Phreatomagmatic and water-influenced Strombolian eruptions of a small-volume parasitic cone complex on the southern ringplain of Mt. Ruapehu, New Zealand: Facies architecture and eruption mechanisms of the Ohakune Volcanic Complex controlled by an unstable fissure eruption

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Abstract

The Ohakune Volcanic Complex is a late Pleistocene tuff ring–scoria/spatter cone complex located south of Ruapehu volcano. This small-volume volcano consists of an outer E-W elongated compound tuff ring edifice, three inner scoria-spatter cones and further volcanic depressions, located on the Ohakune Fault. We quantified accurately the variations of the eruptive styles and processes through time by systematic sampling of key stratigraphic marker beds at proximal and distal locations, and the determination of grain size distribution, componentry, density and vesicularity. Using a Digital Terrain Model coupled with stratigraphic data, we also determined the spatial distribution and volume of each identified unit and individual edifices within the Ohakune Volcanic Complex. Activity began with a shallow phreatomagmatic phase characterized by an almost continuous generation of a low eruptive column, accompanied by wet pyroclastic density currents, together with the ejection of juvenile fragments and accidental lithics from the surrounding country rocks. Subsequent activity was dominated by a variety of Strombolian eruptions exhibiting differing intensities that were at times disrupted by phreatic blasts or phreatomagmatic explosions due to the interaction with external water and/or sudden changes in magma discharge rate. At least three major vent-shifting events occurred during the eruption, which is demonstrated by the truncation of the initial tuff ring and the filling of the truncated area by several coarse grained surge units. Our study indicates that approx. 12 × 10⁶ m³ DRE magma erupted within maximum 2.5 to 5 months through multiple vents. The erupted magma ascended from a depth of 16–18 km, and reached the surface within approximately 50 h. Alternating eruption styles, frequent vent-shifting and a variety of emplacement mechanisms inferred from the deposits of the Ohakune Volcanic Complex demonstrate the unpredictable nature of small-volume volcanism, contributing to our understanding of hazards attributed to this type of activity. Future small-volume eruptions within the Taupo Volcanic Zone may be of similar nature.

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Fissure vent
Tuff ring
DRE volume
Digital Terrain Model

1. Introduction

Monogenetic (or small-volume) basaltic volcanoes occur within every geodynamic regime of the Earth and their associated volcanism is characterized by a range of eruption styles, eruptive materials and geomorphology (Sheridan and Wohletz, 1983; Lorenz, 1986; Head and Wilson, 1989; Valentine and Gregg, 2008; Németh, 2010; White and Ross, 2011; Kereszturi and Németh, 2012; Németh and Kereszturi, 2015). The determining factor of the eruptive behaviour of a volcano and the resulting volcanic products is the type of magma fragmentation (Lorenz, 1973; Walker, 1973; Kokelaar, 1983; Parfitt, 2004). Magmatic or dry fragmentation is characterized by exsolution and quick expansion of initially dissolved volatiles (e.g. H₂O and CO₂). The ascending bubbles coalesce and collapse near the surface and the exiting gas bursts result in Strombolian or Hawaiian eruptions (Parfitt, 2004; Houghton and Gonnermann, 2008; Valentine and Gregg, 2008). The main differences between these two eruptive styles are the speed of the gas flux with respect to magma rise (Parfitt, 2004) and differences between the ejection mechanisms of fragments during eruptions (e.g. the duration of particles remaining as ballistics is greater in Strombolian eruptions, and particles transported by fallout jets is more common in Strombolian activity) (Riedel et al., 2003; Capaccioni and Cuccoli, 2005). Hawaiian eruptions exhibit higher magma flux that allows limited cooling within the fire-
fountain, which results in agglutination or fusing due to the high accumu-
lion rate of the fragments (Head and Wilson, 1989), while the clasts from
Strombolian eruptions cool prior to landing (Walker, 1973; Ver-
gniolle and Mangan, 2000). Phreatomagmatic or wet fragmentation refers to the interaction of magma and external water resulting in the conversion of magmatic heat to mechanical energy (Sheridan and Wohletz, 1983; Lorenz, 1985; Wohletz, 1986; White, 1996; Wohletz and Zimanowski, 2000; Lorenz, 2003; Keresztturi and Németh, 2012). The interaction and the resulting volcanic processes in most cases primarily depend on the water/magma ratio, lending to two basic types of fragmentation: phreatomagmatic (Taalian) and Surtsyean (Kokelaar, 1983; Wohletz and Sheridan, 1983; White, 1996; Keresztturi and Németh, 2012). During Surtsyean-style eruptions the water/magma ratio is high enough (e.g. in a shallow subaqueous environment) to allow continuous bulk mixing and fragmentation. As a result, the Surtsyean-style eruptions are characterized by quickly recurring, short-lived tephra jets, causing the rapid growth of a volcanic pile and the growth of a tuff cone (Thorarinsson, 1965; Kokelaar and Durant, 1983).

In contrast, phreatomagmatic eruptions are characterized by lower water/magma ratios in the subaerial environment, in which fragmentation is triggered by a molten fuel-coolant interaction (MFCI) process causing energetic thermo-hydrodynamic explosions (Sheridan and Wohletz, 1981; Wohletz, 1986; White, 1996; Wohletz and Zimanowski, 2000; Büttner et al., 2002). The phreatomagmatic eruptions commonly form two types of volcanoes, depending on the substrate (Lorenz, 2003; Auer et al., 2007) and the depth of excavation: tuff rings and maars. Both types have an ejecta ring around their craters, but in contrast with tuff rings, maar craters cut deeply into the pre-eruptive surface and are underlain by a mixture of volcanic and country rock fragments composed of debris commonly referred to as the diatreme (Wohletz and Sheridan, 1983; Martin and Németh, 2005; White and Ross, 2011; Keresztturi and Németh, 2012; Agustin-Flores et al., 2014). During phreatomagmatic eruptions the explosion loci often exhibit a downhill or lateral migration according to the hydrologic and local substrate settings or following regional faults as it documented at Ukinrek Maars, Alaska (Kienle et al., 1980), Cerro Xalapaxo (Abrams and Siebe, 1994) and Técuitlapa Maar (Ort and Carrasco-Núñez, 2009), central Mexico, Tihany maars (Németh et al., 2001), and Yangpori diatreme, SE Korea (Son et al., 2012). Newer models presume that MFCI occurs at varying depths within the diatreme during the course of the activity, although it is most effective at shallow depth, where the hydrostatic pressure is less than the critical pressure. Thus, deep-seated explosions are rarely powerful enough to erupt (Valentine and White, 2012; Graettinger et al., 2015). As commonly documented, longer-lived eruptions of subaerial volcanoes and scoria cones quickly deplete the water from the substrate or surface sources and gradually become dry explosive and/or effusive eruptions, such as Ság-hegy, western Hungary (Martin and Németh, 2004), Cosa de Sant Dalmai maar, NE Spain (Pedrazzi et al., 2014b), and Motukorea tuff ring, Auckland Volcanic Field, New Zealand (Agustin-Flores et al., 2015). An alternating pattern of phreatomagmatic and magmatic activity may also be exhibited due to vertical or lateral migration of eruption loci, providing new influx of water from distinct aquifers (e.g. Rothenberg volcano, East Efie; Tihany maars, Hungary; Atexacca maar, eastern Mexico) (Houghton and Schmincke, 1986; Németh et al., 2001; Carrasco-Núñez et al., 2007), or due to change in magma flux as documented at White Island, New Zealand (Houghton and Nairn, 1991).

