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Chapter 6  
Geochemistry and properties of magmas from OVC and TVC

Figure 6.12. Glass compositions (A) OVC and (B) TVC eruptives plotted on the synthetic Qz-Ab-Or-H_2O ternary system of Tuttle and Bowen (1958). The low-SiO_2 rhyodacites (OVC old MSg) and post-26.5 ka TVC dacites plot below the cotectics indicating P_c >400 MPa.
Figure 6.13. Relationships between T, P, (estimated using Fig. 6.13), and ferromagnesian mineralogy in OVC and TVC eruptives. Average depths were estimated using an overburden density of ~2500 kg/m³.

DISCUSSION

Ferromagnesian mineralogy and magmatic relationships

All deposits from the OVC and TVC can be classified by their ferromagnesian mineralogy (Table 6.7). Several observations can be made on the occurrence of particular phases and the estimated properties of the magma (also see Appendix 6.1).

Biotite

Abundant biotite is found in the uppermost ~50 ka Rotoiti deposits and in six younger OVC post-26.5 ka eruption deposits (see Table 6.1). Biotite is not as common at TVC and is only found in trace abundance in the pre-26.5 ka OSg eruptive deposits (see Table 6.2). Biotite commonly occurs with quartz and hornblende, and occasionally with cummingtonite at OVC, and only crystallises in evolved magmas (Fig. 6.6a and b) with high SiO₂ (>77 wt %) and K₂O (>3.8 wt %) melts. The crystal chemistry in these biotite-bearing deposits also reflects the more evolved melt composition, for example, K₂O-rich plagioclase (>0.5 wt %) and FeO-rich hornblende (>15.8 wt % in biotite-bearing OVC deposits and >19.5 wt % in TVC deposits). In addition, FeO-rich hornblende accompanies FeO-rich biotite. The composition of biotite does not show a direct relationship to glass composition, T, or fO₂ (Shane et al., 2003). However, the presence of biotite does. T and P, estimates indicate that biotite crystallises in cool (<780°C), shallow (100-180 MPa) magma storage chambers (Fig. 6.13). The fO₂ of biotite-bearing deposits is -0.2 to -0.4 NNO for OVC deposits and -0.65 to 0.2 NNO for TVC deposits. These differences in fO₂ between the centres could reflect different sources for the OVC and TVC magmas (cf. Carmichael, 1991; and below). It is also noted that all OVC biotite-bearing eruptions tapped multiple magma batches.
Cummingtonite

Cummingtonite only occurs in some OVC deposits (Table 6.1). It commonly occurs with hornblende, orthopyroxene, and/or biotite (in minor abundance). The composition and magmatic properties of cummingtonite-bearing deposits are somewhat distinctive. Their glass composition is typically >77.25 wt % SiO₂, and 3.3-4.1 wt % K₂O (Fig. 6.6b). Cummingtonite-bearing deposits display low T (735-785°C; Fig. 6.9b) and P (110-160 MPa; Fig. 6.13). These parameters partly overlap with those of biotite-bearing deposits. Thus, cummingtonite and biotite commonly crystallise under similar magmatic conditions. However, cummingtonite-bearing magmas are usually more oxidised, forming a different T-f02 trend (Fig. 6.9b), and have a narrower range in fO₂ (+0.1 to +1.1 NNO; Fig. 6.10a).

The oxidation state of the magma influences the composition of cummingtonite (Fig. 6.14), where XFe (= %Fe/(Fe + Mg + Mn + Ca) of cummingtonite) decreases with oxidation. This relationship was first noted by Evans and Ghiorso (1995). Degree of water saturation in cummingtonite-bearing magmas can be estimated using the thermodynamic stability model of Evans and Ghiorso (1995). The activity of water (awater) ranges from 0.6 -1, indicating cummingtonite crystallises in a range of water-saturation conditions (Fig. 6.14). However, OVC cummingtonite-bearing magmas crystallise at shallower depths (110-160 MPa) than the Evans and Ghiorso (1995) model used here (200 MPa). Therefore, awater was probably higher and the cummingtonite-bearing magmas are closer to water saturation. Nicholls et al. (1992) indicated that cummingtonite only crystallises in magmas that are close to water saturated. The 8.1 ka Mamaku, 17.7 ka Rerewhakaaitu, and 13.8 ka Waiohau RSg OVC eruptives do not contain significant cummingtonite but all have a similar composition, T, P, and fO₂ to the OVC cummingtonite-bearing magmas. Thus, it could be deduced that these non cummingtonite-bearing magmas were more undersaturated.

Figure 6.14. T, fO₂, and XFe in OVC cummingtonite-bearing deposits, with activity of water contours from Evans and Ghiorso (1995).
Clinopyroxene occurs in the less-evolved rhyolitic deposits from OVC and TVC (Table 6.1 and 6.2). These magmas have less evolved glass compositions (SiO$_2$ <76.5 wt %, K$_2$O <3 wt %, and CaO >1.35 wt %; Fig. 6.6b), and with CaO-rich plagioclase (>3 wt %) and FeO-poor orthopyroxenes (~18-22.5 wt %). Clinopyroxene in OVC and TVC deposits crystallised in high T magmas (>825°C; Fig. 6.13), at great depths, mostly >300 MPa (>12 km) at both OVC and TVC (Fig. 6.13). These deposits are also the most oxidised, at their respective centres, with fO$_2$ ranges of +0.7 to +1.45 NNO at OVC and -0.5 to +0.3 NNO at TVC (Fig. 6.9b).

Orthopyroxene and hornblende
Most OVC and TVC silicic eruptives in the last 50 kyr contain orthopyroxene. It is commonly found with other ferromagnesian minerals such as biotite, cummingtonite, or clinopyroxene. Properties of orthopyroxene-dominant magmas are not as distinctive as those that crystallise biotite, cummingtonite, or clinopyroxene. All post-50 ka OVC silicic deposits and pre-26.5 ka TVC deposits examined contain hornblende. It only occurs in trace abundance in a couple of post-26.5 ka TVC deposits (Table 6.2). Hornblende is always accompanied by orthopyroxene, and can occur with any other ferromagnesian minerals in OVC and TVC deposits. Hornblende-dominant deposits display slightly lower T and P$_c$ than solely orthopyroxene-bearing deposits.

