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Waipiata Volcanics, Otago, New Zealand

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Abstract

This guide to a one-day pre-conference field trip includes a summary of the eruption style, mechanisms and landform evolution of the monogenetic volcanoes of the Waipiata Volcanic Field in Otago. The trip will address basic preservation styles and types of phreatomagmatic volcanoes (from eroded tuff rings, to maars and exposed diatremes), and visit some informative sites. We will discuss current scientific problems associated with volcanism in intraplate terrestrial settings, including signs of magmatic complexity in small-volume volcanoes and the potential use of erosional remnants of monogenetic volcanoes in landscape-evolution models for broad regions. The main stops will be at the Swinburn volcanic complex, in the context of a trip from Dunedin to Oamaru with over-look stops for erosional remnants of monogenetic volcanoes in the schist-tor landscape near Middlemarch.

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^2) 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).

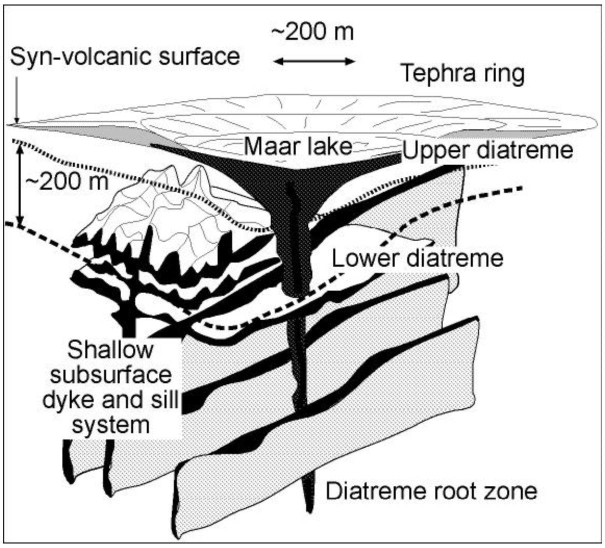


Fig. 1 – Schematic 3D section of a phreatomagmatic maar-diatreme volcano.

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 volcaniclastic 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 diametre (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 - 0s). 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 northem 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).

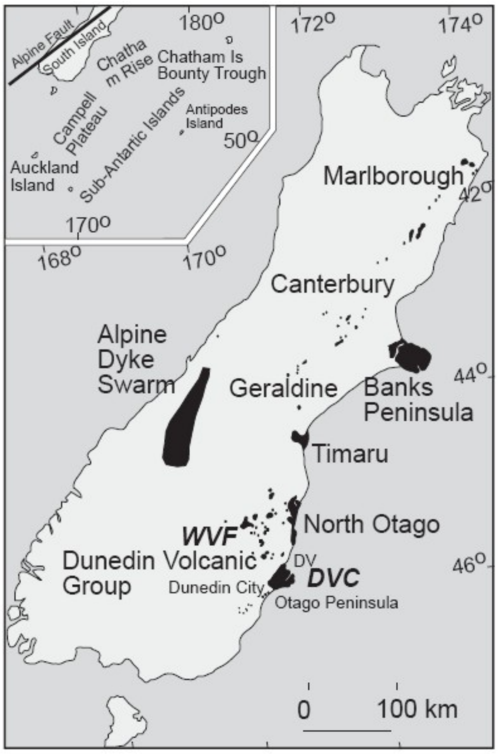


Fig. 2 – Intraplate volcanic fields in the South Island of New Zealand. DV – Dunedin Volcano; DVC – Dunedin Volcanic Complex; WVF – Waipiata Volcanic Field.

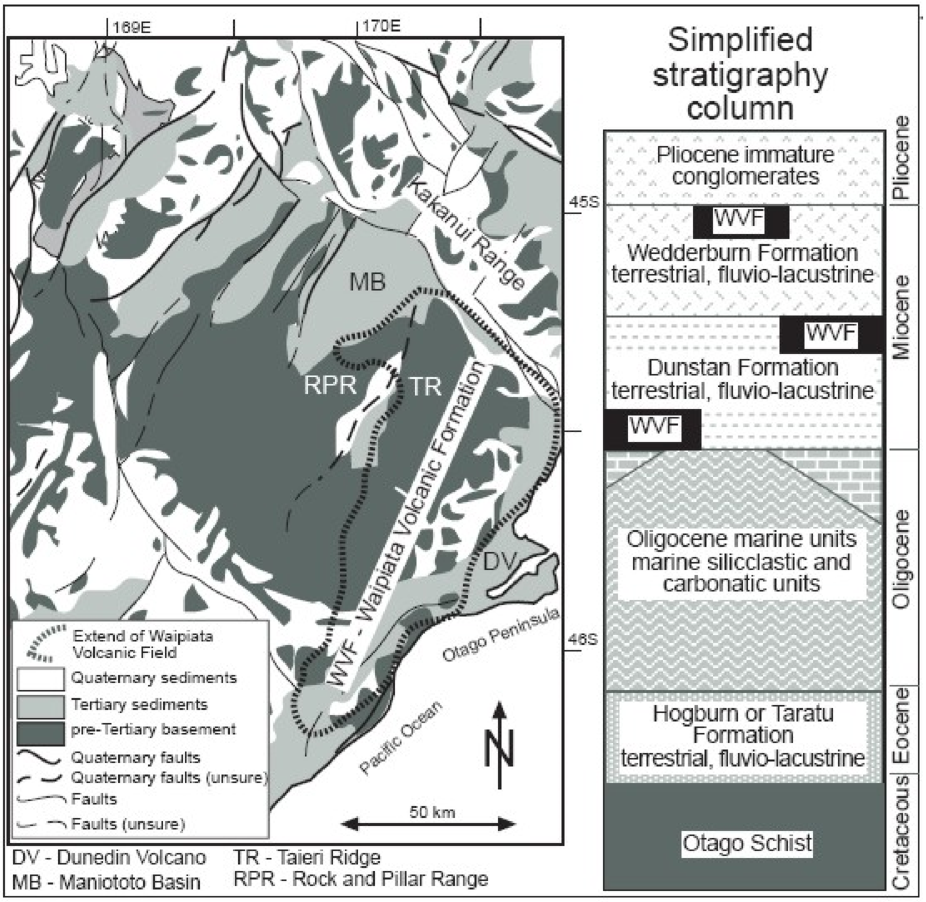


Fig. 3 – Pre-volcanic stratigraphy and sedimentary basins in the Otago region, in the location of the Waipiata Volcanic Field.

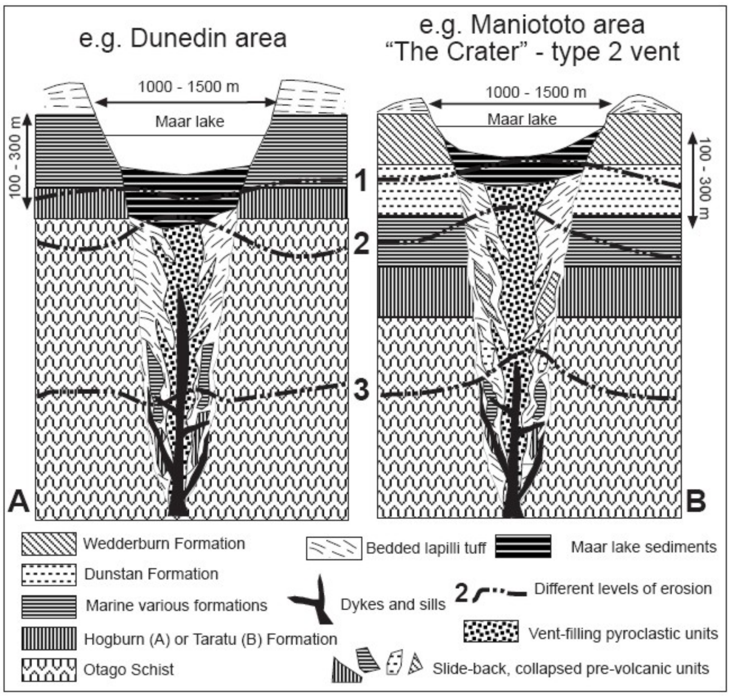
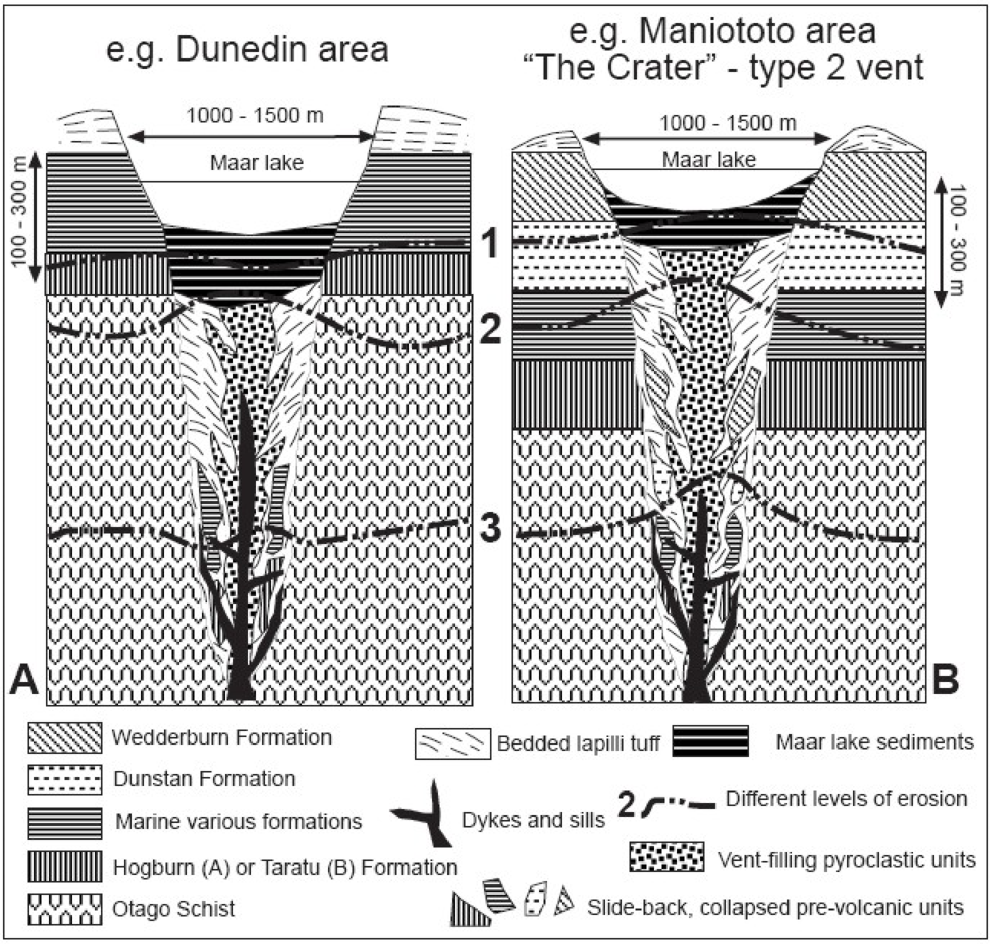


Fig. 4 – Interpretive cross-sections of maar-diatreme volcanoes cut through the pre-volcanic rock units in the Dunedin (A) and in the Waipiata (B) area.

