macgregor *et al*., 1920; McAdam and Tulloch, 1985).

Published analyses of rocks from the Garleton Hills Volcanic Formation in general and their associated intrusions are few, the most recent, including stable isotope data, being those of

Smedley (1986a,b, 1988a). Analyses of trachytic rocks are included only by Livingstone and McKissock (1974), Macdonald (1975) and Smedley (1986a).

Interpretation

Max (1976) and Floyd (1994) have both noted that the Garleton Hills lava field lies over the sub-surface extension of the main Southern Upland Fault, and Upton (1982) suggested that this zone of weakness may have acted as a focus for the development of magma chambers large enough to evolve felsic magmas. Magmas of trachytic composition are considerably more viscous and volatile-rich than those of basaltic and hawaiitic composition. Consequently, the trachytes of the Garleton Hills GCR site are likely to have been erupted as viscous lavas of limited aerial extent, and some may have been emplaced as lava domes. Such eruptions are commonly associated with pyroclastic ash-flow and ash-fall deposits. Thin units of bedded tuff and welded tuff, and also some of volcaniclastic sedimentary rocks are present in the Garleton Hills Volcanic Formation, though none are seen in the Garleton Hills GCR site. The presence of the sedimentary rocks shows that volcanic activity was intermittent, and that during the quiescent intervals plant and animal communities were established (Bateman and Scott, 1990; Scott, 1990). Upton (1994) has suggested that the Holocene cinder cones and domes in the Massif Central of France are good analogues of both the basaltic and the trachytic volcanism in the Garleton Hills Volcanic Formation (see also **North Berwick Coast** GCR site report).

Most geochemical studies of Dinantian volcanic rocks of the Midland Valley of Scotland have concentrated upon the basaltic to hawaiitic members (Macdonald, 1975; Macdonald *et al.*, 1977; MacDonald and Whyte, 1981; Smedley, 1986a). This reflects not only their dominance in almost all sequences across the Carboniferous–Permian Igneous Province of northern Britain, but also that the more basic types are of most use in determining the composition and melting characteristics of the underlying mantle. However, understanding the evolution of the more evolved rocks from these suites, such as those seen in the Garleton Hills GCR site, is critical to our understanding of magmatic processes in the upper crust.

The Garleton Hills trachytes, like other more evolved Dinantian lavas of the Midland Valley, are regarded as the intermediate differentiation products of mildly alkaline and transitional olivine basalt magmas that underwent fractional crystallization in relatively high-level magma chambers (e.g. Macdonald, 1975; MacDonald and Whyte, 1981; Smedley, 1986a). Crustal contamination does not appear to have had a major influence, even in the evolved rocks (Smedley, 1986a). Intrusive rocks of even more evolved composition are represented nearby as the phonolitic trachytes of the Bass Rock and North Berwick Law, and the phonolite of Traprain Law. However, these highly evolved rocks are silica-undersaturated, in contrast to the silica-oversaturated trachytes and quartz trachytes of the extrusive sequence. They have been correlated traditionally with the trachytic rocks of the Garleton Hills, but their only likely extrusive associates are the flows of analcime-bearing hornblende trachybasalt that occur locally at the base of the lava sequence. The significance of this possible association is discussed in the **Traprain Law** GCR site report.

Conclusions

The volcanic rocks exposed in the Garleton Hills GCR site comprise the upper part of the Visean Garleton Hills Volcanic Formation, a sequence of trachyte lavas and minor pyroclastic beds overlying the mainly basaltic volcanic rocks that comprise the lower parts of the formation (see **North Berwick Coast** GCR site report). Though trachytic lavas are known from other Dinantian lava fields of the Midland Valley of Scotland, the Garleton Hills GCR site has been selected to represent this important group of geochemically evolved rocks and their style of volcanism. Individual trachyte lavas are probably more than 20 m thick, and were probably erupted as highly viscous flows or even as steep-sided domes. The extrusive rocks are cut by the remains of a few small volcanic necks that represent the feeders to the volcanoes.

The trachytes probably represent the magma that remained in magma chambers at relatively high crustal levels after the eruption, or crystallization at depth, of basalts (the process known as 'crystal fractionation'). Their abundance in East Lothian suggests that magma chambers of considerable volume were present beneath the lava field for considerable time in order to generate magmas of this composition. Such magma chambers may have been located along the projected continuation at depth of the Southern Upland Fault.

TRAPRAIN LAW, EAST LOTHIAN (NT 582 747)

I.T. Williamson and D. Millward

Introduction

The prominent rocky hill of Traprain Law in East Lothian is composed of phonolite, a rare, evolved, silica-undersaturated igneous rock. The phonolite forms a high-level intrusion, emplaced within the Ballagan Formation of the Inverclyde Group, and forming a marked structural dome beneath the Garleton Hills Volcanic Formation (Strathclyde Group; see **Garleton Hills** GCR site report) (Figure 2.12). The phonolite is believed to be associated with the development of the Garleton Hills Volcanic Formation and therefore of Dinantian age. A K-Ar whole-rock determination of 322 ± 3 Ma (*c.* 328 Ma using new constants) (De Souza, 1974) and a Rb-Sr isochron date of 342 ± 4 Ma (De Souza, 1979) appear to support this. The well-

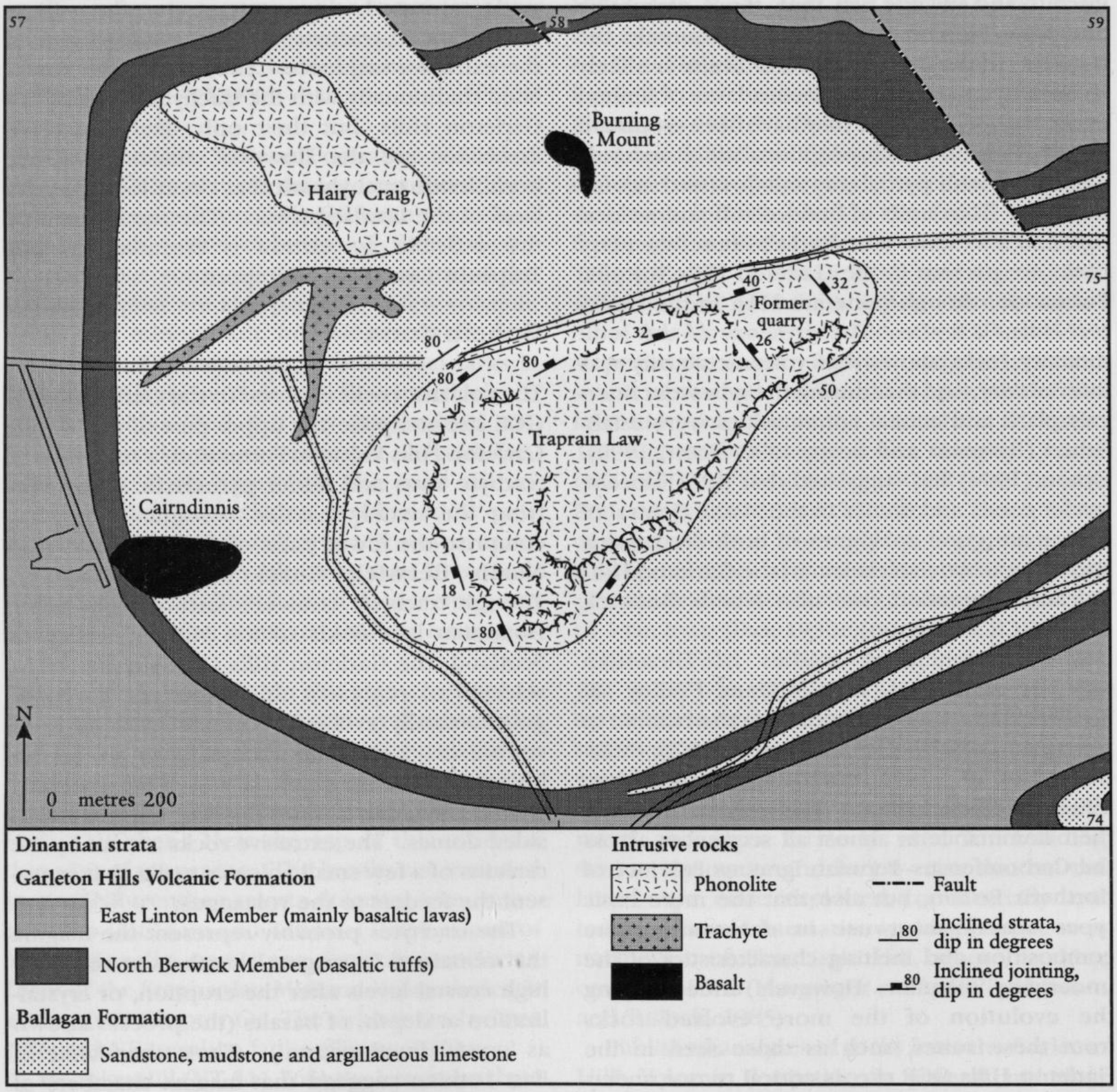


Figure 2.12 Map of the area around the Traprain Law GCR site. Based on Geological Survey 1:10 560 mapping by M.F. Howells (1963–1964) and A.D. McAdam (1964–1967 and 1974–1976).

documented dome-like outcrop is considered to be a particularly fine example of a laccolith, and is one of only a handful of examples of this form of intrusion in the British Isles. Although there are several phonolitic intrusions in East Lothian, rocks of this composition are extremely rare elsewhere in the Carboniferous–Permian Igneous Province of northern Britain and hence the Traprain Law mass provides an additional insight into the magmatic processes involved.

The Traprain Law intrusion was described first by Howell *et al.* (1866) as 'clinkstone', an archaic term for phonolite. Further descriptions are by Geikie (1897) and Bailey (in Clough *et al.*, 1910). Details of the petrography and chemical composition were presented by MacGregor and Ennos (1922), but the most recent descriptions of the intrusion, including details of its structure, petrography and age, are by McAdam and Tulloch (1985). British occurrences of this igneous rock-type are few, and petrographical descriptions of the Traprain Law example are featured in textbooks of igneous petrology such as Hatch's *An Introduction to the Study of Petrology* (1891), Harker's *Petrology for Students* (1895) and Sutherland's *Igneous Rocks of the British Isles* (1982). The considerable educational value of this GCR site is demonstrated by its prominence in field guides to this part of Scotland (McAdam *et al.* in Upton, 1969; Upton and Macdonald in McAdam and Clarkson, 1986). The following description is based on these accounts.

Description

Traprain Law (221 m) rises abruptly from the fertile, undulating farmlands of East Lothian to dominate the surrounding countryside, about 3 km south-west of East Linton (Figure 2.13). The hill has a pear-shaped plan that is partially the result of glacial erosion, and measures about 1 km along its long axis. Its present-day, dome-like outline closely mirrors the original three-dimensional shape of the upper surface of the intrusion. The steeper slopes of the hill are craggy and are covered locally by talus cones. The crags on the south side are a popular rock-climbing venue. The surrounding terrain is formed by rocks of the Ballagan and Garleton Hills Volcanic formations, and a series of escarpments blanketed by till are present. In addition to abundant, glacially sculpted, weathered exposures scattered across the hill, the rock is well exposed in the former roadstone quarry at the north-east end of the hill (NT 5835 7495).

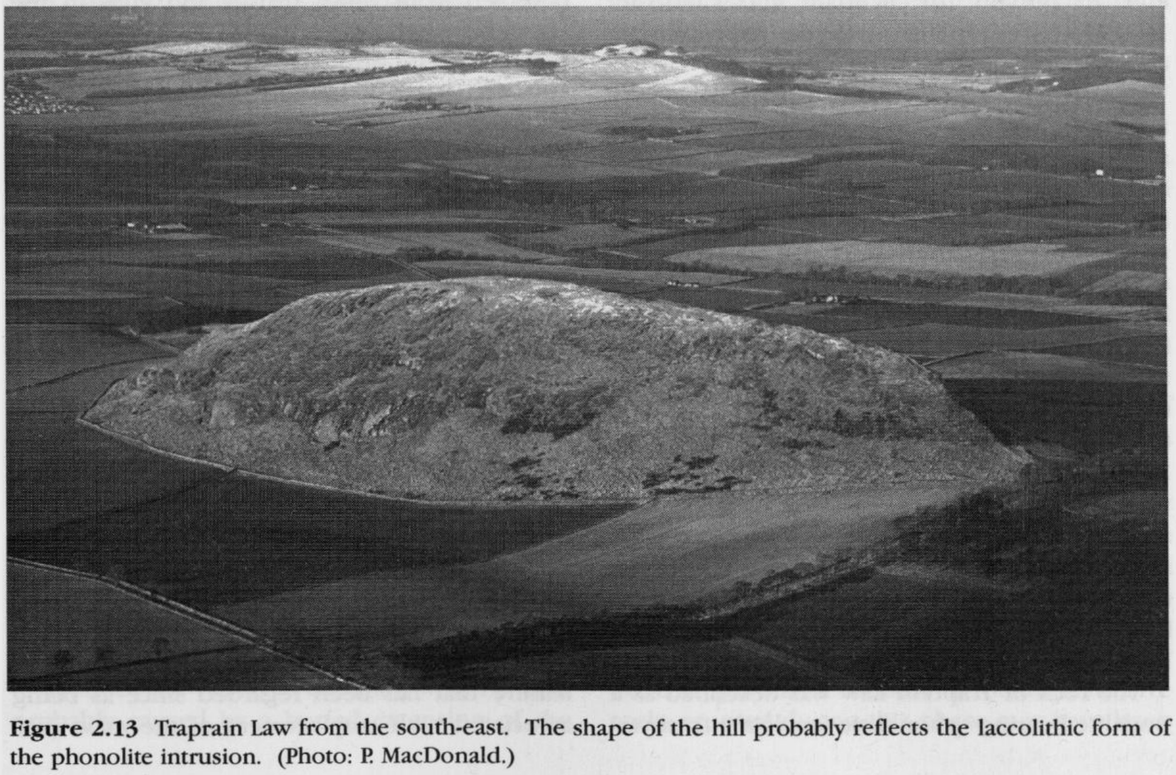


Figure 2.13 Traprain Law from the south-east. The shape of the hill probably reflects the laccolithic form of the phonolite intrusion. (Photo: P. MacDonald.)

Most of the sedimentary rocks covering the phonolite mass have been removed by erosion but, locally, exposures of bedded mudstone, calcareous sandstone and argillaceous limestone are tilted at up to 80° away from the hill (Figure 2.12). On the south-eastern side of the hill, sedimentary rocks are exposed well up the hill-side and are steeply inclined at angles of about 50°. These demonstrate that strata are domed over the intrusion.

The phonolite mass is heterogeneous and shows considerable variation in texture and colour. The outer, glaciated surfaces of the intrusion exposed on the higher parts of the hill, most especially towards the summit, are variably vesicular, and there is an apparent increase in grain size towards the centre of the mass. Colour varies from pale pink and speckled through to darker grey, a feature that is exemplified by flow banding. This is clearly seen in the former quarry where the contact between dark and light rock is generally sharp and complex patterns are displayed. Towards the edge of the intrusion, the banding is more marked and is commonly sharply convoluted. Colour mottling is also noted. Bailey (in Clough *et al.*, 1910) and Tomkeieff (1952) both regarded the variations in colour as due to alteration of the feldspars, and Upton and Macdonald (in McAdam and Clarkson, 1986) attributed this to slight hydrothermal effects.

Jointing is prominent throughout the intrusion. In the former roadstone quarry a set of curvi-linear joints is orientated approximately parallel to the surface outline of the hill. These impart a coarse platy appearance to much of the exposure. The joints become more closely spaced towards the outer parts of the intrusion and are generally sub-parallel to, but locally cut, the flow banding. On the flanks of the hill, platy jointing is more-or-less conformable with bedding in the country rock and dips outwards at moderately high angles (Figure 2.12).

Baked xenoliths of sandstone and fissile mudstone are quite common, and several large masses, up to almost 3 m across, were noted in the quarry by Day (1930d, 1932b). Xenoliths of basic volcanic rock are small and less common (Bennett, 1945; Tomkeieff, 1952).

The rock of Traprain Law was described as a 'sodalite-bearing phonolitic analcime-trachyte' in *Petrology of Igneous Rocks* (Hatch *et al.*, 1961) and as a 'phonolitic trachyte' by Upton in Sabine and Sutherland (1982). However, the former work noted the presence in the rock of about 20% analcime, 4% nepheline and a few crystals of sodalite, which along with the abundant alkali feldspar, indicate that it should be classified as an analcime phonolite after the scheme of Le Maitre (2002). Some parts of the mass are aphyric but others are sparsely porphyritic, containing phenocrysts, up to 5 mm long, of oligoclase and sanidine-cryptoperthite. Also present are opaque oxide pseudomorphs, up to 4 mm long, after amphibole and scattered crystals of augitic pyroxene. The pilotaxitic-textured groundmass is composed of abundant twinned sanidine laths, magnetite, anhedral aegirine-augite, fayalite, which poikilitically encloses alkali feldspar crystals, apatite and the feldspathoids mentioned above. The feldspathoids are wholly interstitial. Alteration of the phonolite is extremely patchy on a millimetre scale, with abrupt gradation from areas in which the alkali feldspar is intensely sericitized to other areas in which it is unaltered. Analyses have been published by MacGregor and Ennos (1922) and Day (1930c).

A number of unusual minerals have been collected from veins, druses and vugs in the phonolite. The most common species found include calcite and alkali feldspar, but Upton and Macdonald (in McAdam and Clarkson, 1986) list analcime, anhydrite, apophyllite, datolite, natrolite, pectolite, prehnite, selenite and stilpnomelane. Batty and Moss (1962) recorded powellite ($CaMoO_4$).

In the same general vicinity as the laccolith, there are a few other intrusions (Figure 2.12). A sheet-like body of phonolite is seen at Hairy Craig (NT 577 751) and small plugs of olivine-dolerite are noted at Cairndinnis (NT 573 745) and Gold Knowe (NT 580 752), all within about one kilometre of Traprain Law.

Interpretation

Early mapping of the area by H.H. Howell of the Geological Survey showed that strata adjacent to the Traprain Law phonolite are steeply inclined away from the hill (Howell *et al.*, 1866), a feature that has been regarded since as being associated with emplacement of the mass (see also Geikie, 1897, fig. 132). The domed outline of the intrusion, together with its contact relationships with the sedimentary rocks, and the arrangement of the flow-banding and jointing, suggested to Bailey (in Clough *et al.*, 1910) that emplacement as a laccolith was likely. This view was re-iterated by McAdam and Tulloch (1985). However, neither Bailey nor McAdam and Tulloch were entirely certain about the interpretation because lower contacts of the mass are not exposed. Nevertheless, there are several pieces of evidence to suggest the possibility of a concordant lower contact supporting the laccolith model. First, Clough *et al.* (1910) recorded vertical columnar jointing at one location on the deeply eroded western side of the mass. From this they inferred an approximately horizontal floor to the phonolite 'at no great distance below the surface'. The second clue is from a temporary excavation made at the entrance to the quarry in 1955 (examined by C.J.S. Stillman and cited by McAdam *et al.* in Upton, 1969). This exposed 12 m of bedded sedimentary rocks, with phonolite both above and below. The contacts were seen to be almost concordant and the bedding was inclined in a similar direction to that of sandstones in the adjacent fields, though at a shallower angle. Stillman concluded that the evidence present in the trench supported the long-held view that the Traprain Law mass has a laccolithic form. The complex patterns exhibited by the flow-banding at the north-eastern end of the outcrop suggest magmatic convection, and McAdam and Tulloch (1985) suggested that a feeder pipe to the laccolith may have been located in this area.

Further evidence for the three-dimensional form of the intrusion comes from ground magnetic profiles run by the Institute of Geological Sciences in 1965, which suggest very steep-sided contacts to the intrusion on the north-west and south-east sides of the hill. A pronounced, elongate aeromagnetic anomaly extends south-westwards from the phonolite outcrop at least as far as Whitelaw Farm (NT 567 720) and this too is compatible with a steep-sided igneous body with high magnetic susceptibility located at a shallow depth. The phonolite has the required susceptibility value to produce the observed anomaly, which is thus probably caused by a buried extension of the exposed intrusion (McAdam and Tulloch, 1985). Whether or not the elongated shape of the intrusion in some way reflects a buried fissure, crudely parallel to deep-seated basement structures such as the Southern Upland Fault, is debatable. However, Upton (1982) has pointed out that the rocks of the Garleton Hills Volcanic Formation lie over the sub-surface extension of the main Southern Upland Fault.

Though the emplacement age of the Traprain Law laccolith, and a large number of other intrusions in East Lothian, cannot be fixed with certainty at present, their geographical association with the Garleton Hills Volcanic Formation suggests that many of them may be sub-volcanic components of this Visean volcanic field. The vesicular nature of the upper surfaces of the laccolith suggests emplacement at a high crustal level and it is possible that this, and other intrusions, may have acted as feeders to the lavas and pyroclastic rocks.

However, highly evolved silica-undersaturated rocks such as the Traprain Law and Hairy Craig phonolites and the nearby intrusive phonolitic trachytes of North Berwick Law and the Bass Rock are not represented within the Garleton Hills Volcanic Formation. Indeed, evolved silica-undersaturated rocks are virtually absent (there is one flow in the Campsie Fells) from all Dinantian volcanic successions of the Midland Valley, which evolve typically along a differentiation trend from alkaline or transitional olivine basalt to silica-oversaturated quartz-trachyte and exceptionally rhyolite. The Garleton Hills succession is unusual however, in that single flows of silica-undersaturated analcime-bearing hornblende trachybasalt (formerly 'kulaite') occur locally at the base of the earliest lava member, including one at Blaikie Heugh, only 2 km from Traprain Law. There are also xenoliths of analcime-hornblende trachybasalt in the Traprain Law phonolite, which both Bennett (1945) and Tomkëieff (1952) regarded as co-genetic with the nearby flow. Therefore, it is possible that the earliest magma chambers to form beneath the East Lothian area did evolve along silica-undersaturated lines, and that the phonolitic intrusions may represent their most extreme products. Almost all of the extrusive rocks erupted during this phase may have been removed by erosion, and the subsequent magmatism evolved entirely along a silica-oversaturated trend as seen in the preserved lava succession. The Traprain Law phonolite is thus potentially of great importance in showing that evolved silica-undersaturated rocks can be associated with a Dinantian basalt–quartz-trachyte series. It is one of only a handful of such examples within the Midland Valley of Scotland.

Conclusions

The Traprain Law intrusion is of national importance both for the form of the intrusion, a laccolith, and for its rock-type, a phonolite, which is rare in Britain. It is representative of a small group of intrusions in East Lothian of silica-undersaturated felsic igneous rocks, apparently associated with the Visean Garleton Hills Volcanic Formation. Other members of the group include the phonolitic trachyte intrusions of North Berwick Law and the Bass Rock. Rocks of these highly evolved compositions are not represented within the Visean extrusive rocks of East Lothian, which range in composition from basalt to quartz-trachyte (i.e. they are mostly silica-oversaturated). The presence of rare phonolitic rocks therefore adds considerable information to the understanding of the origin and evolution of Dinantian volcanic rocks in the Midland Valley of Scotland.

The shape of the upper surface of the intrusion is approximated by the craggy rounded outline of Traprain Law. Locally, around the lower slopes of the hill sedimentary rocks dip steeply away from the hill, indicating that these strata were arched up over the intrusion. The dome-shaped upper surface is mirrored within the intrusion by a characteristic set of curved joints, particularly in the outer part of the mass. The vesicular outer part, formed by gases and vapours escaping from the magma, indicates emplacement at shallow depth. Though the basal contact of the intrusion is not exposed, there are indications that it may be concordant with the sedimentary host rocks. Thus, the inferred form of the intrusion (a domed upper surface and a flat base) is that of a laccolith. Few laccoliths are described from Britain, and Traprain Law is featured in many geological textbooks as a typical example.

ARTHUR'S SEAT VOLCANO, CITY OF EDINBURGH (NT 266 733–NT 283 731, NT 262 742 and NT 251 735)

B.G.J. Upton

Introduction

Within Edinburgh, outcrops of Dinantian volcanic rocks occur at Craiglockhart Hill, Castle Rock, Calton Hill and Holyrood Park. Because each of these forms a high topographical feature, they have had a profound influence on the development of the city from at least Iron Age times to the present day. Edinburgh Castle and the Castle Rock on which it stands are world-famous. Calton Hill lying just north of the eastern continuation of Princes Street has provided sites for, among other things, the old astronomical observatory, Nelson's Monument, the Royal High School and St Andrew's House. Holyrood Palace is sited on a low-lying outcrop of lavas south-east of Calton Hill. The largest of these outcrops, however, is that mainly embraced by Holyrood Park to the south-east of the city centre and containing Edinburgh's highest prominence, Arthur's Seat. As with Castle Rock, the high ground within the park provided sites for human habitation some 6000 years ago.

The above-mentioned localities have major geological, as well as historical and cultural, importance. Castle Rock is the surface expression of a sub-cylindrical stock of basalt, almost certainly representing the infilled conduit of a surface volcano. The Calton Hill and Holyrood outcrops comprise lavas and tuff layers with numerous intrusions and vents occupied by fragmental deposits. The volcanic succession is some 200 m thick on Calton Hill, thickening to between 400 m and 500 m on Whinny Hill on the eastern side of Holyrood Park. These successions form part of the 20°–25° eastward-dipping limb of a synclinal structure occupied by the Midlothian Coalfield, whose eastern limb reveals the 600 m-thick East Lothian volcanic succession (see **North Berwick Coast** and **Garleton Hills** GCR site reports). The extrusive sequence is transected to the south by two major volcanic vents at the Lion's Head and the Lion's Haunch. The smaller Lion's Head Vent is cut by the Lion's Haunch Vent. A basaltic plug in the Lion's Head Vent forms the high point, Arthur's Seat.

Within the western confines of Holyrood Park, the sedimentary sequence underlying the volcanic rocks consists of the Ballagan Formation (sandstones, marls, etc.) of the Inverclyde Group. This sequence contains two sills, the Heriot Mount–St Leonard's Sill and the Dasses Sill, which are regarded as contemporaneous with the extrusive activity. Approximately midway stratigraphically between these lies the much more prominent Salisbury Craigs Sill, which forms one of the most distinctive landmarks around Edinburgh. Although undated, the Salisbury Craigs Sill may be considerably younger than the igneous features mentioned above. Younger still are some (approximately) E–W-trending dykes that were intruded towards the close of the Carboniferous Period.