The Ohakune Volcanic Complex (OVC), North Island, New Zealand was previously examined by Houghton and Hackett (1984). Their study was based on the examination of outcrops in quarries that were cut into the scoria cones and the tuff rings. Since their work the extent of the main quarry has increased significantly, exposing a much larger internal portion of this small-volume volcano. The currently exposed volcanic units show evidence for new vent openings and simultaneous activity of distinct vents characterized by different eruption styles resulting in a very complicated volcanic architecture. Due to the continual influence of external water and the relatively small volume of magmatic eruptions in the final phases, volcanic features from the embryonic stage of the activity were not completely covered, making the OVC one of the few known locations that retains vital information on the evolution of monogenetic volcanoes affected by complex changes of vent loci (Houghton and Schmincke, 1989; Pedrazzi et al., 2014b; Agustin-Flores et al., 2015). Previous studies concentrated on the central vent volcanism and related hazards (e.g. Cronin et al., 2003; McClelland and Erwin, 2003; Graettinger et al., 2010; Pardo et al., 2012), rather than focusing on the hazards produced by such a volcanism within the Taupo Volcanic Zone (TVZ) in terms of eruption cloud height, eruption volumes and potential longevity of an eruption. This study is the first step towards assessing the volcanic hazards associated with small-volume volcanism with respect to the relationship between parasitic/satellite volcanoes and their larger polygenetic counterparts of the TVZ. We focus on the understanding of the eruptive conditions including influencing factors of fragmentation, dominant explosive and depositional mechanisms, the range of the eruptive products created by different eruptive styles and the preserved geochemical characteristics. Furthermore new Geographical Information Systems (GIS) techniques allow utilizing a high resolution Digital Terrain Models (DTM) in order to study the geomorphologic features, providing great support in the understanding of the temporal evolution of this monogenetic volcano and contributing to the identification of the hazards of future small-volume eruptions within the TVZ (Fig. 1).

2. Regional setting

The late Pleistocene monogenetic OVC (also known as Rochfort Crater) is located within the south-southwest ringplain of Ruapehu stratovolcano (Fig. 1). Volcanic activity in the southern end of the TVZ dates back to the eruption of Hauhungatahi volcano (Fig. 1), which has been dated at 921 ± 40 ky by 40Ar/39Ar (Cameron et al., 2010). This earliest activity was followed by the construction of primeval edifices of Ruapehu, which are mostly eroded and overlapped by the present Ruapehu volcano about 250 ky ago (Hackett and Houghton, 1989; Tost and Cronin, 2015). The OVC, along with another monogenetic volcano (Rangataua Craters), are the southernmost appearance of volcanic activity within the TVZ (Fig. 1) (Cole et al., 1986; Hackett and Houghton, 1989). The surrounding county rocks are mostly reworked volcanoclastic sediments of Mt. Ruapehu and andesitic to rhyolitic fallout tephra from various sources (Hackett and Houghton, 1989; Möbis, 2010). Pliocene marine sediments (mostly siltstones and sandstones of the Matemateaonga Formation) underlie the volcanic edifice and volcanoclastic series, and occasionally crop out as uplifted fault-bounded blocks (Townsend et al., 2008). The eruptions of the OVC are most likely linked to the E-W trending Ohakune Fault (Fig. 2) (Houghton and Hackett, 1984). The eruptions started in a flat or slightly sloping laharc depositional environment, where the unconsolidated reworked volcanoclastic sediments could have been divided by newly formed river pathways. The youngest sediments of the pre-eruptive surface sourced from the lavas of the 60–15 ky old Mangawhero Formation while the edifice is overlain with disconformity by the Kawakawa/Oruanui Tephra, which gives a 25.4 ka minimum age for the activity (Hackett and Houghton, 1989; Vandergoes et al., 2013). Radiocarbon dating on charred wood from an OVC associated coarse lapilli layer from a distal outcrop (Fig. 2) yields a date of 31,500 ± 300 y BP (Froggatt and Lowe, 1990).

3. Materials and methods

3.1. Field observations and sampling

The proximal sequence of the ejecta ring (outer edifice) was examined in terms of thickness, colour, bedding structures and grading, componentry and grain morphology. We also examined the distal sequence in order to correlate the existing units (Fig. 2). High resolution georeferenced thickness data of individual eruptive units from within
the eastern ejecta ring were collected by Terrestrial Laser Scanner (TLS) (Fig. 2). The quarry walls were surveyed using approximately 15 mm point spacing at 30 m distance by TLS, along with high resolution photographs. The units named as A–E, G, H, I, K–M, O–Q and U (Fig. 2) were sampled from the proximal facies of the edifice within the quarry. In addition, large juvenile fragments (>10 cm in diameter) were collected from the massive ballistic beds (A, D, H, J, T, X) (Fig. 2) for density, petrography and geochemical analysis. The surface morphology of the ejecta ring was mapped using TLS and RTK GPS to upgrade remotely sensed Digital Terrain Model (DTM) (Gómez-Vasconcelos et al., 2016) for edifice volume and unit volume calculations.

3.2. Sample preparation and analytical techniques

The grain size distribution of the matrix supported units were defined using phi sizes from $-5 \phi$ (32 mm) to $4 \phi$ (0.63 mm) at half phi intervals. Particles smaller than $4 \phi$ were analysed by a Partica LA-950V2 Laser Scattering Particle Size Distribution Analyser at Massey University. Grain size parameters were calculated using the Gradistat 8.0 program (Folk and Ward, 1957; Walker, 1971; Blott and Pye, 2001). The componentry and the morphological characteristics of the fragments were examined at $-3, 0.5$ and $2 \phi$ under a stereomicroscope and a FEI Quanta 200 environmental Scanning Electron Microscope (SEM). The majority of samples had to be cleaned in an ultrasonic bath in order to disaggregate the adhering fine material. The microscopic observations were complemented with point counting of stereomicroscopic images using the JMicr0Vision 1.2.7 software. Five subsamples of blocks and bombs from each of the bomb-bearing units (A, D, H, J, T, X) were examined to determine envelope densities and vesicularities, which was carried out using a Micrometrics GeoPyc 1360 analyser. We also measured the densities for $30–3 \phi$ clast from A, D, K, M, P units. Vesicularity and clast microtextures of the blocks/bombs were determined using 2D macroscopic and SEM image analysis for 10 samples from 6 units (Table 1). Macroscopic high resolution images have been taken from each sample (with an extent of 40 to 60 cm²) by a photo and document scanner (Canon Pixma MP160), while the smaller vesicles were mapped through backscattered electron imaging (BSE) of thin sections striving to the requirement of representativeness. Two macroscopic images and nine microscopic images were captured at 25× and 100× magnifications ensuring that the resolution was sufficient to capture all size ranges of bubbles with appropriate resolution. 2D vesicularity was measured using ImageJ and the Matlab™-based FOAMS (Fast Object Acquisition System) software.
and Measurement System) software by the method of exponential image nesting (Shea et al., 2010). Applying stereological conversion of vesicles by FOAMS provided further details of vesicle textures (e.g. vesicle size and vesicle volume distributions), vesicle size and shape properties which preserve information on the eruptive conditions (e.g. bubble nucleation or degassing) (Shea et al., 2010). Geochemical compositions were determined for single ash particles from matrix-rich (A–E, G–I, L) units (Fig. 2) at the size of 0.5 ɸ with a JEOL JXA-840 Electron Microprobe Analyser (EMPA) equipped with a Princeton Gamma Tech Prism 2000 Si(Li) EDS X-ray detector at Massey University, employing an accelerating voltage of 15 kV and a 10 ɥm defocused beam. Groundmass glass of spatter bombs were analysed with defocused beam and olivine and pyroxene crystals of the scoriaceous fragments from the bomb beds (D, H, J, T, X units) with a focused beam by EMPA. Elemental analysis and mapping of thin sections was accomplished by energy dispersive x-ray spectroscope (EDAX) of SEM.