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 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 fanglomerates 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).



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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 Cambell 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).

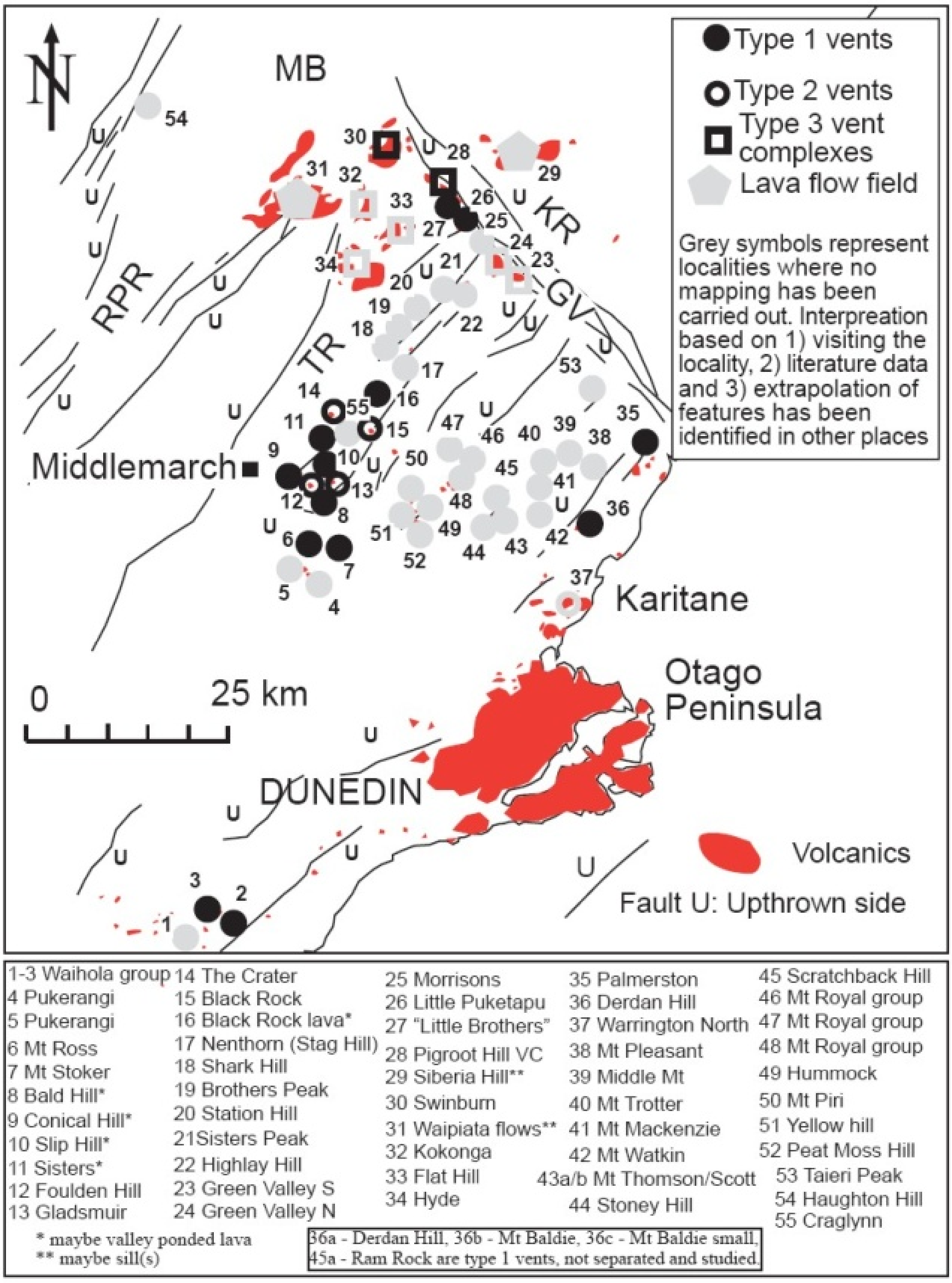


Fig. 6 – Volcanic erosional remnants within the Waipiata Volcanic Field.

In the WVF, three types of volcanic erosion remnants are distinguished (Németh and White, 2003b) according to their volcanic facies associations, ratio of preserved pyroclastic and lava units, and the size and number of identified eruptive centers (Fig. 7).

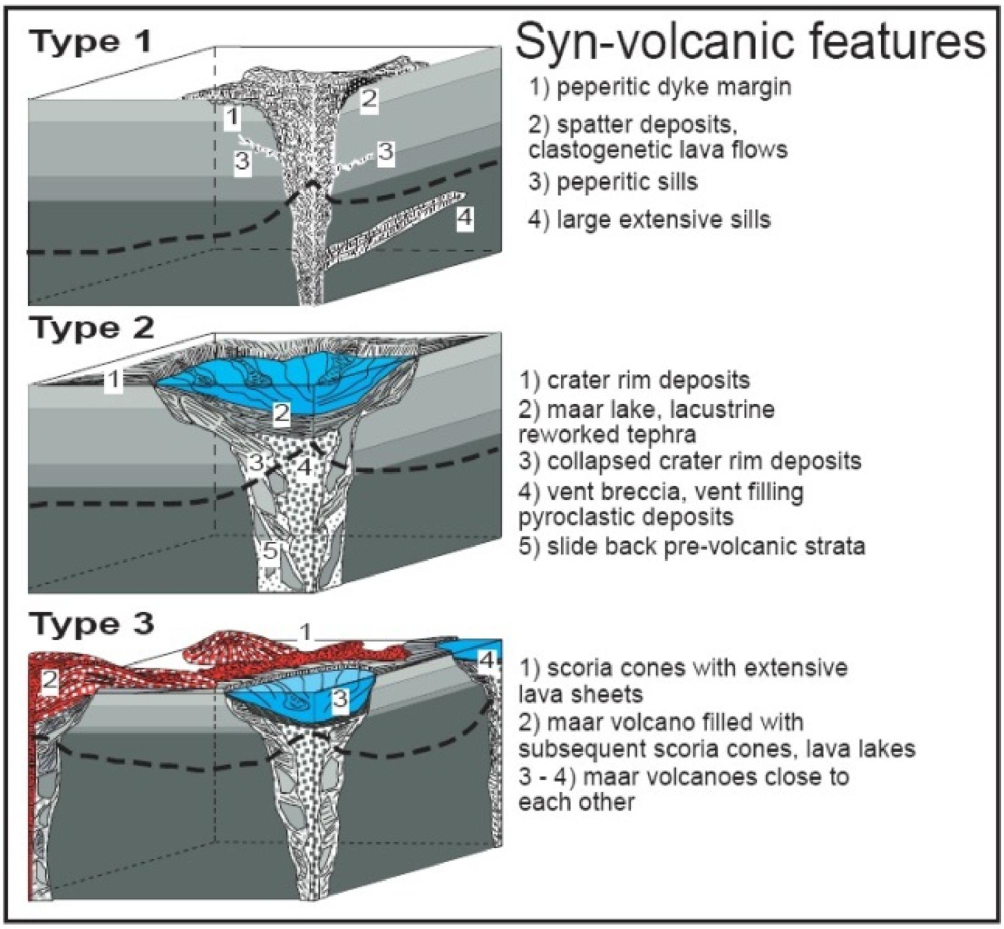


Fig. 7 – 3D reconstruction of the original volcanic landforms the three distinguished volcanic erosional remnants could be associated with (after Németh and White 2003).

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.

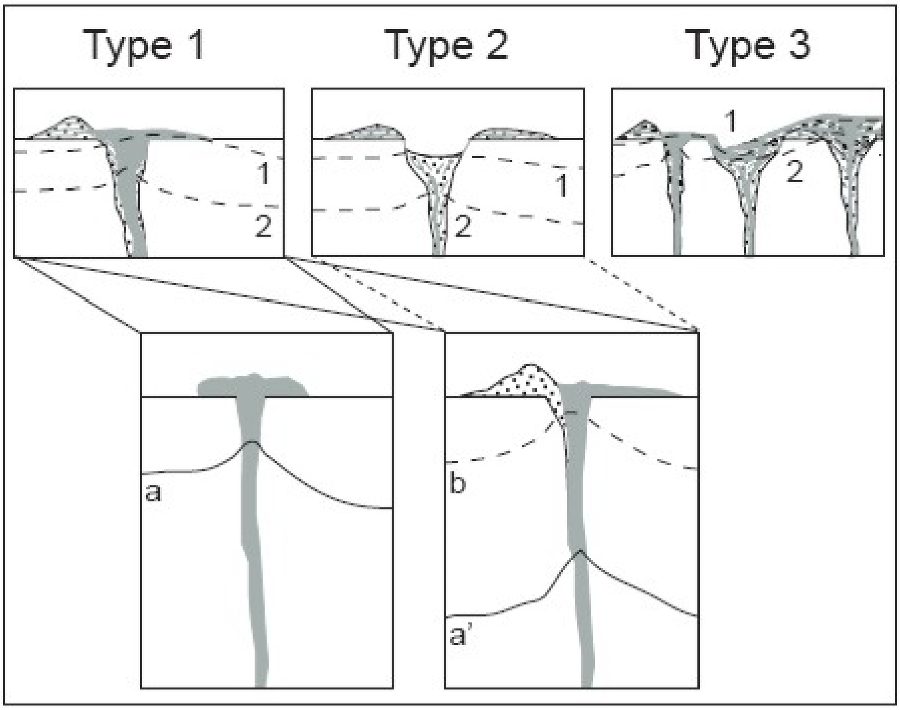


Fig. 8 – Type 1 vents and their potential link to an original volcanic landform.

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.

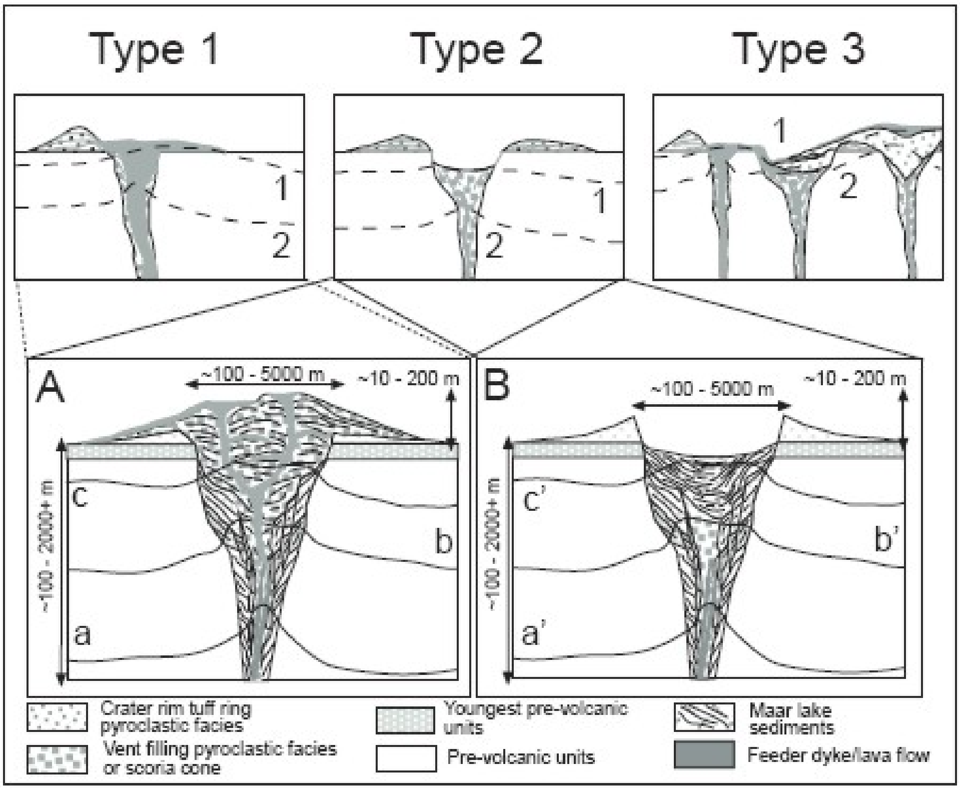


Fig. 9 – Type 2 vents and their potential link to an original volcanic landform; A) maar filled with scoria cone, B) maar filled with lake (after Németh and White 2003).