Following the Late Palaeozoic Earth movements that gave the easterly dip to the strata beneath Edinburgh, the geological record for the next 250 million years has been lost by erosion. The topography bequeathed to modern-day Edinburgh has been sculpted almost entirely by eastward-flowing ice during the past two million years. Among the most obvious results of this glaciation is the classic crag-and-tail topography exhibited by Castle Rock and the ridge of the 'Royal Mile' to the east (see cover photo, this volume), the steep west-facing escarpments of the Salisbury Craigs Sill and Whinny Hill lavas and their dip-slope tail towards Abbey Hill, and the analogous dip and scarp geomorphology of Calton Hill (see Figure 2.17). In each case the steep west-facing escarpments are formed by the igneous rocks, which presented more resistance to the glaciers that differentially eroded the softer sedimentary rocks and fault zones.

The rocks of the Arthur's Seat Volcano have played a major part in the history of geological science. The earliest geological reference appears to have been that of Atkinson in 1619, who mentioned the occurrence of 'Lapis haematite' in Holyrood Park (Clark, 1956). In the late 18th century, James Hutton realized that hot molten rock (magma) had injected sedimentary strata and that Salisbury Craigs represents a sill. Tradition maintains that Hutton used the site at the base of the sill ('Hutton's Section') to demonstrate the intrusive nature of the sill and to refute the widespread belief that 'whinstone' and basalt were marine precipitates as advocated by the popular Neptunist hypothesis. Samples of basaltic rocks from the park were used by Sir James Hall (1805) for the earliest petrological melting experiments. Among the first accounts of the geology of the park are those by Townson (1799), Boué (1820), Maclaren (1834, 1839, 1866) and Howell and Geikie (1861). There was a subsequent burst of scientific interest in the area in the later part of the 19th century which saw publications by Zirkel (1870), Allport (1874), Judd (1875), Bonney (1878), Henderson (1880), Geikie (1880) and Teall (1888). Detailed mapping by the Geological Survey led to the appearance of a revision of Howell and Geikie's 1861 memoir by Peach *et al.* (1910) and the description of the Arthur's Seat Volcano in the memoir was also issued separately in the following year (Peach, 1911). Numerous early 20th century papers include those of Day (1912, 1923, 1933), Campbell (1914), Bailey (1923) and MacGregor (1936), but the definitive accounts that form the basis for most recent descriptions are those of Clark (1956) and Black (1966). Being so accessible, in the centre of a university city, the site must be one of the most popular geological excursion venues in Britain and has featured in all excursion guides to the area (e.g. Cox and Upton in Upton, 1969; Black and Waterston in McAdam and Clarkson, 1986; Land and Cheeney, 2000).

Description

The Arthur's Seat Volcano GCR site includes all the outcrops of Early Carboniferous igneous rocks close to the city centre of Edinburgh, namely Castle Rock, Calton Hill and Holyrood Park; the latter includes Arthur's Seat itself (Figure 2.14). A cross-section of Holyrood Park shows the relationships between many of the individual features of the Arthur's Seat Volcano (Figure 2.15).

Castle Rock (NT 251 735)

The rugged prominence crowned by Edinburgh Castle is composed of a steep-sided basaltic plug, 300 m by 200 m and elongated north-west–south-east in plan (see cover photo, this volume, and Figure 2.16). The plug cuts sandstones of the Ballagan Formation and the contact on the south-eastern side is visible from the road at Johnston Terrace. Castle Rock is an essentially homogeneous, fresh basalt ('Dalmeny' type) containing microphenocrysts of abundant olivine and less abundant augite and plagioclase.

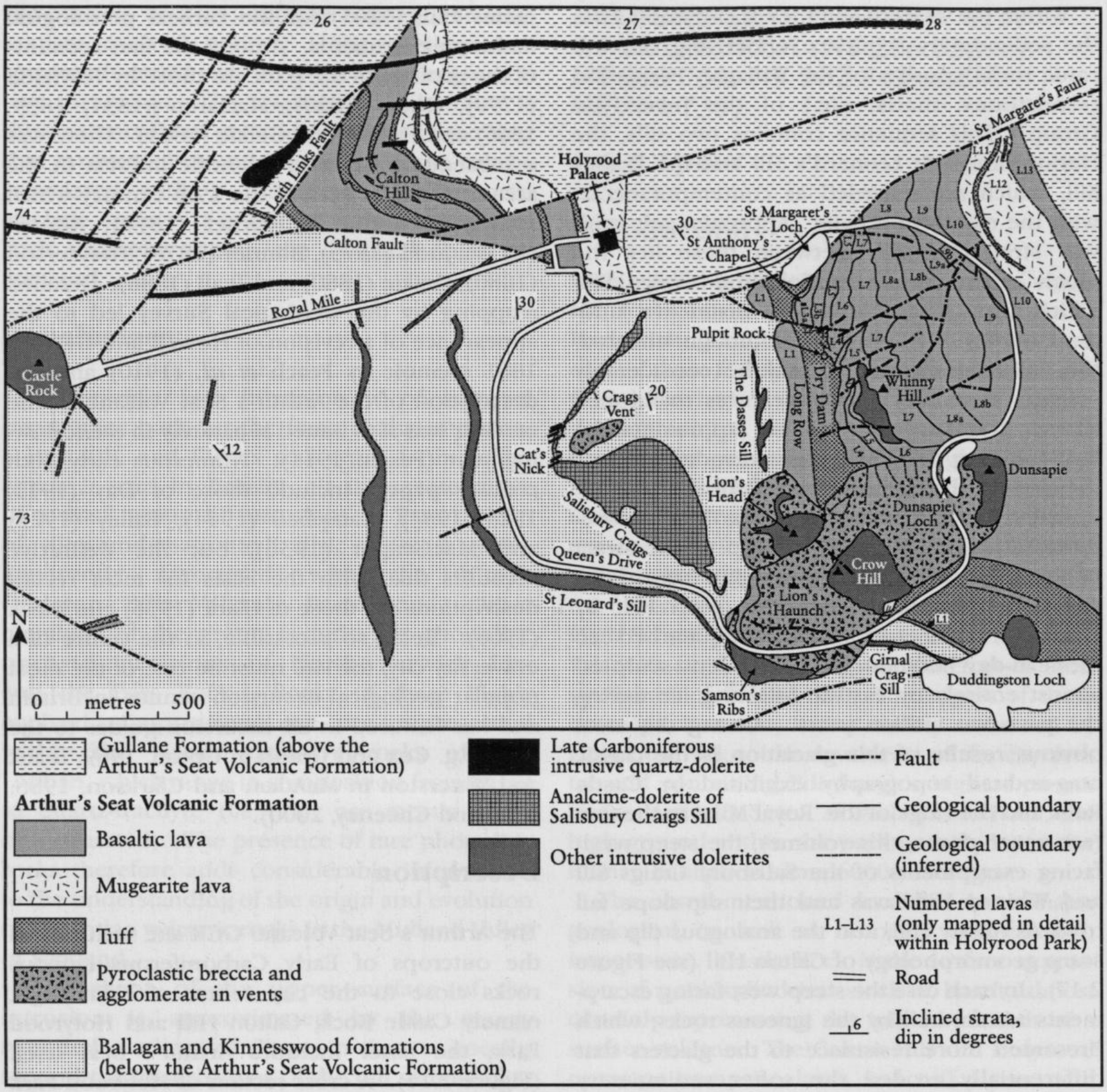


Figure 2.14 Map of the area around the Arthur's Seat Volcano. After Land and Cheeney (2000); and British Geological Survey 1:10 000 Sheet NT 27 SE (2000).

Calton Hill (NT 262 742)

The volcanic succession at Calton Hill is bounded on its north-west side by the NE-trending Leith Links Fault and on its southern and south-eastern margins by the Calton Fault and a WSW-trending fault that passes some 90 m north of Holyrood Palace. The succession comprises a number of lavas with subordinate tuffs, all with the regional eastward dip. As with the other two faulted outcrops of volcanic rocks (a) in Holyrood Park and (b) underlying Holyrood Palace, there is a generalized progression with time from relatively primitive basalts through hawaiites ('Markle' type) to more highly differentiated mugearites. Cessation of volcanism was marked by an erosional disconformity followed by a relative sea-level rise and deposition of the Abbey Hill Shale. Basal tuffs to the west of Calton Hill are succeeded by a basalt flow *c.* 30 m thick, overlain by a tuff several metres thick. Above this are three flows ('Markle' type; probably hawaiites) with intervening thin tuff layers. The uppermost

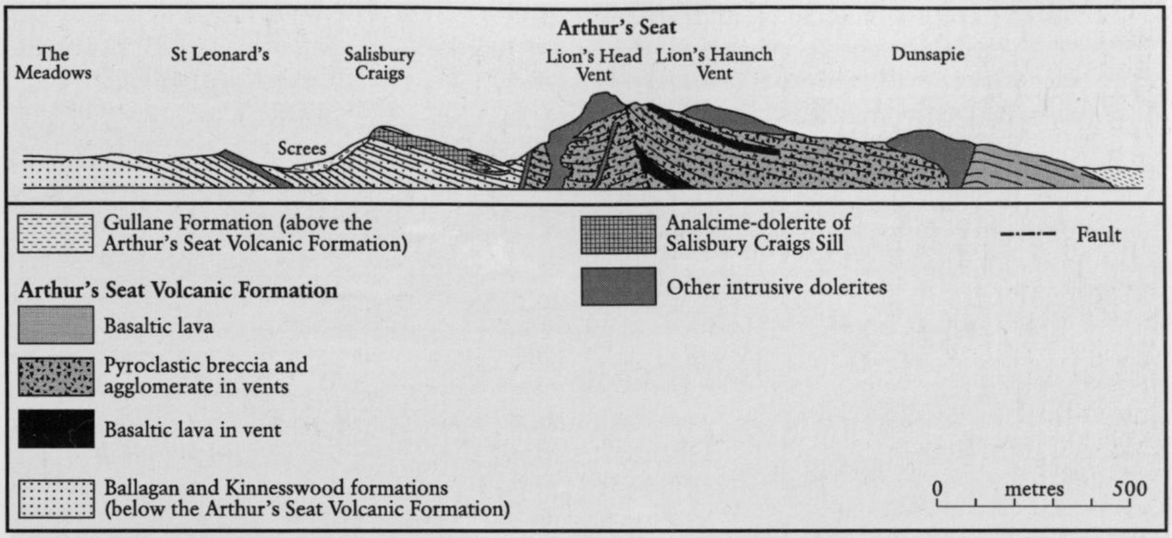


Figure 2.15 Cross-section of the southern part of Holyrood Park, Edinburgh, passing through the Arthur's Seat Volcano. After Mitchell and Mykura (1962).

of these tuffs forms the summit of the hill on which the old City Observatory stands (Figure 2.17). Overlying this are three highly weathered mugearite flows separated by thin tuffs on the north-eastern dip-slope, which are, in turn, overlain by the Abbey Hill Shale. Three E–W-trending Stephanian quartz-dolerite dykes traversing the northern part of the outcrop mark the youngest igneous events in the Calton Hill fault block.

Holyrood Park (NT 266 733–NT 283 731)

At Holyrood Park, as at Calton Hill, volcanism commenced with explosive activity, locally yielding a basal tuff (presently unexposed) immediately beneath the first lava. There have been different opinions concerning the number of lavas in the Whinny Hill succession in the northern area of the park (NT 278 734). Whilst Peach *et al*. (1910) reckoned that there are 19, Clark

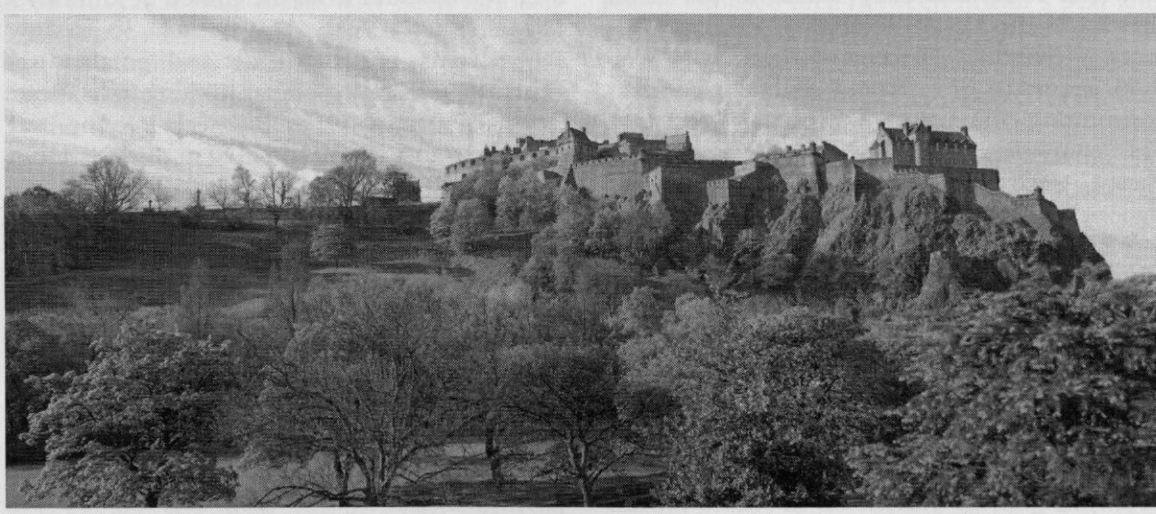


Figure 2.16 Castle Rock from Princes Street Gardens, Edinburgh; a plug of olivine basalt within the Arthur's Seat Volcano GCR site. Note the glacial 'tail' to the left (east), protected by the plug. (Photo: British Geological Survey, No. MNS5624, reproduced with the permission of the Director, British Geological Survey, © NERC.)

Figure 2.17 General view across Calton Hill (old observatory and monument on the summit), towards the Arthur's Seat Volcano and the Salisbury Craigs Sill, Edinburgh. (Photo: P. Macdonald.)

(1956) and Black (1966) concluded that there are only 13. Lava 1, approximately 30 m thick, forms a prominent cliff feature referred to as the Long Row (NT 276 735). It appears to be absent from the Calton Hill succession but crops out again south of the Lion's Haunch Vent in the vicinity of Duddingston Loch (NT 284 725). Lava 1 is compositionally similar to the basalt that forms the Castle Rock plug, although it is more porphyritic. Succeeding Lava 1 is a 30 m-thick sequence of tuffs and sedimentary rocks constituting 'the Lower Ash of the Dry Dam'. This comprises tuffs, fissile mudstones and a cherty limestone about 1 m thick, regarded as a lagoonal evaporite. The volcaniclastic mudstones and limestone contain plant fragments. At the top of this sequence, a tuff layer heralding the resurgence of volcanism is overlain by a highly porphyritic but severely altered basalt flow *c.* 8 m thick. This flow (Lava 2, of 'Craiglockhart'-type basalt) contains macrophenocrysts of pseudomorphed olivine, together with augite and plagioclase. The flow is highly amygdaloidal, with the amygdales occupied by calcite, chlorite, haematite and quartz. The long repose period that followed the Lava 1 eruption and permitted deposition of the Dry Dam mudstones and limestone was not repeated and, from here on upwards, the succession is wholly composed of lavas and pyroclastic rocks. The 'Upper Ash of the Dry Dam', overlying Lava 2, is up to 7 m thick and contains fragments of plant and fish fossils.

Lava 3 consists of an (ankaramitic) alkali basalt ('Craiglockhart' type), with an outcrop confined to a restricted zone extending a little over 200 m south from the ruined St Anthony's Chapel (NT 276 738). This thick (*c.* 30 m) flow shows signs of 'colonnade and entablature' structure; massive columnar jointing in its lower facies gives way above to finer-scale jointing, and the uppermost facies, some 20 m thick, is more blocky with irregular jointing. Whereas the blocks are compositionally similar, they vary in degree of vesicularity, with some blocks evidently having accumulated in a highly scoriaceous condition. Abrupt changes of attitude shown by the jointing in the lower part of the flow, turning from perpendicular to its base to sub-horizontal a few metres above, testifies to a complex cooling history. A small columnar-jointed basaltic plug at Pulpit Rock on the western flank of Whinny Hill may be the source of Lava 3 (Clark, 1956; Black, 1966). Lava 4 (*c.* 8 m thick) is seen to the south of Pulpit Rock, overlying basaltic tuff. It comprises basalt ('Dalmeny' type) with abundant small microphenocrysts of fresh olivine. The flow has well-developed columnar jointing and forms the notable escarpment above the eastern flank of the Dry Dam (NT 277 735).

The higher part of the succession consists wholly of lavas. Lavas 5, 6 and 7 form distinct west-facing escarpments around the top of Whinny Hill. These basaltic to hawaiitic lavas ('Jedburgh' type) have microphenocrysts of olivine and plagioclase. They are overlain by lavas 8, 9 and 10, which are distinctly more porphyritic lavas of hawaiitic composition ('Markle' type). Lavas 11 and 12 are platy-jointed mugearites, overlain by Lava 13, a hawaiite of 'Markle' type, petrographically similar to those beneath Lava 11.

To the south of Arthur's Seat, the volcanic succession is seen on the north side of Duddingston Loch. Lava 1 and the Dry Dam volcaniclastic sedimentary layers can be readily correlated with those of Whinny Hill. Above these is a thick, coarse-grained pyroclastic unit in the vicinity of Duddingston village, within which two thin lavas can be discerned. This unit, which has no counterpart in the northern outcrops, is overlain by a series of hawaiitic and mugearitic lavas approximately correlative with lavas 8 to 13 in the Whinny Hill succession.

The lavas, in particular, have been extensively affected by relatively high-temperature hydrothermal alteration. Olivines have been pseudomorphed by calcite, iron-oxides and/or serpentine minerals. Pyroxenes are commonly chloritized and calcic plagioclases have been variously sericitized, albitized or analcitized. Amygdale and vein infillings include chlorite, haematite, calcite, chalcedony and prehnite.

The Crags (or Western) Vent forms an elongate outcrop (*c.* 200 m by 90 m) on the eastern dip-slope above the Salisbury Craigs Sill, close to the 'Camstone' sandstone quarries. It is filled with basaltic clasts, up to 40 cm across, petrographically identical to the basalts of lavas 1 and 2 (Clark, 1956). The clasts include highly amygdaloidal to scoriaceous types representing juvenile material.

The Lion's Head Vent would, prior to truncation by the Lion's Haunch Vent, have been approximately circular in plan, with a diameter of *c.* 300 m. It is filled with pyroclastic breccia, penetrated by a number of basaltic intrusions. The breccia shows crude bedding, defined by variation in clast size and dipping centrally. Clasts, up to *c.* 6 cm across, mainly of basalts similar to those of the lower lavas, are accompanied by scarcer sandstone clasts. A plexus of 'Dalmeny'-type basaltic dykes in the lower exposures coalesces upwards to form a coherent mass in the centre of the vent, now forming the summit of Arthur's Seat. The latter consists of basalt with fairly well-developed, fine-scale columnar jointing.

The cross-cutting Lion's Haunch Vent (NT 275 729) has an ovoid plan, approximately 1200 m north-east–south-west by 500 m north-west–south-east. The vent includes Dunsapie Hill (NT 282 731) in its north-eastern extremity and the basaltic intrusion of Samson's Ribs in the south-west. A WNW–ESE fault, downthrowing to the north, traverses the south-western part of the vent. The pyroclastic breccia that occupies much of the Lion's Haunch Vent (NT 276 728) is, like that of the Lion's Head Vent, coarsely layered with the bedding dipping towards the vent interior. Much of it is very coarse relative to that of the Lion's Head Vent, with basaltic clasts up to 2 m across (Figure 2.18). Blocks of sandstone, mudstone and limestone are minor components. The breccia differs from that of the Lion's Head Vent in containing fragments of the feldspar-phyric hawaiitic ('Markle' type) lavas. There are several lavas within the vent, whose outcrops are predominantly towards the extreme south-west. Well-bedded sedimentary rocks intercalated between two of these may have been deposited in a crater lake (Black, 1966). A substantial but poorly exposed area between Arthur's Seat and Dunsapie Loch (NT 281 732) is also believed to be underlain by at least three lavas of relatively evolved feldspar-phyric basalt and hawaiite ('Dunsapie' or 'Markle' type).

Three substantial basaltic masses within the Lion's Haunch Vent are regarded as intrusive.

1. The Samson's Ribs mass, intruded along the south-western contact of the vent, shows spectacular columnar jointing. The columns, like those of Pulpit Rock, are curved (Figure 2.19). From the top of the *c.* 30 m-high cliff that forms the north wall of the road west of Duddingston Loch, the SSW-inclined columns steepen as followed down the cliff from *c.* 60° to 75° before turning outwards at much shallower angles to lie almost perpendicular to the rock face. As noted by Black (1966), the lower columns appear to have grown in response to cooling against an

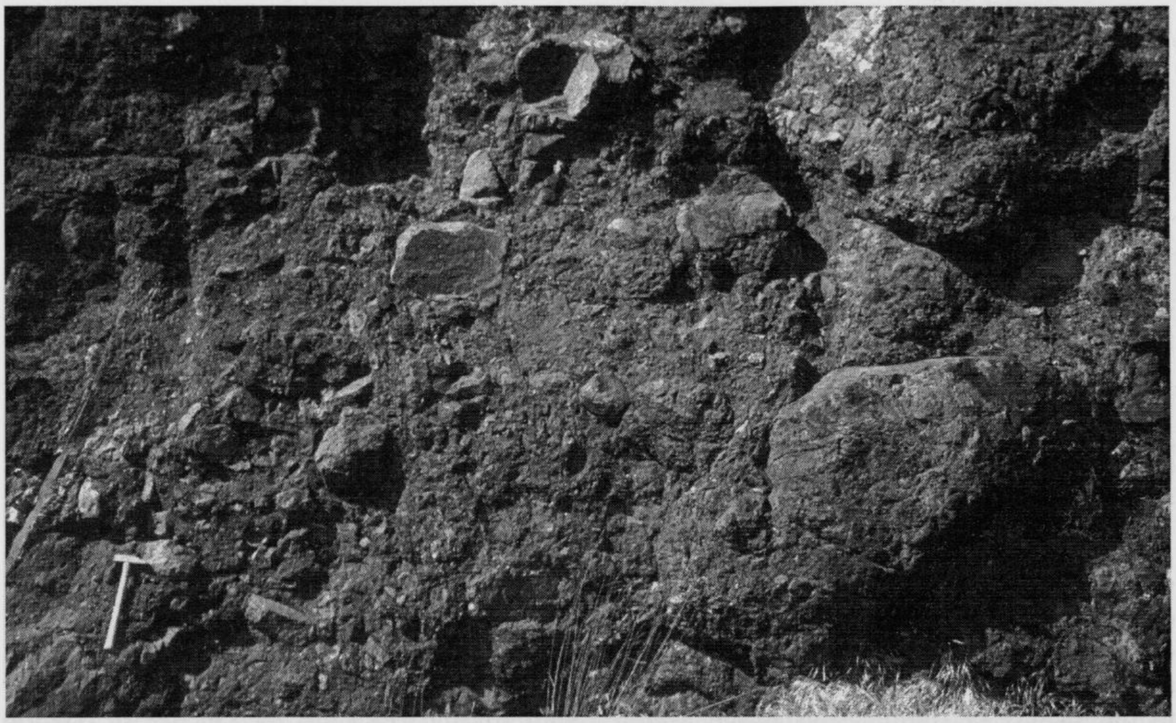


Figure 2.18 Pyroclastic breccias, consisting of blocks of basalt in a matrix of red tuff, Lion's Haunch Vent, Arthur's Seat Volcano GCR site. The hammer shaft is about 35 cm long. (Photo: British Geological Survey, No. D3461, reproduced with the permission of the Director, British Geological Survey, © NERC.)

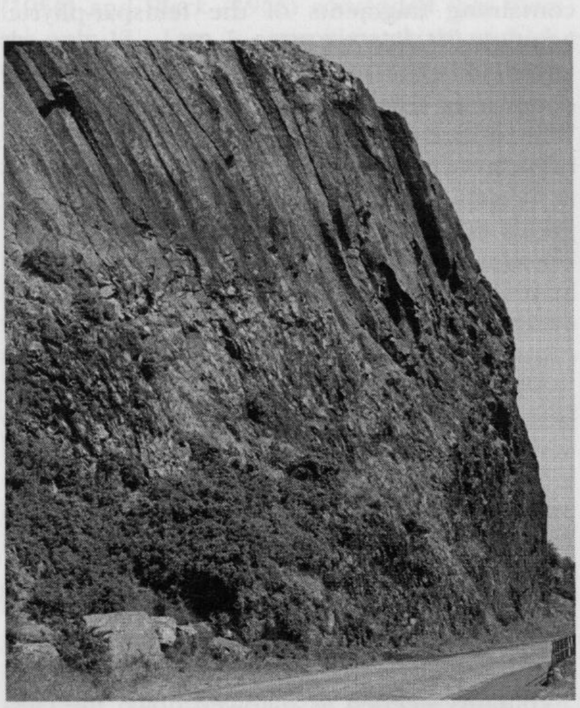


Figure 2.19 Spectacular columnar jointing of basalt in vent intrusion, Samson's Ribs, Arthur's Seat Volcano GCR site. (Photo: British Geological Survey, No. D3465, reproduced with the permission of the Director, British Geological Survey, © NERC.)

almost vertical side-wall whereas the upper portions grew in response to heat loss from a sub-horizontal upper surface.

2. Crow Hill, or the summit area of the Lion's Haunch Vent, consists of a mass of basalt with steep columnar jointing, surrounded by the vent breccia. The relatively fresh basalt ('Dunsapie' type) comprises approximately 30% phenocrysts of olivine, augite, plagioclase and subordinate magnetite in a finer-grained matrix. It is widely used for teaching purposes.
3. The third principal basaltic mass within the Lion's Haunch Vent lies at the north-eastern extremity where it forms Dunsapie Hill (the type locality for 'Dunsapie'-type basalt). It is crudely cylindrical in form and, like the Sampson's Ribs basalt, is thought to have been intruded along the contact zone of the vent, between the pyroclastic breccia and the Whinny Hill lavas.