Evolution of OVC and TVC
Temporal changes in eruption style and magma type
After the OVC caldera-forming Rotoiti episode at ~50 ka, the tempo of rhyolitic activity increased at both OVC and TVC. At OVC there were at least 14 medium-sized (0.1-6 km$^3$) plinian eruption episodes (MSg) between 40 and 27 kyr. The first nine eruption episodes (Old MSg) were generally of smaller volume, while pyroclastic flows accompanied the following Young MSg eruption episodes. Between ~50 ka and 26.5 ka (deposits bracketed by the Rotoiti and Oruanui time planes), OVC erupted a greater volume of magma (~22 km$^3$) than TVC (~9 km$^3$) (Fig. 6.3). TVC was infrequently active, producing six small-medium (<3 km$^3$), plinian eruption episodes (Okaia Subgroup). The second major change in eruptive behaviour at both eruptive centres followed the TVC caldera-forming Oruanui episode at 26.5 ka. Since then, TVC has erupted more frequently than OVC, with ~28 small-medium (0.01–~13.4 km$^3$) plinian eruption episodes, totalling ~22 km$^3$ of magma (Fig. 6.3). In contrast, at OVC there were nine (1-17.5 km$^3$) rhyolitic episodes, erupting ~82 km$^3$ of magma. Many OVC episodes occurred from chains of vents, up to 20 km long, at Tarawera and Haroharo, which run parallel to the dominant fabric of the TVZ (Fig. 6.2a). These episodes commonly produce plinian falls, proximal pyroclastic density currents, and lava domes and flows (Nairn, 2002). The OVC rhyolitic episodes are roughly evenly spaced temporally and erupted volumes appear to be approximately the same for each episode (average ~9 km$^3$; Fig. 6.3).
Changes in the ferromagnesian mineralogy of the eruptive units at both centres broadly correspond to the changes in eruption style (Table 6.1 and 6.2). At OVC, Old MSg deposits are clinopyroxene-bearing, Young MSg deposits are orthopyroxene- and hornblende-bearing, and most OVC rhyolites of the last 26.5 kyr are cummingtonite- and/or biotite-bearing. Pre-26.5 ka TVC eruptives are commonly biotite-bearing, the 26.5 ka Oruanui eruptives are hornblende- and orthopyroxene-bearing, and post-26.5 ka deposits are orthopyroxene-dominant. Eruption styles and ferromagnesian mineralogy differ over time; therefore, several general compositional and magmatic temporal trends are evident (Fig. 6.15). After the ~50 ka Rotoiti episode, OVC erupted less evolved, high T and P$_c$, clinopyroxene-bearing magmas and then changed at ~34 ka to more evolved, lower T and P$_c$ magmas containing greater proportions of hydrous phases (such as hornblende). The more recent OVC rhyolite magmas have begun to crystallise (and reside) in cooler (<800°C), shallower (480 MPa) storage chambers. In contrast, compositional and magmatic changes at TVC do not vary so systematically with time (Fig. 6.15). However, in general the more recent orthopyroxene-dominant rhyolitic magmas have erupted at higher T (>800°C) and crystallised at greater P$_c$ (>200 MPa) compared to the pre-26.5 ka biotite-bearing magmas (<800°C; <150 MPa) (Fig. 6.13 and 6.15). The intervening dacites (20.5-17 ka) are an exception, and were derived from high T (>900°C) and P$_c$ magmas (>400 MPa) following the caldera-forming episode (Fig. 6.13 and 6.15).

The least evolved magmas (dacites-rhyodacites) at both OVC and TVC erupted after the caldera-forming episodes at their centres. In addition, after the caldera forming episodes at both OVC and TVC, magma rose from storage chambers at greater depths. Similar behaviour is seen in the record of activity at Yellowstone caldera in Wyoming, the least evolved magmas erupted following the caldera-forming Lava Creek event (Hildreth et al., 1984). They suggest that the less evolved magma was from deeper levels within a magma chamber that was incompletely tapped during the caldera-forming event. It is unlikely that the TVZ dacites/rhyodacites are derived from residual mafic magmas that directly underlay the large rhyolitic bodies whose eruptions caused caldera collapse, as the dacites resided at much greater depths than the rhyolite magmas (difference >200 MPa). However, both the Rotoiti and Oruanui caldera-forming magmas were largely chemically and physically homogeneous, suggesting vigorous convection driven by mafic underplating. Rare mafic clasts and ash are found in the Oruanui deposits (Sutton et al., 1995), and basalt was erupted immediately before the Rotoiti episode (Pullar and Nairn, 1972). Thus, both caldera-forming episodes may have been primed by basaltic underplating.

There is no simple relationship between SiO$_2$ contents in magmas and eruptive volumes. However, the caldera-forming eruption episodes are high-SiO$_2$ rhyolites, while dacitic eruption episodes are small volume (<1 km$^3$) (Table 6.1 and 6.2). There is strong correlation between T (from Fe-Ti oxides and pyroxenes) and P$_c$ (derived from crystal-melt equilibrium) for all eruption episodes at both centres (Fig. 6.13). This indicates that the melt (glass) was in equilibrium with the Fe-Ti oxides, orthopyroxene,
Figure 6.15. Temporal trends of glass composition, average T, average fO$_2$ (NNO), and average P$_c$ at (A, C, E, G) OVC; and (B, D, F, H) TVC. Solid line is the time of major caldera collapse.
clinopyroxene, and plagioclase. Fe-Ti oxides are very sensitive to magmatic changes and their composition adjusts (with respect to Ti distribution) to be in equilibrium with the melt within a few days to months of temperature change (e.g., Devine et al., 2003). Therefore, Fe-Ti oxides reflect the conditions within the magma storage chamber prior to eruption rather than inheritance from source or earlier thermal events. Since the Fe-Ti oxides are in equilibrium with the rest of the crystal assemblage, most crystallisation must have occurred in shallow storage chambers leading up to eruption. Plagioclase crystals in many OVC and TVC deposits display optical zoning patterns. However, reconnaissance analyses show no simple relationship to composition, with many weak (normal and reverse) chemically zoned plagioclases and homogeneous crystals, which are also optically zoned (see Table 6.1 and 6.2 for ranges), found in the same eruption deposits. Plagioclase does not provide any evidence for major changes in magmatic conditions in the storage chambers.

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**Oxidation state of the magmas**

Oxygen fugacity of OVC and TVC eruptives does not show any overall relationship to ferromagnesian mineralogy, crystal geochemistry, or pressure. However, fO$_2$ varies between the volcanic centres, and the periods of activity (defined by changes in eruptive style and ferromagnesian mineralogy; Table 6.8). All magmas from OVC are more oxidised, displaying higher fO$_2$ at a given T, than those from TVC (Fig. 6.9a). The difference in fO$_2$ between the centres is of long duration (>50 kyr), and is independent of magma temperature. Each period of eruption episodes from both centres has a distinct T-fO$_2$ trend. Changes in fO$_2$ at OVC followed the ~50 ka Rotoiti, c. 39 ka Ngamotu, 36.1 ka Hauparu (last Old MSg episode), and c. 31.5 ka Unit L (last Young MSg episode) eruption episodes. These changes in fO$_2$
also relate to apparent changes in vent location, thus indicating a spatial relationship. The ~50 ka Rotoiti episode is thought to have erupted from the northern part of the Okataina (Haroharo) caldera. Vents of the Old MSg eruption episodes are likely located within central caldera, and those of the Young MSg may be located in Puhipuhi Basin (Fig. 6.2a). A change in vent location also accompanied the change in fO₂ during the Old MSg eruption episode interval; vents of the first two episodes (c. 40 ka Unit A and c. 39 ka Ngamotu) were probably located in southern part of the caldera, and others were thought to be located in the central-northern part (Fig. 6.2a) (Jurado-Chichay and Walker, 2000). Post-26.5 ka RSg episodes were from the Tarawera and Haroharo linear vent zones that cut across the central OVC caldera (Fig. 6.2a). However, deposits from these two vent zones cannot be distinguished by T-fO₂ trends. It is possible that these post-26.5 ka RSg magmas, which erupted along the two linear vent zones, came from a similar uniform source. However, the Haroharo and Tarawera magmas resided in geographically separate locations under their respective vent zones and experienced different physical conditions prior to eruption.

At TVC, changes in fO₂ occurred after the 26.5 ka Oruanui caldera-forming episode, and again after the c. 17-20.5 ka dacitic eruptions (Table 6.8; Fig. 6.9a). Similar to OVC, the changes in fO₂ at TVC appear to accompany changes in vent location. However, vents of TVC eruption episodes are poorly constrained. Pre-26.5 ka and the Oruanui eruption episodes are thought to be from vents within the central-northern present Lake Taupo, the dacitic episodes from north, and the post-26.5 ka rhyolites from a northeast-southwest trending corridor in the east side of modern Lake Taupo (Fig. 6.2b).

The spatial controls on fO₂ and the lack of relationship with ferromagnesian mineralogy, geochemistry, or pressure, suggest a strong link to source. This is consistent with Carmichael (1991) who suggested that the redox state of magmas is inherited from the source region. If this is correct, then the source materials/magmas for OVC and TVC appear to be fundamentally different, with minor spatial differences within each of the centres presumably reflecting a heterogeneous protolith.