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).

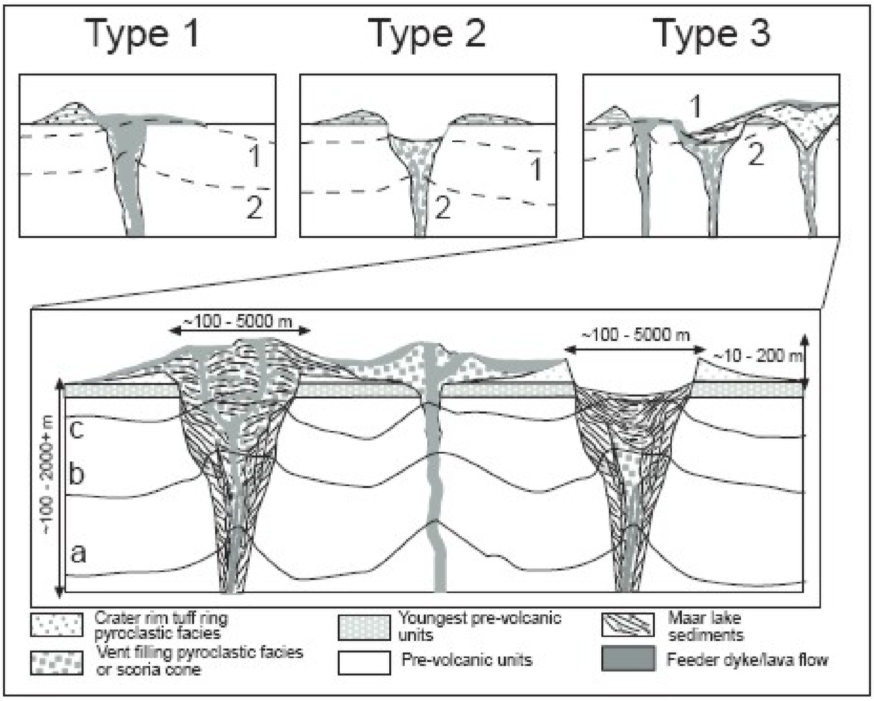


Fig. 10 – Type 3 vent complex and its potential link to an original volcanic landform(after Németh and White 2003).

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).

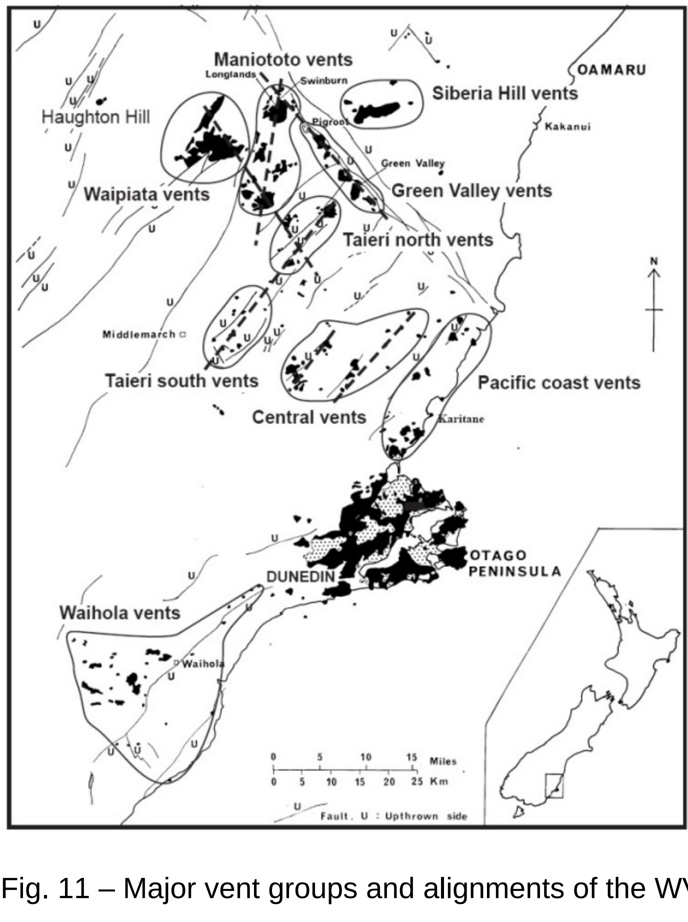


Fig. 11 – Major vent groups and alignments of the WVF.

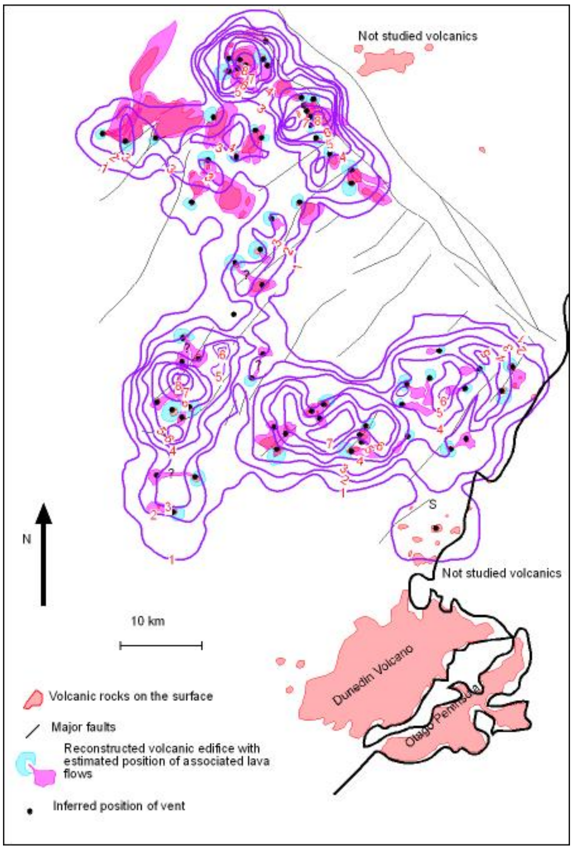


Fig. 12 – Vent-density map of the WVF.

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^3 (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^3 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, 30st locality on Fig 6), and is as recent as 12.8 Ma (Green Valley, 24th 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 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 $^{40}Ar/^{39}Ar$ ages of Hoernle et al. (2006), and concludes that the volcanic activity in the Dunedin Volcano lasted from ***16.0 ± 0.4 to c. 10.1 Ma***, and the volcanic activity of the WVF lasted from ***24.8 ± 0.6 to 8.9 ± 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 Sipiera, 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 100An/An+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 alkalies at <49% SiO2, 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).

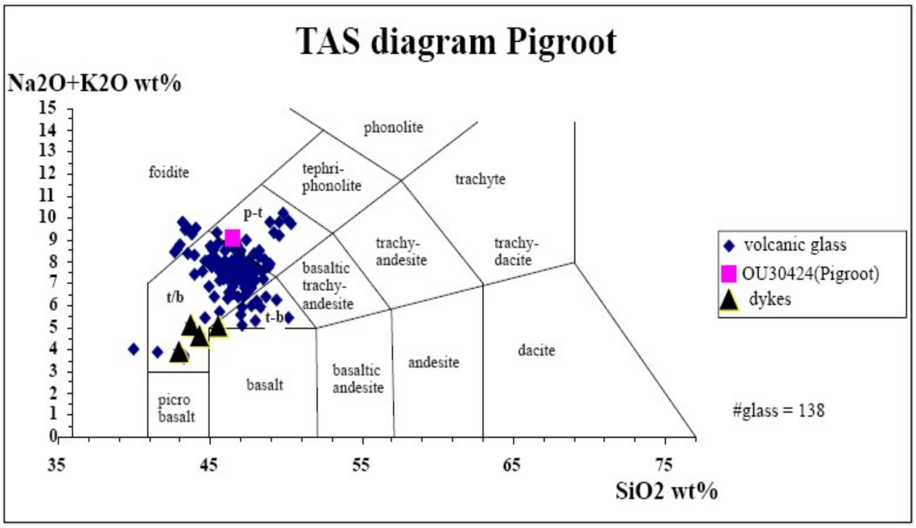


Fig. 13 – TAS diagram of samples from the Pigroot Hill volcanic complex (Type 3 vent complex) of the WVF.

Among lavas, basanites predominate, containing many lherzolite and other mantle-derived xenoliths; olivine nephelinite, mugearite and nepheline hawaiite are also present (e.g., Coombs et al. 1986, 2008). In general, rock types more evolved than nepheline hawaiite 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).

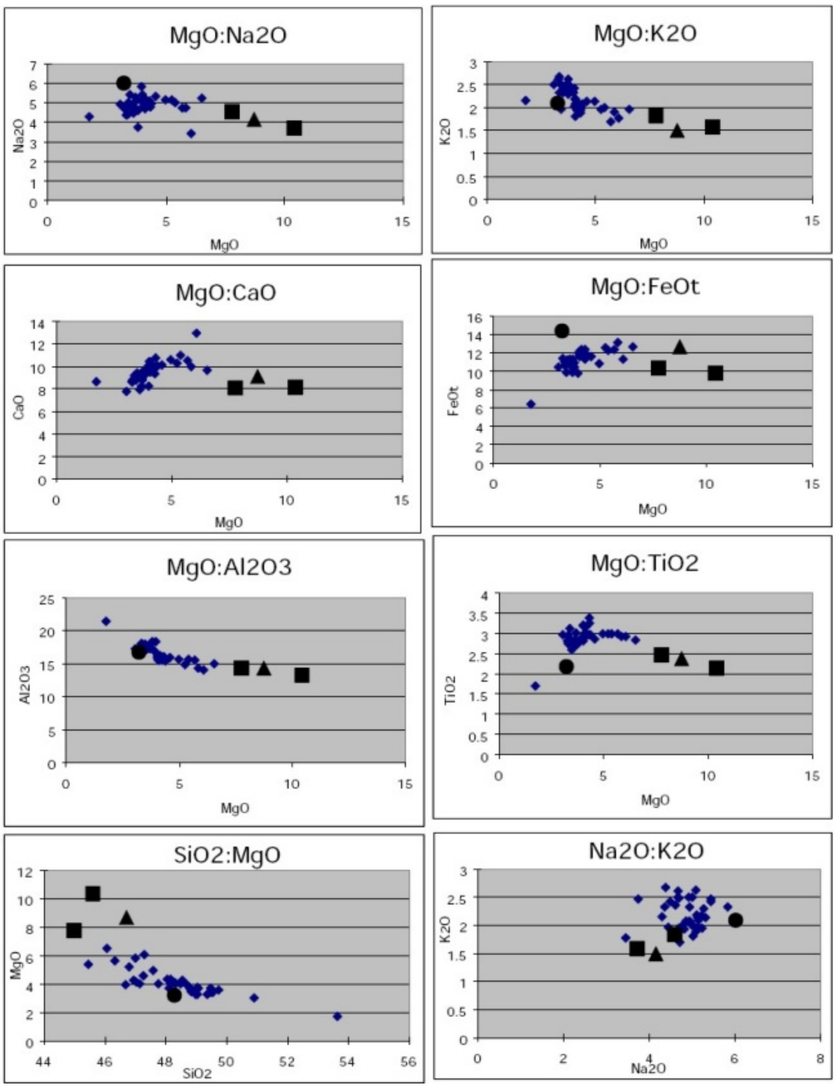


Fig. 14 – Discrimination diagrams of major element composition of volcanic glass shards from early phreatomagmatic successions and subsequent dykes and sills of The Crater diatreme (Type 2 vent).