The Duddingston plug lies outside the Lion's Haunch Vent and cuts the upper Duddingston tuff. It is ovoid in plan (*c.* 250 m across) and consists of olivine-clinopyroxene-felspar-phyric basalt ('Dunsapie' type).

Sills are a prominent feature of Holyrood Park. The lowest is the Heriot Mount–St Leonard's Sill. This is a composite body some 11 m thick, in which a 7 m-thick 'core' of porphyritic basalt ('Dunsapie' type) is surrounded by an envelope *c*. 2 m thick, of an aphyric hawaiite; there is a diffuse boundary, *c*. 85 cm wide, between the two varieties (MacGregor, 1936; Clark, 1956; Boyd, 1974).

The Dasses Sill, considerably higher up the sedimentary succession, is a more complex body, possibly consisting of several lenticular bodies, thickest in the south close to the Lion's Head pyroclastic breccia and pinching out northwards towards the St Margaret's Fault. Here too, a composite character has been shown by Oertel (1952) and Rutledge (1952). The compositional contrast is more extreme than in the St Leonard's Sill, involving a change from basalt to benmoreite (Boyd, 1974). The correlation of a section of sill at Girnal Crag (NT 280 726), farther east between the Lion's Haunch Vent and Duddingston Loch, has been contentious. Mitchell and Mykura (1962) considered the Girnal Crag Sill to be a continuation of the St Leonard's Sill, but Boyd (1974) concluded on petrographical grounds that it correlates with the Dasses Sill.

Another small sill within Holyrood Park that probably accompanied the principal volcanic activity is that on Whinny Hill (NT 278 734). The Whinny Hill intrusion comprises 'a sill-like mass of Craiglockhart-type basalt' (Black, 1966), traceable for some 300 m and lying between lavas 6 and 7. A smaller body of identical basalt, with oval plan and vertical contacts, lying about 50 m to the west, was regarded by Black (1966) as the probable feeder to this sill.

The Salisbury Craigs Sill, attaining a maximum thickness of *c*. 40 m, presents a commanding feature in the park (Figure 2.20). Lying roughly midway between the Dasses and St Leonard's sills, it thins and cuts out south towards the contact of the Lion's Haunch Vent. Over much of its outcrop the sill is generally conformable with the sedimentary rocks into which it is intruded but it steps down northwards through the strata in the vicinity of the St Margaret's Fault. There are, however, some notable disconformities along the well-exposed lower contact,

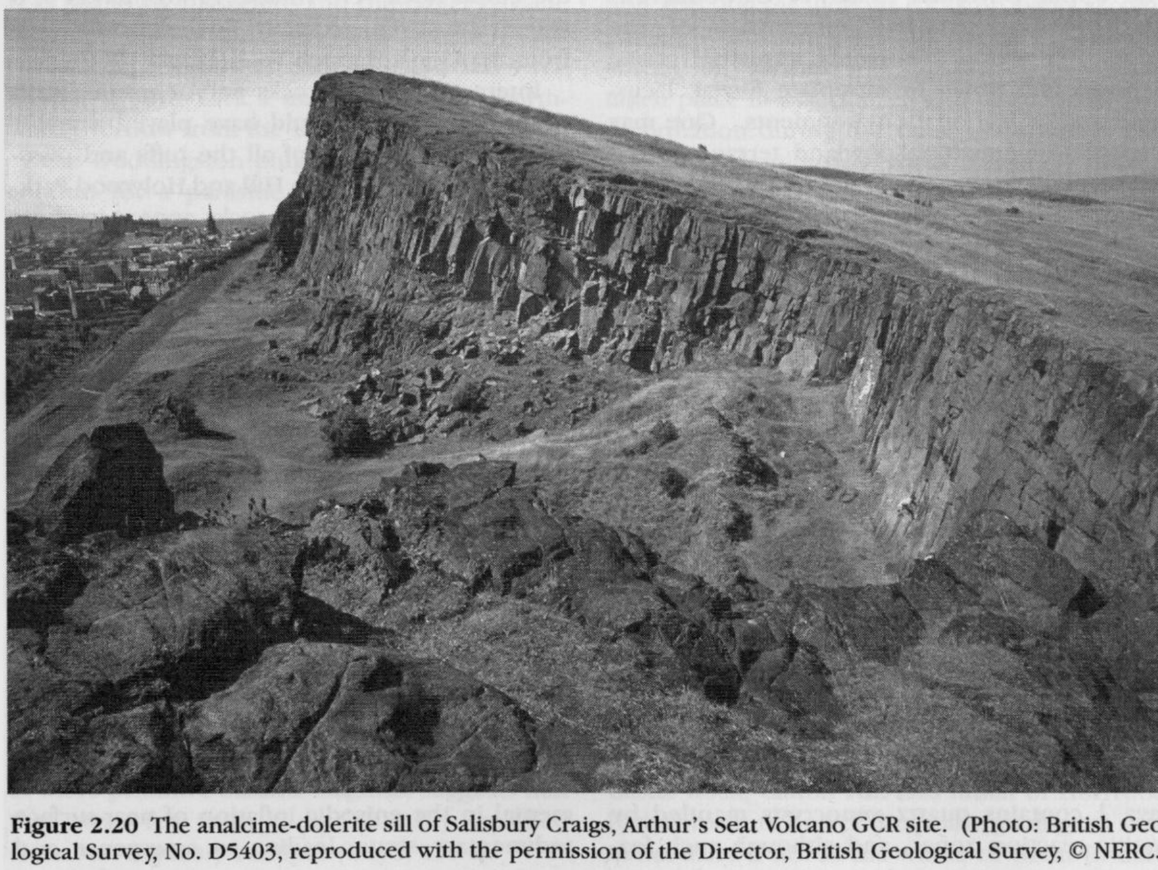


Figure 2.20 The analcime-dolerite sill of Salisbury Craigs, Arthur's Seat Volcano GCR site. (Photo: British Geological Survey, No. D5403, reproduced with the permission of the Director, British Geological Survey, © NERC.)

including the famous 'Hutton's Section', where the magma has prized off a section of the underlying strata in a similar manner to that seen at the **South Queensferry to Hound Point** GCR site (see GCR site report). Towards its southern end, several thin sheets of hornfelsed sediment are intercalated near its upper surface. The contact with overlying sedimentary rocks is visible towards the north. The sill consists of analcime-dolerite (teschenite), which is coarse grained in the interior but fine grained in its marginal facies. Some large-scale layering in the sill, apparent from variations in the colouring, has never been investigated in detail but probably relates to variations in the modal content of olivine. The dolerite comprises plagioclase, olivine, augite, magnetite, apatite and analcime. Thin veinlets of microsyenite represent late differentiates in the southern part of the sill and there are some veins of haematite up to several centimetres wide, including one that has been preserved from quarrying, known as 'Hutton's Rock'.

Interpretation

The Upper Devonian to Lower Carboniferous sandstones, marls and mudstones of the Edinburgh region are mainly terrestrial clastic deposits laid down in intraplate fluvial, lacustrine and/or lagoonal environments. One may envisage an equatorial lowland terrane whose surface lay close to sea level and in which volcanism commenced at around 340 Ma. At the onset of volcanism, rising magmas encountered either standing water or waterlogged sediments at, or close to, surface level and Surtseyan-type phreatomagmatic eruptions resulted.

At near-surface levels the rising magmas would have reached levels of neutral buoyancy (particularly in the low-density Lower Carboniferous sediments) where they spread laterally as sills. Some batches, however, clearly reached surface level and erupted as small basaltic volcanoes. Initial gas release, largely of steam from heated meteoric water, drilled sub-cylindrical conduits that were followed and enlarged by rising magma. Castle Rock probably originated in this manner; a small cinder cone a few hundred metres high and with an external diameter of about 2 km has been eroded away, but the central plug of basalt remains. It has not been dated. Lava 1 contains quartz xenocrysts mantled by augitic reaction zones, and although such features are rare in the other basalts of the region, they are typical of the Castle Rock basalt. This led Black (1966) to conclude that Lava 1 erupted from the Castle Rock volcano and flowed eastwards. If this is correct, it places activity at Castle Rock as the oldest in the north-eastern part of Edinburgh (although it is probably younger than the lavas at Craiglockhart). Black (1966) suggested, on slender evidence, that Lava 2 flowed from a southerly source within the Lion's Head Vent, whereas Lava 3 is inferred to have been supplied through a subsidiary centre to the north at Pulpit Rock (Clark, 1956; Black, 1966). Since the columnar joints of Lava 3 are presumed to have grown normal to the isothermal surfaces as cooling proceeded, the geometry of these surfaces was subject to continuous change, possibly through the action of percolating water. The blocky and scoriaceous nature of the uppermost part suggests proximity to a vent, and a lava fountain may have played above the columnar-jointed Pulpit Rock Vent. Flow 4 is correlative with the basalt plug occupying the core of the Lion's Head Vent and it was inferred that the lava erupted from this vent and flowed northwards until diverted by a lava cone around the Pulpit Rock Vent (Black, 1966). Lavas 5, 6 and 7, however, probably flowed northwards from the Lion's Haunch Vent (Black, 1966).

Interaction of hot rocks and/or magma with near-surface water would have played a major part in the formation of all the tuffs and pyroclastic breccias of Calton Hill and Holyrood Park. Most of the volcaniclastic rocks appear to show signs of some subaqueous reworking, although those of the Crags, Lion's Head and Lion's Haunch vents may have been largely ash-fall tuffs and scree (talus) deposits within the confines of steep-walled craters. The nature of the thick pyroclastic deposits at Duddingston, however, is scarcely known because of lack of exposure. There is a strong asymmetry in the stratigraphy north and south of the Lion's Haunch Vent, with pyroclastic beds to the south taking the place of the dominant lavas in the north. To explain this, Black (1966) surmised that strong northerly winds were responsible for the concentration of ash-fall deposits on the southern side of the main vents. At its maximum development the volcano may have risen to about 1000 m above sea level, with a cone-base of up to 5 km diameter.

Intrusion of sills is likely to have been instrumental in the episodic inflation of near-surface sedimentary strata, leading to emergence and allowing subaerial weathering and plant growth. Since plant fragments are commonly encountered in the volcaniclastic rocks, we may envisage the volcanic hills as having been forested for long periods between the occasional eruptions. The Dasses, Girnal Crag and St Leonard's sills may have been emplaced very early in the volcanic history 'into soft, pliable sediments' (Oertel, 1952). Subsidence and inundation following Lava 1 allowed deposition of the well-bedded ashes with intercalated lagoonal sediments seen in the lower part of the Dry Dam to the north and the lower part of the lower ash at Duddingston. The plant fragments within the Dry Dam sequence are inferred to have been washed down into shallow waters from the adjacent forested volcano flanks. The Dry Dam volcaniclastic mudstone unit thickens south towards the Lion's Head Vent, which may have been growing at the time through explosive action. The Crags Vent, which was probably surmounted by a basaltic cinder cone approaching 1 km diameter, may have developed fairly early, possibly contributing to the ashes of the Dry Dam (Black, 1966). The whole of the Whinny Hill–Lion's Head–Lion's Haunch area is clearly very shallowly dissected, and the larger basaltic outcrops within the two vents probably had surface expressions as confined lava lakes (cf. Oertel, 1952). Lava 4 may have been a northward overflow from the Lion's Head lava lake.

The Duddingston plug was possibly a feeder conduit for a parasitic basaltic volcano developed at a late stage in the volcanism on the south-eastern flanks of the main edifice. The Edinburgh volcanoes may thus have been distributed along a WNW–ESE lineament, *c.* 2.5 km long, exhibiting a very generalized migration of activity over time from Castle Rock in the WNW to the Duddingston plug in the ESE.

The younger products associated with the Lion's Haunch Vent tended to have more highly fractionated hawaiitic and mugearitic compositions, suggesting that magma ascent rates generally decreased with time, allowing time for fractionation to occur. However, analogous compositions did appear earlier if the surmise is correct that the Dasses, St Leonard's and Girnal Crag sills were early intrusions. The observation that these sills are composite, with more highly fractionated magma having been intruded ahead of more primitive basaltic magmas, suggests that they were fed from compositionally stratified chambers (dykes?) at depth (Boyd, 1974).

Basanitic magma originating from comparatively small-fraction melting, at greater mantle depths than the preceding activity, arose to form the Salisbury Craigs Sill. Although the depth at which the sill was intruded is uncertain, the vesicularity near the upper surface makes it unlikely that it was intruded at much more than a kilometre or two beneath a cover of sedimentary and volcanic rocks. Whilst the layering features could be due to in-situ differentiation they more probably reflect differences in the crystal content of successive magma batches as the sill inflated.

Although the Salisbury Craigs dolerite has not been dated, it may be significantly younger than the volcanic rocks. The principal reasons for so thinking are (a) the observation that it thins towards the main Holyrood Park vents and (b) that it is notably more silica-undersaturated than the other rocks (Peach *et al.*, 1910). It is cut (at 'the Cat's Nick') by a thin (*c.* 1 m) E–W-trending quartz-dolerite dyke of the late Stephanian swarm (see Chapter 6). This dyke provides an upper time limit for the Salisbury Craigs Sill, which is probably of Late Carboniferous age.

Whereas some of the secondary mineralization affecting the igneous rocks probably accompanied hydrothermal activity associated with the volcanism, further modification would have taken place in association with deep burial and deformation during the Variscan Orogeny. The Lower Carboniferous rocks revealed at the surface today would formerly have lain at a depth of several kilometres beneath Upper Carboniferous, Permian and possibly younger formations prior to uplift and erosion.

The rocks described above are still not precisely dated. Several of the fresher intrusive bodies could be dated by Ar-Ar methods. Dating of the Salisbury Craigs Sill, while highly desirable, is likely to present difficulties on account of secondary alteration. There is ample scope for further research within the site, such as a proper petrological investigation of the Salisbury Craigs Sill, for which a continuous drill-core would be desirable, and a modern volcanological study of the various volcaniclastic rocks.

Conclusions

The outcrops of igneous rocks forming the Castle Rock, Calton Hill and much of Holyrood Park constitute one of the prime geological sites in Scotland if not in the whole of Great Britain.

The site is representative of (1) early Visean volcanism and (2) the Late Carboniferous suite of alkali dolerite sills in the east of the Midland Valley. The various outcrops of igneous rocks that constitute 'Arthur's Seat Volcano' dominate the landscape of the city and are a vital part of its cultural heritage (the site of the Scottish Parliament is on the edge of the GCR site, as is the architectural World Heritage Site of the 'Old Town'). Splendid examples of many classic volcanic features are easily accessible to the specialist and general public alike, all within the confines of the inner city. These include the lavas and intervening tuffs, penecontemporaneous sedimentary strata, intrusions of various forms (plugs, sills and dykes), as well as volcanic vents infilled with pyroclastic breccia.

The site has great historical significance in the development of geological science and has played a continuing role in the evolution of ideas on volcanic rocks since the days of Hutton and Hall in the late 18th and early 19th centuries. It is undoubtedly of national importance and, from the worldwide interest that it has generated over two centuries, it can be argued that it is also of international importance. In brief, the value of this site, historically, scientifically and scenically, cannot be over-emphasized.

BURNTISLAND TO KINGHORN COAST, FIFE (NT 252 864–NT 280 891)

D.G. Woodhall

Introduction

The Burntisland to Kinghorn Coast GCR site incorporates one of the best-exposed successions of Visean volcanic rocks in the Midland Valley of Scotland. The 485 m-thick succession consists of basaltic lavas with subordinate volcaniclastic rocks (hyaloclastite, pyroclastic rocks and volcaniclastic sedimentary rocks).

Notable aspects of the site are the well-exposed internal structures of individual lava flows, a hyaloclastite unit, and occurrences within some lavas of sedimentary inclusions that locally contain important Early Carboniferous floral assemblages. The latter are described in the *Palaeozoic Palaeobotony of Great Britain* GCR volume (Cleal and Thomas, 1995) from two sites, Kingswood End and Pettycur, which fall within the area of the GCR site described here.

A detailed log of the volcanic succession by Geikie (1900) has formed the basis of most subsequent descriptions (Allan, 1924; MacGregor, 1996). The site was included in a recent re-survey of the Kirkcaldy district by the British Geological Survey (Woodhall, 1998; Browne and Woodhall, 1999, 2000). As a result of this re-survey, the volcanic rocks have been formally designated as the Kinghorn Volcanic Formation, and the exposures at this GCR site constitute the type section. Previously the volcanic rocks were either unnamed or referred to as the 'Burntisland Volcanic Formation' (Francis, 1991). Only a few geochemical analyses of the lavas have been published (Allan, 1924; Macdonald *et al.*, 1977; Smedley, 1986a, 1988a).

Description

The Early Carboniferous age of the succession at the Burntisland to Kinghorn Coast GCR site is constrained, outside of the site area, by Asbian to Brigantian miospore assemblages from fluvio-deltaic, lacustrine and marine sedimentary rocks of the Sandy Craig and Pathhead formations (Strathclyde Group) (Brindley and Spinner, 1987, 1989; Browne *et al.*, 1996). Within the site, these same strata underlie, interdigitate with, and overlie the volcanic succession (Figure 2.21).

Lavas dominate the succession, and range in thickness from 5 m to 30 m. They are typically greyish- or brownish-green-weathered, olivine- or olivine-clinopyroxene-microphyric alkali olivine basalts. Amygdales of dark-greenish-grey chlorite and/or pale-yellow calcite are typically most abundant in the lower and upper parts of individual lava flows. Brecciated flow bases and/or tops are apparent locally, but in many cases have been obscured by weathering. Some of this weathering may have taken place soon after the emplacement of the flow, but it is indistinguishable from Quaternary weathering. Many lavas rest sharply on intercalated volcaniclastic and/or siliciclastic sedimentary rocks, and have a regular basal contact. However, those resting on mudstone tend to have an irregular contact owing to the presence of load structures, and some of these lavas contain inclusions of sedimentary rock that were probably derived from the underlying sediment during the emplacement of the flow. At Pettycur (NT 2608 8625), limestone inclusions have yielded an important Early Carboniferous flora dominated by lycopsids and ferns (Gordon, 1909; Scott *et al.*, 1984, 1986; Rex and Scott, 1987).

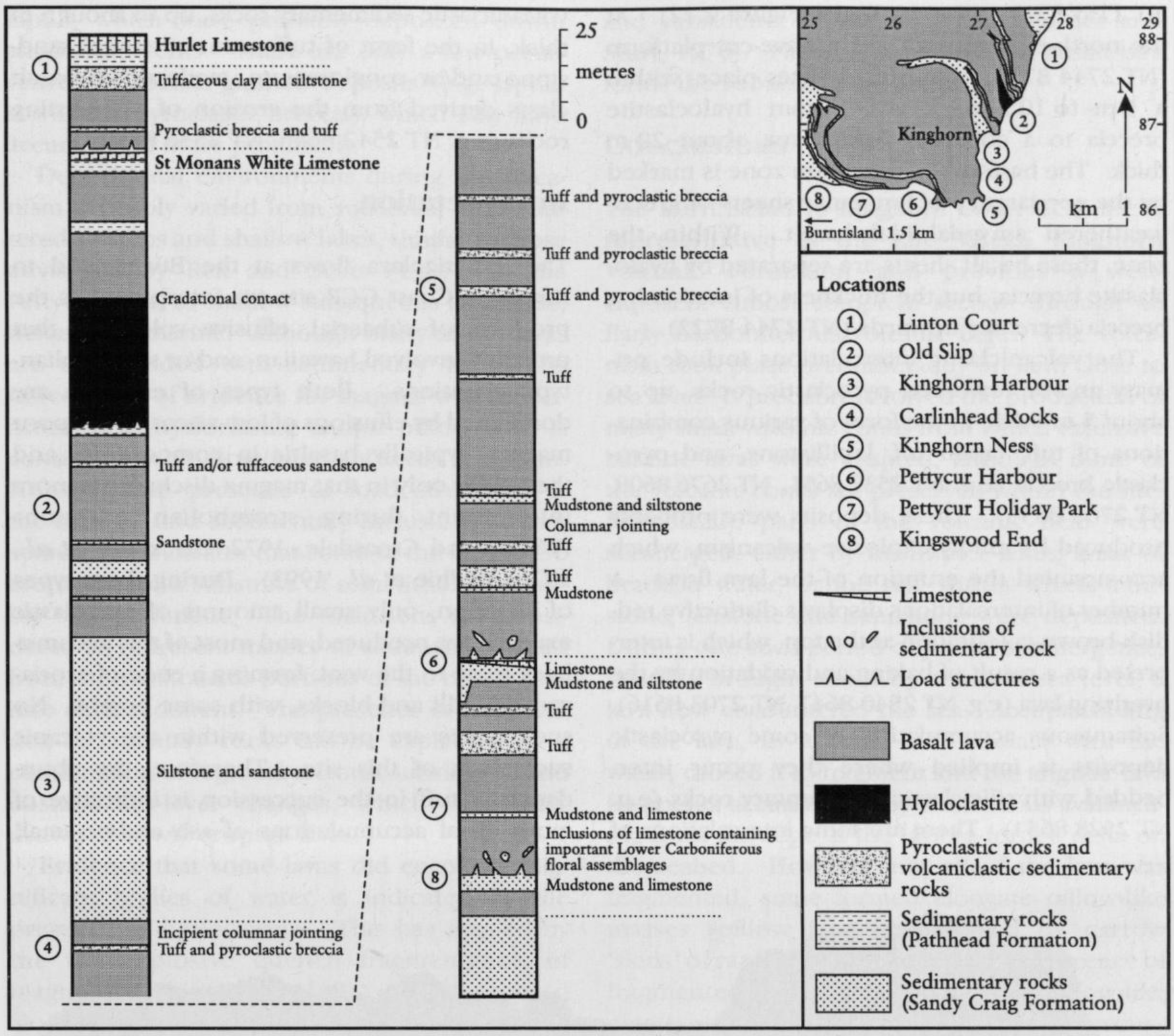


Figure 2.21 The volcanic succession exposed in the Burntisland to Kinghorn Coast GCR site.

The middle parts of the flows are the least weathered, contain the fewest and smallest amygdales, and commonly display cooling joints, which are locally columnar. Columnar jointing occurs in some of the flows exposed in the crags at Kingswood End (Allan, 1924), and on the coast, between Pettycur Harbour and Carlinhead Rocks (e.g. NT 2677 8611). Geikie (1900) described occurrences of pillow lava between Kinghorn Ness and Linton Court (e.g. NT 2751 8728, NT 2753 8740). However, these polyhedral, pillow-like masses lack chilled margins, which would be expected in true subaqueous pillows, and have been produced by weathering along intersecting planar and curvi-planar, horizontal and vertical cooling joints (MacGregor, 1996). In places the joint pattern is enhanced by vein-calcite along joint planes (e.g. NT 2753 8740).

Hyaloclastite, 35–40 m thick in the upper part of the succession, is exposed along the coast north-east of Kinghorn where it forms a 100 m-long wave-cut platform (NT 2734 8717–NT 2744 8722) (Woodhall, 1998). The base of the hyaloclastite is exposed at the southern end of the platform, where it rests on an intercalation of siliciclastic and volcaniclastic sedimentary rocks. The contact is in part irregular due to loading, but in places it truncates bedding in the underlying strata (NT 2734 8717). The hyaloclastite consists of green- to brownish-green-weathered, structureless breccia made up of angular clasts up to 30 cm across, many of which are pillow fragments. It also contains numerous basaltic pillows and pillow fragments, 0.5 m to about 1 m across, which are green with distinctly paler-green or brownish-green chilled margins (e.g.

NT 2735 8717, NT 2739 8723) (Figure 2.22). At the north-eastern end of the wave-cut platform (NT 2744 8722), a transition takes place, within a 5 m- to 10 m-thick zone, from hyaloclastite breccia to a coherent basalt lava, about 20 m thick. The base of the transition zone is marked by the appearance of lenticular sheets of green-weathered amygdaloidal basalt. Within the zone, these basalt sheets are separated by hyaloclastite breccia, but the thickness of intervening breccia decreases upwards (NT 2744 8722).

The volcaniclastic intercalations include primary and/or reworked pyroclastic rocks, up to about 5 m thick, in the form of various combinations of tuff, lapilli-tuff, lapillistone, and pyroclastic breccia (e.g. NT 2532 8651, NT 2676 8608, NT 2708 8633). These deposits were probably produced by mildly explosive volcanism, which accompanied the eruption of the lava flows. A number of intercalations display a distinctive reddish-brown colouration at the top, which is interpreted as a result of baking and oxidation by the overlying lava (e.g. NT 2540 8647, NT 2703 8616). Subaqueous accumulation of some pyroclastic deposits is implied where they occur interbedded with siliciclastic sedimentary rocks (e.g. NT 2528 8651). There are some intercalations of volcaniclastic sedimentary rocks, up to about 5 m thick, in the form of tuffaceous siltstone, sandstone and/or conglomerate, made up of basalt clasts derived from the erosion of pre-existing rocks (e.g. NT 2542 8646, NT 2536 8650).

