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## **MIOCENE PHREATOMAGMATIC MONOGENETIC VOLCANISM OF THE WAIPIATA VOLCANIC FIELD, OTAGO, NEW ZEALAND**

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# Miocene phreatomagmatic monogenetic volcanism of the Waipiata Volcanic Field, Otago, New Zealand: Field Guide

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**Abstract** This one day pre-conference field trip's guide includes a detailed summary of the eruption style, mechanism and landform evolution of the phreatomagmatic monogenetic volcanoes of the Waipiata Volcanic Field in Otago. The trip will concentrate on demonstrating the basic types of preserved phreatomagmatic volcanoes (from eroded tuff rings, to maars and exposed diatremes). Participants will visit some informative sites, and discuss current scientific problems associated with phreatomagmatic monogenetic volcanism in intraplate terrestrial settings including signs of magmatic complexity in small-volume volcanoes (monogenetic versus polygenetic nature of volcanism) and the potential use of erosional remnants of phreatomagmatic monogenetic volcanoes in landscape evolution models for broad regions. The trip will focus on outcrops in the Pigroot Hill area, in the context of a trip from Dunedin to Oamaru with overview stops for examination of erosion remnants of monogenetic volcanoes in the schist-tor landscape near Middlemarch, and in the Maniototo Basin.

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## Introduction

Small-volume volcanic eruptions are commonly associated with monogenetic constructional volcanic landforms such as tephra cones, tephra rings, or tephra mounds consisting of bedded pyroclastic deposits emplaced by fallout, density currents and/or by downslope remobilization of tephra (Connor et al., 2000; Valentine and Gregg, 2008; Vespermann and Schmincke, 2000). Monogenetic volcanic fields often include deposits related to explosive eruptions driven by violent magma-water interaction (phreatomagmatism) where intruding magma encounters shallow or deep groundwater and/or surface water sources (White, 1991a). The amount and availability of surface- and ground-water sources to fuel magma and water (water-saturated sediment) interaction influence the type of volcanic landforms constructed, and can be affected by seasonal weather changes (Aranda-Gomez and Luhr, 1996; Auer et al., 2007; Németh et al., 2001). For these reasons a great variety of volcanic landforms can develop, especially in low lying areas, or where the hydrogeology or/and rheology of the country rocks are complex (e.g. soft versus hard substrate) (Auer et al., 2007; Németh et al., 2008a; Ort and Carrasco-Nunez, 2009; Sohn and Park, 2005). The resulting volcanic landforms in such settings are strongly dependent on the nature of the pre-eruptive surface, the lithology and mechanical properties of

volcanic conduit wall rocks, vent geometry, and the availability and type of external water.

Erosion exposes the internal architecture of monogenetic volcanoes, revealing volcanic lithofacies that provide important information on the eruptive mechanisms involved in construction of the volcanoes (Keating et al., 2008; Valentine and Keating, 2007; Valentine et al., 2005; Valentine et al., 2006). Monogenetic volcanic fields commonly include large numbers of edifice clusters and/or alignments that may include hundreds of single maars or cones (Connor and Conway, 2000; Connor et al., 2000; Conway et al., 1998).

The age-distribution pattern of centres in a monogenetic volcanic field can be used for probabilistic eruption forecasting, determining the structural architecture of the volcanic field as a whole, and understanding the overall plumbing system feeding the volcanic field (Condit and Connor, 1996; Connor, 1990; Connor and Hill, 1995; D'Orazio et al., 2000; Magill and Blong, 2005; Mazzarini, 2004; Mazzarini et al., 2008). Over the life of a volcanic field (thousands to millions of years) individual edifices may erode significantly, leading to a diminishing number of variably eroded landforms preserved on a gradually degraded landscape (Németh et al., 2007a; Németh et al., 2007b; Németh and White, 2003b; White, 1991b). Over longer time periods, a relatively uniform

landscape can be dissected, lowered and commonly inverted, showing preserved clusters of formerly low-lying structures preserved as elevated remnants (Németh and Martin, 1999). In particular, erosion during a field's development may result in remnants of volcanoes of similar age forming clusters, and this may enable a geomorphic horizon representing the land surface of that time to be identified (Németh and Martin, 1999). A refined erosion history of large (hundreds of km<sup>2</sup>) volcanic areas can be reconstructed when there is a substantial number of well-dated age-clustered volcanic remnants. A successful geomorphic reconstruction of this sort requires geomorphic data on erosion trends for certain types of volcanic landforms such as scoria cones, tuff rings, tuff cones and maars (Németh et al., 2007a).

There are many ways to reconstruct the erosional profile of a particular monogenetic landform, and thereby to estimate the erosion stage of the syn-eruptive landscape on which it erupted. In ancient volcanic fields (Ma old), erosional remnants of phreatomagmatic volcanoes provide important information on the syn-eruptive country rock stratigraphy (Németh, 2001a).

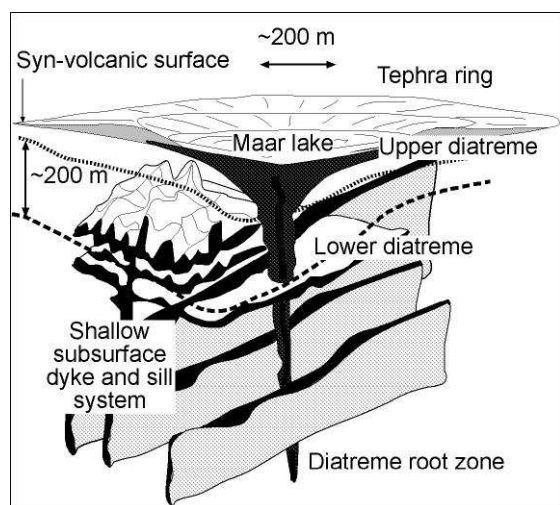


Fig. 1 – Schematic 3D section of a phreatomagmatic maar-diatreme volcano.

In maar-diatreme volcanoes (Fig. 1) a large amount of country rock is fragmented and ejected with juvenile lapilli and bombs (Lorenz, 1987; Lorenz and Kurszlaukis, 2007). The ejected volcanoclastic debris forms a tephra ring surrounding the crater of a phreatomagmatic volcano (Lorenz, 1986). In cases where the explosions take place in the shallow subsurface or at surface levels, only a wide crater may form, commonly referred as a tuff ring (Mastrolorenzo, 1994). Where substantial disruption takes place well below the surface, the ejection of material opens temporary cavities that collapse to form a subsidence feature commonly

referred as a maar (Lorenz, 1986; Lorenz and Kurszlaukis, 2007). Maars craters can be a few kilometres across and be associated with underlying diatremes also up to a few kilometres in diameter (Lorenz, 1986). Subsurface interactions of magma and water that generate tephra are attributed to thermohydraulic explosions at the base of the evolving vent, known as the root zone of a diatreme (Lorenz, 1987; Lorenz and Kurszlaukis, 2007; Lorenz et al., 2002; Ross et al., 2008). Diatremes exposed by erosion reveal the subsurface architecture of a phreatomagmatic volcano (Hoernle et al., 2006b; McClintock et al., 2008; Németh and White, 2009; Ross and White, 2005; Ross and White, 2006; White, 1991a; White, 1991b). The shape, size and fragment populations of diatremes are diverse and reflect wallrock properties and the style of magma – water interaction. Interaction is controlled by properties of the arriving magma and the rates and steadiness with which it is delivered, and the country-rock hydrology (Fig. 1). Study of diatremes can provide information about the syn-eruptive paleoenvironment of a volcanic field and the hydrogeological conditions of the strata encountered by the erupting magma. Analysis of exposed diatremes and crater-filling deposits in the Miocene Waipiata Volcanic Field in Otago has helped us better understand the evolution, eruption mechanisms and paleoenvironment of Miocene intraplate volcanism in the Otago region.

## Geological Setting

The eroded volcanoes visited during this field trip (Fig. 2) belong to the Early Miocene Dunedin Volcanic Group (DVG) (Coombs et al., 1986) and form the Waipiata Volcanic Field (WVF) (Németh and White, 2003b). The pre-volcanic Cenozoic units (Fig. 3) consist of non-marine and marine clastic sediments (Youngson and Craw, 1996; Youngson et al., 1998) deposited on an early Cretaceous erosional surface (Landis et al., 2008; LeMasurier and Landis, 1996) cut into a schist basement (Otago Schist - Os). The oldest terrestrial clastic sediments deposited on the schist form the Hogburn Formation (HF) (Fig. 3). Marine transgression followed Late Cretaceous extension and separation of New Zealand from Gondwana (Carter, 1988; Landis et al., 2008) causing widespread marine clastic sedimentation in the area (Oligocene marine sequences - OMS). These marine units are generally fine sandstones, siltstones with variable amount of glaucony. In the northern part of the Waipiata Volcanic Field, limestone beds are developed above marine clastic units (Green Valley Limestone - GL). Volcanic conduits exposed today have contacts cross-cutting this pre-volcanic marine stratigraphy (Fig. 4).

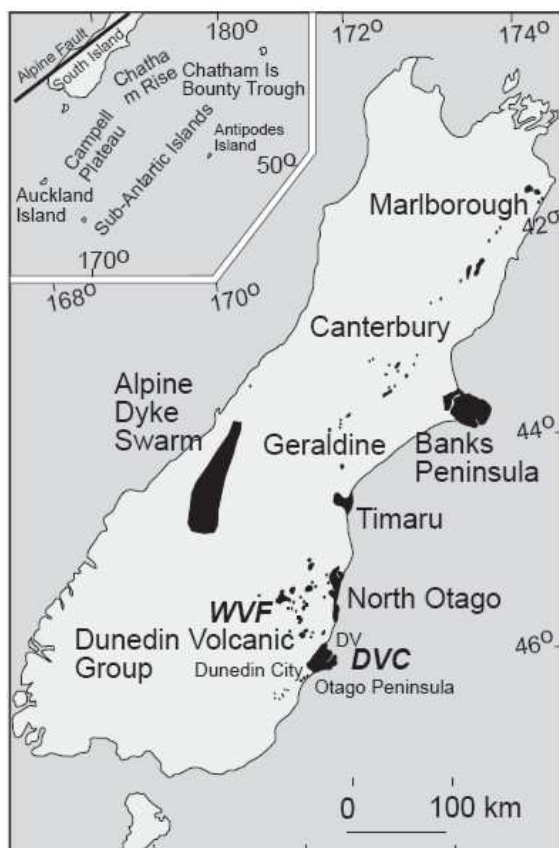


Fig. 2 – Intraplate volcanic fields in the South Island of New Zealand. DV – Dunedin Volcano; DVC – Dunedin Volcanic Complex; WVF – Waipiata Volcanic Field.

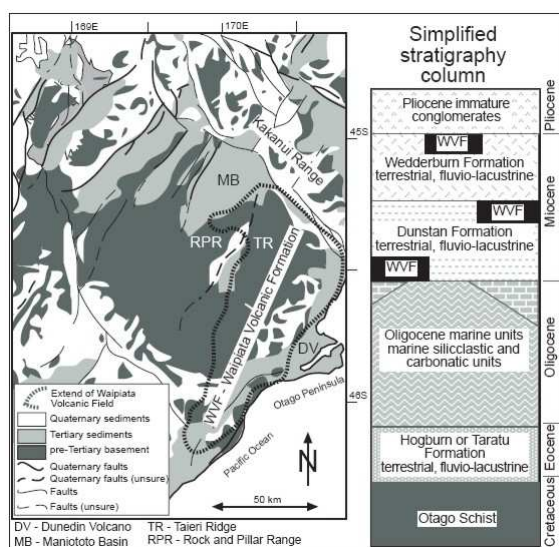


Fig. 3 – Pre-volcanic stratigraphy and sedimentary basins in the Otago region, in the location of the Waipiata Volcanic Field.

The area re-emerged in the early Miocene in response to transpressional tectonics with the inception of the Alpine Fault (Cooper, 1986; Cooper et al., 1987; Hoernle et al., 2006a; Hoernle et al., 2006b). This period is marked by terrestrial fluvi-

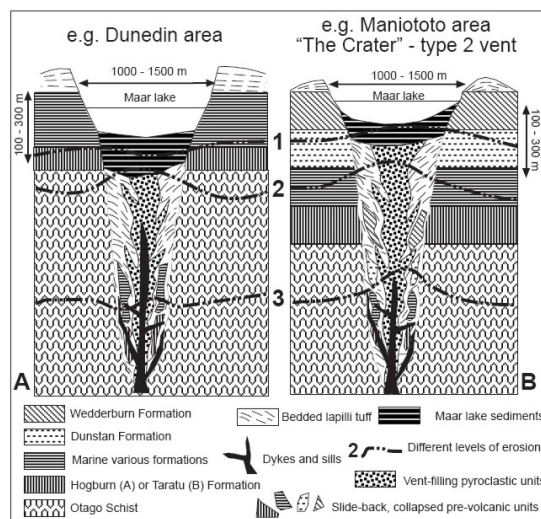


Fig. 4 – Theoretical cross-sections of maar-diatreme volcanoes cut through the pre-volcanic rock units in the Dunedin (A) and in the Waipiata (B) area.

Fluvio-lacustrine clastic deposition (Dunstan Formation - DF). Late Miocene - Pliocene uplift of ranges on the north edge of the schist belt initiated deposition of extensive conglomerates and braidplain deposits to the south of these newly risen ranges (Wedderburn Formation - WF) (Fig. 4). These deposits are overlain by the greywacke clast-bearing Wedderburn Formation and more-voluminous, immature, greywacke-dominated Maniototo Conglomerate in the northern part of the WVF (Youngson et al., 1998).

Erosional remnants of scoria cones, tuff rings and maars of WVF form a characteristic landscape topped by volcanic buttes and knobs. Cenozoic sedimentary units are preserved predominantly along the northern margin of the WVF partly due to thick lava flows. In the central, uplifted fault/fold block areas of the WVF Cenozoic sedimentary units are not preserved, and the only direct information on their previous presence or absence comes from the pyroclastic deposits preserved in various volcanic pipes (Fig. 4), which highlights the importance of studying volcanic fields like the Waipiata Volcanic Field (Németh, 2001a; Németh, 2003; Németh and White, 2003a).

### Waipiata Volcanic Field

Intraplate volcanism in the South Island of New Zealand has taken place in various styles periodically throughout the Cenozoic. In East Otago, during the Paleogene, along the present day coastline, submarine to emergent volcanism took place on the shallow continental shelf, producing Surtseyan volcanic mounds and tuff cones (Cas et al., 1989; Cas and Landis, 1987; Coombs et al., 2008; Coombs et al., 1986; Corcoran and Moore, 2008; Maicher, 2000; Maicher, 2003; Reay et al.,

2002). In the next period of volcanism, the locus of activity was further south and produced the Dunedin Volcanic Group. The Dunedin Volcanic Group traditionally includes the Alpine Dyke Swarm (ADS), the Waipiata Volcanic Formation (WVF) and the Dunedin Volcanic Complex (DVC) in terminology introduced and used by Coombs et al. (1986). The "Waipiata Volcanic Formation" is a somewhat unfortunate name from a volcanological point of view because it emphasises a "pseudo-stratigraphic" unit rather than the time-transgressive and discontinuous eruptive products themselves. It is in any case strictly applicable only to the rocks, but not to the volcanic field and its volcanic landforms. The Waipiata Volcanic Field (*also* WVF) was introduced and used to refer to a Late Miocene volcanic field adjacent to the Dunedin Volcanic Complex, produced by a common mantle anomaly that led to formation of a large number of small-volume eruptions in the Otago region (Németh and White, 2003b). Because it shares much of the same age range, and includes in its geographic boundaries, the large Dunedin Volcano shield volcano, one may take a broader view and include the entire Dunedin Volcanic Complex including small vents around the large shield (Coombs et al., 2008; Godfrey et al., 2001; Martin, 2002; Price et al., 2003) as the product of the same mantle melting anomaly/ies. Taken together the Late Miocene volcanism of the Otago region is comparable to the Cenozoic volcanic fields in eastern Australia such as the Atherton - McBride system in northern Queensland or Buckland - Mitchell provinces of central Queensland where central volcanoes and peripheral vents are viewed together as a single volcanic field (Johnson, 1989). Intrusion of the Alpine Dyke Swarm (ADS) marked the beginning of the this phase of volcanism, 28-16 Ma (Cooper, 1986). The dike rocks of the ADS are more extreme in composition than other volcanic rocks of WVF, with some dykes being carbonatitic lamprophyres (Cooper, 1986). Silica-undersaturated alkaline rocks such as phonolites and benmoreites, and more silica-rich trachytes, are also associated with the ADS (Cooper, 1986). Geophysical evidence, in conjunction with presence in the dykes of feldspathic xenoliths, indicate a crustal magma chamber, whereas the presence of upper mantle xenoliths points to a mantle origin of ADS in other areas (Cooper, 1986). The most extensive group of volcanoes formed during the periods of Cenozoic volcanism in the Otago region comprise the continental alkaline basaltic volcanoes centred on the Dunedin area, particularly considering that buried late Miocene volcanoes are inferred to be exist east of Dunedin on the Campbell Plateau (Fig. 4) (Adams, 1981; Gamble and Adams, 1985; Gamble et al., 1986; Hoernle et al., 2006b).