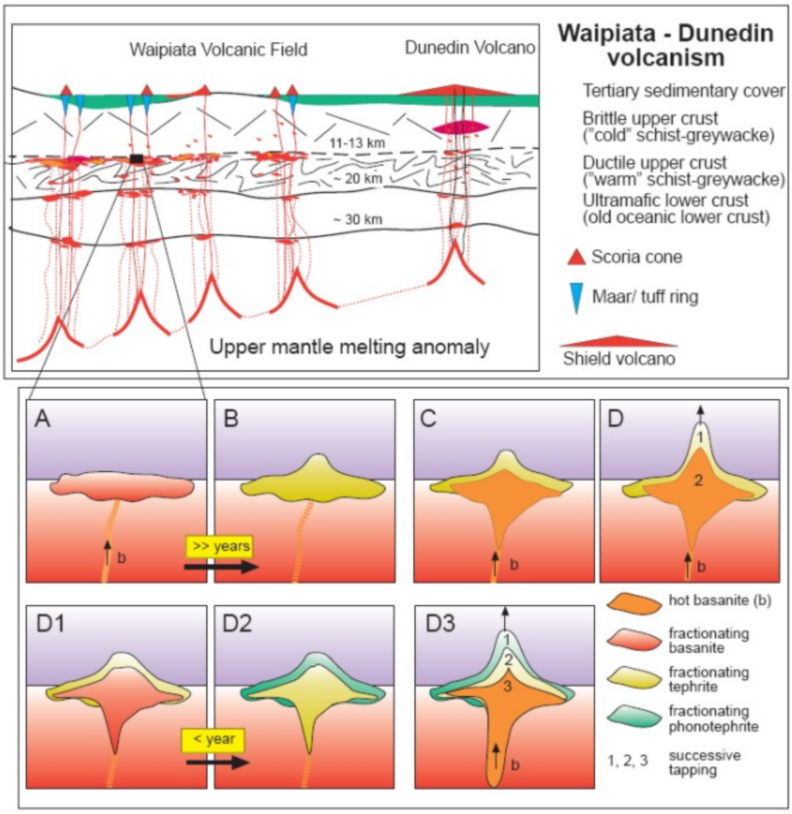


Fig. 15 – A theoretical model for the magma evolution of the individual volcanoes of the WVF.

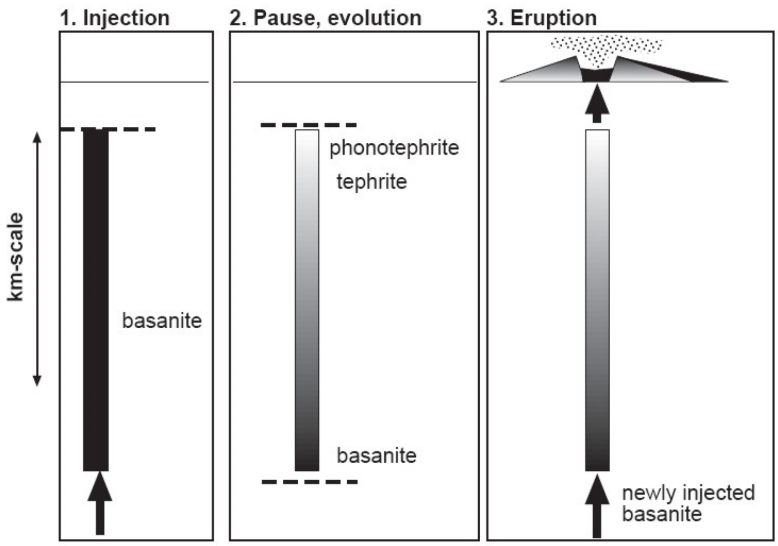


Fig. 16 – Trapped basanite melt evolves to tephrite or phonotephrite. Newly injected basanite pushes evolved melt ahead, so initial eruptive products are evolved, followed by more primitive products.

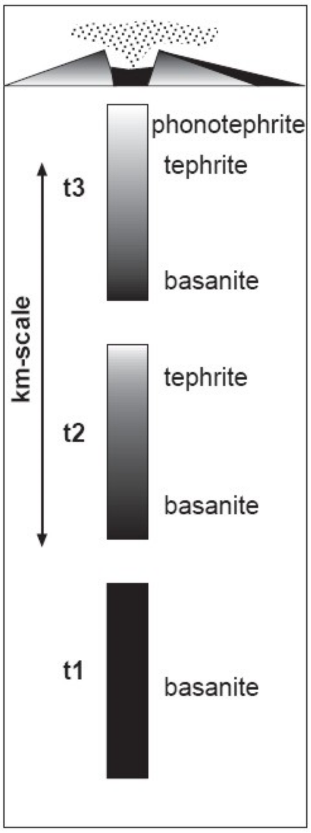
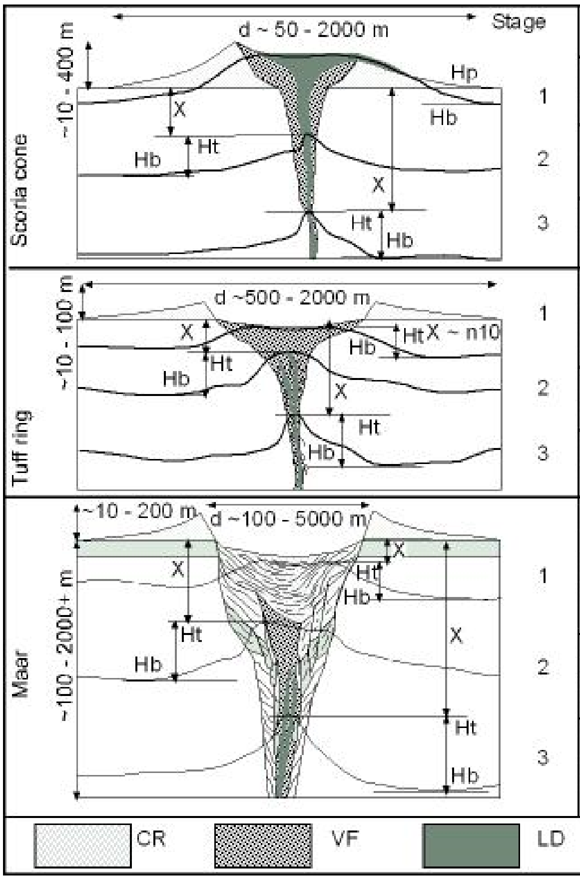


Fig. 17 – Gradual evolving of melt in the melt column.

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.



Type	Stage	Characteristics	Calculation
Scoria cone	1	Distinct hills of bedded scoriaceous lapilli tuff, tuff breccias and agglomerate are characteristic.	Hp - Hb
	2	Butte structure is not wider than few 10 - 100 m. Near-vent pyroclastics are exposed.	Ht - Hb + X X ~ 10 m scale
	3	This stage is represented by exposed feeder dykes. Unbedded, coarse-grained,	Ht - Hb + X X ~ 100 m scale
Tuff ring	1	This stage is represented by outcrops of bedded to non bedded lpyroclastic rocks. Coarse-grained, unsorted, weakly bedded lapilli tuffs crop out in the central part of the	Ht - Hb + X X ~ 0
	2	Buttes are smaller than stage 1 buttes. Pyroclastics are predominantly coarse-	Ht - Hb + X X ~ 10 m scale
	3	Buttes are smaller than stage 2 buttes. Shallow seated lithic clast-rich pyroclastics	Ht - Hb + X X ~ 100 m scale
Maar	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	Ht - Hb + X X > d/5
	2	Maar lake deposits are already eroded, vent-filling pyroclastics are exposed. Collapsed maar crater rim sequences can be identified in complex volcanic stratigraphy.	Ht - Hb + X X ~ 100 m scale
	3	This stage is represented by small dissected outcrops of lvent-filling pyroclastic rocks and crosscutting feeder dykes. Dykes often have	Ht - Hb + X X ~ 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 he 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).

