Bull Volcanol (1999) 60:223–240

ORIGINAL PAPER

S. L. Donoghue · A. S. Palmer · E. McClelland
K. Hobson · R. B. Stewart · V. E. Neall
J. Lecointre · R. Price

The Taurewa Eruptive Episode: evidence for climactic eruptions at Ruapehu volcano, New Zealand

Received: 5 July 1998 / Accepted: 12 March 1999

Abstract The ca. 10,500 years B.P. eruptions at Ruapehu volcano deposited 0.2–0.3 km^3 of tephra on the flanks of Ruapehu and the surrounding ring plain and generated the only known pyroclastic flows from this volcano in the late Quaternary. Evidence of the eruptions is recorded in the stratigraphy of the volcanic ring plain and cone, where pyroclastic flow deposits and several lithologically similar tephra deposits are identified. These deposits are grouped into the newly defined Taurewa Formation and two members, Okupata Member (tephra-fall deposits) and Pourahu Member (pyroclastic flow deposits). These eruptions identify a brief (<ca. 2000-year) but explosive period of volcanism at Ruapehu, which we define as the Taurewa Eruptive Episode. This Episode represents the largest event within Ruapehu's ca. 22,500-year eruptive history and also marks its culmination in activity ca. 10,000 years B.P. Following this episode, Ruapehu volcano entered a ca. 8000-year period of relative quiescence. We propose that the episode began with the eruption of small-volume pyroclastic flows triggered by a magma-mingling event. Flows from this event travelled down valleys east and west of Ruapehu onto the upper volcanic ring plain, where their distal remnants are preserved. The genesis of these deposits is inferred from the remanent magnetisation of pumice and lithic clasts. We envisage contemporaneous eruption and emplacement of distal pumice-rich tephras and proximal welded tuff deposits. The potential for generation of pyroclastic flows during plinian eruptions at Ruapehu has not been previously considered in hazard assessments at this volcano. Recognition of these events in the volcanological record is thus an important new factor in future risk assessments and mitigation of volcanic risk at Tongariro Volcanic Centre.

Key words Ruapehu volcano · Taurewa Formation · Tephra · Pyroclastic flow · Remanent magnetism · Taurewa eruptive episode · Volcanic hazard

Editorial responsibility: W. Hildreth

Susan L. Donoghue (✉)
Department of Earth Sciences, University of Hong Kong, Pokfulam Road, Hong Kong
e-mail: donoghue@hkucca.hku.hk,
Fax: +852-25176912

Alan S. Palmer · Robert B. Stewart · Vincent E. Neall
Jèrôme Lecointre
Institute of Natural Resources, College of Science, Massey University, Private Bag 11-222 Palmerston North, New Zealand

Elizabeth McClelland · Kate Hobson
Department of Earth Sciences, University of Oxford, Parks Road, Oxford, OX1 3PR, UK

Richard Price
School of Earth Science, La Trobe University, AUS-3083 Bundoora, Australia

Present address: Richard Price, School of Science and Technology, University of Waikato, Private Bag 3105, Hamilton, New Zealand

Introduction

Mount Ruapehu is the largest andesitic stratovolcano in North Island, New Zealand. It is one of four major andesitic massifs (Kakaramea-Tihia, Pihanga, Tongariro and Ruapehu) of Tongariro Volcanic Centre (TgVC), which is located at the southwestern end of the Taupo Volcano Zone, a zone of active volcanism extending 250 km northeast across the central portion of North Island (Fig. 1; Cole 1978). Volcanism at TgVC may have commenced as early as ca. 1.7 million years B.P. (Flemming 1953), although the oldest exposed lava flows on Ruapehu and Tongariro are K–Ar dated at 230 and 273 ka, respectively (Patterson and Graham 1988; Hobden et al. 1996).

The geology of Ruapehu is detailed by Hackett and Houghton (1985) who defined four cone-building episodes represented by four formations (Table 1). The volcanic deposits of these formations are grouped into two lithofacies associations: central and flank vent, and proximal cone-forming. The former association com-

Table 1 Lithostratigraphy of Ruapehu composite volcano after Hackett and Houghton (1989). Each of the four formations corresponds to a cone-building episode

Formation	Approximate age (Ka)	Location of vents
Whakapapa	0–15	Summit and flanks of modern cone
Mangawhero	15–60	Summit region of Ruapehu
Wahianoa	60–120	SE quadrant of Ruapehu
Te Herenga	>120	North and NW Ruapehu

prises thin lava flows, remnants of historical lava domes, and welded spatter deposits; the latter largely comprises block lava flows and autobrecciated lavas. Unwelded primary tephra deposits are rare in the cone-forming sequences; most have been rapidly eroded from the steep flanks of the volcano so that presently little evidence remains on the cone of past explosive volcanism (Hackett and Houghton 1989).

A more complete record of Ruapehu's eruptive activity is found on the volcanic ring plain. Here, thick sequences of andesitic tephras mantle, and are interbedded with, Holocene and Pleistocene lahar deposits, lava flows, debris avalanche deposits, fluvial and aeolian sands, and exotic rhyolitic tephras erupted from the Taupo and Okataina volcanic centres north of TgVC (Fig. 1; Topping 1973; Palmer and Neall 1989; Donoghue 1991; Donoghue et al. 1995b). A detailed late Quaternary (22,600 years B.P. to present) tephrostratigraphy now exists for Ruapehu volcano (Table 2) based largely on the eruptive record preserved on the northern and eastern Ruapehu and Tongariro ring plains (Donoghue et al. 1995b).

The ring plain stratigraphy suggests that, in late Pleistocene-Holocene time (post 22,600 years B.P.), small-volume tephra eruptions and lahars dominate events at Ruapehu and debris avalanches and pyroclastic flows are rare. Ruapehu's steep, debris-covered flanks, and summit crater lake and ice caps predisposes Ruapehu to lahar generation, a scenario played out at many of the world's stratovolcanoes, such as at Mount Rainier in the Cascade Range (Washington, USA). Mount Rainier, a lofty (4390 m) snow-capped volcano, has a late Pleistocene-Holocene eruptive history similar to that of Ruapehu, with activity clearly dominated by lahar events, at times voluminous (Scott et al. 1995), a few pyroclastic flows, and numerous tephra eruptions, the majority of which have restricted dispersal (east and north of the volcano) and are of small volume (S. Donoghue and J.W. Vallance, unpublished data).

Two principal eruptive periods are identified within Ruapehu's ca. 22,500 years B.P. eruptive history. The earliest is recorded by Bullot Formation tephras (Table 2), a sequence of at least 50 sub-plinian lapilli and ash beds erupted between ca. 22,600 and 10,000 years B.P. These tephras are preserved mostly in sections east of the volcano, where they mantle older lahar deposits and lava flows. The second, most recent period (ca. 1850 years B.P. to present) is recorded by the late Holocene Tufa Trig Formation tephras (Table 2), a distinctive sequence of black vitric ashes distributed principally east and northeast of Ruapehu which are thought to be the products of phreatomagmatic and strombolian eruptions (Donoghue et al. 1995b; Donoghue et al. 1997).

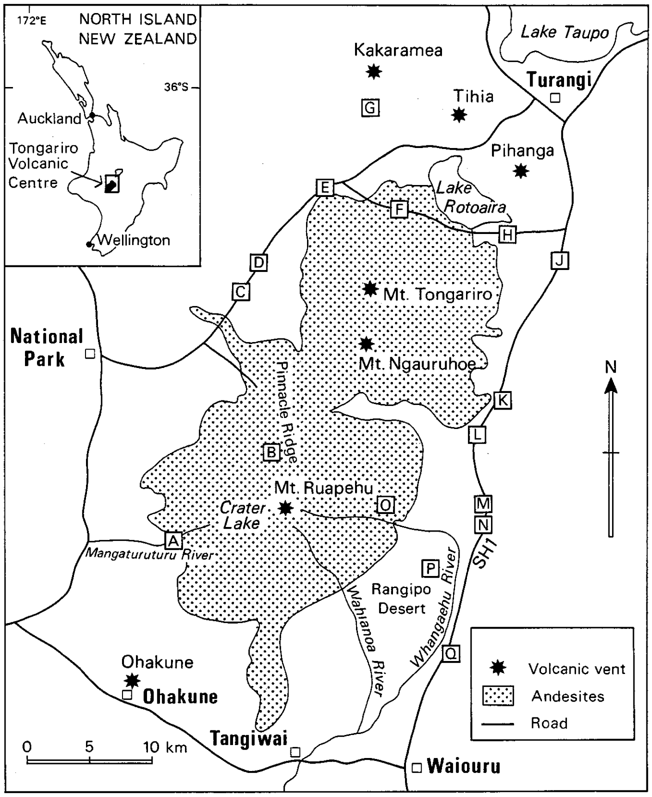

Fig. 1 Location of Mt Ruapehu and other principal volcanic vents of Tongariro Volcanic Centre, North Island, New Zealand; type, reference, and information sections (discussed in text) designated for Taurewa Formation. *A* Mangaturuturu S20/253085; *B* Pinnacle Ridge T20/318152; *C* Okupata type S19/284283; *D* Lava flow S19/287229; *E* Evians T19/352364; *F* Te Rato T19/376352; *G* Otamangaku T19/368414; *H* Poutu T19/4813252; *J* Hydro Access 10 T19/536270; *K* Desert Road 16 T20/481186; *L* Mangatoetoenui Quarry T20/459153; *M* Waikato Stream T20/467102; *N* Desert Road 11 T20/464092; *O* Bullot Track T20/412108; *P* The Chute T20/437045; *Q* Wahianoa Aqueduct T20/435990. Grid references are based on the NZMS 260 Map Series (1:50,000): Sheets S19 (Raurimu); S20 (Ohakune); T19 (Tongariro); and T20 (Ruapehu) drawn on the New Zealand Map Grid Projection and showing coordinates in terms of the New Zealand Geodetic Datum 1949

