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Silicic volcanism in the Scottish Lower Carboniferous; Lavas, intrusions and ignimbrites of the Garleton Hills Volcanic Formation, SE Scotland.

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Extensional tectonics and incipient rifting, on the north side of the Iapetus suture were associated with eruption of (mainly) mildly alkaline olivine basalts. Initially in the Tournaisian (Southern Uplands Terrane), magmatic activity migrated northwards producing the Garleton Hills Volcanic Formation (GHVF) across an anomalous sector of the Southern Uplands. The latter was followed by resumption of volcanism in the Midland Valley Terrane, yielding the Arthur's Seat Volcanic Formation. Later larger-scale activity generated the Clyde Plateau Volcanic Formation (CPVF) and the Kintyre lavas on the Grampian Highlands Terrane. Comparable volcanic successions occur in Limerick, Ireland. This short-lived (c. 30 Myr) phase was unique in the magmatic history of the Phanerozoic of the British Isles in which mildly alkaline basaltic magmatism locally led to trachytic differentiates. The Bangly Member of the GHVF represents the largest area occupied by such silicic rocks. The most widespread lavas and intrusions are silica-saturated /over-saturated trachytes for which new whole-rock and isotopic data are presented. Previously un-recognised ignimbrites are described. Sparse data from the fiamme suggests that the magma responsible for the repetitive ignimbrite eruptions was a highly fluid rhyolite. The Bangly Member probably represents the remains of a central-type volcano the details of which are enigmatic.

Introduction

Following cessation of Iapetus subduction a period of continental rift-related extension in the Laurussian continent north of the Variscan front (Millward & Williamson 2003) gave rise to volcanism in the first thirty million years of the Carboniferous (Tournaisian and Visean). Whilst there were some tholeiitic basalts, most were 'mildly alkaline' or 'transitional olivine basalts'. Relatively small differences in geochemistry between the different volcanic provinces are ascribed to variations in degree of melting and in composition of mantle sources. Overall there is a generalised similarity in the volcanic products with minimal crustal contamination. During intermittent lithospheric ascent, fractional crystallisation commonly led to hawaiitic and mugearitic differentiates. This proceeded further in some provinces to yield silicic residua (trachytes, phonolitic trachytes and rhyolites (Macdonald 1975; Smedley 1986a; Kirstein *et al.* 2006). The sequencing of transitional olivine basalt, hawaiite, mugearite lends a broad commonality to these volcanic provinces that distinguishes the early Carboniferous (Tournaisian and early Visean) as a unique phase in the magmatic history of the British Isles.

This phenomenon occurred across a wide area from the Limerick basin in SW Ireland to the Southern Uplands, Midland Valley and the Scottish Highlands (Fig.1). Comparable activity, generally on a smaller-scale, occurred in England and several other parts of Ireland. Basaltic magmatism was persistent in Scotland through the later Carboniferous and Permian (with a major exceptional tholeiitic episode towards the close of the Carboniferous), and generally involved more silica-deficient magmas representing smaller mantle melt-fractions than those of the Tournaisian and early Visean (Kirstein *et al.* 2006; Read *et al.* 2002; Wallis 1989). These more alkalic basalts were generated in volumes presumed to have been too small to give rise to silicic residues.

This paper focusses on the Garleton Hills Volcanic Formation of East Lothian (SE Scotland) and its late-stage products and shows their relationship to the other early Carboniferous trachytic occurrences in Scotland and south-west Ireland.

Chronology of magmatism

The earliest extensional effects gave rise to small-scale continental rifting in the Southern Uplands, accompanied by eruption of the Birrenswark lavas across a

zone approximately coinciding with the suture resulting from Iapetus closure (Leeder 1974; Leeder *et al.* 1989; Read *et al.* 2002: Fig.1). The Birrenswark lavas (361 ± 7 Ma, De Souza 1982) were associated with extensional faulting and subsidence of the Northumberland-Solway Basin (Millward & Williamson 2003) and were probably pene-contemporaneous with the Kelso lavas further north-east (Fig.1). Most of the broad arc of at least fifty 'necks', roughly colinear with the NE-SW to NNE-SSW-trending Birrenswark and Kelso lava outcrops (Fig.1) is also considered broadly contemporaneous with the lavas. Petrologically these necks have received little study but include some trachytes, together with (rare) peralkaline rhyolites and phonolitic trachytes (McRobert 1920; Lumsden *et al.* 1967). Among them is the rather younger Eildon Hills centre that has been precisely dated by $^{40}\text{Ar}/^{39}\text{Ar}$ at 352 ± 1.4 Ma (Monaghan & Pringle 2004). Collectively these centres may have contributed to a lava field coalescing with the Birrenswark and Kelso lava fields (Read *et al.* 2002). Small-scale volcanism succeeded the latter in the Scottish Borders, yielding the Glencartholm and Kershopefoot basalts and pyroclastic deposits (Lumsden *et al.* 1967: Fig.1) that may have overlapped in time with the Garleton Hills Volcanic Formation (GHVF).

The GHVF volcanism took place approximately 50 km north of the Kelso lavas on a fault block between the Lammermuir fault and a postulated NE-SW trending fault that may be an off-shore extension of the Southern Upland Fault (Read *et al.* 2002) controlling the coast-line between Port Seton and Eyebroughy (Fig. 2). The GHVF magmatism presents the next largest outcrop of Lower Carboniferous extrusive rocks in Britain and Ireland after the Clyde Plateau Volcanic Formation (Fig.1). The Bangly Member is the youngest component of the GHVF. Despite current poor exposure the Bangly Member is considered to have extended over a wide area (at least 20 km across) and may have been the most voluminous of all the Lower Carboniferous silicic suites. Two precise $^{40}\text{Ar}/^{39}\text{Ar}$ dates (342.4 ± 1.1 and 342 ± 1.3 Ma) from its trachytes show the GHVF to be early Visean (Monaghan & Pringle 2004). Following production of the GHVF, renewal of activity in the Edinburgh district gave the Arthur's Seat Volcanic Formation at 341-335 Ma (Monaghan *et al.* 2014). The Arthur's Seat Formation and the GHVF are both considerably earlier than the voluminous outpourings of the Clyde Plateau Volcanic Formation (**Fig.1**).

Distribution of Dinantian trachytes in Scotland and Ireland

There are occurrences of silicic rocks in the Southern Uplands Terrane of Tournaisian age: Skelfhill Pen (McRoberts 1920) comprises both silica-saturated and undersaturated trachytes. It is just one amongst the dozens of igneous centres among the scattered arc of volcanic sites linking the Birrenswark and Kelso lava fields, most of which have not been investigated since the pioneering investigations of Rachel McRoberts a hundred years ago. Near Melrose, the Eildon Hills (**Fig.1**) comprise a near-surface suite of sill-like bodies varying from augite olivine trachytes to alkali rhyolitic varieties (McRobert 1914). These are also Tournaisian, dated ($^{40}\text{Ar}/^{39}\text{Ar}$) at 352 Ma (Monaghan & Parrish 2006). Adjacent to them is the large (c. 800 x 400 m) Chiefswood vent (Eckford & Manson 1928; 1992), probably of phreatomagmatic origin. It is filled with sedimentary country-rock fragments together with basaltic, trachytic and rhyolitic clasts. The low Zr/Nb and high Ce/Y values for the quartz porphyry clasts indicate comenditic compositions (unpublished data). The more silicic Eildon sills contain riebeckite (Eckford & Manson 1928) and the evolution of the whole Eildon suite was clearly towards peralkaline residues.