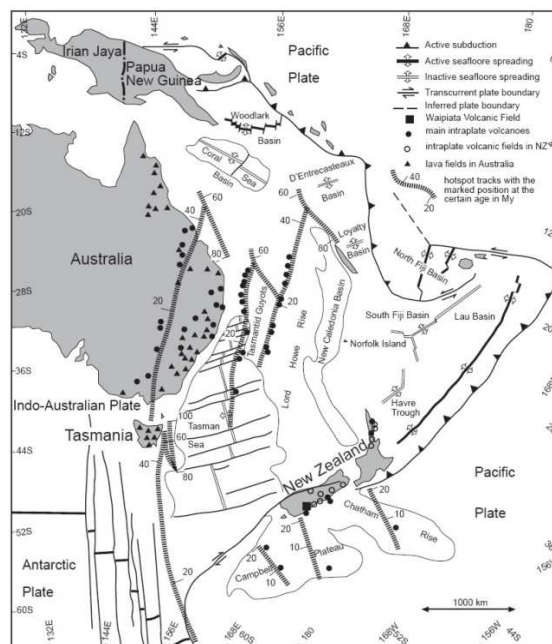


Fig. 5 – Geotectonic situation of the SW Pacific with the marked location of Cenozoic volcanoes/volcanic fields (Modified after Johnson 1989).

Large shield volcanoes similar to Dunedin Volcano exist around the South Island of New Zealand such as Carnley and Ross volcano at the Auckland Islands (Gamble and Adams, 1985; Gamble et al., 1986) or Campbell Island volcano at Campbell Island (Morris, 1984). Intraplate volcanism reported from the Antipodes Islands (Johnson, 1989) and the Chatham Islands (Eocene - Oligocene/40 - 35 Ma and Miocene - Pliocene/6 - 2.6 Ma) (Morris, 1985) east of the Dunedin off shore.

Intraplate volcanics in Eastern Australia have been inferred to be closely related to passing hotspots, because the ages of the volcanics seem to correlate with the inferred positions of certain hotspots (Johnson, 1989)(Fig. 5). A similar relationship is not demonstrable in the New Zealand region, however, although hot spot tracks have been inferred both on the Campbell Plateau and on the Chatham Rise (Johnson, 1989) (Fig. 5). Early dates for major Cenozoic alkali-basalt shield volcanoes on the continental lithosphere of southern New Zealand and the Campbell Plateau suggested ages decreasing from northwest to southeast (25 to 0 Ma) in concert with Pacific plate motion over stationary hotspots (Johnson, 1989), but this apparent age progression does not hold for other tholeiitic-basalt lava-field occurrences in the southern NZ region, and is inconsistent with long-lived volcanism at single sites (6 million years for the Dunedin Volcano). Recent work on the intraplate volcanoes of the South Island, Chatham Rise and Campbell Plateau instead suggests a volcanism occurring sporadically over a large area throughout the Cenozoic, with no

recognizable lineament array (Hoernle et al., 2006b). The cause of volcanism is inferred to be a result of some sort of lithospheric removal with resultant melting of upwelling asthenosphere (Hoernle et al., 2006a).

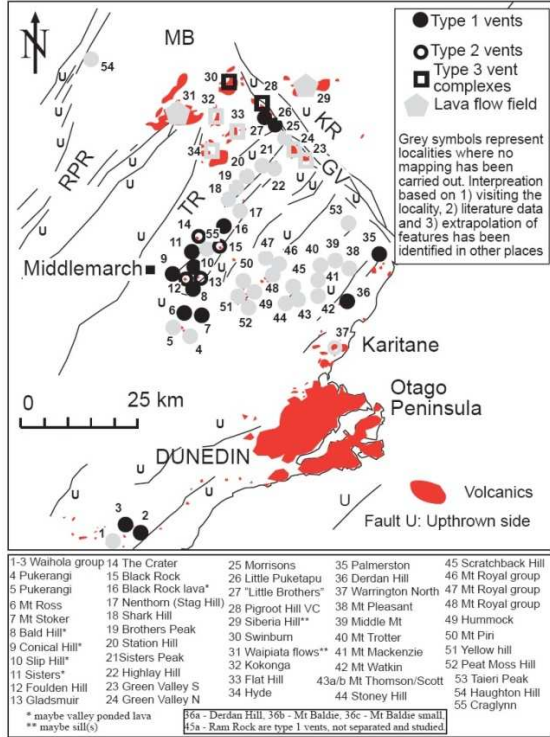


Fig. 6 – Volcanic erosional remnants within the Waipiata Volcanic Field.

In the WVF, three types of volcanic erosion remnants are distinguished (Németh and White, 2003b) according to their volcanic facies

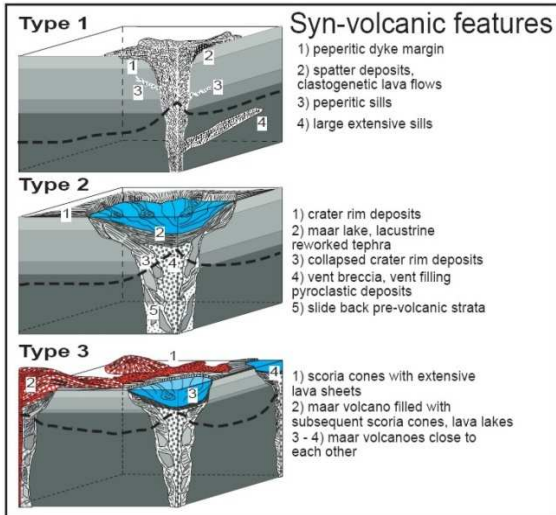


Fig. 7 – 3D reconstruction of the original volcanic landforms the three distinguished volcanic erosional remnants could be associated with (after Németh and White 2003).

associations, ratio of preserved pyroclastic and lava

units, and the size and number of identified eruptive centers (Fig. 7).

Type 1 vents are remnants of single-vent monogenetic volcanoes and consist predominantly of feeder dykes or/and remnants of lava lakes and/or lava flows (Fig. 8). Type 1 vents are concentrated on elevated parts of fold and/or fault blocks. Two major types of type 1 vents can be distinguished; 1) vent remnants filled completely with lava, 2) vent remnants filled with minor pyroclastic deposits which are cross-cut by feeder dykes, covered by solidified lava lakes, and/or coincide with the source sites of lava fields. Thin scoriaceous pyroclastic deposits are often accompanied by thick (>>10 m) lava piles, and it is inferred that these eruptive centers had an explosive eruptive history initially, and very likely represent deeply eroded remnants of former Strombolian scoria cones.

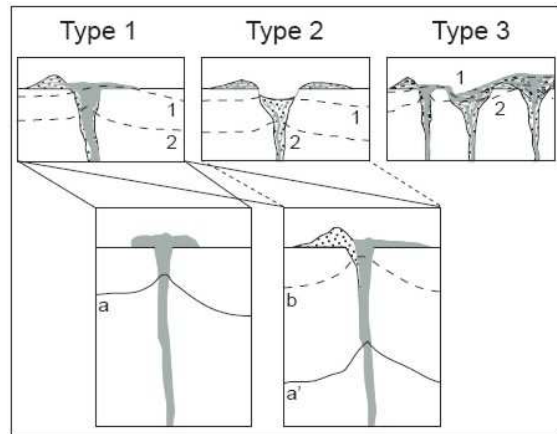


Fig. 8 – Type 1 vents and their potential link to an original volcanic landform.

Type 2 vents are remnants of individual volcanoes and consist predominantly of pyroclastic rocks and/or small-volume lava flows (Fig. 9). Type 2 vents are located in the same areas as type 1 vents. Typical pyroclastic facies are non-volcanic-lithic-rich, grain- or matrix-supported, massive tuff breccias and lapilli tuffs that consist predominantly of accidental lithic blocks and lapilli. These facies are often interbedded with unsorted, matrix-supported, diffusely stratified tuff breccia and lapilli tuff beds. These rocks are volumetrically dominant in these vent remnants, and are interpreted to be vent-filling pyroclastic units formed by phreatomagmatic explosive eruptions. Thinly bedded, locally scour-fill cross stratified, accidental lithic-rich lapilli tuff and tuff beds form decimetre-scale blocks with uniform bedding dips toward the centre of the vent. The blocks are enclosed within other pyroclastic rocks, and indicate syn-eruptive collapse and/or subsidence (Lorenz, 1971; Lorenz et al., 1970). The sedimentary textures of the facies suggest deposition from base surges. Most of the

rocks of the facies contain fresh sideromelane and just a small proportion of tachylite, the proportions indicating phreatomagmatic fragmentation of the magma. The vent remnants are topped with pyroclastic units rich in lava spatter, and these are inferred to indicate exhaustion of the water supply to the explosion sites and hence cessation of phreatomagmatic fragmentation. A sudden increase in mantle-derived xenoliths in the capping units suggests relatively rapid upward movement of the magma in later stages of the eruptions.

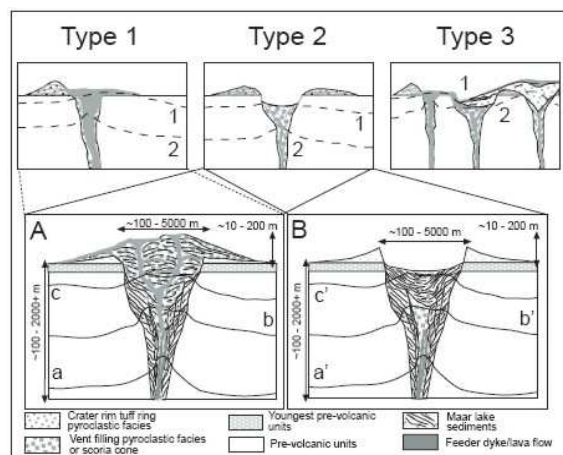


Fig. 9 – Type 2 vents and their potential link to an original volcanic landform; A) maar filled with scoria cone, B) maar filled with lake (after Németh and White 2003).

Type 3 vent complexes are remnants of complex terrestrial volcanoes that consisted of overlapping maars, tuff rings and scoria cones accompanied by extensive lava lakes and/or valley-filling lava flows (Fig. 10). These volcanic centres often had simultaneous eruptions from more than one of their vents, producing intercalated beds of explosive and effusive products. Hence, a type 3 vent complex is a group of coalesced type 2 +/- type 1 vents. Type 3 vent complexes are preserved to high stratigraphic levels, and locally include deposits formed on the paleo-ground surface adjacent to the volcano; they thus represent the best preserved, least eroded volcanic remnants in the field, in which shallow-level and surficial complexities can still be studied. The main criteria used to recognize type 3 vent complexes are: 1) close relationships among neighbouring (hundreds of metres) vents, 2) presence of lava flow units sourced from more than one site. Type 3 vent complexes tend to be located on the northern side of the WVF. The visible pyroclastic infillings of vents in type 3 complexes are very similar to deposits described for type 2 vents, but with a wider variety of facies. Some of these (for instance tuff units characterized by dune-bedding, and/or accretionary lapilli, and/or vesiculated tuff) are typical of medial

to distal deposits of phreatomagmatic tuff rings. Different types of scoriaceous, structureless to weakly bedded, tuff breccia, lapilli tuff, and tuff beds are preserved in thick piles (>10 m) as capping units, indicating that “dry” magmatic explosive phases preceded or accompanied the effusion of lava. Clastogenic lava flows are commonly preserved in dish-like structures. Spatter-rich pyroclastic beds are intercalated with thinly bedded sideromelane-rich pyroclastic beds, suggesting simultaneous deposition from magmatic and phreatomagmatic activity at closely spaced vents (Aranda-Gomez et al., 1992; Houghton and Hackett, 1984; Houghton and Schmincke, 1986).

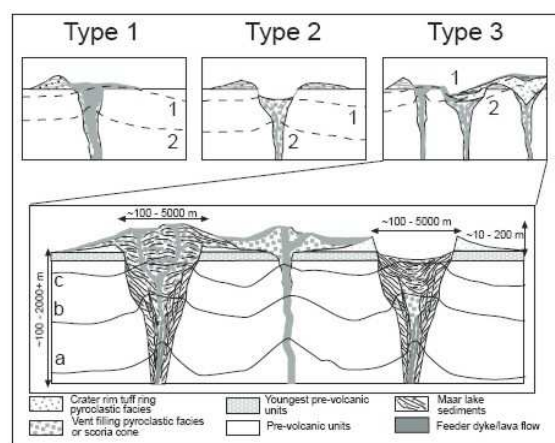


Fig. 10 – Type 3 vent complex and its potential link to an original volcanic landform(after Németh and White 2003).

Identified vent alignments at WVF mimic the major structures of the Otago area, which are predominantly NE–SW and NW–SE-oriented faults and NW–SE shear zones (Fig. 11).

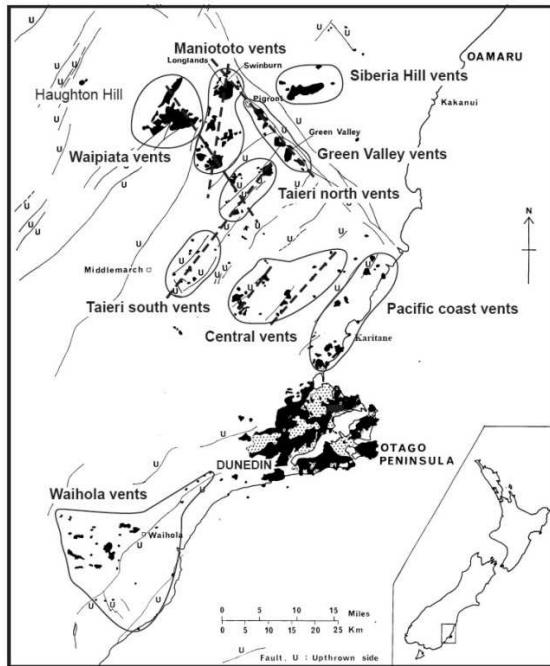


Fig. 11 – Major vent groups and alignments of the WVF.

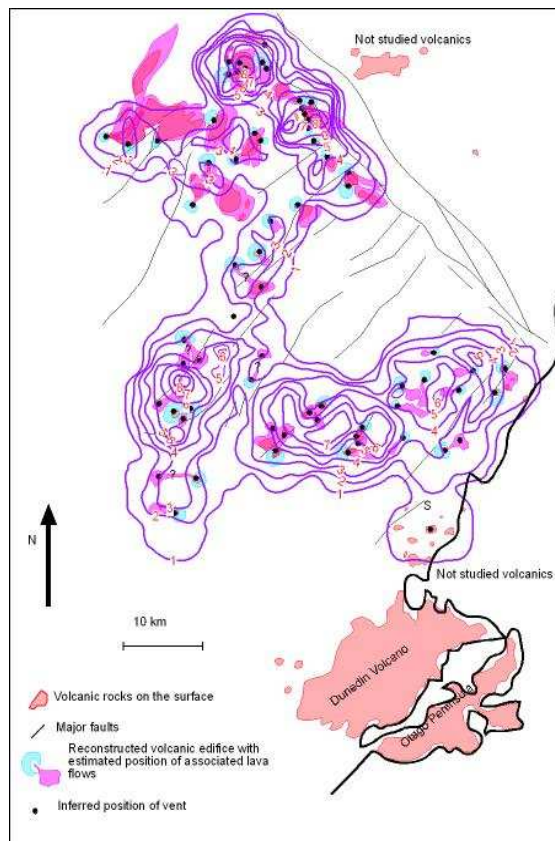


Fig. 12 – Vent-density map of the WVF.

Vents also tend to form clusters, as shown by contouring vent density using rectangular grids with a spacing of 2 km and a search radius of 5 km (Fig. 12). This vent-density map is somewhat

generalized due to uncertainties in the reconstruction of the precise number of vents and their positions, especially of the type 3 vent complexes. There is a significant vent cluster at the northern margin of the WVF, and vent clustering is also apparent in the central zone of WVF.

The total magma erupted in the WVF was calculated by estimating the total juvenile fragment volume for each volcano on the basis of the estimated volume of its lava and pyroclastic rocks. Based on different estimations the juvenile material produced by all volcanoes of the WVF may range from 9 to 40 km<sup>3</sup> (Németh, 2001b).

### Age of Volcanism

The earliest lava flows of the WVF were erupted at 21 Ma during a period of mild crustal extension related to the opening of the Tasman Sea and the separation of New Zealand from Gondwana (Coombs et al., 1986; Coombs and Reay, 1986). The major volcanism of this time occurs about 120 km east of the Alpine Fault, with only minor alkaline volcanism immediately to the east of the Alpine Fault, and with no volcanism between the two areas. All volcanic activity in the Otago area ceased at about 10 Ma as a possible result of the change to compressional tectonics. In fact, intraplate volcanism in the South Island of New Zealand still commenced after 10 Ma producing large shield volcanoes and subsequent monogenetic volcanoes at the Banks Peninsula near Christchurch which area seems to have been active between 11 and 5.8 Ma leading to an accumulation of around 1800 km<sup>3</sup> volcanic rocks (Barley et al., 1988; Stipp and McDougall, 1968; Weaver and Sewell, 1986; Weaver and Smith, 1989). This would imply that either 1) there is no direct relation between the cessation of intraplate volcanism and switch to compressional tectonics or 2) the switch to compressional tectonic regime was slightly delayed in comparison to the Otago region. The latest view on the tectonic regime switch to compressional regime indicates that it has occurred around 6.4 Ma (Sutherland, 1995; Walcott, 1998). The volcanic rocks are frequently in close proximity to known, recently reverse faults, and lineaments (Coombs et al., 1986; Coombs and Reay, 1986).