4. Architecture of Ohakune Volcanic Complex

The OVC has previously been described as a kilometre wide slightly elongated tuff ring with intra-crater scoria cones (Houghton and Hackett, 1984). Based on a DTM (Horizons Regional Council, 2016), aerial photo from 1951 and field observations, the northern section of the tuff ring appears undeveloped due to lack of space between the vents and the Ohakune fault scarp. Only scoriaceous beds are exposed at the eastern section of the edifice that correlate with the pyroclastic successions at higher stratigraphic levels (Figs. 2 and 3). Furthermore dense vegetation makes access difficult, and erosion by the Mangawhero River makes reconstruction difficult in this part of the volcanic complex. It appears that the southern section may also not be a single edifice, but rather consists of coalesced ejecta rings formed by alternating phreatomagmatic and Strombolian eruptions from separate vents (Figs. 2 and 3). For these reasons, the outer complex edifice is not an ordinary tuff ring (Lorenz, 1986; Kereszturi and Németh, 2012), and we will refer to it here as an ejecta ring. There are at least 5 depressions between the ejecta ring structures and the fault plain, which probably reflects the location of the eruptive vents that were active during the final eruptive phases (Fig. 2). Most of these pits form a 600 m long fissure zone subparallel with the Ohakune Fault, which may indicate that the area was penetrated by the extruding tips of a blade-like dyke. Three of the depressions host the craters of a large and two small late stage scoria/spatter cones (Fig. 2).

The proximal sequences of the inner and outer edifices are exposed at several quarries. The largest inner scoria/spatter edifice (CSC) (Figs. 2 and 3) has a bedded structure of coarse scoria and agglutinated scoria units, alternating with clastogenetic lava flow units (Fig. 3). The majority of the edifice of the scoria cone has been quarried away except the north section. The proximal sequence of the ejecta ring is very well exposed in a quarry that is currently active (Figs. 3 and 4); however the contact with the country rock is not visible (Fig. 4). Geomorphologic observations, coupled with Real-Time Kinematic (RTK) GPS measurements, showed that the pre-eruptive country rocks are located at the altitude of 585–590 m above sea level, thus about 10 m of the initial sequence has not yet been exposed. The lowest visible units are light-coloured and matrix-rich, while the upper units appear to be the scoria-dominated units alternating with units of pyroclastic density currents and poorly-sorted units of weathered scoriaceous fragments with denser lava blocks. The distal sequence comprises a 1.8-thick unit that is situated about 600 m south of the OVC on the left side of Mangawhero River (Fig. 2). The distal outcrops have been separated from the OVC.
by fluvial erosion prior to the Oruanui eruption (25.4 ky) and those deeply cut channels have been later occupied by fluvial sediments as part of the formation of an alluvial fan (Fig. 2).

5. Stratigraphy and sedimentology

The active quarry exposes the coalesced eastern ejecta ring (EER) (Figs. 2 and 3). The walls are north-west radially orientated to the ejecta ring, providing a proximal to distal cross-section of the volcano. At the NW corner of the quarry we identified an unconformity. In relation to the unconformity a small fault is also visible with decreasing offsets from the base to the top (Fig. 4). The lowest, barely exposed units are thick-bedded and poorly-sorted with accidental lithics from the laharic basement and dense lava fragments from the OVC eruption alternating with massive ash beds. The upper sequence of the ejecta ring is characterized by dark-coloured bomb beds alternating with light-coloured ash beds and mixed beds through the approximately 22 m thick visible succession. 95% of the fragments are clearly associated with the magma erupted from OVC, the most distinctive feature of which is the high content (10–15%) of pyroxene and olivine phenocrysts up to 0.8 cm in size. The remaining accessory fragments are mostly well-rounded andesitic boulders of the fluvial-laharic basement originated from Ruapehu and a small percentage of baked sedimentary rocks (Fig. 4). The exposed units were labelled (Fig. 4) and classified into 4 lithofacies on the basis of sedimentological characters (e.g. componentry, grain size distribution and sedimentary structures) (Fig. 4): 1. Ab: ash beds (C, E, G, L, N, O, Q, S units), 2. AbC: ash beds with coarse fragments (A, I, U), 3. BbA: bomb beds with ash (B, D, H, K, P, R, XLower), 4. Bb: spatter-scoria bomb beds (J, M, T, XUpper).

5.1. Ash beds (Ab)

5.1.1. Description

Ash beds are matrix-supported massive, diffusively stratified or planar bedded with internal low angle unconformities. The beds are poorly-sorted and predominantly composed of fresh and varying amounts of yellowish to brownish coloured palagonitized ash and lapilli fragments from −3 φ to 7 φ in size with a peak at 3.5 φ. Mud-coating is frequent on the surface of fragments. Some ash beds exhibit stratification by the alternation of finer and coarser layers, where the coarsest layers often show inverse grading (e.g O1 layer). The tephra has a damp, adhesive behaviour, which is brightening after drying. Impact
sags are rare. The biggest clasts (up to 20 cm) can be altered vesicular or dense scoriaceous fragments correlated to the Ohakune eruption and rounded andesite cobbles of Ruapehu origin. The thickness of the units is variable (≤50 cm thick), and visible single layers are in the range of 1 cm to 25 cm thick. Individual beds usually exhibit irregular thickness variations without appreciable thinning trends within the quarry (Fig. 4). The distal counterparts are mostly thinly stratified ash (−1 to 7 ϕ in size) with low angle cross and dune bedding structures. The distal sequence also exhibits soft sediment deformation and impact sags with ballistic bombs up to 10 cm in size.

5.1.2. Emplacement interpretation

The high proportion of mud-coated fragments and the wet and massive nature of beds imply an abundance of water during the eruptions and during deposition (Heiken and Wohletz, 1985; Sohn and Chough, 1989; Houghton and Smith, 1993). The clearly visible cross and dune bedding of the distal deposits confirms the dilute pyroclastic density current origin of these deposits. Palagonitized ash fragments and fine dust is probably derived from the slurry of the crater floor. Hence, minor recycling of juvenile fragments from the earlier products of the OVC activity must have occurred (Houghton and Smith, 1993; D’Oriano et al., 2014). The majority of coarser fragments are likely to have been transported by ballistic trajectories or in pyroclastic density currents at the proximal section, while the coarser fragments clearly have a ballistic origin at the distal sequence (Fisher and Schmincke, 1984; Sohn and Chough, 1989). Therefore, Ab units most likely deposited as a result of discrete phreatomagmatic explosions generating wet pyroclastic surges and some tephra fall.

