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$^{40}Ar/^{39}Ar$ geochronology of magmatic activity, magma flux and hazards at Ruapehu volcano, Taupo Volcanic Zone, New Zealand

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Abstract

We have determined precise eruption ages for andesites from Ruapehu volcano in the Tongariro Volcanic Centre of the Taupo Volcanic Zone (TVZ) using $^{40}Ar/^{39}Ar$ furnace step-heating of separated groundmass concentrates. The plateau ages indicate several eruptive pulses near 200, 134, 45, 22 and <15 ka and, based on our and previous field mapping confirm the lavas of the Te Herenga Formation as the oldest exposed part of the volcanic edifice. The pulse at ~ 134 ka includes an entire >300-m section of lavas in Whangaehu gorge as well as some lavas in Ohinepango and Waihianoa catchments on eastern Ruapehu, and this suite of lavas belongs to the Waihianoa Formation. This pulse of activity is not represented on nearby Tongariro volcano, indicating that the two volcanoes have independent magmatic systems. A younger group of lavas yields dates between 50 and 20 ka and includes lava flows from the Turoa skifield and in the Ohinepango and Mangatoetoenui catchments and is consistent with two pulses of magmatism around the time of the last glacial maximum, relating it broadly to the Mangawhero Formation. Syn- and post-last glacial activity lavas, with ages <15 ka are assigned to the Whakapapa Formation, and include the voluminous flows of the Rangataua Member on southern Ruapehu. Magma flux, integrated over 1000-yr periods, averages 0.6 $km^3\ ka^{-1}$ assuming a volcano lifespan of 250 ka. Fluxes for the Te Herenga, Waihianoa and Mangawhero Formations are consistent at 0.93, 0.9 and 0.88 $km^3\ ka^{-1}$, respectively. These fluxes are broadly comparable with those measured at other modern andesite arc volcanoes (e.g. Ngauruhoe, 0.88; Merapi, 1.2 and Karymsky 1.2 $km^3\ ka^{-1}$). The relatively low flux (0.17 $km^3\ ka^{-1}$) calculated for the Whakapapa Formation may derive from underestimates of erupted volume arising from an increase in phreatomagmatic explosive eruptions in postglacial times. However, using volume estimates for the 1995–1996 eruptions and a recurrence interval of 25 yr has yielded an integrated 1000-yr flux of 0.8 $km^3\ ka^{-1}$ in remarkable agreement to estimates for the prehistoric eruptions. Overall, Ruapehu shows consistency in magma flux, but at time scales of the order of one hundred to some thousands

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of years, field evidence suggests that short bursts of activity may produce fluxes up to twenty times greater. This is significant from the perspective of future activity and hazard prediction.

Keywords: Ruapehu Arc volcano; $^{40}Ar/^{39}Ar$ dating; pulsatory growth; magmatic flux; natural hazards

1. Introduction

Ruapehu volcano (2797 m) is the largest (~110 km^3) active andesite–dacite volcano in the onshore part of the Taupo Volcanic Zone (TVZ) of New Zealand. It has been studied extensively (Clark, 1960; Cole et al., 1986; Graham and Hackett, 1987; Houghton et al., 1987; Hackett and Houghton, 1989) and on the basis of detailed geological mapping Hackett (1985) established a stratigraphy comprising four major formations. From the oldest to the youngest, these are the Te Herenga, Waihianoa, Mangawhero and Whakapapa Formations. Until recently (Tanaka et al., 1997; this work) there were very few (e.g. Stipp, 1968) radioisotopic age data upon which to base the chronology of the past eruptive history or to calculate the magma flux to the volcano. The combined palaeomagnetic and K–Ar dating study by Tanaka et al. (1997) focussed on the Whakapapanui valley on the northwest slopes of the volcano, but details of the sampling are not recorded. Results from this work confirm the ages of the Te Herenga Formation and those of the younger overlying Whakapapa Formation. In this paper we report results from a $^{40}Ar/^{39}Ar$ dating study of key sections through each of the four formations on the Whakapapa and Turoa skifields and Whangaehu and Waihianoa valleys.

A number of recent studies of andesitic to dacitic arc volcanoes, including Mount Adams in the USA (Hildreth and Lanphere, 1994); Mount Tongariro in New Zealand (Hobden et al., 1996, 1999, 2002) and the Tatara–San Pedro Complex in Chile (Singer et al., 1997; Dungan et al., 2001), have shown that much useful information can be gained by integrating high precision dating studies with detailed field observations and geochemical measurements. This information can be used to understand the longevity, periodicity of activity, and magma flux rates on a volcano, and therefore has an important bearing on hazard mitigation strategies and prediction models.

2. Field relations

Mount Ruapehu (2797 m) is the highest mountain in the North Island of New Zealand forming an eroded active strato-volcano at the southern extremity of the TVZ (Wilson et al., 1995; Bibby et al., 1995; Fig. 1). In a benchmark study of the geology of Ruapehu volcano, Hackett (1985) distinguished four chronostratigraphic units (see also Hackett and Houghton, 1989), which, from the oldest to the youngest, were named the Te Herenga, Waihianoa, Mangawhero, and Whakapapa Formations (Fig. 1). Nowhere on the volcano, are all four formations juxtaposed, and until recently only three radioisotopic (conventional K–Ar, two of which were imprecise) age determinations by Stipp (1968) were available to form the basis of chronostratigraphy and magma flux calculations.

The proximal cone-building volcanic rocks include variably dipping lava flow sequences, autoclastic breccias, reworked slope-wash and glacial deposits (Smith et al., 1999). On Pinnacle Ridge (Northwest Ruapehu) these units are cut by dykes and shallow intrusives that were feeder systems to early conduits. Zones of propylitic alteration and disseminated sulphide mineralisation accompany these intrusions. Pyroclastic deposits are rare on the proximal cone. They include welded deposits of the Whakapapa Formation which drape the Te Herenga Formation lavas on Pinnacle Ridge (Hackett and Houghton, 1985, 1989), welded deposits in the Mangatoetoenui catchment (Hackett, 1985; Chapman, 1996) and adjacent to Mangatuturutu Shelter, and localised unconsolidated, reworked pumiceous deposits above 1500 m.