The Glencartholm volcanics of Visean age represent a rejuvenation of magmatism in the Borders (**Fig.1**). Trachyte clasts in the pyroclastic necks were described by Elliot in Lumsden *et al.* 1967. To the north of all these, between the Lammermuir Fault and the extrapolation of the Southern Upland (and/or Crossgate Hall Fault) lies the GHVF on or close to the boundary between the Southern Uplands and the Midland Valley terranes. The Arthur's Seat volcanics, to its west, range from basalt to mugearite but did not attain trachytic composition. Further to the west is the large chevron-shaped outcrop of the CPVF, dated from 335 ± 2 to 329 ± 1.4 Ma (Monaghan & Pringle 2004; Monaghan & Parrish 2006; Monaghan *et al.* 2016), across which extensive fractionation led to trachytic and rhyolitic residua at several centres (Whyte & MacDonald, 1974; MacDonald & Whyte, 1981; Patterson *et al.*, 1990; Monro, 1999). To the west of the main outcrops of the CPVF trachyte dykes, bosses and sheets occur on the island of Great Cumbrae in the Firth of Clyde (**Fig.1**): the broad ENE-WSW trending trachyte dykes may be westwards extensions of those of the Misty Law complex in the CPVF. A short distance west of the Cumbrae islands are three trachyte lava flows in southern Bute (Gun *et al.* 1903; Tyrrell 1918). Whilst the Arthur's Seat and CPVF were unequivocally erupted through the Midland Valley Terrane, the Kintyre lavas are the only volcanic

representatives of lower Carboniferous volcanism north of the Highland Boundary Fault (**Fig.1**). Among the Kintyre lavas, the basalts at Skerry Fell Fad are crowned by trachyte, locally marking the close of activity (Williamson 2003a).

Early Carboniferous trachytes also crop out in the Limerick district of SW Ireland (**Fig.1**). Although lying at over 400 km distance from the Birrenswark basalts of the Southern Uplands, they are approximately along strike from the latter. The Limerick lavas are mainly Visean although some may be Tournaisian (Strogen 1977; Sevastopulo & Wyse-Jackson 2001). A lava succession at Carrigogunnel in the Limerick graben is analogous to that of the GHVF in that a basaltic volcano built up from the sea-floor to form a sub-aerial edifice, culminating in trachytic eruptions (Strogen 1973) and supports the concept that the trachyte magmas were the culminating fractional crystallisation residuals from a mafic parent. The Scottish volcanism occurred to the NW side of the Iapetus suture whereas their Limerick counterparts lay on the SE side, confirming that the differences of the Laurentian and Avalonian lithospheric plates composing the (relatively) new Laurussian Continent were irrelevant to the petrogenesis of the magmas (Macdonald and Walker 1985).

Garleton Hills Volcanic Formation

The eruptions that created the GHVF occurred on an anomalous fault-bounded area (East Lothian), bounded to the SE by the Lammermuir Fault and to the NW by another major fault regarded as the north-eastward extrapolation of the Southern Upland fault (or the related Crossgatehall Fault). This is hypothesized to lie close off-shore, parallel to the coast-line between Port Seton and Eyebroughy (Max 1976; McKerrow & Elders 1989; Floyd 1994: Fig.1). This fault block is structurally complex as the Southern Uplands Terrane is thought to be over-thrust onto the southern margin of the Midland Valley Terrane (Badenszki *et al.* 2009; Badenszki *et al.* 2019). Accordingly, while generally regarded as part of the Scottish Midland Valley Terrane, this fault block should, more properly, be regarded as a marginal part of the Southern Uplands.

Back-arc extensional stresses generated by Variscan plate motions, far to the south, are assumed to have spread northwards at the start of the Carboniferous causing (a) the rift-related volcanism in the Borders before generating (b) the

Garleton Hills Volcanic Formation (GHVF), (c) the Arthur's Seat Volcanic Formation and then (d) the Clyde Plateau Volcanic Formation (CPVF). While there is compositional similarity between the GHVF and the Arthur's Seat Volcanic Formation, some 20 km further west (**Fig.1**), differentiation in the latter did not extend beyond the mugearite stage (Macdonald 1975; Smedley 1986a). Despite their petrological affinities and geographical proximity, recent dating has shown that the two Formations, while relatively close in time, were not contemporaneous (Monaghan *et al.* 2014). Both lie centrally within the broad ENE-WSW zone defined by the Pentland Fault to the NW and the Lamermuir Fault to the SE and both are approximately equidistant from the extrapolation of the Southern Upland Fault (**Fig.1**). Consecutive transtensional motions on the latter (Floyd 1994) may have triggered the volcanism of both Formations. Comparable short-lived tectono-magmatic phases have been recognised in the Northumberland-Solway and Tweed Basins (Monaghan & Pringle 2004).

The GHVF commenced with shallow-water Surtseyan basaltic eruptions but, with an increase in magma production, developed into a sub-aerial, multi-vent lava field comprising transitional olivine basalts, hawaiites, mugearites and benmoreites before culminating in eruptions of trachyte lavas and pyroclastic rocks that compose the Bangly Member. Macdonald (1975) noted that among the Scottish Dinantian volcanic occurrences 'the lines of liquid descent' showed considerable variation. The GHVF for example has a distinctly more potassic flavour, as had been remarked previously by Bailey in Clough *et al.* (1910). The total thickness of the GHVF is 520 m of which the Bangly Member contributes up to 160 m (Williamson 2003a; McAdam & Tulloch 1985). Covering an area of c.40 km², the Bangly Member forms the largest outcrop of Carboniferous silicic rocks in the British Isles.

The late-stage silicic products of the GHVF: The Bangly Member.

This paper considers the differentiated rocks composing the Bangly Member that were erupted in the closing stages of the GHVF. Their relative neglect reflects the sparsity of outcrops away from the high-ground of the Garleton Hills (**Fig.2**). The GHVF outcrop area is mainly low-lying farmland (< 50 m above sea-level). Although exposures are scarce the land rises abruptly some 11 km south of the coast to form an ENE-WSW ridge, the Garleton Hills. These, rising to nearly 190 m high, provide the majority of exposures. When mapped by the Geological Survey in the late 19th

century there were numerous small quarries and railway-cuttings, now mostly infilled. The survey authors considered that some outcrops represented intrusions (mainly sills) as in the former quarry (Craig's Quarry) close to Dirleton (**Fig.2**). Clough *et al.* (1910) considered that some of the other inland exposures are also of intrusions rather than lavas. One prominent outcrop at Dirleton castle is of micro-phyric trachyte, utilised as the site for the mediaeval castle at Dirleton [NT 518 839] about 8 km north of the coast. The homogeneous rock-face, c.10 m high, has crude columnar jointing. It is not obviously extrusive although some trachyte blocks in the castle walls are strongly vesicular. The latter may have come from a nearby quarry, now infilled and the question as to whether the castle is built on a thick lava or an intrusive sheet remains open. An overview of the Bangly Member is given by McAdam & Tulloch (1985) and the geology of the Garleton Hills themselves was described by Williamson (2003b).

There are several thin (poorly exposed) pyroclastic horizons between the lavas referred to by McAdam & Tulloch (1985) as brown/mauve trachytic tuff with coarse trachyte clasts. With the lack of any further description we assume that these were typical fall deposits. However, along the south coast of the Forth estuary, there is an intermittent stretch (c. 500 m) of sea-cliffs composed of what have been simply described as 'welded tuffs' (Upton 1982), 'bedded tuffs' (McAdam & Tulloch 1985) and 'welded air-fall' (Upton 2003). These, referred to here as the Marine Villa tuffs, are exposed in the cliffs and their adjacent wave-cut platform (**Fig.3**).