5.2. Ash beds with coarse fragments (AbC)

5.2.1. Description

Ash beds with coarse fragments are matrix-supported, very poorly-sorted units with variable amounts of coarse lapilli and bomb size clasts (Fig. 6). The matrix is composed of fresh juvenile, mud-coated and palagonitized ash and medium lapilli fragments between −3 to 7 ϕ in size with the main peak at −2.5 to −1.5 ϕ. The larger sized lapilli and bombs are composed of rounded andesite cobbles and fresh olivine/pyroxene-bearing basaltic fragments up to 50 cm in size. Some of the lava fragments are vesicular with a dense core exhibiting conchoidal fractures, others have cauliflower or plastically deformed shapes. None of the clasts have formed visible impact sags. The fraction of
these larger clasts is significantly higher than in the Ab units. AbC units often have internal bedding without sharp depositional or erosional contacts. These units were typically interstratified with subunits that are analogous to Ab-type lithofacies traverses through the outcrop (Fig. 4).

5.2.2. Emplacement interpretation

The lack of the sharp internal structures and impact sags suggest continuous deposition by fallout and debris jets, while the relatively coarse average size as well as the smaller portion of very fine (<3 ɸ) juvenile fragments imply less energetic fragmentation than in the Ab units. The massive beds of AbC interstratified Ab beds may be the result of a sustained, but varying level of interaction of magma with external water at shallow depth, which enables a near-continuous discharge of tephra and gases coinciding with more energetic discrete explosions. The continuous uprush produced a complex deposition of fresh dry ash, recycled damp ash and aggregates from the contact zone of the dyke edge and conduit wall, similar to the 1976–1982 eruption of White Island, New Zealand (Houghton and Nairn, 1991). However, the fresh, sometimes plastically deformed bombs (in unit A) suggest that the middle part of the dyke could have produced drier Strombolian-type fragments occasionally. Strombolian/Hawaiian and phreatomagmatic eruptions usually represent very different types of magma fragmentation (Wohletz and McQueen, 1984), however there are a few documented eruptions that exhibit both fragmentation processes at the same time (Moore et al., 1966; Kienle et al., 1980; Dzurisin et al., 1995; Valentine et al., 2000). There are some alternative models for magma fragmentation and admixture of tephra from two contrasting eruptions. Depending on variations in vent geometry and hydrological conditions the products could have originated from different parts or depths of the same vent or fissure (Houghton and Schmincke, 1986; Dzurisin et al., 1995; Valentine and White, 2012) or from two or more localized vents along a fissure (Moore et al., 1966; Kienle et al., 1980; Pedrazzi et al., 2014a; Belousov et al., 2015). Alternatively, these units can be interpreted as products of a long-lasting transitional phase between phreatomagmatic and magmatic phases of Strombolian/Hawaiian eruption (Belousov et al., 2015), which might have been emplaced by spatter/scoria-bearing pyroclastic density currents (Valentine et al., 2000). One of the AbC-type units (U) is interstratifying...
with a bomb bed (unit X), which indicates that the two units did not form by explosions from a single vent. The fragmentation during stronger discrete eruptions produced a higher amount of fine ash deposited by wet pyroclastic surges and tephra fallout. The slightly higher proportion of accidental fragments that originate from the shallow laharic basement demonstrates that the eruption loci were still of shallow depths without notable excavation.

5.3. Bomb beds with ash (BbA)

5.3.1. Description

The BbA lithofacies is clast-supported and predominantly made up from mud-coated, weathered (oxidised at very proximal locations) and fresh juvenile lapilli and bomb sized fragments with various amounts of fines, usually without stratification (Fig. 4). The ratio of the ash and lapilli/bomb fragments can vary vertically within the same unit. In a few locations the ash rich portions are arranged into hardly distinguishable graduated sheets. The majority of coarse juvenile fragments are 3–10 cm in size with a maximum of 25–30 cm in diameter in loosely packed clast-supported beds. The bigger clasts sometimes display shapes generated by plastic deformation. Some of these flattened clasts have experienced agglutination. Other clasts are dense lapilli and blocks with fractured edges exhibiting various degrees of alteration. At the NW corner of the quarry the fragments are oxidised, but in other locations yellowish to brownish alteration of clasts can be observed. The ash componentry is identical to the other ash-rich units. Mud-coating is common to all clasts and at all grain size classes. Ballistic bomb sags are not visible, but the impact of the biggest andesitic boulders can form minor distortions. The thickness of the units can vary between 0.3 and 3.5 m.

5.3.2. Emplacement interpretation

The tephra associated with BbA units have most likely originated from two sources. The average grain size, the shape of the coarsest fragments, lack of matrix, and the dominance of moderate to highly vesiculated juvenile fragments suggest that these units have originated from Strombolian-style eruptions (McGetchin et al., 1974; Houghton and Hackett, 1984). The documented clast flattening and agglutination imply occasionally higher retained heat, which might have been caused by the fluctuation of the effusion rate during Strombolian activity. However the varying proportions of fine material, componentry (e.g. mud-coated fragments and aggregates), fractured, altered juvenile blocks and accidental boulders all suggest that fresh juveniles erupted through slurry-filled craters. The altered earlier fragments were mixed with Strombolian products. The adhering dust on the surface of clasts across a wide spectrum of grain sizes suggests a broad recycling and in-situ

<table>
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<th>2D BV (Image)</th>
<th>2D BV (%)</th>
<th>Nv (corr) (mm⁻³)</th>
<th>EqDi range (mm)</th>
<th>Mean EqDi (mm)</th>
<th>Mean Area (mm²)</th>
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<td>1.7 × 10³</td>
<td>0.005–14.895</td>
<td>0.308</td>
<td>0.397</td>
<td>0.8147</td>
<td>0.6050</td>
</tr>
<tr>
<td>X2</td>
<td>39.8</td>
<td>38.1</td>
<td>22.1</td>
<td>6.5 × 10²</td>
<td>0.005–2.622</td>
<td>0.177</td>
<td>0.070</td>
<td>0.8333</td>
<td>0.5577</td>
</tr>
<tr>
<td>X3</td>
<td>26.4</td>
<td>44.1</td>
<td>27.2</td>
<td>1.46 × 10³</td>
<td>0.006–1.341</td>
<td>0.172</td>
<td>0.036</td>
<td>0.8481</td>
<td>0.4931</td>
</tr>
</tbody>
</table>
palagonitization of the glassy rinds of pyroclasts (Houghton and Smith, 1993). Strombolian/Hawaiian and phreatomagmatic eruptions repre-
sent usually very different type of magma fragmentation (Wohletz
and McQueen, 1984), however, there are a few documented eruptions
that exhibited both fragmentation process at the same time (Moore
et al., 1966; Kienle et al., 1980; Dzurisin et al., 1995; Valentine et al.,
2000). The Bb unit products did not form by magmatic fragmentation
alone. The high portion of dust-adhered fragments demonstrates an
abundance of water in the eruption plume. There are two alternative
models for the magma fragmentation and mixture of tephra from two
contrasting eruption styles. Depending on the variations of vent geometry
and hydrological conditions the products could have originated from
different parts of the same vent or fissure (Houghton and Schmincke,
1986; Dzurisin et al., 1995) or from two or more localized
vents along a fissure (Kienle et al., 1980; Pedrazzi et al., 2014a;
Belousov et al., 2015).