Extensive pyroclastic fall, laharic deposits, re-

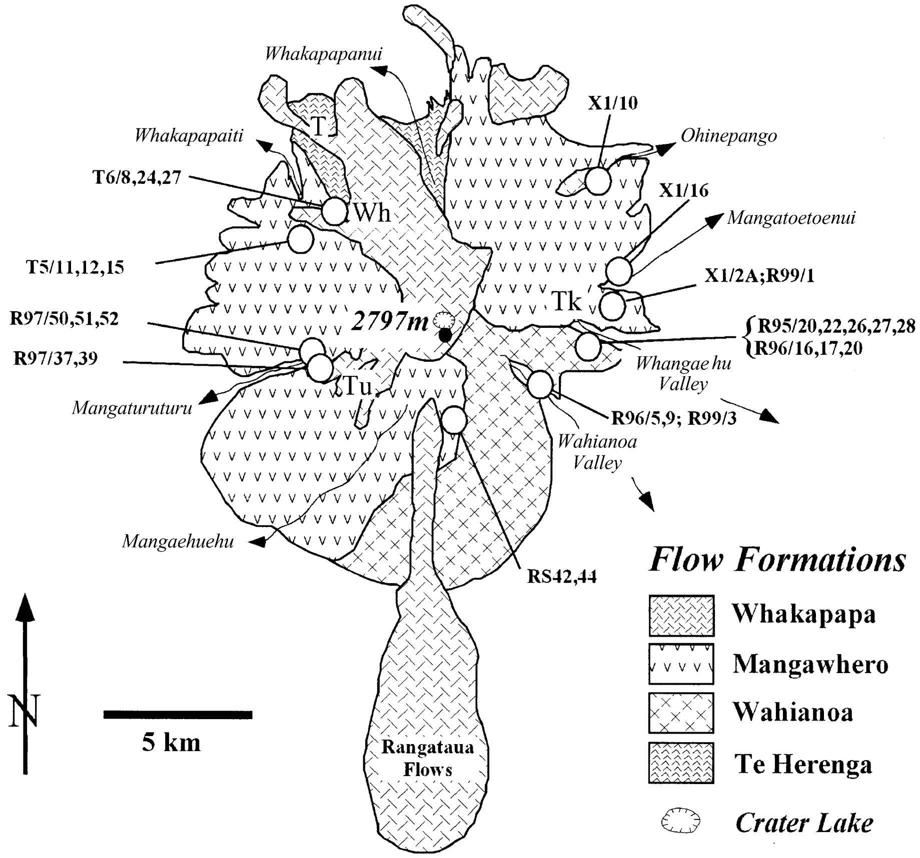

Fig. 1. Map showing distribution of principal volcanic flow sequences (Formations) of Ruapehu Volcano (Hackett, 1985; Price et al., unpublished data). Location of samples used for Ar–Ar dating are also shown. Full details of samples are contained in Table 1. Key to localities on map as follows: T, Te Herenga Ridge; Wh, Whakapapa skifield; Tu, Turoa skifield; Tk, Tukino skifield.

worked sheet flow deposits and rare pyroclastic flows blanket the ring plain surrounding the volcano (Hackett and Houghton, 1989; Donoghue et al., 1995a, 1997, 1999). Here, well dated and intercalated rhyolitic tephras, sourced from the Taupo, Whakamaru and Okataina caldera centres to the north (Wilson, 1993), permit correlation with events elsewhere in the TVZ. Importantly, these distal rhyolitic tephras provide a useful and underpinning chronological framework, based on radiocarbon dating, extending back to around 26 ka. Over the past 2 ka volcanic activity at Ruapehu has been largely focussed through a vent system presently located beneath the Crater Lake. As a result, most recent activity has been phreatomagmatic in character and the record is preserved in the Tufa Trig tephras on the ring plain (Donoghue et al., 1997).

The dynamic erosional environment on the proximal cone has resulted in deep dissection, at various times, through a number of the cone building episodes. As a result, deep gorges have been cut through the volcano flanks, and these offer the potential to sample deep into the internal stratigraphy of the volcano. Of these gorges, the Whangaehu and Waihianoa gorges on the east and southeast flanks (Figs. 1 and 2) are the deepest, the former being the pathway for most lahars during the 1995–1996 eruption (Cronin et al., 1997) and the route of the lahar which caused the Tangiwai rail disaster of Christmas Eve 1953 (Houghton et al., 1987).

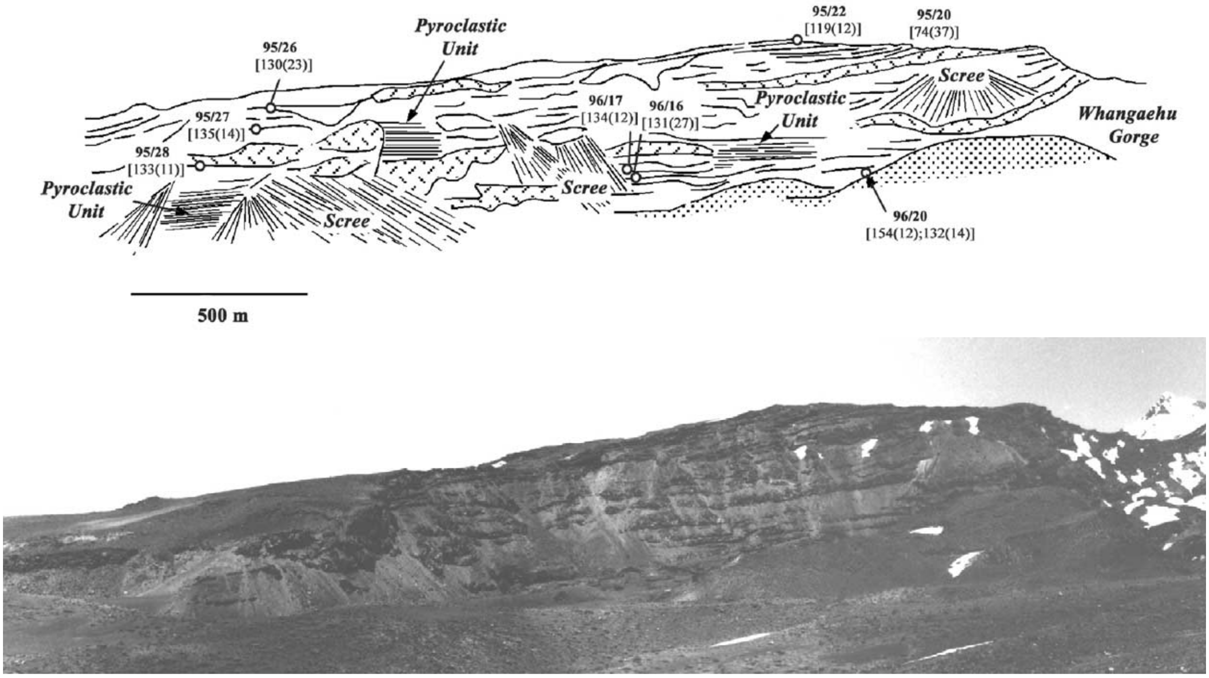

Fig. 2. Photographic panorama of the upper Whangaehu gorge section of the Wahianoa Formation on eastern Ruapehu (see Fig. 1). The sketch shows a geological interpretation and locations of samples. Ages are shown in parentheses as [Age(error)].

For this work samples were collected from mapped and documented sections in the Whangaehu River gorge, the Turoa and Whakapapa skifields and from other localities in Ohinegango, Mangatoetoenui and Waihianoa valleys (Fig. 1).

3. Mineralogy and petrology

Lavas and pyroclastic rocks from Ruapehu volcano are variably porphyritic basaltic andesites, andesites and rare dacites in which the phenocryst assemblage is dominated by plagioclase, with lesser clinopyroxene and orthopyroxene and Fe–Ti oxides. Olivine is a rare constituent of some lavas and is generally jacketted by Ca-poor pyroxene. Amphibole is also rare, but not entirely absent from Ruapehu lavas and tephras (Graham and Hackett, 1987; Gamble et al., 1999; Donoghue et al., 1997; Nakagawa et al., 1999). In a detailed study of the petrogenesis of the volcano, Graham and Hackett (1987) identified six distinctive magma types based upon a combination of petrography and geochemistry, of which the plagioclase+clinopyroxene+orthopyroxene+Fe–Ti oxide assemblage was most common. These authors identified long-term geochemical evolution in the volcano, based on crustal assimilation with fractional crystallisation (AFC). Later, more detailed work on measured stratigraphic sections (Price et al., 1997, 2000; Gamble et al., 1999) has modified this to include open system processes operating on a periodically flushed, continuously fractionating, periodically mixing magma system. Interaction of these processes at various levels in a dispersed magmatic system fed through a plexus of dyke- and sill-like magma reservoirs has led to non-systematic trends on Harker-type geochemical plots as batches of melt were displaced and overwhelmed by new influxes of magma.