Recent investigation of the late Quaternary tephrostratigraphy of the northern Ruapehu ring plain has prompted re-examination of the ca. 10,500 years B.P. tephrostratigraphy, specifically the chronostratigraphic and eruptive relationships of a pyroclastic flow unit (Pourahu Member of the Bullot Formation; Donoghue et al. 1995b) and tephra fall units of the Okupata Tephra (Topping 1973). We present a new interpretation of this eruptive period and describe the events that took place during this time. We identify distal pyroclastic

Table 2 Stratigraphy and chronology of andesitic and rhyolitic (italicised) tephras preserved on the Mt. Ruapehu ring plain. Note stratigraphic position of Taurewa Formation relative to two dated rhyolitic tephras (Karapiti Tephra and Waiohau Tephra) used to constrain the age of the formation. *TVC* Taupo Volcanic Centre; *OVC* Okataina Volcanic Centre; *TgVC* Tongariro Volcanic Centre

Formation	Named members		Source	Age (years B.P.)[a]	14 C no.	Reference to age and stratigraphic definition
Tufa Trig Formation and Ngauruhoe Formation	Tf19–Tf1		Mt. Ruapehu TgVC	ca. 1850 to present		Donoghue et al. (1995b)
Taupo Tephra	*Taupo Ignimbrite*		*TVC*	*ca. 1850 ± 10*[b]		
Mangatawai Tephra			Mt. Ngauruhoe	2500 ± 200	NZ186	Fergusson and Rafter (1959); Donoghue et al. (1995b)
Papakai Formation	Unnamed ash		TgVC			Topping (1973)
	Black ash-1		Mt. Ruapehu			Donoghue et al. (1995b)
	Orange lapilli-2		Mt. Ruapehu			
	Orange lapilli-1		Mt. Ruapehu			
Mangamate Tephra	Poutu Lapilli	Pt	Mt. Tongariro	ca. 9700		Topping (1973)
	Wharepu Tephra	Wp	Mt. Tongariro			Donoghue et al. (1995b)
	Ohinepango Tephra	Oh	Mt. Tongariro			
	Waihohonu Lapilli	Wa	Mt. Tongariro			
	Unnamed tephra	ut'	–			
	Otutere Lapilli	Ot	Mt. Tongariro			
	Te Rato Lapilli	Tt	Mt. Tongariro	9780 ± 170	NZ1372	Topping (1973)
Unnamed tephra, *Karapiti Tephra*		*Kp*	*TVC*	*9820 ± 80*[b] *or 10,100*[c]		
Pahoka Tephra		Pa	Mt. Tongariro	ca. 10,000–9800		Topping (1973); Donoghue et al. (1995b)
Bullot Formation	Ngamatea lapilli-2	Nt–2	Mt. Ruapehu	*ca.* 10,000		Donoghue et al. (1995b)
	Ngamatea lapilli-1	Nt–1	Mt. Ruapehu			
Taurewa Formation	Okupata Member	Ok_1–Ok_3	Mt. Ruapehu			Donoghue et al. (1995b); this paper
	Pourahu Member	Ph	Mt. Ruapehu			
Bullot Formation	L18–L17	Sh	Mt. Ruapehu			Donoghue et al. (1995b)
	Shawcroft Lapilli		Mt. Ruapehu			
	L16		Mt. Ruapehu			
Waiohau Tephra		*Wh*	*OVC*	*11,850 ± 60*[b]		
Bullot Formation	L15–L1	Bt	Mt. Ruapehu			Donoghue et al. (1995b)
Kawakawa Tephra Formation		*Kk*	*TVC*	*22,590 ± 230*[d]		

[a] All ^{14}C ages discussed are conventional ages in radiocarbon years B.P. based on the old (Libby) half-life of 5568 years
[b] Froggatt and Lowe (1990)
[c] Wilson (1993)
[d] Wilson et al. (1988)

flow deposits using remanent magnetisation techniques and correlate tephra and pyroclastic flow deposits based on newly obtained stratigraphic data and lithological similarities of the deposits. The Taurewa Formation is defined, representing a sequence of tephra-fall and pyroclastic flow deposits that identify the largest eruptive events at Ruapehu volcano in late-Pleistocene to Holocene time and the only Ruapehu eruption known to have generated pyroclastic flows. We also examine the significance of this eruptive period in assessing the corresponding potential volcanic risk at Ruapehu.

Tephrostratigraphic and tephrochronological framework: previous work

Okupata Tephra

Okupata Tephra was first identified by Topping (1973) in sections north of Ruapehu (Table 3) and described as a prominent andesitic tephra comprising two lapilli units and an overlying palaeosol. Isopach data for the combined lapilli indicate strong northerly dispersal and eruption from a vent on Ruapehu near Pinnacle Ridge (Topping 1973; Hackett and Houghton 1989).

Table 3 Comparison of the stratigraphy of Taurewa Formation, Okupata Member and Pourahu Member (as defined in this paper) with that of Topping (1973) and Donoghue et al. (1995b)

Topping (1973)		Donoghue et al. (1995a, 1995b)		This study		Reference to age (years B.P.)
Formation	Member	Formation	Member	Formation	Member	
Mangamate Tephra		Mangamate Tephra		Mangamate Tephra		9790–9780 ± 170[b]
Undescribed andesitic tephras		Unnamed andesitic ash with interbedded Karapiti Tephra		Unnamed andesitic ash with interbedded Karapiti Tephra		9820 ± 80[c] 10,100[d]
Pahoka lapilli		Pahoka Tephra		Pahoka Tephra		ca. 9800–10,000[e,i]
	Bullot Formation	Pourahu Member[a]				
Okupata Tephra	Palaeosol Upper lapilli	Okupata Tephra	Palaeosol Upper lapilli	Medial ash Taurewa Formation	Okupata Member and Pourahu Member	– 9790 ± 160 to 2450 ± 340[b] ca. 10,300[f]
	Lower lapilli		Lower lapilli			
Undescribed andesitic tephras			Bullot Formation	Unnamed members	Bullot Formation	Unnamed members ca.10–22,600[g,h]

[a] Exact stratigraphic position relative to Okupata Tephra unknown
[b] Topping (1973)
[c] Froggatt and Lowe (1990) age of Karapiti Tephra
[d] Wilson (1993) age of Karapiti Tephra
[e] Topping (1973)
[f] Lowe and Hogg (1986)
[g] Donoghue et al. (1995b)
[h] Inferred stratigraphic age – not ^{14}C dated

Several radiocarbon ages are reported for this tephra [all ages are conventional ages in radiocarbon years B.P. based on the old (Libby) half-life of 5568 years]. Topping (1973) gives an age between 12,450 ± 340 and 9790 ± 160 years B.P., and Lowe and Hogg (1986) date the tephra at ca. 10,300.

Pinnacle Ridge tuff

Pinnacle Ridge Member is defined by Hackett and Houghton (1985) as a subplinian welded, coarse-grained lapilli bed (Pinnacle Ridge tuff) belonging to the Whakapapa Formation. This formation represents the most recent cone-building episode at Ruapehu volcano, and its deposits include young (ca. 15,000 years B.P. to present) post-glacial lava flows, autoclastic breccias and pyroclastic beds (Table 1).

The Pinnacle Ridge Member is preserved as three lobes, up to 25 m thick, mantling and unconformably overlying Te Herenga Formation on the flanks and crest of Pinnacle Ridge (Figs. 1, 2) near the source vent. We correlate this tuff with Okupata Tephra based on its distribution and preservation on the northern slopes of Ruapehu.

Pourahu Member

Pourahu Member, one of 23 defined members of the Bullot Formation (Table 2), comprises two tephra units, a pumiceous ignimbrite (defined from the type locality in Rangipo Desert east of Ruapehu) and a fall unit (defined from the type section at Desert Road Section 16; Fig. 1; Donoghue et al. 1995b). This member has an estimated age of ca. 11,850–10,000 years B.P. (Tables 2, 3) based on its stratigraphic position relative to local andesitic and distal rhyolitic (Waiohau Tephra ca. 11,850 years B.P.) tephra marker beds on the eastern Ruapehu ring plain (Donoghue et al. 1995b).

Sampling and analytical procedures

Correlation and tephra volumes

We mapped in detail the distributions of Okupata Tephra (Topping 1973) and the Pourahu Member (fall unit; Donoghue et al. 1995b) across the northern and eastern Ruapehu and Tongariro ring plains, describing and measuring 29 sections to establish the stratigraphic relationships and chronology of these units and to construct tephra isopachs.