The lavas of the Bangly Member dip gently to the WSW beneath lower Visean sediments with which they are essentially concordant. In the western and southern areas some exposures of the contact of the trachytes with the younger sediments have been found in the past. One such exposure is described as a conglomeratic rock with rounded trachytic pebbles reflecting late-stage erosion of the extinct volcano prior to its cover by shallow-water Visean sediments (Clough *et al.* 1910). South of the Garleton Hills the trachyte outcrops are truncated by the Alderston Fault. A highly condensed succession including some trachytes crops out in the graben structure between the Dunbar-Gifford and Lammermuir Faults but the Ordovician escarpment of the Lammermuir Fault probably formed the south-eastern barrier to the whole formation (**Fig.2**). A severely altered lava crops out on the coast, c.10 km to the north of the Garleton Hills, in the vicinity of Weaklaw [NT 499.858] (Upton *et al.* 2019) directly overlying the Marine Villa welded tuffs (**Fig.3**). Close-

offshore, and un-affected by the late-stage metasomatism that affected this lava, lies the island of Eyebroughy, composed of trachyte and provisionally regarded as a sill (McAdam & Tulloch 1985). Along the mainland coast to the east on the outskirts of North Berwick a faulted patch of silicified trachytic tuffs is mentioned by McAdam & Tulloch (1985).

The original outcrop extent of the Bangly Member remains indeterminate since its eastern margin has been eroded, its western extensions are overlain by younger strata, its northern limit (beneath the Firth of Forth) is unknown and to the south it is truncated by the Alderston Fault. To the south-west, lavas and tuffs were penetrated by a Geological Survey boring at Spilmersford (7.5 km SW of Haddington, **Fig.2**: McAdam & Tulloch 1985). However, analyses of sixteen of these lavas (Livingstone & McKissock 1974) show them all to be relatively primitive and not correlatable with the Garleton Hills trachytes.

The thickness of the Bangly Member appears to decrease markedly towards the north and south and the coincidence of the highest elevations of the Garleton Hills and thickest part of the succession suggests that the eruptive focus probably lay in this area. The hills are composed of several thick (up to 20 m) lavas (**Fig. 3**). The lavas are conformable with their under- and over-lying strata and are accordingly inferred to have had essentially horizontal surfaces when extruded and were potentially ponded in a caldera. Whilst poverty of exposure makes this impossible to verify possible modern analogues include, for example, Menengai in the Kenyan rift (Leat 1984).

Whereas the more primitive lavas preceding the trachytes are inferred to have erupted from relatively small volcanoes now represented by various plugs in East Lothian, the site or sites through which the Bangly Member magmas were erupted has received little attention.

The transition from mafic to felsic magmatism in the GHVF appears to have been not wholly abrupt since a feldspar-phyric 'Markle Type' basalt (*i.e.* hawaiitic) flow was described as occurring "some little way up the trachytic group" whilst "one band" (*sic*) of mugearite occurs half-way up the trachytic series (Clough *et al.* 1910). The critical outcrops, however, are no longer exposed. In the northernmost outcrops, in the vicinity of Weaklaw [NT 499 858], welded ignimbrites (the Marine Villa tuffs described below) overlie a mugearite flow. While these pyroclastic rocks have generally been accepted as locally representing the base of the Bangly Member they are themselves overlain by a porphyritic mugearite lava, exposed in sea-cliffs. The outcrop of this relatively inaccessible exposure has not previously been recorded and is further evidence that the change from the mafic early products of the GHVF succession to the late-stage silicic magmatism was more complex than formerly recognised.

The Bangly Member trachytes are silica-saturated or silica-oversaturated although there are several silica-undersaturated (phonolitic) intrusions in the vicinity. The most prominent and best studied are the plugs that form Bass Rock (an island east of North Berwick), North Berwick Law (c.1 km south of North Berwick) and the laccolith of Traprain Law c.10 km further south (**Fig. 2**: Upton *et al.* 2007). These have not been radiometrically dated but the Traprain Law intrusion uplifted the surrounding GHVF strata and is clearly younger than the latter. These very shallow-level intrusions are probably sub-volcanic, although no associated lavas or pyroclasts are preserved. Any extrusives that may have overlain the silica-saturated trachytes would have been preferentially eroded. As reported by Cameron & Stephenson (1985) the over- and under-saturated trachytes appear to have been near-contemporaneous as had been suggested by K/Ar dating by De Souza (1982).

► This paper presents new whole-rock, trace element and Sr and Nd isotopic data for the trachytes together with a description of the (previously unstudied) Marine Villa ignimbrites. Despite the very fragmental record accorded by such a poorly exposed area, this paper presents some new perspectives on what are considered to be the eroded relics of a central-type silicic volcano.

Analytical techniques

Major element compositions were determined using an X-ray fluorescence spectrometer at the School of GeoSciences, University of Edinburgh. Samples were fused with a lithium metaborate-lithium carbonate flux containing La_2O_3 as a heavy absorber (Johnson Matthey Spectroflux 105). Sample powders were dried at 110°C over-night, and a nominal but precisely weighed 1 g aliquot was ignited at 1100°C to determine loss on ignition (LOI). The residue was then mixed with flux in a sample flux ratio of 1:5 based on the unignited sample weight. The mixture was then fused at 1100°C in Pt5%Au crucibles in a muffle furnace. After initial fusion, the crucible was re-weighed, and any weight loss from the flux was made up with additional flux to maintain the 1:5 sample flux ratio as closely as possible. After a second fusion and homogenisation over a Meker burner the molten sample flux mixture was cast onto a graphite mould and flattened with an aluminium plunger. During this process, the mould and plunger were kept at 220°C on a hot-plate.

The fused discs were analysed on a Philips PW2404 X-ray fluorescence spectro-node primary X-ray tube. Corrections for matrix effects on the intensities of the major element fluorescence lines were made using theoretical alpha coefficients calculated using Philips software. The coefficients were calculated to allow for the extra flux replacing volatile components in the sample lost during the LOI step, so that analytical totals should be 100% less the LOI. The spectrometer was calibrated using a suite of ten USGS and CRPG standards the values of which are given in Govindaraju 1994. Calibration lines show that accuracy and precision are closely similar with an average 1σ value across the ten elements of 0.03 wt%. Trace elements concentrations were determined on pressed powder samples. Eight grams of rock powder were mixed thoroughly with eight drops of a 2% solution of polyvinyl alcohol. This mixture was formed into a 40 mm disc backed and surrounded by an aluminium foil cup, by compression between two tungsten carbide discs at a pressure of 1.6 tonnes/cm^2 .

Mineral identification and semi-quantitative phase analysis by X-ray diffraction were conducted at the School of GeoSciences, University of Edinburgh. The instrument used was a Bruker-AXS Advance 8 powder diffractometer in a Bragg-Brentano (reflection) geometry, equipped with a Cu tube, (wavelength $K_{\alpha 1} = 1.540596 \text{ \AA}$), and a NaI detector. Variable slits were used, and the samples were measured in 2 to 65° 2θ range. Identification of minerals was made

using Bruker EVA software with reference to the International Centre for Diffraction data PDF-4+ 2015 diffraction database. Quantitative analysis was determined by the Rietveld method (Rietveld 1969) using the computer program Topas 3.1 (from Bruker-AXS). SEM analyses: these were made using polished thin-sections coated using Denton BTT-IV evaporation coater. Backscatter (BSE) imaging and EDS analyses were acquired using a Carl Zeiss SIGMA HD FEG SEM in the School of GeoSciences, University of Edinburgh with an Oxford Instruments Aztec Energy Dispersive Spectrometry (EDS) system using an accelerating voltage of 20Kv and a 7 mm working distance.

Column chemistry: Around 0.2000g of finely crushed rock powder was put into a PFA Teflon (savillex) beaker, and digested in steps using ultrapure HF, HNO₃ and HCl over a hotplate at 140-150°C. Sr and REE fractions were separated by standard cation exchange column chemistry using Bio-Rad analytical grade cation exchange resin AG 50W x 8 hydrogen form in a quartz glass column with approximate column volume of 12 ml. In order to further separate Rb from the Sr fraction, samples were passed through Triskem Sr spec resin columns, using methods from Pin & Bassin (1992), and the Sr fraction collected. Rare earth elements fractions from cation columns were passed through small columns containing Triskem/Eichrom Ln resin (Pin and Zalduegui, 1997), and the Nd fractions collected.