5.4. Spatter-scoria bomb beds (Bb)

5.4.1. Description
Spatter-scoria bomb beds are clast-supported without any internal
bedding. The upper boundary is sharp, but the lower can be sharp or
interfingering with Abc-type units. Bb units are made up from oxidised
or fresh juvenile fragments of plastically deformed and cauliflower-
shaped coarse lapilli and blocks or bombs with minor amounts of medi-
um lapilli (0.5–1 cm in size). The grain size distribution of these beds
implies better sorting than the other lithofacies. The average clast size
ranges between 3 and 10 cm, the maximum clast size is observed up
to 30 cm. These units often exhibit agglutination as well as locally con-
 fined welding (Fig. 4).

5.4.2. Emplacement interpretation
The lack of admixed ash, the plastically deformed fragments and the
short dispersal range of the rapidly thinning beds suggest that Bb beds
originated from vents characterized by low energy Strombolian and/or
Hawaiian-style activity. However, the existence of the large number of
dense cauliflower-shaped blocks suggests rapid cooling prior to eruptions
(Kienle et al., 1980). In addition, the interfingering ash-dominated
lithofacies (Ab, Abc) demonstrates that other phreatomagmatic vents
could have been active at the same time referring to the abundance of
water to other parts of the dyke.

6. Density and vesicle microtexture analysis

Clast densities and vesicle textures (e.g. vesicle size and shape, vesic-
le number densities) provide information on characterizing the condi-
tions of magma ascent and degassing, and may also elucidate the changing conditions within the conduit between consecutive eruptive
phases (Houghton and Wilson, 1989; Perfitt, 2004; Moitra et al.,
2013). The skeletal (absolute) density determined for blocks/bombs
was 2.85–2.90 g/cm². The bombs/blocks exhibit different envelope den-
sities than the lapilli size clasts. The lapilli (−3 4) average densities are
between 1.27 and 1.33 g/cm³, while the bombs have densities of 1.52–
1.54 and 1.72–1.82 g/cm³. The modes show a similar pattern of 1.0–
1.2 g/cm³ for lapilli size and 1.5–1.7 g/cm³ for bomb/block size
(Fig. 6). The 3D vesicularity was calculated applying a 2.85 g/cm² densi-
ty, yielding values significantly higher (26.4–51.4%) than the FOAMS-
calculated integrated 2D vesicularities (22.1–31.6%). The 2D shape of
the bubbles usually exhibits irregularities (polylobate shapes) due to
frequent coalescence inducing a large-scale interconnectivity of the ves-
icles (Fig. 7, Table 1). Most of the equivalent diameters of vesicles are
smaller than 1 mm and range between 0.005 and 14.9 mm with an av-
erage of 0.97–0.308 mm. Typically, 50–70% of bubbles are smaller than
0.1 mm, except for unit X where these small bubbles made up only 16–
20% (Table 1). Vesicle number density values vary from 6.5 × 10² to
4.69 × 10⁴ mm⁻³ (Fig. 9, Table 1).

6.1. Interpretation of observed density and vesicle microtexture features

In comparison with the results of previous density measurements
from OVC, similarities exist between our lapilli densities and the density
measurements of clasts from “S beds” and “Pb beds” from Houghton and
Hackett (1984). Minimum values are 1 g/cm³, but our maximum values
are larger by 0.5 g/cm³ than earlier results. That could be an explanation,
why our mean densities are larger at approx. 0.1 g/cm³. The bomb/block
densities of this study are similar to the bomb density values published
in Houghton and Hackett (1984) from the unwelded parts of the inner
scoria cone. The density differences between the two examined clast
sizes suggest that they originated from different parts of the magma
and may represent different fragmentation and eruption styles. Further-
more, units H and X are more heterogeneous with clasts also having
higher densities. It is clear that the heterogeneous clast assemblage origi-
inated from the same eruption phase; hence any conduit process (e.g.
intensified degassing from a more fluid melt) could have been respon-
sible for the recorded density diversities.

Even the vesicle textures and shapes from the most Strombolian-like
beds (Bb and Bb lithofacies) are completely different to textures origi-
nating from other Strombolian (violent Strombolian) and Hawaiian
eruptions (Lautze and Houghton, 2007; Gurioli et al., 2008; Cimarelli
et al., 2010; Stovall et al., 2011; Moitra et al., 2013; Kereszturi and
Németh, 2016) (Fig. 7). Vesicle textures from the OVC exhibit similar
features to the products of basaltic Plinian eruptions of Etna, 122 BCE
and Tarawera, 1886 (Sable et al., 2006; Moitra et al., 2013) and the
hydromagmatic eruptions of the Ichulbong tuff cone (Murtagh et al.,
2015). The lower part of the VSD slopes often exhibit coalescence (e.g. A1, D3, J4, T3, X2) which may have
occurred at multiple stages. B Cumulative vesicle size distributions (CVSD) are
log(NV /L) as a function of log(L) plots, where NV is the total vesicle number density.
Vesicle density measurements were used based on the objects per cubic mm greater
than L. CVSD trends also suggest multiple stage of nucleation and growth with
coalescence and bubble collapse events in most of the cases (e.g. A1, D3, J4, T3, X2)
(Shea et al., 2010).
Element mapping was executed by EDAX. Comparison with ash fragments (Figs. 11) groundmass glass results display a restricted compositional range in elements (Appendix 1). The main observation is that the blocks/bombs as well as of groundmass glass and phenocrysts of bomb/block fragments have visible vesicles with uneven distributions. We determined ene and olivine crystals and crystal nodules (Fig. 10). Even the densest 0.1–0.8 cm light green to yellowish, mostly isometric or broken pyroxene and olivine crystals and crystal nodules (Fig. 10). Even the densest clasts have visible vesicles with uneven distributions. We determined the major element geochemistry of glass fragments from ash samples as well as of groundmass glass and phenocrysts of bomb/block fragments (Appendix 1). The main observation is that the blocks/bombs groundmass glass results display a restricted compositional range in comparison with ash fragments (Figs. 11–13). Harker diagrams of oxide distributions follow trends of depletion in MgO, FeO, and CaO, enrichment of K2O, and Na2O, and initial enrichment followed by depletion in Al2O3 and Na2O versus SiO2 (Fig. 11), although there is significant scatter in the low- to intermediate silica range. Glass compositions range from basaltic andesite to rhyolite in the TAS classification, i.e. range to much more evolved compositions than the bulk rock (57% SiO2) (Hackett and Houghton, 1989). Strikingly, compositions broadly follow an iron enrichment trend in the AFM diagram (Fig. 12), atypical for continental arc volcanic products (e.g. Ruapehu compositions, Price et al., 2012) and more characteristic for magmas related to decompression induced melting.

We examined compositional variations through time by plotting chemical data from consecutive units (Fig. 13). The ash geochemistry exhibits a great compositional variation within all the examined oxides, and its average values show significant fluctuations through time (SiO2: 62–67 wt%; K2O: 1.2–2.7 wt%; MgO: 5.1–3.1 wt%). In contrast, the averages of groundmass glass compositions measured from blocks/bombs display comparatively minor variations (Figs. 11 and 13).

We calculated crystallization temperatures and depth by thermobarometry utilizing Putirka’s (2008) two-pyroxene calibrations. Our results yield lower temperatures (1034–1053 °C) and pressures (4.8–5.4 kbar) than previous estimates based on cpx–liquid thermobarometry (Table 2) (Deering et al., 2011). Using 2.7 g/cm3 for crustal density (Deering et al., 2011), the two-pyroxene method indicates 16–18 km crystallization depth (Table 2).