Following Wilson's (1993) methodology, isopachs presented in this paper are based on the thickness of primary tephra only, thus identifying "volcanic events". Overlying palaeosols, which may include additions of loess and distal tephra from other sources or which may represent a weathering interval during which material was lost, are excluded from measurement and formation definitions. Where the upper contacts of tephras

Fig. 2a, b Pinnacle Ridge viewed from the northwest, showing lobate Pinnacle Ridge Member (Pinnacle Ridge tuff) deposits unconformably overlying older Te Herenga Formation lava flows. Location of **a** proximal and **b** distal sections through this member are from Hackett and Houghton (1989)

are gradational (thickness of transition is >0.1 m) with overlying andesitic ash or palaeosols, it is difficult to obtain accurate thickness data; therefore, we delineate boundaries using the apparent maximum rate of change in the concentration of pyroclasts with depth.

Tephra volumes are calculated using the log thickness vs area$^{1/2}$ methodology of Fierstein and Nathenson (1992), where:

$$\text{Tephra volume} = \frac{2T_0}{k^2}$$

where T_0 = extrapolated tephra thickness at $A=0$; where A = area enclosed by a selected isopach, and k = slope of the line (expressed as *-k*) on a ln T vs $A^{1/2}$ plot.

Geochemical analysis

Tephras were sampled from type or reference sections and other sites where correlation is established (Figs. 1, 3). Samples were cleaned by treatment with acid-oxalate following the reagent preparation of Blakemore et al. (1987). Samples for X-ray fluorescence analysis were further washed in hot HCl to remove surface cations and sieved at 2 mm to separate the largest pumice lapilli, which are most likely to represent the original magma composition (Paulick and Franz 1997).

Major elements of whole-pumice samples were determined by X-ray fluorescence spectrometry on fused glass disks using a Philips PW 1400 (Philips, Best, The Netherlands) automated spectrometer at the Department of Earth Sciences, La Trobe University, Australia, and the methods of Norrish and Hutton (1969) and Palmer (1990). Precision for each major and minor element (Si, Ti, Al, Fe, Mn, Mg, Ca, Na, K, P and S) is better than ±1% (1 SD) of the reported value. For some samples, due to limited sample size, we analysed only whole-rock samples comprising both pumice and lithics (accessory and juvenile).

Trace elements were determined by X-ray fluorescence spectrometry on pressed powder pellets using a Siemens 303 spectrometer (Siemens, Erlangen, Germany) and methods described by Norrish and Chappell (1977). Precision varies from 2 to 10% with detection limits generally approximately 1 ppm (Price et al. 1992; Stewart et al. 1996).

Clean pumice lapilli were impregnated under vacuum with epoxy resin and thin sectioned. The glass compositions of pumice were determined by wavelength-dispersive electron microprobe analysis using the JEOL-733 Superprobe (JEOL, Tokyo, Japan) housed in the Analytical Facility of Victoria University of Wellington (New Zealand). Instrument analysing conditions used were an 8-nA beam current at 15 kV and a 10-μm beam diameter. Analysis of the glass standard KN-18 provided a check on probe performance. Ferromagnesian minerals were identified from their optical properties in polished thin sections and major element chemistries were determined by electron microprobe analysis.

Distal rhyolite tephras

Distal rhyolitic tephras, erupted from the more northern Taupo and Okataina volcanic centres, are identified throughout the study area on the basis of their stratigraphic positions relative to known and dated andesitic tephras, ferromagnesian mineral assemblages and major element glass chemistries as determined by electron microprobe analysis (Donoghue 1991; Donoghue et al. 1995b).

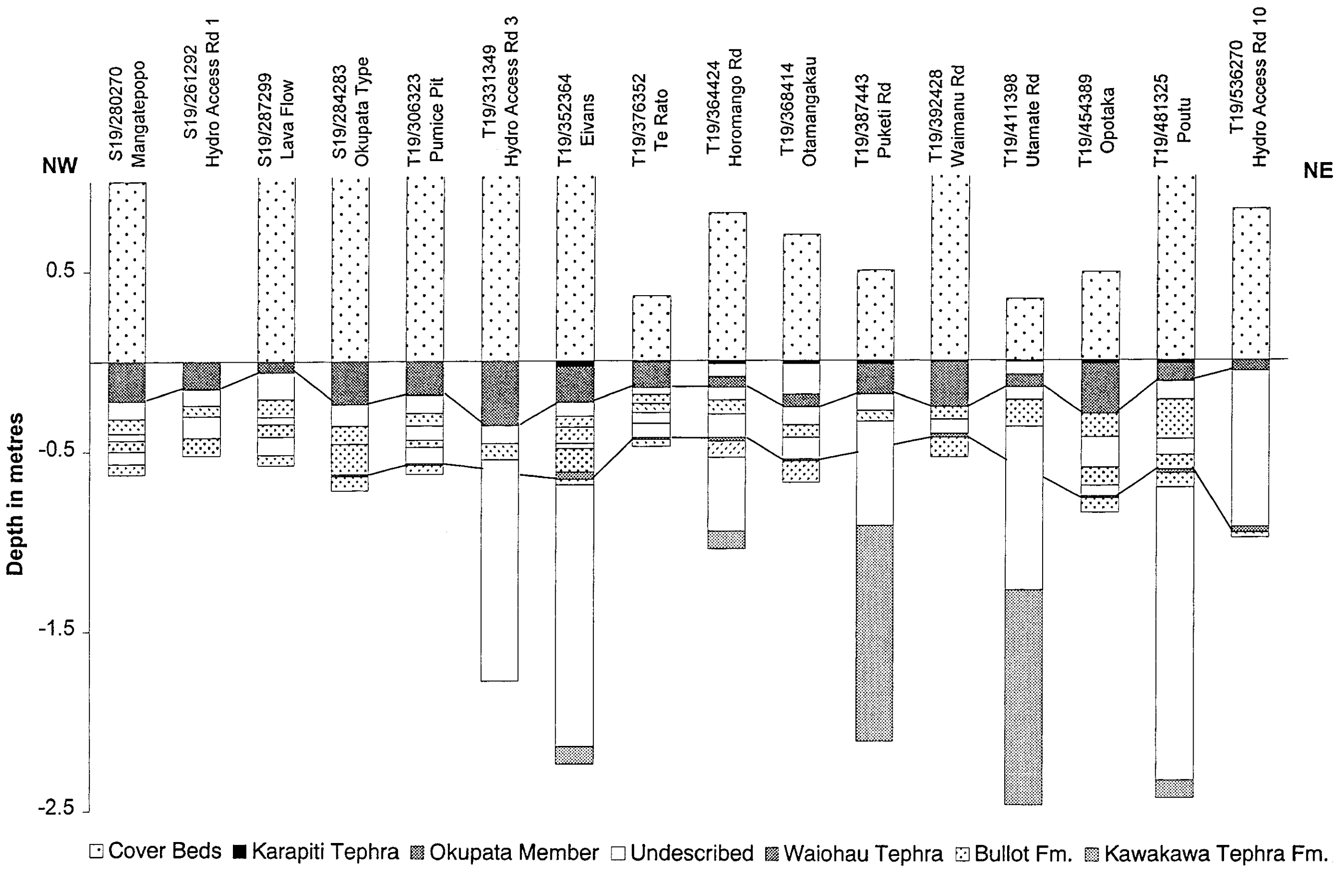

Fig. 3 Summary of Okupata Member stratigraphy across the Ruapehu and Tongariro ring plains. Datum line is placed at the base of coverbeds overlying Okupata Member. Distal rhyolitic tephras used to constrain the age of Okupata Member (Karapiti Tephra, Waiohau Tephra) are also shown. For reference, *correlation lines* define the base of Okupata Member and Waiohau Tephra. Kawakawa Tephra Formation (dated at ca. 22,590 years B.P.; see Table 2) defines the base of the Bullot Formation (see text). Grid references are based on the NZMS 260 Map Series Sheets S19, S20, T19 and T20

Remanent magnetism

The natural remanent magnetism (NRM) of the Pourahu deposits was initially investigated in the field using a portable fluxgate magnetometer. Hoblitt and Kellog (1979) compared field and laboratory measurements and concluded that this rapid field technique could be used with "a fair degree of confidence" without recourse to more time-consuming laboratory studies. At two sites, ten of the largest clasts were measured with an FG Electronics BR-2 portable magnetometer. At each locality, some of the clasts were magnetically aligned (implying a TRM acquired when the deposit was emplaced above the Curie Point) and some showed random magnetic directions (implying deposition after clasts had cooled to close to ambient temperatures). The field results were of sufficient interest to initiate a more detailed laboratory investigation of the Mangaturuturu exposure.