Interpretation

The basaltic lava flows at the Burntisland to Kinghorn Coast GCR site are interpreted as the products of subaerial, effusive volcanism that probably involved hawaiian- and/or strombolian-type eruptions. Both types of eruption are dominated by effusions of low-viscosity, gas-poor magma, typically basaltic in composition, and they differ only in that magma discharge is more intermittent during strombolian eruptions (Walker and Croasdale, 1972; Blackburn *et al.*, 1976; McPhie *et al.*, 1993). During these types of eruption, only small amounts of pyroclastic material are produced, and most of this accumulates close to the vent, forming a cone of scoriaceous lapilli and blocks, with some bombs. No such cones are preserved within the volcanic succession of this site. Therefore, the abundance of tuff in the succession is indicative of more distal accumulations of ash and/or small

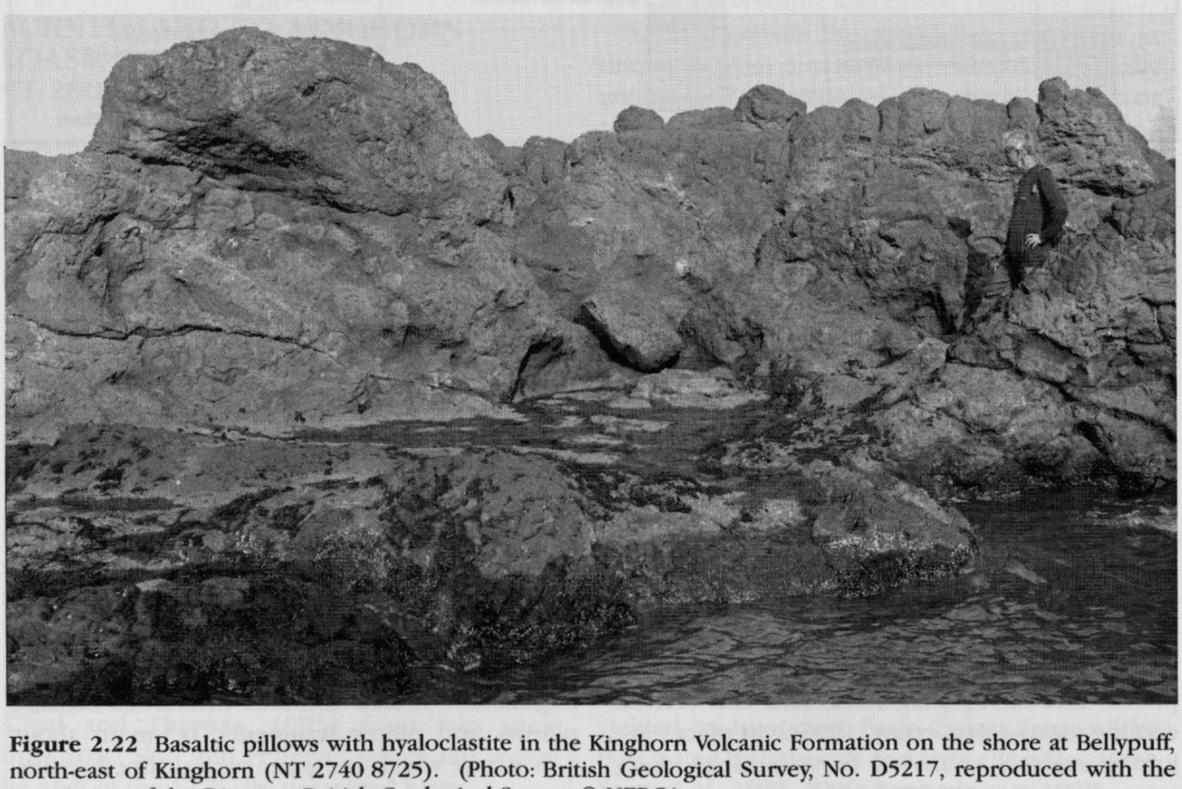


Figure 2.22 Basaltic pillows with hyaloclastite in the Kinghorn Volcanic Formation on the shore at Bellypuff, north-east of Kinghorn (NT 2740 8725). (Photo: British Geological Survey, No. D5217, reproduced with the permission of the Director, British Geological Survey, © NERC.)

lapilli, which were probably deposited downwind from vents. There are only a few occurrences of coarser-grained deposits (e.g. lapillistone and pyroclastic breccia), which may have accumulated more proximally.

Depositional environments during the volcanism probably varied from subaerial, with scattered swamps and shallow lakes, similar to those envisaged by Rex and Scott (1987), in the Pettycur area, to shallow subaqueous lacustrine, deltaic and marine. Although many of the lavas are interbedded with sedimentary rocks, the infrequency of evidence for magma–water interaction (e.g. hyaloclastite and/or pillow lava) is consistent with dominant subaerial eruption. However, the presence of load structures at the base of, and sedimentary inclusions within, some flows indicates that some of the lavas were erupted onto a substrate of semi-lithified, possibly wet, sediment. The inclusions are considered to represent masses of sediment isolated from the substrate as portions of the basalt sank into such sediment. The presence of interbedded sedimentary rocks can be explained by a combination of regional tectonic subsidence and eustatic sea-level changes, which repeatedly drowned newly erupted lavas.

Evidence that some lavas did encounter significant bodies of water is indicated by the presence of hyaloclastite. This has formed by the non-explosive quench fragmentation of magma (McPhie *et al.*, 1993). That exposed along the coast north-east of Kinghorn is capped by basalt lava, which displays no evidence of subaqueous emplacement. This relationship is similar to that present in lava deltas, formed as subaerially erupted lava flowed into water (Jones and Nelson, 1970; Moore *et al.*, 1973; Furnes and Sturt, 1976; Cas and Wright, 1987). Consequently, the hyaloclastite north-east of Kinghorn is interpreted as having formed the lower, subaqueous part of a lava delta. The transition upwards from hyaloclastite to basalt lava marks the approximate water level during delta formation, and consequently the 35–40 m thickness of the hyaloclastite provides an indication of minimum water depth. The presence of the trace fossil, *Rhizocorallium*, in siliciclastic sedimentary rocks immediately beneath the hyaloclastite, suggests that the delta formed during or soon after a marine transgression. The thickness (5–10 m) and complexity (alternating sheets of lava and hyaloclastite) of the transition zone is possibly due to the combined effects of tidal variation and subsidence of the lava delta (Furnes and Sturt, 1976). The overlying coherent basalt lava forms the subaerial part of the delta.

Conclusions

The Burntisland to Kinghorn Coast GCR site is representative of the late Visean Kinghorn Volcanic Formation and provides a well-exposed, almost complete section through an Early Carboniferous volcanic field. The volcanism took place predominantly on land close to sea level. It probably involved the production of many small volcanic cones from which extensive basaltic lavas were erupted, although none of the volcanic cones are preserved within the site. Periodically, parts of the volcanic field were submerged, either by freshwater lakes, areas of brackish water, or the sea, within which mudstone, siltstone and sandstone were deposited. During one such period of partial submergence, possibly brought about by a rise in sea level, a lava flow encountered the sea. Abrupt cooling of the lava, as it came into contact with the water, caused it to fragment and the angular lava fragments accumulated by continuous avalanching from the edge of the lava to form a delta on the seabed. However, not all of the lava was fragmented; some formed elongate pillow-like masses (pillow lava) surrounded by narrow 'skins' of rapidly cooled lava. This occurrence of fragmented lava, with associated pillow-like masses, is one of few within the Carboniferous–Permian Igneous Province of northern Britain.

TOUCH, FINTRY AND GARGUNNOCK HILLS, STIRLING (NS 650 867–NS 626 895–NS 730 934)

I.T. Williamson

Introduction

The Touch, Fintry and Gargunnock hills between Fintry and Stirling comprise the most north-easterly fault-bound block of the Dinantian Clyde Plateau Volcanic Formation. The GCR site representative of this fault block comprises the whole of the spectacular, NNW-facing escarpment of the Touch and Gargunnock hills and the WNW-facing scarp of the Fintry Hills to the west (Figures 2.23 and 2.24). Prominent high points are Stronend (NS 629 895) (511 m), Lees Hill (NS 660 910) (411 m) and

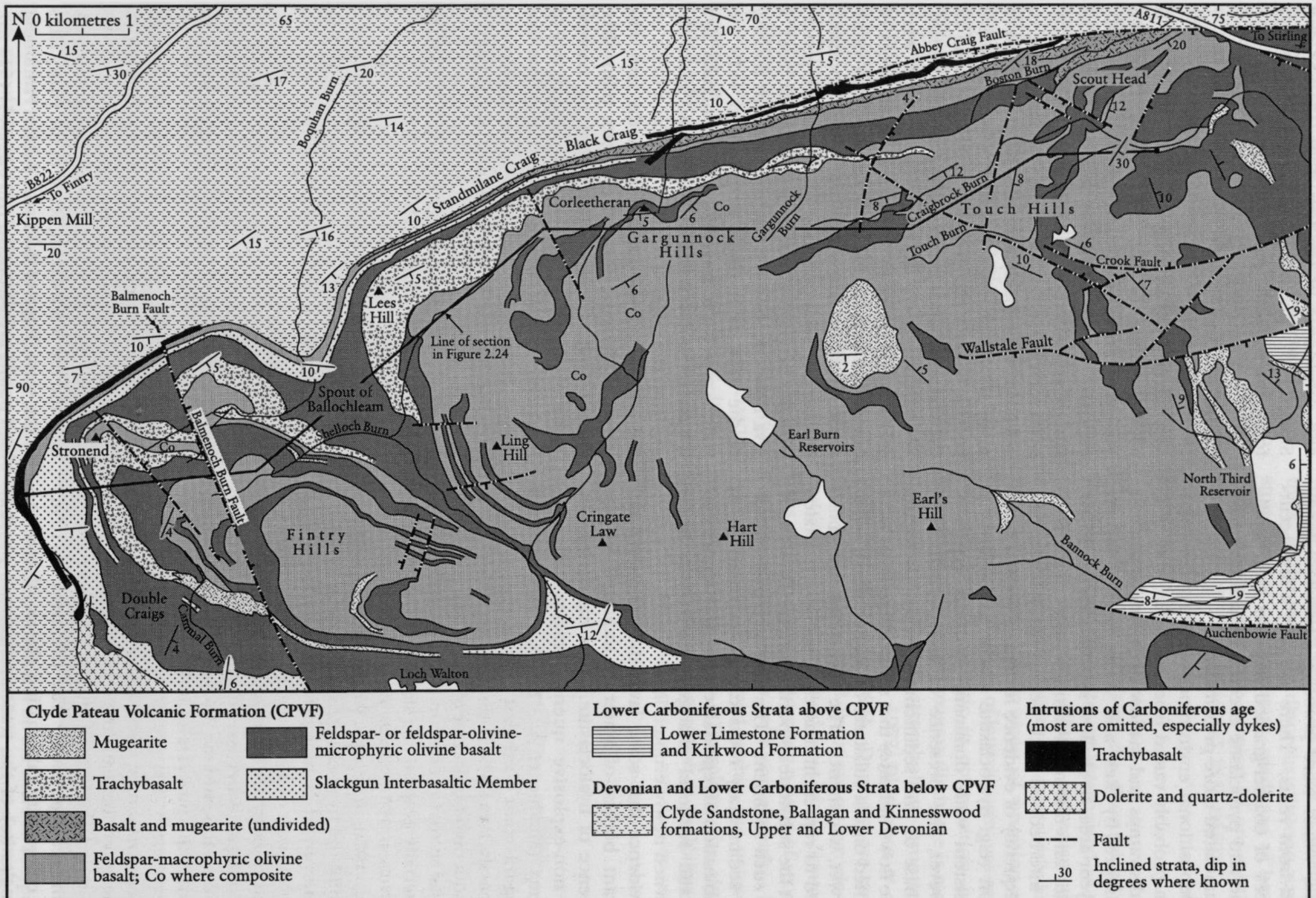


Figure 2.23 Map of the area around the Touch, Fintry and Gargunnock Hills GCR site. Based on Geological Survey 1:50 000 Sheet 39, Stirling (1970).

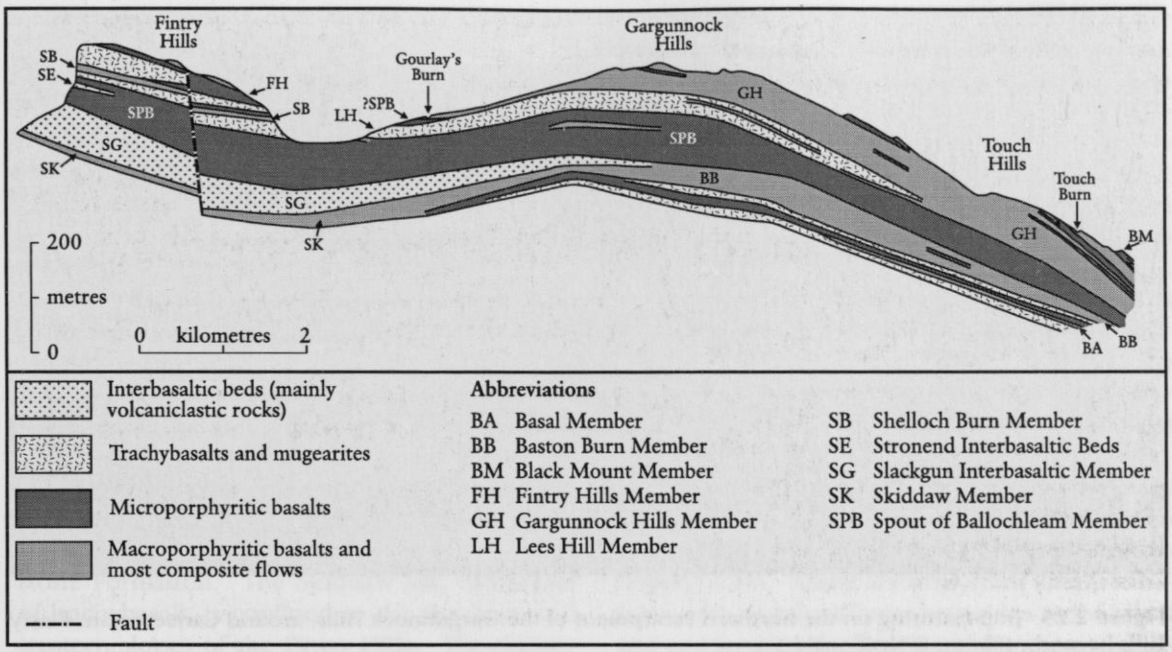


Figure 2.24 Cross-section of the northern part of the Touch, Fintry and Gargunnock Hills GCR site showing the dominant lava-types and boundaries between members of the Clyde Plateau Volcanic Formation. After Francis *et al.* (1970).

Carleetheran (NS 688 919) (485 m). A terraced, or 'trap', topography is particularly well developed along parts of the escarpment and the slopes immediately above (Figure 2.25). The highest ground and dip-slopes form an undulating and relatively poorly exposed moorland plateau. This terrain is mainly covered by a veneer of peat, though streams have cut through this and underlying glacial deposits, locally exposing bedrock. The Touch Hills have less cover of superficial deposits. Landslips are a major feature of the escarpment face, and talus and small alluvial cones have developed in many places.

The volcanic rocks were not sub-divided on early geological maps of these hills and only brief descriptions were given by Geikie (1897), who included them within the Clyde Plateau. Dixon (1938) was the first to give more details and to sub-divide the lavas. Little, if any, advance on this was made prior to the Geological Survey's re-mapping during the 1950s, almost entirely by W.A. Read, and the subsequent publication of the accompanying memoir (Francis *et al.*, 1970). That work remains the most detailed study of the district.

Description

The Clyde Plateau Volcanic Formation exposed along the northern escarpment of the Gargunnock Hills and Fintry Hills is conformable upon the Clyde Sandstone Formation of the Inverclyde Group (Paterson and Hall, 1986). This is predominantly composed of fluvial sandstone and conglomerate (Read and Johnson, 1967) and it is probably of late Tournaisian (Chadian) age. The Gargunnock Burn (NS 7072 9298 to NS 7067 9333) is also selected as a GCR site for the Dinantian strata below the lavas (see Cossey *et al.*, in prep.).

The upper age of the Clyde Plateau Volcanic Formation is less easily defined than its base as, outwith the GCR site, the highest lavas are separated from overlying volcanic detritus by a markedly diachronous erosional regional unconformity (Dinham and Haldane, 1932; Francis *et al.*, 1970). This unit of conglomerates, sandstones and mudstones, derived from the weathering of the volcanic formation, is similar to the Kirkwood Formation in the western blocks of the Clyde Plateau Volanic Formation (Paterson and Hall, 1986) and is probably of late Visean age.

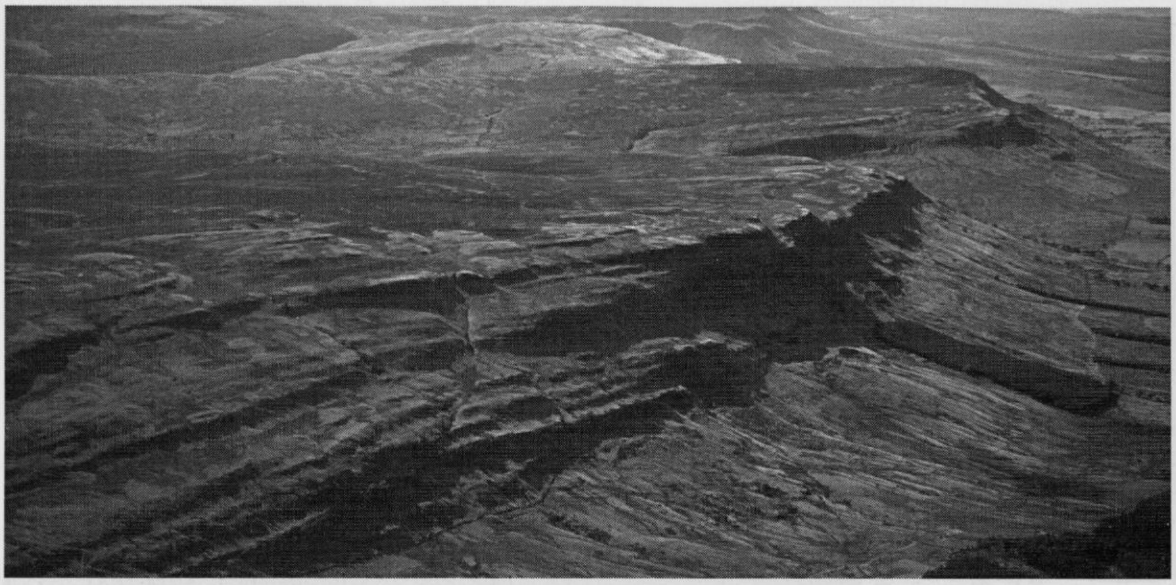


Figure 2.25 Trap-featuring on the northern escarpment of the Gargunnock Hills, around Carleetheran; Fintry Hills beyond. (Photo: P. Macdonald.)

Evidence for Dinantian volcanism prior to eruption of the earliest lavas is contained in beds of reworked volcaniclastic detritus below the oldest flows at several localities. Francis *et al.* (1970) included these within the uppermost part of the Clyde Sandstone Formation and noted increasing amounts of volcaniclastic detritus upward in the formation. Near Craigend, on the northern slopes of the Fintry Hills (NS 6229 8953), for example, 'white pebbly sandstones are reported to grade up into yellowish and greenish volcanic detritus which locally contains plant impressions'. Some beds closely resemble sedimentary units intercalated with the lava succession. In the Gargunnock Burn (NS 7072 9295) dark-red mudstones underlie the lavas and contain grains of plagioclase, pseudomorphs after olivine and decomposed basalt. These were derived either from pre-existing volcanic rocks or from ash-fall deposits.

The lava-dominated main volcanic sequence is 300–400 m thick and thins northwards and eastwards. The lavas are mainly plagioclase-phyric basalts described traditionally as olivine basalts of 'Markle' (macroporphyritic) and 'Jedburgh' (microporphyritic) types. They are more comprehensively described as plagioclase ± olivine ± Fe-oxide-phyric basalts, basaltic hawaiites and hawaiites (Macdonald, 1975). Some flows may be composite. There are also subordinate trachybasalts and mugearites, the latter mainly in the upper half of the succession.

Most of the lavas comprise a central massive facies between marginal rubbly, clinkery and amygdaloidal facies suggesting that they are aa lava. However, many show structures transitional between aa and pahoehoe lava types. Some, for example, exhibit rounded or elliptical masses of either massive or crudely columnar lava enveloped within less structured amygdaloidal lava and autobreccia. These structures may be considered as auto-intrusive features, perhaps representing cross-sections through infilled lava tubes. Such features are more common in pahoehoe lava fields. Some flows are clearly traceable along strike for several kilometres whereas others are more localized, perhaps only a kilometre or less in section. This probably reflects varying volume, viscosity and effusion rates as well as local topographical control during eruption. The lateral persistence, uniform stratigraphy and relatively constant thickness of most of the flows, as seen on the comparatively well-exposed northern escarpment, are typical of continental flood-basalt terrains. Many of the lavas have irregular weathered tops, and lateritic palaeosols (boles) are developed on some. These are seen particularly well at Double Craigs in the Fintry Hills, for example. A thick intercalation of laterite, detrital volcaniclastic sedimentary rocks and possible beds of reworked tephra (the Slackgun Interbasaltic Member), occurs in the lower half of the sequence. This marks a prolonged episode of weathering and the re-distribution of weathered and unconsolidated volcaniclastic materials.

In marked contrast to most of the other regional blocks of the Clyde Plateau Volcanic Formation, there are no agglomerate-filled necks cutting the volcanic sequence in the Touch, Fintry and Gargunnock hills, but there are rare plugs. A good example is the irregular intrusion of microporphyritic olivine basalt within the Slackgun Interbasaltic Member at the Dun, north-east of Fintry (NS 628 873). However, several basaltic and doleritic dykes cut the volcanic sequence and the underlying strata. Most are thin, nearly vertical bodies trending a few degrees either east or west of north. Two major sills intrude the upper parts of the Clyde Sandstone Formation. The Skiddaw Sill, composed of trachybasalt, is confined to the northern and western slopes of the Fintry Hills. The distinctive Downie's Loup Sill, exposed below the northern escarpment of the Touch Hills, is a composite mass of trachybasalt with abundant large feldspar phenocrysts in its lower part, but is aphyric and finer grained in its upper part.

In the west of the GCR site the lavas dip generally between 4° and 8° towards the south-east and south, but farther east the strike swings round so that they generally dip eastwards and north-eastwards at 5° to 15° (Figure 2.23). Along the northern escarpment of the Gargunnock and Fintry hills, the dips of the lower lavas are greater than the dips of those higher in the sequence. There are some large open fold structures, but it is not clear to what extent these reflect major palaeosurface irregularities, volcanotectonic activity, or later tectonic events.

There are relatively few faults in the Gargunnock and Fintry hills. A number trend either north-west or NNW, generally with small throws down to the north-east. One of the largest is the Balmenoch Burn Fault which throws down to the ENE by at least 30 m. In other parts of the GCR site, some of the very large, approximately E–W-trending structures that affect the Stirlingshire Coalfield (e.g. the Wallstale and Auchenbowie faults) also cut the Clyde Plateau Volcanic Formation.

A few major- and trace-element analyses of lavas from the Touch, Fintry and Gargunnock hills have been published in papers by Macdonald (1975), Macdonald *et al.* (1977) and Smedley (1988a), with additional data available in theses by Craig (1980) and Smedley (1986a). As with the Clyde Plateau Volcanic Formation in general, the lavas are mildly alkaline to transitional in character, meaning that basic members of the suite comprise both nepheline- and hypersthene-normative types. They can be assigned to the differentiation series ankaramitic basalt–basalt–hawaiite–mugearite–benmoreite–trachyte–rhyolite, although rocks more evolved than mugearite are not recorded from this GCR site.

The detailed lithostratigraphy of the Clyde Plateau Volcanic Formation in the Touch, Fintry and Gargunnock hills was established by Francis *et al.* (1970). Eleven lava 'groups' were recognized and these are now regarded as members of the formation. The members interdigitate in part and their boundaries may be diachronous (Figure 2.24). There are important lateral variations, with slightly different sequences in the north and east forming the Touch and Gargunnock hills and in the south and west forming the Fintry Hills. Correlation between the Touch, Fintry and Gargunnock hills sequence and the volcanic sequences in the Campsie Fells, Kilsyth Hills and Denny Muir has been established by Forsyth *et al.* (1996) (Figure 2.4). The succession exposed in the GCR site is as follows (Table 2.1).

Basal Member

The Basal Member is only present in the eastern area of the Touch and Gargunnock hills, where it forms minor cliffs at the foot of the main escarpment. It comprises a thin sequence of varied lava types with trachybasalt being the most common. In the Baston Burn, (between NS 7311 9352 and NS 7315 9340), Francis *et al.* (1970) recorded a 31 m-thick section comprising at least four flows. A basal, thick and laterally persistent mugearite is overlain by plagioclase-macrophyric and plagioclase-microphyric flows.

Baston Burn Member

The Baston Burn Member comprises mainly plagioclase-macrophyric flows but, as Francis *et al.* (1970) only refer to them as 'Markle basalts', the sequence may also include basaltic hawaiite and hawaiite. Rare plagioclase-microphyric units may represent parts of localized composite flows. From Baston Burn to Gargunnock Burn, both the thickness and the number of flows increases.

Table 2.1 Succession of the Clyde Plateau Volcanic Formation in the northern part of the Touch, Fintry and Gargunnock hills. (After Francis *et al.*, 1970, table 7.)

Fintry Hills	Touch and Gargunnock hills
Kirkwood Formation Conglomerates, sandstones and mudstones derived from weathering of Clyde Plateau Volcanic Formation	
Regional diachronous erosional unconformity	
	Touch House Member (thickness not known) Feldspar-macrophyric olivine basalts
	Black Mount Member (>24 metres) Microporphyritic basalts and subordinate trachybasalts
Fintry Hills Member (>122 metres) Feldspar-macrophyric olivine basalts with a high proportion of microporphyritic basalts and rare trachybasalts	**Gargunnock Hills Member (91–>152 metres)** Feldspar-macrophyric olivine basalts and composite basalts with subordinate microporphyritic basalts and rare mugearites
Shelloch Burn Member (40–60 metres) Trachybasalts, microporphyritic basalts and feldspar-macrophyric olivine basalts	**Lees Hill Member (0–40 metres)** Trachybasalts
Spout of Ballochleam Member (24–92 metres) Feldspar-microphyric basalts; Stronend Interbasaltic Beds in middle part	
Slackgun Interbasaltic Member (0–79 metres) Tuff (possibly volcanic detritus), laterites and weathered lavas	
Unconformity	
Skiddaw Member (0–37 metres) Feldspar-macrophyric olivine basalts and composite basalts	**Baston Burn Member (9–67 metres)** Feldspar-macrophyric olivine basalts
	Basal Member (30–46 metres) Trachybasalts, feldspar-macrophyric olivine basalts and feldspar-microphyric basalts
Clyde Sandstone Formation Fluvial sandstones and conglomerates; some reworked volcaniclastic detritus in upper part	

The Gargunnock Burn section (NS 7073 9320–NS 7234 9297) shows the maximum development of the sequence, with at least ten flows, totalling about 67 m. Farther west, successive flows are truncated by the overlying Spout of Ballochleam Member, and at Standmilane Craig (NS 6760 9214–NS 6683 9176), the section is reduced to only two flows, totalling 12 m.