Okareka basin (OB) is situated on the western margin of OVC and overlaps with eastern margin of the Kapenga caldera (Fig. 6.2a). Vents in the basin lie on the same lineament as the post-26.5 ka Haroharo vents of the central Okataina (Haroharo) caldera, and have been active during the 25 ka Te Rere and 15.8 ka Rotorua eruption episodes. However, deposits from the OB are mineralogically and compositionally different to those of the Haroharo complex (e.g., Figs. 6.6c), including the Te Rere episode plinian pumice (T1) erupted from a Haroharo vent immediately before the Te Rere eruption at OB. Both the Rotorua and Te Rere episodes erupted two distinct types of magma from vents in the OB: (1) K₂O-rich, biotite-bearing; and (2) less evolved, clinopyroxene-bearing deposits (see Smith et al., 2004) (Fig. 6.6c). In addition, OB deposits differ in fO₂ relative to other RSg OVC eruptives. These features are shared by EFP and the Kapenga Rhyolites from Kapenga caldera (Fig. 6.9c). This
suggests that the deeper source for some OB eruptives is the same for that for the north eastern Kapenga caldera eruptives.

Although there are relationships between vent location, and (a) $fO_2$ and (b) magma composition, there is no simple relationship between $fO_2$ and magma composition. If $fO_2$ (oxidation state) relates to the magma source, as indicated by relationship to vent, it follows that either there were differing degrees of partial melting at source, or that some processes altered magma composition after melt formation. Observed phenocrysts within deposits crystallised within high-level storage chambers after the melt was emplaced (see Origin of Magmas below). Separate fractional crystallisation models of the Old MSg, Young MSg, Tarawera, and Haroharo eruptives (Fig. 6.5c) suggest that fractional crystallisation alone cannot generate the range in melt compositions observed. Similarly, the models of Sutton et al. (2000) suggest that the compositional categories within the post-26.5 ka TVC rhyolites cannot be produced by fractional crystallisation. Instead, magma composition may be strongly influenced by the degree of partial melting of the source region, or by differences in source materials, prior to migration of the melt to storage chambers. Oxidation state ($fO_2$) would be unaffected by partial melting.

Volcanism and tectonics
Changes in style of activity, composition, and vent location at OVC and TVC were contemporaneous and occurred after the caldera-forming episodes, the ~50 ka Rotoiti and the 26.5 ka Oruanui. The tempo of activity appeared to increase post-50 ka. The lack of pyroclastic deposits preserved at OVC and TVC prior to ~50 ka, suggests that both centres were not particularly active. This is further supported by the lack of pre-50 ka rhyolitic tephra in medial-distal settings in the North Island (Shane and Hoverd, 2002; Shane et al., 2002). Post-50 ka tephra are commonly recognised in these more distal settings. This implies either: (1) external, regional controls on volcanism; or (2) the voluminous (>100 km$^3$) caldera-forming eruptions changed the regional stress regime. The variation in volume of rhyolitic magmas generated may be a function of variability in subduction geometry, as this can substantially influence on the generation of mafic melts and thus the crustal temperature field (Laube and Springer, 1998). Nairn et al. (1998) suggested that the concurrent onset of frequent eruptive behaviour at ~12 ka of the andesitic Tongariro centre and TVC (the post-Oruanui rhyolitic volcanism) was likely to have been controlled by tectonic activity; with regional extension inducing concurrent rifting and magmatic fracturing. By analogy, it is possible there were changes in tectonic regime over much of the central TVZ at both ~50 ka and 26.5 ka causing the contemporaneous changes in eruptive behaviour, and vent location at both OVC and TVC. Increased spreading could have allowed more intrusion of basalt magma from the mantle, thus inducing increased crustal melting. Alternatively, if rhyolite magma production at TVC and OVC is not influenced by tectonics, its eruptions may have been. Increased extension could provide more conduits for eruption.
Magma chamber dynamics
Pre-eruptive gradients and zoning
Whole-rock, glass, and mineral compositions, crystal contents, and physical parameters (T and fO2) of stratigraphically-controlled samples through many eruption deposits has found little evidence for pre-eruptive gradients within most OVC and TVC silicic magmas erupted in the last 50 kyr. The exceptions are some old MSg eruption episodes, which display heterogeneous glass and mineral compositions (see Table 6.1 and 6.4). Variation in many other OVC eruption episodes is associated with the tapping of multiple magma batches (see below). However, the individual magma batches lack pre-eruptive gradients (Table 6.1). Even the voluminous, caldera-forming, ~50 ka Rotoiti and 26.5 ka Oruanui, eruption episodes are mostly homogeneous (>95 %) and lack T and fO2 gradients (Sutton et al., 1995; Shane, 1998; Wilson, 2001; Shane et al., 2005) (Table 6.1 and 6.2). In addition, investigations of some TVC eruption episodes by Dunbar et al. (1989a; 1989b) indicate a lack volatile gradients. The absence of gradients is independent of eruption volume (1-550 km3), suggesting that most episode eruptives are from well mixed, and thus presumably rapidly convecting magmatic systems.

Magma batches and magma mingling
The discrete compositional and temperature gaps within many OVC deposits indicate that these episodes tapped multiple, compositionally distinct, separate rhyolitic magmas. These deposits are not continuously zoned (chemically or thermally) as are the large rhyolitic magmas (>500-1000 km3) from continental settings of Long Valley or Yellowstone (e.g. Hildreth, 1981). Instead, the OVC deposits commonly contain mingled rhyolitic pumice clasts with disequilibrium mineral assemblages, indicating short-lived interaction of separate magmas prior to eruption (see Nairn et al., 2004; Smith et al., 2004; Shane et al., 2005). Several different interactions between rhyolite magmas are noted for OVC: (1) During the 0.7 ka Kaharoa eruption episode, a compositionally stratified rhyolitic sill was evacuated after being primed by a basaltic intrusion. The eruption episode sequentially tapped two different rhyolite magmas, plus mingled clasts, along an 8 km vent zone (Nairn et al., 2004). (2) In the 15.8 ka Rotorua eruption episode, a hotter, volatile-rich, rhyolite magma (T1) intruded a cooler, crystal-rich stagnating magma (T2), so that both were erupted. The hotter magma heated and reactivated the crystal-rich magma into eruption. Due to contrasting physical properties (e.g., temperature, viscosity, and density) there was minimal chemical and mechanical mixing (Smith et al., 2004), with only a few mingled clasts found. (3) During the large ~50 ka Rotoiti episode, 100 km3 of homogeneous magma incorporated three separate, smaller magma bodies (which were at various stages of crystallisation). The magmas did not completely mix, and the ejecta contain quenched granitoid clasts and mingled pumices (Burt et al., 1998; Shane et al., 2005).

In the three OVC examples above, the volumetrically subordinate mingled ejecta most likely formed in the conduit during the eruption when the magmas were brought into contact. The individual magmas lacked chemical and physical gradients as indicated by glass and mineral compositions. The
phenomenon of OVC eruption episodes tapping multiple rhyolitic magma bodies appears to be common, as ejecta of other depositional units (e.g., 36.1 ka Hauparu, 21.9 ka Okareka, and 17.7 ka Rerewhakaaitu; Table 6.1) also contain compositionally and mineralogically distinct pumices, a feature we have not observed in TVC deposits (Table 6.2). It is not clear why OVC eruptions tapped multiple rhyolite magma bodies, while TVC eruptions did not. Perhaps the subsurface structure, which affects magma ascent and residence time, beneath the two caldera centres is different.

Post-26.5 ka OVC eruption episodes were mostly larger, crystal-rich (up to 20%), lower T and P_c, plinian episodes accompanied by lava domes and flows of roughly equal volume to the pyroclastics, and containing hydrous crystal phases. In contrast, individual TVC post-26.5 ka episodes were frequent, small, crystal poor (<5%), high T and P_c, plinian episodes with little evidence of associated lava extrusion. TVC deposits are generally chemically and thermally homogeneous, with episodes tapping single magma batches. The crystal-poor character of TVC deposits suggests rapid ascent of magma from depth with little time in shallow storage for crystallisation to occur. The only deposits that display significant chemical heterogeneity are those of the 3.6 ka Unit S (see below) and 26.5 ka Oruanui eruption episodes. Similar to the Rotoiti episode (Shane et al., 2005), the subtle chemical variation within the Oruanui deposits could be associated with the digestion of xenoliths derived from stagnated magmas and country rock as a large volume of homogeneous magma (>500 km³) moved towards the surface.