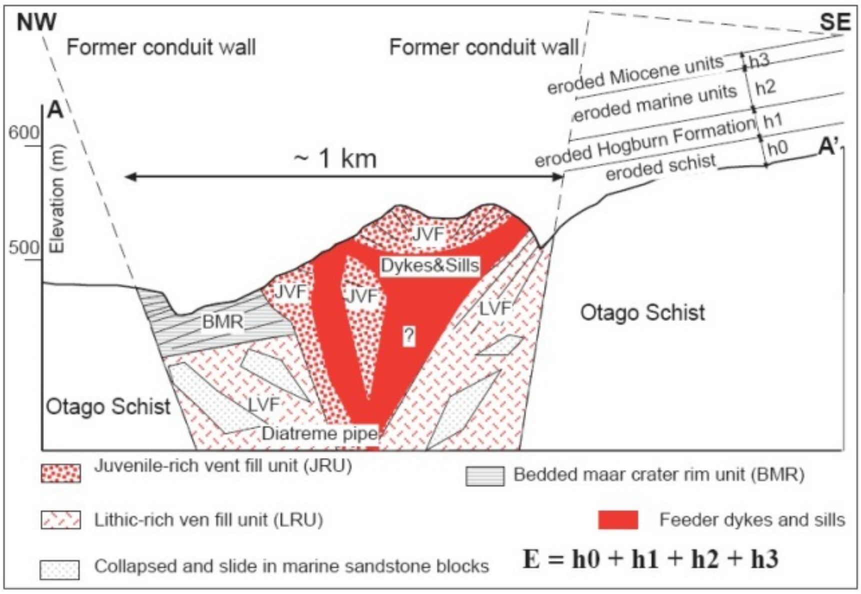


Fig. 18 – Erosion calculation model for The Crater diatreme

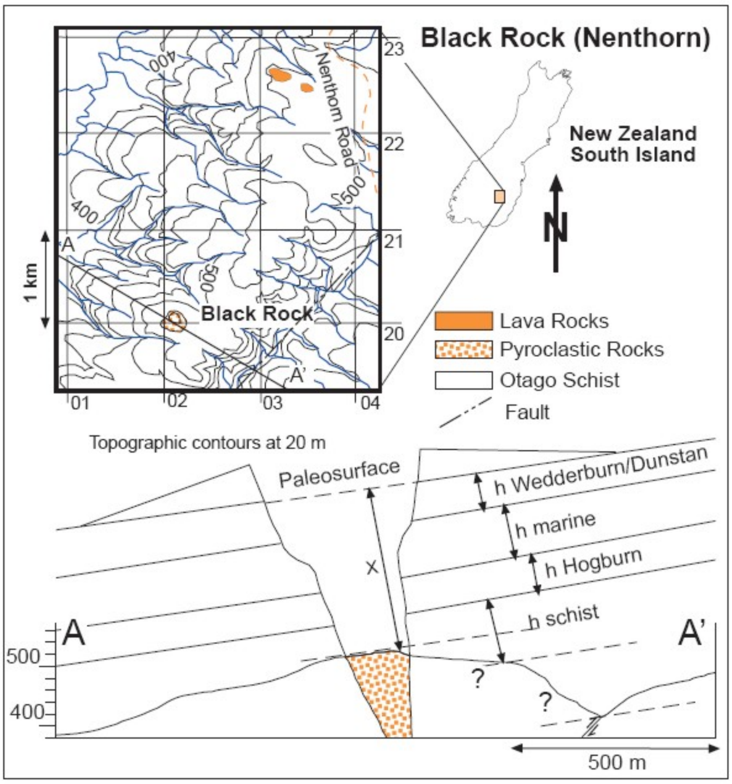


Fig. 19 – Erosion calculation model for the Black Rock diatreme

Field Guide

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.

Stop 2 – Foulden Hill lookout – stop next to road

km - 75

GPS: -45.521541º; 170.220631º

Stop 3 – The Crater lookout – stop next to road

km - 79

GPS: -45.488225º; 170.288371º

Stop 4 – Swinburn – stop next to road

km - 138 (roadcut)

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

-45.137776º; 170.323016º (roadcut)

Stop 5 – Swinburn Plateau – stop next to road

km - 139; walk up paddock track to edge of plateau lavas

Stop 6 (if time) – Red Cutting Summit – stop along road

km - 150

GPS: -45.193160º; 170.41995º

Arrival to Wanaka before 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 Swinburn volcanic complex and the origin of its unusually coarse basaltic rocks.

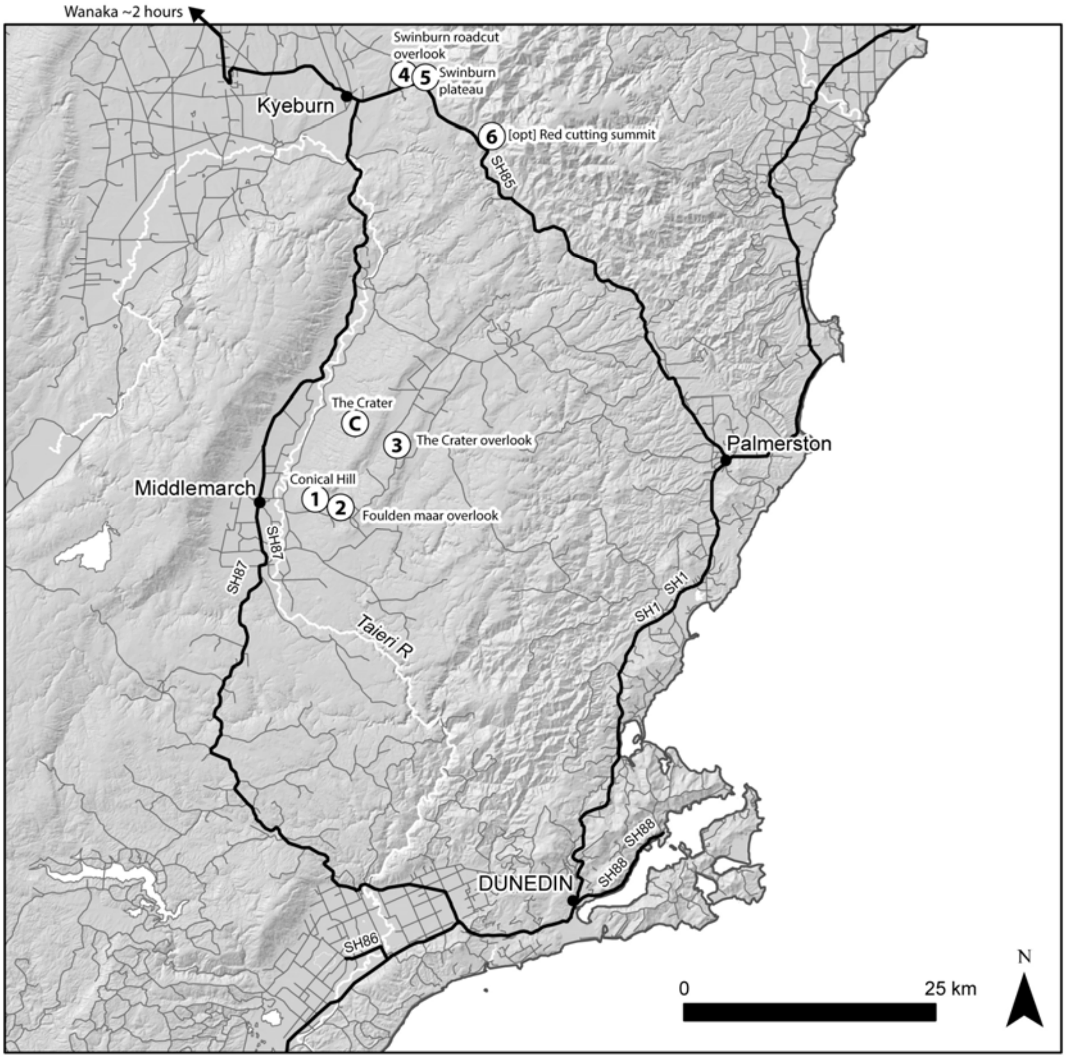


Fig. 20 – Field trip stops

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 then 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

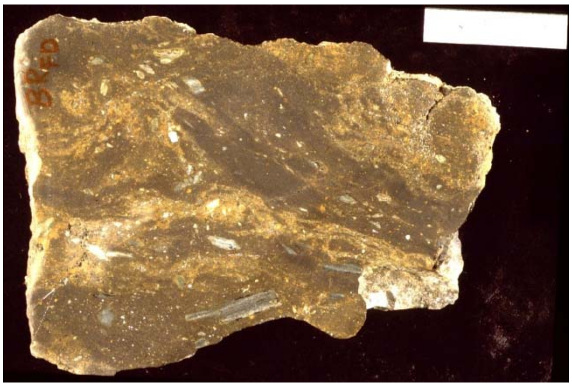


Fig. 22 – Transition from coherent magmatic to fragmented clastic rock facies from the Little Brothers vent suggesting that Little Brothers is a vent that produced fragmental rocks. Such samples have not been identified yet from Conical Hill.

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 againt 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.

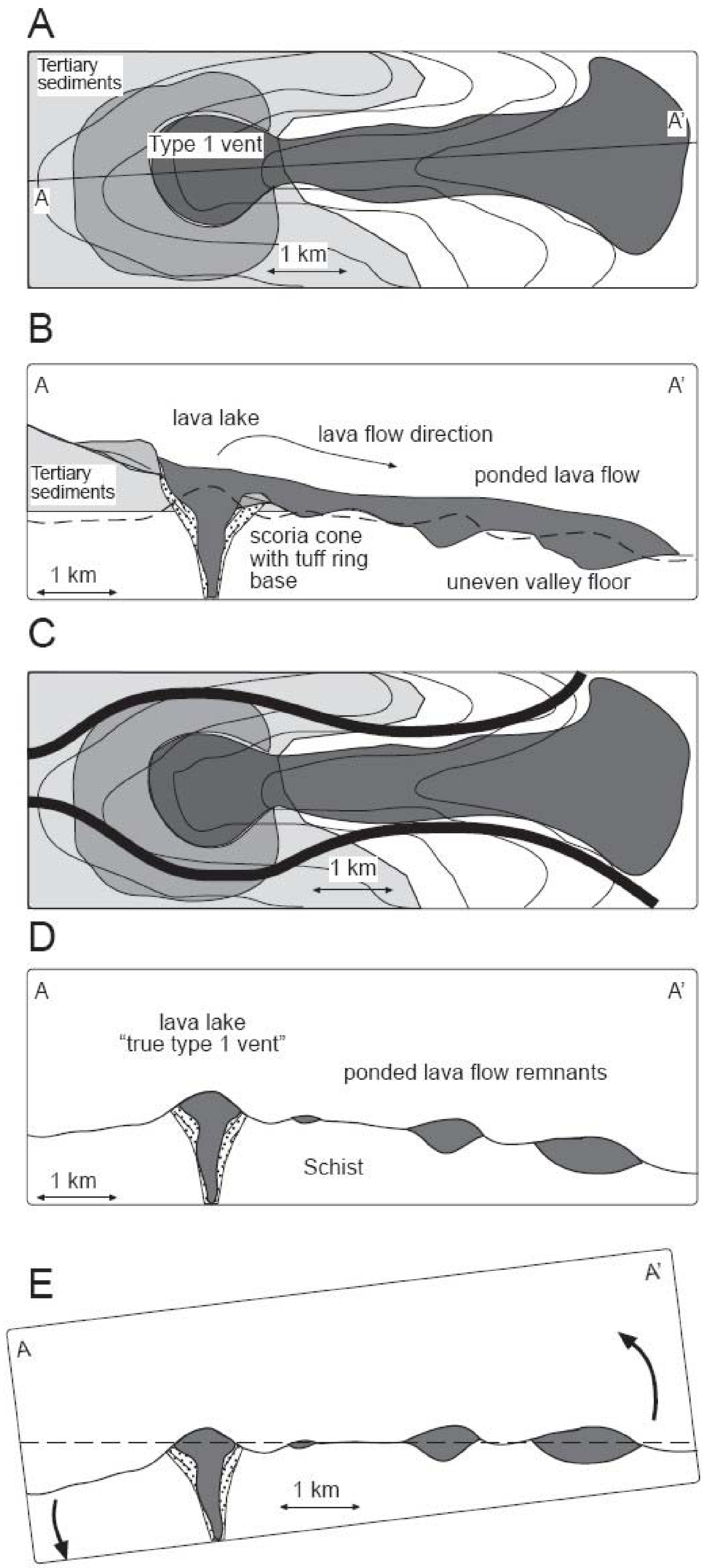


Fig. 23 – 3D relationship between "true" vents and erosional remnants of ponded flow bodies. A – original syn-eruptive morphology, B – cross sectional view, C – potential fluvial reworking (thick line), D – erosional surface, E – tilting effect

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 untis 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 m2) 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 containng 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).