7. Petrology, geochemistry and thermobarometry

The fresh block/bomb fragments of the Ohakune andesite are dark grey in colour with inequigranular/hyalopilitic textures, containing 0.1–0.8 cm light green to yellowish, mostly isometric or broken pyroxene and olivine crystals and crystal nodules (Fig. 10). Even the densest clasts have visible vesicles with uneven distributions. We determined the major element geochemistry of glass fragments from ash samples as well as of groundmass glass and phenocrysts of bomb/block fragments (Appendix 1). The main observation is that the blocks/bombs groundmass glass results display a restricted compositional range in comparison with ash fragments (Figs. 11–13). Harker diagrams of oxide distributions follow trends of depletion in MgO, FeO, and CaO, enrichment of K2O, and Na2O, and initial enrichment followed by depletion in Al2O3 and Na2O versus SiO2 (Fig. 11), although there is significant scatter in the low- to intermediate silica range. Glass compositions range from basaltic andesite to rhyolite in the TAS classification, i.e. range to much more evolved compositions than the bulk rock (57% SiO2) (Hackett and Houghton, 1989). Strikingly, compositions broadly follow an iron enrichment trend in the AFM diagram (Fig. 12), atypical for continental arc volcanic products (e.g. Ruapehu compositions, Price et al., 2012) and more characteristic for magmas related to decompression induced melting.

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7.1. Evaluation of the results of petrology, geochemistry and thermobarometry

The observed iron enrichment trend suggests that magmatism at the OVC is related to decompression-melting with little contribution from the subducting slab. This is consistent with the location of the OVC at the southernmost end of the TVZ at the tip of the southwards propagating extension zone. Further, the relatively minor groundmass glass compositions measured from blocks/bombs suggest that the compositional characteristics of the eruptive products are not attributable to multiple magma batches with different chemical compositions. Instead, variations in ash compositions may reflect changes in magma ascent.
rate accompanied by variations in the degrees of cooling, degassing, decompression rates and resulting microlite crystallization (Cashman and Blundy, 2000). Turning to the phenocryst phases, the OVC andesitic magma may represent a mixture of crystals and melts as a result of polybaric fractional crystallization and magma mixing, similar to other erupted andesitic to dacitic rocks from Tongariro Volcanic Centre (Price et al., 2012). For example, the lack of plagioclase phenocrysts in these intermediate composition eruption products suggest that the nodules and phenocrysts of opx, cpx and olivine may represent antecrystic material picked up from a crystal mush of an earlier stalled and cooled magma (e.g. Zellmer et al., 2014). One possible petrogenetic scenario is that this crystal mush is remobilized by intrusion of andesitic to dacitic melt, which ultimately carries the earlier formed crystals of olivine, opx, and cpx to the surface. Corona textures of the olivines may have formed during mush remobilization at depth. The final crystallization event occurred at shallow (tens to hundreds of meters) depth, forming the microlitic groundmass during the relatively fast magma ascent at the onset of eruption. The calculated crystallization depths are slightly shallower than the previous calculations by Deering et al. (2011) (Table 2) yet both results are within the acceptable depth range of a crystal mush zone or magma source.

8. Eruptive volume calculations

The eruptive volumes are essential for understanding the evolution, eruptive styles and related hazards of small-volume volcanoes (Guilbaud et al., 2012; Le Corvec et al., 2013; Kereszturi et al., 2014; Bebbington, 2015; Németh and Kereszturi, 2015). Monogenetic
volcanoes usually have a wide variety of eruptive styles and products as demonstrated by the volcanic sequence of the OVC. To account for this complexity, we employed the eruptive volume scheme used in the Auckland Volcanic Field, New Zealand (Kereszturi et al., 2013). Due to the overall small proportion of lithics in the exposed sequences, we do not expect diatreme structures beneath the edifice, thus we calculate volumes for proximal ejecta rings and scoria cones as well as distal tephra deposits (Fig. 14). The edifice volume was calculated from a DTM using uniformly 585 m asl plane as the pre-eruptive surface, whereas the distal tephra volume was calculated by applying exponential thinning (Pyle, 1989) with the considerations of thinning trends measured within the quarry and thickness and sedimentological observations at distal outcrops. Using our componentry, density, and vesicularity data (Fig. 6, Table 1) and the earlier density measurement for the central scoria cone (Houghton and Hackett, 1984), we could convert bulk distal tephra and edifice volumes to DRE volumes (Fig. 14, Table 3), using the method developed by Kereszturi et al. (2013).

The obtained $12.36 \times 10^6$ m$^3$ DRE total erupted volume includes $2.99 \times 10^6$ m$^3$ (24.2%) rough DRE distal volume and $9.36 \times 10^6$ m$^3$ (75.8%) edifice DRE volume. It seems that the DRE volume of the phreatomagmatic and magmatic products is broadly equal, but scoriaceous deposits are dominant within the edifices (Fig. 4). Due to the morphological differences between the west ejecta ring (WER) and east ejecta ring (EER) (Fig. 2) we calculate the volume of the WER edifice only with phreatomagmatic successions. Uncertainties exist surrounding the possible volumes of eroded sequences of the eastern part of the volcanic complex. The calculated distal volume should be considered only as a minimum value due to the small number of distal thickness measurement. The accuracy of topographic data and data availability for pre-eruptive surface determination can also contribute to the uncertainties associated with volume calculations.

9. Discussion

9.1. Fragmentation and eruptive styles

Despite the absence of accretionary lapilli and the rare occurrence of blocky glass fragments, other features of pyroclastic fragments (e.g. glass alteration – palagonitization, aggregates of ash, presence of mud-coating and secondary minerals inside the cavities) is indicative of a hydrovolcanic origin of the ash-rich beds (as evidenced by Ab and AbC lithofacies) (Sheridan and Wohletz, 1983; Cioni et al., 1992). Furthermore, sedimentary features such as plastically deformed bomb sags and soft sediment deformations support a water-saturated nature of pyroclastic density currents (Fisher and Schmincke, 1984; Németh et al., 2001). At first glance, the coarser bomb beds (BbA and BbC lithofacies) of the ejecta rings show a magmatic (likely Strombolian) origin through their grain size properties (coarse fragments with better sorting) and plastically deformed shapes. However these beds commonly contain cauliflower-shaped bombs (Fig. 4), which are usually characteristic of hydrovolcanic fragmentation (Kienle et al., 1980; Fisher and

![Fig. 12. AFM diagram shows glass compositions measured on ash fragments (red circles) and groundmass glass of blocks and bombs (blue rectangles) in comparison with whole rock compositions of Ruapehu volcano (grey field) (Price et al., 2012, e-appendix 10).](image1)