Radiometric age datings produced huge number of data in the past 50 years. Earlier three K/Ar age data published from the WVF shows that volcanism extends as far back as 15-16 Ma (Swinburn, Longland Station, 30<sup>st</sup> locality on Fig 6), and is as recent as 12.8 Ma (Green Valley, 24<sup>th</sup> locality on Fig. 6) (McDougall and Coombs, 1973). Subsequent K/Ar age datings confirmed an age of 13.4 +/- 0.3 Ma (Youngson et al., 1998) suggested that the volcanism in the WVF may have been



relatively long-lived. A poorly constrained age of 21 Ma from a drill core from 50 km from Haughton Hill suggested that volcanism may have started far earlier than it is expected in the WVF (Youngson et al., 1998). A recent synthesis summarized fifty-six previously unpublished K-Ar ages for the Dunedin Volcanic Group, previously published K-Ar and newly obtained  $^{40}\text{Ar}/^{39}\text{Ar}$  ages of Hoernle et al. (2006), and concludes that the volcanic activity in the Dunedin Volcano (e.g. as a lithostratigraphic unit) lasted from  **$16.0 \pm 0.4$  to c.  $10.1$  Ma**, and the volcanic activity of the WVF lasted from  **$24.8 \pm 0.6$  to  $8.9 \pm 0.9$  Ma** (Coombs et al., 2008). The peak of volcanic activity in the WVF climaxed at c. 16-14 Ma, exactly when volcanic activity of the Dunedin Volcano was beginning (Coombs et al., 2008). The WVF may also have outlasted the Dunedin Volcano by c. 1 m.y. (Coombs et al., 2008), though this is probably within the margin of uncertainty for the newly presented K-Ar dates obtained in the 1980's.

### Petrology and Geochemistry

Whole-rock geochemical analyses demonstrate the exclusively alkalic nature of the WVF and Dunedin rocks, with the Waipiata Volcanics being more strongly alkalic than most of the mafic members of the central volcano around Dunedin (Coombs et al., 2008; Coombs et al., 1986). As for other intraplate Cenozoic volcanism in the New Zealand region, rock compositions ranging overall from tholeiitic to highly alkalic, major- and trace-element patterns support an origin from a garnet-bearing ocean island basalt source region with high U/Pb mantle characteristics (Coombs et al., 2008; Coombs et al., 1986).

The outlying vents, which produced the deposits comprising the Waipiata Volcanic Formation of Coombs et al. (1986), contain the majority of lherzolite xenolith-bearing lava flows of the Dunedin Volcanic Group. Within Dunedin Volcanic Complex only few lherzolite xenolith-bearing lava flows have been identified, such as at Murdering Beach (McIntosh, 1989) (Martin, 2000) and Taiaroa Head (Martin, 2000). Lherzolite xenolith-bearing lava flows in the Waipiata Volcanic Field are subaerial, of limited extent, and are generally located along major faults that control the topography of the fold ranges (Coombs et al., 2008; Reay et al., 1991; Reay and Sipiara, 1987). Although olivine, orthopyroxene, chromian diopside, and spinel xenoliths are the dominant inclusions, websterites and gabbros are also relatively common, and fragments of the underlying quartzofeldspathic schist and Tertiary sediments are ubiquitous (Reay et al., 1991). A feature common to all lava flows is the presence of megacrysts, predominantly Ca-tschermakitic clinopyroxenes with varying

combinations of amphibole, anorthoclase, and spinel (Reay et al., 1991). Lava flows of the WVF, which are rich in lherzolite nodules, are all nepheline normative ranging from just undersaturated to evolved rocks with as much as 28% normative nepheline (Reay et al., 1991). All the WVF lava flows plot on the differentiation index versus  $100\text{An}/(\text{An}+\text{Ab})$  diagram [classification diagram by (Coombs and Wilkinson, 1969) close to the fields defined by Coombs & Wilkinson (1969) for the rocks from the Dunedin Volcano. A series of oxide plots against MgO exhibits trends consistent with fractionation controlled by the removal of olivine and pyroxene. The two most evolved lherzolite nodule-bearing lava flows are amongst the world's most strongly differentiated basaltic rocks derived directly from the mantle, containing 9.05 and 11.28% total alkalis at <49% SiO<sub>2</sub>, and with DI of 55 and 62, respectively (Reay et al., 1991). The most primitive WVF lava flows contain lherzolite nodules (Stoney Hill, 44th locality on Fig. 6). The lava has the second highest Mg number, the lowest abundance of the LILE, and the highest abundance of most of the High Field Strength Elements of all WVF lavas. The most basic members of WVF lherzolite-bearing lava flows show similarities to lava flows from Victoria (Australia) (Frey and Green, 1974).

The range of composition shown by the lherzolite-bearing lava flows of WVF suggests crystal fractionation was the major process controlling differentiation, and since all the rocks contain mantle xenoliths, the fractionation must have occurred at high pressure (Reay et al., 1991). The reason for this argument is that the xenoliths would settle out if the magma stopped moving upward, therefore xenoliths must be sourced from the site of magma origin/differentiation from depth (high pressure).

The Stoney Hill (locality 44 on Fig. 6) lherzolite-bearing lava flow is the most primitive of the WVF, and has been modelled as the product of 19% fractionation of a primary basanitic magma, the fractionation being dominated by aluminous pyroxene with minor olivine and titanomagnetite (Reay et al., 1991). The two most evolved lava flows must represent even greater degrees of fractionation under upper mantle conditions. The calculations suggest that these "mafic phonolite" lava flows are the product of over 60% crystal fractionation of the basanite parent melt. The fractionating phases are 37% clinopyroxene, 14% kaersutite, 4.6% spinel, 4.5% olivine, and 1.6% apatite. The modelling demonstrates that the lherzolite-bearing lava flows of WVF can be produced by the upper mantle fractionation of a primary basanitic liquid, the dominant fractionating phase being aluminous clinopyroxene, with olivine being a major accessory

phase during early fractionation, and kaersutite becoming important during later stages (Reay et al., 1991).

Electron microprobe studies of volcanic glass from initial phreatomagmatic pyroclastic deposits of many vents and vent complexes of the WVF clearly demonstrate that the pyroclastic products are significantly more evolved than any of the associated, subsequent lava flows or dykes (Németh and White, 2003a; Németh and White, 2003b; Németh et al., 2003b) (Fig 13).

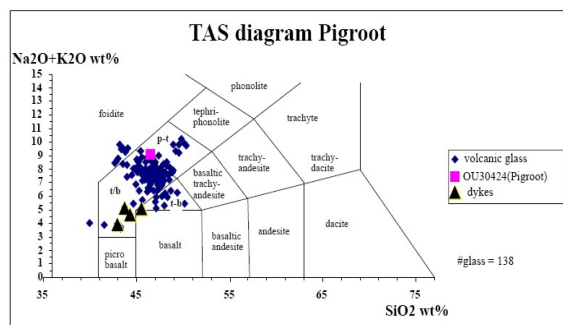


Fig. 13 – TAS diagram of samples from the Pigroot Hill volcanic complex (Type 3 vent complex) of the WVF.

Among lavas, basanites predominate, containing many lherzolite and other mantle-derived xenoliths; olivine nephelinite, mugearite and nepheline hawaiiite are also present (e.g., Coombs et al. 1986, 2008). In general, rock types more evolved than nepheline hawaiiite appear to be rare, although nepheline mugearite and lherzolite-bearing “mafic phonolite” (phonotephrite) are also known (Price and Coombs, 1975). In contrast to the unevolved compositions of lava flows, volcanic glass shards from initial phreatomagmatic units were determined to be tephrite or phonotephrite (Németh et al., 2003b). Discrimination diagrams for individual vents show well-separated fields for volcanic glass and lava; the rocks are interpreted to be related to each other via crystal fractionation, predominantly controlled by settling of olivine and clinopyroxenes (Fig. 14). Because the early-tephrite, late-basanite pattern is present at many individual volcanoes, magma evolution must have taken place in some sort of magmatic plumbing system for each of these individual monogenetic volcanoes (Figs 15, 16, 17) (Németh et al., 2003b). Overall, it can be concluded that the recorded geochemical variation characterizes volcanoes throughout the Waipiata Volcanic Field, regardless of their position within the field or degree of complexity (Németh et al., 2003b).

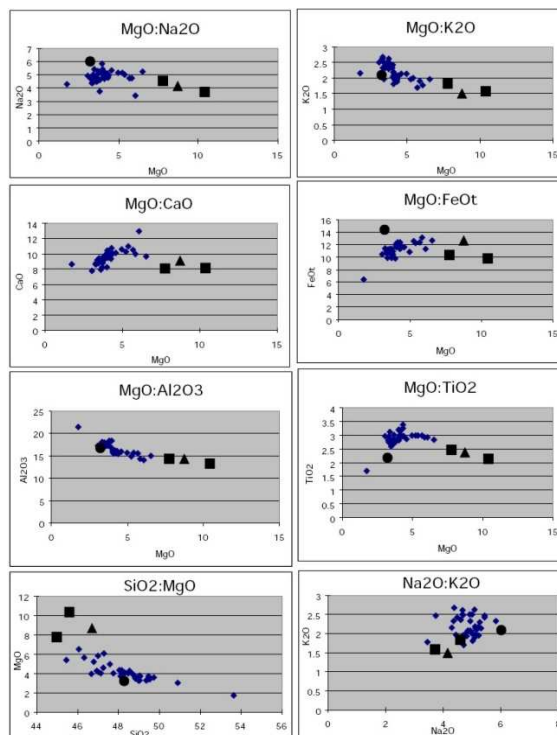


Fig. 14 – Discrimination diagrams of major element composition of volcanic glass shards from early phreatomagmatic successions and subsequent dykes and sills of The Crater diatreme (Type 2 vent).

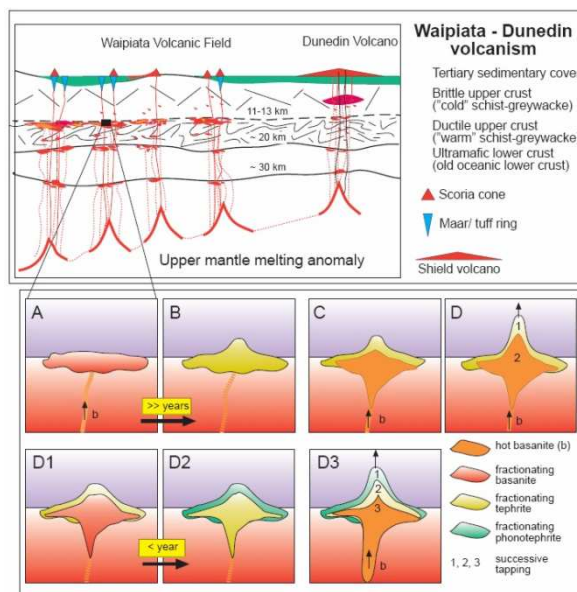


Fig. 15 – A theoretical model for the magma evolution of the individual volcanoes of the WVF.

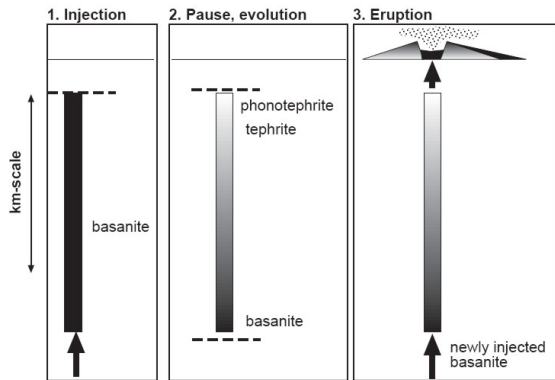


Fig. 16 – Trapped basanite melt evolves to tephrite or phonotephrite. Newly injected basanite pushes evolved melt ahead, so initial eruptive products are evolved, followed by more primitive products.

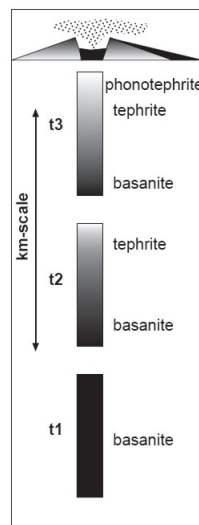


Fig. 17 – Gradual evolving of melt in the melt column.

## Erosion Calculation

Long-term erosion rate is defined as the erosion of a prevolcanic landscape "preserved" over a timescale of millions of years by the record in eruptive products of small, terrestrial volcanoes. Estimation of different stages of erosion of volcanoes is possible with the identification of preserved pyroclastic facies in the eroded remnants of volcanoes as demonstrated from the Mio/Pliocene Bakony- Balaton Highland Volcanic Field in western Hungary (Németh and Martin, 1999). Relative "erosional depth" of exposed volcanic units can then be estimated. The background elevation (Hb) is a reference elevation around the volcanic remnants. This is the elevation of the area (km-scale) surrounding the erosional remnant. The elevation of the pre-volcanic/volcanic contact is abbreviated by Hp (pre-volcanic surface elevation), and the elevation, hills capped by various rocks, by Ht. An X value is introduced to correct the

calculated values of elevation difference based on measurable elevation data. Estimating X is the most critical value in the calculation of erosion rates, and possible criteria for estimation of X can be given (Németh, 2001a; Németh, 2003; Németh and Martin, 1999; Németh et al., 2006)

Studying erosional remnants of scoria cones, tuff rings and maars is useful for paleogeographical reconstruction. Erosion of these volcanoes follows general stages (Table. 1).

	Stage	Characteristics	Calculation
Scoria cone d ~ 50 - 2000 m ~10 - 400 m	1	Distinct hills of bedded scoriaceous lapilli tuff, tuff breccias and agglomerate are characteristic	$H_p - H_b$
	2	Butte structure is not wider than few 10 - 100 m. Near-vent pyroclastics are exposed.	$H_t - H_b + X$ $X \sim 10$ m scale
	3	This stage is represented by exposed feeder dykes. Unbedded, coarse-grained,	$H_t - H_b + X$ $X \sim 100$ m scale
Tuff ring d ~ 500 - 2000 m ~10 - 100 m	1	This stage is represented by outcrops of bedded to non bedded pyroclastic rocks. Coarse-grained, unsorted, weakly bedded lapilli tuffs crop out in the central part of the	$H_t - H_b + X$ $X \sim 0$
	2	Buttes are smaller than stage 1 buttes. Pyroclastics are predominantly coarse-	$H_t - H_b + X$ $X \sim 10$ m scale
	3	Buttes are smaller than stage 2 buttes. Shallow seated lithic clast-rich pyroclastics	$H_t - H_b + X$ $X \sim 100$ m scale
Maar d ~ 100 - 5000 m ~100 - 2000 m	1	This stage is represented by few hundred metres wide butte structures. Exposed maar lake suspension sediments in the central areas are intercalated with coarse-grained	$H_t - H_b + X$ $X > 0$
	2	Maar lake deposits are already eroded, vent-filling pyroclastics are exposed. Collapsed maar crater rim sequences can be identified	$H_t - H_b + X$ $X \sim 100$ m scale
	3	This stage is represented by small dissected outcrops of vent-filling pyroclastic rocks and crosscutting feeder dykes. Dykes often have	$H_t - H_b + X$ $X \sim 10$ to 1000 m scale

Table 1 - Steps of erosion on terrestrial monogenetic volcanic landforms, recognition criteria, and general calculation of the erosion based on these landforms. CR - crater rim pyroclastic units; VF - vent-filling pyroclastic units; LD - lava flows and dykes; ML - maar lake sediments; PV - pre-volcanic units

Scoria cones are constructional landforms, in contrast to maars, which are local sediment traps. Tuff rings represent a transitional form between scoria cones and maars from a geomorphic point of view. Scoria cones are relatively quickly degraded on a geological time scale (few hundred thousand years) and erosion leaves only a welded scoriaceous pile of pyroclastic rock or lava flows. Rim beds of tuff rings and maars can be eroded as fast as scoria cones unless they are capped by lava flows. Only the deep parts of maars and tuff rings are preserved after a few million years of erosion. In this late stage of erosion it is generally hard to distinguish erosional remnants of tuff rings and maars, though the latter typically has higher contents of accidental lithic fragments including originally deep-seated accidental lithic clasts. The calculation of erosion can be summarised as requiring three major steps: 1) the original landform has to be reconstructed based on available field data; 2) the possible stage of erosion should be estimated from the mapped lithofacies associations, their facies relationships, thickness, relative abundance and position, and; 3) the possible "missing" pre-volcanic units and their thickness should be established based on the accidental lithic-clast population of exposed and mapped pyroclastic units of the erosional remnants.

The calculations from the Miocene Waipiata Volcanic Field (New Zealand) indicate erosion rates of a few tens of meters per million years (Németh et al., 2003a; Németh and White, 2003a). Erosion rates based on volcanic remnants located close to a subsequently uplifted fault/fold block ("The Crater" on Fig. 18 and Black Rock on Fig. 19) gave higher values of erosion (30 - 46 m/My) than calculations based on volcanic remnants located far from an uplifted fault/fold block (Swinburn - 3.75 - 15.6 m/My) (Németh, 2001a).