Isotope Analyses: Sr samples were loaded onto Re filaments using a Ta emitter solution, and Nd samples were loaded onto Ta side-filaments in a triple filament arrangement. Total procedure blanks were <300 pg. Concentration data were obtained by isotope dilution mass spectrometry (Sr, Nd and Sm). Samples were analysed on a VG Sector 54 mass-spectrometer operated in multi-dynamic mode for isotope ratios and Sr concentrations and static mode for Sm and Nd concentrations. The NIST SRM987 standard gave 0.710268 ± 20 (2SD n = 29) and the JNd(i)-1 standard gave 0.512078 ± 8 (2SDn = 17) during this study.

Composition of the Garleton trachytes

There are many published compositional data for the Carboniferous basaltic lavas and intrusions but not for the silicic products. Although data on the phonolitic products of the Lower Carboniferous have been published (Upton *et al.* 2007) there are very few published for the silica-saturated and oversaturated lavas and

intrusions. Table 1, presented here, gives the only recent full analyses of a selection of the GHVF trachytes. The freshest trachytes are mainly from the Garleton Hills lavas and some of the (possible) Garleton intrusions (Table 1). They are commonly feldspar-phyric, some containing sanidine phenocrysts and, others with sodic plagioclase cores and alkali feldspar outer zones. Diagrams illustrating a compositional continuum from basalts to trachytes in East Lothian show this proceeding via mugearites and benmoreites (Macdonald 1975). Our trachyte samples, from localities ranging from Eyebroughy in the NE to the Garleton Hills in the south (**Fig.2**), show little change in composition. Apart from feldspar phenocrysts they contain micro-phenocrysts of hedenbergitic pyroxene, apatite and, more rarely, iddingsite pseudomorphs after fayalite. The matrices comprise alkali feldspar, aegirine-augite, quartz, apatite and magnetite. Fluorite is recorded from intersertal patches in quartz trachytes from several areas e.g. on Eyebroughy, a fine-grained quartz trachyte body in the far north-west of the Bangly Member and also, in one case, together with drusy quartz (McAdam & Tulloch 1985). The lava above the welded ignimbrite at Weaklaw ('Smugglers' Cave') had been assumed to be part of the trachyte series. However, despite it having undergone alteration, analysis shows a loss-on-ignition of 4.8% and the contents of TiO_2 , P_2O_5 , Ni, Cr, V and Sc are too high for the flow to have been a trachyte (Table 1). Its low MgO and Na_2O and very high K_2O and Ba contents are unlikely to represent magmatic values and are almost certainly due to a) leaching and b) enhancement of large-ion elements by later fluids related to a nearby (Namurian?) vent (Upton *et al.* 2019). We surmise that the lava is probably a metasomatized mugearite.

The contrasted compositions of the Bangly Member trachytes and the East Lothian silica-undersaturated trachyte intrusions (North Berwick Law, Bass Rock and Traprain Law (Upton *et al.* 2007) are shown in terms of alkali vs. silica **Fig.4**. Their trace-element data emphasize differences between the two groups: the Zr, REE and Y concentrations are higher in the oversaturated trachytes than in the phonolitic trachytes although their Nb contents are lower (Table 1). The mean Zr/Nb values are 6.62 and 3.09 respectively for the two groups.

The Sr and Nd isotope initial ratios are shown in Table 2. Initial $^{87}\text{Sr}/^{86}\text{Sr}$ values show extreme variations which are unlikely to be representative of magmatic ratios and probably indicate post-crystallisation exchange of Rb and/or Sr. Initial $^{143}\text{Nd}/^{144}\text{Nd}$ appear more robust, as has previously been noted for the phonolitic trachytes of the Midland Valley (Upton *et al.* 2007). Epsilon Nd³⁴⁵ (Table 2) ranges from +2.0 to +3.8: these positive ϵNd^{345} values clearly confirm that the parental magmas were mantle-derived.

The Sr and Nd isotopic differences between the silica-saturated/oversaturated trachytes and the phonolitic trachytes are small. The ϵNd^{345} of the Bangly trachytes varies from +2.0 to +4.6 (mean +3.5) whilst that for the phonolitic trachytes is from +1.99 to + 3.3 (mean +2.5: Upton *et al.* 2007). Such differences are regarded as negligible and the difference in the Sr isotopes appears insufficient to ascribe the differentiation to silica saturated/oversaturated magmas solely to crustal contamination.

The Marine Villa ignimbrites

Apart from some welded pyroclastic rocks in the Clyde Plateau Volcanic Formation (McPherson & Phillips 1998) the welded tuffs cropping out on the north coast of the Bangly Formation, from Marine Villa westwards (**Fig.5**) appear to be the only other known ignimbritic occurrence in the U.K. Carboniferous. These Marine Villa ignimbrites, some 10 m thick, are excellently exposed on the coast between [NT 502 861] and [NT 499 859] (**Fig.5**).

There being no previous description of these ignimbrites an introductory account of them is given below. The outcrops can be followed west from Marine Villa on the foreshore and adjacent sea-cliffs for c. 200 m. until their outcrop is replaced by a younger unrelated suite of (unwelded) tuffs. After another 300 m westwards along the coast at [NT 499 859], the ignimbrites re-appear beyond two prominent cliff features known as the Hanging Rocks, close to Weaklaw (**Fig.5**). These welded ignimbrites have complex structural features and a plethora of lithic clasts, with some, intercalated, fall deposits.

The suite commenced with stratified ignimbrites that cover the underlying mugearite lava (**Fig.6**). In places, this grades to a pumice-lapilli fall deposit, with

roughly equally sized pumice fiammé and few, smaller lithics. This fall deposit is often truncated by the overlying cross-stratified ash-rich tuff with occasional block-sized lithics and well-defined bedding planes. The bedding planes, well exposed at Marine Villa, exhibit broad undulatory ('swaley') surfaces with wave-lengths measurable in tens of metres and are interpreted as unmodified depositional features (**Fig. 7**). Above them is a structurally and granulometrically complex welded ignimbrite sequence comprising laterally discontinuous beds and lenses of eutaxitic massive lapilli tuff, bedded lapilli tuff, cross-stratified tuff, and massive ash-rich tuff.

Distinct beds within the ignimbrites are defined by granulometry, structure and variation in polymictic lithic concentration. In the lower portion of the ignimbrite sequence, a particularly lithic rich region of bedded lapilli tuff, contains metre-scale 'crypto-clasts' of earlier phases of the same eutaxitic bedded lapilli tuff; identifiable as distinct regions where the bedding is rotated in comparison to horizontal bedding of the surrounding ignimbrite. This shows that the deposit became cohesive shortly after emplacement, and capable of whole-scale re-entrainment by later phases of the same density current; indicating that pyroclasts were sufficiently hot and fluidal to weld on collision (Mahood 1984; Branney & Kokelaar 1992). Pumices in the Marine Villa tuffs are also deformed, forming elongated metasomatised fiammé (**Figs. 8a and 8b**) and their development in welded ignimbrites may be attributed to simple shear during emplacement and pure shear during compaction and alteration (Bull & McPhie 2007). Apart from rare fragments derived from the underlying mugearite lava, the lithic clasts in these ignimbrites cannot be matched to any of the other lithologies exposed in the GHVF.

Scanning electron microprobe (SEM) analyses were performed on fiammé in polished thin-section from two welded ignimbrite samples. MV6, from outcrops at Marine Villa [NT 502 861] and MV13 beside the Hanging Rocks [NT 499 859]. Back-scattered images of both are shown in Figs. 8a and 8b.

