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1 Revision 3

New petrological, geochemical and geochronological perspectives 2 on andesite-dacite magma genesis at Ruapehu volcano, New 3 Zealand 4 5 6 Chris E. Conway^{1,2}*, John A. Gamble^{1,3}, Colin J. N. Wilson¹, Graham S. Leonard⁴, Dougal 7 B. Townsend⁴, Andrew T. Calvert⁵ 8 9 10 ¹ School of Geography, Environment and Earth Sciences, Victoria University, PO Box 600, 11 12 Wellington 6140, New Zealand ² Department of Geology and Paleontology, National Museum of Nature and Science, 4-1-1 13 14 Amakubo, Tsukuba, Ibaraki 305-0005, Japan ³ School of Biological, Earth and Environmental Sciences, University College Cork, Ireland 15 ⁴ GNS Science, PO Box 30-368, Lower Hutt 6315, New Zealand 16 ⁵ US Geological Survey, 345 Middlefield Road, MS-937, Menlo Park, CA 94025, USA 17 18 19 20 *Email: chrisconway@kahaku.go.jp

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ABSTRACT

22 Time-composition relationships in eruptive sequences at composite volcanoes can 23 show how the ongoing intrusion of magmas progressively affects the lithosphere at 24 continental convergent margins. Here, new whole-rock and microanalytical major and trace 25 element data from andesite-dacite lava flows are integrated with previous studies and existing 26 isotopic data, and placed within the framework of a high-resolution chronostratigraphy for 27 Ruapehu volcano (southern Taupo Volcanic Zone, New Zealand). The geochemical evolution 28 of lavas erupted over the ~200 kyr lifetime of the exposed edifice reflects variable degrees of 29 fractionation and systematic changes in the type of crustal assimilation in the Ruapehu 30 magma system. Lavas erupted from $\sim 200-150$ ka have previously been distinguished from 31 those erupted <150 ka based on Sr-Nd isotopic characteristics, which indicate that the oldest 32 lavas were sourced from magmas that assimilated oceanic crust. Such source rocks underlie 33 the regionally widespread Mesozoic meta-sedimentary greywacke-argillite basement, which 34 was conversely assimilated by <150 ka magmas. New results from this work reveal that since 35 150 ka, an upper limit of magma differentiation occurred from \sim 50–35 ka. High K₂O (\sim 6 wt. 36 %) and Rb contents (~270 ppm) in melt inclusions, interstitial glass, and glass from *in situ* 37 quenched melts of partially fused crustal xenoliths are reported for andesite-dacite lavas 38 erupted during this period. In addition to crystal fractionation, selective partial melting and 39 assimilation of K-and Rb-rich mineral phases (e.g. biotite, K-feldspar) that are significant 40 components of the meta-sedimentary basement rocks is inferred to explain these geochemical 41 characteristics. These processes coincided also with the effusion of high-MgO andesite-dacite 42 lavas that display petrological evidence for mixing between andesite-dacite and more mafic 43 magmas. An influx of hotter mafic magma into the system explains why the extent of crustal 44 assimilation recorded by Ruapehu lavas peaked during the ~50–35 ka eruptive period. From 45 26 ka to the present, andesite lavas have reverted to more mafic compositions with less

46	potassic melt inclusion and whole-rock compositions when compared to the \sim 50–35 ka lavas.
47	We suggest that the younger lavas assimilated less-enriched melts because fertile phases had
48	been preferentially extracted from the crustal column during earlier magmatism. This
49	scenario of bottom-up heating of the lithosphere and exhaustion of fertile phases due to the
50	progressive intrusion of magma explains the geochemical evolution of Ruapehu lavas. This
51	model may be applicable to other long-lived composite volcanoes of the circum-Pacific
52	continental arcs.
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54	Keywords: Ruapehu; andesite; dacite; petrogenesis; arc magma; crustal contamination; high-
55	Mg andesite
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pathways of differentiation in the crust (e.g. Hobden et al. 1999; Gamble et al. 2003; Frey et
al. 2004; Eichelberger et al. 2006; Sisson et al. 2013).

72 Beneath arc volcanoes, the processes of crustal anatexis, magma mixing, and crystal 73 fractionation are the primary agents of magma differentiation (e.g. Price et al. 2012; Kent 74 2013; Lee and Bachmann 2014). Two commonly documented signatures of intermediate arc 75 magmas that arise from these processes are: (a) the generation and entrapment of melts that 76 are substantially more silicic than their host whole-rock compositions (see compilations by 77 Reubi and Blundy 2009; Kent 2013), and (b) disequilibrium petrographic textures that reflect 78 mixing between felsic and mafic components and their crystal cargoes, together with 79 evolving intensive parameters such as pressure, temperature and degassing (e.g. Clynne 1999; 80 Dungan and Davidson 2004; Kent et al. 2010; Koleszar et al. 2012). Because the felsic and 81 mafic end-members seldom erupt independently at composite volcanoes, their cryptic 82 existence within crustal magma systems is recorded within mineral-hosted melt inclusions 83 and interstitial glass (e.g. Reubi and Blundy 2008) and the crystal cargo of the derivative 84 intermediate composition magmas (e.g. Eichelberger 1975; Streck et al. 2007). Thus, the 85 nature and origin of felsic and mafic sources, and the processes by which they interact, may 86 be elucidated by microanalytical investigations (Davidson et al. 2007). Constraining these 87 magma sources and the subsequent mixing parameters and processes is critical to 88 understanding the assembly of arc magmas.

89 Compositions of eruptive products from arc volcanoes record not only snapshots of 90 crustally dominated magma processes, but also the cumulative effect of progressive heating 91 and "conditioning" of the lower, middle, and upper crust in response to the magmas that have 92 passed through these regions. How magmatic sources and processes vary throughout time and 93 what controls those variations represent fundamental questions for understanding 94 petrogenesis at and the eruption potential of arc volcanoes. Critical to addressing these

95	questions is the collection of high-resolution eruptive histories for arc volcanoes that are
96	constrained by field studies and radiometric age data so that eruptive products can be
97	accurately sequenced (e.g. Dungan et al. 2001; Frey et al. 2004). Here, we apply this
98	approach to Ruapehu volcano in the southern Taupo Volcanic Zone (TVZ; New Zealand).
99	We combine a revised chronostratigraphy of Ruapehu volcano (Conway et al. 2016) with
100	detailed geochemical sequencing of lava flow compositions in order to investigate the genesis
101	of and esite-dacite magmas at this volcano throughout the \sim 200 kyr lifetime of the exposed
102	edifice.

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GEOLOGICAL SETTING

105 The Taupo Volcanic Zone (TVZ; Fig. 1) is a zone of continental magmatism in the 106 North Island of New Zealand related to the subduction of the Pacific Plate beneath the 107 Australian Plate (Cole 1990). The basement geology of the TVZ comprises greywacke-108 argillite meta-sedimentary rocks of the Torlesse Terrane, which are exposed to the east of the 109 volcanic area (Adams et al. 2009; Fig. 1). Waipapa Terrane meta-sedimentary rocks crop out 110 to the west of the TVZ (Fig. 1), but are not considered to play a significant role in TVZ 111 magmatism (e.g. Price et al. 2015). Caldera-forming rhyolitic eruptions have dominated the 112 style of volcanic activity within the TVZ for 2 Myr, however, the northern and southern 113 extents of the rifted arc are marked by the active andesite-dacite stratovolcanoes of White 114 Island and Ruapehu (Wilson et al. 1995; Fig. 1). 115 Ruapehu volcano is a $\sim 150 \text{ km}^3$ complex composite edifice primarily composed of

115 Ruapehu volcano is a ~150 km² complex composite edifice primarily composed of 116 stacked sequences of lava flows and associated breccias, intercalated with till and subsidiary 117 volumes of proximal pyroclastic flow and fall deposits (Hackett and Houghton 1989). The 118 exposed edifice has a lifetime of ~200 kyr, although volcanism in the area may have started

119 as early as ~340 ka (Gamble et al. 2003; Tost and Cronin 2015). Ruapehu lavas were grouped 120 by Hackett (1985) into four formations that represent the main phases of edifice growth 121 (Table 1): Te Herenga (200–150 ka), Wahianoa (160–115 ka), Mangawhero (50–15 ka), and 122 Whakapapa (<15 ka). The absolute ages and durations of these formations were first 123 constrained by Gamble et al. (2003), who primarily concentrated on the ~300 m-thick stack 124 of Wahianoa Formation lavas exposed on the southeast flank of Ruapehu. Their data were recently combined with new ${}^{40}\text{Ar}/{}^{39}\text{Ar}$ ages by Conway et al. (2016) to provide a new 125 126 geological map (summarised in Fig. 2) and to outline the revised edifice construction history 127 summarised here. Sub-glacial to ice-marginal effusive eruption of medium-K basalticandesites and andesites constructed the northern portion of the exposed edifice between ~200 128 129 and 150 ka (Te Herenga), and a wide southeast planèze as well as parts of the northern, 130 eastern and western flanks between ~160 and 80 ka (Wahianoa). No lava flows on the edifice have ages in the range of 80–50 ka, however, ${}^{40}\text{Ar}/{}^{39}\text{Ar}$ eruption ages measured for lava clasts 131 132 within debris flow deposits indicate that there were some effusive eruptions during this 133 period (Tost and Cronin 2015). The lack of dated lavas on the edifice for this period is likely 134 to be the result of erosion and burial of lavas and/or syn-eruptive glacial conveyance of lava 135 flows as debris to the ring-plain (Conway et al. 2016). In addition to bridging the apparent hiatus and extending the maximum age for volcanism at Ruapehu, other ⁴⁰Ar/³⁹Ar ages for 136 137 clasts within distal debris flow deposits generally overlap with the timing of effusive volcanism recorded by in situ lavas of the Te Herenga and Wahianoa formations (Tost and 138 139 Cronin 2015). From ~50–15 ka edifice growth occurred via effusive eruptions onto the 140 glaciated flanks of the volcano, resulting in construction of ice-bounded planèzes and ridges 141 (Mangawhero). Large-scale retreat of flank glaciers since ~15 ka has resulted in intra-valley 142 lava flow emplacement at elevations below ~1500 m on the edifice. Conway et al. (2016) reported the first ⁴⁰Ar/³⁹Ar eruption ages for lava flows of the post-glacial Whakapapa 143