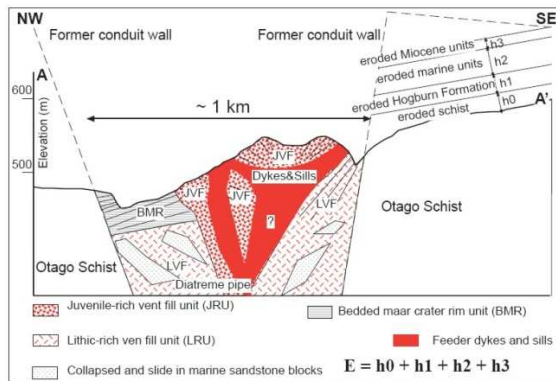


Fig. 18 – Erosion calculation model for The Crater diatreme

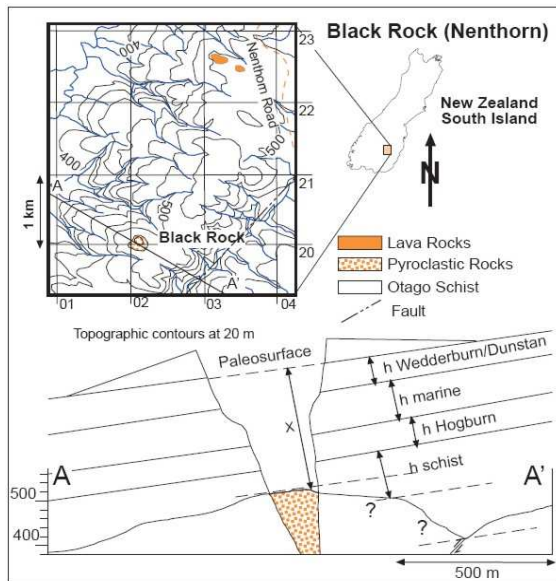


Fig. 19 – Erosion calculation model for the Black Rock diatreme

## Itinerary

*Kilometre measurements start from the BP service station in downtown Mosgiel.*

**Start 8.45 AM** Dunedin Railway Station

**Star 9.30 AM** Dunedin Airport

**Stop 1 – Conical Hill – stop next to road**  
km - 73

GPS: -45.515102°; 170.192233°

**Stop 2 – Foulden Hill lookout – stop next to road**  
km - 75

GPS: -45.521541°; 170.220631°

**Stop 3 – The Crater lookout – stop next to road**  
km - 79

GPS: -45.488225°; 170.288371°

**Stop 4 – Swinburn – stop next to road**

km - 133 (lookout); 138 (roadcut)

GPS: -45.153500°; 170.262084° (lookout);

-45.137776°; 170.323016° (roadcut)

**Stop 5 – Red Cutting Summit – stop next to road**  
km - 150

GPS: -45.193160°; 170.41995°

**Lunch break 12.00 noon – Carpark Pigroot #Hw85**  
km - 151

**Stop 6 – Trig 634 – 15 minutes walk from lunch**  
stop [alternative stop]

GPS: -45.199889°; 170.418821°

**Stop 7 – Pigroot Hill 1 – 15 minutes walk road**

GPS: -45.205428°; 170.420199°

**Stop 8 – Pigroot Hill 2 – 10' walk from Stop 7**

GPS: -45.204184°; 170.421941°

**Stop 9 – Pigroot Hill 3 – 10' walk from Stop 8**

GPS: -45.205683°; 170.423110°

**Stop 10 – Conclusion – Pigroot Hill top – 5' walk**  
from Stop 9

GPS: -45.204102°; 170.423228°

**Arrival to Oamaru 6.00 PM**

The first half (morning) of the fieldtrip will give an overview of the deeply eroded volcanic remnants and the present day landscape of the southern part of the Waipiata Volcanic Field. This part of the trip will cover about 150 km driving through Otago's scenic schist-tor landscape along the Taieri Ridge and surrounding valleys. Before lunchtime, the trip will pass through the Waipiata, Swinburn and Maniototo region, in the northern extremities of the Waipiata Volcanic Field. This part of the trip will offer superb photo opportunity of the rugged landscape of Otago. The second part (afternoon) of the field trip will focus on one of the best exposed erosional remnants of a phreatomagmatic volcanic complex (Type 3 vent complex), the eroded and deeply exposed nested maar/diatreme volcano of the Pigroot Hill. This part of the trip will involve about 6 km walking through grass and farm land with cliff faces exposing various level of a phreatomagmatic volcano. Prior to departure for Oamaru, the trip will conclude with a brief summary and explanatory presentation of the volcanic sights the participants can look at on the way to Oamaru.

## Field Guide

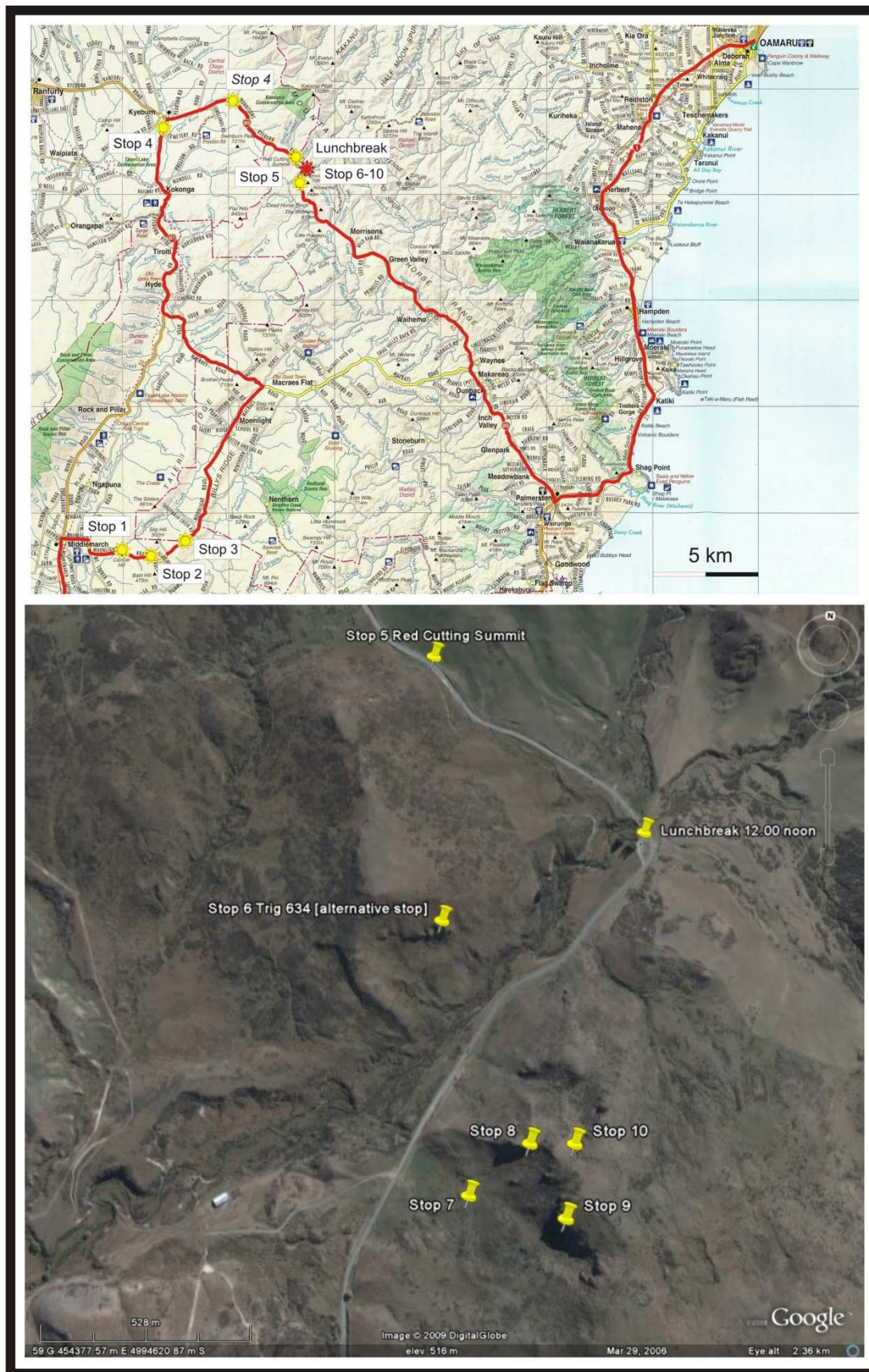


Fig. 20 – Field trip stops on KiwiMaps 250 series background (top) and on Google Earth Image (bottom).

## Stop 1 – Conical Hill – Lava Plug, Type 1 Vent

Coherent lavas are volumetrically the most important rock facies of type 1 vents. They are often columnar jointed and rarely platy jointed. Joint systems of columnar jointed lavas can be very different from place to place. Lava can be radial columnar jointed, with the columns extending radially from the axis of lava bodies; cross-sectional dimensions of the columns gradually diminish inward from outer surfaces. Well-developed radial columnar jointed lava is present at Swinburn (NZMS 260 I42/989585), and at Trig 634 (NZMS 260 I42/068529) within vent complexes (type 3 vent complexes), and slightly less well developed radial columnar jointing characterizes some lavas from the type 1 vents "Little Brothers" (NZMS 260 I42/086496), Little Puketapu (NZMS 260 I42/092479), and Conical Hill (NZMS 260 I43/904175). The bulk of type 1 vents have jointed lava with 3-, 4-, 5-, 6-, and 7-sided columns. The column diameters range from ~5 cm to 55 cm. It was not possible to determine the pattern of jointing in columnar jointed lava sheets over large (tens of metres) scales because outcrops are of limited extent. Regular, well-developed wide columns are more characteristic for the basal zones of exposed lava flows among type 1 vents (e.g., Mt Watkin - NZMS 260 I43/205128, Mt Mackenzie - NZMS 260 I43/201167, Slip Hill - NZMS 260 I43/939189, Mt Stoker - NZMS 260 I43/955087, Palmerston - NZMS 260 I43/290211). The tops of most of the type 1 vents are capped by lava with thinner, less regular columns commonly having complex structures and orientation (e.g., top of Little Puketapu - NZMS 260 I42/093478, Sister - NZMS 260 I43/935218, Bald Hill - NZMS 260 I43/936154). In general, no significant composition or textural differences have been identified among columnar jointed lava units. The well-developed columnar jointed parts of lava units seem to be slightly coarser grained than the marginal, less developed columns within the same unit.

A coherent nephelinite lava at **Conical Hill** (Fig. 21) has radial columnar joints with a spacing of up to 30 cm. There are no remnants of sedimentary rocks or peperitic margins within individual joints. The lava is topographically and stratigraphically above weathered schist. The contact between schist basement and lava is not exposed, but it is inferred to be steep. There is no evidence of Cenozoic rock between the schist and lava. There is no further stratigraphic unit above the lava cap.

Coherent nephelinite lava at "Little Brothers" near to Pigroot Hill volcanic complex is a fine-grained, microcrystalline black, grey, non-

vesicular rock with upper, radial (entablature) and lower, vertical (colonnade) columnar joints. There are places where large numbers of vertically aligned clasts, derived from prevolcanic units, are enclosed in the lava. Around non-volcanic inclusions "flow banding" and/or areas of elongate vesicles occur. Many large schist fragments were captured in un-vesiculated melt grading to vesicle-rich outer melt zones. This suggests that capture of the xenoliths cooled the melt prior to vesiculation, and hence must have occurred at some depth. Peridotite lherzolite and/or megacrysts are rare and usually small in size (<cm). Mantle nodules are generally concentrated in the same areas as other, crustal, lithic fragments. Coherent lava facies in places clearly demonstrate transition between coherent to clastic rock facies indicating that the erosional remnant is a former vent/conduit of a volcano produced explosive eruptions. **Conical Hill** is the basanite-filled conduit of a former volcano that lacks identified pyroclastic deposits (Fig. 22).



Fig. 21 – Conical Hill lava plug



Fig. 22 – Transition from coherent magmatic to fragmented clastic rock facies from the Little Brothers vent suggesting that Little Brothers is a vent produced fragmented rocks during its activity. Such samples have not been identified yet from Conical Hill.

Coherent mafic lava at Little Puketapu is very similar in texture to lavas from "Little Brothers", and also encloses crustal lithic fragments in well-distinguished zones as well as megacrysts and/or small (<cm) peridotite lherzolite xenoliths.

The lava is irregularly, radially columnar jointed (entablature) in the exposed areas.

Thick basaltic lavas, especially lava lakes comprising most of the circular lava fields of type 1 vents commonly develop a typical zones of jointing with 1) a very regular, well-developed vertical jointed basal zone (colonnade), 2) an irregular medial zone with thinner columns than in colonnade (called entablature) and 3) a similar but less developed vertical jointed zone as in the basal zone (called upper colonnade).

The reason that the described and identified lava units are predominantly topped by an irregular zone, or just a very regular columnar jointed zone in the top of the remnants is that 1) upper colonnades were eroded so that irregular entablature is exposed in the top of the remnant (e.g., Little Puketapu) or 2) both the upper colonnade and the middle entablature are eroded and only the lower colonnade remains. Therefore in areas where predominantly colonnade zones are exposed (no other zones have been identified and it can be inferred that none were originally present) it can be inferred that those zones represent the lower zone of a lava flow exposing the lower colonnade of e.g. ponded lava flow.

Unfortunately the interpretation of type 1 vents where only a lava flow is exposed can be difficult, with ambiguous results. This ambiguity is pronounced at **Conical Hill** (Fig. 23) where only a small (ten metres scale) lava cap is preserved in exposure.

At **Conical Hill**, three equally valid interpretations can be given of the origin of the lava-capped butte because there is no information due to lack of exposure of the style of contact between prevolcanic and volcanic units ;

1) reconstructing an inferred sharp and steep contact with pre-volcanic schist, the absence of any Cenozoic sediment between schist and lava, the present elevation of the schist/lava contact, and the radial columnar jointing of the lava suggest that the lava represents an eroded remnant of a lava plug. The plug cuts schist basement, thus it is inferred that erosion must have down-cut to the level of schist, stripping away all the younger overlying units. It is not possible to reconstruct whether the crater rim of the original volcanic landform was breached and lava flowed over to the surrounding areas or lava flows were confined to its narrow conduit and/or vent.

2) since there is no field evidence of pyroclastic units around the Conical Hill it is impossible to know whether Conical Hill, represents the feeder dyke of a former explosive eruptive centre or is a remnant of an intruded dyke which might have or might have not reach the surface. The depth of erosion and absence of any outcrop linearity argues against this possibility.

3) it is also should be considered that a small butte like Conical Hill could be a fortuitously preserved segment of a ponded lava flow that filled a paleovalley. Because there are no Cenozoic units between the lava flow and the basement schist, if Conical Hill is a remnant of a ponded lava flow bottom zone, the contact should represent a paleosurface, and would also imply a deep valley cut through the whole of the Tertiary section at this site. The lack of any outcrop or geomorphological linearity here argues against this interpretation.

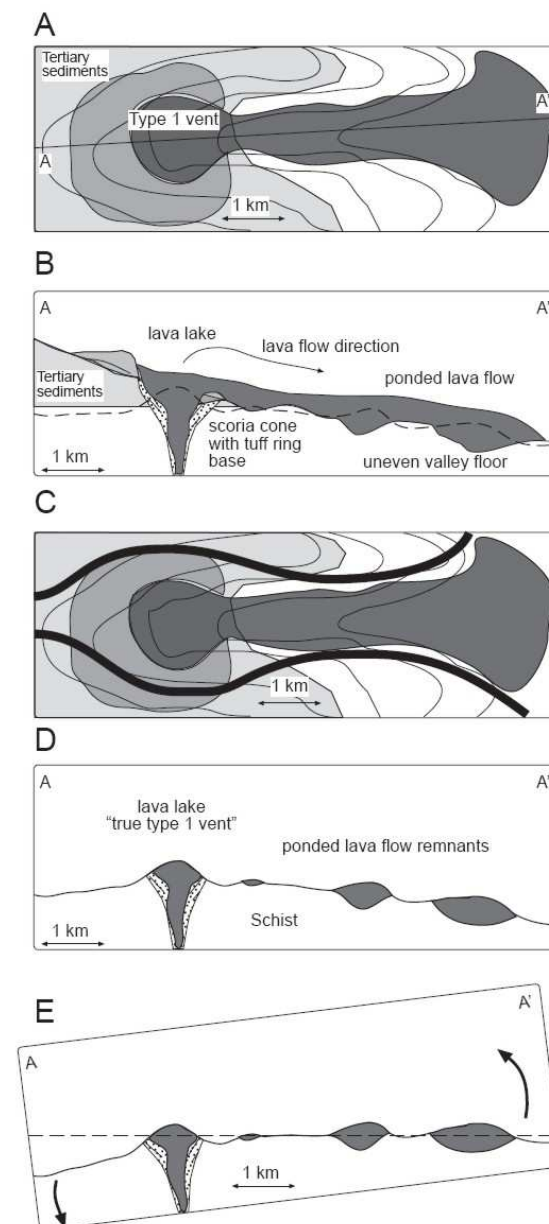


Fig. 23 – 3D relationship between “true” vents and erosional remnants of ponded flow bodies. A – original syn-eruptive morphology, B – cross sectional view, C – potential fluvial reworking (thick line), D – erosional surface, E – tilting effect

These three solutions for the Conical Hill highlight the possible reconstruction of a lava capped sequence in the WVF. In the case of Conical Hill the first interpretation seems to be the more realistic, because 1) Cenozoic units are known to have covered the area at the time of eruption, though they may not have been very thick (tens of metres scale?) and 2) to preserve a very small (few tens of m<sup>2</sup>) lava cap from a big ponded lava flow unit is rather unusual (but not impossible).