The lack of physical and chemical gradients in most OVC and TVC magmas requires effective mixing to maintain homogeneity. Thus, storage chambers are likely to have been sill-like, with a low-aspect ratio allowing rapid heat loss at the roof, promoting convective mixing (De Silva and Wolff, 1995). The studies of Blake et al. (1992), Nairn et al. (2004), Smith et al. (2004), and Shane et al. (2005) have presented evidence of sill-like magma storage chambers for OVC and TVC rhyolites.

Triggering

The collision and mixing of a hot mafic magma with a cooler more silicic magma has been widely recognised as a potential trigger mechanism for explosive eruptions, induced by heat, mass, and volatile transfer (e.g., Sparks et al., 1977; Pallister et al., 1992). At OVC, basalt intrusion apparently triggered the eruption of one or more rhyolite magmas during the ~50 ka Rotoiti, 21.5 ka Okareka, and 0.7 ka Kaharoa eruption episodes (Nairn, 1992; Leonard et al., 2002; Shane et al., 2005). At TVC, the 3.6 ka Unit S episode is the only episode obviously primed by mafic magma (Blake et al., 1992).

At OVC, many post-26.5 ka eruption episodes each tapped multiple rhyolitic magma batches (Table 6.1). The 15.8 ka Rotorua eruption episode is an example of rhyolitic intrusion triggering the eruption of another rhyolitic magma (see Magma mixing and magma mingling above). Mafic or silicic magma intrusion may not have triggered all of the eruptions that tapped multiple magmas. For example, during
the 25 ka Te Rere eruption episode (see Nairn, 2002), vents in the central Okataina (Haroharo) caldera erupted (as plinian pyroclastics and lavas) magma (T1) that was compositionally different to that erupted from the OB vents (T2 and T3) immediately after the plinian eruptions (Fig. 6.6c). The Te Rere eruptives do not display mingling (unlike other OVC eruption episodes) and the vents were widely separated (~16 km) suggesting the compositionally distinct magmas were never in contact. Thus, the sequential triggering of two, spatially separate, chemically different magmas from the same linear vent zone (situated above a basement fracture zone that is a NE-continuation of the axis of the Ngakuru Graben) (Nairn, 2002) is most likely to have been tectonic. No basaltic clasts have been found in the Te Rere episode eruptives, thus it is unlikely the magmas were primed by a basaltic intrusion along the structure.

Xenocrysts and earlier crystallisation episodes
The presence of granitoid xenoliths within eruptive deposits demonstrates that ascending magmas encountered remnants of crystallised magma bodies at depth (Brown et al., 1998; Burt et al., 1998; Nairn et al., 2004). Magmas with contrasting physical properties may not completely mix and thus produce compositionally bimodal ejecta as described above. In other cases, colliding rhyolite magma batches may have had similar physical properties, so that complete mixing and hybridisation occurred prior to eruption, producing chemically and petrographically homogeneous ejecta, i.e. the 13.8 ka Waiohau eruption episode from OVC (Table 6.1). Evidence for mixing may only be found in refractory phases such as zircon, which can become incorporated into a new magma. The zircons are not easily isotopically reset. Often new zircon crystallises around the rim of another, older xenocryst (e.g., Brown and Smith, 2004), not resetting it and thus preserving the magmatic history. Xenocrysts are commonly entrained into the magma from surrounding country rock either during ascent or during eruption and can represent a volumetrically significant population (see Gansecki et al., 1996; Gansecki et al., 1998; Gardner et al., 2002). Thus, wide age ranges from zircons (e.g. Charlier et al., 2003) and other phases (e.g., allanite, seeVazquez and Reid, 2004) in some silicic deposits do not necessarily reflect long crustal residence times for the bulk of the erupted magma. Instead, they could be derived from plutons or country rock engulfed by the ascending magma. Wide ranges in U-series zircon ages have been observed in numerous OVC and TVC eruptives (Charlier et al., 2003; Charlier et al., 2005). Many TVC eruption episodes have bimodal age spectra (Charlier et al., 2005) and suggested that the old peaks represented the zircons derived from older silicic mush (see below), while the younger peaks represented the zircons that crystallised in the chamber just prior to eruption. The concentration of zircons within TVC deposits is miniscule (usually <0.0001%) (Charlier et al., 2005), this amount is easily inherited and possibly is not comagmatic.

Origin of the magmas
The source materials and formation processes of TVZ rhyolitic magmas are uncertain. Silicic melts can be formed from crustal melting by heat transfer, from the intrusion of mantle-derived mafic magmas into
a semi-ductile crust, or by fractional crystallisation of mafic magma together with crustal assimilation. In the TVZ, there is evidence to support all of these processes to certain degrees (Graham et al., 1995). The wide range of zircon ages in many OVC and TVC deposits (Charlier et al., 2003) indicates that melts can entrain material of different ages (i.e., country rock, stagnated magmas, or plutons) and thus magma formation may be multi-cycle, including the remelting of igneous protoliths. Mafic magma is intruded and erupted in the TVZ. Fractionation of the basalt could produce residual silicic melts. However, it would require crystallisation of a vast volume of mafic melt to produce >750 km$^3$ of rhyolite and there is no geophysical evidence for such mafic accumulations. Graham et al. (1995) show inconsistencies in trace and REE, for both fractional crystallisation of basalt and crustal melting of Mesozoic meta-greywackes. However, ages of some inherited zircons in TVZ eruptives are consistent with the magmas originating from Mesozoic greywacke basement rocks (e.g., Brown and Smith, 2004; Charlier et al., 2005). It is also possible, and highly favoured by Graham et al. (1995), that the TVZ magmas are generated from the melting of granulites; however, there is no independent evidence to suggest that granulites underlie the TVZ. Conrad et al. (1988) and Nicholls et al. (1992) indicate that the rhyolites are not primary melts and have undergone extensive fractionation.

Whole-rock and glass compositions of OVC and TVC magmas plot along the same geochemical trend (Fig. 6.4). However, the whole-rock compositions are less evolved, reflecting their phenocryst content. Whole-rock compositions are that of the melt (glass) and phenocrysts combined, and thus could be seen as the bulk composition of a ‘proto-melt’ prior to phenocryst crystallisation (see Blake et al., 2002; Hawkesworth et al., 2004). Agreement between glass-quartz-plagioclase equilibria, Fe-Ti oxides, and orthopyroxene-clinopyroxene equilibrium in most of the OVC and TVC rhyolites examined here, indicate nearly all phenocrysts form under similar conditions within the storage chamber shortly before eruption. Since there is a positive correlation between glass composition and temperature (Fig. 6.10b), a temperature for the ‘proto-melt’ can be estimated by extrapolating the glass-temperature trend to the less evolved whole-rock compositions (Blake et al., 2002). These ‘whole-rock’ model temperatures ($T_w$) represent that of the initial melt before it rose to its storage chamber where the phenocrysts formed (Table 6.1 and 6.2). The relationship between glass composition and temperature varies according to ferromagnesian mineral assemblage or geochemical affinity. Thus, the compositional/mineralogical eruptive groups were modelled separately. The $T_w$ estimates are ~4 to 181°C (mostly ~60°C) higher than the Fe-Ti oxide equilibrium temperatures ($T$) (Table 6.1 and 6.2), which suggests that the rhyolite magmas were derived from hotter source regions. Variation in $fO_2$ between both volcanic centres and within the centres suggests that the source regions are heterogeneous. However, melt (glass) compositions vary independently of $fO_2$ suggesting the melt composition is not strongly inherited from source. Instead, it could be the result of varying degrees of partial melting of broadly similar sources.