144 Formation in addition to 25 ages for lavas of the Mangawhero Formation, which together 145 provide a comprehensive volcanic sequence to investigate time-composition relationships. 146 The greatest volume of lava flows on Ruapehu are plagioclase and pyroxene phyric 147 andesites, however, basaltic-andesites and dacites have also been erupted at Ruapehu. 148 Complex crystal and lithic cargoes are found throughout the eruptive products, which have 149 been described in detail by Clark (1960), Cole (1978), Graham (1987), and Graham and Hackett (1987). The last authors defined six petrographically and geochemically distinct lava 150 151 types with corresponding models of magma formation. These models accounted for the crust-152 like geochemical and isotopic composition of Ruapehu lavas by invoking partial melting and 153 assimilation of mid-upper crustal greywacke-argillite basement during magma genesis (see 154 also Graham 1987). 155 The overall development of the magma system beneath Ruapehu was later explored 156 within stratigraphic contexts by Donoghue et al. (1995), Gamble et al. (1999, 2003), Waight 157 et al. (1999, 2017) and Price et al. (2005). A model for the polybaric generation of derivative 158 magmas by independent episodes of melting, mixing, assimilation and fractionation within an 159 open magmatic system was generated from these studies (Price et al. 2012 for summary). 160 This model allows for the non-systematic geochemical and isotopic variations observed over 161 timescales ranging from short-lived explosive episodes and cycles (1–50 yr; Gamble et al. 162 1999; Auer et al. 2013; Kilgour et al. 2013) to successive lava flow emplacement phases (1-163 10 kyr; Waight et al. 1999; Gamble et al. 2003). Over the 200 kyr lifetime of the exposed 164 Ruapehu edifice, however, a general progression towards more evolved compositions with 165 time was proposed based on whole-rock geochemical variations using the available 166 stratigraphic constraints (Graham and Hackett 1987; Price et al. 2012). In addition, the oldest exposed lavas (Te Herenga Formation) have distinct ¹⁴³Nd/¹⁴⁴Nd–⁸⁷Sr/⁸⁶Sr characteristics that 167 168 set them apart from all subsequent lavas (Price et al. 2005). On this basis, Graham et al.

(1990) and Price et al. (2005, 2012) argued that Te Herenga Formation lavas were
contaminated by Mesozoic oceanic basaltic rocks of the lower crust, whereas post-Te
Herenga lavas assimilated partial melts of the basement greywacke-argillite sequence in the
mid-upper crust.

Within the context of prior studies there are three key topics that we address here:

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174 1. Time-composition relationships. Incomplete sampling of the edifice and a lack of 175 chronological control hampered previous studies of the geochemical evolution of the 176 Ruapehu magma system. Complex stratigraphic relationships as a result of lava-ice 177 interaction (Conway et al. 2015) and sparse radiometric eruption age constraints previously 178 precluded accurate definition of the relative and absolute ages for lava flows of the 179 Mangawhero and Whakapapa formations that form much of the edifice. Combined with previous dating constraints by Gamble et al. (2003), new ⁴⁰Ar/³⁹Ar ages for Ruapehu lava 180 181 flows have provided a high-resolution framework with which to interrogate geochemical data (Conway et al. 2016). By using geochemical data acquired for the same samples that have 182 183 been directly dated and collected within chronostratigraphic constraints, we are able to assess 184 the long-term evolution of magmatism at Ruapehu within closely defined spatial and 185 temporal contexts, as has been done at other arc stratovolcanoes globally (e.g. Dungan et al. 186 2001; Hora et al. 2007; Jicha et al. 2012). Data from lava clasts in distal deposits and tephra 187 in pyroclastic deposits have been integrated into this study where it is appropriate and 188 possible to do so.

2. Nature and origin of silicic melts. Melt inclusion and matrix glass data
 compositions ranging from dacite to rhyolite have previously been described from Wahianoa
 Formation lavas, and Whakapapa Formation and historical lavas and pyroclasts (Donoghue et al. 1995; Gamble et al. 1999; Price et al. 2005; Kilgour et al. 2013; Pardo et al. 2014). In this

193	contribution we present the first glass geochemistry data for Te Herenga Formation eruptive
194	products and a number of spatially distinct packages identified within the Mangawhero
195	Formation. These new data contribute to models of magma genesis at Ruapehu.
196	3. Felsic-mafic interactions. In light of the abundance of rhyolitic melt compositions,
197	is there evidence for mafic recharge in the generation of intermediate (andesite) magmas, as
198	postulated from local (Donoghue et al. 1995; Nakagawa et al. 2002) and global (e.g. Kent et
199	al. 2010) studies? Is this a common process involved in the generation of magnesian
200	andesites and continental crust at volcanic arcs globally? Petrological and geochemical
201	characteristics of a suite of high-MgO lavas are presented here to explore these concepts.
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203

SAMPLES AND METHODS

204 Samples used in this study were collected from lava flows on Ruapehu between 2012 205 and 2015. A full sample list is available in the Electronic Supplementary Material. 206 Petrographic descriptions for Ruapehu lava flows have been presented by Cole (1978), 207 Graham (1987), Graham and Hackett (1987), Graham et al. (1990), and Price et al. (2012). 208 Brief descriptions are provided for the sample suite used here for completeness and because 209 new portions of the edifice have been sampled. Ruapehu lava flows are in general moderately 210 to highly porphyritic, with only rare examples with <10 % phenocrysts (term used herein to 211 define macrocrystals with sizes $>100 \mu m$, but generally in the range of 0.3–1.0 mm). 212 Plagioclase phenocrysts occur in all samples examined and most commonly have sieved 213 cores with complexly zoned rims. Orthopyroxene, clinopyroxene and olivine phenocrysts are 214 present throughout Ruapehu lavas in varying proportions. For the basaltic-andesite lavas of 215 the Te Herenga and Whakapapa formations, clinopyroxene and orthopyroxene are equally 216 abundant. Orthopyroxene is the major ferromagnesian phase in lavas of the Mangawhero

217 Formation. Amphibole occurs within Mangawhero dacite lavas studied here only as rare 218 pseudomorphs that are replaced by symplectite assemblages of plagioclase, pyroxene, oxides 219 and glass. The groundmasses of Ruapehu lavas typically consist of felted aggregates of 220 microlites (i.e. <100 µm) of plagioclase and orthopyroxene with volumetrically minor but 221 ubiquitous oxide phases and small patches of glass. Magnetite is the dominant oxide phase 222 present throughout the groundmass of Ruapehu lavas, although rare ilmenite and spinel are 223 also present. Patches of glass in microlite-free areas \sim 5–10 µm-wide are present only in 224 samples acquired from the margins of rapidly chilled ice-bounded Mangawhero Formation 225 flows and some glassy Whakapapa Formation flows.

226 Meta-sedimentary and meta-igneous xenoliths are present in all collected samples and 227 are particularly common in lavas of the Mangawhero and Whakapapa formations. Meta-228 sedimentary types are typically fine-grained assemblages of plagioclase, orthopyroxene and 229 magnetite, which are likely to be metamorphic equivalents (or refractory residua) of the 230 greywacke crust beneath Ruapehu (Price et al. 2005). Some meta-sedimentary fragments also 231 contain glass and biotite (e.g. Graham 1987). Meta-igneous granulite xenoliths are composed of plagioclase, orthopyroxene and ilmenite with minor clinopyroxene and olivine and display 232 233 relict igneous or cumulate textures (Graham et al. 1990). A summary of the mineral 234 chemistry data acquired for this study is presented in Table 2. Typical zoning features of the 235 major mineral phases are displayed in Fig. 3. Further petrographic descriptions and full 236 mineral chemistry data are provided in the Electronic Supplementary Material.

Fresh interiors of samples were acquired by using a diamond saw to cut away any remaining weathered surfaces. Samples were crushed and then milled using an agate or tungsten-carbide ring mill. Major element compositions for 238 whole-rock powder samples were determined using an ARL[®] 8420+ dual goniometer wavelength dispersive X-ray fluorescence (XRF) spectrometer at the Open University, United Kingdom. These XRF data

242 have been presented by Conway et al. (2016) for the purposes of defining chronostratigraphic 243 eruptive packages, and are used here to investigate magmatic processes. A subset of 73 244 whole-rock powders used for major element analysis was selected for trace element analysis 245 and prepared at Victoria University of Wellington (VUW). Trace element analysis was 246 carried out by solution inductively coupled plasma mass spectrometry (ICP-MS), on a 247 Thermo-Fisher Element2 sector-field ICP-MS equipped with an ESCI auto sampler at VUW 248 using methods similar to those of Eggins et al. (1997). Major element compositions for 249 minerals and glass in 15 lava samples were analysed by electron probe microanalysis 250 (EPMA) using the JEOL JXA 8230 instrument at VUW. Thin sections were prepared from 251 lava samples and carbon coated before analysis. Analysis of crystal phases was performed 252 using a focussed electron beam (~1um-wide) at a current of 12 nA and accelerating voltage 253 of 15 kV. The beam was defocussed to a width of 5 µm during analysis of melt inclusion and 254 groundmass glasses. Due to the commonly small sizes of melt inclusions and interstitial glass 255 patches, a 5 µm spot was used (rather than 10 µm). Only melt inclusions in pyroxene hosts 256 were analysed; plagioclase-hosted inclusions were not well-enough preserved for viable 257 analysis. The current was reduced to 8 nA and counting times for Na were reduced to 15 258 seconds on the peak and 10 seconds for the background to minimise potential 259 devolatilisation. Laser ablation ICP-MS analyses of interstitial glass and xenolith-hosted 260 glass in dacite lava sample CC415 were carried out at VUW using a Resonetics 261 RESOlutionS155-SE laser ablation system equipped with an ATL ArF Excimer laser 262 (wavelength 193 µm), coupled to an Agilent 7500CS ICP-MS. Additional details of sample 263 preparation and analysis can be found in the Electronic Supplementary Material and in 264 Conway (2016).