There are large number of individual plugs in the Dunedin Volcanic Group, Waipiata Volcanic Field e.g., The Sisters (NZMS 260 I43/935217) and Mt Watkin (NZMS 260 I43/205130), that are similar to Conical Hill.

### Stop 2 – Foulden Hill Maar Lookout, Type 2 vent

The **Foulden Hill** area includes a down-faulted semicircular depression on the Otago schist surface up to 1000 m in diameter (Figs 24, 25). There is only one very limited location exposure of Cenozoic sediments inferred to be part of the Pakaha Group (non-marine and/or shallow marine deposition; close to the westward limit of Cretaceous-Palaeogene marine transgression) in the vicinity of the depression. There are no pyroclastic rocks known around Foulden Hill (Lindqvist and Lee, 2009). The depression is mostly filled by diatomite (Lindqvist and Lee, 2009; Travis, 1965). Drilling of the diatomite showed that at least 80 m of sediment occur in the depression at Foulden Hill (Gordon, 1959a; Gordon, 1959b). Newly obtained seismic reflection surveys suggest that the preserved thickness of the diatomite could exceed 100 metres (Gorman et al., 2006). Gordon's (1959) original drill report describes 280/18 dip direction of bedding of the diatomite at near surface. The north-westerly dip increases with depth. Ripple marks on bedding planes were reported (Gordon, 1959a; Gordon, 1959b). The recovered diatomite is dominantly siliceous diatom casts (90 %) (Travis, 1965). An ongoing study (Lindqvist and Lee, 2009) of the diatomite in exposed section reveals two depositional facies; 1) thinly laminated and 2) dark brown speckled beds. The speckled beds are interpreted to represent sediments deposited by gravity flows from the crater wall similar to other crater lakes (Németh et al., 2008b). Layers of quartz grains, micaceous silt, spinel, palagonite and brown vesicular pellets were reported earlier (Travis, 1965). Fossils other than diatom tests (*Cymbella* and *Navioula*) are common from the Foulden Hill diatomite (Travis, 1965), predominantly from the thinly laminated facies (Lindqvist and Lee, 2009). Macrofossils include a large variety of well preserved leaves (Lindqvist and Lee, 2009), fossil

flowers (Bannister et al., 2005), fish (*Galaxias*) (Lee et al., 2007) and various insects (Kaulfuss et al., 2008).

The general characteristics based on earlier descriptions suggested that the Foulden Hill diatomite is a maar crater containing a lacustrine sequence. The changing (steeping) dip values suggest continuous down-sagging of a maar basin during lacustrine sedimentation which is widely reported from other recent or ancient maar lakes (Jámbor and Solti, 1976; Mingram et al., 2004; Mrlina et al., 2009; Németh et al., 2008b; Pirrung et al., 2008). Deformation of lacustrine strata can also result from the post-eruptive subsidence and re-settling of the underlying diatreme (Lorenz, 2007; Suhr et al., 2006).

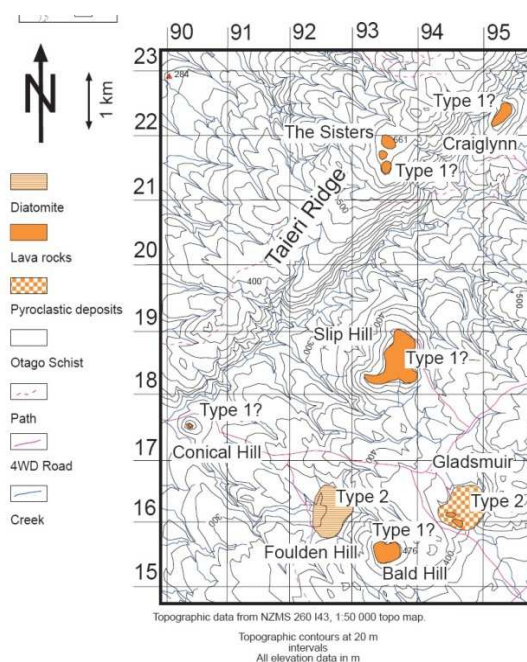


Fig. 24 - Location of Foulden Hill and other Waipiata volcanic rocks on the surface.

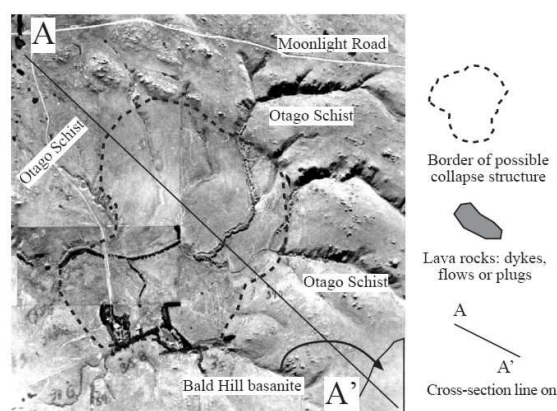


Fig. 25 – Airphoto of the Foulden Hill depression.



The original floral community reconstructed from fossils from the Foulden Hill diatomite indicates that it was probably moderately diverse and growing on a nutrient-rich substrate (Pole, 1996). On the basis of the floristic assemblage, the recovered macroflora and fauna, and the style of sedimentation in the region inferred at the time suggest that the Foulden Maar was located in a low-relief (coastal) plain under warm to sub-tropical conditions with seasonally dry, maritime climate (e.g. Lindqvist and Lee 2009 and references therein).

This deep infilled cavity in the Otago Schist shares characteristics of other nearby examples (e.g., Gladsmuir, "The Crater"), which are of clearly volcanic origin, and by analogy it was inferred that the Foulden Hill itself represents a former maar/diatreme eruptive centre. There is no direct radiometric age data from volcanic sediments encountered at the base of recent drill holes, but a coherent magmatic body exposed within the outline of the inferred structure recently yielded a  $^{40}\text{Ar}/^{39}\text{Ar}$  age of 23.2 Ma (Lindqvist and Lee, 2009). While this age is ambiguous with respect to the Foulden Maar, it is in the age range other volcanic rocks in the area (Coombs et al., 2008; Lindqvist and Lee, 2009). A similar interpretation as infill of a maar crater was given for the Hindon diatomite ~ 25 km west from Dunedin, which underlies another depression on the Otago Schist surface (Youngson, 1993).

The present-day bottom of the valley at Foulden Hill is about 120 metres below the schist surface at nearby Bald Hill (Fig. 24). The known thickness of the diatomite is at least 80 m (Gordon, 1959a; Gordon, 1959b), consistent with recent geophysical surveys (Gorman et al., 2006), thus at least 120+80=200 m of rock must have been excavated during the phreatomagmatic explosions and the ongoing and/or subsequent down-sagging at Foulden Hill (Fig. 26). The 200 m crater depth fits with empirical data for other maar craters' depth and width (Mertes, 1983). Usually maar crater depth, immediately after eruption and before compaction, is approximately a fifth of the crater diameter (Lorenz, 1986; Mertes, 1983). An at least 1000 m wide crater most likely a good estimate, thus a 200 m deep maar crater is also quite plausible.

At this stop, a discussion about the 3D landscape reconstruction will be initiated. The nearby coherent magmatic bodies as well as the preserved phreatomagmatic pyroclastic rocks (Gladsmuir) indicate a very complex syn-eruptive scenario, which could be reconstructed in many ways (Figs 26, 27).

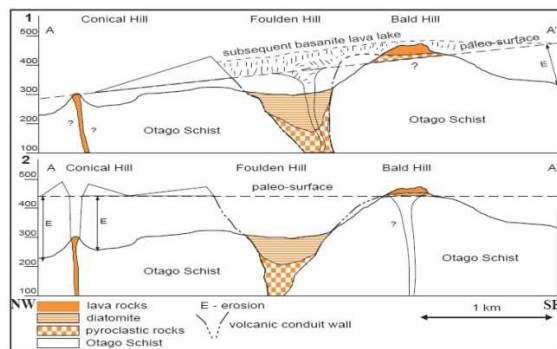


Fig. 26 – Foulden Hill - Bald Hill problem.

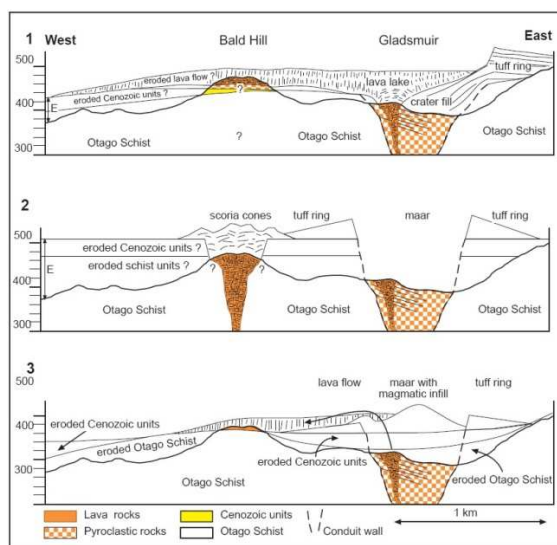


Fig. 27 – Gladsmuir – Bald Hill problem.

### Stop 3 – The Crater – Black Rock Diatreme Lookout, Type 2 Vent

In this short stop we will look at the Otago Schist-dominated landscape cut through by two diatremes (Fig. 28) (Type 2 vents). Due to complications with access we are not able to examine the rocks, but will discuss the significance of the diatremes in a syn-eruptive landscape reconstruction.

Type 2 vents are erosional remnants of monogenetic volcanic landforms such as tuff rings, tuff cones, maars, and/or Strombolian scoria cones. Type 2 vents are predominantly located in the central part of the WVF between the Pacific coastline and the westerly Rock and Pillar Range in a similar geographical position than type 1 vents. Type 2 vents are characteristically near elevated ridges in a very similar situation to type 1 vents. Type 2 vents are predominantly filled by pyroclastic rocks but occasionally small volume lava flows or

cross cutting dykes are common and very likely helped to preserve these volcanic conduits.

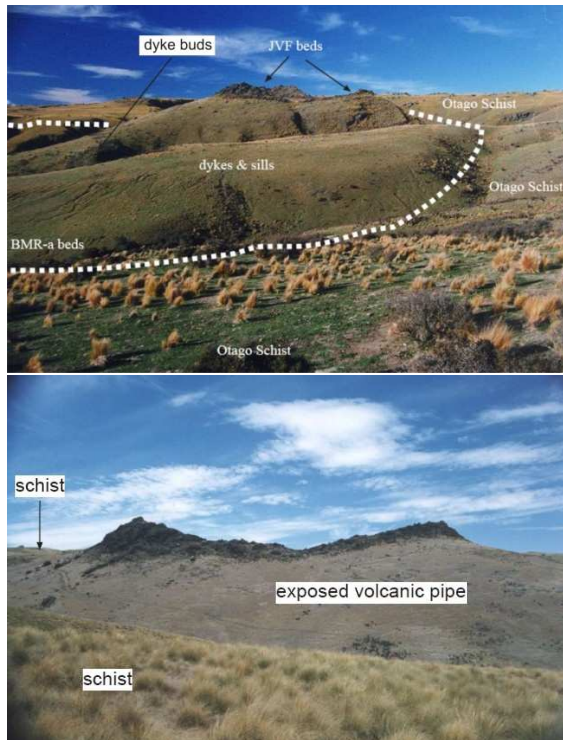


Fig. 28 – Overviews of “The Crater” diatreme.

Type 2 vents are rarer than type 1 vents which could be related to either 1) type 2 vents are already eroded completely due to the less resistant characteristics of pyroclastic rocks in comparison to lava rocks or 2) the volcanism in general produced significant amounts of lavas in most of the localities in the Waipiata Volcanic Field, therefore pyroclastic rock-dominated volcanic pipes are relatively rare.

Pyroclastic rock facies are named and referred to as stated in Table 2.

Four type 2 vents were studied in detail by Nemeth (2001). The most detailed study was carried out at “The Crater” a circular pyroclastic rock filled volcanic pipe on the Taieri ridge. The features, facies and facies associations have been identified at “The Crater” largely similar to features have been identified at Black Rock or Gladsmuir therefore “The Crater” can be viewed as a type locality for type 2 vents.

Lithofacies associations were possible to be separated at “The Crater” because “The Crater” is the only studied type 2 vents which is large and diverse enough to exhibit great variety of facies.

There are four major facies association have been identified (Fig. 29); 1) lithic-rich vent fill (LVF); 2) juvenile-rich vent fill (JVF), 3) bedded maar rim basal (BRM-a) and 4) bedded maar rim-capping (BRM-b) lithofacies association.

Volcanism related facies	Tuff breccia (TB)	Lapilli tuff (LT)	Tuff (T)
<b>Clast-supported</b>			
<b>Non-volcanic lithic-rich</b>			
1 Massive	TB1	LT1	T1
2 Weakly-bedded	TB2	LT2	T2
3 Well-bedded	TB3	LT3	T3
<b>Juvenile-rich</b>			
4 Massive	TB4	LT4	T4
5 Weakly-bedded	TB5	LT5	T5
6 Well-bedded	TB6	LT6	T6
<b>Matrix-supported</b>			
<b>Non-volcanic lithic-rich</b>			
7 Scour-fill bedded	TB7	LT7	T7
8 Channel-fill massive	TB8	LT8	T8
9 Unsorted massive	TB9	LT9	T9
10 Strongly lithified pisolitic massive	TB10	LT10	T10
11 Pisolitic	TB11	LT11	T11
12 Diffusely stratified	TB12	LT12	T12
13 Thinly bedded	TB13	LT13	T13
14 Cross-stratified	TB14	LT14	T14
15 Undulatory-bedded	TB15	LT15	T15
16 Dune-bedded	TB16	LT16	T16
17 Inverse-to-normal graded	TB17	LT17	T17
<b>Juvenile-rich</b>			
18 Scour-fill bedded	TB18	LT18	T18
19 Channel-fill massive	TB19	LT19	T19
20 Unsorted massive	TB20	LT20	T20
21 Strongly lithified pisolitic massive	TB21	LT21	T21
22 Pisolitic	TB22	LT22	T22
23 Diffusely stratified	TB23	LT23	T23
24 Thinly bedded	TB24	LT24	T24
25 Cross-stratified	TB25	LT25	T25
26 Undulatory-bedded	TB26	LT26	T26
27 Dune-bedded	TB27	LT27	T27
28 Inverse-to-normal graded	TB28	LT28	T28

Table 2 – Lithofacies table used to distinguish pyroclastic rock types across the WVF.

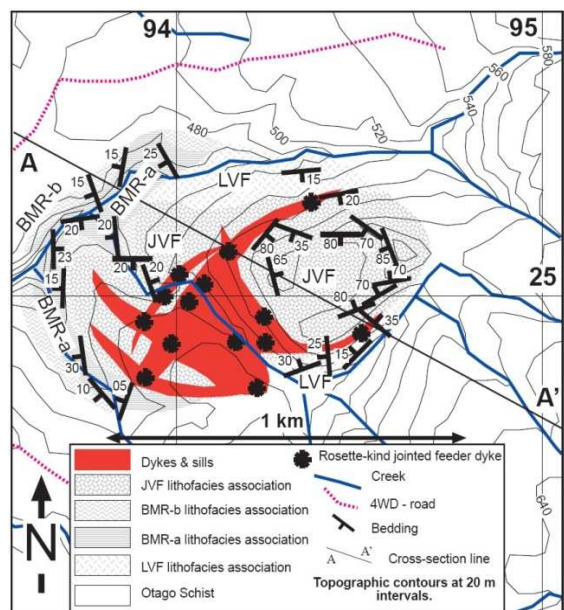


Fig. 29 – Volcanic lithofacies distribution of “The Crater” diatreme

The presence of a large number of intact, large (up to 35 cm in diameter) marine sandstone fragments in the pyroclastic rocks of "The Crater" suggest intensive un-roofing of these beds, causing significant slump back processes. The finely dispersed, fine grained quartzofeldspathic grains and marine deposit derived glauconite (Fig. 30) suggest that these marine beds must have occurred in the vent zone in a relatively homogeneous distribution and probably in a (semi-) loose state. Blocks (mega-blocks) of the marine deposit most likely were slide back to the vent zone due to repeated explosive activity then they were recycled and reworked by each individual explosion, producing a homogeneously dispersed distribution in the deposited pyroclastic rocks.

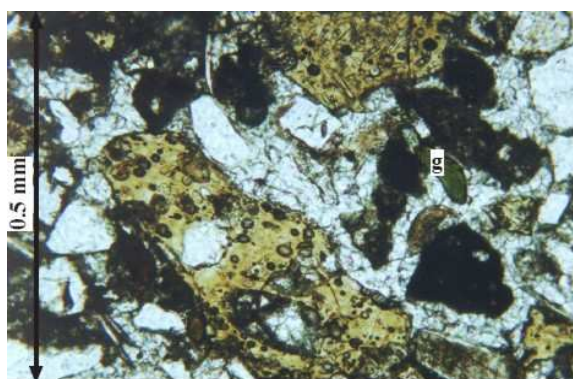


Fig. 30 – Sideromelane glass shards and glaucony in a fine grained tuff from "The Crater" diatreme.