Recent work by Bachmann and Bergantz (2004) couples the evolution of granitoids and volcanic rocks, suggesting that all crystal-poor rhyolites are expelled from crystal-rich mush zones. Hildreth (2004)
proposes a similar mechanism for the generation of voluminous rhyolite magmas at Long Valley caldera. Both the Bachmann and Bergantz (2004) and Hildreth (2004) models suggest the rhyolitic melts are derived from intermediate magmas which have experienced large amount of crystallisation (~45-50%) of mainly anhydrous feldspars and quartz. The interstitial rhyolite melt in these crystal-rich magmas (mush) escapes upwards due to compaction (associated with tectonic loading, earthquakes, and eruptive disturbances) and settling (hindered and micro-settling) of the mush (Bachmann and Bergantz, 2004; Hildreth, 2004). These workers suggest that the mush zones have a long-lived melting, mixing, crystallisation, and fractionation history, and are constantly rejuvenated by the influx of hotter, mafic. These models of rhyolite magma generation explain: (a) the stronger crystal fractionation signature in rhyolite eruption deposits compared to the plutonic counterparts; (b) multiple eruptions of compositionally and mineralogically homogeneous silicic magmas from a volcanic centre over long periods of time, e.g., the Coso Volcanic Field which tapped the same magma composition across an area >80 km² over >100 kyrs (Manley and Bacon, 2000), and (c) extensive zones of low P-wave velocity in the lower crust below silicic volcanic centres, e.g., at Yellowstone (Miller and Smith, 1999).

Charlier et al. (2005) developed a model for the generation and storage of rhyolitic melts at TVZ using zircon model-age spectra, zircon-dissolution modelling, and data from previous TVZ studies. The model is similar to that of Hildreth (2004) and suggests that there are three main entities to magmatic systems in TVZ: a source zone, root zone, and holding chamber. The source zone is where the silicic melts are generated by complex assimilation and fractionation processes, driven by inputs of mantle-derived mafic melts. The silicic melts then crystallise forming a mush zone (root zone) and the residual melt then percolates its way to the top of the mush and forming a holding chamber, directly above a sharp interface.

Our evidence supports a similar model for the generation of rhyolites at OVC (Fig. 6.16). Beneath the TVZ, primary magmas are most likely generated from melting of crustal materials, such as greywacke, plutons, and/or granulites, perhaps due to mafic intrusion as modelled by Huppert and Sparks (1988). The magmas appear to then crystallise and fractionate, as suggested by the strong fractionation signatures (especially of plagioclase) in TVZ rhyolites (Conrad et al., 1988; Nicholls et al., 1992), most likely generating a crystal-rich mush zone. Geophysical studies by Ogawa et al. (1999) and Bannister et al. (2004) have revealed that there are very conductive mid-crustal zones (between ~15 and 20 km depth) beneath the TVZ which have low S-wave velocities, suggesting that they are zones of partial melt. We suggest the zones of partial melt are the mush zones. Our higher Tₜₑₐ estimates for OVC and TVC magmas are consistent with the rhyolitic proto-melts forming at these greater depths.

Our model for the storage of magma beneath the TVZ is somewhat different to that of Charlier et al. (2005). T and Pₑ data indicate that the phenocrysts formed in equilibrium with the melt (in the storage
Figure 6.16. Schematic illustration of the magmatic systems that resided under both OVC and TVC during the last 50 kyr. The upper ~15 km of crust is dominated by quartzo-feldspathic compositions, such as greywacke or silicic intrusives, with low P-wave velocities ($V_p \sim 5.6$ km/s) (Harrison and White, 2004; Stern and Davey, 1987). The lower crust (between ~15 and 30 km depth) displays seismic velocities of 6.9-7.9 km/s and is interpreted to be anomalously conductive and heavily intruded (Harrison and White, 2004; Henrys et al., 2003). The recent geophysical studies by Ogawa et al. (1999) and Bannister et al. (2004) suggest there zones of partial melt in the mid-crust. We suggest the zones of partial melt are the mush zones where the rhyolitic proto-melts are generated from the crystallisation and fractionation of less evolved melts and crustal assimilation. The proto-melts then ascended and crystallised within high-level crustal storage chambers immediately prior to eruption.
chambers) at various crustal depths (particular to the eruption), in the range 4-16 km with many at 5-7 km. This suggests that once the rhyolitic proto-melts are expelled from the mush zones they move up to shallow levels in the crust, and do not sit directly above the mush. The lack of any crystal fractionation trends (Fig. 6.5c) and the different pressures and temperatures of the magmas, supports the idea that they are not derived from a single common magma and indicates they were separately derived from mush zones. Low crystal contents in most TVZ deposits (~5-15%), and the lack of pre-eruptive gradients, suggest the magmas did not reside at shallow depths for any significant lengths of time. Many post-26.5 ka OVC deposits display evidence for interaction of multiple magma batches (mingled pumice clasts). It appears that the magmas accumulate in the upper crust and often interact. However, disequilibrium crystal assemblages suggest that interaction between the magmas was short-lived. Granitoid lithics, in the Rotoiti and Kaharoa deposits, suggest that some storage chambers cooled and crystallised over longer periods were intersected at depth. Zircons in OVC and TVC eruptives with ages older than their eruption age could be relicts from source rocks, xenocrysts from country rock, and inputs from magma mixing (Shane et al., 2005). Xenocrysts are commonly inherited by magmas at any stage during ascent or eruption, and can represent volumetrically important populations (Gansecki et al., 1996; Gardner et al., 2002).

**Comparison with other volcanoes**

Even though the frequency, volume, eruptive style, and magmatic properties at OVC and TVC differ in detail, they are broadly similar when compared to other rhyolitic centres on Earth. Intra-caldera rhyolite activity at both OVC and TVC is more frequent than that of resurgent North American calderas (e.g., Yellowstone and Long Valley), reflecting differences in tectonic setting and crustal architecture. TVZ is located above thin crust (~30 km) at an active, convergent plate boundary; whereas the North America calderas, Yellowstone and Long Valley, are located above thick, continental lithosphere. North American volcanoes experience caldera-forming episodes rarely (periodicity ~600 kyr for each centre), the subsequent intra-caldera activity is marked by effusive episodes and often small pyroclastic eruption episodes. Examples of this type of intra-caldera activity are observed at Yellowstone and Coso Volcanic field (e.g., Christiansen, 1979; Gansecki et al., 1996; Manley and Bacon, 2000). Coso volcanic field has experienced ~37 moderately sized (~2 km³) rhyolitic eruption episodes in the last 300 kyr (Manley and Bacon, 2000), and Yellowstone has gone through periods of intense rhyolitic intra-caldera activity; for example, a total of >900 km³ of magma was erupted during 21 episodes between 70-170 kyr. Although, the intra-caldera activity within North American calderas is analogous to that of TVZ calderas, the activity, frequency, and volumes of eruptions are significantly different. Eruptions within North American resurgent calderas are typically effusive, and of larger volume than those from TVZ. However, intra-caldera eruption episodes at both OVC and TVC are far more frequent than at North American calderas.
TVZ also differs in eruptive frequency and style from large rhyolitic volcanic centres at other convergent plate boundaries, e.g., Toba Caldera in Sumatra, Indonesia. Both Toba and TVZ are located in rifting arcs behind subduction zones. However, the crust beneath Toba is ~50-70 km thick compared with crustal thicknesses of ~30 km beneath TVZ. Volcanic activity at Toba is similar to that of Yellowstone and appears to be characterised by infrequent, voluminous events. Rhyolitic episodes at Toba occurred at 74 ka erupting 2800 km³ of magma, at 500 ka with 60 km³, and at 840 ka during which 500 km³ of magma was erupted (Chesner et al., 1991; Chesner, 1998). The deposits of the three episodes display continuous compositional and crystal content gradients, consistent with the eruption tapping zoned magma chambers similar to Yellowstone.