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266

RESULTS AND DISCUSSION

267	This section presents new geochemical data that are placed within the context of a
268	revised chronostratigraphy and existing data for Ruapehu, as well as regional and global
269	studies of arc volcanism. As such, the results are presented together with a discussion of their
270	relevance. New whole-rock major and trace element data are presented in Figs. 4–7, and
271	existing whole-rock isotopic data from Price et al. (2012) are displayed in Fig. 8. New glass
272	major element data are presented in Fig. 9 and compared to existing data from Ruapehu and
273	Taupo volcanoes. The extended sample coverage and improved chronological sequencing of
274	lava flows presented here makes it possible to advance more detailed models of the evolution
275	of the Ruapehu magmatic system, and bring to light several key questions for global studies
276	of arc volcanism, including:
277	• What are the scales of temporal and spatial variability in magma composition at
278	continental composite volcanoes?
279	• Do long-term time-composition trends recorded in arc lavas reflect density filtering of
280	magmas via edifice loading and sector collapse?
281	• What is the nature and origin of felsic (rhyolitic) melts that are captured in
282	intermediate arc lavas?
283	• What effect does the progressive throughput of magma have on the crustal
284	contamination of continental arc magmas?
285	• Are high-MgO and esites-dacites erupted at continental arc volcanoes the products of
286	mantle melting processes, or magma mixing during transit through the crust?
287	

288 Compositional diversity of lavas in time and space at Ruapehu

289 Composite volcanoes are the surficial structures of polygenetic magmatism (Davidson 290 and de Silva 2000). Volcanic maps and stratigraphies that are coupled with geochemical data 291 can constrain the temporal and spatial scales of compositional heterogeneity within 292 polygenetic magma systems that feed composite volcanoes (e.g. Frey et al. 2004; Hora et al. 293 2007). A summary of whole-rock compositions for Ruapehu volcano lava flows was 294 presented by Price et al. (2012), however, the major and trace element variations between and 295 within formations are re-evaluated here in light of the extended sampling coverage and 296 improved geochronological control. Compositional heterogeneity within the Ruapehu magma 297 plumbing system is investigated in this section, with a particular focus on the time window 298 between 50 and 35 ka. Ruapehu lava whole-rock major and trace element bivariate diagrams 299 are shown in Fig. 4, and a representative suite of analyses is presented in Table 3. A complete 300 list of whole-rock major and trace element data is available in the Electronic Supplementary 301 Material.

302 Te Herenga lavas exhibit a relatively narrow compositional range for all major and 303 trace elements that is generally distinct from younger lavas, excepting some overlap with the 304 Wahianoa Formation (Fig. 4). The most mafic lava flow sampled on Ruapehu volcano 305 (CC348: 53.8 wt. % SiO₂, 6.75 wt. % MgO) was erupted at 186.2 ± 6.8 ka (2 s.d: Conway et 306 al. 2016). A notable characteristic of Te Herenga lavas is their low K₂O contents (Fig. 4) 307 when compared to younger lavas, and incompatible trace element contents are essentially 308 constant over their ~3 wt. % SiO₂ range (Fig. 4). Wahianoa Formation lavas are basaltic-309 andesites and andesites that exhibit slightly elevated SiO₂, and incompatible trace element 310 contents when compared with Te Herenga Formation lavas (Fig. 4). Previously un-sampled 311 lavas erupted from ~100-80 ka have major element compositions that generally match ~166-312 100 ka lavas of the Wahianoa Formation. Within the Wahianoa Formation, a distinct group of 313 flows is identified by low-MgO and high-Al₂O₃ contents consistent with higher modal

abundances of plagioclase relative to pyroxene (Graham and Hackett 1987). Mangawhero Formation lavas are separated on the basis of the age data into early (50–35 ka), middle (35– 26 ka) and late (26–15 ka) divisions in this contribution. Lavas of the ~50–35 ka suite are compositionally diverse, covering an 8.2 wt. % range in SiO₂ from basaltic-andesite to dacite. These flows show strong positive trends for incompatible trace element abundances and a distinct trend toward higher K₂O and Rb contents when plotted against SiO₂ and compared to older and younger lavas (Fig. 4).

321 Despite the broadly linear relationship between SiO₂ and incompatible trace elements, 322 lavas erupted between 50 and 35 ka can be further grouped into high- and low-MgO types. 323 Reasons for the high-MgO compositions are discussed in a later section of the discussion, but 324 their features are described here. The high-MgO trend is defined by lavas of the 325 Mangaehuehu and Te Kohatu eruptive packages, and low-MgO andesites and dacites are grouped within the Ngahuinga and Mangaturuturu eruptive packages (Fig. 6). The Mananui 326 327 dacite package, which consists of a spatially isolated flow on the northern flank of Ruapehu, has an intermediate MgO content. High-MgO and esites (Mg $\#_{60-69}$, where Mg# = 100Mg/[Mg 328 329 + Fe]) and dacites (Mg $\#_{54-57}$) also exhibit relatively high Ni and low Al₂O₃ contents at 330 equivalent SiO₂ values when compared to all other eruptive products from Ruapehu volcano 331 (Figs. 4, 5). The timing of eruption for these high-MgO andesites, which span a 5.2 wt. % 332 range in SiO₂ contents, is constrained by maximum and minimum ages of 45.4 ± 2.0 ka and 333 41.8 ± 1.8 ka (Conway et al. 2016). This is a noteworthy feature when considering that a 334 comparable range of \sim 4.8 wt. % SiO₂ is defined by the \sim 300 m-thick sequence of Wahianoa 335 Formation lavas exposed on southeast Ruapehu, which reflects a much longer period of 336 volcanism from ~160–115 ka (Gamble et al. 2003). The trends defined by our new data imply 337 that at least four geochemically distinctive magma types (high-MgO andesite; high-MgO 338 dacite; mid-MgO dacite; low-MgO andesite and/or low-MgO dacite) were feeding spatially

339	dispersed lava flow packages with volumes of several km ³ to the south, west and north flanks
340	of Ruapehu during the period from ~50–35 ka (Conway et al. 2016). This implies that magma
341	batches with volumes substantially greater than those tapped by historical pyroclastic
342	eruptions were able to remain compositionally isolated in the crust beneath Ruapehu (Gamble
343	et al. 1999; Kilgour et al. 2013).

344 Lava flows emplaced on the western flank of Ruapehu volcano from ~35 to 26 ka 345 exhibit SiO₂, K₂O and trace element contents that are intermediate between the low-MgO 346 andesites and dacites erupted at 50-35 ka. Late Mangawhero Formation lava flows erupted at 347 \sim 26–15 ka are andesites and rare dacites. These flows define a less steep trend for K₂O and incompatible trace elements (e.g. Rb, Zr) plotted against SiO₂ when compared to middle 348 349 (~35–26 ka) and early (~50–35 ka) Mangawhero Formation lavas (Fig. 4). The major and 350 trace element compositions of $\sim 26-15$ ka and esites generally overlap with younger flows of 351 the Whakapapa Formation (Fig. 4), which are andesites and basaltic-andesites. Numerous 352 eruptive packages within the Whakapapa Formation that were sourced from discrete vents 353 and summit cones have been defined (Price et al. 2012; Conway et al. 2016). Taken together, 354 Whakapapa Formation lavas define a SiO₂-K₂O trend contiguous with \sim 26-15 ka 355 Mangawhero Formation lavas (Fig. 4) and broadly overlap with post-glacial Plinian eruptive 356 deposits (see Pardo et al. 2014; Conway et al. 2016).

Previously published bivariate major element plots showed a scatter of overlapping data for lavas of variable age (e.g. Fig. 10 in Price et al. 2012), however, the data presented here allows some key discriminations to be made between lavas of different ages. Pre-80 ka lavas do not show strong linear trends towards a unique end-member, but are tightly clustered around relatively low SiO₂ and K₂O values (Fig. 4). Lavas that were erupted between ~50 and 26 ka and <26 ka define distinct SiO₂-K₂O compositional trends: Te Kohatu package dacite (sample CC508; 37.6 \pm 1.4 ka) forms a silicic end-member for 50–26 ka lava flows,

364	whereas Whakapapaiti package member dacite (sample CC513; 25.7 ± 3.8 ka) forms a silicic
365	end-member for <26 ka lavas (Fig. 4). Although the preserved portion of the Whakapapaiti
366	package dacite flows is volumetrically minor (see Fig. 10 in Conway et al. 2016), their
367	eruptions delineate a change in the nature of magma differentiation after ~ 26 ka, which is
368	addressed in the following sections.

369

370 **Potential role of density filtering of magma compositions**

371 The eruptive products of composite volcanoes can cover the compositional spectrum 372 from basalt to rhyolite. Although time-sequenced trends generally do not follow unique 373 pathways of progressive evolution toward the silicic end of the spectrum (cf. Eichelberger et 374 al. 2006), a common concept is that magma systems evolve toward producing more silicic 375 eruptive products over long time periods (>100 kyr). This idea is founded on the association 376 of non-evolved (mafic) eruptive products with stratigraphically old portions of volcanic 377 edifices, and the eruption of more evolved lavas and, in some cases, the incidence of climactic explosive eruptions of silicic magma later in the history of composite volcanoes 378 379 (e.g. Bacon and Lanphere 2006; Hora et al. 2007; Singer et al. 2008; Escobar-Wolf et al. 380 2010). Throughout the cycle of volcanism represented by the eruptive products, the 381 corresponding development of evolved, relatively low-density intrusive complexes in the 382 crust will impede higher-density, more mafic magmas from reaching the surface. This 383 provides a feedback loop that may explain general trends toward more evolved magma 384 compositions over time at some volcanoes (e.g. Bacon and Lanphere 2006). An alternative 385 consideration is that the load imparted on a magma source region during the construction of a surficial volcanic edifice allows only the eruption of progressively more buoyant magmas as 386 387 the load increases with cumulative growth of the volcano (Pinel and Jaupart 2000; Hora et al. 388 2007). With time the density of magmas will be lowered via crystal fractionation, enabling

389 them to overcome the critical density threshold imposed by loading of the edifice in order to 390 ascend and ultimately erupt (Pinel and Jaupart 2000). Conversely, unloading of an edifice via 391 deglaciation, erosion or sector collapse will reduce the confining pressure on magma reservoirs and enable relatively dense magmas to ascend to the surface (Pinel and Jaupart 392 393 2005; Rawson et al. 2016). In this way, a composite volcano acts as a density filter for 394 eruptible magma compositions. The time-sequenced variation of Ruapehu lava compositions provides a type study for 395 396 assessing whether density filtering affected erupted magma compositions, by using time-397 composition plots (Fig. 6) and multi-element diagrams (Fig. 7) for illustration. For the latter, 398 whole-rock trace element contents have been normalised to values for average continental 399 crust (Rudnick and Gao 2003) in order to reflect the extent of crustal contamination. Existing 400 isotope data for Ruapehu lavas from Price et al. (2012) are also reviewed in this section (Fig. 401 8). Ruapehu lavas display a general increase in SiO₂, K₂O, and the incompatible trace 402 element ratio Rb/Zr from ~200 to 50 ka (i.e. lavas of the Te Herenga and Wahianoa 403 formations; Figs. 6, 7). A key difference between lavas of the Te Herenga and Wahianoa formations is their Sr-Nd isotopic characteristics (Price et al. 2005, 2012). ⁸⁷Sr/⁸⁶Sr isotopic 404 405 compositions of Wahianoa Formation lavas overlap with the most radiogenic Te Herenga Formation samples (~ 0.7050), however they have distinctly lower ¹⁴³Nd/¹⁴⁴Nd ratios (Fig. 8). 406 407 Eruptive compositions show extreme diversity between ~50 and 35 ka (Fig. 6). Lava 408 compositions range from basaltic-andesite to dacite during this period, and display a wide 409 range in incompatible trace element contents, which extend to the most continental crust-like values for Ruapehu samples (Figs. 6, 7). Similarly, ⁸⁷Sr/⁸⁶Sr values are highest in dacites of 410 411 this period (Fig. 8). High Rb/Zr values for these dacites likely indicate addition to derivative 412 magmas of Rb from a potassic phase because Zr is unlikely to be fractionated by the mineral 413 assemblage. Lavas that have erupted since ~35 ka, however, define a trend opposite to that of