Oriented samples of eight pumice clasts, two pumice bombs and 19 lithic clasts were taken from the lower lithic-rich and the upper pumice-rich deposits at the Mangaturuturu exposure (Fig. 1) for detailed laboratory study of their remanent magnetism. Clast dimensions ranged between 20 and 150 mm. In the laboratory, cores were cut from the blocks for palaeomagnetic analysis. Remanence components were determined by progressive thermal demagnetisation of the samples in a Magnetic Measurements (Aughton, Lancashire, UK) furnace with a residual magnetic field of less than 5 nT at increments of 50 to 600 C; remanence was measured using a Minispin (New Castle-upon-Tyne) spinner magnetometer or a Cryogenic Consultants (London, UK) cryogenic magnetometer after each thermal treatment. Magnetic components were determined using the IAPD palaeomagnetic analysis package written by T. Torsvik. Magnetic mineralogy was determined by analysis of unblocking temperature spectra obtained from thermal demagnetisation data and from the variation of magnetic susceptibility with temperature obtained using a CS2 furnace attachment to a KLY2 Kappa Bridge.

Results and discussion

Tephrostratigraphy

Pourahu Member, Okupata Tephra, and enclosing palaeosols and andesitic tephra units have been mapped in detail across the Ruapehu and Tongariro ring plains to determine their stratigraphic relationships (Fig. 1).

Detailed mapping demonstrates stratigraphic equivalence and correlation of Okupata Tephra (Topping 1973) to Pourahu Member (fall unit). This correlation necessitates redefinition of these units and revised interpretations of previously published isopach data. Redefinition is further warranted given our new understanding of the significance of these tephras in the volcanic record at Ruapehu. We thus define the new Taurewa Formation and define two new members.

Taurewa Formation

Taurewa Formation is a new name for a sequence of tephra-fall and pyroclastic flow deposits erupted from Ruapehu volcano. The formation is named from the settlement of Taurewa, 17 km northeast of National Park on State Highway 47.

The Formation is dated between ca. 11,850 and 9500 years B.P. based on the ages of two distal rhyolitic tephras found interbedded with andesitic tephras that bracket the formation. Karapiti Tephra (radiocarbon dated at 9820 ± 80 years B.P.; Froggatt and Lowe 1990) occurs above the formation, and Waiohau Tephra (radiocarbon dated at 11,850 ± 60 years B.P.; Froggatt and Lowe 1990) occurs below it (Fig. 3; Table 2; Donoghue et al. 1995b). Based on the closer stratigraphic proximity of Taurewa Formation to Karapiti Tephra, we assign Taurewa Formation an age of ca. 10,500 years B.P.

The formation comprises two newly defined members; Pourahu Member which comprises all tephra-flow deposits, and Okupata Member, which comprises all lithologically similar tephra-fall deposits of equivalent age.

Okupata Member

We redefine and rename Topping's (1973) Okupata Tephra as a formal member, Okupata Member of Taurewa Formation. Topping's (1973) type section for Okupata Tephra on the northern Ruapehu ring plain is retained here for Okupata Member and replaces that defined by Donoghue et al. (1995b) for Pourahu Member (fall unit). Reference sections are defined at Lava Flow Section, Eivans Section, Otamangakau Section and Te Rato Section (Figs. 1, 3), where the stratigraphy of the member units and their relative stratigraphic positions to dated rhyolitic tephra marker beds are evident.

Okupata Member is typically of lapilli grade. At the type section two lapilli beds (Ok_1 and Ok_2) are identified (Fig. 4), both 0.12 m in thickness, and the contact between them is gradational. They comprise fine and medium pumice lapilli and dark grey, fine andesite lithic lapilli with minor dark-brown greasy ash matrix. Pumice lapilli in the lower bed (Ok_1) are characteristically stained orange and cemented by iron oxide. Scattered banded pumice lapilli, comprising pale yellow and grey pumice and similar to those identified in the Pourahu Member (Donoghue et al. 1995a; this paper), are identified in this member.

Okupata Member is overlain in turn by a palaeosol developed in medial deposits (representing intermittent accumulation of volcaniclastic material), Pahoka Tephra (erupted from Tongariro volcano ca. 10,000–9800 years B.P.) and unnamed ash in which the

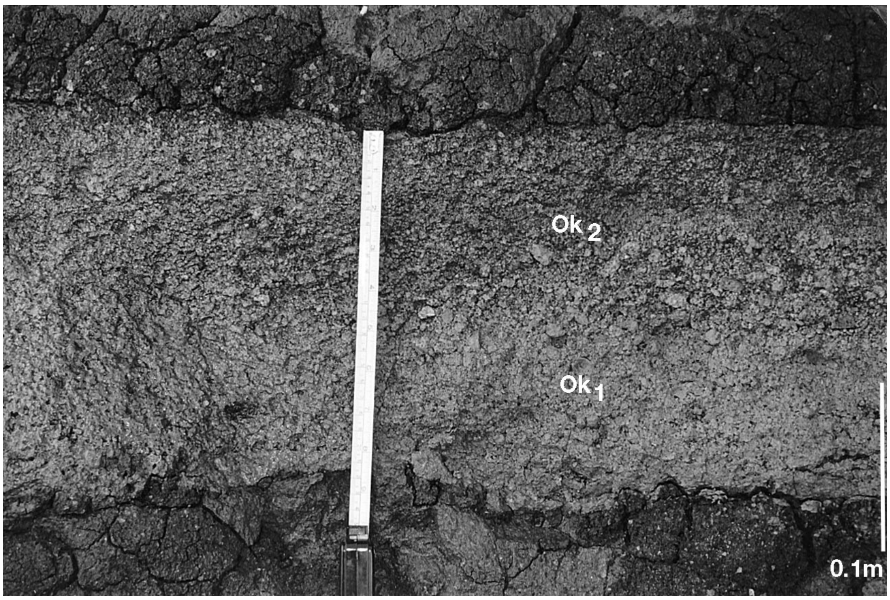

Fig. 4 Okupata Member at the type section S19/284283, indicating stratigraphic positions of units Ok_1 and Ok_2. Bracketing the Okupata Member are prominent palaeosols developed in medial ash

distal rhyolitic Karapiti Tephra is interbedded (Table 3). The term "medial" is defined by Soil Survey Staff (1990) as fine earth with andic soil properties, in which rock fragments make up less than 35% by volume. Topping (1973) included this overlying palaeosol in his definition of Okupata Tephra, but on the basis of Wilson's (1993) arguments regarding the inclusion of palaeosols in formation definitions, we exclude this from our definition of Okupata Member. The palaeosol does, however, aid correlation by broadly defining the upper stratigraphic limits of Okupata Member at more distal sites where it is of ash grade. Okupata Member overlies a palaeosol above a sequence of unnamed Ruapehu eruptives (Bullot Formation tephras; Table 2) within which a chronostratigraphic marker bed, the Waiohau Tephra (ca. 11,850 years B.P.), is interbedded.

North and east of the type section Okupata Member lapilli beds thin and occur variously as fine lapilli or predominantly coarse ash, with individual units not so readily discerned. Where Okupata Member occurs only as a coarse ash we are unable to determine unequivocally if this ash represents the distal equivalent of only the upper lapilli bed seen at the type section, or both beds.

At our most northeastern section, Hydro Access Road 10, where there are five distinct units within the ca. 11,000–10,000 years B.P. stratigraphic interval, we correlate the central 0.06 m coarse pumiceous ash unit to Okupata Member.

In sections along SH1 south of Hydro Access Road 10 to Wahianoa Aqueduct (Fig. 1), lithologically similar lapilli and ash beds occurring in the same stratigraphic interval as Okupata Member are correlated with this Member (Donoghue et al. 1995b) and designated unit Ok_3. At these sites unit Ok_3 is a reasonably prominent tephra, occurring variously as a pale yellow and black coarse pumiceous (>30 mm diameter) and crystal-rich ash, or as a lapilli bed dominantly comprising pale mostly yellow to white vesicular pumice within a coarse ash matrix.

Due to the absence of interbedded datable material we are not able to obtain a radiometric (^{14}C) age for Okupata Member. However, from the tephrostratigraphy and chronology at source, an age between ca. 11,800 and 9800 years B.P. and close to ca. 10,500 years B.P. is indicated.

Pourahu Member

We redefine Pourahu Member (ignimbrite unit; Donoghue et al. 1995a) as the Pourahu Member of Taurewa Formation (Table 2). A detailed stratigraphy and description of Pourahu Member at the type locality in the Rangipo Desert (Fig. 1) on the eastern Ruapehu ring plain are given in Donoghue et al. (1995a, 1995b). Banded clasts representing mingling of light and dark andesite (Donoghue et al. 1995a, 1995b) are a distinctive feature of the deposit.

We recognise a lithologically similar deposit at another site, on the western Ruapehu ring plain near the headwaters of Mangaturuturu River. The deposit is partially exposed in the bed and banks of the river. Two distinct lithological units are recognised, an upper (0.5 m depth) pumice-rich lithofacies and a stratigraphically lower lithic-rich lithofacies. The upper deposit is a matrix-supported pumice-rich diamicton. The dominant clasts are pale-yellow pumice lapilli. Also present are grey-white colour-banded pumice lapilli (20% by volume of clasts), scattered radially jointed pumice bombs that are also colour-banded and indicative of magma-mingling processes (Donoghue et al. 1995a), and few (5% by volume) white and red andesite clasts. The matrix is a pumiceous and lithic sand. The lower (1.5 m) deposit is a matrix-supported lithic-rich diamicton comprising 60–80% (by volume of clasts) andesite lithics (pebbles to cobbles) together with pumice clasts and a few colour-banded breadcrusted pumice bombs. The base of this deposit is not exposed.