One of the major differences in tectonic setting between Toba, Yellowstone, and TVZ is the thickness of the crust. It appears the thick crust beneath both Toba and Yellowstone allows incremental additions of silicic magma to accumulate in chambers without erupting, producing large, long-lived magma bodies that often become zoned as they reside and cool in the crust. In contrast, the crust beneath the TVZ is now thin and cannot provide a large accommodation space. In addition, the crust is weak, attenuated, intensely faulted, and the active rifting (~7 mm/yr) allows the rapidly generated magmas to escape to the surface easily.

CONCLUSIONS

During the past 50 kyr, rhyolitic volcanism at OVC and TVC has been more frequent and voluminous than at any other rhyolitic caldera on Earth. The high frequency of episodes appears to reflect the thin crust, high heat flow, and active tectonics of the central North Island. The two contemporaneously active calderas, OVC and TVC, that have been active in the last 50 kyr display contrasting volcanic and magmatic histories. Following the last caldera-forming episode at ~50 ka (Rotoiti) from OVC, intra-caldera episodes have tapped progressively more evolved (~71-77 wt % SiO₂), shallower (>400-150 MPa), and cooler (940-730°C) magmas. Directly following the caldera-forming episode, eruptions were rhyodacites and derived from deep, hot magmas, containing clinopyroxene. The most recent OVC eruption episodes (post-26.5 ka) were from shallow, cool magmas, and are dominated by hydrous mineral phases. In general, the frequency of episodes decreased and volumes appear to have increased over time at OVC. At TVC, pre-26.5 ka eruption episodes were relatively infrequent, and were derived from shallow (~100 MPa), cool (~750°C) magmas, containing hydrous mineral phases. After the TVC caldera-forming 26.5 ka Oruanui episode, initial eruption episodes were dacitic, and derived from small volume, deep (~400 MPa), and hot (>900°C) magmas. At 12 ka TVC rhyolitic activity re-commenced, episodes since have been more frequent and progressively have tapped deeper (~140-300 MPa) and hotter (~790-850°C) magmas. Following caldera-formation, erupted magmas at both OVC and TVC were the least evolved and from hot (>900°C), deep (~400 MPa) magmas. This suggests that these magmas rose rapidly after formation and exploited new conduits generated during the caldera collapse.
The resumption of TVC rhyolitic activity at 12 ka coincides with increased andesitic volcanism at Tongariro volcanic centre, located ~45 km south, suggesting a tectonic influence (Nairn et al., 1998). Overall, changes in eruptive style, mineralogy, composition, and vent location at both OVC and TVC were contemporaneous and occurred at ~50 and 26.5 ka, following the caldera-forming episodes. This suggests that the contemporaneous changes were in response to fundamental changes in the OVC and TVC tectonic regime, which were likely associated with increased rates of spreading.

$fO_2$ of OVC and TVC eruptives varies temporally and spatially, and not with geochemistry or mineralogy, implying a strong relationship to source (cf. Carmichael, 1991). Thus, source rocks are heterogeneous under each centre and sources under both calderas are distinctly different. At any given temperature, OVC magmas are more oxidised than those from TVC.

Both OVC and TVC eruptive deposits are generally crystal-poor (usually 1-15% crystals). Equilibrium between glass and phenocrysts (including Fe-Ti oxides) suggest crystallisation occurs in upper crustal storage chambers immediately prior to eruption. Most deposits lack pre-eruptive mineralogical, compositional, and physical gradients, suggesting rapid convection and/or short crustal residence times. The lack of temporal whole-rock and trace element (fractionation) trends, and the distinct temperatures and pressures, suggests that the magmas are not derived from a single common magmatic system/source at either OVC or TVC. Volumetrically subordinate mingled clasts in some OVC ejecta suggest that these eruption episodes tap multiple rhyolitic magma batches. The mineral assemblages of the mingled clasts commonly have not re-equilibrated suggesting contact between the magmas was short-lived. Crystal-rich stagnating magmas have become reactivated by new intrusions or engulfed into larger magma bodies. Rhyolitic magma intrusions into another silicic body and basaltic intrusions have triggered episodes at OVC (e.g. Smith et al., 2004). Such pre-eruptive interactions are less evident at TVC.

APPENDIX 6.1: MINERAL CHEMISTRY

Plagioclase

Plagioclase occurs in all OVC and TVC eruptive deposits. It generally occurs as euhedral to subhedral tabular laths usually 2-4 mm in size. Crystals display twinning. However, EMPA shows that the optical zoning commonly observed, does not always reflect compositional variation. OVC and TVC plagioclase occupies a wide compositional range (Fig. 6.17), $An_{15}$-$An_{58}$, Or$_1$-Or$_6$ (Table 6.1 and 6.2) and classifies as oligoclase ($An_{10}$-$20$), andesine ($An_{20}$-$50$), and labradorite ($An_{50}$-$70$) (Deer et al., 1966). Plagioclase from TVC eruptives occupies a slightly narrower range, $An_{20}$-$53$ (mainly andesine). Plagioclase is compositionally variable within deposits of individual eruptions, with ranges of approximately $An\pm6$, Or
±1.5 (±1.5 wt % CaO, and ±0.2 wt % K₂O), somewhat greater than analytical error. Bimodal compositions within single pumice clasts are observed in ~50 ka Rotoiti, 15.7 ka Rotorua, and 0.7 ka Kaharoa, which reflects the tapping of multiple magma batches (Shane et al., 2005; Shane et al., 2003; Smith et al., 2004). Most plagioclase compositions of OVC and TVC deposits overlap and are not unique (Fig. 6.17). The only distinctive plagioclase compositions are those of the Old MSg (Fig. 6.17a; ~An₄₀₋₅₈). This corresponds to their less evolved compositions (70-76.5 wt % SiO₂ glass) and reflects higher temperatures (see section 6.1).

![Graph A](image1)

![Graph B](image2)

**Figure 6.17.** (A) Composition of plagioclase in OVC eruptive deposits. Rotoiti (~50 ka) displays multiple populations associated with distinct magmas (Tables 6.1 and 6.2). (B) Composition of plagioclase in TVC eruptive deposits.

**Quartz**

Quartz is found in Rotoiti Pyroclastics, EFP, most post-26.5 ka OVC deposits, pre-Oruanui TVC deposits, and Oruanui tephra (see Table 6.1 and 6.2). Quartz is common in high-SiO₂ rhyolites (>76 wt %) and with ferromagnesian mineral assemblages that contain hornblende and biotite. It is not found in
any clinopyroxene-bearing deposits. In Rotoiti and EFP, quartz crystals are large (up to 5 mm) and bi-
pyramidal; they commonly contain rounded glass inclusions up to $300 \mu m$ in diameter. In the other
deposits, quartz is usually anhedral to subhedral and 1-3 mm in size. Rounded glass inclusions ($\sim150-
250 \mu m$ in diameter) within crystals are common.

Orthopyroxene
Orthopyroxene is found in all eruptive deposits from OVC and TVC. It displays a mixture of euheiral to
subhedral, and acicular to lath shaped habits. The crystals lack optical and compositional zoning and
are usually $<2$ mm in length. Orthopyroxene crystals commonly contain titanomagnetite (spinel) and
ilmenite (rhombohedral) inclusions. The compositional range orthopyroxene is wide ($En_{40-70}$, $Wo_{1-3}$),
classifying as ferro-hypersthene to hypersthene (Table 6.1 and 6.2). Most deposits display wide
compositional ranges ($\pm4.5$ wt % FeO; $\pm2.5$ wt % MgO; Fig. 6.18) that are not unique to eruption or
eruptive centre. Deposits of the 25 ka Te Rere and 15.8 ka Rotorua eruptive episodes display the
widest variation, reflecting magma mingling (Smith et al., 2004). The TVC dacites and Old MSg have
higher En ($En_{59-65}$; Fig. 6.18), this corresponds to the presence of clinopyroxene and less evolved glass
compositions.