Skiddaw Member

From the western end of the Gargunnock Hills into the Fintry Hills, the earliest lavas of the Clyde Plateau Volcanic Formation are assigned to the Skiddaw Member. The lavas are plagioclase-macrophyric types, similar to those of the Baston Burn Member, along with some composite basalt and microporphyritic basalt flows. At Slackgun (NS 6572 9122), below Lees Hill, Francis *et al.* (1970) recorded a 34 m-thick section. Partial sections are seen in the nearby Boquhan Burn and there are intermittent exposures along the western slopes of the Fintry Hills below Stronend. South of Skiddaw (NS 6210 8905), the member appears to be absent, so that the succeeding Slackgun Interbasaltic Member and beds within the upper part of the Clyde Sandstone Formation possibly merge and are difficult to distinguish from each other.

Slackgun Interbasaltic Member

The Slackgun Interbasaltic Member is a heterogeneous unit. There are intermittent exposures along parts of Standmilane Craig, but the best are at Slackgun (NS 6576 9118–NS 6577 9112), where Francis *et al.* (1970) recorded some 34 m. There are scattered exposures along the northern and western slopes of the Fintry Hills. Here there are three important sections: volcaniclastic rocks occur in the Boquhan Burn (NS 6526 8994–NS 6523 9014) (48 m), and the Cammal Burn (NS 6433 8697) (probably in excess of 61 m), at much the same stratigraphical level as thick and massive red laterite in the Balmenoch Burn (NS 6482 8694). The Slackgun Interbasaltic Member is not present in the eastern Gargunnock Hills nor in the Touch Hills, where an unconformity separates the Baston Burn and Spout of Ballochleam members.

There are two main lithofacies associations. The lower part is characterized by thick lateritic deposits and rare, thin and laterally impersistent olivine-phyric basaltic lavas, whereas the upper parts form a stratified sequence composed of tuff and volcaniclastic sedimentary rocks. Some of the coarser-grained beds may be cross-bedded and some contain large spindle-shaped clasts that may be volcanic bombs. The lateritic deposits mostly derive from deep subaerial weathering of volcanic rocks *in situ*, but locally may show evidence of reworking. The member possibly formed, at least in part, during a relatively quiescent interlude in the development of the lava field. Craig (1980) interpreted it as the degraded remains of a line of ash cones.

Spout of Ballochleam Member

The type locality for the Spout of Ballochleam Member is the Boquhan Burn at the Spout of Ballochleam (NS 6490 8963–NS 6526 8994), where there are nine lavas, totalling 85 m. At least eight flows are exposed in the cliffs of the western Fintry Hills. All but the lowest lava are plagioclase-microphyric basalts and hawaiites. At NS 6425 9047, the lowest lava, scoriaceous olivine-microphyric basalt up to 20 m thick, envelops large rounded bodies of massive basalt interpreted as infilled lava tubes. A similar sequence, more than 100 m thick, occurs in the southern Fintry Hills. In the eastern Touch Hills there are fewer flows, totalling about 30 m, but farther west both their number and thickness increase. At Easter Blackspout (NS 6912 9252–NS 6911 9242) and east of Standmilane Craig the sequence is more than 80 m thick.

The Stronend Interbasaltic Beds, comprising a laterite up to 2.75 m thick, occur in the middle part of the member and are seen best below Stronend at NS 6288 8885 and in a tributary of the Cammal Burn at NS 6399 8764.

Lees Hill Member

The Lees Hill Member generally forms the crest of the Gargunnock Hills escarpment at the top of cliffs formed by the Spout of Ballochleam Member. There are a few rare olivine-plagioclase-macrophyric flows, but otherwise the Lees Hill Member is dominated by trachybasalt. Francis *et al.* (1970) recorded a 40 m section of at least two thick trachybasalt flows in the Gargunnock Burn (between NS 7065 9249 and NS 7059 9222).

Shelloch Burn Member

The Shelloch Burn Member comprises a variable sequence of trachybasalts and mugearites, along with olivine-plagioclase-macrophyric basalts ('Markle' type) and microporphyritic basalts. Its fullest development is in the northern Fintry Hills where it varies in thickness from 40 m to 60 m. Representative sections are exposed in the Shelloch Burn (NS 6509 8913–NS 6523 8924), the Boquhan Burn (NS 6468 8967–NS 6490 8963) and in the crags along the western side of the Fintry Hills (NS 6281 8910–NS 6304 8870).

The Shelloch Burn and Lees Hill members cannot be correlated despite their similar stratigraphical positions, their lithological similarities and their close proximity. Regionally, they probably correlate with the Langhill and Lower Lecket Hill lavas in the Campsie Fells block (Craig, 1980; Forsyth *et al.*, 1996).

Gargunnock Hills Member

The Gargunnock Hills Member forms the greater part of the dip-slope of the Touch and Gargunnock hills. It comprises mainly plagioclase-macrophyric ('Markle' type) basalt (at least four flows of which are composite), several microporphyritic basalt lavas and rare mugearite. There are also significant lateral variations in the sequence, with microporphyritic basalt more common towards the Fintry Hills and the local development of the composite flows. In the eastern Touch Hills the member is more than 90 m thick, but it may have originally exceeded 150 m in the Gargunnock Hills. It can be divided into two units. Part of the upper unit is well exposed on Craigbrock Hill (NS 7385 9295) and in the Touch Burn below Gilmour's Linn (NS 7395 9252). The remainder is seen in the area around Scout Head (NS 7354 9337–NS 7350 9313), where Francis *et al.* (1970) recorded a 75 m-thick section.

Fintry Hills Member

The Fintry Hills Member is the youngest unit of the Clyde Plateau Volcanic Formation in the western parts of the Touch, Fintry and Gargunnock Hills GCR site, where it forms the highest ground of the Fintry Hills, south-east of Stronend. The sequence consists of plagioclase-macrophyric ('Markle' type) lavas intercalated with a high proportion of microporphyritic basalts and rare mugearites, and is at least 122 m thick. It may be equivalent to the Gargunnock Hills Member in the Gargunnock and Touch hills, and regionally may correlate with the Denny Muir, Kilsyth Hills and Holehead lavas in the Campsie Fells block (Craig, 1980; Forsyth *et al.*, 1996).

Black Mount Member and Touch House Member

The youngest flows of the Clyde Plateau Volcanic Formation in the Touch, Fintry and Gargunnock hills are not seen within the GCR site. The Black Mount Member comprises at least 24 m of microporphyritic basalt and subordinate trachybasalt and the Touch House Member is composed entirely of plagioclase-macrophyric lavas.

Interpretation

Correlation of the Touch, Fintry and Gargunnock hills sequence with other sequences in the Clyde Plateau Volcanic Formation is illustrated in Figure 2.4. The lower members are not easily correlated with sequences in the Campsie Fells and are perhaps localized developments. Although substantial fragmental deposits occur at the base of the formation in the Kilpatrick Hills, they thin eastwards and are not seen to continue into the Campsie Fells block. In the Kilsyth Hills and Denny Muir area of the latter, the lowest member comprises interbedded proximal facies lavas and thick tuffs, the latter comprising up to half the total thickness. It is possible that the volcaniclastic beds within the upper parts of the Clyde Sandstone Formation in the Fintry and Gargunnock hills, are distal equivalents of these tuffs.

Evidence for the existence of small shield volcanoes, vents and caldera structures is abundant throughout the northern Clyde Plateau; many eruption sites lie along linear features. However, unlike the sequences in the Campsie Fells and the Kilsyth, Kilpatrick and Renfrewshire hills, there are only rare plugs and no recorded agglomerate-filled vents or calderas within the Touch, Fintry and Gargunnock hills block. Also, the dykes present seem to be too few and insignificant to be considered the main sources of the lavas. Hence, it is more likely that most of the laterally continuous pahoehoe flows with 'continental flood-basalt' characteristics are the distal products of eruptions along the linear vent–fissure systems in the Campsie and Kilsyth hills (Whyte and MacDonald, 1974; Craig and Hall, 1975; Hall *et al.*, 1998). Other flows, exhibiting proximal features and interbedded tuffs, may have been derived more directly from less voluminous eruptions centred upon small composite shield volcanoes south-west of the area.

Some members of the Clyde Plateau Volcanic Formation apparently thicken south-westwards from the Gargunnock Hills towards the Campsie Fells. This is especially clear in the Spout of Ballochleam Member and its proposed equivalent, the Campsie lavas, as illustrated in Figure 2.4. This could suggest the existence, in addition to the vents and structures such as the Waterhead Central Volcanic Complex (Craig, 1980; Forsyth *et al.*, 1996), of at least one major volcano in the Campsie Fells. However, there is considerable evidence to support, at least locally, the views of Tyrrell (1937) that the Clyde Plateau was more the result of coalesced lava flows erupted from a large number of small, closely spaced volcanoes. The absence of clear evidence of vent structures in the Touch, Fintry and Gargunnock hills block is in stark contrast to the well-documented **Campsie Fells** block to the south-west (see GCR site report).

Conclusions

The Touch, Fintry and Gargunnock Hills GCR site represents the most north-easterly of several large, fault-bound blocks that make up the overall outcrop of the Visean Clyde Plateau Volcanic Formation, by far the thickest and most extensive outpouring of Carboniferous or Permian volcanic rocks in the whole of Britain. Subparallel terracing, or 'trap' topography, is well developed and is seen particularly well on the long, spectacular, north-facing escarpment of these hills. The GCR site includes rock-types ranging from basalt to mugearite in composition, and there are several examples of ash-fall tuffs, probable remnants of ash cones, interflow sedimentary deposits derived from the reworking of the volcanic rocks, and soils that developed between eruptions. A detailed stratigraphy has been established, enabling comparisons with other blocks within the outcrop of the Clyde Plateau Volcanic Formation, thus allowing realistic three-dimensional models to be constructed for the evolution of the entire lava field through time.

CAMPSIE FELLS, STIRLING and EAST DUNBARTONSHIRE (NS 572 800–NS 535 825–NS 609 867)

J.G. MacDonald

Introduction

The Visean lavas and pyroclastic rocks of the Clyde Plateau Volcanic Formation, together with associated vents and intrusions, form the hilly areas that lie to the north, west and south of Glasgow. In the northern part of the Clyde Plateau the steep slopes that bound the Campsie Fells provide extensive exposures (Figures 2.26 and 2.27). Glacial erosion has produced escarpments, which afford well-exposed sections through the lower part of the lava succession. The spectacular escarpment on the north-western margin of the Campsie Fells is carved out of the largest concentration of vents and intrusions in the Clyde Plateau, the North Campsie Linear Vent System (Figure 2.28).

The Campsie Fells GCR site extends for 8 km along the NW-facing escarpment, from Dunmore, above the village of Fintry, as far as the twin volcanic plugs of Dumgoyne and Dumfoyne. From there it continues south-eastwards along a 4.5 km stretch of the SW-facing escarpment of the Strathblane Hills between Dumfoyne and the Spout of Ballagan. As well as the volcanic features of the area, localities of stratigraphical and palaeontological interest have been notified at the Balglass corries in the north and at Ballagan Glen in the south (see Cossey *et al.*, in prep.).

The earliest detailed description of the geology of the Campsie Fells by Young (1860) proved to be so popular that it was reprinted in 1868 and 1893. More recent sources of information (Clough *et al.*, 1925; MacDonald, 1967; Whyte and MacDonald, 1974; MacDonald and Whyte, 1981; Hall *et al.*, 1998) provide a general picture of the petrography and geochemistry of the volcanic rocks, which vary in composition from mafic basalt to trachyte. Most of these accounts draw upon more detailed information in PhD theses by MacDonald (1965) and Craig (1980). Excursions to parts of the Campsie Fells outwith the GCR site are described by MacDonald and Whyte (in Upton, 1969) and MacDonald (in Lawson and Weedon, 1992). The mildly alkaline chemistry of the rocks, their range of composition, the relationships of vents to lava flows, and the presence of a flow of relatively fresh hawaiite that exhibits interesting internal variations, afford a potential for further research in the area, which could be of international significance.

Whole-rock K-Ar radiometric dates from vent intrusions in the western Campsie Fells (De Souza, 1979) include 329 ± 7 Ma (Dumgoyne) and 316 ± 5 Ma (Dunmore) (*c.* 336 Ma and 323 Ma respectively using new constants). Ages in the range 315 ± 7 Ma to 303 ± 7 Ma were obtained from lavas farther east but De Souza considered that, given the probability of argon loss during alteration, the age of eruption is likely to be in the vicinity of, or older than, 330 Ma (*c.* 337 Ma using new constants).

Figure 2.26 Corrie of Balglass on the northern escarpment of the Campsie Fells, with the Fintry Hills in the background. Largely microporphyritic basalts and hawaiites of the Lower North Campsie Lava Member forming the steep wall of the corrie, overlie volcaniclastic rocks derived from the North Campsie Linear Vent System. (Photo: P. Macdonald.)

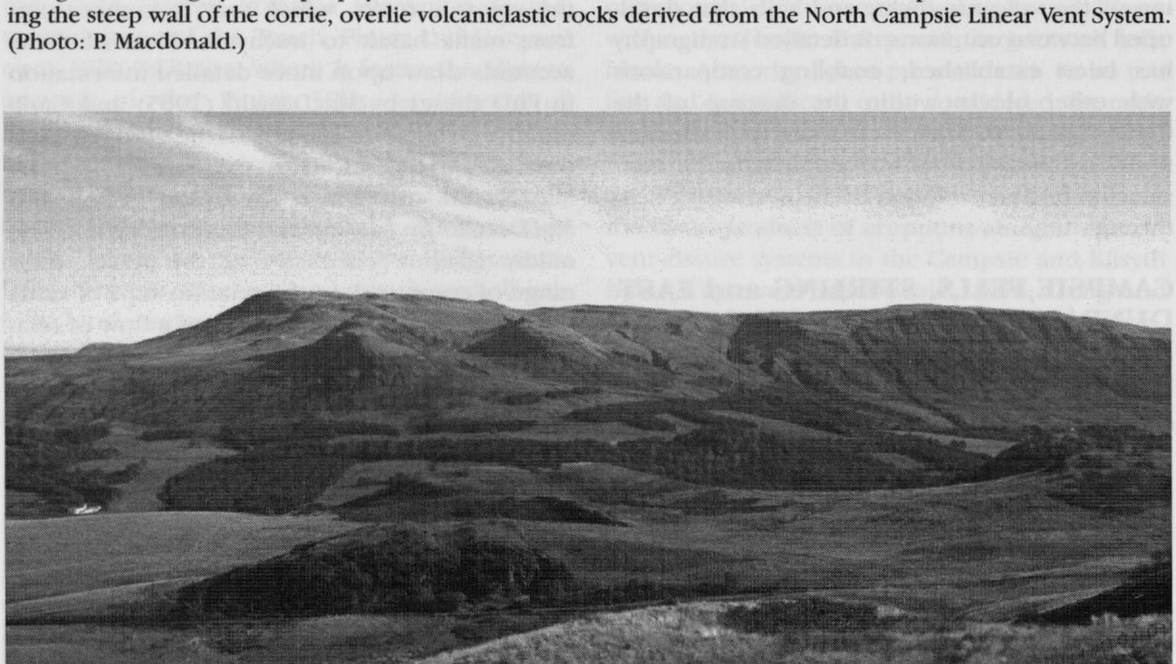


Figure 2.27 The western end of the Campsie Fells viewed across Strath Blane from the south-west. The Dumfoyne Vent is the feature in the centre of the photograph; the Dumgoyne Vent is to the left of it. The high ground on the skyline above Dumgoyne marks the south-west end of the North Campsie Linear Vent System. (Photo: J.G. MacDonald.)

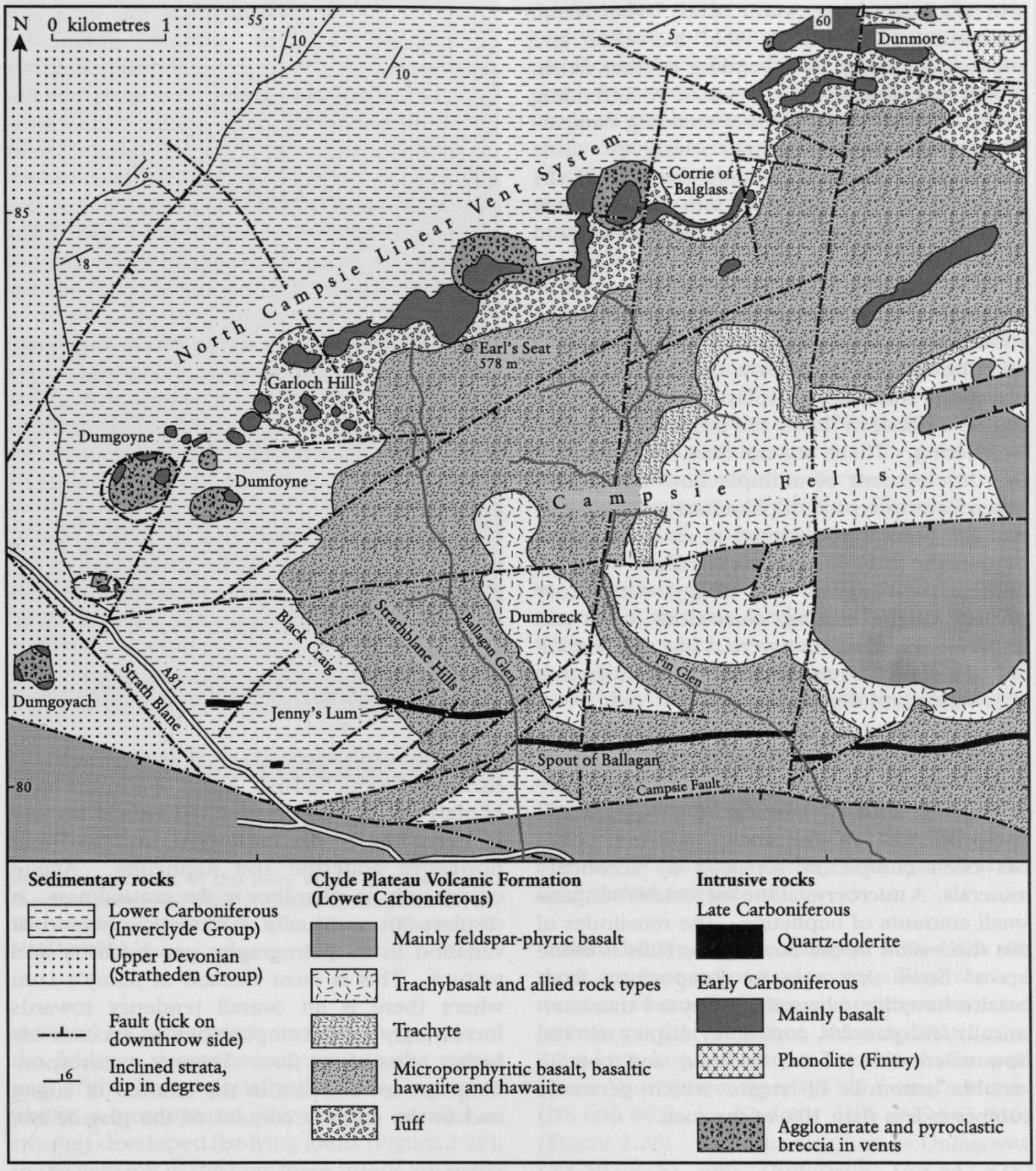


Figure 2.28 Map of the area around the Campsie Fells GCR site. Based on British Geological Survey 1:50 000 Sheet 30E, Glasgow (1993).

Description

The lava succession

Throughout the Campsie Fells the lavas rest directly upon the Clyde Sandstone Formation, the topmost division of the Tournaisian Inverclyde Group (Figures 2.4 and 2.28). The lavas form a plateau, dipping gently to the south-east and deeply dissected by Ballagan Glen (NS 565 830–NS 573 795) and Fin Glen (NS 583 833), where exposures of extensive parts of the succession supplement a continuous section in the steep cliffs at Black Craig

(NS 552 818), above Strath Blane. The overall succession is shown in Table 2.2. The lava plateau is cut by ENE-trending normal faults, which are displaced by a set of N–S-trending faults. To the south, the E–W-trending Campsie Fault downthrows to the south by some hundreds of metres, bringing the top of the Kilpatrick Hills succession into juxtaposition with the lower part of the Campsie succession (Figure 2.28).

The lavas vary in thickess from flow to flow and within flows, averaging about 10 m but in some instances exceeding 20 m. They generally take the form of single flow units, vesicular on top and in some cases displaying well-developed red bole between flows, the product of lateritic weathering. There are no clear indications of the development of multiple flow units or of other features typical of pahoehoe, so it is likely that the lavas were erupted as aa. The upper parts of the three lowest flows at Black Craig are particularly vesicular and slaggy, with drusy cavities, calcite veining, chalcedony veins and jasper lenses. The latter were at one time exploited as a source of material for the manufacture of jewellery in Edinburgh.

The basal lava at Black Craig is the most basic in the western Campsie sequence, being an olivine basalt containing abundant, randomly orientated, laths of labradorite (An_{66}), which comprise 55% of the rock. Olivine (12%) has been completely replaced by secondary minerals. A microcrystalline mesostasis contains small amounts of nepheline. The remainder of the succession in the Strathblane Hills is made up of flows that vary in composition from basaltic hawaiite to hawaiite. They are characteristically feldspar-rich, commonly display marked flow orientation of the feldspar, and contain variable amounts of augite which generally comprises less than 10% of the rock.

Studies of the petrology of the lavas are complicated by the almost ubiquitous replacement of olivine by secondary minerals such as 'serpentine', green pleochroic bowlingite and, in extreme cases, calcite. The oxidation state of the opaque oxides has also been affected to varying degrees so that titanomagnetite, which in some flows exceeds 9%, has undergone alteration resulting in the transformation of exsolved magnetite to maghemite (Goswami, 1968). This has the effect of distorting the ratio of ferrous to ferric iron in whole-rock chemical analyses to the extent that some nepheline-bearing rocks appear to be silica-oversaturated in their normative composition. Most of this alteration can be explained by reaction of the early-formed minerals with the volatile fraction of the magma during the late stages of crystallization.

The distinctive texture of a hawaiite flow that occurs near the base of the succession in the Strathblane Hills allows it to be traced along the Campsie escarpment for at least 2.4 km (MacDonald, 1967). At Jenny's Lum (NS 562 806) it has a thickness in excess of 18 m of which the lowest 15 m are fresh, almost free of vesicles and display better developed columnar joints than is normal in the Campsie lavas (Figure 2.29). It is rich in andesine feldspar, much of it in the form of platy microphenocrysts; it has a small amount of nepheline in the mesostasis, and has both normative andesine and nepheline. Augite occurs only as microlites in the groundmass. It displays an unusually systematic gradational variation in its petrography, geochemistry and texture. This is most marked at Jenny's Lum where there is an overall tendency towards increasingly basic compositions in successively higher parts of the flow. There is a corresponding upward increase in the amount of augite and in the calcium content of the plagioclase,

Table 2.2 Succession of the Clyde Plateau Volcanic Formation in the western Campsie Fells. (After Hall *et al.*, 1998, table 4)

	Lava types	Source
Holehead Lava Member	Mainly feldspar-macrophyric basalt ('Markle type')	Waterhead central volcano
Fin Glen Lava Member	Microporphyritic basalt, mugearite, trachybasalt and a persistant phonolitic trachyte	Local centres and North Campsie Linear Vent System
Upper and Lower North Campsie lava members	Microporphyritic basalt, basaltic hawaiite and hawaiite	North Campsie Linear Vent System

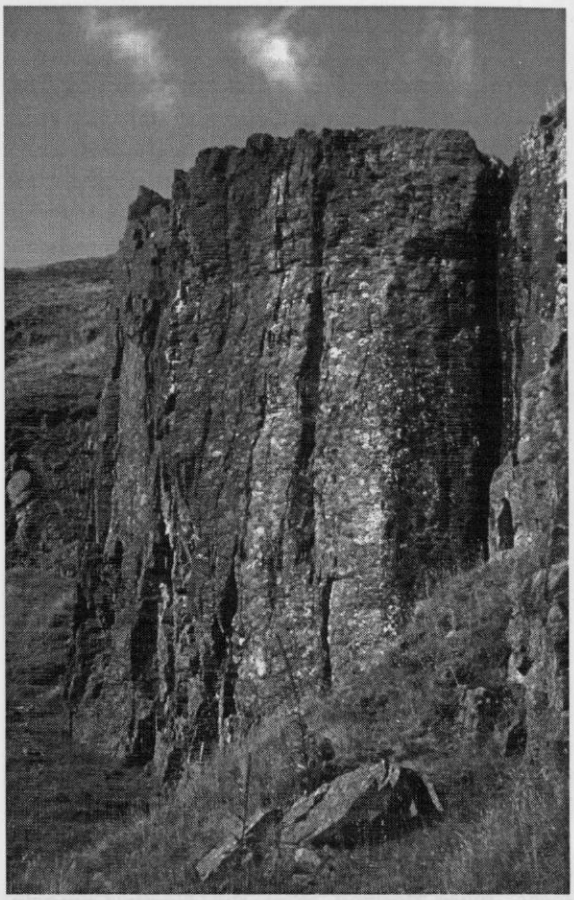


Figure 2.29 Hawaiite lava at Jenny's Lum, western Campsie Fells. Note the flat-lying joints, particularly in the upper part of the flow, which are parallel to the flow texture of platy andesine microphenocrysts. The height of the cliff is over 15 m. (Photo: J.G. MacDonald.)

matched by a systematic increase in the size of the feldspar microphenocrysts. In the upper part of the flow especially, the feldspar crystals impart a pronounced platy fabric, parallel to strongly developed flat-lying joints (Figure 2.29). In the very top of the flow these trends are partly reversed.

The gradational variations and apparent absence of internal discontinuities are very similar to those observed in hawaiitic flows in the northern part of the Renfrewshire Hills (Kennedy, 1931; see **Dunrod Hill** GCR site report). However, they are in contrast with composite lava flows involving two markedly different components, such as those described by Kennedy (1933) from elsewhere in the Renfrewshire Hills. One could argue that the Jenny's Lum and Dunrod Hill hawaiites are not truly composite but display gradational variations in composition, suggesting an orderly mode of emplacement of a magma that progressively changed in composition during the course of eruption. Such progressive variation in the composition of lava during the course of an eruption has been observed in historical activity in Iceland (Thorarinsson and Sigvaldason, 1972).