LVF represents a vent filling succession, a possible lower diatreme, where diatreme is a pipe like volcanic conduit filled with pyroclastic debris (tuff and lapilli tuff) and blocks of wall rock. Diatreme deposits remain poorly sorted - which is the case in this lithofacies - because tephra falling back into the vent from the eruption column is emplaced at the base of the funnel shaped vent. White (1991) separated three zones in a diatreme: 1) root zone, which is a mixture of hypabyssal intrusive bodies and un-bedded tephra, 2) lower diatreme, un-bedded tephra deposited within the vent, 3) upper diatreme, bedded tephra deposited in the upper level of some vents.

According to a classification of White (1991) LVF represents a lower diatreme zone. Characteristic features of lower diatreme include near-absence of bedding, a crudely concentric arrangement of rock types, and common sub-vertical lithologic boundaries (White, 1991b). Larger schist fragments seem to arrange a concentric distribution along the lower level of the sequence along the steep contact zone with a country rock (schist) and the weak bedding characteristic seems to support the lower diatreme origin of the deposits.

BRM-a lithofacies association is a series of matrix-supported, accidental lithic-rich, bedded, weakly-to moderately sorted lapilli tuff and tuff beds. BMR-a lithofacies has been identified and mapped at "The Crater". No other type 2 vents exhibit facies grouped into BMR-a. BRM-b is a relatively uniform lithofacies association which can be separated into two stratigraphic units according to the matrix-to-other large clast size ratio; 1) basal ash-supported, fluidal bomb rich lapilli tuff unit and 2) lapilli-supported, scoriaceous, lava spatter rich lapilli tuff capping unit.

Juvenile-rich Vent-filling (JVF) is a uniform lithofacies association which crops out in the central part of "The Crater" on topographic heights (NZMS 260 I43/945252). The JVF at "The Crater" seems to cover and/or cut through LVF. There is no visible contact, but the distribution of the facies grouped into JVF implies that it must have a steep contact with earlier units. JVF has been identified in both Black Rock and Gladsmuir locality. At Black Rock the entire pyroclastic sequence could be grouped into this lithofacies association.

The volcanic lithofacies associations around "The Crater" are interpreted as primary, pyroclastic sequences which are directly related to phreatomagmatic explosive activity. In general, the volcanic lithofacies associations of "The Crater" can be grouped into two major group according to their sedimentary features. In the western side of the field in a semicircular distribution, bedded pyroclastic deposits occur mostly as base surge deposits and phreatomagmatic near-vent air fall origin, that gradually transform into beds representing Strombolian-type fall-out origin (BMR-a and BMR-b). In contrast, the centre part of the area, pyroclastic deposits exhibit mostly non- or weakly bedded characteristics, rich in accidental lithic (LVF) and/or vitric clasts (JVF). These two type of pyroclastic units have a steep dipping toward the centre part of the field, suggesting that they belong to a similar volcanic - probably collapse - structure. It is noteworthy, that the vitric-rich pyroclastic sequences in the middle part of the area (JVF) seems to cut through the other lithofacies associations (LVF and BMR). The textural and compositional characteristics of the pyroclastic deposits, their semicircular, steep bedding around the centre part of the field suggest that "The Crater" is a complex pyroclastic debris filled volcanic pipe, complex collapse structure, a diatreme similar to those ones have been described from the Hopi Buttes, Arizona (White, 1991b), Montana (Hearn, 1968) or in the Pannonian Basin (Martin and Németh, 2004).

The 3 major pyroclastic units (BMR, JVF, LVF) of "The Crater" represent a complex history of phreatomagmatic to magmatic explosive eruption,

with intermittent collapse and subsidence of bedded crater rim sequences into active vents. The BMR-a and BMR-b lithofacies association exhibit a sequence of a crater rim beds formed by phreatomagmatic explosive eruptions transforming into magmatic explosive eruption. The latest stage of this eruption was clearly magmatic explosive, forming Strombolian scoria beds inter-bedded with spatter deposits.

In contrast the central part of “The Crater” comprises a basal tuff breccia and lapilli tuff series representing a vent-opening stage of the eruption that formed accidental lithic-rich, coarse-grained pyroclastic beds by sub-surface phreatomagmatic explosions. In the top of the section in the central part of “The Crater”, presence of juvenile-rich (sideromelane) lapilli tuff beds indicate a sudden change in the vent morphology and indicate a clearer stage of an erupting vent and possible shallower level of explosion locus. The pyroclastic beds (JVF) most likely developed in an earlier formed volcanic depression (e.g. maar crater). The top-section of this unit (JVF) exhibits strongly tilted blocks with inward dip direction which indicates inward subsidence of large pyroclastic blocks in this stage of the eruptive history.

#### **Stop 4 – Swinburn, Type 3 Vent Complex Lookout and Roadcut section**

In this brief overlook stop the participants can view the extensive Swinburn lava plateau. This region is cut by at least 4 pyroclastic filled pipes inferred to be diatremes. Swinburn is interpreted to be a type 3 vent complex, a closely spaced phreatomagmatic volcanic vent complex. The individual volcanic centres are inferred to have erupted more or less simultaneously and produced a complex assemblage of explosive and effusive products with locally intercalated non-marine sedimentary rock. Hence, a Type 3 vent complex such as Swinburn is a group of coalesced type 2 +/- type 1 vents. Type 3 vent complexes are preserved to high levels, locally including deposits formed on the ground surface adjacent to the vent, and thus represent the best preserved, least eroded volcanic remnants in the field, in which shallow-level and surficial complexities can still be studied. The main criteria used to recognize type 3 vent complexes are: 1) close relationships among neighbouring (hundreds of metres) vents, interpreted by the common facies associations require to identify a vent, 2) presence of lava flow units sourced from more than one site.

Type 3 vent complexes tend to be located in the northern side of the Waipiata Volcanic Field, close to the Maniototo basin's eastern margin (Figs

31-33), and have been identified in the Swinburn area, along the northern sector of the Green Valley, and in a basinal structure between the northern end of the Taieri and Rock and Pillar Ranges near Hyde



(NZMS 260 I42/960430), Flat Hill (NZMS 260 I42/020490) and Kokonga (NZMS 260 I42/970510).

Fig. 31 – Swinburn Volcanic Complex from the North.



Fig. 32 – Typical finely bedded tuff hand sample from one of the Swinburn diatremes near Longland Station.

Fig. 33 – Overview of tilted Swinburn lavas at the northern edge of the complex. Tilting is along a still-growing schist ridge, but there is evidence of rugged syn-volcanic topography.

Each of these volcanic structures represents multiple-vent complexes including adjacent voluminous lava flows. At Hyde (Hyde Volcanic Complex - HVC) the number of individual vents is uncertain due to poor outcrop. At Kokonga (Kokonga Volcanic Complex - KVC), only limited deposits of vitric tuff and lapilli tuff suggest the presence of explosive vents in the complex. The thick and extensive lava flows at both Hyde and Kokonga are inferred to indicate both 1) significant ponding in paleo-lows, and 2) large volume of erosion.

**Stop 5 – Red Cutting Summit, Base Surge and Phreatomagmatic Fall, Type 3 Vent Complex**

The Pigroot Hill Volcanic Complex is located approximately 60 km west of Palmerston along the Pigroot Highway (Fig. 20). The area is close to uplifted greywacke of the Kakanui Range, and is marked by hillsides covered with volcanic rock and sloping generally north-eastward. The area has 3 major hills. The northernmost is Flat Hill – (771 m abs), comprising the so called "mafic phonolite dome" (Price and Green, 1972). Southeast of Flat Hill is a smaller but well-defined hill referred to here as Trig 634 (634 m abs). It is almost entirely composed of pyroclastic rocks covered by minor lava, especially at the northwestern edge of the hill. The major morphological high of the area is a southern group of hills including Pigroot Hill (758 m asl) and Round Hill (743 m asl). These hills form a large southeastwardly sloping ridge largely formed of lava along its southeastern side, but with tens of metres of exposed pyroclastic rock cliffs in the northwest side.

The major part of the Pigroot Hill Volcanic Complex is the Pigroot Hill and Round Hill massif itself. The massif is a well-defined hill approximately 3 km long with a north to south axis. On the top of the hill two peaks consisting of radially columnar jointed feldspar-phyric nephelinite. Each hill is elevated ~40 - 60 m above the lava-covered southeast sloping surface of Scattered glauconitic limestone debris, probably pieces of Green Valley Limestone, lies at the surface on the western flank of Pigroot Hill peak. Its origin is uncertain, but it may indicate gaps in the lava that provide a source of colluvial limestone blocks. The eastern side of the massif is truncated by the Waihemo Fault.

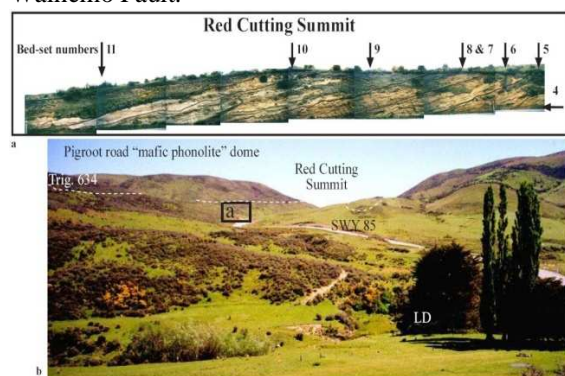


Fig. 34 – Red Cutting Summit road cut exposes a complex phreatomagmatic base surge and fall bed succession, part of a tuff ring around a phreatomagmatic volcano.

Approximately 500 m northwest from the “picnic area” a complex series of fine grained finely bedded

lapilli tuffs and tuffs (Fig. 34) is exposed along the Pigroot Highway (NZMS 260 I42/070540).

The bedding dip directions seems to be more less perpendicular or even opposite to those in the southern facing cliffs of Trig 634, making it difficult to establish the exact stratigraphy position of these beds relative to those of the southeastern areas. The most prominent features of these outcrops on the northern side are 1) accretionary lapilli rich beds, 2) vesiculated tuffs (Fig. , 35) a great variety of soft sediment deformation features (Fig 36), 4) dune bedded pyroclastic rocks (Fig. 37) and 5) thickly bedded lapilli tuffs (LT9) rich in peridotite lherzolite and/or megacrysts.

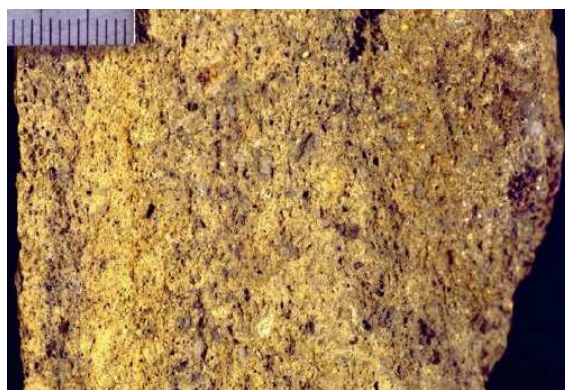


Fig. 35 – Vesicular (vesiculated) tuff.

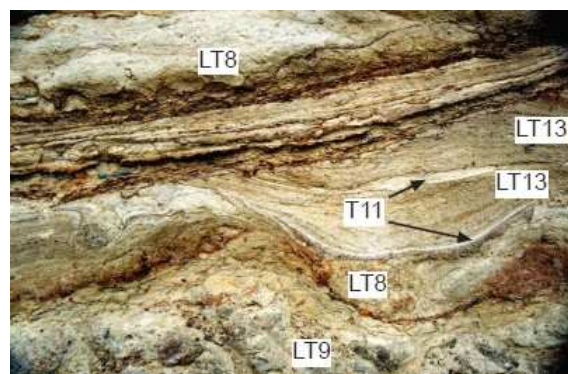


Fig. 36 – Soft-sediment deformation and impact sags in tuff and lapilli tuff beds of the Red Cutting Summit.



Fig. 37 – Dune-bedded pyroclastic beds of the Red Cutting Summit phreatomagmatic succession.

The presence of abundant soft sediment features, vesiculated tuffs and accretionary lapilli at NZMS 260 I42/070540 suggest that damp pyroclastic density currents (already condensing steam in the passing surge, about 700 metres from their source) formed the majority of those pyroclastic sequences (Fig. 37). The transportation direction inferred from dune structures and the asymmetry of impact sags indicates southeast to northwest movement of base surges and ballistic bombs at this location potentially sourced from a region close to the “picnic area”.

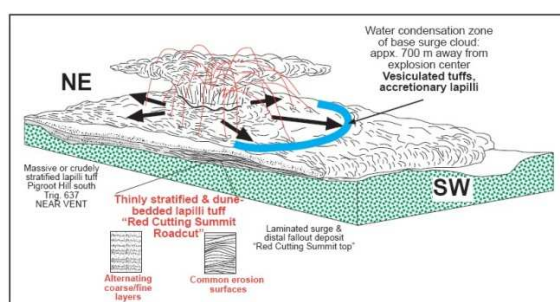


Fig. 38 – A potential 3D reconstruction of the depositional environment of the pyroclastic succession of the Red Cutting Summit road-cut.

### Stop 6 – Trig 634 – Diatreme Filling Tuff Breccia Succession, Part of a Type 3 Vent Complex [alternative stop]

Trig 634 (Fig. 39) consists largely of volcanic rocks, with Hogburn Fm. and marine sediments cropping out along the western and southern sides of the hill. Marine sandstone is exposed along Pigroot Creek, where they are cut by dykes and overlain by pyroclastic rocks.

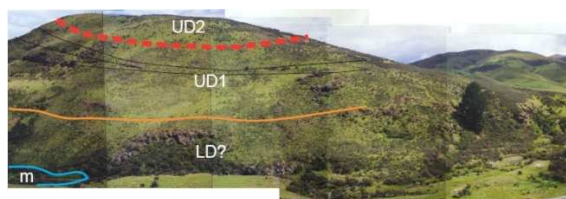


Fig. 39 – Interpretative view of Trig 634.

Pyroclastic rocks of Trig 634 also crop out in the lower and middle sections of the southeastern cliff. The pyroclastic rock along Pigroot Creek appears to be collapsed blocks, because their dip direction varies greatly from place to place. In the lower and middle section of the Trig 634 cliff is an

in situ, subhorizontal to gently northwestward dipping, pyroclastic succession, covered and/or cross cut by lava and dykes. Pyroclastic rocks crop out on the north side of Trig 634 along the Pigroot Highway (NZMS 260 I42/070540).

Three stratigraphic units (LD, UD1, UD2) have been recognised on the basis of 1) their positions relative to one another, 2) their relationships to volcanic conduits (vents), and 3) their typical juvenile (sideromelane and tachylite) and accidental lithic componentry.

Rocks of unit LD 1) do not overlie other volcanic rocks, 2) are predominantly unsorted, structureless tuff breccia or lapilli tuff facies, 3) exhibit a steep angular unconformity with pre-volcanic and other volcanic units, and 4) commonly overlain by tuff or bedded lapilli tuff. This stratigraphic unit forms the lower section of Trig 634 along the northern part of the Pigroot Creek valley (Fig. 39). The volumetrically most significant facies of this unit is LT9, which forms beds a few tens of cm to ~2 m thick. LT9 beds in this unit almost contain variable amount of accidental lithic clasts in variable size limit from few cm up to 50 cm blocks. The large blocks are randomly oriented and have no related impact sags. Smaller platy schist clasts occasionally form slightly oriented (imbricated) features. Type locality of is the lower section of the southeastern valley bottom along Pigroot Creek (NZMS 260 I42/073532), where it shows relatively uniform characteristics of the unit comprising LT9 facies occasionally inter-bedded with T16. Similar facies relations can be drawn for other places especially next to the "picnic area", where structureless undifferentiated LT9 is exposed. Unfortunately in places along the Pigroot Creek large pyroclastic rock blocks show very chaotic orientations, and it is likely that the valley is filled by slumped metre-scale blocks of pyroclastic rocks derived from the pyroclastic sequence of the eastern cliff of Trig 634. Between the "picnic area" and the NZMS 260 I42/073532 locality, a thick, uniform, sequence is exposed in the Pigroot Creek valley bottom, exhibiting uniform diffusely stratified lapilli tuffs (LT12) rich in glauconite grains and fresh, angular sideromelane glass shards. A continuation of this pyroclastic sequence can be traced upstream along Pigroot Creek for a few tens of metres (NZMS 260 I42/077537). Afterward the pyroclastic sequence seems to have a sharp contact with lignite-rich terrestrial sandstone beds that closely resemble rocks of the Hogburn Formation. The LT12 pyroclastic sequence grades laterally(?) into a thickly bedded, strongly tilted, predominantly LT9 succession similar to that exposed around the "picnic area" (NZMS 260 I42/075536). Pyroclastic rocks assigned to unit LD grade abruptly into a better bedded, finer grained lapilli tuff and tuff series

approximately halfway between the Pigroot Creek valley floor to the top of the hill on the southeastward facing Trig 634 cliffs. Bedding varies along the Pigroot Creek valley without any preferred orientation. Higher in the section the bedding, where it is clear, shows a northwestward dip of 15-25 degrees. These bedding characteristics are more prominent in the overlying UD1 unit. The volumetrically most significant facies of this unit (LT9) has been interpreted as the deposit of unsteady high-concentration pyroclastic density currents. These pyroclastic density currents must have been generated by subsurface phreatomagmatic explosions which interpretation is supported by the abundance of sideromelane glass shards and accidental lithic clasts from different pre-volcanic rock units. The structureless characteristics of the beds of this unit and the crudely stratified texture suggest deposition in a near vent setting. Unfortunately there is no exposed contact between pre-volcanic and volcanic units, so it can not be determined whether the unit lies within a vent or was deposited on the surrounding ground surface. The coarse, chaotic deposits exposed at the lowest topographic levels around the "picnic area" (NZMS 260 I42/075536) may truly represent a conduit filling unit produced by inward flowing high-concentration pyroclastic density currents and/or fall back processes.