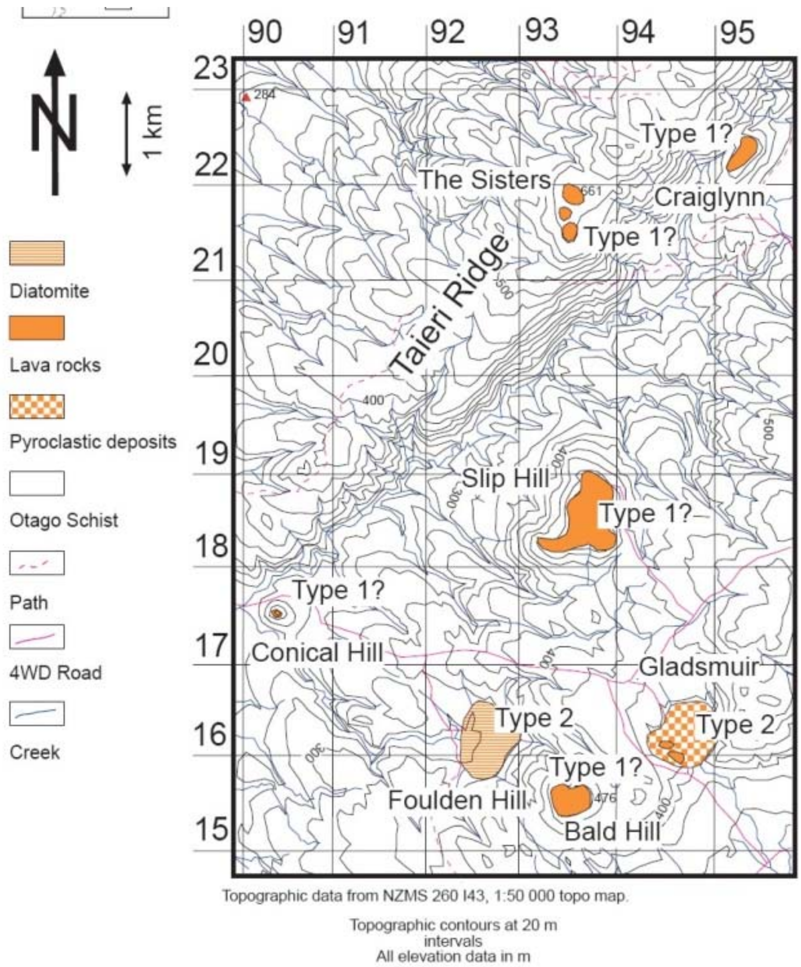


Fig. 24 - Location of Foulden Hill and other Waipiata volcanic rocks on the surface.

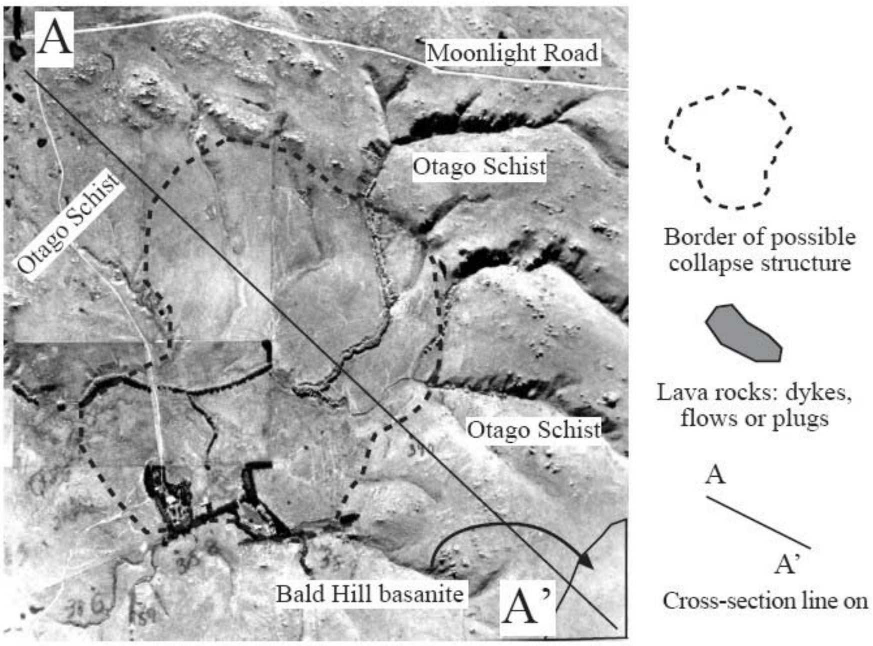


Fig. 25 – Airphoto of the Foulden Hill depression.

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. Lindquivist 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 $^{40}Ar/^{39}Ar$ 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).

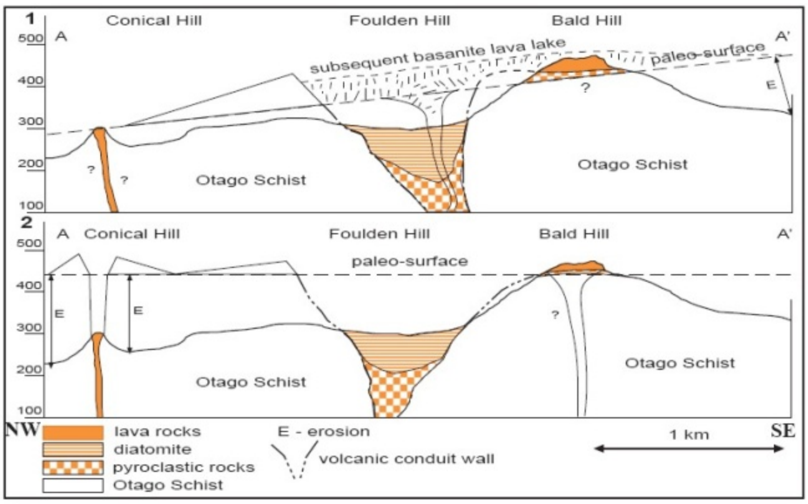


Fig. 26 – Foulden Hill - Bald Hill problem.

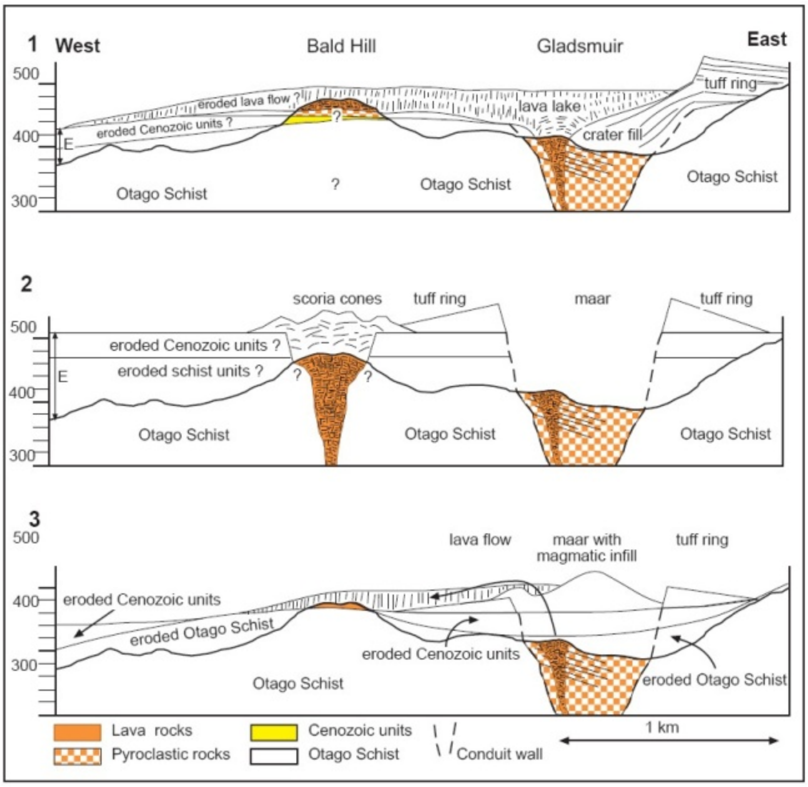


Fig. 27 – Gladsmuir – Bald Hill problem.

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.

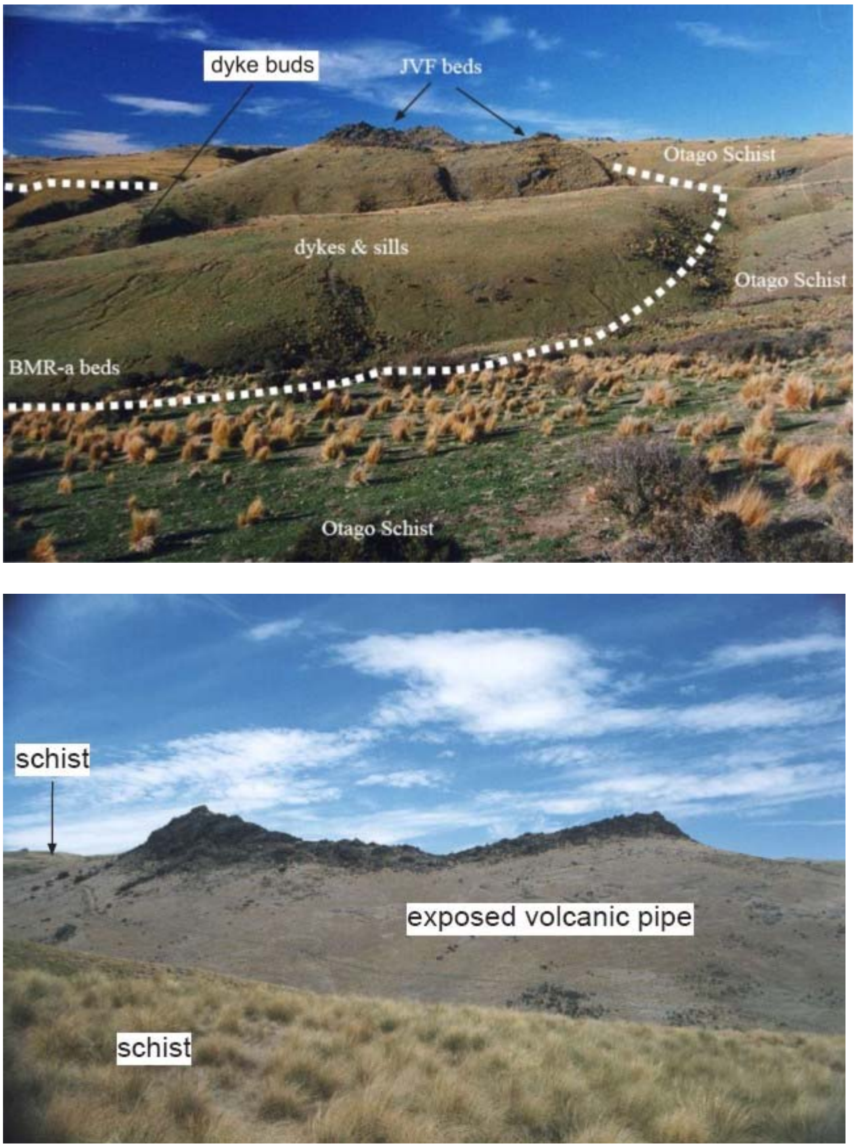


Fig. 28 – Overviews of “The Crater” diatreme.