![Fig. 13. Compositional variations of SiO2, K2O and MgO of ash (yellow fields) and blocks and bombs through the sequence.](image2)
Schmincke, 1984). Moreover the ash fragments of the BbA beds show identical features with fragments from the ash beds. The examined vesicle microtexture properties of the largest clasts from the Strombolian beds do not exhibit the typical micro-textural characteristics of Strombolian fragments (Fig. 7). The size and shape of vesicles as well as the bubble number density values might reflect an increased degree of cooling and higher viscosities similar to the properties of hydrovolcanic eruptions. It seems that throughout the course of the activity water was abundant, which implies the eruptions were unable to permanently dry out the shallow basement (Lorenz, 1986; Lorenz, 2007). The componentry analyses exhibit small percentages of accidental fragments dominantly from the subsurface laharic deposits, indicating that excavation was minor and the fragmentation likely occurred at very shallow depth. The detected heterogeneities of clast densities in units H and X indicate mingling of degassed dense magma with hotter gas-rich melt within the conduit. This suggests that the ascent rate extensively fluctuated and the stalling degassed upper part was broken through from time to time by fresh magma with varying intensity (Lautze and Houghton, 2007). The higher ascent rates of fresh magma resulted in sustained Strombolian, occasionally Hawaiian eruptions with little influence from water (water/magma ratio around 0.1), while the lower ascent rates dominantly led

<table>
<thead>
<tr>
<th>Two-pyroxenes geothermobarometry, sampled units</th>
<th>Eqn 36 T (°C)</th>
<th>Eqn 37 T (°C)</th>
<th>Eqn 38 P (kbar)</th>
<th>K_D (Fe-Mg)</th>
<th>Estimated depth (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>D</td>
<td>1046</td>
<td>1034</td>
<td>5.4</td>
<td>0.981</td>
<td>18</td>
</tr>
<tr>
<td>D</td>
<td>1041</td>
<td>1052</td>
<td>4.8</td>
<td>0.850</td>
<td>16</td>
</tr>
<tr>
<td>H</td>
<td>1035</td>
<td>1053</td>
<td>5.3</td>
<td>0.970</td>
<td>17.8</td>
</tr>
<tr>
<td>X</td>
<td>1043</td>
<td>1036</td>
<td>4.8</td>
<td>0.847</td>
<td>16</td>
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<table>
<thead>
<tr>
<th>Cpx-liquid geothermobarometry (Deering et al., 2011)</th>
<th>Eqn 33 T (°C)</th>
<th>Eqn 34 T (°C)</th>
<th>Eqn 31 P (kbar)</th>
<th>Eqn 35 K_D (Fe-Mg)</th>
<th>Calculated depth (km)</th>
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<td>Ohakune</td>
<td>1137</td>
<td>1152</td>
<td>6.4</td>
<td>0.281</td>
<td>21.4</td>
</tr>
</tbody>
</table>

Table 2
Results of geothermobarometry calculations utilizing Putirka's (2008) two-pyroxenes calculations on phenocrysts of bombs and blocks. Earlier calculations (Deering et al., 2011) indicate deeper source region for the erupted magma.

**Fig. 14.** Delineation of distinct edifices of OVC. The obtained area and thickness values were applied for bulk volume calculations. Pie charts represent the DRE correction scheme applied for volume estimation from the different parts of the volcanic complex (Kereszturi et al., 2013). As the first step of the DRE corrections, the pie bigger charts imply the proportion of juvenile, non-juvenile and interparticle void space. Secondly, the smaller pie charts show the remaining juvenile content was corrected with the available vesicularities.
to phreatomagmatic eruptions with higher water concentrations. The transitions from phreatomagmatic to Strombolian eruptions and vice versa were rapid.

9.2. Eruption history

Although the initial craters are highly obscured, the geomorphology, calculated volume and stratigraphy is inconclusive for the existence of a kilometre wide tuff ring (Houghton and Hackett, 1984). In contrast, we envisage an elongated composite ejecta ring around an at least 600 m long eruptive fissure aligned to the Ohakune Fault, similar to several examples such as Izu-Oshima, Japan (Sumner, 1998); Hverfjall eruptive fissure, Iceland (Mattsson and Höskuldsson, 2011); Easy Chair, Lunar Crater Volcanic Field, USA (Valentine and Cortés, 2013); and Tolbachik, Russia (Belousov et al., 2015). The eruptive centre was initiated by hydrovolcanic eruptions characterized by a sustained low eruptive column due to the abundance of surface and groundwater. The resulting ejecta ring was 15–20 m high, 300–400 m wide along the fissure and characterized by dominantly debris jets and fallout deposits interbedded by layers related to wet pyroclastic density currents (Figs. 4 and 15). The subsequent eruptions were localized into widened and disjointed vents usually inducing changes in mass flux within other areas of the vent zone (Valentine and Gregg, 2008). The usual fluctuating pressure conditions and ascent rates (Keating et al., 2008) and the result of alternations of magmatic and phreatomagmatic eruptions which may have been facilitated through the increased entry of groundwater into the shallow plumbing system. This, and increasing load on the crater rims could have easily led to the destabilisation of the loose surface deposits and the solidifying dyke margins as well (Gutmann, 2002) and demonstrated by the crater rim subsidence (syn-eruptive faulting) at the quarry (Fig. 4). The OVC sequence exhibits the relocation of the active vent(s) within the fissure zone (Fig. 4) probably due to the collapse of destabilized areas from time to time. The proximity of the source is clearly implied by the thermal oxidation along the fault (Fig. 4) similarly to the welded and agglutinated units of the CSC (Fig. 3). The frequent vent-shifting prevented the long term development of a single edifice. Rather, the products were accumulated at the southern margin of the fissure, forming a multisource ejecta ring structure (EER). This was constructed from water-influenced Strombolian bomb beds (Bb, BbA lithofacies) alternating with dense and wet pyroclastic surge and fallout deposits (Ab lithofacies). The deposited pyroclasts within the fissure zone were

Table 3
Results of DRE volume estimations for the OVC based on bulk edifice volume calculated from DTM of Ohakune and DRE correction schemes presented at Fig. 14.

<table>
<thead>
<tr>
<th></th>
<th>Bulk volume (10⁶ m³)</th>
<th>DRE index</th>
<th>DRE volume (10⁶ m³)</th>
</tr>
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<tbody>
<tr>
<td>Central scoria cone (CSC)</td>
<td>8.07</td>
<td>0.487</td>
<td>3.98</td>
</tr>
<tr>
<td>South scoria cone (SSC)</td>
<td>3.51</td>
<td>0.208</td>
<td>0.73</td>
</tr>
<tr>
<td>North scoria cone (NSC)</td>
<td>3.34</td>
<td>0.208</td>
<td>0.69</td>
</tr>
<tr>
<td>West ejecta ring (WER)</td>
<td>3.99</td>
<td>0.45</td>
<td>1.80</td>
</tr>
<tr>
<td>East ejecta ring (EER) (Total)</td>
<td>5.12</td>
<td></td>
<td>2.16</td>
</tr>
<tr>
<td>EER (magmatic)</td>
<td>2.38</td>
<td>0.39</td>
<td>0.93</td>
</tr>
<tr>
<td>EER (phreatomagmatic)</td>
<td>2.74</td>
<td>0.45</td>
<td>1.23</td>
</tr>
<tr>
<td>Total Edifice</td>
<td>24.03</td>
<td></td>
<td>9.36</td>
</tr>
<tr>
<td>Distal</td>
<td>4.80</td>
<td>0.623</td>
<td>2.99</td>
</tr>
<tr>
<td>Total</td>
<td>28.83</td>
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<td>12.36</td>
</tr>
<tr>
<td>Total magmatic</td>
<td>17.30</td>
<td></td>
<td>6.34</td>
</tr>
<tr>
<td>Total phreatomagmatic</td>
<td>11.53</td>
<td></td>
<td>6.02</td>
</tr>
</tbody>
</table>

Fig. 15. Cartoon showing the evolution of facies architecture and eruptive mechanism during the OVC formation (see text for explanation).
removed by subsequent eruptions leaving a hollow, the smaller pits indicat-
ing the area where the dyke intersected the surface (Fig. 14). Similar
morphological features have been reported from many mono-volcanic vol-
canoes within the Auckland Volcanic Field, such as Crater Hill (Houghton
et al., 1999) and Three Kings (Kereszturi et al., 2014). Because the eastern
part of the volcanic complex was eroded by the Mangawhero River, it is
difficult to determine if there were any eruptions (possibly small
phreatomagmatic blasts) along the eastern extension of the fissure. The
late stage evolution was characterized by localization of stable vents pro-
ducing Strombolian/Hawaiian fountains due to higher ascent rates and
decreased water influence, however the axis of the fissure remained un-
buried indicating the existence of vent-clearing phreatomagmatic blasts
along the other part of the fissure (Fig. 15).