The pyroclastic rocks of UD1 1) are in medial stratigraphic positions in pyroclastic sequences, 2) are predominantly accidental lithic and/or sideromelane shard rich, predominantly diffusely, thinly, undulatory or dune bedded lapilli tuff and tuff facies, 3) are commonly overlain by tuffs and scoriaceous coarse grained tuff breccias, lapilli tuffs, and lava flows or dykes. UD1 is represented by moderately to well bedded, non-volcanic accidental lithic-rich lapilli tuff and tuff sequence in the middle stratigraphic section of Trig 634. The total exposed thickness of the UD1 unit is estimated to be ~80 m. The UD1 unit is more diverse than the LD unit. In the southeastern facing cliff of Trig 634, the transition seems to be continuous between LD and UD1 unit. The lower part of the unit comprises structureless beds of LT9 inter-bedded with T16. The repeating sequence continuously grades into a cyclic sequence of LT12 and LT14 beds upward. Up-section a significant unconformity can be identified in the southeast facing cliff of Trig 634 (NZMS 260 I42/071533) (Fig. 40). Below the unconformity the dominant facies in the section is thickly bedded LT9, but above the unconformity a series of LT12 and LT14 occur. In each case the dip direction is approximately northward with abruptly varying dips from sub-horizontal to 35 degrees.

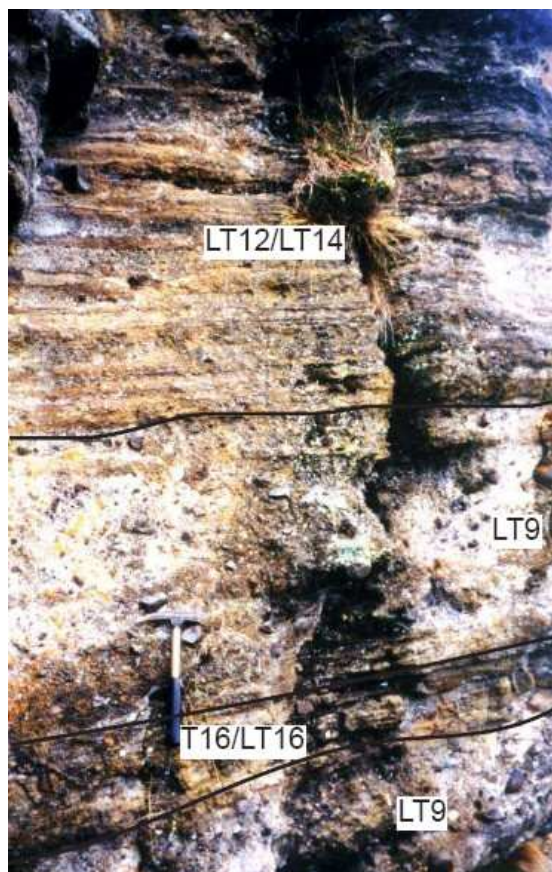


Fig. 40 – Medial section of the Trig 634 location.

In both described localities UD1 is interpreted to be a former crater rim sequence developed around an active phreatomagmatic vent.

The stratigraphic relationship between UD1 units in the southeastern cliff of Trig 634 and the northern side along the Pigroot Highway is not fully understood. The perpendicular or even opposite dip direction allows different explanations of the facies relationships between pyroclastic rocks of these two nearby localities. Two likely alternatives are considered to be: 1) the two localities are parts of a semicircular tephra ring possibly centred somewhere near the "picnic area", and the beds' changing dip direction define the radially outward inclination of the ring beds, or 2) the two localities part of an inward-dipping pyroclastic sequence within a conduit centred at Trig 634 itself. Because there is no clear evidence of steep pre-volcanic/volcanic contacts that would be predicted by second model, the UD1 sequence of Trig 634 is probably part of a crescent-shape tephra ring around a vent may be centred at the "picnic area".

Beds grouped into the UD2 unit at Trig 634 because of their common juvenile-rich (scoriaceous), clastogenic lava spatter-rich weakly or non-bedded characteristics and their highest stratigraphic position. The exposed total thickness of

the UD2 is ~40 m. These beds are covered by lava flow(s) and/or cross cut by dykes in the northwestern side of the Trig 634. The beds of UD2 dip northward in increasing values from 20 to 55 degrees. In up-section, large (metre-scale) lava spatters are more prominent. In lower stratigraphic position scoriaceous lapilli tuffs are the volumetrically dominant facies of the unit. The unit is exposed only in the southeastern side of the Trig 634. In the northern areas it is covered by lava rocks. Interpretation UD2 is a product of repeated, probably short-lived Strombolian explosive eruptions. The eruptions turned into more Hawaiian style lava fountaining in the late stage, producing lava spatter deposits, and occasionally small clastogenic lava flows. The topping lava rocks might be originated from vents produced the spatter deposits during high magma discharge rate period. There is no clear field evidence of the relation between lava rocks from the top of the Trig 634 and Flat Hill "mafic phonolite" dome, their compositional similarities and the general facies relations at Trig 634 may suggest that the two lava fields genetically may relate to each other. The relation might be as close as the source of the Flat Hill "mafic phonolite" dome can be the spatter cones of Trig 634.

### **Stop 7 – Pigroot Hill 1 – Lapilli Tuff and Tuff Breccia of a Lower Diatreme, Part of a Type 3 Vent Complex**

In the following stops a near-continuous traverse will give a good summary for the participants to see the internal architecture of a phreatomagmatic vent complex (Type 3 vent complex). While the guide separates 3 distinct stops, the outcrops and key features continuously will provide fine details of the interior of a phreatomagmatic volcano.

Pyroclastic rocks exposed in this location exhibiting juvenile-rich (sideromelane), unsorted structureless texture lapilli tuffs that are grouped into LD (Lower Diatreme) stratigraphic unit. The exact aerial distribution of this unit is unknown due to lack of continuous outcrops. Because debris is frequent in the western foothill areas in the deepest exposed areas, it is inferred that this deposits representing the basal zone of the pyroclastic cliffs in the western side of the Pigroot Hill (Fig. 41). Similar pyroclastic rocks in the northern side of the Pigroot Hill area next to the Waihemo Fault are also grouped into LD because of their unsorted, coarse grained characteristics with angular unconformity to the pre-volcanic marine sedimentary units. The pyroclastic rocks in this locality contain large number of rounded gabbroid texture lava clasts with

variable size. Pyroclastic rocks (Fig. 42) similar in texture with rocks presented in this locality are also exposed in the southern margin of the Round Hill (NZMS 260 I42/084509) and are grouped into LD because of their unsorted, juvenile clast (altered sideromelane) rich characteristics, and their possible angular unconformity with the pre-volcanic sedimentary units.



Fig. 41 – Interpretative view toward Pigroot Hill.



Fig. 42 – Typical lapilli tuff sample from the basal zone of Pigroot Hill.

These pyroclastic rocks can be interpreted as;

1) pyroclastic rocks (LT20b) of this unit are very likely to represent initial products of a phreatomagmatic eruptive activity or they are itself the lower zone of a phreatomagmatic vent filling pyroclastic sequence. Grouping these rocks into LD – lower diatreme unit is rather a try to point out of their initial product status than an interpretation to interpret them as conduit filling pyroclastic rocks.

2) Pyroclastic rocks at the northern side of the Pigroot Hill very likely represent a vent filling tuff breccia which interpretation is supported by i) the presence of large number accidental lithic clasts derived from different pre-volcanic units, ii) the unsorted, chaotic texture of the rocks, iii) the angular unconformity between pre-volcanic marine and volcanic units and iv) the presence of large number of altered and non-altered sideromelane glass shards in a same pyroclastic facies.

3) Pyroclastic rocks at the southern side of the Round Hill are very likely represent a phreatomagmatic vent filling unit because i) the



unsorted, chaotic, structureless characteristics of most of the exposed pyroclastic rocks, ii) the angular unconformity between pre-volcanic and volcanic units, iii) the presence of strongly altered sideromelane glass shards, and iv) the presence of accidental lithic fragments, mostly derived from marine units. The interpretation of the southern locality is very loose due to lack of large outcrops to determine the detailed sedimentological features of the pyroclastic rocks seem to fulfill a phreatomagmatic vent.

**Stop 8 – Pigroot Hill 2 – Upper Section of Diatrema Filling Lapilli Tuff and Tuff Breccia Succession, Part of a Type 3 Vent Complex**

In this location the medial section (UD1) of the pyroclastic succession of Pigroot Hill is presented (Figs 43, 44).

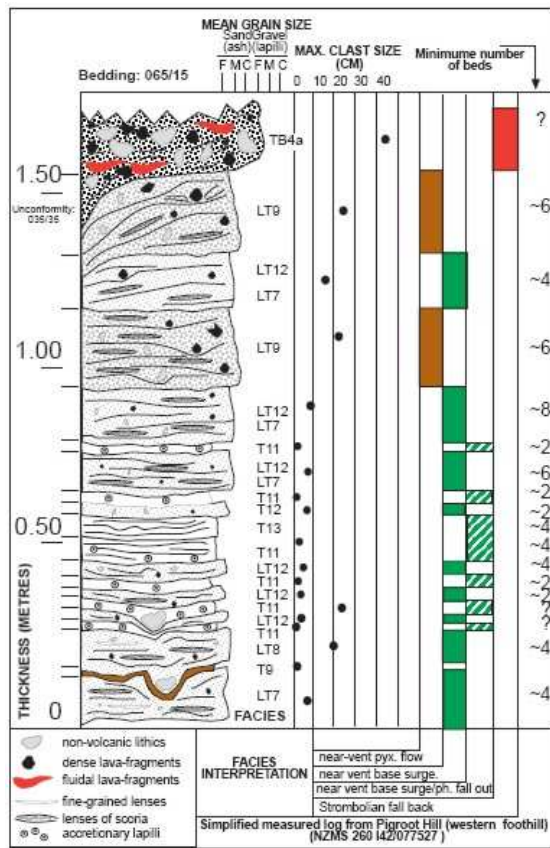


Fig. 43 – Medial pyroclastic section of Pigroot Hill.

The criteria to group pyroclastic rocks into UD1 are 1) medial stratigraphic position in a pyroclastic sequence, 2) pyroclastic rocks are predominantly accidental lithic and/or sideromelane shard rich predominantly diffusely, thinly, undulatory or dune bedded lapilli tuff and tuff facies, 3) either steep angular unconformity or

continuous contact between pre-volcanic and volcanic units, 4) these sediments often overlain by scoriaceous coarse grained tuff breccias, lapilli tuffs and tuffs of any kind, and lava flows or dykes.



Fig. 44 – Unconformity in the medial section of Pigroot Hill

Facies are grouped into UD1 forming the medial stratigraphic level at the western cliffside of the Pigroot Hill. The total thickness of this unit ranges between 40 - 60 m. The facies of UD1 are near-vent to medial accidental lithic clast-rich lapilli tuff and tuff beds. Transportation direction based on bedforms in base surge deposits, where it was possible to determine suggest north to south transportation but in the southern areas the bedding features not well enough developed to say surely the transportation directions. In the southern areas bedding tends to be well developed in contrast to north where unsorted structureless, thickly bedded lapilli tuffs are common in the sequence. The largest grain size of the beds is decreasing from 120 cm in diameter to 25 cm beside a km distance from the NZMS 260 I42/075528 point to NZMS 260 I42/076522. The abundance of large bombs and blocks also gradually decreasing in a similar distance. Transportation direction calculated from the asymmetric impact sags indicates north to south transportation in each case. Impact sags tend to be more asymmetric toward south. The depth of the

impact sags gradually decreasing toward south measuring in a same size bomb. Impact sags are plastically deformed similar to impact sags have been described from the northern part of the Trig 634 from the "road-cut", but they are shallower at the Pigroot Hill areas. R-type accretionary lapilli are common, and their abundance seems to increase toward south. Vesiculated tuffs are more prominent in the south as well in contrast cored bombs are more common in the northern areas. UD1 unit seems to thinning toward south.

UD1 in the southern foothill of the Pigroot Hill exhibit a complex history of a phreatomagmatic volcano. It can be interpreted as 1) a crater rim sequence developed around a phreatomagmatic vent and deposited on a pre-volcanic sedimentary unit or 2) a crater rim sequence developed around a phreatomagmatic vent and accumulated in a nearby depression e.g. maar crater. Because of the presence of LD facies below UD1 is not fully well constrained by its origin both interpretations can be valid. Because LD1 underlying the UD1 in the western foothill side at Pigroot Hill was rather interpreted as an initial pyroclastic sequence around a phreatomagmatic vent, the 1st interpretation is favoured.

The general thinning of the UD1 toward south and the north to south transportation direction calculated from impact sag asymmetry or base surge bedforms indicate that the source of the UD1 must have been not too far (hundred metre scale) from the northwest segment of the Pigroot Hill area, approximately around the recent "picnic area" (NZMS 260 I42/075535). The lack of UD1 unit in the southern areas and in the lava fields indicates that UD1 is very likely thinned out quickly away from its source and has not developed in significant distance (more than 3 km) away from its source, or it is covered by subsequent lava flows or both.

### **Stop 9 – Pigroot Hill 3 – Upper Diatreme Succession of Clastogenic Lava Flows and Lava Spatter-rich Pyroclastic Density Current Deposits, Part of a Type 3 Vent Complex**

This stop will explore the transition of phreatomagmatic pyroclastic units to more magmatic explosive eruption-dominated pyroclastic capping units.

Volcanic rocks of the facies listed above crop out in the upper section of Trig 634 (NZMS 260 I42/070536), in the western cliffs of the Pigroot Hill - Round Hill massif (NZMS 260 I42/077527), in a few scattered, poorly exposed outcrops around the southern areas (NZMS 260 I42/083510), and near the Waihemo Fault to the east (NZMS 260 I42/088528). All facies predominantly consist of

tightly packed fluidally shaped black to red volcanic bombs and lapilli with a small proportion of fine ash matrix (Fig. 45).

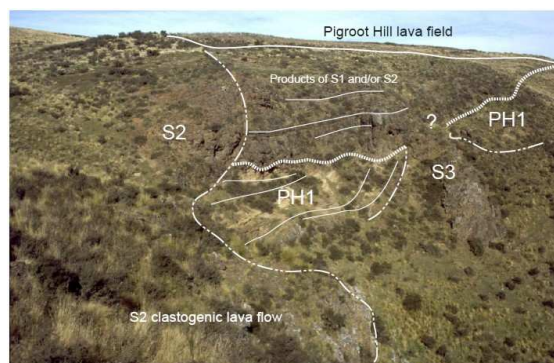


Fig. 45 – Complex 3D relationship between phreatomagmatic pyroclastic rock units and capping scoriaceous pyroclastic rock units at Pigroot Hill.

Beds where they developed, laterally continuous over at least tens of metres (Fig. 46).

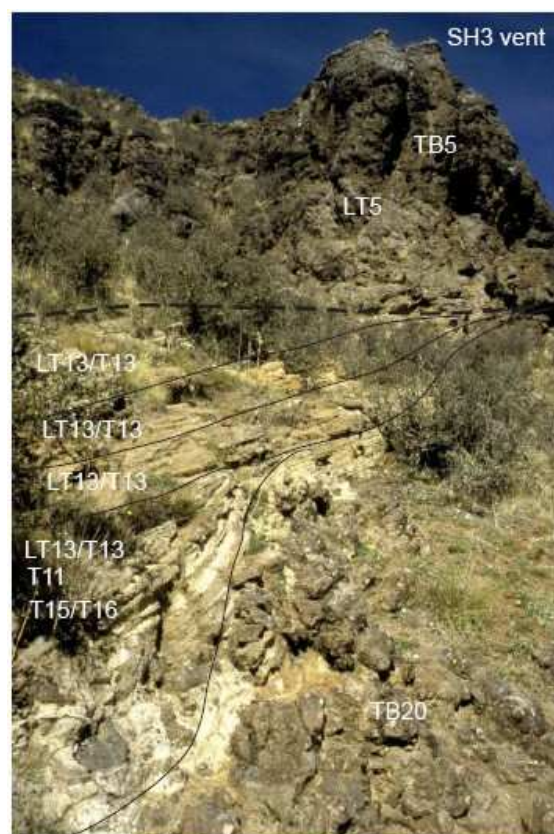


Fig. 46 – Capping scoriaceous beds above phreatomagmatic lapilli tuff and tuff beds.

Large irregularly shaped lava clasts tend to be flattened parallel to the paleosurface, especially in the upper outcrops in each locality (Fig. 47).



Fig. 47 – Ribbon bombs in fine tuff matrix in the base of the capping scoriaceous, lava spatter-dominated succession in Pigroot Hill.