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 bye 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)
Clast-supported			
Non-volcanic lithic-rich			
1 Massive	**TB1**	**LT1**	T1
2 Weakly-bedded	TB2	LT2	T2
3 Well-bedded	TB3	LT3	T3
Juvenile-rich			
4 Massive	**TB4**	**LT4**	T4
5 Weakly-bedded	TB5	**LT5**	T5
6 Well-bedded	TB6	LT6	T6
Matrix-supported			
Non-volcanic lithic-rich			
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
Juvenile-rich			
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.

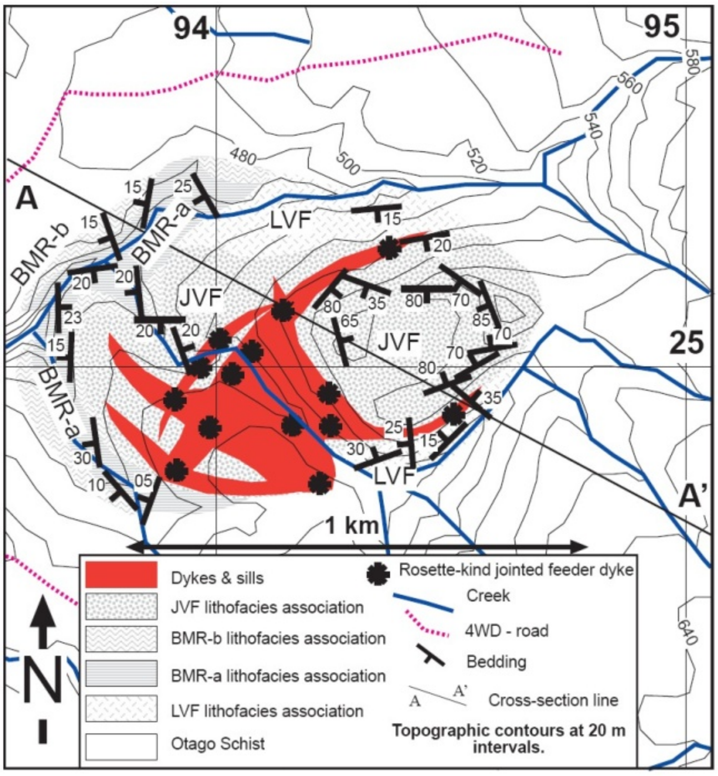


Fig. 29 – Volcanic lithofacies distribution of "The Crater" diatreme

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.

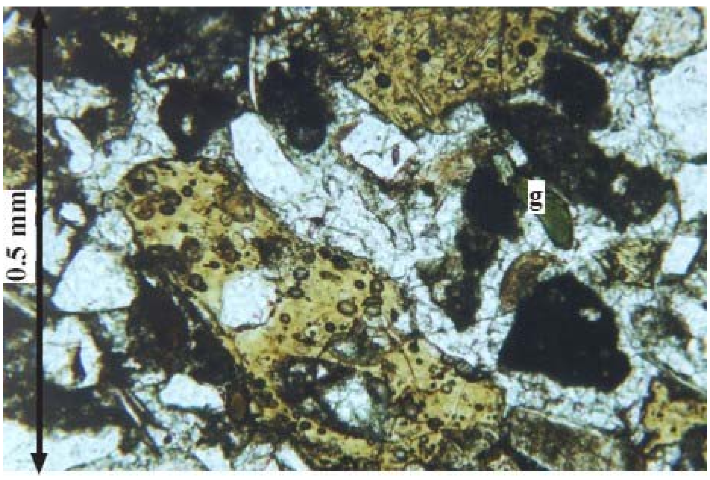


Fig. 30 – Sideromelane glass shards and glaucony in a fine grained tuff from "The Crater" diatreme.

LVF represents a vent filling succession, a possible lower diatreme, where diatreme is a pipe like volcanic conduit filled with pyroclstic 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 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 (main stop) – Swinburn lava sequence

Here we will walk up a paddock track to examine some textural aspects of the main Swinburn Plateau lavas and associated rocks, and to have an overview of the area including the large landslide that helped form the low cliffs with our outcrop.

At the truncated edge of the plateau here we can see a succession of scoriaceous lapilli tuff with small bombs (very poorly exposed) overlain by basalt with a coarsely vesicular base. The interior of the basalt, once > ~30 cm above the basal contact, is sufficiently coarsely crystalline that it has been mapped as dolerite. Locally cutting this generally coarse rock are zones of even more coarsely crystalline rock present in segregation structures. The most striking examples of this rock, exposed at the southern edge of the Swinburn Plateau, have crystals exceeding a centimetre in long dimension and can be considered basaltic pegmatites.

The scoriaceous lapilli tuff, which contains some bombs up to about 20 cm in length (Fig. 34), is important in demonstrating that the coherent igneous rocks of the plateau cannot have been intruded at depth. The Swinburn basalt body has been mapped in the past as dolerite sill, presumably based on its coarsely crystalline texture. There is no evidence of these Miocene basalts having intruded older, Oligocene, marine sedimentary rocks, however, and even if the coherent rocks did form as a "sill" intruding slightly older volcaniclastic deposits, it is extremely unlikely that intrusion depth could have been significant (i.e. that there was a high-standing, now-eroded, volcanic edifice into which a sill intruded). Accepting this, we must seek reasons other than slow cooling at depth to explain the textural characteristics of the rock.

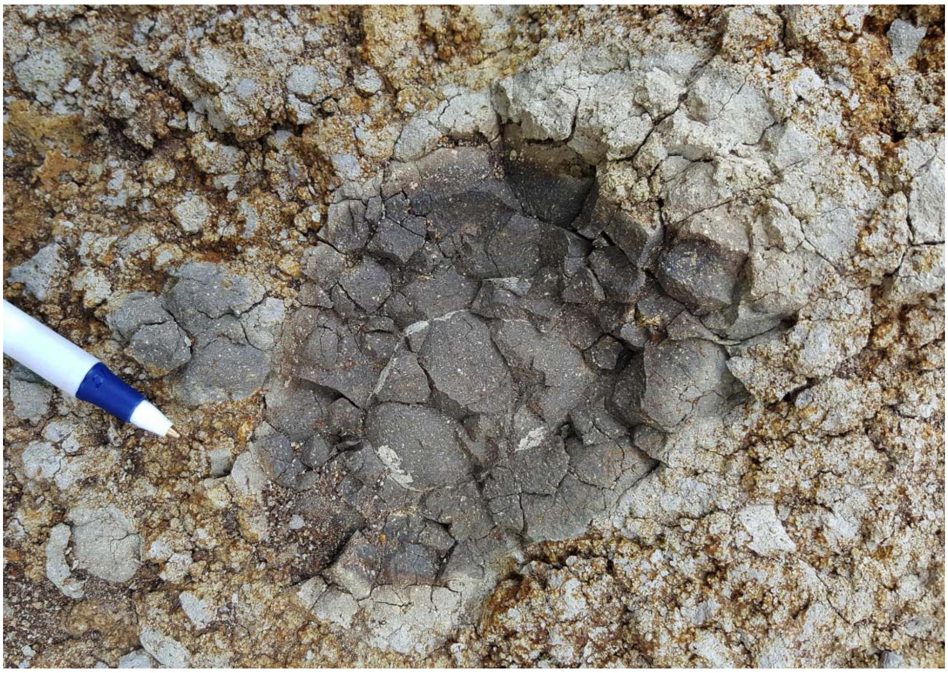


Fig. 34 – Small bomb in lapilli tuff underlying coherent basalt of the Swinburn Plateau.

Immediately overlying the lapilli tuff is about 30 cm of altered vesicular basalt, which grades upward into the "doleritic" rock. We interpret the succession of lapilli tuff overlain by coherent basalt with a strongly vesicular base as being a lava flow emplaced above pyroclastic rocks deposited by subaerial eruption. The thickness of this lava flow was probably modest – current cliffs are only a few metres high, and the low relief of the ground surface behind the cliff argues against deep incision/strong erosion of the rock. Despite the apparent modest thickness of the lava, however, much of it is rather coarsely crystalline (Figs. 35 and 36).

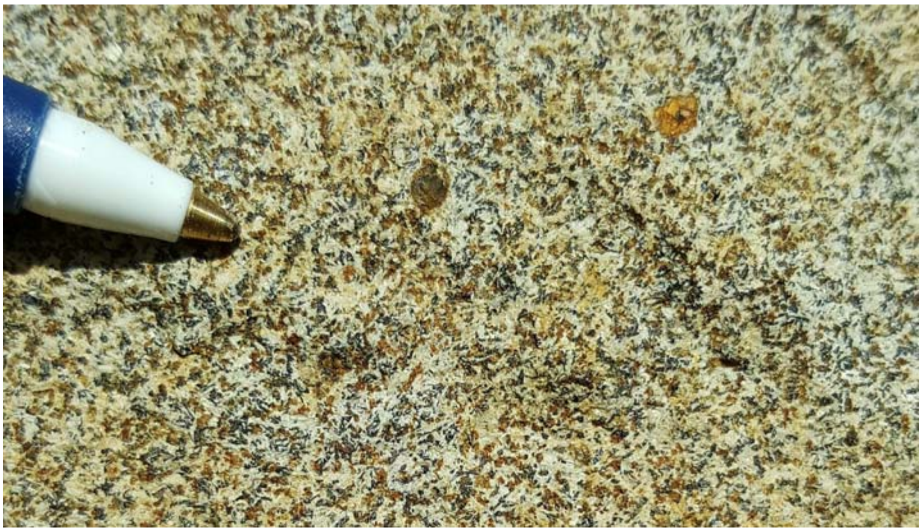


Fig. 35 – Close-up showing texture of most of the basalt making up Swinburn Plateau.

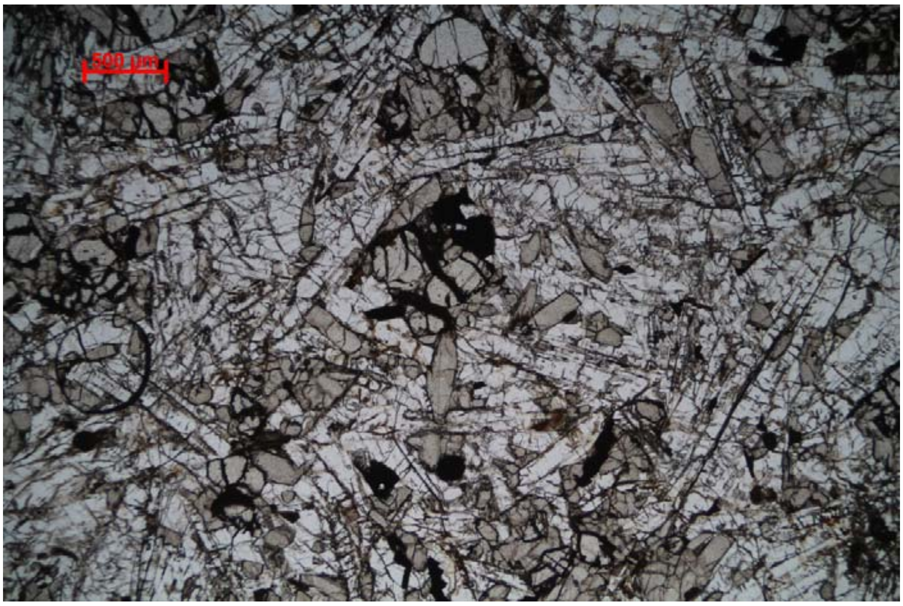


Fig. 36 – Photomicrograph of a "standard basalt" of the Swinburn Plateau.