The fiammé (and other interstitial matter) are presumed to represent former glassy pumice pseudomorphed mostly by kaolinite (XRD composition containing 45 wt% SiO₂ and 40 wt% Al₂O₃) and subsidiary dolomitic and ankeritic carbonate. XRD mineral identification of hand-picked fiammé indicates only kaolinite and minor quartz but SEM scans on polished sections shows vestigial evidence of former phenocrysts that comprise small corroded fragments of quartz and also 'ghost' pseudomorphs of

feldspar (**Fig.9**), SEM scans across the latter reveal minute relics of material for which the semi-quantitative analysis gives a composition of SiO_2 c.65%, Al_2O_3 , c.19% and K_2O c.16 wt %, *i.e.* approximating end-member K-feldspar. From these, admittedly very sparse, data we tentatively conclude that the erupted magma contained both quartz and alkali feldspar phenocrysts.

Since the Marine Villa ignimbritic tuffs lie directly above a mugearite lava (**Fig.6**) they appear to be earliest manifestations of the Bangly Member. However, the ignimbrites contain some highly amygdaloidal (and feldspar-phyric), lava clasts (**Fig.10**) that differ from all the other known GHVF lavas. These distinctive clasts are considered to be from a trachytic lava although this could not be verified petrographically or analytically because of the pervasive metasomatism by younger fluid(s) that pseudomorphed virtually all former vitric and crystalline materials by clays, carbonates and iron oxides. Thus, despite the stratigraphic evidence that the Marine Villa tuffs were early manifestations of the Bangly Member the presence of such lava clasts demonstrates that eruptions had occurred in the interval between the mugearite and Marine Villa tuff eruptions. Because there are no surviving matching outcrops, we can only speculate that they came from the flanks of an eroding volcano. The Marine Villa ignimbrites contrast with the preceding lavas and intrusions, not merely because of their wholly different volcanological character but compositionally. Other than the scarce clasts of the underlying mugearite there is little or no compositional match between the lithic clasts in the Marine Villa tuffs and the lithologies of the exposed Bangly Member lavas and intrusions (Tables 1, 3).

The coarser strata of the Marina Villa tuffs contain lithic clasts of variable size and colour. Most lithics are fine-grained aphyric or micro-phyric igneous rocks with minimal vesiculation (**Fig.8a**) that may signify derivation from hypabyssal intrusions rather than lavas. Most clasts are angular although a degree of rounding in some is attributed to abrasion during transport. The lowest strata tend to contain clasts of <15 mm but many of the younger units are notably coarser with closely-packed clasts up to 300 mm. There is notable heterogeneity among the clasts; their colours range from black to shades of brown, yellow or green. In some strata composite clasts (up to 1 m across) contain fragments of earlier ignimbrites that had consolidated prior to incorporation in younger eruptive pulses. Some of the larger lithic clasts from two

outcrops at [NT 502 861] were separated for analysis. A sample of ignimbrite that is virtually 'lithic-free', taken from immediately west of the Hanging Rocks [NT 499 859] was also analysed (**Fig.8b**: Table 3).

There are all gradations between well-compacted lithic-poor strata and others that are dominantly composed of lithics. In the coarser clast-rich facies the interstitial white carbonate-clay material has a complex morphology and is interpreted as juvenile pumice caught between the lithic fragments that compacted under shear. Very fine-grained strata, free of clasts, form distinctive features in the stratigraphy. These reddish-brown, homogeneous layers, up to 300 mm thick, occur at irregular intervals are regarded as co-ignimbrite ash accumulations formed during periods of repose.

Structures in the ignimbrites and mode of deposition

Some contacts between strata have small-scale flame structures indicative of loading, implying that the underlying strata were still ductile when the next deposit was being laid down (**Fig.11a**). Elsewhere there is clear evidence of earlier strata being consolidated and capable of brittle fracture prior to being overlain by the next deposit. In one example, a large (c.1m) block of ignimbrite, presumably derived from an earlier unit, has been rotated up to 45° within surrounding layered tuffs and points to its rapid consolidation to a coherent rock followed by its breakage and movement by a younger pyroclastic density current.

Other features indicative of lateral flow during deposition include cross-bedding, channelling and rip-up clasts (**Fig.11b**). Very clast-rich units dominate in the succession leading to the conclusion that they resulted from a succession of pyroclastic density currents, separated by brief time intervals. The Marine Villa ignimbrites are complex and offer much scope for further study.

Composition of lithic clasts in the Marine Villa ignimbrites

The textures within the clasts are well preserved but secondary alteration has obscured or obliterated their original mineralogy and chemistry. The alteration is ascribed to pervasive hydrothermal metasomatism that preserved textures but materially affected their chemistry. One chemical anomaly is the notably high K/Na ratios that characterise the tuffs as well as the 'Smuggler's Cave' lava overlying them

at 'Smugglers' Cave' (Table 1). This is suspected to be mainly due to sodium leaching but LILE enrichment cannot be ruled out.

Two samples (MV1 and MV2) from the lapilli-grade ignimbrites at Marine Villa and one (MV3) from [NT 499 858] close to Weaklaw were investigated by X-ray diffraction. The approximate modes obtained showed them to comprise 40-10% feldspar, 9-3% dolomite with subsidiary ankerite and siderite, 7-2% muscovite, 26-11% clay (kaolinite + illite), 9-1% haematite, quartz 37-8%. MV3 also contains small zircon crystals, not seen elsewhere, reflected in their higher whole-rock Zr values (c.1490 ppm: Table 3). Analyses of eleven individual lithic clasts, large enough to be sawn out, are presented in Table 4. A twelfth analysis is a composite sample made up of (small) clasts of uniform aphyric material, hand-picked from a crushed sample (MV28).

The compositions of all twelve are highly variable and, using discriminant function analyses (Roser & Korsch 1988), they were divided into those of igneous and sedimentary provenance. Four of the twelve clasts analysed from the Marine Villa tuffs are of sedimentary origin (**Fig.10a and b**). All samples show a clear negative correlation between SiO₂ wt.% and loss on ignition (LOI): samples with the lowest SiO₂ wt.% show the highest LOI values (**Fig.10a**). The igneous clasts from the welded ignimbrite can be divided into two groups based on major element abundances – Group 1 with high SiO₂ and K₂O, and low MnO, CaO and MgO is similar in composition to the welded tuff as sampled at Marine Villa and are intermediate to felsic in composition; Group 2 has low SiO₂ and K₂O, and high TiO₂, Al₂O₃, MnO, MgO, CaO and P₂O₅, and plot in the mafic field (Fig. 10b). Group 2 also has higher LOI, which is consistent with many of the samples having been affected by alteration. Similar variability is evident in the trace elements of the two groups: Group 1 has lower concentrations of Ni, Cr, V, Sc, Cu and Zn, but generally higher contents of more incompatible elements (Rb, Ba, Nb, Zr, Th, La and Ce) than the Group 2 clasts (Table 4; **Fig.10b**). This is consistent with the Group 2 clasts having been initially more primitive in composition.

Discussion

Trachyte genesis: The genesis of trachytic magmas confined to the 30 Myr of the early Carboniferous makes this period exceptional in UK magmatic history. Subsequent Carbo-Permian basaltic magmas (other than the Stephanian tholeiites)

were more strongly silica-undersaturated: such (basanitic) magmas are held to represent lower-degrees of mantle melting than the majority of the earlier basalts and thus typically generated in smaller volume. In contrast the earlier Dinantian 'transitional basalts' could produce trachytic residues as at the Skelfhill Pen, Eildon, the GHVF, Glencartholm and numerous sites in the Clyde Plateau Volcanic Formation, Kintyre as well as in SW Ireland. Magmatic evolution of the GHVF is deduced to have taken place under polybaric conditions (Macdonald 1975; MacDonald & Whyte 1981; Russell 1985; Stephenson *et al.* 2003) and that were capable of generating both silica over- and under-saturated trachytic residua.

Caldera formation: According to Williamson (2003a) the thick flows exposed in the Garleton Hills may have constituted lava domes. However, their conformity and inferred original horizontality militates against this and are more compatible with having been caldera in-fillings. Whether eruption of the Marine Villa ignimbrite led to caldera collapse is speculative but it is a reasonable assumption that their evacuation caused some collapse.