At the type locality, Pourahu Member has an age of ca. >10,000 years B.P. (Table 2). We are unable to better constrain its age based on the stratigraphy of the Mangaturuturu deposit.

Other deposits associated with Taurewa Formation

Hackett and Houghton (1985) correlate proximal Pinnacle Ridge tuff of Whakapapa Formation to Topping's (1973) Okupata Tephra, proposing that it is a coeval, welded, primary tephra equivalent.

In proximal sections (1–2 km from vent; Figs. 1, 2), Pinnacle Ridge tuff deposits are densely welded (Hackett and Houghton 1985). We identified few pale yellow and white pumice clasts (10–60 mm) within the proximal deposits; large (up to 0.5 m diameter) grey and red accessory andesite blocks dominate, with fewer dark-brown juvenile lithic clasts and some radially jointed lithic blocks, all in a moderately hard, welded crystal ash matrix.

Although the pumice clasts are relatively uncommon, they are similar in appearance to Pourahu Member pumice (phenocryst rich and pale coloured) but are distinctly more weathered, crumbling when extracted from the welded ash matrix. In distal sections (Fig. 2) Pinnacle Ridge tuff consists of massive, poorly bedded, moderately sorted lapilli and bomb deposits, dominated by highly vesicular scoria, with subordinate angular lithic blocks derived from the underlying Te Herenga Formation and breadcrusted juvenile blocks (Hackett and Houghton 1985, 1989).

Correlation with Topping's (1973) Okupata Tephra is based on the distribution of the tuff deposits across the northern slopes of Ruapehu and extrapolation of Okupata Tephra isopachs to identify a source in the vicinity of Pinnacle Ridge. Hackett and Houghton (1985,

1989) provide no additional information to support this correlation. No age is available for the Pinnacle Ridge tuff, and the petrology of the deposits was not described. We agree, however, that the distribution of Pinnacle Ridge tuff and the isopachs of Okupata Tephra (Topping 1973) support their correlation and therefore we refer to Pinnacle Ridge tuff as an associated deposit of Okupata Member.

These tuff deposits, although of restricted distribution, provide an important link between the parent volcano, where little evidence remains of explosive volcanism, and distal records of explosive volcanism (Hackett and Houghton 1985). They also provide useful insights into the nature of the Taurewa Formation eruptions.

Genesis of Pourahu Member

Donoghue et al. (1995a, 1995b) cite both sedimentologic and lithologic evidence to support an origin from a pyroclastic flow. The most significant observations are: overall poor sorting; the occurrence of juvenile lithic lapilli in a poorly sorted crystal-rich ash; radially jointed vesiculated pumice bombs; the pink colouration of some lapilli and blocks indicating a high-temperature origin and suggesting that at least this component was hot; and the near monolithological composition of the deposit, which contrasts with the strong heterolithological nature of other sediment-flow deposits at Ruapehu.

Some sedimentological features observed in the type section (Fig. 5a, b) are, however, more difficult to interpret. For example, the main part of the deposit (Fig. 5a) is weakly stratified and contains numerous fine laminations and sandy intercalations. It laterally interfingers with heterolithological, lithic-dominant deposits interpreted to be deposits of hyperconcentrated sediment flows and debris flows (Fig. 5b). Possibly, therefore, the Pourahu Member deposit originated by other mechanisms or flow types. Many sedimentological features of pyroclastic flow deposits (e.g. poorly sorted, structureless deposits, matrix support of large clasts, presence of fine-grained basal layers) are similar to those found in deposits from other high-particle concentration sediment gravity flows, such as debris flows, and it can therefore be difficult to distinguish pumice-rich debris flows from pyroclastic flows (Sparks 1976; Carey 1991).

From close examination of the deposit, we believe that the internal stratification is indicative of either several flow units (Sparks 1976; Fisher and Schmincke 1984) preserved within the main body of the deposit or post-depositional reworking. The laminations and sandy intercalations suggest partial reworking of primary but distal flow deposits either during or following emplacement (Buesch 1991).

The interfingering of the deposit with sandy hyperconcentrated flow and debris flow deposits (Fig. 5b) also raises the question as to whether it is a genetically associated, pumice-rich, debris flow deposit. We believe, however, that the occurrence of perfectly preserved pumice bombs (with jigsaw jointing and of Pou-

Fig. 5a, b Distal pyroclastic flow deposits of Pourahu Member exposed at the type locality. **a** Overlying, discontinuous grey Tangatu Formation hyperconcentrated flow deposits, Holocene tephra beds and young (<ca. 1850 years B.P.) lahar deposits; **b** pumice-rich Pourahu Member pyroclastic flow deposits exposed farther northwest up channel toward source. Note the interfingering of Pourahu Member deposits with sandy lithic-dominant hyperconcentrated flow deposits containing rafts of Pourahu Member pumice

a **b**

rahu Member lithology) embedded within these sandy deposits suggests that they are small debris flows generated at the distal margins of a pyroclastic flow where there is a transition from thin pyroclastic flow units to lahars (Pérez-Torrado et al. 1997). It seems unlikely that the pumice bombs would be preserved unless the debris flows were generated proximal to a pyroclastic flow and contained hot clasts. Later movement of cooled clasts generally causes disintegration along radial thermal contraction joints (Hoblitt and Kellog 1979; Smith and Lowe 1991).

Similarly, at the Mangaturuturu exposure the lithic-rich basal facies could be interpreted as the deposit of a debris flow but possibly also as a lithic-rich basal facies of a pyroclastic flow. Sparks et al. (1997b) describe deposits from pyroclastic flows as comprising both pumice-rich facies and lithic-rich facies, where the pyroclastic flows are inferred to have segregated based on density differences during movement. Large lithic clasts found in lithic-rich facies are believed to have been eroded from the substrate (such as unconsolidated colluvium and talus), and the pumice-rich facies typically occur at the margins and distal parts of the pyroclastic fans (Sparks et al. 1997b).

We have used remanent magnetism techniques to better interpret the emplacement mechanisms of the type locality and Mangaturuturu deposits.

Remanent magnetisation of Pourahu Member deposits

Laboratory analysis of thermoremanent magnetisation (TRM) of rocks has been widely used in volcanic studies to distinguish emplacement mechanisms and to determine the temperature of emplacement of volcanic deposits (Hoblitt and Kellog 1979; McClelland and Druitt 1989; Arguden and Rodolfo 1990). Typically, pyroclastic flows contain lithic clasts of lava that were extruded, solidified and magnetised prior to their incorporation into the pyroclastic flow. The effect of a hot pyroclastic flow (but not so hot as to exceed the Curie temperature of the magnetic minerals in the lava clasts; 580 °C for magnetite) is to partially remagnetise the remanence in the lava clasts, giving rise to a two-component remanence where the lower unblocking temperature (T_{ub}) component was acquired during cooling after deposition of the flow, and the higher T_{ub} component was retained from the original formation of the lava. The low T_{ub} component will be uniformly directed in all clasts, and the higher temperature component will be randomly directed between clasts. The laboratory temperature to which samples have to be heated to remove the low temperature, uniformly directed magnetic overprint reflects the equilibrium temperature of the pyroclastic flow. Cold deposition will lead to random orientations of remanence between lithic clasts (Hoblitt and Kellog 1979; Smith and Lowe 1991). Remanence in pumice in primary pyroclastic flow deposits carries only one component of remanence parallel to the ambient Earth's magnetic field at the time of deposition (Zlotniki et al. 1984; McClelland and Druitt 1989).

Field measurements of remanence directions in volcanic deposits using a fluxgate magnetometer have been used to identify hot pyroclastic flows (Crandell and Mullineaux 1973; Miller 1978). However, when the flow temperature is below 580 °C, not all of the remanence in lithic clasts within the volcanic deposits will have been thermally reset into the ambient magnetic field direction, and a randomly oriented higher unblocking temperature component will be retained. The magnetic vector in such clasts, termed Natural Remanent Magnetisation or NRM, will not be well grouped if the inherited randomly oriented component is of significant magnitude. Randomly oriented field measurements of NRM directions are reported from deposits the origins of which are attributed to emplacement by pyroclastic flows and hot lahars (Crandell and Mullineaux 1973), suggesting that field measurements need to be interpreted with care and should be backed up by laboratory analysis of remanence components.

Furthermore, the juvenile component (pumice or scoria) may carry a chemical remanent magnetisation (CRM) acquired at a time significantly after deposition when new magnetic minerals were formed by authigenic formation or mineral transformations (McClelland and Druitt 1989; McClelland 1996). If a field fluxgate magnetometer study were to be carried out on a deposit from a cold flow dominated by juvenile material carrying a secondary CRM, uniform magnetisation would be observed that would lead to an erroneous interpretation that the deposit was a hot pyroclastic flow. This is important, because a significant proportion of the natural remanence found in igneous rocks is now believed to be due to secondary CRM magnetisation rather than a primary process such as TRM magnetisation (McClelland 1996; Valet et al. 1998).