414	the earlier volcanic history: ~35–26 ka lava flows have a relatively narrow range of
415	compositions intermediate within the range defined by \sim 50–35 ka lavas, and younger flows
416	(<26 ka) exhibit progressively less silicic compositions with time (Fig. 6). Trends for K_2O
417	contents and Rb/Zr values display the same decreasing pattern as SiO_2 for post-26 ka lavas,
418	which also generally have lower 87 Sr/ 86 Sr values than 50–35 ka lavas (Fig. 6, 8).
419	In light of the hypothesised consequences of density filtering of magmas at growing
420	volcanic edifices, three observations within the geochemical evolution of Ruapehu lavas are
421	significant. First, the effusion of dacite following >150 kyr of edifice construction by
422	eruption of basaltic-andesites and andesites of the Te Herenga and Wahianoa formations.
423	Second, the reversion after ~ 18 kyr to more mafic lava compositions, which broadly overlaps
424	with the timing of deglaciation of the edifice, as well as Holocene sector collapse events
425	(Eaves et al. 2015; Conway et al. 2016). Third, more evolved samples generally have more
426	crust-like ⁸⁷ Sr/ ⁸⁶ Sr values.
427	With respect to the first observation, the 50-35 ka effusive episode occurred after the
428	construction of a composite edifice comprising the Te Herenga and Wahianoa formations that

429 contributed $\geq 110 \text{ km}^3$ of total eruptive products to the surface above the Ruapehu magmatic

430 system, during which time lava compositions generally became gradually more evolved (Fig.

431 6; Conway et al. 2016). Mangawhero Formation dacite lavas were erupted at ~40 ka, possibly

432 indicating that lower density magmas were able to overcome the critical density threshold

433 imposed by loading of the edifice in order to ascend and ultimately erupt. However, basaltic-

434 and esites were also erupted at the beginning of the 50-35 ka eruptive period with SiO₂

435 contents comparable to pre-80 ka lavas. The eruption of basaltic-andesite thus contradicts the

436 premise that a density filter biased the eruptible magma compositions at this time.

437	With respect to the second observation, there is a clear trend toward more mafic lavas
438	erupted since ~26 ka, and Holocene lava flows are and esites to basaltic-and esites with SiO_2
439	contents \leq 59 wt. % (excepting one late Holocene flow with 61.5 wt. % SiO ₂). Although not
440	exactly coincident with the onset of this compositional change, wholesale glacial retreat on
441	Ruapehu from ~18 ka is inferred from regional paleoclimate records (Newnham et al. 2003),
442	and corroborated by local moraine geochronology (Eaves 2015) and lava-ice interaction
443	features (Conway et al. 2015). Furthermore, sector collapse of the northwest summit and
444	upper flank at ~10.5 ka (Palmer and Neall 1989; Eaves et al. 2015) and the southeast summit
445	at ~4.6 ka (Donoghue and Neall 2001) would have also contributed to a reduction of the
446	vertical stress on the sub-volcanic system. The models of Pinel and Jaupart (2000, 2005)
447	suggest that the reduction of the confining pressure on the magma reservoirs would have
448	enabled relatively dense magmas to be erupted at Ruapehu since 18 ka. However, the loss of
449	ice cover and the volume of material removed from the edifice during the sector collapse
450	events ($\sim 1 \text{ km}^3$) is minor when compared to the total edifice volume ($\sim 150 \text{ km}^3$).

451 With respect to the third observation, the matching trends for the different elements 452 suggest that some common process, or combination of processes, have controlled the 453 compositional variation through time (Fig. 6). The general trend of more evolved lavas tending to have higher ⁸⁷Sr/⁸⁶Sr values (Fig. 8) indicates that assimilation of basement meta-454 455 sedimentary crust had a primary role, along with crystal fractionation, in the compositional 456 development of post-Te Herenga Formation magmas (Graham and Hackett 1987; Price et al. 457 2012). Crust-normalised multielement diagrams also reflect this trend (Fig. 7). These 458 relationships do not preclude the influence of a density filter, but are interpreted here to 459 reflect a coupling between magma diversity and crustal assimilation, which relates more to 460 the thermal condition of the lithosphere than the load imposed by construction (or 461 destruction) of the volcanic edifice. Flux of melt from the mantle and tectonic activity are

462	also key controls of the timescales of magma generation, residence and eruption (e.g.
463	Schmidt and Grunder 2009). Data are not available to quantify the effect of these factors for
464	this study, however, because (1) past mantle-derived melt flux rates cannot be constrained by
465	surficial volcanic records, and (2) fault slip rate data are unable to resolve volcano-tectonic
466	processes on timescales of \sim 5–10 kyr in the Ruapehu Graben (Rowland et al. 2010; Gómez-
467	Vasconcelos et al. 2017).
468	In summary, it is considered unlikely that systematic trends within geochemical
469	evolution of Ruapehu lavas studied here were influenced primarily by near-surface processes.
470	Eruptions tapped magmas that had experienced variable degrees of differentiation in the crust
471	via crystal fractionation and crustal assimilation (Price et al. 2012). Thus, despite evidence
472	for compositional heterogeneity within eruptive sequences on annual to 10 kyr timescales
473	(e.g. Gamble et al. 1999, 2003), this study has defined a broadly coherent geochemical trend
474	for Ruapehu eruptive products over the \sim 200 kyr lifetime of the exposed edifice (Fig. 6). The
475	mechanisms of magma differentiation and the reasons for their varied influence through time
476	are now considered.

477

478 Nature and origin of silicic melts in Ruapehu lavas

Silicic glass compositions (often rhyolitic) for groundmass and melt inclusions in phenocrysts of arc andesite-dacite volcanic rocks have been widely reported (e.g. Reubi and Blundy 2009; Kent 2013; Lee and Bachmann 2014). Silicic melts can originate via one or more of: assimilation of plutonic crystal cumulate roots (e.g. Reubi and Blundy 2008); partial melting of subducted oceanic lithosphere and sediment (e.g. Tatsumi 2001); partial melting of crustal rocks (e.g. Jackson et al. 2003); or fractionation of mantle-derived primary basaltic magmas (e.g. Sisson et al. 2005). We focus on the latter two processes through the

486	compositions of glasses that represent trapped portions of melt from the Ruapehu magma
487	system. Dacite to rhyolite compositions ($\sim 67-78$ wt. % SiO ₂) were measured for Ruapehu
488	lava groundmass glass and pyroxene-hosted melt inclusions in this study (Fig. 9) and are
489	compared to data from Ruapehu pyroclasts (Pardo et al. 2014) and Taupo volcano (Barker et
490	al. 2015). Representative major element (Table 4) and trace element data (Table 5) are
491	provided here, and full data are available in the Electronic Supplementary Material.

492 The silicic composition of Te Herenga Formation melt inclusions may not be a 493 surprising result considering the ubiquity of such inclusion compositions in younger Ruapehu 494 lavas (Price et al. 2005) and arc volcanic rocks globally (Reubi and Blundy, 2009). However, 495 the low-K₂O contents are notable, given that Te Herenga Formation lavas also have low-K₂O 496 whole-rock compositions (Figs. 4, 9). Melt inclusion and groundmass glasses within the high-497 MgO andesites and dacites of the early Mangawhero Formation have significantly higher 498 K₂O contents in comparison to Te Herenga samples (Fig. 9). Discordance between melt 499 inclusion and groundmass glass compositions in the high-MgO andesite lavas may reflect 500 late-stage magma mixing processes (see next section). K₂O contents of up to 5–6 wt. % were 501 measured for groundmass glasses, melt inclusions and xenolith-hosted glasses within dacite 502 lavas. The pattern of the quenched melt network in the xenolith fragment within dacite 503 sample CC415 that was analysed indicates that the melt was derived from partial melting of 504 the xenolith (Fig. 3). This glass is also characterized by Rb contents of up to 270 ppm (Table 505 5). Melt inclusions in Whakapapa Formation lava flows exhibit a range of compositions that 506 return to lower K₂O contents (Fig. 9). The compositions presented here are broadly consistent 507 with groundmass glass and melt inclusion data for historically erupted samples (Gamble et al. 508 1999; Kilgour et al. 2013) and previous analyses from Whakapapa Formation lavas (Price et 509 al. 2012) and pyroclastic deposits (Pardo et al. 2014). Furthermore, the glass data are

510	intersected by the lower SiO ₂ – K_2O trends defined by <26 ka lava whole-rock compositions

511 when compared to earlier Mangawhero Formation lavas (Fig. 9).

512 Glass data for the different formations broadly align with their respective whole-rock 513 trends that are age-constrained (Fig. 9), and therefore represent the products of different 514 patterns of assimilation and fractional crystallization (AFC: DePaolo, 1981). The distinctive whole-rock ¹⁴³Nd/¹⁴⁴Nd isotopic compositions of Te Herenga lavas (Fig. 8; Price et al. 2005, 515 516 2012) have been inferred to reflect an early stage of the thermal evolution of the magma 517 conduit system beneath Ruapehu attended by assimilation of mafic oceanic crust, which is 518 considered to form the lower crust underlying the meta-sedimentary greywacke-argillite 519 Torlesse Terrane sequence of middle to upper crustal rocks (Graham 1987; Price et al. 2012; 520 Waight et al. 2017). Despite the limited amount of data, the distinct compositions of Te 521 Herenga glasses support the idea that melts and magmas were generated by distinct 522 petrogenetic processes during the $\sim 200-150$ ka time period when compared to subsequent 523 magmatism. The role of assimilation of continental crust in the genesis of younger magmas has been previously documented, and high ⁸⁷Sr/⁸⁶Sr values for Mangawhero Formation 524 525 dacites indicate that evolved magmas at Ruapehu experienced more mid-upper crustal 526 contamination (Graham and Hackett 1987; Price et al. 2005, 2012; Fig. 8).