Additional features found in the lavas, which we will see at this site, are coarsely crystalline domains of rock that cut through the interior of the lava flow. Figure 37 shows a small example of this texture, with a zone containing crystals commonly 2-5 mm in length enclosed within "standard" Swinburn basalt. These domains are interpreted to be basaltic segregations, crystallised from volatile-enriched melt that formed as the lava was cooling and crystallising, and which then intruded its parent lava.

The most coarsely crystalline rocks at Swinburn, associated with poorly outcropping "standard" Swinburn basalt, have crystals on the order of a centimeter in length, and can be interpreted as basaltic pegmatites (Fig. 38).

Fig. 37 – Domain of coarse crystals cutting across "standard" Swinburn basalt.

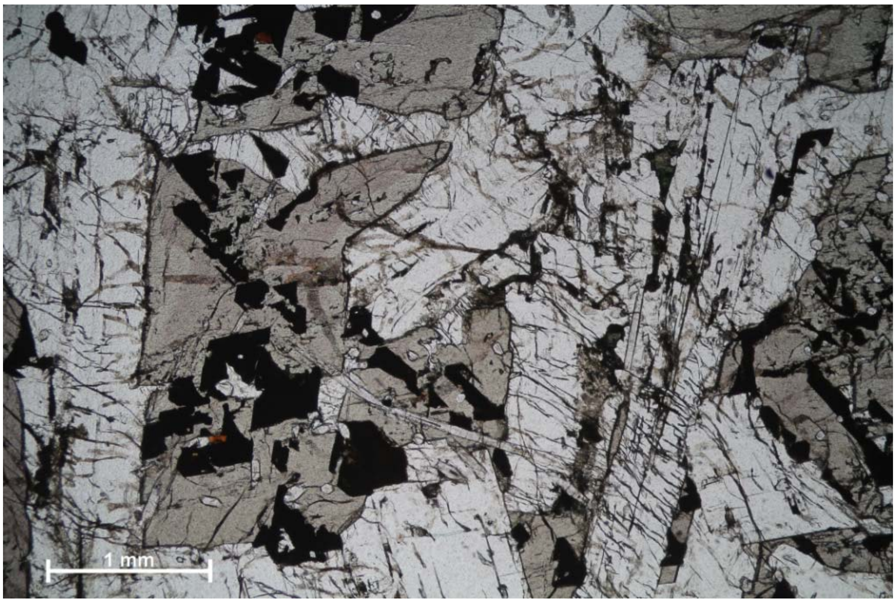


Fig. 38 – Among the most coarsely crystalline of basaltic rocks round at Swinburn plateau, these pegmatitic rocks are tentatively interpreted as having formed by separation of volatile-rich melt from the crystallising lava, which then intruded its parent/host lava by rising buoyantly to form large segregation structures.

Stop 6 (if time) – 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 absl), 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 absl). 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 masl) 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.

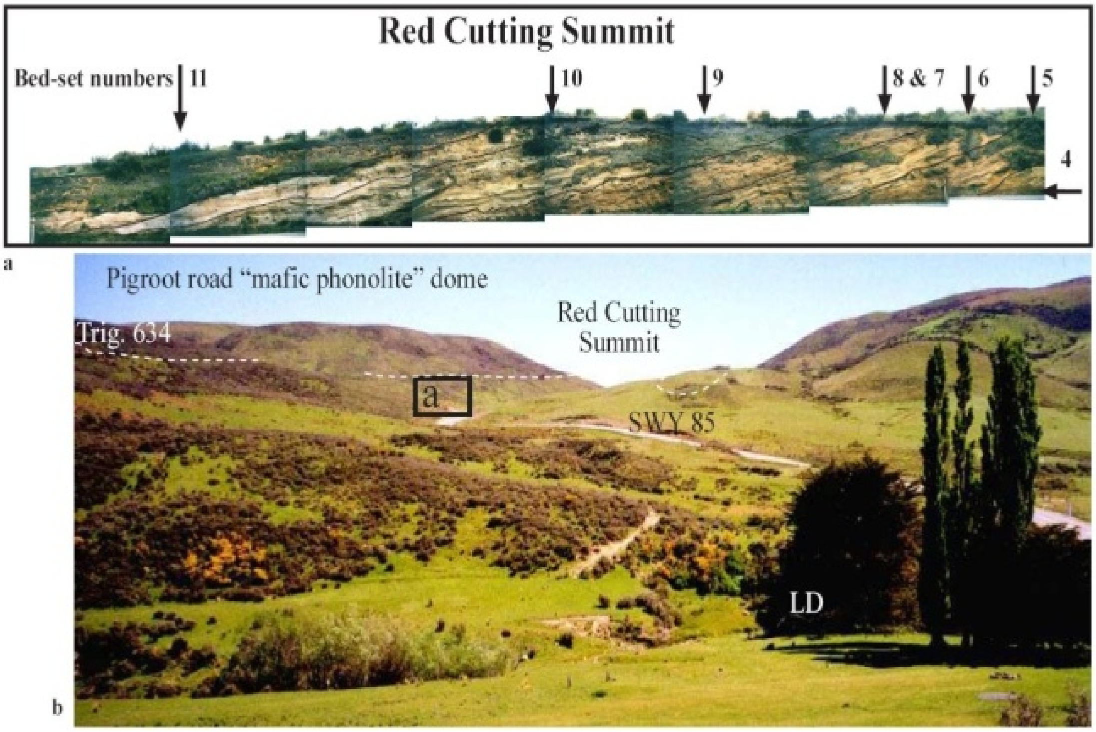


Fig. 39 – Red Cutting Summit road cut exposes a complex phreatomagmatic base surge and fall bed succession, part of a tuff ring around a phreatomagmatic volcano.

Approximately 500 m northwest from the “picnic area” a complex series of fine grained finely bedded lapilli tuffs and tuffs (Fig. 39) 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. , 40) a great variety of soft sediment deformation features (Fig 41), 4) dune bedded pyroclastic rocks (Fig.42) and 5) thickly bedded lapilli tuffs (LT9) rich in peridotite lherzolite and/or megacrysts.

Fig. 40 – Vesicular (vesiculated) tuff.

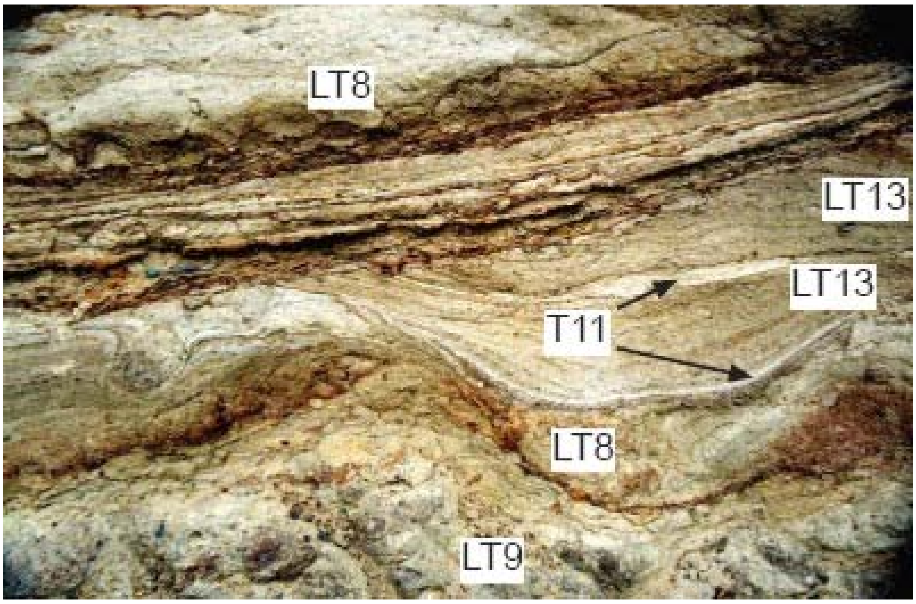


Fig. 41 – Soft-sediment deformation and impact sags in tuff and lapilli tuff beds of the Red Cutting Summit.

Fig. 42 – 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. 43). 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”.

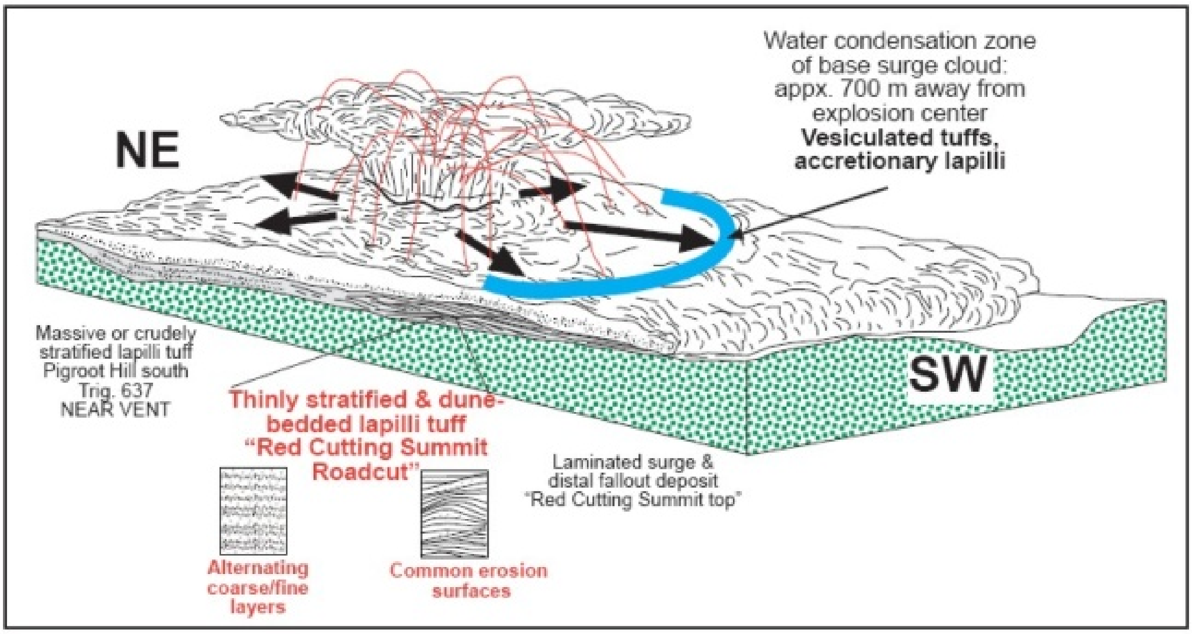


Fig. 43 – A potential 3D reconstruction of the depositional environment of the pyroclastic succession of the Red Cutting Summit road-cut.

Acknowledgements

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