Results of Fluxgate–NRM

In 1995, before the September 1995 to June 1996 eruptions at Ruapehu, we undertook preliminary field NRM measurements of pumice blocks within the flow deposits at both exposures. At the Mangaturuturu exposure, fluxgate NRM measurements indicated that pumice clasts in the upper pumice-rich facies had strong magnetic alignment and that the lithic clasts in the lower lithic-rich facies had random NRM orientations. At The Chute, the NRM of pumice clasts in the upper unit showed that they were magnetically aligned, but the large pumice clasts found throughout the deposit and also smaller pumice clasts sampled from the middle and lower units showed variable, though not entirely random, NRM values.

Our initial interpretations from the fluxgate–NRM data were that, in the Mangaturuturu deposit, the aligned NRM measurements suggested that the pumice-rich deposit was emplaced hot and represented a primary pyroclastic flow deposit. The random NRM orientations in the lithic clasts suggested that they could have been eroded from the substrate and entrained in the base of the flow (Sparks et al. 1997b), and were thus accidental clasts.

Similarly, our initial interpretation of the magnetic alignment of pumice clasts in the upper unit at The Chute was that the deposits were primary and were emplaced by a pyroclastic flow. The range in NRM of pumice clasts sampled from the middle and lower units suggests the deposits were either the distal portions of a pyroclastic flow that had cooled considerably below the blocking temperature during transport or were the result of post-depositional reworking.

However, our NRM data need to be interpreted with caution because of the problems outlined previously; thus, we cannot argue conclusively that at these two localities Pourahu Member deposits were emplaced by a pyroclastic flow. The laboratory analysis discussed in the next section modified our conclusions for the Mangatururturu deposit.

Results of laboratory analysis of remanence

Compared with the uncertainties in the field-based NRM work, the results of our laboratory palaeomagnetic analysis allow us to very clearly determine the genesis of the Mangatururturu deposit. Remanence in the lithic clasts from both the upper pumice-rich layer (0.5 m thick) and the lower lithic-rich layer (1.5 m thick) is usually two-component, indicating that the clasts have been partially heated at some time, but both low- and high-temperature components are randomly distributed (Table 4). This random distribution indicates that the clasts have not been heated in the deposit in situ. Both units are sufficiently thick for the small lithic clasts to have reached thermal equilibrium before significant cooling of the deposit occurred; hence, the tephra was cold when it was deposited. The two-component nature of the remanence in the lithic clasts implies that they were once part of a hot pyroclastic flow but have been reworked after the hot flow cooled and the low temperature component was acquired.

Our data show that the strong magnetic alignment of the pumice clasts in the upper unit of the deposit, determined by fluxgate measurements, is in fact caused by a post-emplacement formation of titanomagnetite (Curie temperature of ca. 300 °C) that carries a CRM, not by a primary TRM acquired during cooling of a hot flow. The magnetic mineralogy of the pumice clasts is dominated by titanomagnetite, which carries a well-grouped component of magnetisation close to the present Earth's magnetic field direction (Table 4). However, there is a high temperature, primary remanence carried by magnetite just identifiable in eight clasts, which is determined to be randomly oriented from a great-circle analysis. This suggests that the pumice clastss came to rest and cooled somewhere else (acquiring a remanence in magnetite parallel to the Earth's field) and were then reworked to their present position. The deposit is therefore reworked, and the pumice clasts have been remagnetised by post-deposition chemical weathering. Features of the Mangatururturu deposits consistent with at least partial reworking include: a high proportion (80%) of large, heterolithological non-juvenile clasts; fragmented andesitic bombs; a lithic and pumiceous ash matrix; and a transitional zone containing up to 50% heterolithological lithic clasts.

We thus conclude that the deposit at Mangatururturu was emplaced by a pumice-rich debris flow and can refute a primary pyroclastic flow origin. The field-based fluxgate measurements, which suggest the existence of primary pyroclastic flow deposits, are flawed because they cannot discriminate between uniformly directed primary TRM acquired during cooling and uniformly directed CRM acquired after deposition.

We are not able to undertake a full palaeomagnetic analysis of Pourahu Member pumice clasts from the type locality (The Chute), because access to the area (Rangipo Desert) on military land has been restricted. Although our field NRM data are inconclusive in determining the genesis of the deposit at the type locality, we argue in favour of an origin from a small-volume pumice flow based on the field characteristics of the deposits preserved at this locality. The abundance of low-density pumice of high-silica composition, the large quantity of loose fine ash in the matrix of the deposit, and low abundance of juvenile andesite lithic clasts is

Table 4 Magnetisation components of Pourahu Member clasts from Mangatururturu section. *T* temperature; α cone of 95% confidence about mean direction; *N* number of specimens; *R* length of resultant sum of *N* vectors; *k* precision parameter

Lithology	Component	Declination	Inclination	α_{95}	N	R	k	Grouping
Pumices	Low T	331.7	–36.5	20.8	10	8.59	6.4	Grouped
	High T	Unresolvable			8			Random (great circles)
Lithics (1)	Low T	134.5	–50.4	90.0	6	2.7	1.5	Random
	High T	59.8	–19.5	180.0	7	1.45	1.1	Random
Lithics(2)	Low T	254.4	–60.1	70.2	10	3.74	1.4	Random
	High T	4.0	–46.8	180.0	3	1.45	1.3	Random

Lithics (1) from pumice-rich upper layer; lithics (2) from lithic-rich lower layer consistent with such a genesis (Crandell and Mullineaux 1973; Crandell et al. 1984; Fisher and Schmincke 1984; Carey 1991). Features of the Mangaturuturu deposits that are consistent with reworking (e.g. a high proportion of large heterolithological non-juvenile clasts and fragmented andesitic bombs) are not observed in the type locality.

The eruption of Pourahu Member

Petrology and geochemistry of Pourahu Member

The petrology and geochemistry of Pourahu Member at the type locality is described in detail in Donoghue et al. (1995a). A summary of these data are presented herein, together with additional data obtained from the newly identified Mangaturuturu deposit.

At the type locality three compositionally different types of pumiceous clasts are identified within the flow deposits. In order of abundance they are: light-coloured vesicular andesite (58–61% SiO_2; all data normalised to loss free) with more siliceous rhyolitic (71–73% SiO_2) groundmass glass compositions; dark-coloured vesicular andesite (60% SiO_2), with quench olivine and andesitic to dacitic (60–68% SiO_2) groundmass glass compositions; and fewer clasts consisting of interbanded light and dark andesite (Fig. 6a; see Table 6; Donoghue et al. 1995a). The colour variations in these pyroclastics are attributed to mingling of two compositionally different magmas (Donoghue et al. 1995a). The ferromagnesian mineral assemblage of pumice clasts comprises orthopyroxene (dominant), clinopyroxene and rare quench olivine. Whole-pumice compositions of these clasts are andesite (Donoghue et al. 1995a).

The deposit at Mangaturuturu River is of similar lithology. Blocks of light pumice dominate, and the groundmass glass chemistry of these blocks is rhyolitic (71% SiO_2; Fig. 6a; Table 5). Within the groundmass glass are phenocrysts of orthopyroxene, clinopyroxene, plagioclase and titanomagnetite. Also shown (Figs. 6a, 7; Table 6) are whole-pumice compositions of loose Pourahu Member pumice blocks found on the eastern slopes of Ruapehu and sampled from a small pyroclastic flow deposit in a tributary of the Mangatoetoenui Stream (R. Price and J. Chapman, pers. commun.; see discussion below). Compositions are basaltic andesite and andesite.

Donoghue at al. (1995a) modelled the eruption of Pourahu Member based on petrological investigations and geochemical analysis of pumice clasts from type locality deposits. This model proposed that injection of a small volume of hot andesite magma into a magma chamber beneath Ruapehu volcano triggered magma vesiculation and eruption. Vesiculation of a cooler, less dense melt brought about by injection of hot mafic magma into the base of a magma chamber has been implicated as a mechanism for triggering explosive eruptions at numerous other volcanic centres (Sparks et al. 1977; Gourgaud et al. 1989; Venesky and Rutherford 1997).