Clinopyroxene
Clinopyroxene found in OVC and TVC eruptive deposits classifies as augite ($En_{36-43}$; $Wo_{40-45}$). It
commonly occurs as subhedral, light green, lath shaped, crystals about 1 mm in length. It is an
abundant ferromagnesian mineral in the Old MSg low-silica rhyolites, where is comprises up to 35 % of
the ferromagnesian mineral assemblage. It also occurs in the Okareka Basin deposits (25 ka Te Rere
and 15.7 ka Rotorua) but only in trace abundance ($<5\%$). At TVC, augite occurs in the post-Oruanui
dacite subgroup, rhyolite subgroup 1 (eruptions B-E), and Unit Y (Taupo). TVC rhyolite subgroup 1
augite is lower in MgO (12-13.1 wt %) and slightly higher in CaO (11.3-13.75 wt %) compared to augite
in OVC tephra and those in the post-Oruanui dacite subgroup (8.9-12.2 wt % CaO; 13.3-15 wt % MgO;
Table 6.1 and 6.2; Fig. 6.19). Variation in augite composition within either subgroup is approximately
equal to analytical error. Glass compositions of augite-bearing deposits are commonly less evolved,
with low SiO$_2$ (mostly $\sim72-76$ wt %), high CaO ($>1.1$ wt %), and low K$_2$O ($<3.3$ wt %).

Amphibole
Amphiboles in OVC and TVC eruptives are calcic ($Ca_b >1.5$ atoms; Mg/Mg $+Fe^{2+} >0.5$), and classify as
magnesio-hornblendes ($(Na + K)_a <0.5$ atoms) following Leake et al. (1997). Analytical totals are
typically $\sim97 \%$, with the difference from 100% attributed to unanalysed OH and F anions. Hornblende
is the most common hydrous phase in OVC and TVC eruptives; however, it is uncommon in post-26.5
ka TVC units (Table 6.1 and 6.2). It occurs as large ($\sim2-3$ mm), dark, elongate euheiral crystals and
commonly displays dark green-brown pleochroism. Dark hornblende crystals within EFP are often up to
7 mm in length.
Figure 6.18. (A) Composition of orthopyroxene in OVC eruptive deposits. (B) Composition of orthopyroxene in TVC eruptive deposits.

Figure 6.19. Composition of clinopyroxene (augite) in OVC and TVC deposits.
Both inter- and intra-eruptive unit compositions of OVC and TVC hornblends occupy a wide range (4.9-10.4 wt % Al₂O₃, 10.9-22.2 wt % FeO, and 9.4-15.9 wt % MgO; Fig. 6.20; Table 6.1 and 6.2). Hornblends in TVC eruptives overlap, but some are generally more FeO-rich than in OVC deposits (Fig. 6.20). Hornblende compositions are not unique to particular eruptive deposits. Some deposits display bimodal hornblende populations. This reflects the tapping of multiple, chemically different magma batches, e.g. 15.7 ka Rotorua (Smith et al., 2004) and ~50 ka Rotoiti (Shane et al., 2005). Compositional zoning within individual crystals is occasionally observed. Most of the compositional variation, in Al, Mn, Ti, and Si, is associated with pressure and temperature induced element substitutions (Bachmann and Dungan, 2002), and thus reflect minor changes in magmatic conditions rather than bulk compositional variations. Substitutions of Si for ³Al (Al in the tetrahedral site) in Rotoma hornblends suggest that the compositional variations are influenced by temperature changes.
Hornblendes in biotite-bearing units are significantly more FeO-rich relative to those in non-biotite bearing units (e.g., high-K₂O Okareka basin eruptives; Fig. 6.20a), reflecting the more evolved melt composition (high-K₂O >3.8 wt %). In addition, those from biotite-bearing, TVC deposits are more FeO-rich than those from OVC. Edenitic and Pargasitic hornblendes in Poihipi (pre-26.5 ka TVC; Fig. 6.20b) are most likely detrital, perhaps derived from an andesitic source.

Cummingtonite
Cummingtonite is only found in OVC deposits. It is common in the ~50 ka Rotoiti, 9.5 ka Rotoma, and 5.6 ka Whakatane eruptives (Table 6.1 and 6.2). It occurs in trace amounts in the 21.8 ka Okareka, 17.7 ka Rerewhakaaitu, and 0.7 ka Kaharoa eruptives biotite-bearing Tarawera eruptive deposits. Cummingtonite occurs as pale yellow, elongate, prismatic crystals, displaying weak or no pleochroism and can reach up to 3 mm in length. It occurs in high-SiO₂ rhyolites (>77 wt %) and commonly occurs with quartz, and with biotite if glass K₂O >3.8 wt %. The compositional range of cummingtonite in OVC deposits is 15.5-19 wt. % MgO and 18.3-23.4 wt. % FeO (Fig. 6.21; Table 6.1 and 6.2). Compositional variation within individual eruptive deposits is within analytical error. MgO and FeO contents allow separate eruptive episodes to be distinguished (Fig. 6.21). These differences reflect fO₂ (see Temperature and oxygen fugacity section).

Figure 6.21. Composition of cummingtonite in OVC deposits. Most of the eruptions from Haroharo, where cummingtonite dominates, have unique compositions. However, those from Tarawera display similar compositions.
Chapter 6

Geochemistry and properties of magmas from OVC and TVC

Biotite

Biotite EMPA totals are typically 94-96%. The deficiency from 100% represents unanalysed components OH\textsuperscript{-}, F, and Ba (e.g. Righter et al., 2002). Cation totals of I= 1.8, M= 5.6-5.8, and T= 8 (I-interlayer cations, M-octahedral cations, T-tetrahedral cations), calculated assuming idealised anions and based on 22 oxygens, are typical of biotite compositions (e.g. Deer et al., 1966). Biotite is common in OVC and TVC eruptives and is useful for the identification and fingerprinting of tephra beds (Shane et al., 2003). Biotite occurs in pre-26.5 ka TVC deposits and in Rotoiti Pyroclastics (≈50 ka), EFP, and post-26.5 ka OVC deposits (25 ka Te Rere, 21.8 ka Okareka, 17.7 ka Rerewhakaaitu, 15.7 ka Rotorua, and 0.7 ka Kaharoa tephra). It commonly occurs as flakes and as euhedral, hexagonal 'books', displaying a mottled green-brown pleochroic appearance. Crystals are usually 3 mm but are often 5-7 mm in the EFP and Rotoiti Pyroclastics, and commonly contain zircon, apatite, and Fe-Ti micro-inclusions. There is no evidence of compositional or optical zoning within crystals.

The presence of biotite is controlled by the melt composition (Shane et al., 2003). It only occurs in deposits with evolved glass compositions (>77 wt % SiO\textsubscript{2}) that is K\textsubscript{2}O-rich (>3.8 wt %) (Fig. 6.22b). Biotite commonly is accompanied by hornblende and quartz. MgO and FeO display the most variation between depositional units (Shane et al., 2003). An inverse relationship is observed between MgO and FeO, with a FeO range of 18-24.5 wt % and a MgO range of 7.9-12.5 wt % (Fig. 6.22). The compositional variation of biotite within individual tephra or eruptive units is usually within analytical error (±0.5 wt %). However, some RSg deposits, display wide compositional ranges and often multiple compositional groups (Shane et al., 2003). This reflects the compositional variation observed in the glass. Biotite heterogeneity in the 21.8 ka Okareka and 0.7 ka Kaharoa tephra reflect the tapping multiple magma batches defined by differing glass compositions (Nairn, 2002; Shane et al., 2003). Although many individual tephra can be distinguished by their biotite composition, there are no fundamental compositional differences between volcanic centres or time intervals (Fig. 6.22).