4. Sample selection and $^{40}Ar/^{39}Ar$ analytical methods

Accurate and precise K–Ar or $^{40}Ar/^{39}Ar$ dating of young (< 1 Ma) volcanic rocks, particularly those low in K as are typical of arc volcanoes, has proven a difficult challenge to geochronologists, demanding careful sample selection and preparation and optimum analytical procedures (Hildreth and Lanphere, 1994; Singer et al., 1997). In the course of this study, we have tried several strategies to obtain reproducible, precise data. We first analysed whole rock samples, with limited success. Next we tried pure plagioclase separates, in the belief that melt inclusions (occult in plagioclase phenocrysts from Ruapehu) or plagioclase phenocryst rims would contain high K. The imprecise results obtained from this experiment led us to search for alternative procedures and to eventually select crystalline groundmass separates for dating experiments.

In general, samples selected for analysis were collected from the central parts of relatively thick (up to 10 m) lava flows. Hildreth and Lanphere (1994) adopted similar strategies in their study of Mount Adams. Thin sections were studied to select samples with crystalline groundmass and polished and carbon-coated sections of selected samples were then surface mapped for K-content by electron microprobe analysis. In the most crystalline samples most of the K was found to be held in glass or groundmass μ-scale sanidine crystals formed during late stage groundmass crystallisation. This approach permitted a quantitative estimate of the proportions of groundmass glass to crystalline material, and led us to classify groundmass samples into four categories: glassy, moderately glassy, moderately crystalline, and crystalline (Table 1). In the moderately crystalline to crystalline samples, crystallisation of groundmass glass resulted in a vermicular intergrowth of quartz, sanidine ($\sim Or_{65}$) and Fe–Ti oxides, where individual grains measured 1–20 μ across. The traces of residual interstitial glass in the moderately crystalline samples had rhyolitic (> 70% SiO_2, ~4–5% K_2O) compositions.

Groundmass concentrates were prepared from selected samples by crushing, sieving, washing, drying and hand picking a 0.3–1-mm size fraction under a stereo microscope. Great care was taken to ensure that the hand picked grains were both phenocryst- and xenolith-free.

Samples were irradiated at the Ford Research Reactor in Lansing, MI, and analysed by the $^{40}Ar/^{39}Ar$ incremental-heating resistance-furnace method at the New Mexico Geochronology Research Laboratory in Socorro, NM. Details of irradiation and analytical procedures are in the footnotes to Table 1, which also summarises the analytical results. Complete analytical data are in McIntosh and Gamble (2002).

5. Results

Table 1 summarises results from a total of 33 analyses of 28 samples; replicate analyses were made of five samples. Representative spectra are shown in Fig. 3, along with plots of radiogenic yield (percent of non-atmospheric ^{40}Ar), K/Ca ratios (calculated from K-derived ^{39}Ar and Ca-derived ^{37}Ar), and K/Cl ratios (calculated from K-derived ^{39}Ar and Cl-derived ^{38}Ar). The quality of the age spectra varied widely among samples. Groundmass concentrates from samples with fully crystallised groundmass, as well as some of the samples with small amounts of groundmass glass, yielded generally flat spectra having relatively precise individual step ages, radiogenic yields as high as 20%, K/Ca ratios near 1.0, and high K/Cl ratios (e.g. Fig. 3a–c). Groundmass concentrates from glassier samples tended to yield age spectra that were slightly to strongly discordant, with less precise individual step ages, low radiogenic yields, similar K/Ca ratios, and lower K/Cl ratios than crystalline groundmass (e.g. Fig. 3d). For 30 analyses, plateau ages were calculated for flat portions of the age spectra (selected using the criteria of Fleck et al., 1977). Total gas or preferred ages were calculated for the remaining three more discordant age spectra. Calculated ages for all analyses are summarised in Table 1. We have rejected results from eight analyses (seven samples and one replicate analysis) because of low-precision or strongly discordant age spectra (Table 1). We have accepted results from 25 analyses (21 samples and four replicate analyses). For each of the four accepted pairs of replicate analyses, plateau ages agree within 2σ uncertainty (Table 1). As detailed below, none of the accepted age determinations conflicts with established stratigraphic relationships. The incremental heating results were also plotted on isotope correlation diagrams. In all cases, isochron intercept ages agree with plateau ages, and $^{40}Ar/^{36}Ar$ intercepts are near-atmospheric, suggesting that groundmass concentrates from our samples of Ruapehu volcano do not contain significant extraneous ^{40}Ar. The plateau

Table 1
$^{40}Ar/^{39}Ar$ age determinations from Mount Ruapehu lavas

Sample number	Location	Map reference	Groundmass crystallinity	Laboratory number	Age analysis	n	K/Ca	Age (ka) $\pm 2\sigma$
Postglacial Lavas								
R97-37	Sunset Ridge (young lava)	S20-24091	Xtalline	50447-01	total gas	9	0.4	−336 ± 177[a]
X1/2A	Mangatoetoenui	T20-357115	Glassy	8903-01	plateau	7	0.4	3 ± 17[a]
Mangawhero B Formation								
X1/16	Mangatoetoenui	T20-365111	Mod. glassy	8091-01	preferred	8	0.7	22 ± 7
X1/16	Mangatoetoenui	T20-365111	Mod. glassy	8885-01	plateau	11	0.7	29 ± 3[b]
X1/6	Ohinepango	T20-387149	Xtalline	8900-01	plateau	7	0.3	23 ± 10
X1/6	Ohinepango	T20-387149	Xtalline	8889-01	plateau	5	0.4	21 ± 6[b]
R99-3	Waihianoa, north side	T20-351089	Mod. xtalline	50445-01	plateau	8	0.2	28 ± 60[a]
R97-39	Mangaturuturu 'D'	S20-292093	Mod. xtalline	50437-01	plateau	6	0.5	23 ± 8
R99-1	Mangatoetoenui, below X1/2A	T20-357115	Mod. xtalline	50427-01	plateau	7	0.7	23 ± 4
R97-52	Turoa	S20-273093	Mod. xtalline	50421-01	plateau	5	0.3	26 ± 16
R97-51	Turoa	S20-273095	Mod. glassy	50433-01	plateau	6	0.3	−10 ± 22[a]
Mangawhero A Formation								
R-S44-10	Mangaehuehue/ Wahianoa	T20-320064	Xtalline	50431-01	plateau	5	1.1	47 ± 4
R-S42-10	Mangaehuehue/ Waihianoa	T20-320064	Xtalline	50429-01	plateau	3	1.2	53 ± 3[b]
T5-11	Whakapapaiti (lower)	S20-396941	Mod. glassy	50439-01	plateau	8	1.2	46 ± 5
T5-12	Whakapapaiti (lower)	S20-540846	Mod. glassy	50441-01	total gas	9	0.6	37 ± 84[a]
T5-15	Whakapapaiti (upper)	S20-798363	Mod. glassy	50449-01	plateau	5	1.2	−9 ± 35[a]
T5-15	Whakapapaiti	S20-798363	Glassy	50450-01	plateau	5	1.2	−36 ± 34[a,b]
Waihianoa Formation								
X1/10	Ohinepango	T20-354135	Xtalline	8899-01	plateau	9	0.3	1.38 ± 14
R95/22	Whangaehu	T20-340093	Mod. xtalline	8904-01	plateau	6	0.3	119 ± 12
R96/20	Whangaehu	T20-339097	Mod. xtalline	8902-01	plateau	3	0.3	132 ± 14
R96/20	Whangaehu	T20-339097	Mod. xtalline	8886-01	plateau	5	0.3	154 ± 12[b]
R95/20	Whangaehu	T20-338094	Glassy	9227-01	plateau	6	0.2	74 ± 37[a]
R95/26	Whangaehu	T20-357094	Mod xtalline	9229-01	plateau	9	0.2	130 ± 23
R95/27	Whangaehu	T20-358095	Xtalline	9230-01	plateau	6	0.3	135 ± 14
R95/28	Whangaehu	T20-359096	Xtalline	9231-01	plateau	9	0.3	133 ± 11
R96/17	Whangaehu	T20-348094	Xtalline	9232-01	plateau	10	0.3	134 ± 12
R96/16	Whangaehu	T20-348095	Xtalline	9225-01	plateau	9	0.2	131 ± 27
R96/5	Wahianoa	T20-348071	Mod. xtalline	9228-01	plateau	10	0.3	129 ± 15
R96/9	Wahianoa	T20-350067	Mod. glassy	8898-01	plateau	7	0.3	147 ± 12
R97-50	Turoa	S20-275092	Mod. xtalline	50423-01	plateau	6	0.2	147 ± 10
Te Herenga Formation								
T6-27	Te Herenga (top)	S20-305163	Mod. glassy	50435-01	plateau	6	0.2	205 ± 27
T6-24	Te Herenga (middle)	S20-305162	Xtalline	50425-01	plateau	5	0.3	197 ± 12
T6-8	Te Herenga (base)	S20-309158	Xtalline	50443-01	plateau	5	0.2	183 ± 13