A distinctive phonolitic trachyte flow to the east of Fin Glen has petrographical similarities to an irregular intrusion of phonolite near Fintry (Hall *et al.*, 1998). This flow and a few analcime trachybasalts near North Berwick are the only silica-undersaturated evolved lavas known within the Carboniferous and Permian volcanic sequences of Britain. Elsewhere, any more evolved compositions trend towards quartz-trachytes and rhyolites (for discussion see **Traprain Law** GCR site report). The phonolitic trachyte marks the base of the Fin Glen lavas, which are, on average, more felsic than the underlying flows; in addition to basalt they include trachybasalt and mugearite. The topmost part of the succession in the western Campsie Fells consists of feldspar-macrophyric basalts ('Markle' type) of the Holehead Lava Member.

The North Campsie Linear Vent System and associated intrusions

Within the area of the Campsie Fells GCR site there are four major agglomerate-filled vents, a number of smaller ones and many associated intrusions that together form a continuous 7 km-long linear feature. This North Campsie Linear Vent System trends WSW from Dunmore (NS 606 865) to Garloch Hill (NS 553 836) (Figure 2.28). The volcanic plugs of Dumgoyne (NS 542 828) and Dumfoyne (NS 547 825) (Figure 2.27) are situated on this trend, which continues beyond the confines of the GCR site, through Dumgoyach, and to the WSW through the Kilpatrick Hills as far as Dumbarton (the Dumbarton–Fintry Line of Whyte and MacDonald, 1974; Figure 2.3; see **Dumbarton Rock** GCR site report). In close proximity to the vents, outcrops of bedded tuff and scoria represent the remains of cinder cones produced by lava fountaining.

The intrusive rocks associated with the vents vary in composition from basalt to hawaiite and mugearite. The basaltic types most commonly include microlitic and feldspar-microphyric varieties ('Jedburgh' type), and less commonly feldspar-macrophyric 'Markle' types. The more mafic varieties, rich in phenocrysts of olivine and augite, which occur in vents in the Kilpatrick Hills to the south of the Campsie Fault, are not represented in the Campsie vents. In general, the basalts and related rocks of the vent intrusions have suffered less immediate post-eruptive alteration than the lavas; olivine is much more commonly preserved, for example (see **Dumbarton Rock** GCR site report). It is likely that many vent intrusions represent fractions of magma that were emplaced at a late stage in individual eruptive sequences. As such they would commonly have been depleted in volatile constituents that had escaped to the surface through the open vent or had risen as gas bubbles to higher levels, now removed by erosion.

Whereas the North Campsie vents have an almost continuous outcrop that forms the northwestern boundary of the lava plateau, Dumgoyne, Dumfoyne and a number of smaller vents and intrusions lie to the west of the main mass of lavas, forming isolated features. These plugs cut sedimentary rocks of the Ballagan Formation and, in the case of Dumgoyne, the underlying red and white cross-bedded sandstones of the lowermost Carboniferous Kinnesswood Formation. No part of the subaerial cone of these volcanic edifices is preserved *in situ* but both Dumgoyne and Dumfoyne are composed mainly of agglomeratic material, some of which may have slumped back into the volcanic conduit from higher levels at the end of eruptive episodes. At Dumgoyne, basaltic intrusions cut the agglomerates, especially on the eastern side of the vent where a major dyke-like mass occurs. Dumfoyne has only one small vent intrusion on the north side.

Within the western Campsie Fells a number of dykes have trends similar to that of the North Campsie Linear Vent System and coincide with ENE-trending normal faults. Some of these dykes are of feldspar-macrophyric ('Markle' type) basalt. Whyte and MacDonald (1974) have suggested that these could have been feeders for fissure eruptions of feldspar-phyric lavas, the latter having been subsequently removed by erosion of the top of the succession in the western Campsies.

Interpretation

The underlying structural control of the ENE-trending North Campsie Linear Vent System is probably related to a Caledonian lineament in the pre-Carboniferous basement (see 'Introduction' to this chapter). This trend is sufficiently similar to that of the feldspar-phyric dykes and associated normal faults to suggest that all three features are related to a common stress system. The high concentration of magmatic activity along the linear vent system is likely to have been accompanied by corresponding local swelling of the Earth's crust during periods of maximum magmatic activity. Such conditions are conducive to normal faulting, facilitating the intrusion of dykes, parallel to the elongation of the vents, as is seen in many areas of recent active volcanism. The swelling could also have created the palaeoslope, down which the lavas flowed away from the vents.

The similarity in petrography and geochemistry between the vent intrusions and the lavas of the western Campsie Fells (MacDonald and Whyte, 1981) make it appear likely that the bulk of the succession, comprising the Fin Glen Lava Member and the Upper and Lower Campsie lavas, were derived from the North Campsie Linear Vent System and its continuation in Dumgoyne and Dumfoyne. It is difficult to correlate individual lavas precisely with particular vents, but Dumfoyne, although mainly composed of agglomerate, features a vent intrusion of hawaiite on its north side that is similar in its geochemistry and petrography to the Jenny's Lum hawaiite (MacDonald and Whyte, 1981). It is thus possible that the latter could have been erupted from this vent and hence flowed to the south-east for a minimum distance of nearly 4 km from its point of eruption. The phonolitic trachyte flow that marks the base of the Fin Glen Lava Member has a present-day extent of about 10 km^2. If the source of this flow is the Fintry phonolite intrusion (Hall *et al.*, 1998) it could originally have had an aerial extent in excess of 20 km^2 and flowed south for at least 6 km from its source. The dominantly felspar-phyric basaltic Holehead lavas at the top of the succession were most probably erupted from a large central volcanic complex at Waterhead, some 3.5 km to the east of the area covered by Figure 2.28, which has been described by Craig (1980) and Forsyth *et al.* (1996).

Conclusions

The Campsie Fells GCR site exhibits the lower part of the volcanic succession in the Campsie Fells; it is typical in many respects of the northern outcrops of the Visean Clyde Plateau Volcanic Formation. The lava pile is bounded to the north-west by a line of deeply eroded volcanic vents, representing the roots of small volcanoes, and a continuous apron of fragmental rocks formed from the ash and cinders of the volcanic cones. The vents consist of coarse blocky material that collapsed back into the conduit of the volcano at the end of each eruption, and many are intruded by volcanic plugs, formed as fresh magma forced its way towards the surface.

The North Campsie Linear Vent System is the most concentrated example of multiple volcanic vents preserved in Dinantian times in the Midland Valley, and the lava sequence of the Campsie Fells is one of few for which the general source area and hence the type of eruption can be clearly identified. It has even been possible to tentatively suggest specific vents as the sources for some individual lavas. Some of the more distinctive lavas can be traced for considerable distances. The lavas and intrusions have been the subject of several geochemical investigations and could provide material for a variety of further studies into magmatism in the Midland Valley and the origin and evolution of magmas in general. The volcanic plugs in particular could provide fresh rocks suitable for radiometric dating, which would have wider significance for the timing of events in the Midland Valley.

DUMBARTON ROCK, WEST DUNBARTONSHIRE (NS 400 745)

J.G. MacDonald

Introduction

The prominent landmark of Dumbarton Rock, on the north bank of the River Clyde at its confluence with the River Leven, has been a fortified site since the 5th century AD or earlier. Its situation, visual impact and state of preservation make it particularly significant as an example of a volcanic plug. It is the westernmost of a series of basaltic plugs and necks, extending in a broad belt for more than 25 km in an ENE direction as far as the Campsie Fells, that define the Dumbarton–Fintry volcanotectonic line (Whyte and MacDonald, 1974; Craig and Hall, 1975; Figure 2.3; see **Campsie Fells** GCR site report). The plug cuts Tournaisian sedimentary rocks of the Inverclyde Group, and the nearest outcrops of possible associated lavas (the Visean Clyde Plateau Volcanic Formation) occur about 2.5 km to the north-east in the Kilpatrick Hills, and 1.5 km to the south, on the opposite bank of the River Clyde.

Despite its easy accessibility, the excellent exposures of fresh glacially smoothed rock and the well-exposed relationships of the intrusive basalt with adjacent tuffs, agglomerate and sandstone, very little was published on Dumbarton Rock prior to a detailed investigation by Whyte (1966). Its popularity as a geological excursion locality is reflected by descriptions in field guides, based largely upon Whyte's account (Whyte and Weedon in Lawson and Weedon, 1992). Further geochemical aspects were discussed by Whyte (1980) and a single K-Ar whole-rock age determination of 302 ± 8 Ma (*c.* 308 Ma using new constants) was reported by De Souza (1979).

Description

Dumbarton Rock rises to a height of 73 m above the reclaimed intertidal mudflats at the mouth of the River Leven (Figure 2.30). It is roughly oval in plan having an east–west elongation of 275 m and a north–south width of 200 m. A NW-trending gully divides the summit area. There is a distinctive pattern of columnar jointing. The columns, averaging about 60 cm in diameter, fan downwards and outwards at steep angles, with a tendency in some places for the inclination to become shallower near the base of the Rock (Figure 2.31).

The relationship between the intrusive basalt, adjacent pyroclastic rocks and associated sedimentary rocks is seen only on the north-west side of the plug (NS 399 746) where bedrock is exposed along about 80 m of shoreline. At the southern end of the section, sandstone, beds of fissile mudrock and carbonate rocks ('cementstones') of the (Tournaisian) Ballagan Formation dip steeply towards the contact. The sedimentary rocks are in the form of isolated blocks up to about 20 m in length, which are either faulted against tuffs and agglomerates or have fallen into them in the volcanic vent (Whyte, 1966).

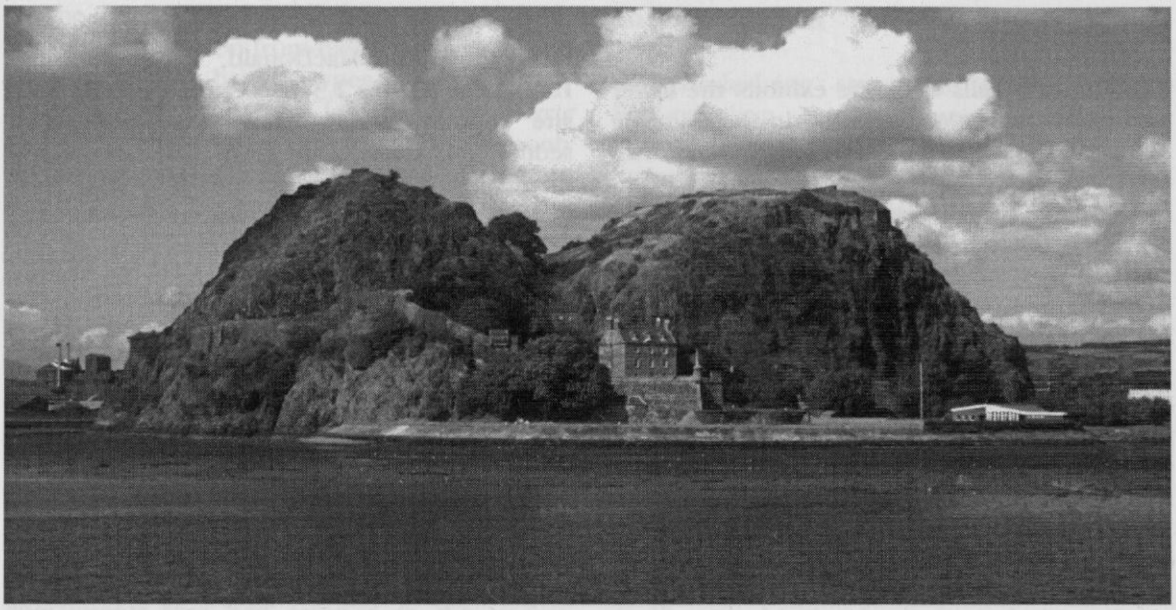


Figure 2.30 Dumbarton Rock, a plug of olivine basalt, from the River Clyde. (Photo: J.G. MacDonald.)

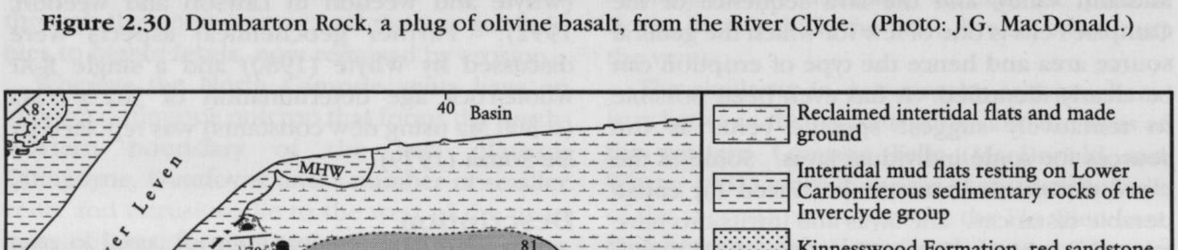


Figure 2.31 Map of the area around the Dumbarton Rock GCR site. After Whyte (1966).

There is a narrow zone of contact alteration, up to 15 cm wide, in the sandstones and 'cementstones' adjacent to the plug, and sandstone xenoliths up to about 30 cm in diameter have been incorporated in the basalt. In the xenoliths, quartz may have been altered to tridymite in reaction rims. The basalt close to the contact is chilled and highly altered. The groundmass olivine and augite are completely altered to chlorite and the feldspar to albite; abundant amygdales are also present, containing spherulitic green chlorite and in some cases a little calcite. Chemical analyses indicate the incorporation of a significant amount of water into the basalt at, or closely following, the time of intrusion. This effect decreases away from the contact and is hardly discernible a metre or so into the plug. Within this 1 m zone, barium and strontium decrease in concentration towards the margin of the basalt and there is a corresponding increase in concentration of barium and, to a lesser extent, strontium in the immediately adjacent sedimentary rocks, especially in sandstone (Whyte, 1980).

Away from the contact zone the basalt of the plug is fine grained and has a uniform microporphyritic texture with microphenocrysts of labradoritic plagioclase and olivine (Fa_{31}) set in a fine-grained groundmass of plagioclase, generally granular augite and opaque iron oxides. Accessory minerals include chlorite, analcime and apatite. In addition to the xenoliths of country rock incorporated in the margins, a few dunitic xenoliths have been found. Although no mineral layering has been detected, flow texture of microphenocrysts and groundmass feldspar laths is common. There is an apparent increase in the ratio of plagioclase to augite from an average of 2.3:1 at the base of the rock to 4.2:1 at the summit. Such variations in mode are common within lavas and plugs of basaltic affinity in the Clyde Plateau Volcanic Formation.

Compositions of fresh basalts from Dumbarton Rock (Whyte, 1966, table 2) are similar in most respects to those found in lavas and plugs of Visean age in the nearby Kilpatrick Hills. The magnesium content, although unusually low, is within the range of variation found within the Clyde Plateau Volcanic Formation (MacDonald and Whyte, 1981). The fresh rock is nepheline-normative and De Souza (1979) suggested that it should therefore be regarded as basanitic. However, no modal nepheline and only accessory amounts of analcime have been identified. On these grounds the rock should be termed an alkali olivine basalt. It differs petrographically from many lavas of the Kilpatrick Hills and other parts of the Clyde Plateau Volcanic Formation only in the freshness of the olivine. The freshness is reflected by relatively high FeO/Fe_2O_3 ratios, a feature that it shares with other plugs associated with the formation. This contrasts with the lavas, in which late-stage alteration, attributable to autometasomatism, increases the proportion of ferric iron. In turn this affects the norm calculation in such a way that the lavas often appear to be more silica-saturated than the plugs.

Interpretation

It has commonly been assumed that Dumbarton Rock was emplaced as a vent intrusion at or near the top of the conduit of a volcano that was active during the time of eruption of the lavas of the Kilpatrick Hills (Whyte, 1966). However, whereas the northern outcrops of the Clyde Plateau Volcanic Formation appear to have been erupted between about 330 Ma and 320 Ma (*c.* 337–327 Ma with new constants), a single K-Ar determination of a sample of fresh basalt from Dumbarton Rock yielded a date of *c.* 308 Ma with new constants (De Souza, 1979). This, together with the perceived basanitic nature of the plug, led De Souza to suggest that the emplacement of the plug took place in Late Carboniferous time and hence that it was not related directly to the Visean volcanism. There is nothing particularly distinctive about Dumbarton Rock in its geographical setting that would set it apart from other vent intrusions in the northern Clyde Plateau and, given the apparent petrographical affinities with the adjacent lavas and plugs, it seems appropriate, for now, to regard the single Late Carboniferous K-Ar date as an anomaly. This fresh intrusion would, however, be a good subject for further dating by the Ar-Ar method, not only to clarify its own association but also to date the age of volcanism in this part of the Midland Valley more accurately.

The joint pattern of Dumbarton Rock is consistent with the base of the cooling body of intruded basalt liquid having the form of an inverted cone. Hence Whyte (1966) deduced that the form of the base of the intrusion could have been determined by the shape of the volcanic crater into which it was emplaced. The basalt of the plug contains abundant amygdales in the contact zone but these become much less common away from the contacts. The presence of amygdales is a clear indication that the basalt was emplaced at low pressure and most likely at a time when there was a connection with the surface. From this it can be inferred that the plug infilled the vent of an active volcano and might indeed represent the lower part of a lava lake that formed in the active crater. Lava lakes formed in such conditions can remain liquid for sufficient time to allow degassing to take place; hence the paucity of amygdales in all but the chilled margins of the plug. The degassing of the magma could account for the freshness of the olivine, which in Visean volcanic rocks is almost invariably replaced by secondary minerals. Mineralogical and geochemical evidence also suggests the subsequent outward migration of volatiles, resulting in chloritization, albitization and some leaching of trace elements from the basalt into the immediately adjacent country rock (Whyte, 1980).

The marginal ring of pyroclastic rocks contains tilted blocks of the Ballagan Formation that are at a structural level below its inferred base in this area. Hence the Ballagan Formation beds most likely collapsed into the open vent, along with parts of the cone, during an interval of decreased eruptive intensity, and are preserved in a subsided cylindrical block within an inferred ring-fault (Figures 2.31 and 2.32). From the above it appears highly probable that the plug was emplaced at a time when the crater of the volcano was open. The lava lake so formed could well have been overflowing to produce a lava flow, now removed by erosion. Similar historical monogenetic volcanoes commonly erupt continuously for periods of many weeks or months and so there would have been adequate time to supply the fluids, and maintain the temperatures needed to account for the observed contact phenomena.

Conclusions

The Dumbarton Rock GCR site combines historical significance as a fortified site with geological importance as the remnant of a mass of lava that solidified in the crater of an Early Carboniferous volcano. It is composed of alkali olivine basalt and has close petrographical and geochemical affinities with the Visean volcanic rocks of the Clyde Plateau Volcanic Formation in the adjacent Kilpatrick Hills. It has special significance as an example of a volcanic plug. Its isolated position, standing above the mudflats of the Clyde estuary, provides a three-dimensional view of the columnar joint pattern that is unrivalled in the west of Scotland. The level of erosion also exhibits the relationships between the basalt plug, remnants of the volcanic cone and country rocks that are seldom seen in the remains of Palaeozoic volcanoes. A single K-Ar radiometric date that suggests a Late Carboniferous age appears to be anomalous in the light of the other evidence and highlights the need for further dating using more accurate modern methods.

DUNROD HILL, INVERCLYDE (NS 236 741–NS 246 721)

D. Stephenson

Introduction

Dunrod Hill, 5 km south of Gourock, is part of the Renfrewshire Hills succession of the Clyde Plateau Volcanic Formation, of Dinantian age. The lavas are typical representatives of the Strathgryfe Lava Member, which comprises the thickest part of the formation and dominates the northern part of the Renfrewshire Hills. It has been selected for the GCR because of excellent exposures of composite lava flows of hawaiite composition, first described by W.Q. Kennedy in 1931. Such flows, comprising a markedly feldspar-phyric upper part overlying an aphyric base, are fairly common in the Strathgryfe Lava Member and have also been described else-

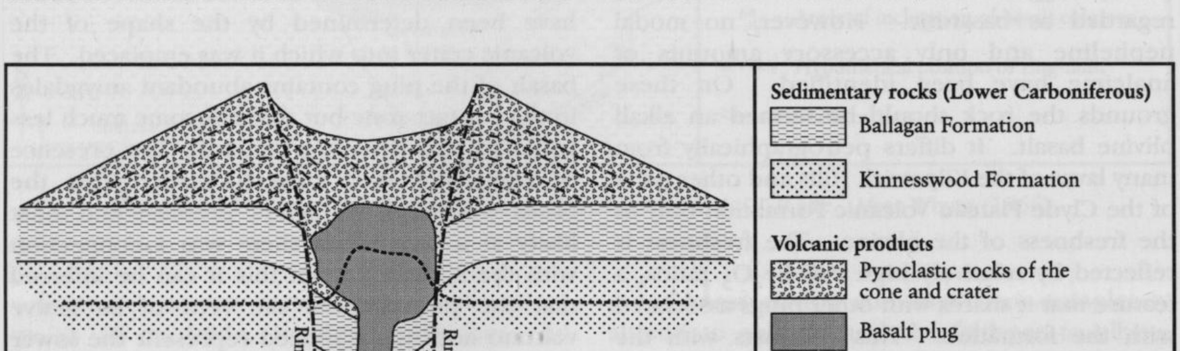


Figure 2.32 Diagrammatic cross-section illustrating possible structures associated with the Dumbarton Rock volcano. After Whyte (1966, fig. 4).

where in the Clyde Plateau Volcanic Formation. They also occur in the Paleocene lava sequence of the Isle of Skye, where they are represented by the Roineval GCR site in the *British Tertiary Volcanic Province* GCR Volume (Emeleus and Gyopari, 1992).

The Renfrewshire Hills succession is typical of the Clyde Plateau Volcanic Formation in that it comprises mainly lavas of a mildly alkaline to transitional, alkali-basalt series, dominated by olivine basalts, hawaiites and mugearites, but ranging locally through to trachyandesites, trachytes and rhyolites (Johnstone, 1965; Paterson *et al.*, 1990; Monro, 1999). The whole succession may be up to 1000 m thick in the north of the hills, of which up to 750 m are the Strathgryfe Lava Member. This member is characterized by feldspar-phyric flows ('Markle' type in the local classification), mostly of hawaiitic composition, and aphyric mugearites in approximately equal proportions; there are few basalts. The eruptions were entirely subaerial and the presence of reddened flow tops and lateritic beds are indicators of tropical weathering between eruptions. Pyroclastic rocks are rare in the main basalt–hawaiite–mugearite sequences and there are no features to suggest a central volcano. The lower parts of the lava pile are cut by numerous dykes with a predominant northeastward trend, particularly along the projected continuation of the Dumbarton–Fintry Line (see 'Introduction' to this chapter), which may mark the site of fissure eruptions.

A general description of the volcanic succession in the northern part of the Renfrewshire Hills is given in the British Geological Survey memoir (Paterson *et al.*, 1990), and many analyses from the district have been included in a geochemical and petrological study of Dinantian lavas of the Midland Valley by Smedley (1986a,b, 1988a). Subsequent to the description by Kennedy (1931), the composite lavas were included in a general study of composite bodies by Boyd (1974), who included 18 analyses from this GCR site. The radiometric age of the Clyde Plateau Volcanic Formation as a whole has been suggested as 335 Ma to 325 Ma, based upon K-Ar whole-rock and mineral dates of the freshest lavas and associated intrusions (De Souza, 1982). This is broadly compatible with its lithostratigraphical position (within the Strathclyde Group), which, in the absence of any reliable biostratigraphical data, suggests a Visean age.

Description

The area between Dunrod Hill and Greenock is a fault-bound block of lavas, separated from the main outcrop of the Clyde Plateau Volcanic Formation in the northern Renfrewshire Hills by the Largs Fault Zone. This structure is a major NNE-trending splay off the Highland Boundary Fault and has a complex history of movement, mainly prior to the eruption of the lavas, but with some post-lava movement (Paterson *et al.*, 1990). To the north-west, the lava outcrop is cut by the Spango Valley along the line of the NE-trending Inverkip faults, which are also probably related to the Highland Boundary Fault. The south-western flank of Dunrod Hill is controlled by the Dunrod Fault, juxtaposing the lava sequence against the stratigraphically lower Clyde Sandstone Formation which forms slightly lower hills to the south-west.

The name 'Dunrod Hill' was formerly applied to the whole of the hill above and to the north-east of Shielhill Glen (cf. Kennedy, 1931), whereas on modern maps it is restricted to the 298 m hilltop with a triangulation pillar (NS 240 726). The GCR site is centred upon the hilltop now known as 'Cauldron Hill' (NS 236 729) and is bound to the south-west and north-west by the aqueduct that takes water from Loch Thom to Greenock (Figure 2.33).

Within this area the volcanic succession dips gently to the north or NNE and consists entirely of hawaiite and mugearite lavas. The former are notably feldspar-macrophyric with a dark-purple matrix. The central parts of flows are generally massive and these tend to form low crags and the more obvious topographical features of the area. The mugearites are pale grey, fine grained and aphyric. Their exposures are characterized by closely spaced jointing broadly parallel to the flow surfaces, and some weathered surfaces have lines etched in a similar orientation. In thin section these planar features are seen to reflect an orientation of the groundmass feldspars, which is almost certainly due to flow foliation. Both types of lava have amygdaloidal zones, most notably, but not exclusively, at the top and bottom of the flow, where they are commonly associated with autobrecciation and hydrothermal alteration. Such zones weather more easily than the more massive central parts of flows and are exposed mainly in stream sections.