9.3. Volcanic hazard assessment

The eruptive styles associated with small-volume volcanoes are usu-
ally strongly influenced by magma ascent rates (Wilson and Head,
1981; Parfitt, 2004; Valentine and Gregg, 2008) and the availability of
external water (Kokelaar, 1986). Despite the lack of direct observations
of magma ascent rates, there are various methods for its estimation such
as field observations of facies architectures (Houghton et al., 1999),
average size of mantle xenoliths (Sparks et al., 1977) and water–content
and zonation of olivine crystals (Peslier and Luhr, 2006; Jankovics
et al., 2013). The Ohakune sequence exhibits transitions between
phreatomagmatic and Strombolian as well as Strombolian and Ha-
awaiian eruptive styles, the latter implying ascent rates varying
around 0.1 m/s (Parfitt, 2004) in the case of a point source stabilized
vent. The earlier stage fissure or multiple source eruptions could
have been characterized by fluctuating high ascent rates coincident
with significant water influx and increased water/magma ratios.
Near surface ascent rates might have exhibited differences to the
lower feeding systems due to guidance of the fault and syn-
intrusive extension. We assume an 1 m width for the feeder dyke
as an average of observations (Geshi et al., 2010), and note that the
facies architecture (Houghton et al., 1999) and possible effects of
fault guidance suggest a similar or shorter dyke length for the lower
feeding system. Combining, the possible source depth estimated
of 18 km obtained by two-pyroxene geothermobarometry with an
average ascent rate of approx. 0.1 m/s, we argue that the magma may
have reached the surface within 50 h. Considering the range of as-
cent rates for the shallow plumbing system (0.01–1 m/s), and the ex-
tent of the active vent area, it is possible to calculate effusion rates.
This analyses suggests peak effusion rates of 25 m$^3$/s for the initial
hydrovolcanic phase, 8 m$^3$/s for the late stage Hawaiian fountains
and <0.1 m$^3$/s for the lowest discharge periods. Using the volume
calculations and ascent/effusion rates it is possible to estimate the
duration of the activity of the OVC. The duration using an inferred
discharge rate of 1–2 m$^3$/s corresponds to a maximum duration of
2.5 to 5 months for the entire event. Even though this duration seems
to be rather long for this type of volcanism however the lack of erosional
structures throughout the eruptive sequence would suggest a shorter
duration. In addition, the OVC erupted from multiple vents, the timing
and duration of each vent remains unclear due to scarce outcrops to ex-
amine stratigraphic relationships and geochemical characteristics. It is
also unclear whether each vent operated simultaneously or were
there time gaps between such activities. These features of the OVC
have to be studied in more detail in the future.

It is difficult to evaluate if there is any relationship between the
OVC’s feeding system and the main conduits of Ruapehu volcano, be-
cause the deep structure of southern slope and ringplain of Ruapehu
volcano is not known. Magnetotelluric surveys did not propose large
scale accumulation of magma shallower than 14 km beneath Ruapehu
(Ingam et al., 2009). However, the plumbing system might include
the network of small-volume dispersed magma storages in the mid-
and shallow crust (Price et al., 2012). Our proposed 16–18 km depth
for the accumulation of magma erupted at OVC corresponds to the
upper part of mush zones of the heavily intruded lower crust (Deering
et al., 2011). In spite of the analogous petrogenesis of the OVC and
some of Ruapehu magmas (Price et al., 2012) we do not expect the
OVC to have a closer relationship with the plumbing system of main
conduit of Ruapehu than to any of the other three composite volcanoes
of the Tongariro Volcanic Centre (Fig. 12). On the basis of the available
major element oxide data, it appears that the eruptions around
Ohakune represent the first precursors of the southward expansion of
the anedetic arc of the TVZ, and that the OVC conduit is independent
of the main feeder system of Mt. Ruapehu.

10. Conclusion

The OVC provides an opportunity to examine an eruptive sequence
characterized by alternating eruptive styles. The resulting deposits indi-
cate that the rising magma erupted through multiple vents by fire-
fountaining to water-influenced Strombolian explosions alternated with
phreatomagmatic blasts. It appears that after the initial vent opening,
the most significant controlling factors were the changing vent geomet-
ries and variable interaction with external water, while the final phase
was characterized by higher effusion rates, which enabled the isolation
from the source of external water inducing fire-fountaining eruptions
and the generation of short clastogenetic lava flows. This study has re-
vealed that the volcanic architecture of the OVC is much more complex
than previously believed, and that the phreatomagmatism was constant-
ly present during its activity. This study indicates that the active vents can
be unstable and abrupt changes in eruption location (likely influenced by
faults) and eruption styles can occur, significantly increasing the hazard
of such small-volume eruptions. The duration of several months of such
events is also significant in terms of volcanic hazards and considerably
longer than previously considered at OVC. It is believed, there is a very
small likelihood of a future volcanic event in the same location yet for fu-
ture similar eruptions in the southern TVZ (south of Ruapehu) they
would have a similar duration under the same tectonic influences.
The eruptive behaviour of the OVC is also considered typical for future
mono-
 generar eruptions within the other parts of Taupo Volcanic Zone.

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References

Abrams, M.J., Siebe, C., 1994. Cerro Xalapaxco: an unusual tuff cone with multiple explo-
183–199.
Agustin-Flores, J., Németh, K., Cronin, S.J., Lindsay, J.M., Kereszturi, G., Brand, B.D., Smith,
I.E.M., 2014. Phreatomagmatic eruptions through unconsolidated coastal plain se-
quen
Geotherm. Res. 276, 46–63.
Agustin-Flores, J., Németh, K., Cronin, S.J., Lindsay, J.M., Kereszturi, G., 2015. Shallow-seated
explosions in the construction of the Motukorea tuff ring (Auckland, New Zealand): ev-
edence from lithic and sedimentary characteristics. J. Volcanol. Geotherm. Res. 304,
272–288.
Auer, A., Martin, U., Németh, K., 2007. The Fekete-hegyi (Balaton Hills Hungary) "soft-
substrate" and "hard-substrate" maar volcanoes in an aligned volcanic complex – im-
plications for vent geometry, surfac
e stratigraphy and the palaeoenvironmental
Bebington, M.S., 2015. Spatio-volumetric hazard estimation in the Auckland volcanic