Caldera-ponded trachyte flows are well-known, e.g. in the Kenyan Rift volcanoes (Leat 1984). Had there been a caldera enclosing these thick lavas its area would have been restricted to only a few kilometres across. If the caldera hypothesis is accepted the area may offer a case of inverted topography, where once low-lying but comparatively erosion-resistant massive lavas ultimately formed the highest parts of the topography.

Possible GHVF analogues in the Clyde Plateau Volcanic Formation: The relative abundance of trachytic extrusives in parts of the Clyde Plateau Volcanic Formation, particularly between Greenock and Strathaven, led McPherson *et al.* (2001) to note that the volcanoes may have been higher in those areas. Among the better known is the Misty Law centre (in Renfrewshire) with its association of trachytic pyroclasts and plugs and its massive trachyte and rhyolite lavas (Tyrrell 1917; Johnstone 1965; Stephenson, in Paterson *et al.* 1990; Monroe 1999) with smaller occurrences in the Dunlop-Eagelsham-Strathaven uplands (McPherson & Phillips 1998). Trachytes and rhyolites characterise the Meikle Bin volcano in the Campsie Hills (Whyte & MacDonald 1974). The best example of caldera formation comes from the Whitehead Volcanic Complex of the Campsie Fells where multiple necks and a wide variety of rock types lie within an oval ring-fault measuring 2 x 2.5 km (Craig 1980; Forsyth *et al.* 1996, summarised by Stephenson 2003).

The instances cited above emphasize the singularity of the Lower Carboniferous magmatism over a very wide area in the generation of silicic (predominantly trachytic) magmas from parental magmas that, despite their geochemical 'provincialism' could all be described as 'mildly alkaline' olivine basalts.

Ignimbrite occurrences: As noted above, ignimbrites are rare in the Carboniferous. The Marine Villa ignimbrites of the GHVF provide evidence of violent eruptions contrasting with the low-energy expulsion of relatively degassed magma as lavas and shallow-crustal intrusions.

A suite of stratified and unwelded tuffs, formerly considered part of the GHVF, transects the Marine Villa ignimbrite at what appears to be a crater margin (McAdam 1996). These younger tuffs, cropping out on the foreshore for several hundred metres between [NT498.857] and [NT 502.860] are more probably Late Carboniferous and related to the Weaklaw Vent (Upton *et al.* 2019) and genetically unrelated to the Marine Villa ignimbrites. The pervasive hydrothermal metasomatism that preserved textures but materially affected the chemistry of the ignimbrites are noted above.

Summary

The Dinantian eruptions that took place over c.30 Myr, were the earliest anorogenic magmas in the Phanerozoic of the British Isles. In Scotland they were mainly concentrated between the Iapetus suture and the Highland Boundary Fault with the Kintyre eruptions, lying to the north of the fault, being the sole exception.

Locally, fractional crystallisation of larger volumes of mildly alkaline (transitional) basaltic magmas led to trachytic (and possibly rhyolitic) residua in the Garleton Hills Volcanic Formation. The Bangly Member comprising the culminating products of this Formation is made up of silica-saturated to over-saturated trachyte lavas and (suspected) shallow-crustal intrusions together with the rhyolitic (?) Marine Villa ignimbrites.

The silicic rocks (mainly trachytes) of the GHVF appear to cover a greater outcrop area than at any of the other Lower Carboniferous sites. The maximum thickness of the Bangly Member coincides with the principal outcrops composing the Garleton Hills. Inferentially these may represent the axial region of a central-vent trachytic volcano in which the several thick, conformable and originally horizontal trachyte lavas were possibly caldera-ponded.

The relative uniformity of the trachytic compositions representing the most silicic components of the Formation, probably crystallised at a cotectic analogous to that of the sanidine-diopside-water experimental system (Yoder & Upton 1971). Their silica-saturated to silica-over saturated compositions contrasts with the several (inferred younger) phonolitic trachyte intrusions in the vicinity. Neodymium and strontium isotopic data suggest that, while both the trachytes and the phonolitic trachytes are mantle derivatives, their differences in silica saturation probably owes less to varying degrees of crustal contamination than to polybaric fractional crystallisation.

The Marine Villa ignimbrites were the products of repeated, high-energy, pyroclastic density currents. Their welding was responsible for their preservation in relatively erosion-resistant sea-cliffs. The lithic clasts comprise a heterogeneous assemblage of both igneous and sedimentary rocks. Some igneous clasts are derived from the underlying mugearite but the majority are exotic in that they cannot be correlated with any other rock types exposed in the GHVF. Most tend to be aphyric and lacking in amygdales and could come from hypabyssal intrusions while others, notably porphyritic and amygdaloidal, were clearly lava samples, possibly acquired from the flanks of the volcano.

Whole-rock analyses of lithic clasts provide very limited information but emphasize the heterogeneity. It is inferred that the Marine Villa tuffs experienced pervasive metasomatism by fluids that left textures intact but severely modified their chemistry. The colourless clays that compose the fiammé and interstitial materials are interpreted as pseudomorphed glass. The morphology of the fiammé as indicated by **Fig.8a**, implies that they were extremely malleable. The presence of fluorite in the Garleton trachytes provides some support for considering that the silicic magmas of the Bangly Member had low-viscosities. Some indication of magma composition comes from the whole-rock analysis of the lithic-poor and essentially parataxitic ignimbrite MV3 from 'Smugglers' Cave' (Table 1) that has a whole-rock Zr content of c. 1500 ppm, total REE of 864 ppm and a low Zr/Nb ratio of 5.1. Such enrichment in Zr and REE is consistent with involvement of melts enriched in halogens which would have lowered their viscosity (Kirstein *et al.*, 2001).

While the suggestion that the Marine Villa ignimbrites related to a major pyroclastic eruption from a shallow-level magma chamber leading to caldera

collapse (Upton 2003) remains unprovable, it is a reasonable assumption that it caused collapse somewhere in its roof-zone.

Evolution and tectonics

Although the Bangly Member presents the largest area of Lower Carboniferous silicic extrusives in the U.K. Carboniferous, the original area covered must have been considerably greater. Its relative neglect can be ascribed to the poverty of exposure. The GHVF as a whole lies adjacent to an off-shore extension of a fault postulated to mark the boundary between the Midland Valley and Southern Uplands Terranes. Adiabatic melting of MORB-source mantle (cf. Smedley 1988) resulting from trans-tensional movement of the fault is suggested to have generated the 'transitional' (mildly alkaline) basaltic magmas parental to the GHVF. Fractional crystallisation of these (Smedley 1986a; Macdonald 1975) gave rise to a (generalised) 'line of liquid descent' from relatively primitive to increasingly fractionated compositions. This generally terminated at the alkali feldspar-clinopyroxene cotectic at the appropriate pressure (the natural analogue of the sanidine-diopside cotectic e.g. at 10 kb, Yoder & Upton (1971). While we make tentative identification of the ignimbrite magmas as rhyolitic, its genesis cannot be judged in the absence of compositional data.

Currently the trachyte outcrops are scattered over a distance of > 20 km from Eyebroughy in the far NNW to the vicinity of Gifford in the SSE, suggesting the likelihood of a substantial trachytic edifice. The total volume of trachyte magma cannot be calculated but is likely to have been of several cubic kilometres. Assuming that all was a differentiate from basalt magma, the latter would have had a very large volume. If all was derived from fractional crystallisation from a single underlying mafic body the latter would have had to be of correspondingly great volume. This appears inherently highly unlikely. An alternative mode of growth for such a large quantity of silicic magma late in the evolution of the GHVF may, more feasibly, have involved slow crystallisation of numerous small mafic magma chambers (dyke-like geometry?) undergoing side-wall fractionation over a long interval. Small volumes of residual trachytic melt would have been buoyant and could theoretically have arisen through incompletely crystalline cumulates to accumulate incrementally at a shallow crustal level, providing a source for late-stage eruptions. Such a hypothesis could also explain evolution of chambers from which

volcanic centres such as Misty Law or Meikle Bin in the CPVF were supplied. The scattered fragments of evidence left in East Lothian are consistent with the concept of a large trachytic volcano having been the final product of the GHVF.