Fe-Ti Oxides

Fe-Ti oxides are abundant as inclusions within most ferromagnesian phenocrysts. Spinel often occurs as a free phase, crystals are usually subhedral to euhedral and up to 500 μm in diameter. The rhombohedral phase is not as abundant; it usually occurs as tabular to elongate inclusions up to 300 μm in length. Both phases lack compositional zoning. Fe-Ti oxides display exsolution features in some lava flows and pyroclastic flow deposits. Some of the crystals contain glass and apatite inclusions. Compositional trends with co-existing glass indicate the Fe-Ti oxides are co-magmatic (Shane, 1998). Compositions of Fe-Ti oxides are a function of melt composition, temperature, and oxygen fugacity (Carmichael, 1991; Ghiorso and Sack, 1995; Shane, 1998).
Figure 6.22. (A) Compositions of biotite in OVC eruptive deposits. Some eruptions (e.g., Kaharoa) have multiple populations, which are associated with different magmas (Table 6.1 and 6.2). (B) Composition of biotite in TVC deposits.

The spinel phase (titanomagnetite) in OVC and TVC eruptions is compositionally variable ($X_{ulvss} = 0.19-0.43$; Table 6.1 and 6.2). Two OVC and four TVC compositional categories are distinguished (Table 6.9 and 6.10; Fig 6.23a and b). TVC spinels are generally more TiO$_2$-rich (>9 wt %) than OVC spinels (Fig. 3.23a). However, Old MSg are also TiO$_2$ rich (7.1-11.8 wt %) (Fig. 6.23a). Variation of spinel composition within individual eruptive units is usually about ±0.5 wt % TiO$_2$, ±0.2 Al$_2$O$_3$, and ±1 wt % Fe$_2$O$_3$ (Table 6.9). Some of the post-26.5 ka OVC eruptive deposits display wide ranges in spinel composition (e.g., 25 ka Te Rere) and often bimodal groups are observed (e.g., 50 ka Rotoiti and 15.7 ka Rotorua eruptives; Fig. 6.10b and 6.23c). These wide compositional ranges probably result from the tapping of multiple magma bodies. Not all eruptive units have unique spinel compositions, although some deposits can be distinguished, e.g., 0.7 ka Kaharoa and 15.7 ka Rotorua.
The rhombohedral phase in OVC and TVC eruptives is ilmenite ($X_{ilm} = 0.74-0.93$; Table 6.1 and 6.2). Two ilmenite compositional categories can be distinguished in both OVC and TVC eruptives (Fig. 6.24a; Tables 6.9 and 6.10). Post-26.5 ka RSg eruptives occupy a very wide ilmenite compositional range, overlapping with Rotoiti Pyroclastics, EFP, Young MSG, and TVC eruptives. Ilmenites in Unit A and Ngamotu deposits are compositionally distinct from those in other Old Mangaone units, with lower $X_{ilm}$ (0.77-0.84) and $\text{TiO}_2$ (39.6-42.8 wt %), and higher $\text{Fe}_2\text{O}_3$ (21-27 wt %) (Fig. 6.24b). The variation within individual deposits is usually about ±0.9 wt % $\text{TiO}_2$, ±0.3 MgO, and ±1 wt % $\text{Fe}_2\text{O}_3$. Similar to the spinel phase, the rhombohedral phase of some post-26.5 ka OVC eruptive deposits display wide compositional ranges, with multiple populations in the 0.7 ka Kaharoa, 15.7 ka Rotorua, 21.9 ka Okareka, and 25 ka Te Rere eruptives associated with the tapping of multiple magmas.

Figure 6.23. (A and B) Composition of spinel in OVC and TVC eruptive deposits. Numbers relate to categories in Table 6.9.
Figure 6.24. (A) Composition of the rhombohedral phase in OVC and TVC eruptives. Numbers relate to categories in Table 6.9. (B) Composition of rhombohedral phases in Old Mangaone deposits (category 3, Table 6.9). The rhombohedral phases in the two oldest Old MSg deposits (Unit A and Ngamotu) are compositionally distinct from those in other Old MSg deposits.
### Table 6.9 Fe-Ti oxide compositional categories

<table>
<thead>
<tr>
<th>OVC</th>
<th>Ferromagnesian mineralogy</th>
<th>Whole-rock</th>
<th>Glass</th>
<th>Sp</th>
<th>Ilm</th>
<th>T &amp; fO2</th>
</tr>
</thead>
<tbody>
<tr>
<td>Low-K2O Rotoiti (~50 ka)</td>
<td>cum &gt; fe-ti &gt;&gt; hb+opx</td>
<td>WROVC1</td>
<td>GOVC1</td>
<td>S1</td>
<td>R1</td>
<td>TOVC1</td>
</tr>
<tr>
<td>High-K2O Rotoiti &amp; EFP (~50 ka)</td>
<td>bio &gt; hb ± cum &gt; fe-ti ± opx</td>
<td>WROVC2</td>
<td>GOVC2</td>
<td>S1</td>
<td>R1</td>
<td>TOVC2</td>
</tr>
<tr>
<td>Old MSg (40-35 ka)</td>
<td>opx + hb &gt; cpx + fe-ti</td>
<td>WROVC3</td>
<td>GOVC3</td>
<td>S2</td>
<td>R2</td>
<td>TOVC3</td>
</tr>
<tr>
<td>Young MSg (35-31.5 ka)</td>
<td>opx + hb &gt; fe-ti</td>
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<td>GOVC4</td>
<td>S1</td>
<td>R1</td>
<td>TOVC4</td>
</tr>
<tr>
<td>RSg (post-25 ka)</td>
<td>opx + hb ± cum ± bio &gt; fe-ti</td>
<td>WROVC5</td>
<td>GOVC5</td>
<td>S1</td>
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<td>TOVC5</td>
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<tr>
<td>Low-K2O OB (25 ka &amp; 15.7)</td>
<td>opx + hb &gt; fe-ti &gt; cpx</td>
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<td>GOVC6</td>
<td>S1</td>
<td>R1</td>
<td>TOVC4</td>
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<td>GOVC7</td>
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<td>R2</td>
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</tr>
<tr>
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<td>GTVC1</td>
<td>S1</td>
<td>R1</td>
<td>TTVC1</td>
</tr>
<tr>
<td>Oruanui (26.5 ka) &amp; 29 ka Okaia</td>
<td>opx &gt; hb &gt; fe-ti</td>
<td>WRTVC1</td>
<td>GTVC2</td>
<td>S3</td>
<td>R1</td>
<td>TTVC1</td>
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<tr>
<td>Dacites (20.5 ka Ψ, 18.8 ka Ω, 17 ka A)</td>
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<td>GTVC3</td>
<td>S4</td>
<td>R3</td>
<td>TTVC2</td>
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<tr>
<td>Units B-E (11.7-10 ka)</td>
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<td>GTVC4a</td>
<td>S5</td>
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<td>TTVC3</td>
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<tr>
<td>Unit F-S (7.05-3.55 ka)</td>
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<td>GTVC4b</td>
<td>S5</td>
<td>R1</td>
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<tr>
<td>Units U &amp; V (2.85-2.8 ka)</td>
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<td>GTVC4c</td>
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<td>Units X &amp; Y (2.2-1.8 ka)</td>
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<td>R1</td>
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<tr>
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<td>GTVC4e</td>
<td>S5</td>
<td>R1</td>
<td>TTVC7</td>
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Fe2O3, X\textsubscript{a\textsubscript{uv}}\textsubscript{v}, and X\textsubscript{m} calculated using the method of Carmichael (1967).

### Table 6.10 Compositional categories observed in phases from TVZ deposits

<table>
<thead>
<tr>
<th>OVC</th>
<th>Ferromagnesian mineralogy</th>
<th>Whole-rock</th>
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<td>R1</td>
<td>TOVC1</td>
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<tr>
<td>High-K2O Rotoiti &amp; EFP (~50 ka)</td>
<td>bio &gt; hb ± cum &gt; fe-ti ± opx</td>
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<td>GOVC2</td>
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<td>R1</td>
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<tr>
<td>RSg (post-25 ka)</td>
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<td>GOVC5</td>
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<td>TOVC5</td>
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See footnote of Table 1 for mineral abbreviations. a see Table 6.3, b see Table 6.6, c see Table 6.8, and d see Table 6.9.