ages for the 21 best-behaved samples are considered to be accurate estimates of eruption ages. These data are summarised in Fig. 4 together with the published data of Tanaka et al. (1997) and the unpublished analyses of Stipp, 1968. The combined data sets are consistent with major pulses of volcanism around 200, 160–115, 55–45, 30–15, and less than 15 ka (Fig. 4). The 200, 160–115, and <15 ka are broadly consistent with the Te Herenga, Waihianoa, and Whakapapa Formations of Hackett and Houghton (1989). The data also indicate that the Mangawhero Formation (Hackett and Houghton, 1989) comprises two events, i.e. at ~ 50 and ~ 23 ka. This is consistent with mapping in the Turoa–Mangaturuturu catchments (Waight et al., 1999).

For the Te Herenga Formation, we dated selected lavas from the type section (Hackett and Houghton, 1989) on Te Herenga Ridge. The ages of the three samples range from 205 ± 27 to 183 ± 13 ka. This ridge marks an erosional remnant of Te Herenga Formation lavas that confine the Whakapapa Formation lavas on Northwest Ruapehu. Tanaka et al. (1997) dated samples collected from the Whakapapanui gorge and Te Herenga Ridge but precise sample locations are not given. The ages overlap with those of Tanaka et al. (1997), but with significantly lower errors ($\pm$13–27 ka or better as distinct from $\pm > 30$ ka). We conclude that our dated samples are likely to be representative of the youngest Te Herenga Formation. The data of Tanaka et al. (1997) suggest that the age of the Te Herenga Formation may extend back to around 250 ka. Combining our data set with Tanaka et al. (1997) suggests the best estimate for the duration of the eruption of the Te Herenga Formation is 250–180 ka.

More than half of the samples dated in this study are assigned to the Waihianoa Formation of Hackett and Houghton (1989). Results from 11 samples, including two samples from the Waihianoa Formation type section and seven samples from the 300-m section of lavas and pyroclastic rocks exposed in the south face of Whangaehu gorge (Fig. 2), fall into a narrow interval ranging from 154 ± 12 to 119 ± 12 ka. The ages of samples from the Whangaehu gorge show remarkable agreement with stratigraphic position through the section (Fig. 2). The lowest stratigraphic samples (Figs. 2 and 5d) give ages (R96/20 154 ± 12 and 132 ± 14 ka for replicate analyses) which are measurably older than the uppermost samples (e.g. R95/22 119 ± 12 ka) suggesting that the entire sequence erupted between 150 and 120 ka. These ages are similar to the ages of samples from the Ohinepango (X1/10 138 ± 14 ka), Wahia-

Legend for Table I. Groundmass crystallinity assessed by electron microprobe in backscattered electron mode. *n* is the number of heating steps used to calculate age. K/Ca is calculated from measured $^{37}Ar_{Ca}/^{39}Ar_K$.
Sample preparation and irradiation. Groundmass concentrates prepared by crushing, sieving, and removal of plagioclase phenocrysts. Samples packaged in Cu foil and irradiated in machined Al discs for 3 h in L67 position, Ford Research Reactor, Lansing, MI. Neutron flux monitor Fish Canyon Tuff sanidine (FC-1). Assigned age = 27.84 Ma (Deino and Potts, 1990) equivalent to Mmhb-1 at 520.4 Ma (Samson and Alexander, 1987).
Instrumentation. Mass Analyzer Products 215-50 mass spectrometer on line with automated all-metal extraction system. Samples step-heated in Mo double-vacuum resistance funace. Heating duration 8 min. Reactive gases removed by reaction with 3 SAES GP-50 getters, two operated at $\sim$400°C and one at 20°C. Gas also exposed to a W filament operated at $\sim$2000°C.
Analytical parameters. Electron multiplier sensitivity averaged 1×10^{-16} mol/pA. Total system blank and background for the furnace averaged 460, 5, 0.6, 1.4, 2.2×10^{-18} mol at masses 40, 39, 38, 37, and 36, respectively, for temperatures <1300°C. J-factors determined to a precision of $\pm 0.1\%$ by CO_2 laser-fusion of four single crystals from each of three radial positions around the irradiation tray. Correction factors for interfering nuclear reactions were determined using K-glass and CaF_2 and are as follows: $(^{40}Ar/^{39}Ar)_K = 0.034 \pm 0.01$; $(^{36}Ar/^{37}Ar)_{Ca} = 0.00026 \pm 0.0002$; and $(^{39}Ar/^{37}Ar)_{Ca} = 0.00070 \pm 0.00005$.
Age calculations. Plateau definition: three or more analytically indistinguishable contiguous steps comprising at least 50% of the total ^{39}Ar (Fleck et al., 1977). Preferred age calculated for indicated steps when the sample does not quite meet plateau criteria. Plateau or preferred ages calculated by weighting each step by the inverse of the variance. Plateau and preferred age errors calculated using the method of Taylor (1982). Decay constants and isotopic abundances after Steiger and Jaeger (1977). All errors reported at $\pm 2\sigma$, unless otherwise noted.
[a] Results rejected because of poor precision of disturbed age spectra.
[b] Replicate of preceding analyses.