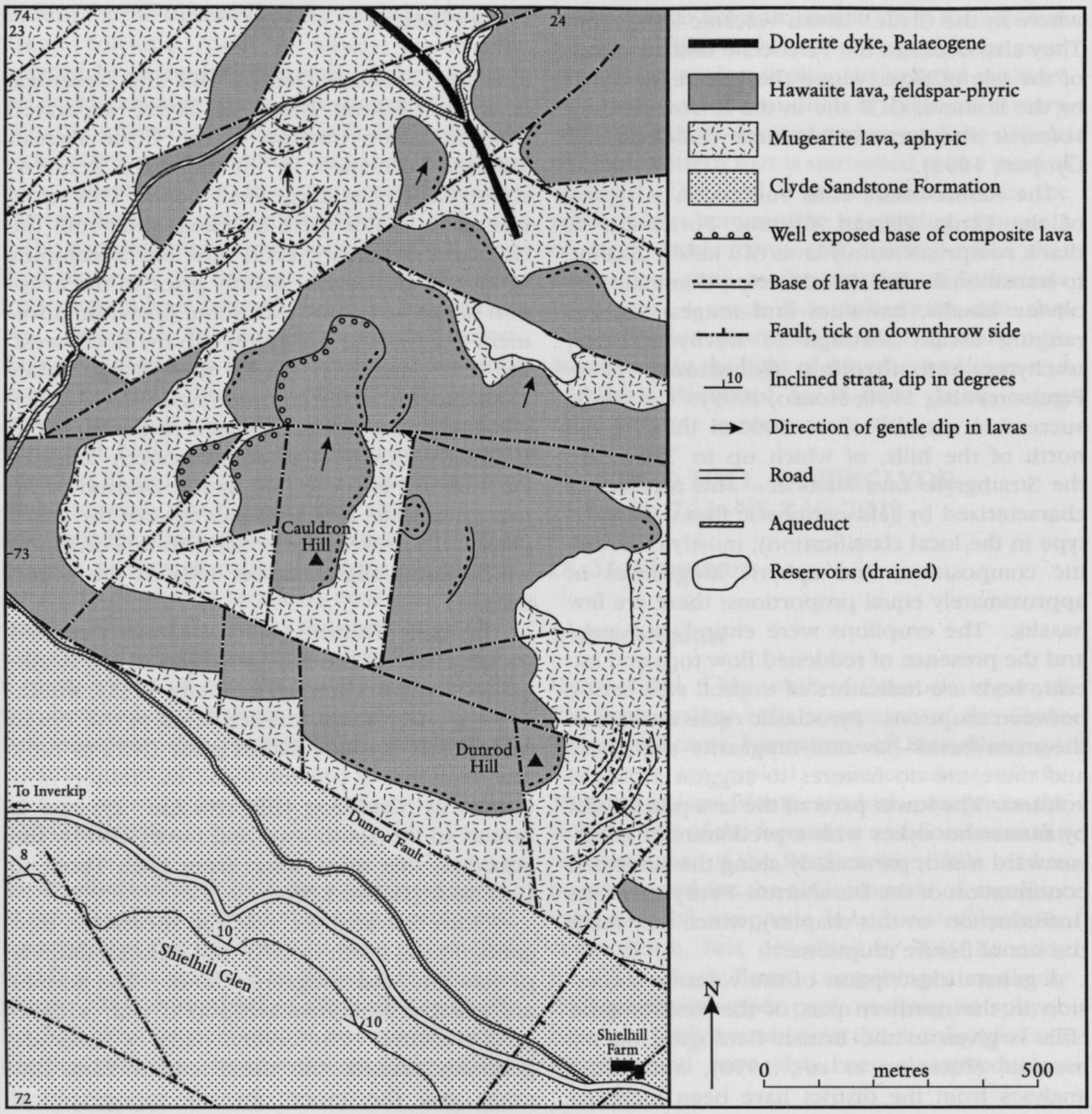


Figure 2.33 Map of the area around the Dunrod Hill GCR site. Based on British Geological Survey 1:10 000 Sheet NS 27 SW (1987).

The composite lavas that are the main feature of this site are best exposed high on the western flank of Cauldron Hill, where they form conspicuous crag features (Figure 2.34). Kennedy (1931) identified two such flows on Cauldron Hill and a further one around the headwaters of the Hole of Spango. The current 1:10 000 scale British Geological Survey map (NS 27 SW, 1987) shows far more faults than the original (Kennedy) mapping, so that the correlation of flows between fault blocks is less certain, but there are certainly at least two composite flows in this area. Similar flows also occur a short distance to the north of the site.

The lower part of each composite flow is aphyric and usually exhibits the platy jointing and foliation that is characteristic of mugearite flows. This is overlain by macroporphyritic lava with abundant (15 to 22%) feldspar phenocrysts up to 15 mm in diameter and rare (less than 1%) microphenocrysts of clinopyroxene and titanomagnetite. The relative proportions of the two

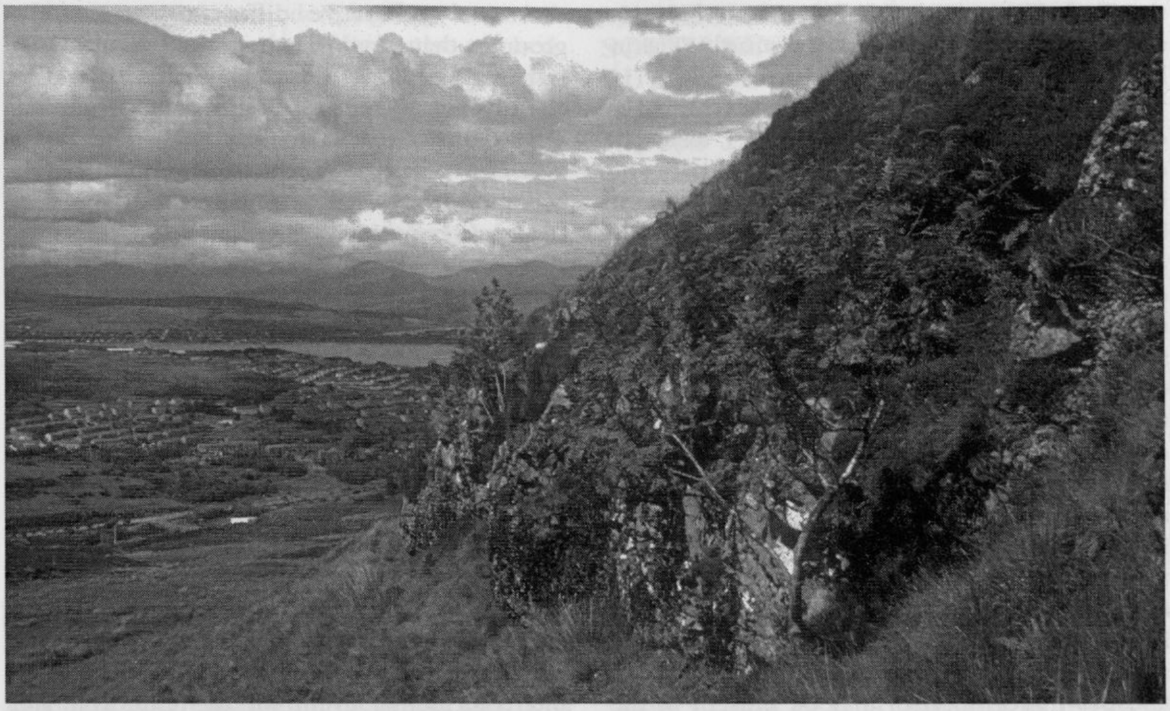


Figure 2.34 View from the north-west flank of Cauldron Hill, towards Gourock and the River Clyde. The low crag is typical of the composite hawaiitic lavas in the Dunrod Hill GCR site. (Photo: D. Stephenson.)

facies vary, but the flows are always asymmetrical with typical thicknesses of about 1 m for the lower facies and 5 m for the upper. The junction between the facies is gradational in places, but more commonly it occurs abruptly over a distance of only a few centimeters. There is no sharp contact and neither facies is chilled. The fine-grained groundmass of the upper facies is indistinguishable from the lower facies and seems to be in continuity with it. In some places, rare macrophenocrysts of feldspar, similar to those of the porphyritic facies, are observed in the upper 80 cm of the aphyric facies. The junction occurs within the massive, central part of the flow and there is never any intervening amygdaloidal or slaggy, brecciated zone, such as is usually seen at flow margins. In most exposures the junction is planar and parallel to the flow surface, with only minor irregularities. However, Kennedy (1931) recorded interfingering in places and cites several instances where one or other of the facies is pinched out. In most cases, this absence of one facies is based on correlation of flows between exposures and is difficult to substantiate. However, at one place the aphyric lower facies is seen to cut up through the porphyritic upper facies, with platy jointing parallel to the junction and clearly dipping at a higher angle than is usual (Kennedy, 1931, locality B; NS 2333 7316). Junctions with underlying and overlying flows, where seen, exhibit the slaggy, brecciated and amygdaloidal zones that are typical of the lava sequence and leave its extrusive nature in no doubt.

Almost all analyses from the composite flows (Kennedy, 1931; Boyd, 1974) fall within the field of hawaiite, whether in the classification based on normative composition (as used for Scottish Dinantian lavas by Macdonald, 1975; Smedley, 1986a; Paterson *et al.*, 1990; Monro, 1999), or in the TAS (total alkalis/silica) system based on oxide percentages (as favoured by Boyd, 1974, and the IUGS classification of Le Maitre, 2002). In the normative classification the more basic compositions could be classed as 'basaltic hawaiites'. Boyd (1974) analysed both whole-rock and groundmass from several samples, confirming field and petrographical observations that the groundmass of the porphyritic facies is very similar to analyses of aphyric rocks close to the junction of the facies. Slightly more fractionated aphyric rocks occur in the lowest parts of the flows, farthest from the junction; these fall just within the mugearite field in the TAS classification. In common with most lavas of the Clyde Plateau Volcanic Formation, the rocks are transitional alkaline and are mostly olivine-hypersthene-normative.

The petrography of the two facies has been described in detail by Kennedy (1931). Feldspar compositions are particularly instructive and have been studied by Boyd (1974). Despite the hawaiitic whole-rock compositions, the plagioclase macrophenocrysts in both facies are very calcic. In the porphyritic facies, complexly zoned cores of bytownite ranging from An_{78-68} are surrounded by normally zoned rims of labradorite, An_{70-58}. The rare macrophenocrysts in the top of the aphyric facies have compositions of An_{76-70}, identical to the cores in the porphyritic rock. Scattered microphenocrysts in the aphyric facies (An_{55-28}) are andesine, identical in composition to the groundmass feldspars of the porphyritic facies (An_{55-30}). Groundmass feldspars in the aphyric rocks are strongly zoned in the andesine range, An_{40-28}. Normative feldspar compositions suggest that significant alkali feldspar (?anorthoclase) may be present in the groundmass.

Interpretation

There can be little doubt that where the two rock-types at Dunrod Hill are juxtaposed as described above, they are parts of a single composite body. The massive, fresh exposures pass upwards and downwards into typically rubbly and amygdaloidal marginal zones, but the internal junction is near planar, undisturbed, unaltered and the aphyric facies seems to be continuous with the groundmass of the porphyritic facies. The two facies cannot have originated as separate flows. Although the transition from porphyritic to non-porphyritic is abrupt, there is no sign of a chill or any other manifestation of a sharp intrusive contact. Nor are any other sharp contacts observed within the bodies, so the possibility of either (or both) of the components having been intruded as sills into a pre-existing lava is unlikely.

The mechanisms whereby composite lava flows may be generated and preserved are more difficult to envisage than those responsible for composite intrusions, in which pulses of magma from either the same or from various sources are channelled up a common conduit. Kennedy (1931) discussed the possibility of some form of in-situ separation of crystals that were suspended in the magma on extrusion. He dismissed gravitative differentiation on the grounds that plagioclase crystals should sink, rather than float, in a magma of hawaiitic composition. He also reasoned that complete separation would be unlikely in the short time between eruption and the cooling magma becoming too viscous to allow movement of crystals. Separation due to liquid or viscous flow also seems unlikely to have produced such clear and complete separation without any sign of turbulent flow patterns, however slight, or of intermingling at the junction of facies.

Kennedy (1931) concluded that the differentiation must have occurred prior to extrusion and that the eruption involved two types of magma from 'separate bodies within the magma basin'. Although near-simultaneous eruption and intermingling of two distinct magmas can be inferred at Craigmarloch Wood (NS 345 719), from another composite flow in the Renfrewshire Hills described by Kennedy (1933), it does not seem necessary to invoke such a complex and coincidental event for the Cauldron Hill flows. The detailed geochemical and mineralogical data of Boyd (1974) confirm the impression gained from field relationships and petrography that the two rock-types are not only close in whole-rock composition, but also show evidence of a close genetic inter-relationship. The distribution of trace elements, particularly Sr, Ba and Rb, strongly suggests plagioclase fractionation, and numerical modelling is able to predict the observed compositional range simply by fractionation of the observed phenocryst phases. In this respect the Cauldron Hill composite flows differ from others studied by Boyd, which require more complex processes.

So, it is likely that the flows were erupted from magma chambers in which the crystallization and settling of plagioclase and, to a much lesser extent, clinopyroxene and titanomagnetite had resulted in zoning in terms of both mineral proportions and bulk magma composition. The upper, phenocryst-free portion was erupted first, followed by phenocryst-bearing magma, possibly as the phenocryst-free magma became exhausted. The continuous nature of the groundmass at the junction between the two phases suggests that full crystallization of the first pulse had not occurred when it was over-ridden by the second, possibly within a few hours by analogy with modern flows. Kennedy (1931) cited localities where the later pulse completely over-ran the earlier pulse to rest directly on the underlying flow, and other localities where only the earlier pulse reached. Such occurrences are highly likely, but are difficult to recognize and substantiate on the ground.

Given the limited variation in whole-rock composition seen in the composite flows and the inferred rapid sequential changes during the eruptions, it is probable that the magma chambers were small local developments, quite close to the surface. Maybe they were similar in form to compositionally zoned dykes that are exposed elsewhere in the world (e.g. South Greenland; Bridgwater and Harry, 1968). These in turn were probably fed from deeper magma chambers where the hawaiitic magmas were produced by higher pressure fractionation of mantle-derived alkali olivine basalts; the bytownite cores to the macrophenocrysts may be relics of this early stage. The compositional range throughout most of the Strathgryfe Lava Member is not much greater than that seen in the composite flows, which may therefore provide a model for magmatic differentiation in the whole lava field and possibly even for similar fields worldwide.

Conclusions

The lava sequence exposed in the area around Dunrod Hill is typical of the Strathgryfe Lava Member, which comprises by far the greatest part of the Visean Clyde Plateau Volcanic Formation in the Renfrewshire Hills. The member is characterized by a restricted range of lavas which are almost all either hawaiites, with phenocrysts (large crystals) of plagioclase feldspar, or slightly more evolved mugearites, which have no crystals visible to the naked eye.

On Cauldron Hill both rock-types can be seen, one above the other, in the same 'composite' lava flows. The relatively sharp but uninterrupted transition between the two rock-types and the close geochemical and mineralogical relationships between them suggest that they were emplaced in rapid succession as pulses of the same eruption. It is likely that the pulses tapped different levels of a near-surface magma chamber that had become compositionally zoned as some of the earliest minerals to crystallize (mainly feldspars) settled out. The Cauldron Hill composite flows are some of the best in Britain and have potential international importance for further studies on the evolution of magmas in high-level magma chambers that feed surface eruptions.

MACHRIHANISH COAST AND SOUTH KINTYRE, ARGYLL AND BUTE (NR 625 201–NR 640 208, NR 629 192–NR 651 182 and NR 688 171–NR 707 155)

I.T. Williamson

Introduction

The Dinantian volcanic rocks that crop out in the south of the Kintyre peninsula belong to the Clyde Plateau Volcanic Formation (Strathclyde Group), the most extensive lava succession within the Carboniferous–Permian Igneous Province of northern Britain. The succession is typical of continental lava plateaux formed by the accumulation of overlapping lava sequences. However, unlike the other lava fields that make up the Clyde Plateau Volcanic Formation the Kintyre sequence lies entirely north of the Highland Boundary Fault and, like the nearby Visean volcanic rocks at Ballycastle in Northern Ireland, is therefore structurally outside the graben of the Midland Valley. The Kintyre sequence is also separated by a considerable distance from coeval lava fields in central Scotland and it is not known if these were ever in physical continuity.

The Machrihanish Coast and South Kintyre GCR site (Figure 2.35) contains some lithologies that are either absent or poorly represented elsewhere within the Clyde Plateau Volcanic Formation. Low-potassium tholeiitic basalt lavas are known only from the Kintyre succession, and benmoreites and trachytes are under-represented elsewhere.

The earliest accounts of the igneous rocks of this district are rather sketchy and appear in the writings of John MacCulloch (1819); they were shown on his general geological map of Scotland in 1840 as a band extending from Campbeltown to the west coast. These igneous rocks figure briefly in papers by Nicol (1852) and Thomson (1865). Nicol divided them into 'Porphyries' and 'Augitic traps' and noted their association with the 'red sandstones and Carboniferous strata' of the district. However, it was Geikie (1897) who was the first to recognize that the Kintyre lava field forms an outlying portion of what he termed the Clyde Plateau.

The Kintyre area was first surveyed in detail by R.G. Symes and the map was published at the 1:63 360 scale by the Geological Survey

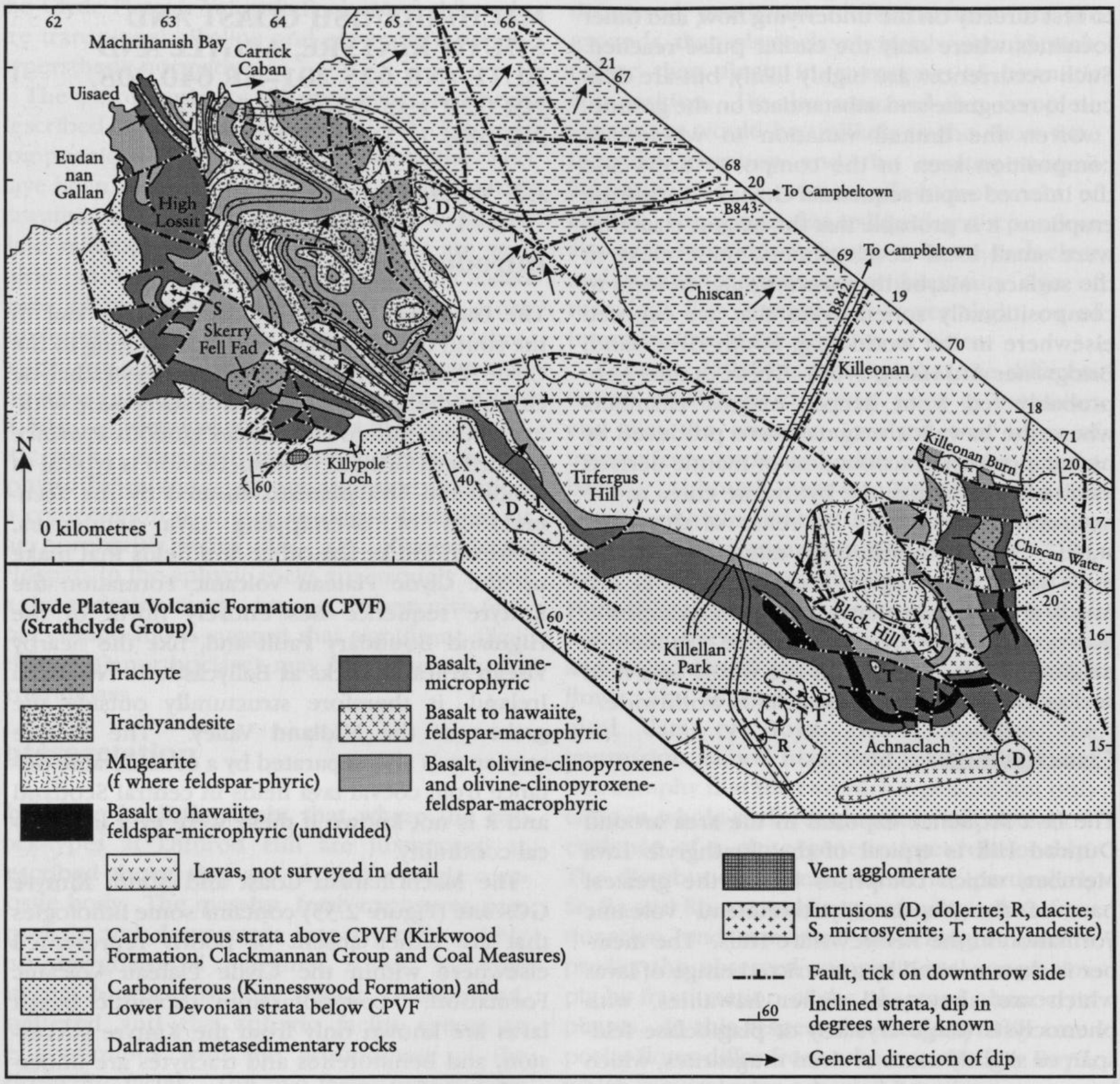


Figure 2.35 Map of the area around the Machrihanish Coast and South Kintyre GCR site. Based on British Geological Survey 1:50 000 Provisional Series Sheet 12, Campbeltown (1996).

in 1894. Details of the volcanic succession were not known until considerably later, when many of the area's pertinent features were described by McCallien (1927). The most recent mapping is by M.M. Avison and M.K. Carr in 1976, and H.M. Ayles and S.P. Duckworth in 1977, undergraduates at the University of Lancaster supervised by R. Macdonald. Their work has been incorporated in the 1:50 000 scale Provisional Series map of the British Geological Survey (Sheet 12, Campbeltown, 1996).

The Kintyre lavas have been featured in several wide-ranging studies of the geochemistry of Dinantian volcanism across the Scottish part of the Carboniferous–Permian Igneous Province of northern Britain by Macdonald (1975, 1980) and Smedley (1986a,b, 1988a) and consequently over 60 whole-rock analyses are available. They were also the main subject of a specific geochemical investigation by Smedley (1988b), highlighting differences in the mantle source of Dinantian magmas across the Highland Boundary Fault.

Description

There are five outcrops of the Clyde Plateau Volcanic Formation in south Kintyre. They comprise a broad belt extending south-eastwards from Machrihanish on the west coast to Tirfergus Hill and Killellan (the 'Machrihanish Lavas' of McCallien, 1927), three relatively small outliers north of Campbeltown, and an outlier on the south-east coast, east of Southend (the 'Southend Lavas' of McCallien, 1927 and Smedley, 1986a). The Machrihanish Coast and South Kintyre GCR site is a composite site made up of three areas selected from the first, and largest, of these outcrops (Figure 2.35). These are the coast section west and north of Machrihanish village (NR 640 208–NR 625 201), the area around the hill of Skerry Fell Fad (NR 638 183) and the area around Black Hill (NR 692 162).

In all three areas of the GCR site the Kintyre lavas overlie an unconformity. Along the western margin of the Machrihanish coast section and in the Chiscan Water on the eastern flank of Black Hill, the volcanic sequence rests upon the Kinnesswood Formation of the Inverclyde Group. This comprises white and red cross-bedded sandstones, and red-brown siltstones with nodules and beds of pedogenic limestone ('cornstone'). At Skerry Fell Fad the Kinnesswood Formation is only present in a small area in the west. Elsewhere, the lavas rest directly upon the Stonefield Schists, a unit of the Neoproterozoic Dalradian Supergroup.

The top of the Clyde Plateau Volcanic Formation in Kintyre is marked by a non-sequence, above which there is, in places, a reddish bauxitic deposit. This, the Kirkwood Formation, mostly comprises a diachronous lateritized, coarse- to fine-grained detrital volcaniclastic deposit that is interpreted as having formed by the contemporaneous weathering of the volcanic rocks. Elsewhere, the overlying rocks belong to the Lower Limestone Formation and Limestone Coal Formation of the Machrihanish Coalfield (McCallien and Anderson, 1930).

The Clyde Plateau Volcanic Formation in southern Kintyre comprises up to 400 m of volcanic rocks, predominantly lavas, dipping to the south-east (Figure 2.35). At some stratigraphical levels the lavas are complexly interdigitated, especially in the lower parts of the Black Hill section. Differences in the mineralogical composition and internal structure of the flows are exploited by weathering to produce a terraced landform ('trap topography') which is particularly well developed on Tirfergus Hill (NR 6645 1722), between the Skerry Fell Fad and Black Hill areas, and near Killellan. By contrast, the areas around Machrihanish and Skerry Fell Fad are characterized by much more rounded and lenticular landscape features, the difference resulting from variations in composition and morphology of the flows. The lava field is cut by a number of NW-trending faults.

The lower part of the lava succession is dominantly olivine basalt that is commonly macroporphyritic, along with some interbedded olivine-pyroxene-macrophyric (ankaramitic) basalt. The upper part contains most of the more evolved rock-types, including hawaiite, mugearite, benmoreite, trachyandesite and trachyte. Some of these lavas may be composite bodies. Several plugs, sills and dykes with compositions ranging from olivine-dolerite to microsyenite and dacite are also exposed (Macdonald, 1975).

Machrihanish coast

The coastal section from Machrihanish Bay westwards offers the most readily accessible part of the GCR site. Within it, a range of lithologies and structures typical of the Clyde Plateau Volcanic Formation may be examined.

A west to east traverse along the coast from Eudan nan Gallan (NR 6260 2027) to east of Carrick Caban (NR 6400 2085) passes up through the volcanic sequence of more than 20 lavas. The lower two-thirds of the sequence comprises basalt with various olivine, clinopyroxene and plagioclase phenocryst assemblages, interbedded with trachyandesites. The higher part of the sequence, east of Big Stone (NR 6350 2090), comprises up to seven flows of trachyte, mugearite and trachyandesite with a thin porphyritic basalt near the base. McCallien (1927) stated that 'tuffs and agglomerates are common associates of the lavas' in this section, but these lithologies are not shown on the latest British Geological Survey map.

Most of the lavas are tabular, sheet-like bodies. The wide range of characteristic lava-flow features exhibited include glassy (in places devitrified) flows, porphyritic variants including some with glomeroporphyritic textures, amygdaloidal and massive flow-units, blocky autobreccias, flow-base and flow-top breccias, ramp structures, and a variety of flow-induced structures. Some flows are remarkably fresh, but most are severely weathered or intensely altered. Olivine is almost always seen as pseudomorphs.