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Figures:

Fig.1. Map illustrating the geographical relationships between the localities for Dinantian volcanic rocks in Scotland. (The dots indicating the many plugs and necks in the Southern Uplands are diagrammatic except for the Eildon Hills). Key-map indicates the area of the Limerick outcrops.

Fig.2. Geological map showing the distribution of the Bangly Member (silicic rocks) within the Garleton Hills Volcanic Formation, East Lothian. The Late Carboniferous quartz dolerite dykes have been omitted. (After McAdam & Tulloch 1985).

Sample sites are abbreviated: B= Bangly (NT 493 756); P= Phantasie (NT 509 758); SH = Skid Hill (NT 506 835); CQ= Craig's Quarry, (MT 505 835); EY = Eyebroughy (NT 494 863); C = Chesters (NT 510 785).

Fig.3. View from Kae Heughs (NT 507 764) in the Garleton Hills depicting two of the thick trachyte lavas.

Fig.4. Compositions of the contrasted GHVF trachytes and the East Lothian phonolitic trachytes shown in the alkali-silica ternary diagram.

Fig. 5. Geological sketch-map of the coast between Weaklaw and North Berwick

Fig.6. Well laminated ignimbrites infilling a hollow on the surface of a mugearite lava.

Fig.7. The undulating (swaley) surface of a tuff unit at Marine Villa, identified as an unmodified palaeo-surface of a cross- stratified ignimbrite.

Fig.8 Back-scattered SEM images of two contrasted ignimbrite strata (cut normal to the bedding). a) sample close to Marine Villa showing sub-angular lithics of fine-grained volcanic rocks .b) sample of lithic-poor facies of ignimbrite from immediately west of Hanging Rocks at (NT 497 858). It is dominantly composed of ovoid to lenticular creamy-white de-vitrified fiamme up to c. 5 mm in length.

Fig.9. Alkali feldspar pseudomorph in fiamme.

Fig.10 Amygdaloidal, porphyritic lava clast identified as probable trachyte.

Fig.11a. Contrasting facies of the Marine Villa tuffs. Lower portion of the image shows conspicuous, moderately deformed, fiamme. It is overlain by ashy laminations, penetrated by 'flames' arising from the underlying layer, consequent on loading while still ductile. This, in turn, is overlain by a coarse-grained tuff.

Fig.11b. The lower part of the image shows a cross-stratified lapilli tuff. Centre; c. 20 cm thick, finer-grained diffusely cross stratified lapilli tuff overlain by coarser, lithic rich massive lapilli tuff. This unit contains a plethora of lithic clasts including a large homogeneous rip-up clast of homogeneous tuff.

Fig.12a. Plot of loss-on-ignition (LOI) - SiO_2 for all samples high-lighting the clear negative correlation between LOI and SiO_2 . Note the bimodal composition of the sedimentary clasts and the different groupings of clasts in the welded ignimbrite based on element mobility.

Fig.12b. Discriminant function figure to establish the provenance of the clasts in the welded ignimbrite and volcanoclastic sediments. The discriminant functions are;

1: $-1.773 \text{ TiO}_2 + 0.607 \text{ Al}_2\text{O}_3 + 0.76 \text{ Fe}_2\text{O}_a - 1.5 \text{ MgO} + 0.616 \text{ CaO} + 0.509 \text{ Na}_2\text{O} - 1.22 \text{ K}_2\text{O} - 9.09$.

2: $0.445 \text{ TiO}_2 + 0.07 \text{ Al}_2\text{O}_3 - 0.25 \text{ Fe}_2\text{O}_3 - 1.142 \text{ MgO} + 0.438 \text{ CaO} + 1.475 \text{ Na}_2\text{O} - 6.861$ (After Roser & Korsch 1988).

Plot of immobile elements Y vs. Nb high-lighting the potential petrogenetic relationship between the different clasts sampled from the tuff and trachyte lavas and Intrusions.

TABLES

Table 1. Whole-rock analyses of some trachyte lavas and intrusions from the GHVF.

Table 2. Sr and Nd isotopic compositions of trachytes from the GHVF.

Table.3. Whole-rock analyses of welded (lithic-poor) tuffs. MV1 and MV2; coast at Marine Villa (NT 502.861) and MV3 Weaklaw (NT 499.859).

Table 4. Whole-rock compositions of lithic clasts separated from the Marine Villa tuff. [NT 505.860 – 502.859]. No trace element data are given for MV31 as the sample was too small.

B. (i)

(ii)

Table 1. Whole-rock analyses of the trachyte lavas and intrusions from the Garleton Hills Volcanic Formation

Grid Ref. NT	Bangly quarry 485.756	Phantasie 512.757	Skid Hill 505.766	Craig's quarry 509.836	Eyebroughy 485.756	Chesters 506.782	Smuggler's Cave 501.861
SiO ₂	61.43	62.36	62.29	63.14	66.26	66.90	51.74
TiO ₂	0.66	0.43	0.51	0.25	0.25	0.36	1.45
Al ₂ O ₃	19.86	16.00	16.62	16.29	15.97	15.17	15.67
Fe ₂ O ₃ *	4.05	5.76	5.09	4.35	5.17	5.38	9.38
MnO	0.05	0.13	0.14	0.13	0.10	0.01	0.20
MgO	0.87	0.38	0.87	0.15	0.31	0.36	1.60
CaO	1.68	0.96	2.08	1.39	0.88	0.43	4.20
Na ₂ O	4.97	5.52	5.64	5.36	5.20	4.16	0.45
K ₂ O	5.47	5.45	5.40	5.71	5.76	5.79	6.24
P ₂ O ₅	0.16	0.07	0.09	0.03	0.03	0.05	0.43
LOI	0.35	2.78	0.72	2.89	4.08	1.16	8.32
Total	99.55	99.84	99.45	99.69	100.01	99.78	99.67
Dis. F. 1	0	0	0	0	0	-2	-2
Dis. F. 2	9	9	9	10	9	7	2
Ni	1.4	3.2	3.2	0.9	0.4	2.1	69.7
Cr	3.7	4.2	6.2	1.7	3.0	2.4	43.1
V	7.7	9.9	14.6	n.d.	0.8	1.3	93.9
Sc	6.6	3.3	3.7	1	0.4	0.9	13.4
Cu	14	11.9	13.2	7	1.7	12.4	13.9
Zn	132	176	227	92.6	88.2	337	108
Pb	7.5	11.1	234	15.9	12.4	19.8	11.2
Sr	186	82.9	94.4	85.1	19.1	25.0	81.8
Rb	129	166	158	179	199	167	88.8
Ba	831	434	472	398	43.9	127	543
Nb	87.3	137	125	137	169	125	61.8
Zr	576	986	902	791	803	1199	322
Y	44.9	78.7	66.6	65.3	68.9	59.3	32.3
La	70.2	110	110	95.1	90.6	128	46.1
Ce	138	227	225	191	184	207	103
Nd	55.6	93.4	92.6	76.6	73.1	99.8	45.3
U	3.7	4.8	5	5.9	5.2	4.1	4.0
Th	14.1	20.2	13.4	20.2	19.9	17.8	8.5

Dis. F. 1 = $-1.773\text{TiO}_2 + 0.607\text{Al}_2\text{O}_3 + 0.76\text{Fe}_2\text{O}_3^{\text{T}} - 1.5\text{MgO} + 0.616\text{CaO} + 0.509\text{Na}_2\text{O} - 1.224\text{K}_2\text{O} - 9.09$ Dis. F. 2 = $0.445\text{TiO}_2 + 0.07\text{Al}_2\text{O}_3 - 0.25\text{Fe}_2\text{O}_3^{\text{T}} - 1.142\text{MgO} + 0.438\text{CaO} + 1.475\text{Na}_2\text{O} + 1.426\text{K}_2\text{O} - 6.861$