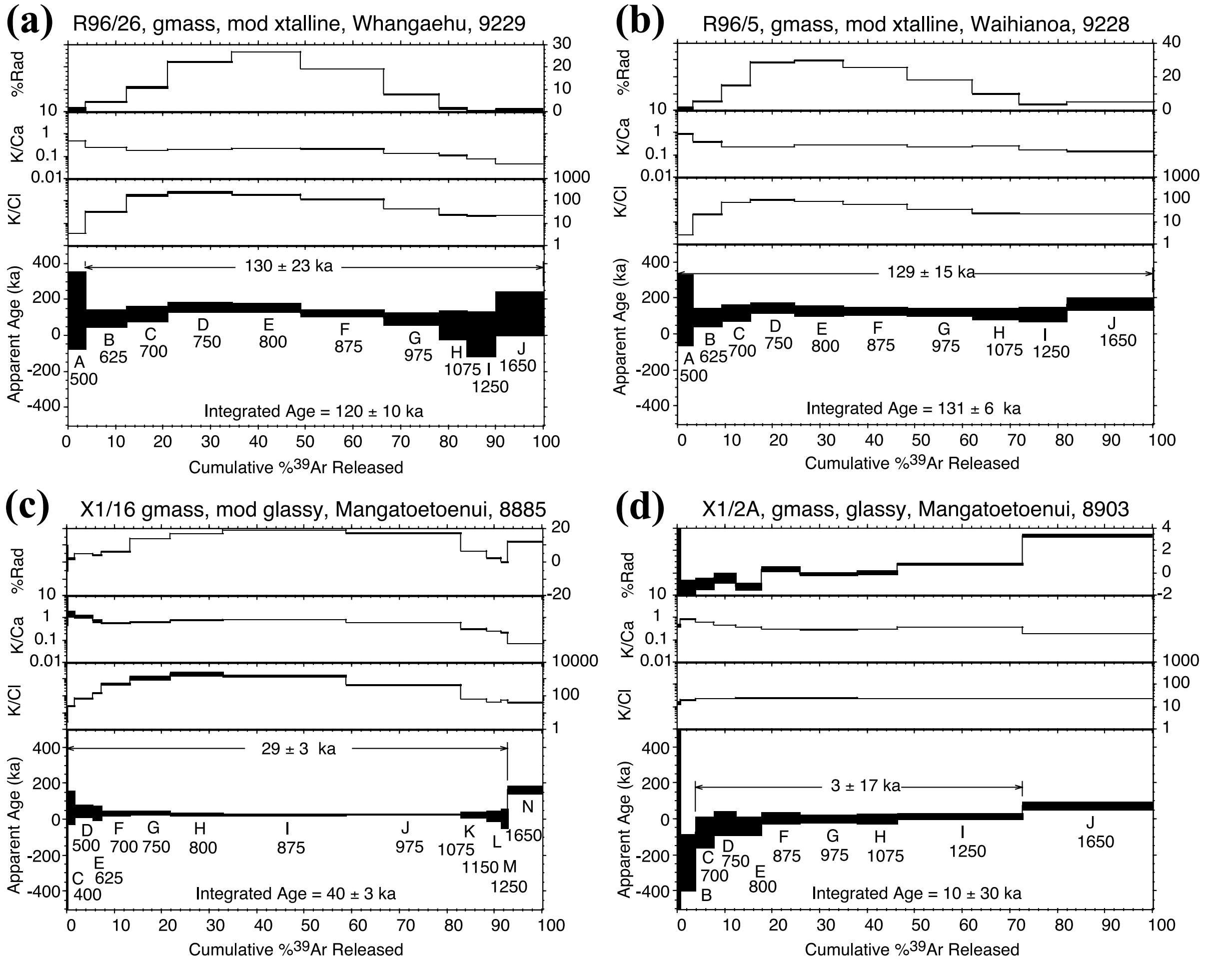

Fig. 3. Representative Ar–Ar age spectra for analysed samples from Ruapehu. See Table 1 for details of sample locations, age data, and analytical conditions.

noa (R96/5 129 ± 15 ka; R96/9 147 ± 12 ka) and Turoa catchments (R97/50 147 ± 10 ka), indicating that this time window produced a laterally extensive, voluminous lava flow sequence on the volcano. Sample R95/20 is from the upper part of the section and major and trace element data suggest that the flow represented by this sample may in fact be part of the younger Mangawhero Formation. The imprecise, but young age for R95/20 (74 ± 37 ka) is permissive of this interpretation.

The plateau ages of samples assigned to the Mangawhero Formation form two tight clusters, i.e. near 50 and 23 ka. For most of these plateau ages, 2σ errors are less than $\pm < 10$ ka. Reproducible replicate analyses from two samples (21 ± 6 and 23 ± 10 ka, Sample X1/6; and 22 ± 7 and 29 ± 3 ka, Sample X1/16) from the Ohinepango and Mangatoetoenui catchments are consistent with stratigraphy and the juxtaposition of overlying lavas. Similarly, samples R97/39 (23 ± 8 ka) and R99/1 (23 ± 4 ka) from the well mapped Turoa skifield area (Waight et al., 1999) overlie and are consistent stratigraphically with the older age (147 ± 10 ka) for the underlying sample R97/50. Furthermore, to the northeast of the Turoa skifield, samples R44/10 (47 ± 4 ka) and R42/10 (53 ± 3 ka) are from erosional inliers of a flow sequence bounding the younger lavas of the Tu-

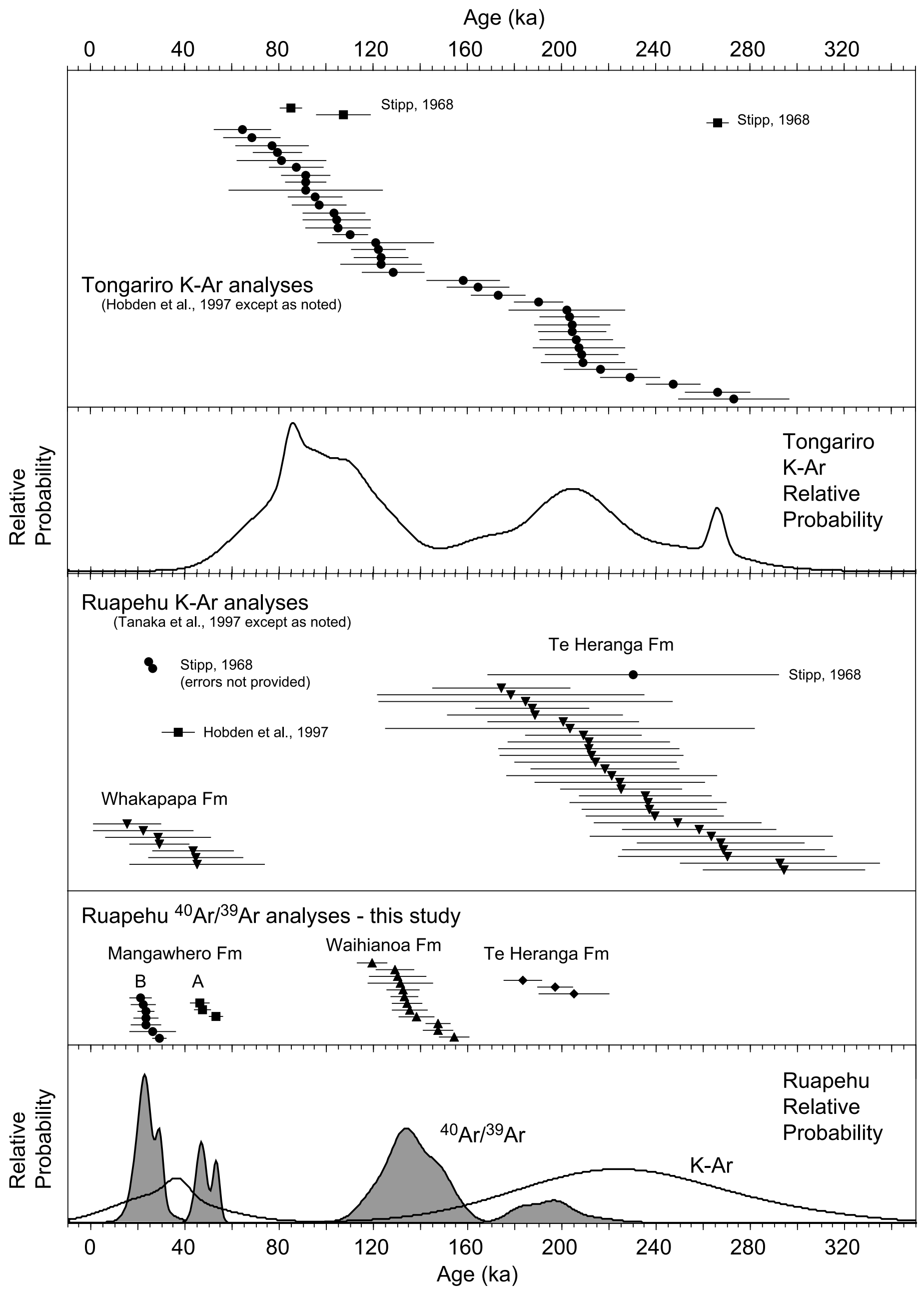

Fig. 4. $^{40}Ar/^{39}Ar$ data from this study and published K–Ar age determinations from Mount Ruapehu and Mount Tongariro. Upper panels show individual dates with $\pm 1\ \sigma$ errors and lower panels show age-probability curves, which are uncertainty weighted histograms constructed by summing probability distributions of individual analyses (Deino and Potts, 1992). The better precision and accuracy of the $^{40}Ar/^{39}Ar$ method reveals four discrete eruptive intervals at Mount Ruapehu.

roa skifield area. Based on these results and local field relationships in the Turoa skifield area, we concur with Waight et al. (1999) that temporally and geochemically distinct subdivisions exist within the Mangawhero Formation. Based on a clustering of the age data we identify the Mangawhero A (~50 ka) and Mangawhero B (~23 ka) groups.