527 When compared to younger lavas, the high-K₂O nature of glasses within the high-528 MgO lavas of the Mangawhero Formation is also considered here to represent AFC 529 processing that involved relatively greater assimilation of mid-upper crust. This inference is 530 supported by the observed Rb/Zr trends (Fig. 6): elevated values for evolved lavas would 531 require addition of Rb from a crustal source given that Zr is unlikely to be fractionated by the 532 phases observed in these samples. Metasedimentary crust assimilation is corroborated by 533 radiogenic isotope data (Price et al. 2012; Waight et al. 2017), and evidence of this process 534 was fortuitously captured by the analysis of a partially fused xenolith in sample CC415 (Fig.

535 3). We interpret that the entrained glass represents an *in situ* partial melt of a meta-

sedimentary crustal rock with a significant contribution from melting of K-bearing minerals
(such as biotite and K-feldspar) that are known to occur within the Torlesse Terrane basement
rocks (Graham and Hackett 1987; Graham et al. 1990; Adams et al. 2009; Price et al. 2012).
Published whole-rock ¹⁴³Nd/¹⁴⁴Nd-⁸⁷Sr/⁸⁶Sr systematics (Price et al. 2012; Waight et al.
2017) and trace element data indicate that Whakapapa Formation lavas were derived from
magmas that were contaminated by continental crust, although to a lesser degree than the 50–
35 ka Mangawhero Formation lavas (Figs. 7, 8).

543 Modelled chemical evolution pathways are presented in Fig. 10 in order to replicate 544 these distinct conditions of magma differentiation. The La/Sm ratio is used to indicate the 545 degree of evolution, and should be insensitive to compositional changes caused by crystal 546 fractionation. Plotted against this ratio, Rb/Zr values should generally increase due to crystal 547 fractionation because Rb is more incompatible than Zr. Significant differences in Rb/Zr 548 versus La/Sm trends between lavas of different ages, however, will reflect variable addition 549 of Rb from potassic phases in crustal rocks during assimilation. In general, post-26 ka lavas 550 lie along a flatter trend than 50–35 ka lavas, which extend to higher Rb/Zr values at higher 551 La/Sm values (Fig. 10). These broadly high and low trends for whole-rock Rb/Zr versus 552 La/Sm variations are seen in lavas erupted from (1) 50–35 ka and (2) \leq 26 ka, respectively. 553 These trends can then be compared against modelled compositions for AFC evolution that 554 involve: (1) a relatively high Rb/Zr contribution from a high-K crustal assimilant; and (2) a 555 relatively low Rb/Zr contribution from a low-K crustal assimilant (Fig. 10). The compositions 556 of the assimilants are constrained by *in situ* analyses of glass within meta-sedimentary 557 xenoliths in lavas of the Mangawhero Formation (this study) and Whakapapa Formation 558 (Price et al. 2005). Thus, assimilants were chosen because they have been observed and 559 analysed within glasses in the host lavas of interest.

560	While there is overlap between lavas of the different periods, the modelled pathways
561	aid the interpretation of two key features of the dataset. Firstly, the highest Rb/Zr values,
562	which are observed for dacite lavas erupted from 50-35 ka, cannot be produced by AFC
563	involving a low-K assimilant. Secondly, the low Rb/Zr values for post-26 ka lavas over a
564	range of La/Sm values are best-approximated by magma differentiation via AFC involving a
565	low-K assimilant. Post-26 ka magmas were likely generated during differentiation in a crustal
566	melting scenario that was more advanced than at ~40 ka beneath Ruapehu due to the
567	continued flux of magma through the lithosphere. In this regard, our modelling suggests that
568	post-26 ka magmas interacted with crust that was relatively depleted in elements such as K
569	and Rb when compared with those that fed the \sim 50–35 ka lavas. The difference is inferred to
570	have been due to the progressive heating and partial melting of the crustal column, which is
571	interpreted to have led to either or both of: (1) the exhaustion of fertile, K- and Rb-rich
572	mineral phases that were consumed during earlier (50-35 ka) stages of upper crust
573	assimilation; and (2) dilution effects reflecting the higher degrees of partial melting as
574	magma flux continued through the crust beneath Ruapehu.
575	
576	A case for mafic recharge recorded in high-Mg andesite-dacites at Ruapehu

577 High-magnesian andesites (HMAs) are intermediate arc volcanic rocks (54–65 wt. %

578 SiO₂) with high Mg# (\geq 50) and Cr and Ni concentrations (Kelemen et al. 2003a). At an

579 extreme, the eruption of primitive HMAs with $Mg\# \sim 70$, although rare, is of particular

580 importance because such magmas may be produced by partial melting of hydrous mantle

- peridotite (e.g. Grove et al. 2002, 2005; Wood and Turner 2009) and therefore play a
- fundamental role in the genesis of arc magmas and continental crust (Kelemen et al. 2003b).
- 583 Other models for the petrogenesis of HMAs invoke equilibration of slab-derived melts with

584	ultramafic mantle wedge material (e.g. Kay 1978; Shimoda et al. 1998; Tatsumi 2001;
585	Yogodzinski et al. 2001) and felsic-mafic magma mixing combined with entrainment of
586	ultramafic crystal material within arc lithosphere (e.g. Kawabata and Shuto 2005; Streck et al.
587	2007). HMAs have been studied at Mt Shasta in the Cascade arc (Anderson 1973; Grove et
588	al. 2002; Streck et al. 2007) and the Setouchi volcanic belt in southwest Japan (Shimoda et al.
589	1998; Kawabata and Shuto 2005). New Zealand examples have erupted at White Island
590	(Mg# ₇₀ : Heyworth et al. 2007), Pukeonake (Mg# ₅₀ : Graham and Hackett 1987; Beier et al.
591	2017) and Hauhungatahi (Mg#70: Cameron et al. 2010). Ruapehu high-MgO lavas were
592	described by Graham and Hackett (1987), but corresponding models of their generation
593	lacked the chronostratigraphic context (Conway et al. 2016) and microanalytical data
594	discussed here.
595	Andesite and dacite lavas of the Mangaehuehu and Te Kohatu eruptive packages,
596	respectively, are HMAs with whole-rock Mg# for the lavas ranging from 60-69 (andesites)
597	and 54–57 (dacites). Mangaehuehu package and esites exhibit high MgO contents over a ${\sim}5$
598	wt. % SiO ₂ range, and Te Kohatu package dacites (~65 wt. % SiO ₂) have MgO contents that
599	are notably higher than other Ruapehu dacite lavas (Fig. 4). The Ruapehu HMAs also show
600	high Ni contents (up to 148 ppm) compared to other Ruapehu lavas at equivalent SiO_2
601	contents (Fig. 4). A petrographic feature that distinguishes these HMA lavas from lower-
602	MgO flows at Ruapehu is their high proportions of resorbed orthopyroxene that are mantled
603	by euhedral rims with high Mg# values (~85–90: Fig. 3). The HMAs also contain
604	metasedimentary xenoliths, indicating that their source magmas interacted with and
605	assimilated crustal rocks. The petrographic and geochemical characteristics indicate that
606	derivative magmas underwent substantial fractionation and assimilation processes in the
607	lithosphere and do not represent near-primary melting products of the mantle wedge.

608	Comparisons are made in Fig. 5 between the composition of Ruapehu lavas and the
609	relatively primitive Waimarino and Kakuki basalts from the central Taupo Volcanic Zone to
610	the north (Gamble et al. 1993), as well as the Pukeonake and Ohakune andesites, which were
611	erupted from peripheral vents located ~20 km from the summit of Ruapehu (Figs. 1, 2, 5).
612	Eruptive activity at the Ruapehu peripheral vents pre-dates the Oruanui tephra (25.4 ka;
613	Vandergoes et al. 2013) as indicated by stratigraphic relations (Hackett, 1985; Froggatt and
614	Lowe, 1990), and is likely to have occurred within the period ~40–30 ka,
615	penecontemperoneous with the high-Mg lavas from Ruapehu (50-35 ka). The high-MgO
616	suite of Ruapehu lavas trends toward the composition of Waimarino basalt, and is
617	approximately intercepted by the composition of Pukeonake and Ohakune andesites (Fig. 6).
618	Low-MgO lavas erupted from 50-35 ka define a trajectory from Mangaturuturu package
619	dacite toward Kakuki basalt, which is broadly intercepted by all post-35 ka and pre-80 ka
620	lavas from the Ruapehu edifice.
621	Pukaanaka layas haya baan dasaribad as hybrid andositas and their compositions can

Pukeonake lavas have been described as hybrid andesites and their compositions can 621 622 be modelled by mixing between dacite and Waimarino basalt (Graham and Hackett 1987; 623 Beier et al. 2017). Ruapehu HMAs are similarly interpreted here to have been produced via 624 mixing between a deeper-sourced mafic magma and a mid- to upper crustal-level felsic 625 magma. Clusters of orthopyroxene preserved in the HMAs display high-MgO rims only on 626 the outermost exposed margins of the crystal clots (Fig. 3), indicating that the mafic magma 627 encountered a crystal-rich magma and interacted with crystal margins that were in contact 628 with the mafic magma during disaggregation of the crystal assemblage (cf. Nakagawa et al. 629 2002; Fig. 3). Subsequent crystal growth occurred within the hotter magma and accounts for 630 the higher MgO rims around the resorbed cores. Evidence for interaction between recharging 631 higher temperature magmas and stagnant, more felsic magmas has also been reported in the 632 ~11 ka Pourahu (Donoghue et al. 1995) and 1995-1996 Ruapehu eruptives (Nakagawa et al.

633 2002). Whole-rock Mg# values of ~60 for such samples are not as high as for Ruapehu 634 HMAs with equivalent SiO₂ contents reported in this study, which may reflect a greater 635 contribution from the mafic end-member within the Mangawhero Formation HMAs. The 636 comparable crystal zonation features are interpreted to reflect a similar process for older lavas 637 and younger examples, however, and has been documented elsewhere. In particular, similar 638 pyroxene zoning features and magma mixing processes have been described for HMAs from 639 Mt Shasta (Streck et al. 2007, figure 2) and the Setouchi volcanic belt (Kawabata and Shuto 640 2005, figure 7).

641 Constraining the nature of primitive magmas in the southern end of the Taupo 642 Volcanic Zone is difficult because even relatively primitive basaltic-andesites have generally 643 undergone textural and chemical modification in the crust (Waight et al. 2017). The reverse-644 zoned crystal cargo of HMAs provides only a cryptic record of mafic magma intrusion into 645 the crust beneath Ruapehu. Comparison to regional primitive lavas, however, yields insight 646 into the spatial distribution of such an end-member with respect to regional volcanism. The 647 shared petrogenetic origin of hybrid andesites at Ruapehu and within its periphery 648 (Pukeonake and Ohakune) may indicate the presence of a common mafic end-member within 649 the Tongariro Volcanic Centre as well as the broader Taupo Volcanic Zone.