Table 2. Whole rock strontium and neodymium isotope data for trachyte lavas and intrusions from Garleton Hills Volcanic Formation

Sample Location	$^{87}\text{Sr}/^{86}\text{Sr}_{(0)}$	Rb (ppm)	Sr (ppm)	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}_{(i)}$	ϵSr_{345}	$^{143}\text{Nd}/^{144}\text{Nd}_{(0)}$	Sm (ppm)	Nd (ppm)	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}_{(345)}$	ϵNd_{345}
Craig's Quarry	0.734403 ±18	178.6	102.6 (2)	5.050	0.709604	+75	0.512650 ± 8	12.82 (3)	68.14 (8)	0.1138	0.512393	+3.9
Bangly Quarry	0.713900 ±23	129.2	151.3 (3)	2.472	0.701759	-36	0.512556 ± 12	10.47 (1)	55.40 (8)	0.1143	0.512298	+2.0
Phantasie	0.732527 ±21	165.7	117.4 (2)	4.094	0.712424	+116	0.512685 ± 17	16.33 (2)	87.79 (2)	0.1124	0.512431	+4.6
Skid Hill	0.725847 ±22	148.2	115.7 (2)	3.713	0.707617	+47	0.512635 ± 11	11.63 (4)	61.34 (6)	0.1147	0.512376	+3.6

Figures in parentheses are 2 SE errors on least significant decimal case including error magnification. Errors on isotope ratios are also 2 SE.

Table 3. Whole-rock analyses of ignimbrites

	MV1	MV2	MV3
SiO ₂	53.93	61.44	52.42
TiO ₂	0.52	0.59	0.54
Al ₂ O ₃	18.66	16.19	28.79
Fe ₂ O ₃ *	7.56	8.46	2.13
MnO	0.17	0.03	0.09
MgO	1.70	0.61	0.68
CaO	3.05	0.94	1.42
Na ₂ O	0.05	0.15	0.05
K ₂ O	2.30	7.24	0.56
P ₂ O ₅	0.10	0.13	0.07
LOI	10.90	3.94	12.85
Total	98.96	99.71	99.49
Dis. F. 1	4	-3	8
Dis. F. 2	-4	3	-4
Ni	6.0	6.0	35.0
Cr	9.0	5.0	2.0
V	11.0	5.0	15.0
Sc	1.0	1.0	1.0
Cu	10.0	4.0	3.0
Zn	27.0	19.0	5.0
Pb	4.4	9.6	6.4
Sr	163	336	30.0
Rb	51.0	109	13.0
Ba	7300	430	52.0
Nb	160	126	291
Zr	528	329	1488
Y	51.0	26.0	119
La	113	130	184
Ce	182	213	384
Nd	73.0	69.0	177
U	0.8	0.1	7.0
Th	16.1	10.4	35.5

Dis. F. 1 = $-1.773\text{TiO}_2 + 0.607\text{Al}_2\text{O}_3 + 0.76\text{Fe}_2\text{O}_3^{\text{T}} - 1.5\text{MgO} + 0.616\text{CaO} + 0.509\text{Na}_2\text{O} - 1.224\text{K}_2\text{O} - 9.09$

Dis. F. 2 = $0.445\text{TiO}_2 + 0.07\text{Al}_2\text{O}_3 - 0.25\text{Fe}_2\text{O}_3^{\text{T}} - 1.142\text{MgO} + 0.438\text{CaO} + 1.475\text{Na}_2\text{O} + 1.426\text{K}_2\text{O} - 6.861$

Table 4. Major- and trace-element data from clasts within the welded ignimbrites

	MV27	MV28	MV29	MV30	MV31	MV32	MV35	MV36	MV37	MV38	MV39	MV40
SiO ₂	54.22	63.06	58.02	26.91	35.70	56.73	34.42	35.41	56.39	25.00	31.35	70.56
TiO ₂	0.40	0.44	1.23	2.64	2.06	1.20	0.27	0.19	1.37	2.58	0.24	0.38
Al ₂ O ₃	17.62	15.08	14.41	18.26	21.63	14.23	11.93	15.19	18.17	19.30	12.13	12.81
Fe ₂ O ₃ *	8.57	8.82	12.26	16.34	6.61	12.09	6.42	8.21	9.21	7.91	9.08	5.77
MnO	0.06	0.04	0.05	0.27	0.26	0.06	0.41	0.37	0.06	0.38	0.51	0.03
MgO	1.34	0.65	1.18	4.26	3.87	1.42	6.72	5.24	0.65	5.28	6.37	0.64
CaO	1.65	0.85	1.78	9.93	8.05	2.34	12.43	10.51	2.48	13.61	12.90	0.77
Na ₂ O	0.28	0.19	0.25	0.10	0.32	0.25	0.89	0.73	4.60	0.06	0.05	0.22
K ₂ O	10.25	5.35	4.23	1.64	2.87	4.26	5.77	5.44	5.03	0.56	3.93	3.76
P ₂ O ₅	0.06	0.09	0.40	0.82	0.83	0.41	0.04	0.03	0.49	1.14	0.04	0.08
LOI	5.34	4.96	5.78	18.57	17.44	6.52	20.59	18.51	1.30	23.90	22.80	4.64
Total	99.79	99.54	99.59	99.74	99.64	99.52	99.86	99.82	99.74	99.72	99.40	99.65
Dis. F. 1	-6	-1	1	8	1	1	-6	-2	3	4	-2	-3
Dis. F. 2	7	0	-3	-7	-2	-2	0	0	7	-6	-4	-2
Ni	9.1	7.0	12.4	32.6		13.0	11.5	82.1	30.4	59.5	4.2	2.3
Cr	6.2	11.9	29.2	52.8		24.6	4.7	419	n.d.	37.2	2.4	19.4
V	14.5	19.5	74.9	177		75.8	7.1	45.6	81.8	200	29.4	17.3
Sc	2.9	3.3	5.8	15.0		6.3	4.2	1.4	1.8	18.5	8.7	1.7
Cu	5.5	4.3	5.0	9.3		6.0	6.5	5.7	8.3	14.8	17.0	5.5
Zn	53.1	30.6	64.0	163		68.0	87.5	112	74.0	118	110	23.9
Pb	8.3	7.8	12.2	21.3		11.1	10.0	8.6	4.4	6.9	17.3	8.7
Sr	34.3	33.3	124	87.3		160	64.9	51.7	93.0	101	549	115
Rb	144	86.8	206	35.3		200	114	149	62.0	11.2	123	89.0
Ba	385	247	244	182		249	164	167	277	119	1685	263

Nb	129	132	100	66.6	112	82.6	147	86.7	62.5	91.8	149
Zr	438	679	421	184	476	152	886	580	242	379	667
Y	45.6	67.6	42.3	21.4	48.8	17.2	55.2	44.6	50.1	35.7	52.2
La	104	111	78.5	55.0	95.3	53.0	121	73.3	57.0	66.5	133
Ce	206	221	136	126	152	110	240	178	128	132	254
Nd	87.0	93.8	66.2	61.8	68.1	52.7	97.5	57.4	63.5	52.9	96.4
U	1.2	1.9	2.7	0.4	3.0	0.5	4.4	0.8	1.0	1.6	1.1
Th	15.1	15.1	8.8	3.7	9.1	11.1	19.1	13.8	6.9	9.6	15.1

Dis. F. 1 = -1.773TiO₂ + 0.607Al₂O₃ + 0.76Fe₂O₃^T - 1.5MgO + 0.616CaO + 0.509Na₂O - 1.224K₂O - 9.09

Dis. F. 2 = 0.445TiO₂ + 0.07Al₂O₃ - 0.25Fe₂O₃^T - 1.142MgO + 0.438CaO + 1.475Na₂O + 1.426K₂O - 6.861