Our efforts (see also Tanaka et al., 1997) to date samples younger than 20 ka have met with limited success despite care in sample selection and particular attention to analytical strategy. One sample (R97-37) with a crystalline groundmass gave a highly discordant age spectrum with an imprecise and total gas age, possibly related to incomplete removal of reactive gasses during argon analysis. A second sample (X1/2A) with a glassy groundmass yielded an age of 3 ± 17 ka, which is too imprecise to be useful. As a result, our chronology for the Whakapapa Formation lavas remains sketchy and the detail is still best resolved by correlating data from tephra sections on lavas with those of the surrounding ring plain (see Donoghue et al., 1995a) which are generally well constrained by radiocarbon dating.

6. Discussion

Our results demonstrate the potential of dating young andesitic volcanic rocks by the $^{40}Ar/^{39}Ar$ method. With the high precision achieved through careful sample selection, microprobe characterisation, sample preparation and step-heating, new insights into the growth history of an arc volcano are now possible. Sections below consider some of the implications of the $^{40}Ar/^{39}Ar$ chronology established by this study for Ruapehu volcano.

6.1. Inception of volcanism at Tongariro Volcanic Centre

Although the oldest lavas dated in this study are the ~200 ka Te Herenga flows, there is evidence for older volcanism at Ruapehu volcano. Some of the Te Herenga Formation lavas dated by Tanaka et al. (1997) were collected from outcrops in Whakapapanui gorge and are demonstrably lower stratigraphically than the Te Herenga lavas dated herein. Results from these samples extend the age of Te Herenga lavas to at least 300 ka. Moreover, the occurrence of andesite clasts (> 90%) in conglomerates (O'Leary Conglomerate) of the Kai-iwi Group (Fleming, 1953) near Wanganui, 100 km southwest of Ruapehu (Parish, 1994) also are directly relevant. These conglomerates are part of the cover bed sequence of the Brunswick Terrace, dated at ~310 ka (by a combination of strandline reconstruction, tephrochronology palynology and amino acid racemisation dating; Pillans, 1990; Bussell and Pillans, 1992). Clast petrography and geochemistry match those of andesites from the Te Herenga and Waihianoa Formations of Ruapehu (Parish, 1994). As such they testify to the existence of volcanism in the Tongariro Volcanic Centre prior to our measured radioisotopic record. Furthermore, rare partially welded plagioclase+quartz+hornblende+biotite+Fe–Ti oxide-bearing rhyolitic ignimbrite clasts in the same deposit are petrologically and geochemically similar to Whakamaru Group ignimbrites, dated at ~340 ka (Brown et al., 1998; Wilson et al., 1995), giving a maximum age to the deposit. We therefore suggest that 340 ka is a reasonable maximum age for the inception of volcanism at Ruapehu and Tongariro volcanoes.

6.2. Implications for magma flux

Hackett and Houghton (1989) estimated a volume of 147.6 km^3 for Ruapehu volcano. Using this volume and 250 ka for the duration of activity yields an average magma flux of 0.6 $km^3\ ka^{-1}$ (Table 2). Using a volume of 300 km^3 as representative of the cone plus the reworked deposits and tephras of the ring - plain, yields an average flux of 1.2 $km^3\ ka^{-1}$ (using 300 km^3 and a maximum age of 340 ka gives 0.88 $km^3\ ka^{-1}$). These estimates are broadly similar to the flux (0.9 $km^3\ ka^{-1}$) that produced the 2.2 km^3 cone of Ngauruhoe volcano in the past 2.5 ka (Hobden and Houghton, 2000; Hobden et al., 2002).

When we compare the chronologies for Ruapehu and Tongariro volcanoes (Fig. 4), it can be seen that the two have grown in parallel, but

Table 2
Calculated magma fluxes for Ruapehu volcano

Formation	Volume[a] (km^3)	Age Range[b]	Duration (ka)	Magma flux
Te Herenga	65	250–180 ka	70	0.93 km^3 ka^{-1}
Waihianoa	45	160–115 ka	45	1.0 km^3 ka^{-1}
Mangawhero	35	55–45 ka	35	0.88 km^3 ka^{-1}
		30–15 ka		
Whakapapa	2.6	< 15 ka	15	0.17 km^3 ka^{-1}
Average flux 1[c]	147.6	250–0	250	0.59 km^3 ka^{-1}
Average flux 2[d]	300	250–0	250	1.2 km^3 ka^{-1}

[a] Based on Hackett and Houghton (1989).
[b] Based on $^{40}Ar/^{39}Ar$, this work.
[c] Based on Formation volume estimates of Hackett and Houghton (1989).
[d] Based on cone volume estimates (above), plus equivalent volume ring plain.

with distinct pulses (time windows) of enhanced magma production. For Ruapehu these pulses occurred at ∼10, 20, 50, 115–160, and 180–250 ka. For Tongariro, the pulses occurred at < 2.5 ka (Ngauruhoe cone), 10, 25, 70–130, 180–230, and 260–280 ka (Hobden et al., 1996). With the volume estimates of Hackett and Houghton (1989) for the various stages of growth of Ruapehu (Table 2) we can use our Ar–Ar data to calculate fluxes of magmatism over the history of the volcano. Thus for Te Herenga Formation we calculate a flux of 0.93 km^3 ka^{-1}; for Waihianoa Formation we calculate 1.0 km^3 ka^{-1} and for Mangawhero Formation 0.88 km^3 ka^{-1}. Our calculated flux for the Whakapapa Formation (0.17 km^3 ka^{-1}) is significantly lower than the other formations. We note that this may be related to the relatively short time span (15 ka) and to the fact that over a significant portion of this time, activity has been pyroclastic, with poor preservation on the cone. However, using the small volume of the 1995–1996 eruptions (0.02 km^3) and a 25-yr repeat interval yields a 1000-yr flux of 0.8 km^3, within the limits of our calculated time averaged fluxes of 0.59 and 1.2 km^3 ka^{-1} (see above and Table 2). Overall we note that our calculated fluxes will be underestimates because we have not considered the associated pyroclastic and reworked deposits or magmas trapped in the crust below the volcano. For Ruapehu our mapping and radioisotopic dating suggest that maximum growth occurred during the Waihianoa stage (115–160 ka) of activity. Within the limitations of the dating precision and field sampling, Tongariro appears to have been relatively quiet during the Waihianoa stage on Ruapehu.