650

651 **IMPLICATIONS**

We have presented new whole-rock, mineral and glass compositions for edificeforming lavas within a high-resolution eruption chronology for Ruapehu volcano. Together with existing isotopic data and petrogenetic models, these data further clarify the evolution of the Ruapehu magma system and provide a case study for continental arc volcanism. The long-term compositional variability of lava flows during the ~200 kyr lifetime of the exposed 657 Ruapehu edifice offers snapshots into the crustal dynamics of an evolving arc magmatic

658 system. Major implications are as follows.

659 •	The broadly coincident effusion of at least 4 compositionally distinct lava flow
660	packages during the \sim 50–35 ka time window at Ruapehu indicates the existence of a
661	compositionally heterogeneous central magma system. Reconstructed vent locations
662	for each lava package indicates that discrete magma batches were spatially and
663	compositionally distinctive despite (in some cases) being vented within 2 km of each
664	other. These results indicate that relatively large-scale heterogeneities can exist within
665	the magma systems feeding composite volcanoes, as well as the more commonly cited
666	small-scale heterogeneities that are tapped by explosive eruptions of relatively low-
667	volume and ephemeral magma batches at Ruapehu (e.g. Gamble et al. 1999; Auer et
668	al. 2013; Kilgour et al. 2013).

669 The systematic geochemical evolution of Ruapehu lavas is unlikely to have been 670 influenced primarily by density filtering of eruptible magma compositions due to 671 loading and unloading of the edifice. Instead, eruptions tapped magmas that had 672 undergone variable degrees of differentiation in the crust via crystal fractionation and 673 crustal assimilation. The effect of crustal assimilation on magma compositions was 674 greatest for the period from \sim 50–35 ka, due to the progressive advection of heat 675 through the lithosphere leading to partial melting and assimilation of mid- to upper 676 crustal meta-sedimentary country rocks. Post-26 ka magmas are interpreted to have 677 assimilated less enriched melts due to the prior consumption of fertile phases in the crustal column and/or higher degrees of partial melting of continental crust associated 678 679 with ongoing magma and heat flux through the magma system. A similar ~200 kyr 680 time-composition trend has been shown for Volcán Parinacota (Hora et al. 2007).

6 81 •	The compositional similarity between the groundmass glass of a dacite lava and a
682	xenolith-hosted partial melt within that lava (sample CC415) confirms the importance
683	of crustal assimilation in the geochemical characteristics of eruptive products at
684	continental arc volcanoes. However, to better constrain the processes involved, further
685	work is required to characterise the mineralogy and isotopic composition of the
686	crustal lithologies beneath Ruapehu volcano, as well as the composition of the
687	primary magmas.

High-MgO andesites and dacites that were erupted between ~50 and 35 ka at Ruapehu
 were produced via interaction between felsic and mafic magmas. Clusters of pyroxene
 were entrained from the more felsic magma reservoirs by the influx of hotter, more
 mafic magma, and were then resorbed and overgrown by Mg-rich rims. Ruapehu
 high-Mg andesites and dacites represent magmas resulting from middle to upper
 crustal-level assimilation and mixing processes rather than being primary magmas
 derived directly by mantle melting.

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- 706

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- 989 Table 1. Summary of chronostratigraphic formations and eruptive packages for edifice-
- 990 forming lava flows at Ruapehu volcano (modified from Conway et al. 2016).

Formation	Eruptive Package
	Crater Lake
	lwikau
	Saddle Cone
Whakapapa (<15 ka)	Tureiti
	Paretetaitonga
	Rangataua
	Turoa
	Makotuku
	Waitonga
Mangawhero late (26-15 ka)	Te Piripiri
	Horonuku
	Whakapapaiti
Mangawhero middle (35-26 ka)	Manganuiotaeo
	Mananui
	Te Kohatu
Mangawhero early (50-35 ka)	Mangaturuturu
	Mangaehuehu
	Ngahuinga
Wahianoa (160-80 ka)	Undifferentiated
Te Herenga (200-150 ka)	Undifferentiated

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1001 Table 2. Summary of EPMA data for major mineral phases in Ruapehu lavas.

	Sample	Information	L	Whole-ro	ck	Phenocryst compositional range (rim)				
Number	Fmtn	Package	Age (ka)	SiO ₂ (wt.%)	Mg#	Opx (Mg #)	Cpx (Mg #)	Plg (An %)	Ol (Mg #)	
CC281	WH	SC	10	56.0	56	62-77 (68-80)	65-85 (59-81)	52-81 (67-80)	73-75	
CC260	WH	TR	12	57.4	52	63-68 (68-78)	67-81 (71-75)	50-90 (56-73)	na	
CC335	WH	РТ	15	58.5	51	61-77 (65-70)	69-84 (71-73)	54-89 (58-88)	70-72	
CC193	WH	RT	15	59.3	47	61-69 (65-71)	70-74 (71-73)	50-87 (53-79)	na	
CC415	MA	TK	38	64.2	55	59-86 (58-90)	72-77 (67-86)	38-64 (45-62)	72-75	
CC479	MA	TK	38	64.9	54	59-86 (58-73)	70-87 (67-88)	40-70 (42-63)	na	
CC313	MA	TK	42	64.4	57	49-85 (54-74)	67-83 (66-88)	38-81 (33-65)	na	
CC069	MA	ME	42	59.3	65	68-90 (68-85)	72-88 (72-85)	48-82 (40-70)	82-84	
CC077	MA	ME	42	60.1	65	65-84 (69-90)	68-89 (72-87)	50-83 (52-60)	na	
CC154	MA	ME	42	60.9	64	69-82 (61-89)	70-85 (74-80)	42-76 (46-61)	78-80	
CC089	MA	ME	43	58.7	66	70-78 (77-89)	72-78 (69-87)	53-68 (51-70)	78-80	
CC125	MA	ME	44	56.8	69	69-90 (67-89)	72-86 (80-84)	48-60 (55-63)	na	
CC308	TH	-	170	57.0	54	67-70 (60-61)	71-83 (70-71)	56-82 (40-67)	na	
CC326	TH	-	170	55.2	53	65-72 (54-60)	70-74 (72-74)	61-82 (78-79)	69-70	
CC348	TH	-	187	53.8	59	67-76 (66-77)	63-86 (73-84)	68-86 (49-70)	70-85	

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1003 Notes. Formation abbreviations are WH (Whakapapa), MA (Mangawhero), and TH (Te

1004 Herenga). Package abbreviations are SC (Saddle Cone), TR (Turoa), PT (Paretetaitonga), RT

1005 (Rangataua), TK (Te Kohatu) and ME (Mangaehuehu). SiO₂ (in weight %) and values of

1006 Mg# [100Mg/(Mg + Fe)] are shown for whole-rock lava. Ranges of Mg# values determined

1007 by EPMA are shown for orthopyroxene (Opx), clinopyroxene (Cpx) and olivine (Olv)

1008 phenocrysts and their rims (in parentheses). Anorthite % shown for analyses of plagioclase

1009 (Plg) phenocrysts and rims (in parentheses). NA indicates not analysed. Full data are in the

- 1010 Electronic Supplementary Material.
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1017 Table 3. Whole-rock major and trace element data for representative Ruapehu lavas.

Sample	GL1983	GL1030	CC335	CC130	CC226	CC364	CC408	CC143	CC513	CC462	CC547
Package	CL	IW	РТ	TR	MK	ТР	MK	WT	WH	MG	MG
Age (ka)	0±2	6±2	15±3	15±2	18±2	21±6	21±3	23±2	26±4	31±2	31±5
SiO ₂	61.45	57.30	58.51	60.08	61.37	58.29	60.87	62.26	65.61	60.73	62.46
TiO ₂	0.72	0.68	0.73	0.70	0.89	0.70	0.78	0.67	0.62	0.86	0.81
Al2O ₃	15.71	16.71	17.09	15.91	17.25	16.08	17.28	16.20	14.88	16.87	17.00
Fe ₂ O ₃	6.41	7.57	7.51	6.59	6.03	7.57	6.42	5.84	5.02	6.49	5.74
MnO	0.10	0.12	0.12	0.10	0.09	0.12	0.10	0.10	0.08	0.10	0.09
MgO	3.94	5.20	3.96	4.95	2.52	5.22	2.64	3.58	3.43	3.23	2.44
CaO	6.03	7.67	7.16	6.51	5.91	7.34	6.02	5.59	4.50	6.10	5.15
Na ₂ O	3.25	3.21	3.22	3.25	3.68	3.00	3.75	3.56	3.69	3.41	3.56
K ₂ O	2.23	1.39	1.57	1.76	2.07	1.54	1.97	2.03	2.06	2.03	2.56
P_2O_5	0.16	0.14	0.14	0.14	0.18	0.14	0.17	0.17	0.11	0.19	0.19
LOI	-0.13	0.12	-0.15	0.01	-0.17	-0.07	-0.10	0.05	0.36	-0.16	0.24
Total	99.19	100.41	99.63	99.44	99.95	98.92	99.08	99.89	100.25	100.00	99.49
Li	22.1	19.9	15.9	20.7	22.2	13.6	25.7	26.7	27.0	21.6	31.3
Sc	19.7	25.2	24.4	22.1	17.2	27.2	17.9	17.7	15.4	19.6	15.8
V	159	194	199	161	195	205	172	128	120	172	144
Ni	34.7	65.2	18.2	66.8	13.4	41.0	7.7	52.4	49.9	19.0	14.5
Cu	21.6	34.6	26.3	33.3	14.6	50.8	32.7	29.9	18.4	28.9	23.5
Zn	61.5	66.3	77.5	59.3	58.8	68.4	66.7	59.9	73.1	68.2	60.6
Ga	17.0	18.3	19.2	17.6	18.8	16.9	19.4	18.1	17.4	18.5	18.6
Rb	86.3	46.8	59.2	62.7	78.4	57.4	69.0	75.5	84.7	72.6	110.2
Sr	251	299	257	264	304	225	305	266	229	305	258
Y	19.7	17.9	21.2	16.8	20.8	19.8	19.6	20.0	13.7	21.9	23.2
Zr	151	96	119	123	153	113	136	133	149	153	172
Nb	5.86	4.03	4.81	5.31	6.17	4.54	5.13	6.23	6.77	6.53	7.18
Cs	3.68	1.79	2.02	2.70	2.62	2.07	2.13	2.80	3.45	2.66	2.96
Ва	408	337	338	355	452	311	417	409	410	425	454
La	15.0	10.0	13.7	12.5	16.0	12.2	14.9	15.6	8.7	17.9	19.9
Ce	32.9	22.2	29.4	26.7	35.1	26.7	32.9	34.7	17.7	38.7	41.6
Pr	4.03	2.77	3.64	3.29	4.35	3.35	3.95	4.26	2.21	4.74	5.27
Nd	15.8	11.9	14.7	13.1	17.1	13.6	16.1	17.1	9.1	19.0	21.0
Sm	3.52	2.97	3.38	3.02	3.//	3.12	3.58	3.69	2.21	4.22	4.48
Eu	0.90	0.87	0.92	0.80	1.01	0.85	0.97	0.90	0.71	1.08	1.05
Ga	5.78	2.94	5.62	5.05	3.82	5.51	5.72	5./1	2.31	4.55	4.08
10 D	0.55	2.00	2.50	0.47	2.46	0.55	0.54	0.38	0.57	2.03	0.00
Dy H-	5.59	2.99	5.39	2.97	5.40	5.56	0.70	5.42	2.54	5.62	4.07
П0 Бъ	0.09	1.77	0.74	1.76	0.75	0.71	1.00	0.08	1.49	0.77	0.81
Er Tm	2.04	0.27	2.14	0.27	2.10	2.07	0.20	0.20	0.22	0.22	2.36
1 III Vh	0.50	1.79	0.52	1.75	0.50	2.02	1.04	1.86	0.22	0.52	0.55
10	0.20	0.27	2.15	0.26	2.03	2.05	0.20	0.27	0.22	2.07	0.22
LU	0.29	0.27	0.55	2.44	0.50	2.12	2.62	2.62	4.01	4.02	0.55
пі	4.00	2.71	0.27	5.44 10.5	4.12	7.40	5.05 0.47	17.1	4.01	4.02	4.49
ть	12.0 8.12	1.15	5.67	5.07	7 22	5 25	5.47	6.42	6.42	7.24	9.65
III II	2 21	1.22	1.43	1.60	1.85	1.42	1.63	1.73	1.91	1.76	2 /1