Skerry Fell Fad

The lower part of the volcanic succession also crops out on the slopes of Skerry Fell Fad (NR 6375 1820). The sequence includes olivine-microphyric basalts ('Dalmeny' type), olivine-clinopyroxene-feldspar-macrophyric basalts ('Dunsapie' type) and an olivine-clinopyroxene-macrophyric (ankaramitic) basalt ('Craiglockhart' type), succeeded by interbedded basalt and trachyandesite. The ankaramitic basalt is the same flow that occurs close to the base of the Clyde Plateau Volcanic Formation sequence at Eudan nan Gallan on the coastal section. The more evolved rock-types seen in the upper part of the succession in the coastal section are not present. However, a pale-weathered trachyte forms the summit of Skerry Fell Fad (Figure 2.36).

Apart from supplementing the range of lithologies and volcanic structures observed in the Machrihanish coastal section, this area exhibits the following features:

- The disconformable nature of the base of the Clyde Plateau Volcanic Formation may be demonstrated.
- The trachyte that caps Skerry Fell Fad has an unusual form. The body has both steep-sided and gently inclined basal contacts with at least two different basaltic lavas.
- Close to the base of the succession, south-west of Skerry Fell Fad, a single lava of tholeiitic (hypersthene-normative) basalt is interbedded with aphyric or microporphyritic basaltic lavas (Macdonald, 1975; Smedley, 1986a, 1988b). This basalt has the lowest K_2O content of any analysed Dinantian basalt and also has distinctive trace-element, rare-earth-element and isotopic compositions.
- Locally, there are minor intrusions and pyroclastic breccias, the latter possibly representing the sites of volcanic vents (Figure 2.35). Minor intrusions, mostly thought to be contemporaneous with the Dinantian volcanism, are not particularly widespread in the southern Kintyre lava field. Of note is a sheet, possibly a sill, of albite-phyric microsyenite exposed in a disused quarry (NR 6308 1907), 400 m SSW of High Lossit, which appears to have been intruded along the unconformity below the lavas. South-west

Figure 2.36 Skerry Fell Fad, Macrihanish Coast and South Kintyre GCR site. The summit rocks are trachyte, either infilling an earlier valley feature or possibly forming a lava dome. Note the terracettes due to soil creep on the steep slopes below the summit. (Photo: C. Bond.)

of Killypole Loch (NR 6415 1757), a small outcrop of igneous breccia, containing mostly basaltic clasts and mapped as 'vent' agglomerate, is associated with a small plug of olivine-dolerite. Its exact relationship to the rest of the lavas is unclear.

Black Hill

Black Hill, SSW of Campbeltown, is separated from the Skerry Fell Fad area by an extensive tract of relatively poorly exposed ground. The Clyde Plateau Volcanic Formation sequence around Black Hill is broadly similar to that in the other areas, with the proportion of lavas of more evolved composition generally increasing stratigraphically upwards. However, in contrast to the other areas, the lower part of the succession also includes some evolved flows. Trachyte near the base of the formation in the Chiscan Water (NR 7025 1680) about 1.25 km north-east of Black Hill, and benmoreite south of Black Hill (NR 691 163) may in fact be parts of the same flow (Smedley, 1986a). Also, within the higher units on Black Hill, there are aphyric and feldspar-phyric mugearite lavas.

Between Killellan Park (NR 6815 1640) and Killeonan Burn (NR 6970 1770–NR 7055 1740), there are lateral stratigraphical variations and other complex relationships between the lavas. The outcrop is much faulted and some lavas appear to be restricted to certain fault blocks. This is especially true of the sequences either side of the major NW-trending fault that runs from Carrick Caban on Machrihanish Bay (NR 6390 2080) to a point (NR 7045 1525) about 1 km east of Achnaclach (Figure 2.35).

Interpretation

The Kintyre sequence of the Clyde Plateau Volcanic Formation shows a broad, two-fold subdivision, with basic lithologies more common in the lower part and more evolved rocks dominating the upper part. Smaller scale variations in lithostratigraphy, such as those suggested by the different fault-block successions in the Black Hill area, may be explained in part by pre-existing topography. Lavas may have been channelled through low-lying areas between older flows and in graben-like structures between contemporaneous volcanic fault scarps. Some of the faults may be re-activations of older, deep-seated (?Caledonian) structures. Similar abrupt variations in local successions are also characteristic of the Lower Dinantian Birrenswark Volcanic Formation in the Anglo-Scottish Borders Region (see **Langholm–Newcastleton Hills** GCR site report). Small-scale fault-control of this type within the lava field may be common throughout the Clyde Plateau Volcanic Formation in the Midland Valley of Scotland.

The basal contact relationships of the trachytic body capping Skerry Fell Fad suggest that it is either a shallow intrusive body or that the underlying lavas had been eroded prior to its emplacement. Trachytic lavas are normally highly viscous and typically do not travel any great distance from their source. Hence, the trachyte could be interpreted as a lava dome capping its feeder pipe.

The south Kintyre lavas are typical of the transitional tholeiitic to mildly alkaline suite of Dinantian age in northern Britain. The sequence in the Macrihanish Coast and South Kintyre GCR site preserves some of the most basic (olivine-pyroxene-phyric basalt) and the most evolved rocks in the formation. Overall, this is a mildly silica-undersaturated and mildly sodic suite, which includes low Fe-variants and exhibits strong P_2O_5 and TiO_2 enrichment. Macdonald (1975) drew an important petrographical distinction between flows from the Kintyre and Campsie sequences of the Clyde Plateau Volcanic Formation, stating that, while in the Kintyre rocks clinopyroxene persisted as a phenocryst phase into the hawaiites, it is completely absent as phenocrysts from even the most basic Campsie lavas. Further sampling and analytical work by Macdonald (1980) and Smedley (1986a, 1988b) established that the Kintyre lavas may also be discriminated from many other Dinantian suites by geochemical parameters such as incompatible trace elements, rare-earth elements and isotopes. The additional data reported by Smedley (1988b) showed that the basaltic lavas from Kintyre are isotopically distinct from their counterparts in the Midland Valley and southern Scotland, having, in particular, lower Nd and higher Sr values. This correlates well with differences seen in the Siluro–Devonian calc-alkaline igneous rocks over the same terranes (e.g. Thirlwall, 1986) and implies long-term differences in the mantle source regions, which were more enriched in incompatible elements north-west of the Highland Boundary Fault.

Xenocrysts of quartz, surrounded by complex reaction rims of clinopyroxene, are a feature of some Kintyre basalts, and according to Smedley (1986a) represent tangible evidence that a degree of crustal assimilation may have occurred. However, isotopic evidence for crustal contamination has been detected in only one trachyte sample.

The low-potassium tholeiitic basalt flow within the lowest part of the Kintyre sequence is unique in the Clyde Plateau Volcanic Formation and its presence is fundamental to the overall understanding of the development of the Carboniferous magma types of the Carboniferous–Permian Igneous Province of northern Britain. Macdonald (1975) and Smedley (1986a) have both shown that it is relatively depleted of incompatible elements. This suggests that it was derived from a depleted mantle source, probably from the same type of spinel lherzolite as the other Clyde Plateau Volcanic Formation basalts, but by substantially greater degrees of partial melting. Contamination by crustal material is not thought to have been significant in this case.

Conclusions

Visean lavas in the Kintyre peninsula, represented by the Machrihanish Coast and South Kintyre GCR site, are the only significant occurrence of the widespread Clyde Plateau Volcanic Formation north of the Highland Boundary Fault. The 400 m-thick succession comprises at least 20 lavas mainly of olivine basalt, but more evolved rock compositions, including mugearite, benmoreite and trachyte, are more common in the upper part. Most of the lavas are extensive sheet-like bodies, but some are lenticular. Some appear to have flowed into contemporaneous topographical hollows, possibly small fault-bound grabens. A trachyte forming the summit of Skerry Fell Fad is a splendid example of a lava dome.

Geochemically, the Kintyre lavas are broadly similar to the transitional to mildly alkaline rocks that constitute the Clyde Plateau Volcanic Formation in the Midland Valley of Scotland, though with some subtle differences. In particular, a single lava of tholeiitic basalt with unusually low potassium content, near the base of the sequence on Skerry Fell Fad, is unique within the formation.

HEADS OF AYR, SOUTH AYRSHIRE (NS 279 183–NS 296 186)

I.T. Williamson and A.A. Monaghan (nee Sowerbutts)

Introduction

The Heads of Ayr, a prominent headland and coastal cliffs, is situated on the Ayrshire coast 5 km south-west of the town of Ayr (Figures 2.37a and 2.38). It is formed of a succession of volcaniclastic rocks and minor basalt intrusions, interpreted as a volcanic neck and believed to be related to the growth of a moderately sized volcano in Dinantian times. The rocks exposed at the GCR site allow the internal structure of the neck and its lithologies to be investigated in some detail. Some of the included rock fragments are probably derived from the mantle and hence provide an insight into the source region of the magma responsible for the volcanism.

Early, brief accounts of the geology of the area were by Geikie *et al.* (1869) and Geikie (1897). However, the first detailed observations were made by Tyrrell (1920) and these were added to by Eyles *et al.* (1929). Many of their observations were confirmed by Whyte (1963b) in his comprehensive study of the neck. The area has been re-described recently by Sowerbutts (1999) as part of a re-survey of the area by the British Geological Survey. The Heads of Ayr GCR site is frequently used for educational purposes and features in field guides to the region (e.g. Whyte in Lawson and Weedon, 1992).

Description

The volcaniclastic rocks at the Heads of Ayr are well exposed along 850 m of foreshore (NS 284 187–NS 295 188), on a wavecut platform and in the adjacent sea cliffs. The cliffs reach a maximum height of 75 m and form a double headland separated by a central embayment (Figures 2.37b and 2.38). The volcaniclastic rocks are juxtaposed against sedimentary rocks of the Dinantian Ballagan Formation (Inverclyde Group) (Sowerbutts, 1999). These generally gently dipping strata comprise interbedded micaceous sandstone, calcareous siltstone and dolomitic limestone ('cement-

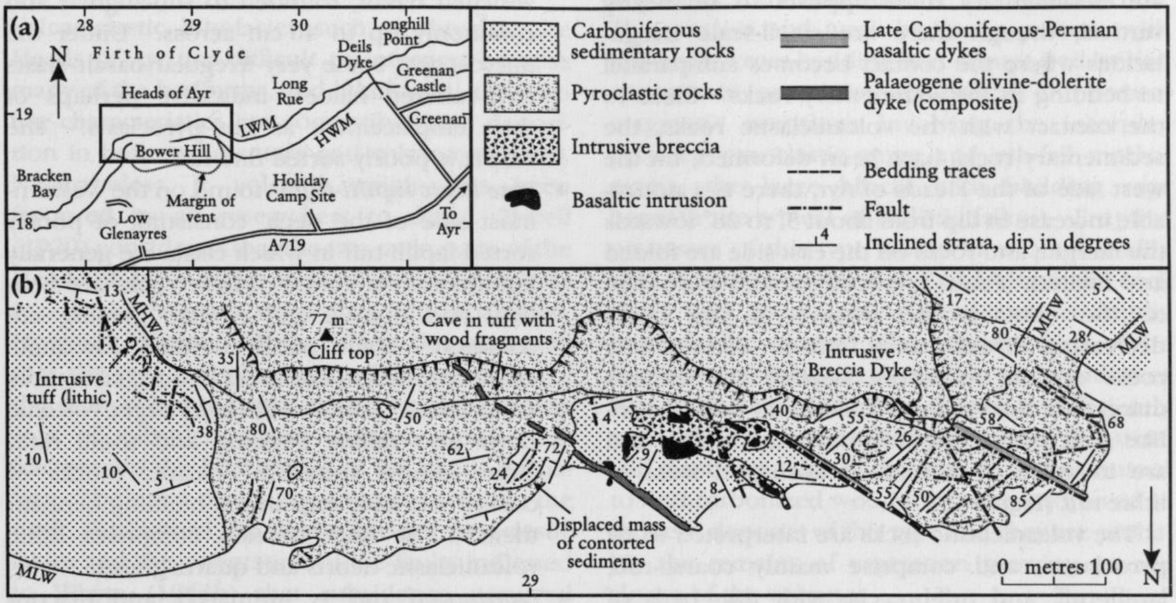


Figure 2.37 Geological map of the area around the Heads of Ayr GCR site. After Whyte (1964); and Lawson and Weedon (1992). Note the unconventional orientation of 2.37b (north at bottom) for easy comparison with Figure 2.38.

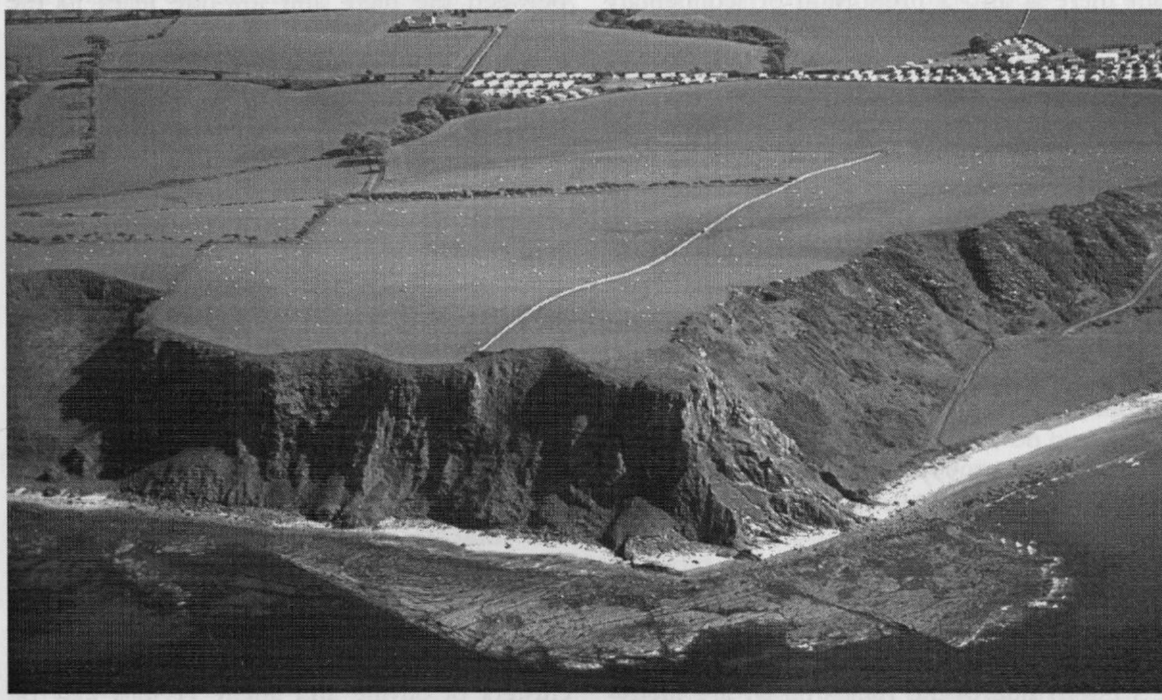


Figure 2.38 The West Cliff, at Heads of Ayr. Note the folded bedded tuffs within the Heads of Ayr Neck and the straight dykes on the wave-cut platform (compare with Figure 2.37b). (Photo: P. Macdonald.)

stone'). The contact between the volcaniclastic and sedimentary rocks appears to be nearly vertical, though there are small-scale irregularities where the contact becomes sub-parallel to bedding in the sedimentary rocks. Close to the contact with the volcaniclastic rocks, the sedimentary rocks have been deformed; on the west side of the Heads of Ayr, there is a noticeable increase in dip from about 5° to 28° towards the margin, and rocks on the east side are folded and faulted. Faults are both parallel and radial to the contact, the throws of the latter decreasing abruptly away from it. Sedimentary rocks on the foreshore beyond the eastern margins of the vent are folded into small basin-like structures a few tens of metres across and are intruded by a small boss-shaped mass of lithic tuff (Figure 2.37b).

The volcaniclastic rocks are interpreted to be pyroclastic, and comprise mainly coarse-tuff, lapilli-tuff, and tuff-breccia, with thin beds of fine-tuff. They vary in colour from green to blue-grey to purplish, generally reflecting the colour of the dominant basaltic clast-type and the degree of alteration of the finer-grained constituents. Most pyroclasts are igneous in origin, but there is also a proportion of sedimentary rock. According to Whyte (1963b), bedded tuff in the western part of the outcrop comprises 96.1% volcanic fragments, commonly cemented by carbonate, whereas rocks in the eastern part contain 97.6% volcanic clasts in a matrix of fine-tuff and crystals. There is a small population of dark-green clasts of serpentinized lherzolite up to 30 cm across. Other clasts include baked 'cementstone', 'cementstone' conglomerate and calcareous sandstone derived from the Ballagan Formation, Devonian pebbly and siliceous sandstones, and Lower Palaeozoic chert and siltstone. Also found are carbonized fragments of wood.

Sowerbutts (1999) identified several volcaniclastic lithofacies within the Heads of Ayr GCR site:

- *Coarse lithic lapilli-tuff and tuff-breccia* is most common and comprises massive and crudely bedded, poorly sorted rocks. On the western side of the Heads of Ayr are metre-thick massive beds of this lithofacies. On the eastern side, some larger slabs of disrupted Ballagan Formation rocks also occur, including one up to 20 m across (Whyte, 1963b). Clasts in this lithofacies are dominantly of vesicular basalt, rounded to sub-angular and commonly up to 40 cm across. Under the microscope, some very irregular basalt clasts have altered haloes, indicative perhaps of their emplacement as hot pyroclasts. The matrix is poorly sorted fine-tuff.
- *Fine lithic lapilli-tuff* is found on the westernmost side of the neck, consisting of poorly sorted lapilli-tuff in which clasts are generally less than 2 cm across. These rocks are bedded, and normal and reverse grading are common. Some poorly defined, low-angle erosional trough- and cross-bedded structures are present. Locally, strings of larger volcanic clasts also define bedding and some rare, large, almost spherical volcanic blocks of cobble size have sags in the bedding beneath them. The finer fraction comprises lithic volcaniclastic debris and quartz grains.
- *Laminated tuff* is intimately and discontinuously interbedded with the fine lithic lapilli-tuff. In places these beds contain weakly developed troughs and erosion surfaces.

Between the high- and low-tide mark in the central embayment at Heads of Ayr, Whyte (1963b) described mudstone and 'cementstone' country rocks that are cut by an irregular mass of intrusive breccia, and irregular small intrusions of basalt (Figure 2.37b). The outcrop of sedimentary strata and its intrusive mass is constrained between two NW-trending faults along which dykes have been emplaced *en échelon*. The western marginal dyke is an analcime-basalt considered to be Carboniferous in age, whereas the larger eastern dyke is a composite body with a marginal facies of tholeiitic basalt intruded by olivine-dolerite and probably of Palaeogene age.

Whyte (1963b) described the intrusive breccia as consisting mainly of angular clasts of igneous and much sedimentary rock in a highly chloritized groundmass. Cutting these rocks are small, irregular masses of olivine-augite-phyric 'monchiquitic basalt' (probably a basanite) with fresh nepheline in some samples; clusters of olivine and augite crystals may be derived from ultramafic xenoliths. The margins of these masses are fragmented and grade into the breccia, forming an igneous matrix to the clasts.

Interpretation

Volcaniclastic lithofacies such as those at the Heads of Ayr are difficult to interpret because many of the bedforms, and lithological and sorting characteristics are compatible with deposition in both sedimentary and volcanic regimes. Nevertheless, a volcanic origin has been favoured by all researchers to date. Tyrrell (1920) considered that the two main parts of the Heads of Ayr outcrop formed independently as two necks beneath vents that later coalesced. However, Whyte (1963b, 1968) demonstrated that the two units represent different facies of the same volcanic structure and emphasized the role of subsidence and collapse along ring-fractures in the later stages of the vent's formation. The steep dips in the country rock near the neck, the inward dip of the bedded pyroclastic facies and the orientation of the faults indicated to Whyte (1963b) that subsidence occurred along ring-fractures when magma was withdrawn from beneath the vent. He suggested that the layering in the tuffs could be interpreted as being due to either normal sedimentary processes or fluidization, especially at the neck margins and close to minor intrusions. He also indicated that a fluidization model could explain the emplacement of the intrusive lithic tuff east of the neck.

According to Whyte (1963b), the Heads of Ayr volcano may have reached a height of 600–900 m with a diameter at its base of about 3 km. He proposed the following five-stage model for its development:

1. Neck emplacement and deformation of adjacent sediments (brecciation and upturning) beneath a volcanic vent. A tephra cone was constructed, but was rapidly degraded and re-deposited in a subsiding marginal basin.
2. A larger volcanic cone grew, but subsidence in the basin continued.
3. Subsidence continued with the lateral spread of the volcanic cone and subaerial deposits. There was some deformation of bedded tuffs due to subsidence.
4. Vent collapse and further deformation of the bedded tuffs.
5. Further faulting and intrusion of basaltic magma, with marginal brecciation to produce intrusive breccias and tuffs; followed by regional tilting.

Sowerbutts (1999) concurred that volcanic processes were dominant at the Heads of Ayr. The massive and weak bedforms present with low-angle cross-bedding and abrupt bed-by-bed changes in grain size are typical of phreatomagmatic eruptions involving the interplay between pyroclastic surge and ash-fall mechanisms; the large blocks with bedding sags beneath are typical of ballistic fallout. Eruptive processes of this type would have been readily able to gather sedimentary and ultramafic rock clasts and quartz grains from various depths beneath the volcano. The coherent form of the fragments of Ballagan Formation rocks indicates that at the time of eruption they were sufficiently well lithified to resist disaggregation, but were still plastic enough to deform. It is not unusual to find carbonized wood fragments preserved in modern deposits of this type and they are probably the remains of vegetation growing on the slopes of the volcano.

However, it is unlikely that the volcanic rocks at the Heads of Ayr are the relics of a tephra cone. Such structures have very low potential for preservation within the geological record. The steep, cross-cutting contacts seen at the Heads of Ayr suggest that these rocks represent levels beneath the substrate, probably within the volcanic neck of a tuff-ring. In such a dynamic volcanic environment, collapse of the vent rim along faults and slumping of blocks of strata into the crater are typical features, and localized erosion, mixing with newly fragmented magma and re-deposition of the pyroclastic sediment is commonplace (cf. Kokelaar, 1983). The diameter of modern maar craters, which are regarded as having similar dimensions to tuff-rings, varies up to 3 km with the mode around 800 m (Cas and Wright, 1987). At about 850 m the diameter of the Heads of Ayr structure is consistent with this. Thus, based on both lithofacies and size, the Heads of Ayr Neck may be considered as the concealed part of a substantial tuff-ring. Similar structures have also been deduced from necks in the eastern Midland Valley (see **East Fife Coast** and **North Berwick Coast** GCR site reports).

Though a Dinantian age has long been assumed for the volcanic rocks, the strong affinities between the 'monchiquitic basalt' intrusions that cut them and Permian volcanism nearby have confused discussions on the age of the neck from the earliest investigations in this area.

Recent palynological studies on the sedimentary rocks of the Ballagan and Lawmuir formations exposed a short distance east of the Heads of Ayr have thrown new light on this debate (M.H. Stephenson, 2000). Separating these two formations are volcaniclastic rocks, named the 'Greenan Castle Member', which are considered to be associated with the Heads of Ayr Neck. Evidence for the correlation of these two volcanic outcrops includes the lithofacies similarity between the Greenan Castle Member and the lowest tuffs in the western part of the Heads of Ayr, cited by Whyte (1963b). Moreover, the most common clast in the member is a greenish basalt that is distinctly different to basalts in the underlying Devonian strata, and therefore unlikely to have been derived from those earlier volcanic beds. Beds below and above the Greenan Castle Member contain a palynological assemblage indicative of a latest Tournasian and early Asbian age respectively. If the Greenan Castle Member is coeval with the volcanism at the Heads of Ayr, this gives a constrained age for the volcanism within the mid-Dinantian, and for the first time confirms correlation with the Clyde Plateau Volcanic Formation.

Thus, a scenario can be envisaged where, early in the development of the Dinantian lava fields in the Midland Valley of Scotland, magmas rising through the near-surface crust would probably have encountered groundwater, wet, unconsolidated sediments and standing bodies of water. Interaction between these and magma would have resulted in intense phreatomagmatic activity, building cones along the line of the controlling fissures. Apart from the Greenan Castle Member there are no other tuffs, lavas nor volcaniclastic sedimentary rocks in the area known to have been derived from the Heads of Ayr vent, supporting the view that such volcanoes in Dinantian times were generally short-lived structures.

Megacryst and ultramafic xenolith assemblages within fragmental deposits, lavas and intrusions are a common feature of alkali basalts across the Midland Valley of Scotland (see **East Fife Coast**, **North Berwick Coast** and **Dubh Loch** GCR site reports). Such assemblages provide valuable clues to magmatic processes and the nature of the crust and mantle beneath the Carboniferous–Permian Igneous Province of northern Britain. The discovery of some ultramafic xenoliths in the Heads of Ayr Neck, and megacrysts in the associated basaltic intrusions, lends added importance to the site (see Chapter 1).

Conclusions

The superb coastal exposures in the Heads of Ayr GCR site beautifully demonstrate the rock-types and internal three-dimensional structure of a major volcanic neck of latest Tournaisian to Visean age and its relationships with the surrounding strata. This neck was emplaced through sedimentary rocks of the Ballagan Formation (Inverclyde Group) and is believed to correlate with volcaniclastic rocks that occur between the Ballagan and Lawmuir formations a little to the east of the GCR site. It is therefore thought to be contemporaneous with the western part of the extensive Clyde Plateau lava field.

The poor size-sorting of fragments, the presence of volcanic bombs (ejected in a molten or plastic state) and the general characteristics of bedding within the neck are indicators of the type of eruption. Together they suggest an explosive (phreatomagmatic) eruption due to the interaction of basaltic magma with water-saturated sediment and/or bodies of water, giving rise to a mixture of pyroclastic surge deposits from the lateral blast and ash-fall deposits that settled out of the ash-cloud. At the time of the eruptions the Ballagan Formation substrate was sufficiently lithified to be fragmented and incorporated in the ejected material and yet still plastic enough to deform around the margin of the neck. Large slumped blocks within the neck testify to the collapse of an overlying vent structure. The Heads of Ayr Neck is thought to have underlain a tuff-ring of substantial size, but there is no evidence to suggest that large volumes of lava were erupted.

Some of the rocks in the neck contain sparse inclusions of rocks and individual crystals that are thought to have been transported from deep within the Earth's crust or from the source region of the magma in the underlying mantle.