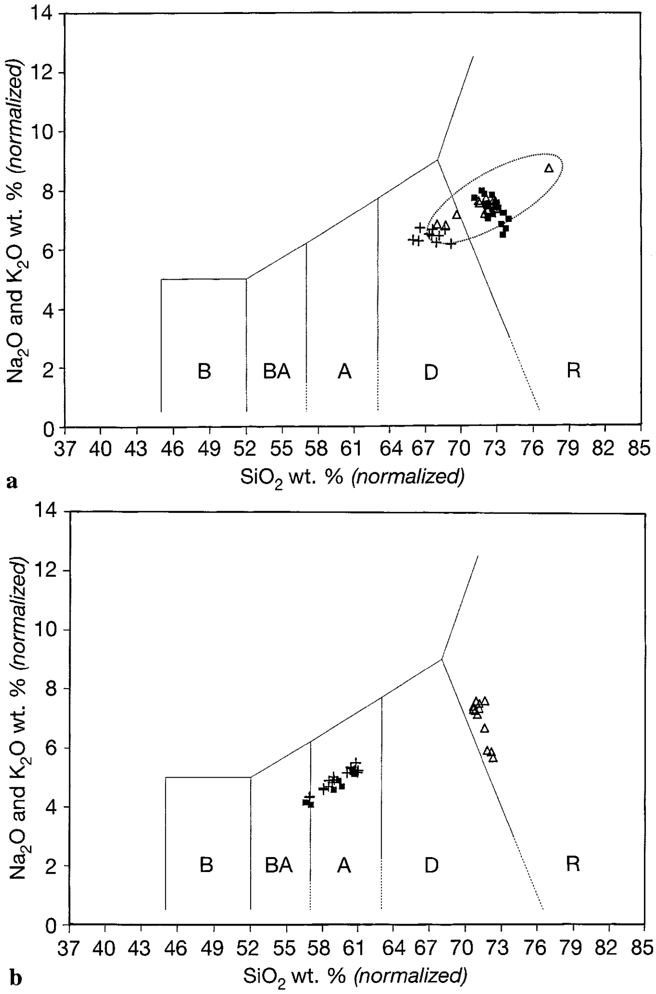

Fig. 6 Total alkali silica diagram (after Le Bas et al. 1986 and Le Maitre 1984) showing **a** whole-pumice and whole-rock compositions of Pourahu Member (*crosses*) and Okupata Member (*squares*) and glass compositions of Pourahu Member at Mangaturuturu (*triangles*); **b** glass compositions of Okupata Member unit Ok_1 (*triangles*; outlined), Okupata Member unit Ok_3 (*squares*) and a more distal coarse ash (*crosses*) sampled from Hydro Access Rd. 10 section and correlated to Okupata Member. All data normalised to a loss-free basis prior to plotting. See text for explanation of sample sites. Data from Donoghue (1991); Donoghue et al. (1995b); R. Price and J. Chapman (unpublished data); this paper. Compositional fields are basalt (*B*), basaltic andesite (*BA*), andesite (*A*), dacite (*D*) and rhyolite (*R*)

Distribution and volume of Pourahu Member

The distribution of in situ Pourahu Member deposits indicates that at least one pyroclastic flow was generated and directed east down the Whangaehu Valley (Fig. 1). Based on the thickness and restricted distribution of this deposit, we propose that the flow was a

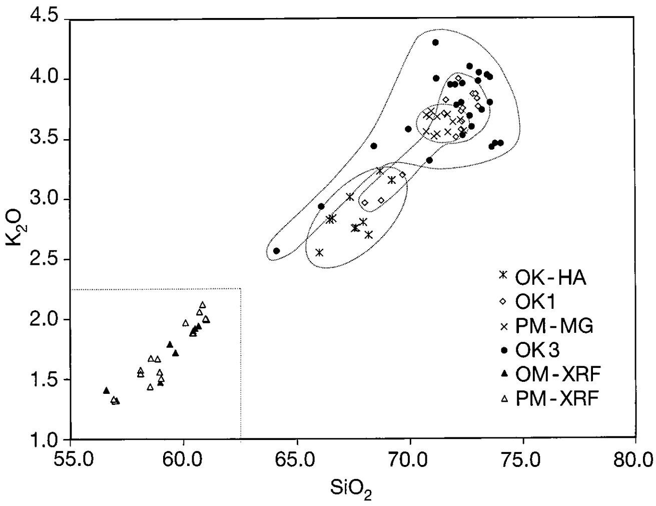

Fig. 7 SiO_2 vs K_2O variation diagram showing compositional fields defined by whole-pumice, whole-rock and glass composition data for Pourahu Member and Okupata Members. *Triangles* whole-pumice and whole-rock data (XRF field defined in *lower left corner* of graph). All other symbols indicate glass data. *PM* Pourahu Member; *OM* Okupata Member; *OK3* Okupata Member unit OK_3; *OK1* Okupata Member unit Ok_1; *OK–HA* Okupata Member sampled from Hydro Access Rd. 10 section; *PM–MG* groundmass glass of Pourahu Member pumice clasts sampled from Mangaturuturu Stream section. All data normalised to a loss-free basis prior to plotting. See text for explanation of sample sites. Data from Donoghue (1991); Donoghue et al. (1995b); R. Price and J. Chapman (unpublished data); this paper

small- to medium-volume channelised flow (Sparks 1976; Carey 1991) that, as it emerged onto the lower-lying ring plain, may have expanded laterally and segregated into several thin-flow units (Pérez-Torrado et al. 1997), reducing opportunities for its preservation. We have not found more proximal (<10 km from vent) flow deposits, suggesting that most of the deposit has been lost through erosion. Streams active at that time would have eroded much of the deposit from ring plain surfaces as well as the lower flanks of the volcano. Attesting to this erosion is the widespread occurrence of reworked Pourahu Member deposits within younger (ca. 10,000–8500 years B.P.) volcaniclastic sequences, principally within lahar and debris avalanche deposits, in catchments west, north and east of Ruapehu. Such erosion is also indicated by thick pumice-block deposits found on the eastern slopes of Ruapehu (near Tukino), where the source appears to be in the Mangatoetoenui headwaters, on the southern slopes (near Turoa; J. Gamble, pers. commun.) and on Holocene surfaces south of Tangiwai near Whangaehu River.

This distribution of reworked Pourahu Member pumice suggests that the primary deposits were once more extensive than is indicated by the present exposures, and that there may have been smaller flows down several major catchments in addition to the main flow that became channelised and deposited east of the volcano. At least one flow was directed down the Mangaturuturu Valley, being emplaced near the headwaters of Mangaturuturu River, and another entered the headwaters of the Mangatoetoenui River. We propose that, at some time following emplacement, these pyroclastic flow deposits were readily eroded and remobilised over short distances as cold pumice-laden debris flows, possibly generated by the melting of summit glacial ice by hot pyroclastic ejecta (Mothes et al. 1998).

The limited preservation and exposure of primary Pourahu Member deposits on the ring plain unfortunately precludes detailed mapping of this member and an estimation of its volume.

The eruption of Okupata Member

Okupata Member is represented by three tephra deposits preserved on the ring plain and the Pinnacle Ridge tuff deposits preserved on the northern flanks of

Table 5 Electron microprobe analyses of groundmass glass in members of Taurewa Formation. *n* number of analyses

	Pourahu Member ($n=12$)[a]	Okupata Member Ok_1 ($n=13$)[b]	Okupata Member Ok_1 ($n=3$)[c]	Okupata Member Ok_3 ($n=10$)[d]	Okupata Member OK_3 ($n=10$)[e]	Okupata Member OK_3 ($n=9$)[f]
SiO_2	71.32 (0.58)	71.42 (1.61)	74.17 (2.70)	67.52 (1.00)	72.40 (0.73)	73.11 (0.56)
Al_2O_3	14.32 (0.22)	14.66 (1.35)	13.24 (0.85)	16.14 (1.08)	14.08 (0.40)	13.54 (0.14)
TiO_2	0.71 (0.06)	0.53 (0.06)	0.39 (0.16)	0.79 (0.19)	0.59 (0.04)	0.55 (0.03)
FeO	3.15 (0.24)	2.61 (0.27)	2.11 (0.95)	3.86 (0.66)	2.71 (0.09)	2.73 (0.16)
MnO	0.19 (0.03)	0.20 (0.04)[g]	0.27 (0.02)[g]	0.20 (0.06)[g]	0.13 (0.02)[g]	0.00 (0.00)[g]
MgO	0.71 (0.03)	0.50 (0.09)	0.38 (0.28)	0.78 (0.14)	0.54 (0.03)	0.48 (0.05)
CaO	2.56 (0.18)	2.55 (0.91)	1.52 (1.03)	4.10 (0.59)	2.13 (0.14)	2.01 (0.20)
Na_2O	3.31 (0.70)	3.79 (0.10)	3.77 (0.22)	3.57 (0.27)	3.48 (0.35)	3.51 (0.32)
K_2O	3.62 (0.07)	3.59 (0.34)	4.07 (0.56)	2.86 (0.21)	3.75 (0.18)	3.85 (0.24)
Cl	0.17 (0.03)	0.21 (0.04)[g]	0.17 (0.05)[g]	0.24 (0.04)	0.27 (0.08)	0.22 (0.03)
Water	1.46 (0.99)	2.54 (1.73)	0.25 (0.69)	1.30 (1.14)	3.12 (2.33)	1.42 1.01

All statistics are for normalised values (to 100% loss free and meaned) above detection limit only; values in parentheses are standard deviations
[a] Mangaturuturu section
[b] Okupata-type section pale pumices
[c] Okupata-type section grey pumices
[d] Hydro Access Rd.10
[e] Desert Road Section 16
[f] Bullot Track
[g] At least one analysis gave a result below detection limit (not included in these statistics)

Table 6 Chemical analyses (whole-pumice and whole-rock) of major elements in clasts from Taurewa Formation members