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1020 Table 3 (continued).

Sample	CC508	CC570	CC125	CC216	CC424	CC470	CC562	CC209	CC213	CC326	CC348
Package	ТК	MN	ME	ME	NG	MT	WA	WA	WA	ТН	ТН
Age (ka)	38±1	40±2	43±1	43±1	40±3	41±2	88±6	121±4	121±3	170±8	186±7
SiO ₂	65.03	63.57	56.80	60.19	59.34	63.49	57.55	58.36	58.40	55.24	53.79
TiO ₂	0.71	0.85	0.70	0.77	0.76	0.83	0.62	0.68	0.70	0.65	0.67
Al2O ₃	15.27	15.84	14.25	15.00	16.81	16.74	16.51	19.70	17.25	17.21	16.62
Fe ₂ O ₃	4.74	5.51	7.44	6.55	6.52	5.27	7.38	6.06	7.40	9.16	9.20
MnO	0.07	0.08	0.12	0.10	0.10	0.08	0.12	0.09	0.12	0.15	0.15
MgO	2.86	2.80	8.61	5.53	4.22	2.21	5.60	2.34	4.36	5.30	6.75
CaO	4.45	4.87	7.67	6.36	6.81	4.87	7.82	7.49	6.89	8.42	9.17
Na ₂ O	3.42	3.50	2.85	3.28	3.48	3.64	3.37	3.92	3.31	3.12	2.82
K ₂ O	3.27	2.78	1.41	2.05	1.80	2.68	0.93	1.22	1.44	0.66	0.71
P_2O_5	0.18	0.20	0.14	0.18	0.16	0.20	0.11	0.13	0.13	0.09	0.11
LOI	0.10	-0.02	-0.10	-0.02	-0.11	0.20	-0.09	0.06	0.11	0.01	-0.25
Total	100.28	99.44	99.89	100.20	99.07	99.62	99.54	100.16	99.90	99.27	99.39
Li	36.7	31.0	19.5	23.6	20.0	24.8	14.9	13.6	21.7	10.2	8.34
Sc	14.3	15.8	27.1	21.6	19.9	16.0	27.0	16.8	23.9	32.0	34.5
V	121	143	191	164	170	146	196	175	188	233	249
Ni	30.4	36.7	175.9	123.3	49.2	12.1	54.9	15.4	30.1	28.6	54.7
Cu	27.9	39.1	39.2	88.5	37.4	12.3	47.6	93.8	43.1	74.0	38.3
Zn	30.2	68.0	61.8	59.2	60.8	66.5	70.1	64.6	81.0	87.1	81.4
Ga	17.8	18.6	15.5	17.3	18.4	19.0	17.8	20.9	18.8	17.9	17.5
Rb	144	111	52.8	79.7	61.8	105	25.4	36.2	44.9	16.6	17.5
Sr	203	232	227	241	252	243	289	327	250	211	226
Y	23.2	23.5	17.5	21.8	18.5	21.3	14.6	19.1	19.2	18.9	18.7
Zr	215	199	114	150	130	186	69.3	92.8	101	59.4	58.5
Nb	9.00	8.36	4.16	6.26	4.58	7.22	3.00	3.79	4.46	1.83	1.89
Cs	5.15	3.83	2.32	2.38	2.00	3.30	0.61	1.07	1.56	0.42	0.37
Ba v	552 20.0	509	302	404	345	495	236	316	332	184	222
La	20.9	20.2	10.6	10.7	12.8	10.0	0.25	22.8	11.4	5.44	0.01
Ce D-	45.5	44.0 5.21	23.5	55./ 4.40	28.4	30.5	14.5	22.8	25.7	11.5	15.5
rr Na	21.0	21.5	12.90	18.1	14.6	4.55	1.95 8.42	12.7	12.5	7.67	1.94 8.07
Nu Sm	4 53	4 56	2.05	3.02	3 20	10.0	8.45 2.10	3.05	3 22	2.17	2 38
5m Fn	0.91	1.01	0.82	0.93	0.87	1.03	0.71	0.92	0.87	0.74	0.81
Ed	4 59	4.60	3.08	4.09	3 40	4 18	2 29	3.12	3 27	2 55	2.81
оu Th	0.64	0.66	0.48	0.62	0.50	0.59	0.39	0.49	0.52	0.45	0.46
Dv	3.88	4.11	2.92	3.75	3.22	3.67	2.58	3.11	3.29	3.09	3.08
Ho	0.80	0.82	0.60	0.75	0.65	0.74	0.53	0.65	0.68	0.68	0.67
Er	2.23	2.35	1.78	2.18	1.83	2.16	1.51	1.90	1.94	1.97	1.98
Tm	0.32	0.34	0.26	0.33	0.27	0.32	0.23	0.28	0.30	0.30	0.29
Yb	2.15	2.23	1.67	2.08	1.78	2.15	1.52	1.80	1.95	1.99	1.92
Lu	0.32	0.33	0.25	0.31	0.26	0.32	0.24	0.28	0.29	0.31	0.30
Hf	5.40	5.30	2.97	4.17	3.41	4.92	1.99	2.57	2.85	1.73	1.73
Pb	17.1	17.5	7.69	12.4	12.4	12.8	4.99	7.37	27.3	6.00	3.63
Th	12.1	10.0	4.91	7.08	5.82	9.82	2.36	3.39	4.15	1.40	1.63
U	3.58	2.70	1.29	2.02	1.49	2.59	0.65	0.93	1.10	0.44	0.43

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1022 Notes. Major oxide values are in weight % and are normalised to anhydrous totals (100 wt.

1023 %) with LOI (loss on ignition) and original analytical totals displayed. Trace elements

1024	contents are in ppm. Package and Formation abbreviations are: Crater Lake (CL);
1025	Paretetaitonga (PT); Tureiti (TR); Makotuku (MK); Waitonga (WT); Whakapapaiti (WH);
1026	Manganuioteao (MG); Te Kohatu (TK); Mananui (MN); Mangaehuehu (ME); Ngahuinga
1027	(NG); Mangaturuturu (MT); Wahianoa Formation (WA); Te Herenga Formation (TH). Full
1028	data are in the Electronic Supplementary Material. Ages are ⁴⁰ Ar/ ³⁹ Ar weighted mean plateau
1029	ages from Conway et al. (2016).
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1043 Table 4. Summary of ment menusion and Achonich and groundinass glass composition

Group	Value	SiO ₂	TiO ₂	Al ₂ O ₃	FeO	CaO	Na ₂ O	K ₂ O	Total
WH	avg	74.0	0.5	14.9	1.1	1.9	3.6	3.8	99.6
mi	max	77.0	1.4	19.0	2.2	3.6	4.5	5.6	100.5
n= 45	min	70.3	0.2	13.0	0.5	1.0	0.8	1.8	98.4
	2sd	3.7	0.4	2.6	0.8	1.4	1.3	2.2	1.3
WH	avg	73.8	1.2	12.2	3.1	1.0	3.4	5.0	99.7
gms	max	75.1	1.4	12.5	3.8	1.1	4.1	5.2	100.3
n=16	min	73.3	1.0	11.7	2.5	0.7	3.2	4.1	98.1
	2sd	1.0	0.2	0.5	0.6	0.2	0.4	0.5	1.4
ТК	avg	73.0	0.7	13.7	2.1	1.3	3.3	5.4	99.9
mi	max	75.6	1.2	16.2	3.3	2.8	4.2	6.2	100.8
n=67	min	70.0	0.2	12.1	1.0	0.7	1.0	4.0	98.5
	2sd	3.0	0.5	2.5	0.9	1.0	0.9	1.1	1.1
ТК	avg	73.8	1.0	12.7	2.3	1.0	3.0	5.8	98.0
i/x	max	75.2	1.3	15.5	2.8	1.4	3.7	6.3	99.0
n=17	min	70.5	0.7	12.1	1.7	0.9	2.7	5.5	97.1
	2sd	2.7	0.4	1.9	0.6	0.2	0.4	0.4	1.1
ТК	avg	74.4	0.8	12.6	2.2	1.2	3.0	5.5	101.0
gms	max	75.4	0.8	15.1	2.7	2.3	3.8	5.9	101.8
n=13	min	71.7	0.6	12.0	1.7	0.8	2.7	4.4	100.3
	2sd	2.1	0.1	1.7	0.4	0.9	0.6	0.8	0.8
ME	avg	72.8	0.6	15.1	1.3	1.8	4.0	4.2	100.0
mi	max	67.6	0.1	13.0	0.4	0.6	3.0	3.2	99.0
n=30	min	76.6	1.9	17.5	3.0	3.6	7.2	5.6	100.6
	2sd	4.9	1.0	2.0	1.4	1.6	1.4	1.1	0.8
ME	avg	69.4	1.7	13.0	5.1	2.5	4.5	3.0	99.9
gms	max	70.1	1.8	13.8	5.8	2.9	5.0	3.4	100.4
n=13	min	69.0	1.6	12.3	4.3	2.4	4.1	2.6	99.3
	2sd	0.7	0.1	0.8	0.8	0.3	0.5	0.5	0.5
TH	avg	72.7	0.5	15.9	1.5	2.4	4.3	2.4	100.4
mi	max	78.7	1.1	17.8	2.1	3.4	5.9	3.5	101.8
n=24	min	69.7	0.1	12.9	0.6	1.3	1.6	1.7	98.1
	2sd	5.6	0.8	3.0	0.8	1.1	1.6	0.8	2.0

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1047 Notes. Major oxide values are in weight % and are normalised to anhydrous totals (100 wt.