Also relevant to this discussion, Nairn et al. (1998) have documented a short, intense period of activity around 10 ka (the Pahoka–Mangamate event) from a multiple vent system located between Tongariro and Ruapehu volcanoes. This activity is suggested to have occurred over a time period of only a few hundred years. Using the isopach constrained pyroclastic deposit volumes and 400-yr duration (Nairn et al., 1998), a magma flux of ∼14 km^3 ka^{-1} can be calculated. This is more than an order of magnitude greater than our estimates for the Waihianoa ‘pulse’ (see above) and nearly two orders greater than the present average. Several lines of evidence suggest a major, and probably short-lived, episode on southern Ruapehu about the same time: (1) the Murimoto debris avalanche and lahar deposits (Palmer and Neall, 1989) were emplaced about this time (9.8 ka); (2) the pyroclastic flow and tephra deposits of the Pourahu Formation (Donoghue et al., 1995a,b) were erupted at this time; (3) tephrochronology associated with the voluminous Rangatau flows to the southeast of the Turoa skifield (Fig. 1) suggest that these were emplaced at this time (R.B. Stewart, pers. commun.; Luther, 1999); and (4) partially consolidated fall deposits have been preserved in the proximal vent region (Lockett, 2001). These short pulses of greatly increased activity are outside the resolving power of our Ar–Ar method (typically ± 10 ka). Nevertheless, they serve as a reminder that future predictive models must consider possibilities of eruptions of short duration that are orders of magnitude greater than those of historic times.

Table 3 allows comparison with fluxes at other arc volcanoes. Our estimates for Ruapehu are larger than the maximum flux of 0.33 km^3 ka^{-1} for Volcan Tatara–San Pedro, in Chile (Singer et al., 1997), but within the ranges of 0.05–5.0 km^3 ka^{-1} for Mount Adams in the Cascades Ranges of the western USA (Hildreth and Lanphere, 1994); 1.2 km^3 ka^{-1} for Merapi, Indonesia (based on the 100-yr record between 1890–1992, Siswowidjoyo et al., 1995); 1.2–1.5 km^3 ka^{-1} for Tungurahua Stage III, Ecuador (Hall et al., 1999); and 1.2 km^3 ka^{-1} for Karymsky, Kamchatka (Braitseva and Melekestev, 1990). Significantly, these fluxes are all greater than the average present-day global fluxes for arc volcanoes of 0.2 km^3 ka^{-1} km^{-1} (Arculus, 1996) or 0.08 km^3 ka^{-1} km^{-1} (Cosca et al., 1998). However, these comparisons are tempered by differences in the calculation procedures, where fluxes for discrete volcanoes can be calculated with greater precision than global averages, that are averaged over a global arc length of $\sim 3.7 \times 10^4$ km (Arculus, 1999). Taken in a global context our calculated magma flux rates are high but comparable to many other arc volcanoes worldwide (Table 3). Moreover, perhaps this should not be surprising given that central TVZ has been identified as 'the most frequently active and productive rhyolitic system on Earth' (Houghton et al., 1995).

When viewed in comparison with fluxes for some basalt–basaltic andesite dominated arc volcanoes such as Fuji and Kluychevskoy (Table 3), which are comparable to oceanic island volcanoes such as Mauna Loa (30 km^3 ka^{-1}) (Lockwood and Lipman, 1987), the Tongariro Volcanic Centre fluxes are an order of magnitude less.

In summary, our calculations show average magma flux rates of $\sim$1.0 km^3 ka^{-1} for the andesite volcanoes of the Tongariro Volcanic Centre. During periods of peak magmatic activity, which may be of short duration (hundreds of years), this figure may vary by more than an order of magnitude, but typically by a factor of two times. These rates are high by comparison with global averages and confirm that the andesite volcanoes of the Tongariro Volcanic Centre are among the most productive on Earth.

6.3. *Relations between magma type and age*

In their study of Ruapehu lavas Graham and Hackett (1987) identified overall increases in $^{87}Sr/^{86}Sr$ with increasing SiO_2 content and with time. These variations were related to a progressive increase in AFC with time as a crustal magmatic plumbing system became established. Using a much larger and more comprehensive data base, Price et al. (1997, 2000) confirmed the general nature of these observations (Fig. 5a–d). In Fig. 5a,b the plots of K_2O content vs. SiO_2 and MgO

Table 3
Magmatic fluxes for global andesite volcanoes

Volcano	1000-yr magma flux (km^3 ka^{-1})	Source
Mount Adams, 1994, USA	0.05–5.0	Hildreth and Lanphere, 1994
Tatara San Pedro, Chile	0.06–0.29	Singer et al., 1997
Tungurahua, Ecuador	1.2–1.5	Hall et al., 1999
Merapi, Indonesia	1.2	Siswowidjoyo et al., 1995
Karymsky, 1990, Kamchatka	1.2	Braitseva and Melekestev, 1990
Shiveluch, Kamchatka	5.0	Crisp, 1984; Khrenov et al., 1991
Avachinsky, Kamchatka	1.7	Crisp, 1984
San Juan, Cerro Ato, Mexico	1.8	Luhr, 1999
Hakone, Japan	0.37	Crisp, 1984
Sakurajima, Japan	1.8	Crisp, 1984
Klyuchevskoy, Kamchatka	27.0	Crisp, 1984
	36.0	Kersting and Arculus, 1994
Fuji, Japan	5.0	Crisp, 1984
	8.0	Togashi et al., 1991