In the western Pigroot Hill outcrops interbedded coarse- and – fine-grained beds give a layered appearance to the outcrops. These spatter-rich deposits are crudely bedded, but have a planar fabric defined by flattened spatter clasts. Tabular bodies of TB4a & TB20a seem to cut through bedded lapilli tuff layers in places in the northwestern part of Pigroot Hill (NZMS 260 I42/075528). Large (tens of cm scale) lava fragments tend to have few cm thick chilled crust in contrast their interiors is highly vesicular with irregular vesicles. Small (<10 cm) rugged lava fragments are very irregular in shape, and they are slightly to extremely vesicular. In the stratigraphically lower section at the western outcrops of Pigroot Hill, the tuff breccias have few or no spindle shaped bombs, but they become more abundant up-section. The tuff breccia in places is completely matrix free (TB4a) and clast supported, but grades over a few metres laterally to a matrix supported rock (TB20a) with tens of vol % matrix by visual estimate. The matrix-rich tuff breccia (TB20a) has a brown matrix rich in accidental lithic clasts derived from Tertiary sedimentary rocks, particularly from the glauconitic marine units. The matrix of these rocks contains both sideromelane

and tachylite shards of mm-scale and angular shape. Most of the sideromelane shards have a palagonite rim. Sideromelane shards are microvesicular, or non-vesicular. Microlites are often present in glass shards. Clinopyroxene and olivine are rare and occur as broken, angular, small (submillimetre) crystals.

Both types of tuff breccia in western outcrops of Pigroot Hill contain large "chunks" of marine sandstone fragments of irregular shape and up to 3 m in diameter. Most sandstone fragments have margins with fused grains, suggesting thermal alteration. Large holes in places (NZMS 260 I42/074531) in the tuff breccia, especially in the clast-supported type, seem on the basis of presence of marine sand in these irregularities to be places from which soft marine sandstone clasts have weathered out. In places limonitic, highly irregularly shaped sandstone "blobs" are present that have cores of coarse white sand that closely resembles some of the marine sandstone. Both types of tuff breccia at Pigroot Hill also contain large numbers of irregularly shaped clasts up to 50 cm in diameter of lapilli tuff or tuff that closely resembles phreatomagmatic tuff and lapilli tuff that underlie the tuff breccia units. Especially in the upper stratigraphic level of the tuff breccia at Pigroot Hill basaltic, slightly ellipsoid cored bombs are common. Each bomb has a core, usually of thermally recrystallised quartz sandstone that closely resembles the pre-volcanic marine sandstones. The core of the bombs is rimmed by glassy basalt that tends to be more vesicular away from the core and with a rugged outer surface. Occasionally bombs are cored by crystalline lava fragments. These cored bombs are more prominent high in beds of the clast-supported, juvenile-rich (scoriaceous), weakly bedded lapilli tuff (LT5a) or matrix-supported, juvenile-rich (scoriaceous), diffuse stratified lapilli tuff (LT23a). On the west flank of Pigroot Hill, lava clasts vary from highly vesicular to non-vesicular, almost "chilled-looking", clasts often having flow banded textures. In places thick (a few metres) interbedded coherent lava is highly vesicular. Highly vesicular zones have ellipsoid vesicles that are flattened parallel to bedding (NZMS 260 I42/978528). There are finely interconnected juvenile fragments, often connected to each other through fine lava "bridges". Lherzolite and pyroxene megacrysts up to 1 cm in diameter are present rarely. Angular clasts of crystalline lava are also present, especially in the lower stratigraphic level. Accidental lithic clasts are predominantly small (cm scale), angular, platy schist clasts.

On the western flank of Pigroot Hill clastogenic lava flows (Wolff and Sumner, 2000) form thick layers associated with at least two main source vents around NZMS 260 I42/077527 and NZMS 260 I42/077522. These areas can be

interpreted as remnants of Hawaiian-type spatter cones (Wolff and Sumner, 2000).

UD2 unit is interpreted to be a product of subsequent magmatic explosive eruptions followed the initial phreatomagmatic eruptions. The well-distinguished areas filled by scoriaceous tuff breccias and lapilli tuffs are interpreted as individual vent zones of a scoria and/or spatter cone. The common vesicular lava flows with flattened vesicular clast rich interior, associated to these well-distinguished dish-shape structures are interpreted as clastogenic lava flows. Clastogenic lava flows usually produced in eruption on scoria or spatter cone when the magma discharge rate suddenly increases. The emplaced fall-back, still melt spatters of melt can melt together around the vent and retain the heat enough to cause flow movement in the melt aggregate and produce rootless, secondary lava flows. At least the highly vesicular clast-rich lava flows represent this type of rootless lava flows. In the upper section, the increased amount of well-developed spindle bombs in the scoriaceous ash rich lapilli tuffs indicate higher energy Strombolian explosion that was able to throw out large lava clasts, which clast travelled a significant time in the air until they impacted to the surface. UD2 units, especially in the northwest area of the Pigroot Hill contain large number of accidental lithic clasts from different pre-volcanic sedimentary units as well as lapilli tuff and tuff clasts from earlier developed UD1 units. The presence of these exotic clasts in the tuff breccias in that area suggests that an active magmatic explosive vent much have cut through an area where these sediments were ready to thrown out by an explosive eruption. The tilted nature of UD1 beds may indicate that magmatic explosive vents generated UD2 tuff breccias erupted in a vent zone where phreatomagmatic lapilli tuffs and tuffs collapsed and slide into. It is also a reasonable interpretation that the magmatic explosions itself caused a significant collapsing of the underlying phreatomagmatic units (UD1) giving substantial source material to be exploded for the subsequent explosions.

### **Stop 10 – Discussion at the Top of Pigroot Hill Sequence - Departure to Oamaru**

Before departure a short summary will be given about the syn-eruptive landform reconstruction have been made for the Pigroot Hill (Németh and White, 2003b).

Correlation between different volcanic facies and/or stratigraphic units of Trig 634 and the Pigroot Hill - Round Hill massif is not simple because of 1) a possible subsequent tilting of the entire area, 2) lack of continuous outcrops between these localities,

3) sudden and dramatic changes of individual volcanic pyroclastic facies according to their position compares to a vent zone and 4) disturbance in the first phreatomagmatic sequences caused by development of multiple magmatic explosive eruptive centres (Fig. 48).

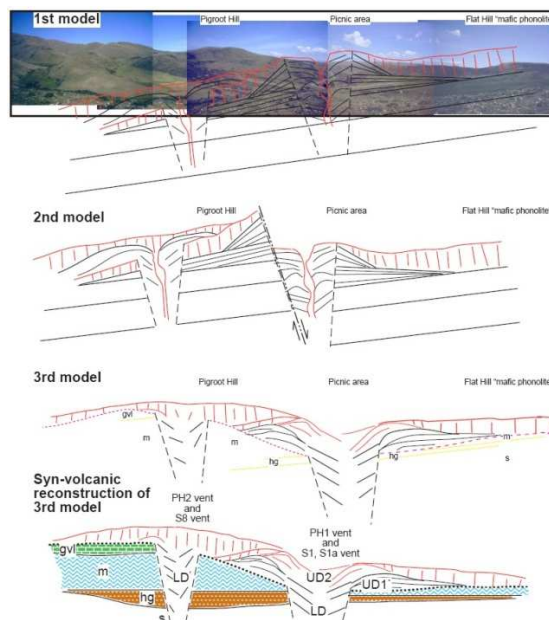


Fig. 48 – Potential 3D reconstruction of volcanic facies relationships around Pigroot Hill.

1) general pre-volcanic bedding tend to dip toward southeast and ii) the orientation of platy joints of the lava field on top of the Pigroot Hill - Round Hill massif tend to follow the recent morphology and mimic the dip direction of the pre-volcanic sediments. Tilting is also suspected according to the general distribution of the pre-volcanic Tertiary sediments mapped around the Pigroot Hill Volcanic Complex. The highest stratigraphic position limestone beds (GVL) tend to crop out in the southeastern side of the field in a same elevation then the deepest stratigraphic position Hogburn beds in the northwest side. This facies distribution is required an at least 5-10 degrees general tilting of the area, which tilting has to be calculated for any correlation attempt for correlating volcanoclastic units in the area.

2) there are no continuous outcrops between Trig 634 and the Pigroot Hill northern area. Similar problem exists in areas between the northeast sites next to the Waihemo Fault and the central part of the Pigroot Hill area. Other problem is that large areas covered by lava flow in both hill sides, and there are no exposed outcrops of any pyroclastic rocks between these large vast areas. However, the presence of large "basal" formations close or between pyroclastic rock cliffs may indicate a presence of a former disturbed zones that may have

been eroded away easier than any other areas. In a same logical way, the presence of large lava fields may also indicate significant paleogeographical interpretations. Lava flows largely tend to fill paleotopography, so today presence is a proof that the areas they covered must have been areas where they were able to accumulate, e.g. paleo-valleys. Lava accumulation not necessarily occurs in pre-volcanic valley systems, especially not in a quickly changeable volcanic landscape. It can occur in zones where "something" controlled the lava flow movements (from which solutions pre-volcanic valley systems can be only one explanation). Lava flow can be blocked by syn-volcanic grows of scoria and/or spatter cones, earlier emplaced syn-volcanic lava flows, or tuff ring rim beds.

3) sudden and gradual changes of individual pyroclastic facies in relation to their depositional position compare to their source vent is well known in most of the monogenetic volcanic fields. Near vent to distal facies changes can be demonstrated in the western foothill of Pigroot Hill where a more less 1 km long pyroclastic sequence is exposed, representing UD1 stratigraphic unit. In this section a general north to south transportation of former base surges and ballistic trajectories has been identified. The northern part of the UD1 unit in this section represents near vent in contrast to the south a distal area relative to the source of UD1. In this short distance the sequence is quickly changing to a more well-bedded, fine grained lapilli tuff and tuff sequence. Assuming a similar range of horizontal changes, it is a reasonable assumption that very different pyroclastic units should crop out in a short distance at nearby hill sides, unless they are in a similar distance-range from their source.

4) it is clear from the western cliff side of Pigroot Hill that small magmatic explosive vents cut through often the UD1 pyroclastic unit. In areas close to these vent zones complex tilting and/or collapsing structures are disturbing the UD1 unit. Few of these magmatic explosive vents surely existed before UD1 unit of the Pigroot Hill west side emplaced thus even more complicated interrelations developed between phreatomagmatic and magmatic pyroclastic sequences.

UD1 units in Trig 634 and Pigroot Hill west may be correlated to each other, and represent a complex tuff ring sequence developed around a phreatomagmatic vent (Fig. 49). With an average of 5 - 10 degrees post-volcanic tilting of the entire area different stratigraphic units can be correlated. Tilting back the entire area by this angle can explain very well the facies and unit distribution has been described in Trig 634 and Pigroot Hill western and northern area.

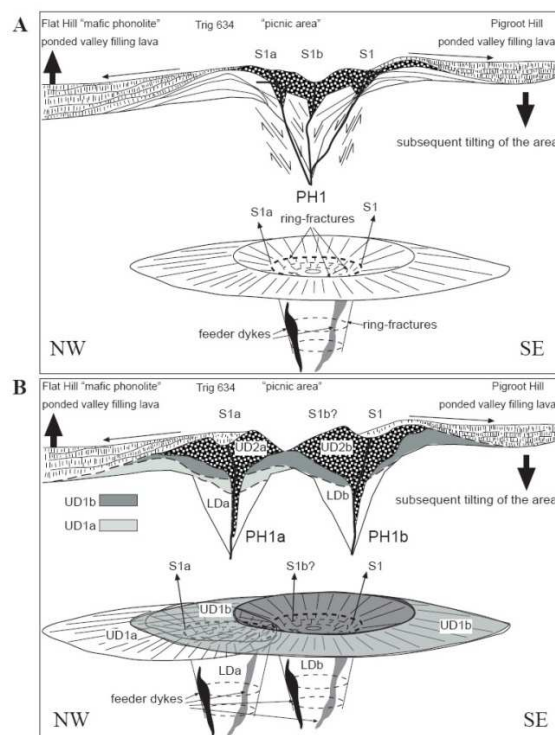


Fig. 49 – 3D model for the vent locations at Pigroot Hill.

LD unit around the "picnic area" (NZMS 260 I42/075535), which area is a deepest exposure of pyroclastic rocks in this area, represents a former vent, possible maar (Fig. 50). This interpretation is supported by the presence of common disturbed, tilted, disoriented mostly coarse grained pyroclastic rocks always rich (5 - 80 vol %) in accidental lithic clasts derived from prevolcanic rock unit. The presence of weakly vesicular sideromelane glass shards indicates magma/water interaction.

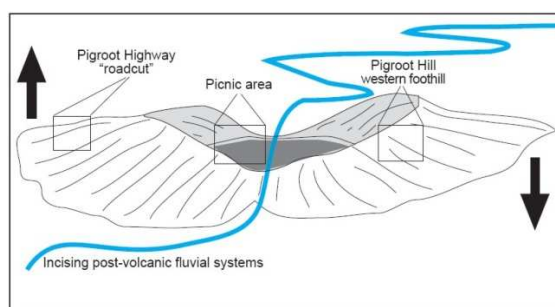


Fig. 50 – Possible vent located around the "picnic area".

These facts indicate that a significant subsurface phreatomagmatic explosive eruption took place in this area. The coarse grained (tuff breccia, lapilli tuff) pyroclastic rocks often tilted disoriented position must represent syn-volcanic, collapsed and

possible subsided blocks into a vent zone. The correlation to LD unit in the western foothill of the Pigroot Hill is not evident. By a reconstructed paleosurface and the position of a vent around the "picnic area" rather indicate that LD unit in that area should have been deposited near a phreatomagmatic vent, probably early in the development of a tuff ring around the vent (Fig. 50). It is also reasonable interpretation that LD unit in the western foothill of Pigroot Hill is a true vent filling unit and the UD1 and UD2 units developed in a former vent zone sourced from a nearby vent around the "picnic area". Unfortunately there is no positive evidence to support either of these reconstructions. Very similar problem can be realised in the southeastern foothill of the Trig 634. The lowermost LD unit in that area seems to be developed already on a pre-volcanic surface according to field relations to the pre-volcanic units and the drawn cross-sections. LD unit in this area very likely represents a near vent initial section of a phreatomagmatic pyroclastic sequence similarly to LD unit in the western foothill of the Pigroot Hill. LD units in the northern area of the Pigroot Hill (NZMS 260 I42/083534) and in the southern area of the Round Hill (NZMS 260 I42/510) represent individual vent zones. Their rim deposits either covered by the extensive lava flows of the Pigroot Hill Volcanic Complex (very likely at the northern side), or they have been already eroded away (very likely in the southern side), or both.

After re-tilting the entire area approximately 10 degrees to northwest, UD1 units in both hills place into the same elevation. UD1 unit in the Pigroot Highway "road-cut" represents a distal mostly wet base surge sequence deposited approximately 700 m away from its source. These beds largely can be correlated to the lower sequences of bedded lapilli tuffs and tuffs in the UD1 unit in the western side of the Pigroot Hill. In the western side of Pigroot Hill this sequence exhibits a proximal to distal facies transition of pyroclastic facies. UD1 unit in the southern foothill of Trig 634 can be correlated to UD1 units in the western foothill of Pigroot Hill. UD1 at Trig 634 south is exhibit features characteristic for near vent deposition from high concentration pyroclastic surges. The different textural and facies characteristics of the UD1 units can be explained well with a deposition of a thick phreatomagmatic pyroclastic sequence around a phreatomagmatic vent (e.g. maar). The crater rim beds around the vent formed a semicircular pile of tephra. Because the individual phreatomagmatic blasts (or longer eruptive periods - hour scale) may have been travelled into different directions instead of uniformly radially outward moving mode, the deposited tephra very likely deposited in an asymmetric, semicircular way around the vent, possible with sudden changing dip directions

(according to the paleorelief) and developed volcanoclastic facies. Correlation between UD2 units is not evident. UD2 units seem to develop lensoid or funnel-shaped structures cut through the phreatomagmatic UD1 units. There are evidences that few magmatic explosive vents developed before the phreatomagmatic units (UD1), and caused a topographic barrier for the passing base surges. The best example is located in the western foothill of Pigroot Hill (NZMS 260 I42/0795324) where a steep spatter deposit pile (up to 15 m thick) forced to climb upward a passing base surge, producing a gradually steepening bed surface in the developing pyroclastic sequence. Magmatic explosive vents may have been simultaneously active with phreatomagmatic vents as well because in the upper stratigraphic position scoriaceous, and spatter-rich deposits seem to be inter-bedded with thin phreatomagmatic lapilli tuffs and tuffs. The magmatic explosive events producing UD2 units in the western foothill of Pigroot Hill must have been active shortly after the phreatomagmatic tephra layers deposited because 1) the magmatic explosive units especially around point NZMS 260 I42/076528 rich in large "chunks" of irregular shape phreatomagmatic lapilli tuff and tuff clasts up to a metre in diameter. In further south the UD2 units contain smaller enclosed lapilli tuff and tuff fragments. In each place the phreatomagmatic units are thermally altered near to beds of UD2 unit. UD2 unit on top of Trig 634 seems to blanket the underlying phreatomagmatic units. UD2 unit is very likely the source of lava flows covering the Trig 634. UD2 units in the eastern side of the Pigroot Hill, close to the Waihemo Fault may be related to a separate vent than vents formed the UD2 units in the western side of Pigroot Hill. Unfortunately the top of the Pigroot Hill is covered by lava, therefore direct correlation between eastern and western side is not possible. Because the UD2 beds in the eastern side of the Pigroot Hill similarly coarse-grained, scoriaceous, unsorted then in the western side, it is closely resembles to near vent spatter deposits, and mark a position of a vent zone in that area. Pyroclastic rocks of UD2 unit in the upper section in the western cliff of Pigroot Hill tend to form a continuous blanketing layer immediately under the lava cap. It may indicate that extensive scoria cone eruptions occurred in immediately before the lava effusion started.

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