1048 %). Table is divided into groups of analyses from melt inclusions (mi), groundmass glass

1049 (gms), and interstitial and xenolith glass (i/x) from within different formations or packages:

1050 Whakapapa Formation (WH), Te Kohatu package, Mangawhero Formation (TK),

1051 Mangaehuehu package, Mangawhero Formation (ME), Te Herenga Formation (TH). From

1052 multiple analyses of glass within each group, values are displayed for average (avg),

1053 maximum (max), minimum (min) contents for each element, as well as 2 s.d. compositional

1054 variability. Full data are in the Electronic Supplementary Material.

1055 Table 5. LA-ICPMS trace element data for xenolith and interstitial glass.

Sample	X1-1	X2-1	X2-2	IG40-1	IG12 (n=7)
Sc	11.6	11.1	9.9	10.6	12 ± 2
V	64.6	52.6	32.7	70.3	57 ± 28
Zn	40.7	33.2	46.9	50.9	50 ± 7
Rb	255	274	273	272	271 ± 12
Sr	56.3	57.4	49.5	90.5	55 ± 6
Y	35.5	36.1	31.9	31.4	39 ± 4
Zr	402	435	420	397	486 ± 22
Nb	13.1	13.7	14.8	11.8	15 ± 2
Cs	18.0	17.8	18.3	17.4	18 ± 2
Ba	789	786	741	1410	834 ± 44
La	33.9	35.5	30.8	35.2	41 ± 5
Ce	73.2	74.5	69.4	77.6	86 ± 9
Pr	8.3	9.2	8.0	8.2	9 ± 1
Nd	32.1	31.3	29.3	29.7	37 ± 4
Sm	7.0	7.8	5.7	9.1	8 ± 2
Eu	0.5	0.6	0.7	1.0	1 ± 0.5
Gd	7.5	7.6	3.7	8.5	8 ± 3
Tb	0.9	0.7	0.6	1.2	1 ± 0.2
Dy	6.6	4.9	5.8	5.3	6 ± 2
Но	1.4	1.0	0.6	1.0	1 ± 0.2
Er	3.6	2.8	2.9	2.5	4 ± 1
Tm	0.7	0.5	0.7	0.2	0.5 ± 0.3
Yb	2.6	2.9	2.2	4.2	4 ± 1.5
Lu	0.4	0.5	0.3	0.3	0.5 ± 0.2
Hf	12.2	12.4	11.9	11.8	14 ± 3

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1057 Notes. Values are in ppm. Sample prefixes are: X = xenolith glass; IG = interstitial glass. X2-

1058 1 and X2-2 are analyses on different spots from the same xenolith. Average values ± 2 s.d.

1059 compositional variability are shown for 7 analyses from IG12.

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1066 FIGURE CAPTIONS

Figure 1. Location of the Taupo Volcanic Zone (TVZ) and associated features within the
North Island of New Zealand. TVZ basement rock outcrops for the Torlesse and Waipapa
terranes have been modified from Price et al. (2015). Locations for features referred to in the
text are shown for White Island (WI), Kakuki basalt (K), Taupo volcano (TV), Waimarino
basalt (W), Pukeonake (P), Tongariro (T), Hauhangatahi (H), Ruapehu (R) and Ohakune
craters (O).

1073 Figure 2. Geological map of chronostratigraphic formations at Ruapehu, modified from

1074 Conway et al. (2016). Marginal ticks and numbers are those of the NZTM 2000 grid, in

1075 meters; a reference point for latitude and longitude is marked in the top right of the map.

1076 Figure 3. Petrographic features of representative Ruapehu lava flows. Red numbers indicate

1077 Mg# ([100Mg/(Mg + Fe)]) of EPMA spots. (a) Clinopyroxene in Te Herenga Formation

1078 basaltic-andesite sample CC348. (b) Cluster of zoned plagioclase crystals in Whakapapa

1079 Formation andesite sample CC281. (c) Reverse-zoned orthopyroxene in Mangawhero

1080 Formation (Mangaehuehu package) andesite sample CC077. (d) Cluster of reverse-zoned

1081 orthopyroxene crystals in Mangawhero Formation (Mangaehuehu package) andesite sample

1082 CC089. (e) Resorbed olivine crystal, with symplectite of magnetite (mt) and orthopyroxene

1083 overgrowth, in Whakapapa Formation sample CC335. (f) Quenched glass (gl) in fragment of

1084 feldspathic xenolith (plag) within Mangawhero Formation (Te Kohatu package) dacite

sample CC415. For sample details and locations, see Conway (2016).

Figure 4. Bivariate plots of whole-rock K₂O, Al₂O₃, MgO, Ni, Rb and Zr versus SiO₂.

1087 Values of 2 s.d. uncertainty from XRF and ICP-MS analyses are smaller than the size of the

1088 symbols. Sample and standard data are in the Electronic Supplementary Material.

1089 Abbreviations for end-member samples referred to in the text are Te Kohatu package, high-K

1090 dacite (TK), Whakapapaiti package, low-K dacite (WH) and Mangaehuehu package, high-1091 MgO andesite (ME). Gray dots are data from Price et al. (2012). Type 2 plagioclase-phyric 1092 andesites of the Wahianoa Formation (Graham and Hackett, 1987) are the green diamonds 1093 labelled T2. Lines of best fit have been drawn in panel (a) and its inset plot to show the 1094 distinct trends for 50–35 ka lavas compared to <26 lavas. Arrows point towards 1095 representative mineral compositions to indicate azimuths of crystal accumulation (opposite 1096 direction indicates fractionation). 1097 Figure 5. Whole-rock compositional characteristics (a) and spatial distribution (b) of early 1098 Mangawhero Formation lava flows. (a) MgO and versus SiO₂ plot for lavas erupted from 50-1099 35 ka. Colour key for lava packages is shown in (b). Shaded fields show compositional 1100 ranges for pre-80 ka Ruapehu lavas (i), post-35 ka Ruapehu lavas (ii), and the ~26 ka 1101 Whakapapaiti dacite package (WH). Lines of best fit (dashed) for the low- and high-MgO 1102 suites are extended towards possible mafic end-members K (Kakuki basalt) and W 1103 (Waimarino basalt): basalt compositions are from Gamble et al. (1993). Average 1104 compositions of P (Pukeonake andesite) and O (Ohakune andesite) shown for comparison; 1105 data are from Cole (1978) and Hackett (1985). Locations of P and O are labelled in (b). 1106 Figure 6. Time-composition relationships for Ruapehu lavas (after Conway et al. 2016). Lava ages are 40 Ar/ 39 Ar weighted mean plateau ages (WMPA) with associated 2 s.d. 1107 1108 uncertainties shown by bars. Both ages and whole-rock compositions were determined for the 1109 same sample. The timing of the onset of deglaciation (light grey bar) and the sector collapse 1110 of the northern flank (dark grey bar) are shown on each panel. Data are also shown for lavas 1111 dated by Gamble et al. (2003) and lava clasts by Tost and Cronin (2015). Geochemical data 1112 for lava clasts is from Tost et al. (2016).

- 1113 Figure 7. Multielement diagrams for Ruapehu lavas. Whole-rock trace element
- 1114 concentrations are normalised to average values of continental crust from Rudnick and Gao
- 1115 (2003). Grey line repeated in all panels is Te Herenga basaltic-andesite sample CC348.
- 1116 Approximate eruption age is labelled for each sample.
- 1117 **Figure 8.** Whole-rock 143 Nd/ 144 Nd and 87 Sr/ 86 Sr isotopic compositions and SiO₂ contents for
- 1118 Ruapehu lavas with the samples allocated to the formations and age data from Conway et al.
- 1119 (2016). Isotopic data from Price et al. (2012). Note that 50–35 ka lavas with highest ⁸⁷Sr/⁸⁶Sr
- 1120 values do not appear in (a) because they do not have Nd isotopic data.
- 1121 **Figure 9.** SiO₂-K₂O compositional variations for whole-rock and glass compositions for
- 1122 Ruapehu lavas. Symbols are compositions for whole-rock (wr: coloured diamonds), melt
- 1123 inclusions (mi: open diamonds), groundmass glass (gm: coloured crosses) and xenolith glass
- 1124 (xg: open blue circles). Data are in the Electronic Supplementary Material. Fields are shown
- 1125 for Taupo volcano glass, post-Oruanui (TV) based on data from Barker et al. (2015), and
- 1126 pyroclast glass from post-glacial Ruapehu Plinian eruptions (RP) based on data from Pardo et
- al. (2014). Lines of best fit are shown for whole-rock data.
- 1128 Figure 10. Plot of Rb/Zr variation with La/Sm for whole-rock lavas compared against
- 1129 inferred chemical evolution pathways (dotted lines, 10 % increments of fractionation).
- 1130 Trajectories represent AFC models for Ruapehu magmas using lava samples CC125 and
- 1131 GL1030 as starting compositions for 50-35 ka and <26 ka lavas, respectively. Following
- 1132 Price et al. (2012), a fractionating assemblage of plagioclase (0.59), clinopyroxene (0.25),
- 1133 orthopyroxene (0.16) was used to constrain bulk distribution coefficients (D) of 0.04 (Rb),
- 1134 0.38 (Zr), 0.11 (La) and 0.29 (Sm) that were calculated using the values available at the Earth
- 1135 Reference Data and Models website (EarthRef: <u>http://www.earthref.org</u>), as collated in the
- 1136 Electronic Supplementary Material. High-K (6 wt. % K₂O) and low-K (4 wt. % K₂O) crustal

- assimilants with corresponding ratios of assimilation to fractionation (R) of 0.5 and 0.25 were
- 1138 used. The trace element compositions of the assimilant were constrained by analyses of
- 1139 xenolith glass from metasedimentary fragments within a 40 ka dacite lava (this study;
- 1140 CC415-X2) and within a Whakapapa Formation lava (Price et al. 2005; 104x). Modelling
- 1141 was performed using the PetroGraph software program (Petrelli et al. 2005). See text for
- 1142 discussion.
- 1143







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