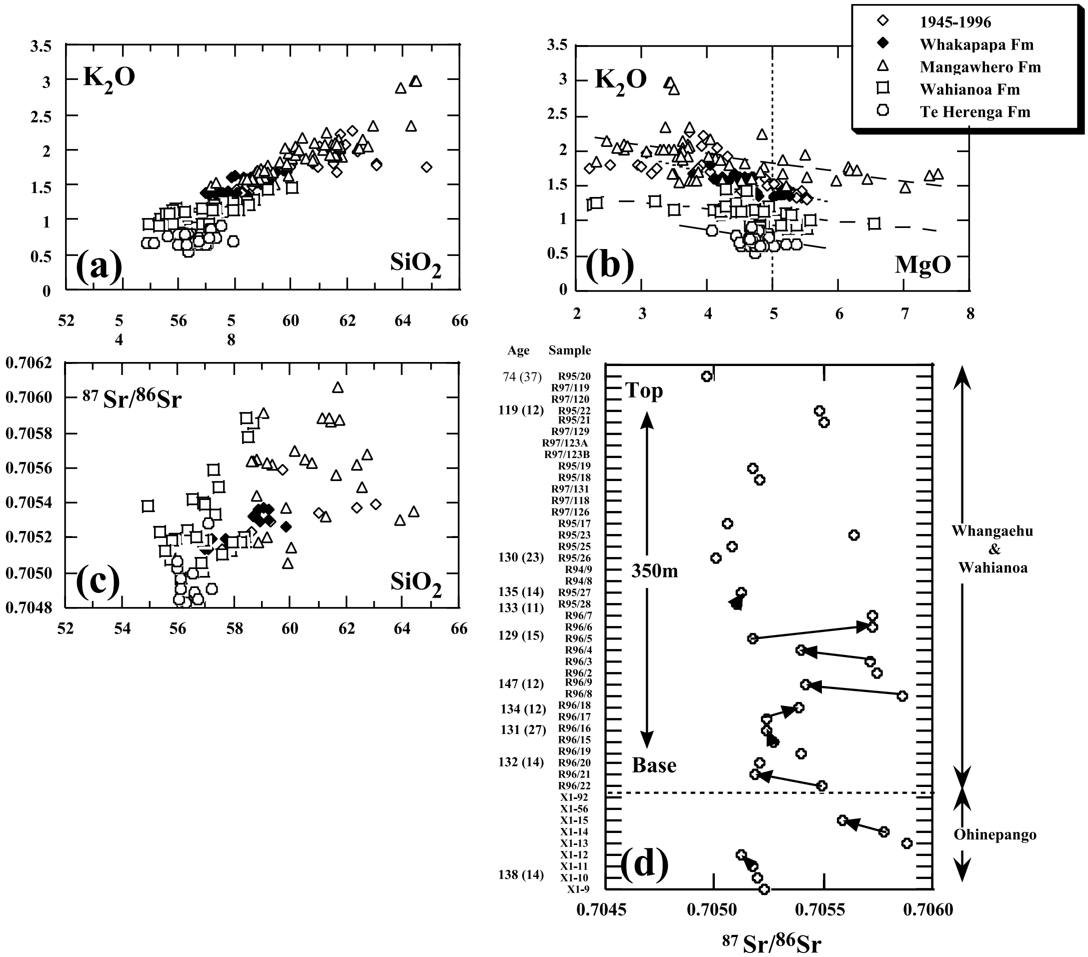

Fig. 5. Selected geochemical variation diagrams for lavas from Ruapehu. (a) K_2O abundance versus SiO_2 content. Note overlapping fields for Te Herenga, Wahianoa, Mangawhero, and Whakapapa Formation lavas. (b) K_2O vs. MgO showing progressive change of K_2O content at 5% MgO. Te Herenga lavas have the lowest K_5 and Mangawhero and Whakapapa the highest. (c) $^{87}Sr/^{86}Sr$ ratio vs. SiO_2 content. Positive correlation has been interpreted to indicate AFC (e.g. Graham and Hackett, 1987). Note strong scatter of data and non-systematic variation with time. (d) Variation of $^{87}Sr/^{86}Sr$ ratio with stratigraphic position in the lava flow sequence of the Waihinoa Formation. Samples from the Whangaehu, Wahianoa, and Ohinepango catchments are arranged vertically in stratigraphic order. Ages of samples (this study) are shown, with errors in parentheses.

abundances show the distinctive character of lavas from the main stratigraphic formations. The MgO vs. K_2O plot (Fig. 5b) shows four overlapping, but distinctive, groups with K_5 (K_2O content at 5% MgO) contents of 0.69, 1.12, 1.81 and 1.42%, respectively, for the Te Herenga, Waihianoa, Mangawhero and Whakapapa lavas. Importantly, Price et al. (1997, 2000) also noted that when geochemistry was studied at a flow by flow scale in time-related sequences, the picture of simple geochemical change with time became increasingly blurred (Fig. 5d). In detail, within a 300-m measured section of lavas from Whangaehu gorge, which is now precisely dated, Price et al. (1997, 2000) and Gamble et al. (1999) identified a series of hiatuses where abrupt reversals in chemical trends occurred. These were interpreted as replenishment events in an open, fractionating and mixing magmatic system. At nearby Tongariro volcano Hobden et al. (1999, 2002) identified similar non-systematic shifts in the chemical composition of lavas from Ngauruhoe and Red Crater cones. Similarly, Nakagawa et al. (1999) and Gamble et al. (1999) identified shifts in the compositions of the 1995–1996 and 1945–1996 eruptives from Ruapehu, which suggested chemical changes on time-scales in the order of hours to years.

Collectively, these data appear to support a model where temporally and chemically distinctive batches of melt aggregate and ascend from a mantle source and experience variable and non-systematic mixing and contamination events during transit through the lithosphere. At the surface this produces compositions which may be distinctive (e.g. with distinct K_5, Rb_5, $^{87}Sr/^{86}Sr_5$, etc.) for a given time window, but with non-systematic trends between successive eruptives. This process highlights the dynamic and continuously evolving nature of processes beneath arc volcanoes.

6.4. Implications for hazard assessment on Ruapehu volcano

Through an ever-widening use of the internet and non-technical reports, such as the IAVCEI Crisis Protocols Subcommittee (1999), the IAVCEI is endeavouring to promote the key role of communication between scientists, emergency coordinators, media and the general public. However, past experience tells us that the time intervals between volcanic eruptions are commonly much longer than the human lifespan so that within several generations communal memory and awareness decline and areas and regions that volcanologists consider to be high risk areas are reoccupied and, in some cases, intensively resettled. In New Zealand, the last andesitic volcanic eruptions of significance were the 1995–1996 and 1945 eruptions from Ruapehu and the 1949–1954 series of eruptions from Ngauruhoe. Prior to these events, the most significant eruption was the 1886 basaltic eruption of Mount Tarawera. Major eruptions such as Tarawera 1886 are now part of history and therefore relevant to past generations. Consequently, few people now living in New Zealand have any appreciation of the effects of a large-scale volcanic eruption. It is up to the scientists to work with the community to build rational models that provide accurate predictions and workable risk management plans. In evaluating the hazard and working towards a plan for hazard mitigation, volcanologists need precise information about magma flux rates, patterns of eruption style and explosivity, and factors such as physiography and hydrology.

From our detailed geochronology we now have reliable estimates for magma fluxes and eruption recurrence times for Ruapehu volcano. These show overall fluxes that are amongst the highest on Earth. Over the last two millenia activity has been characterised by more or less continuous small volume, phreatomagmatic, explosive eruptions (Donoghue et al., 1997) but past history also incorporates large volume pulses of short (<1 ka?) duration (e.g. Nairn et al., 1998; Nakagawa et al., 1998). From a hazard perspective we need to know what may happen in the future and how we can use past events to predict future activity. For Ruapehu, a key question is whether we can anticipate a future Waihianoa-scale or Pahoka–Mangamate-scale event.

A remarkable aspect of Ruapehu magma production is the consistency measured in the major time windows. On average, these are roughly half as much again as the average flux (0.6 $km^3\ ka^{-1}$). For example, the flux estimated for the Te Herenga Formation (0.93 $km^3\ ka^{-1}$) compares closely with the rate for the last 50 years integrated over 1 ka (0.86 $km^3\ ka^{-1}$). Also of interest is the periodicity of eruptions. For example, volcanoes such as Unzen (1990–1902), Pinatubo (1991) and Soufrière Hills (1997–present), which have erupted andesite–dacite magmas during their recent phases of activity, show eruption recurrence times in the order of hundreds of years (e.g. Unzen 198 yr, Pinatubo $\sim$500 yr; Soufrière Hills 350–400 yr). Petrologic studies of these volcanoes (e.g. Nakada et al., 1999; Pallister et al., 1992; Murphy et al., 2000) have, in each case, implicated injection of mafic magma as the trigger for eruptions. As such, they contrast with the andesite volcanoes of Tongariro Volcanic Centre, which, with recurrence times of 25–30 yr, are more or less continually active and would appear to involve more open magmatic systems where replenishment events involve reinjection of fresh andesite magma (Gamble et al., 1999). In Unzen, Pinatubo and Soufrière Hills volcanoes relatively long-lived crustal magma reservoirs are reactivated by injections of fresh mafic magma from deeper levels. At Ruapehu the system appears to have remained open with continuous fluxing of magma through the lithosphere. This model is supported by the geochemical data (Price et al., 1997, 2000, 2001; Gamble et al., 1999) and is also consistent with U–Th disequilibrium data (Hughes, 1999).

Our flux calculations and petrological observations lead us to conclude that the Ruapehu magmatic system has operated in much the same fashion throughout the volcano's life span. The last major event producing voluminous lava flow sequences occurred around 10 ka (Price et al., 2000). Our $^{40}Ar/^{39}Ar$ geochronology does not enable us to resolve events at time scales less than this interval; we have no means of determining, for example, how rapidly the Whakapapa Formation was emplaced. It is therefore reasonable to predict that eruptions at Ruapehu will continue in much the same mode as they have for the past 2–3 ka but the possibility of future Wahianoa- or Pahoka–Mangamate-scale events cannot be discounted. It is also worth noting that the style of eruptive activity over the past 2–3 ka has been dominated by the effects of Ruapehu's crater lake (Donoghue et al., 1997). The disappearance of the lake at some time in the future would substantially change the eruptive style. One might expect a switch from phreatic, surtseyan and phreatomagmatic tephra producing eruptions to lava flows and dome emplacement, bringing a new array of potential hazards.

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