	Okupata Member Ok_1[a]	Okupata Member Ok_2[b,e]	Okupata Member Ok_3[c]	Okupata Member Ok_3[c,e]	Okupata Member Ok_3[d,f]	Pourahu Member dark[f,g]	Pourahu Member dark[f,g]	Pourahu Member light[f,g]	Pourahu Member light[f,g]	Pourahu Member R95/7b[f,h]	Pourahu Member X1-P1[f,h]
SiO_2	52.15	52.55	57.62	56.85	57.73	59.77	60.01	58.63	59.70	55.85	59.43
Al_2O_3	17.98	18.28	16.98	17.40	16.99	17.01	17.07	16.50	16.18	16.52	16.31
TiO_2	0.87	0.85	0.73	0.75	0.73	0.67	0.68	0.74	0.70	0.74	0.68
Fe_2O_3	2.60	2.08	1.51	2.36	6.77	6.54	6.51	6.73	6.46	7.93	1.91
FeO	4.83	4.67	4.84	4.31	–	–	–	–	–	–	4.20
MnO	0.14	0.14	0.10	0.13	0.11	0.11	0.11	0.10	0.10	0.14	0.12
MgO	4.05	3.92	3.86	4.11	3.77	3.51	3.41	3.76	3.54	5.08	3.66
CaO	5.58	5.75	6.27	5.94	6.22	5.90	6.12	5.95	5.90	7.52	6.00
Na_2O	2.51	2.52	2.86	2.98	3.00	3.37	3.43	3.10	3 .16	2.92	3.07
K_2O	1.30	1.22	1.66	1.42	1.74	1.88	1.87	1.92	1.95	1.31	1.95
P_2O_5	0.14	0.15	0.17	0.15	0.14	0.14	0.14	0.13	0.13	0.12	0.13
SO_3	0.01	0.02	0.01	0.00	–	–	–	–	–	0.03	0.12
CO_2	2.52	2.76	1.80	1.48	–	–	–	–	–	0.05	0.30
H_2O^+	3.84	2.64	1.08	1.19	–	–	–	–	–	0.85	0.97
H_2O^-	2.25	2.41	0.55	0.55	–	–	–	–	–	0.43	0.66
LOI (volatiles)					2.50	1.40	0.76	2.57	1.64		
Total	100.78	99.96	100.04	99.61	99.69	100.30	100.10	100.30	99.50	99.50	99.51

[a] Okupata-type section
[b] Lava flow section
[c] Waikato Stream section
[d] Bullot Track
[e] Whole-rock analysis
[f] Fe_2O_3 is total iron
[g] Representative data from Donoghue et al. (1995b); LOI is loss on ignition to 1000 °C
[h] Representative data from R. Price and J. Chapman (unpublished data)

Ruapehu volcano between 1500 and 1900 m elevation. Correlation of the ring plain tephra deposits is established by detailed field mapping of the units and demonstration of stratigraphic equivalence.

Petrology and geochemistry of Okupata Member

We have analysed the major element compositions and glass chemistries of the three member units (Ok_1, Ok_2, Ok_3) to determine whether the petrology of the tephras is useful for correlation.

Unit Ok_1 has a basaltic andesite (56% SiO_2) whole-pumice composition, with two pumice types indicating a probable origin from a mingled melt. The light-coloured pumice clasts have rhyolitic glass chemistry (71% SiO_2) and contain phenocrysts of orthopyroxene (dominant), clinopyroxene, titanomagnetite and plagioclase. Similarly, the grey-coloured pumice clasts have rhyolitic glass (74% SiO_2), and the phenocryst assemblage differs only in that it contains rare quench olivine crystals that, significantly, are also identified in the "dark" component in Pourahu Member pumices (Fig. 6b; Donoghue et al. 1995a).

Unit Ok_3 has an andesite (57.8% SiO_2) whole-pumice composition (Donoghue et al. 1995b) and rhyolitic (70% SiO_2) glass chemistry. The same unit, sampled from Bullot Track, has an andesite whole-pumice composition (Donoghue 1991) and also a rhyolitic glass chemistry (72% SiO_2).

We have only whole-rock data for Unit Ok_2. The whole-rock sample contains pumice and subordinate lithic clasts (assumed to be both juvenile and accessory). Its composition is basalt to basaltic-andesite (57% SiO_2), consistent with whole-rock data for unit Ok_3 [also basalt to basaltic–andesitic (57% SiO_2)]. The similarity in whole-rock and whole-pumice compositions (Fig. 6a) suggests that either data set could be used to characterise the tephras in this case.

We have also analysed a coarse ash collected from Hydro Access Road 10 that we correlate with Okupata Member but believe represents ash deposited at the margins (off the dispersal axis) of lobes Ok_1, Ok_2 and Ok_3 and possibly therefore is a composite tephra layer. The glass chemistry of this ash is dacitic (67% SiO_2; Fig. 6b), and the ash contains phenocrysts of orthopyroxene (dominant), clinopyroxene, titanomagnetite, ilmenite and plagioclase.

The groundmass glass in pumice fragments from the welded Pinnacle Ridge tuff is highly weathered and devitrified. Analysis of the glass chemistry yields low-analysis totals (90–92%) but indicates that the glasses are siliceous (71% SiO_2; unnormalised data). Pumice clasts contain phenocrysts of orthopyroxene, clinopyroxene, plagioclase and titanomagnetite.

Overall, the whole-pumice, whole-rock and groundmass glass compositions of units Ok_1, Ok_2 and Ok_3 are similar. The whole-pumice compositions, glass compositions and ferromagnesian mineral assemblages are also similar for Okupata Member and Pourahu Member, except for a more mafic (basalt whole-pumice) component in Okupata Member. The evident compositional and stratigraphic relationships of these members strongly suggest correlative and coeval events (Figs. 6a, b, 7; Tables 5, 6).

Distribution and volume of Okupata Member (Ok_1, Ok_2, Ok_3 and Pinnacle Ridge tuff)

We present new isopach maps for Okupata Member tephras (Fig. 8a, b), based on 29 measured sections. Variations in grain size and thickness of this member suggest that it was erupted in three strongly directional, overlapping lobes, with dispersal axes to the north and northeast. The strongly directional lobes suggest Okupata Member units were erupted in pulses associated with brief plinian activity.

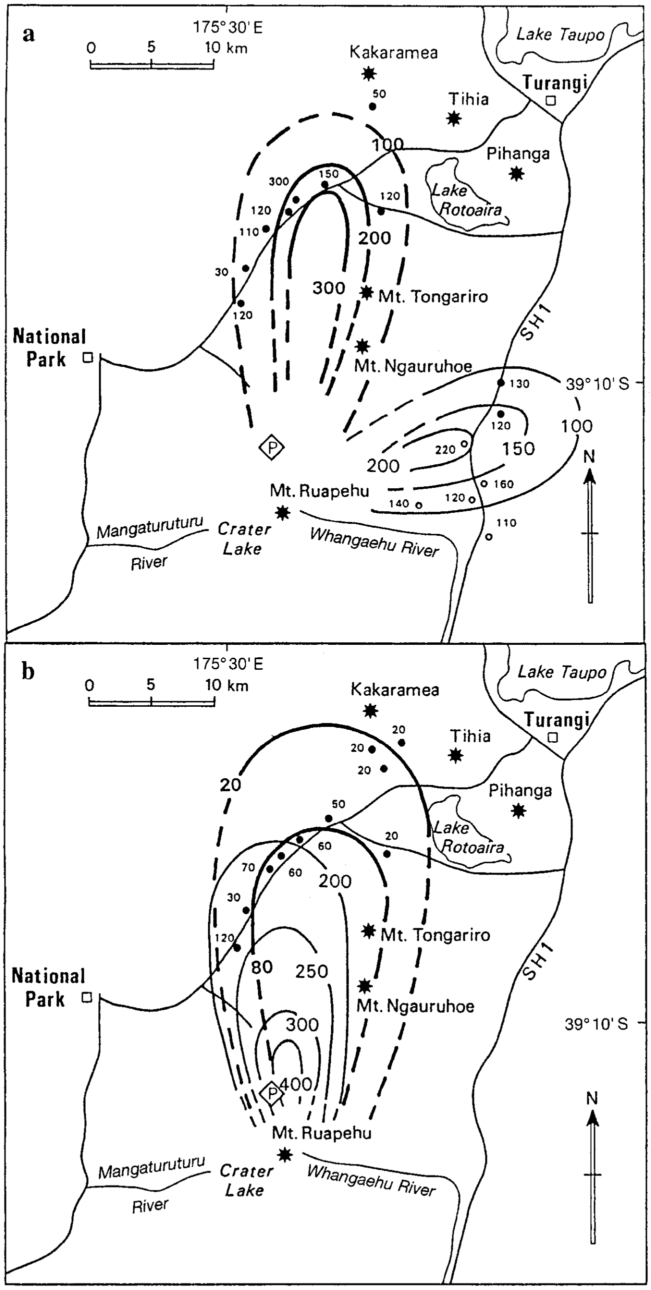

Fig. 8a, b Isopachs of Okupata Member units. **a** Ok_3 is indicated by *thin lines, Ok_2 is indicated by thick lines;* **b** Ok_1 is indicated by *thick lines,* and isopachs of Okupata Tephra are indicated by *thin lines,* adapted from Topping (1973). Contours are in millimetres. *Circles* indicate measured principal sections for isopach data. Also shown is location of Pinnacle Ridge (*P*); see Fig. 2 for distribution of tuff deposits

Based on the eruptive model of Donoghue et al (1995a), we propose that activity at this time (ca. 11,850 and 9800 years B.P.) commenced with the eruption of the Pourahu Member pyroclastic flow (or flows), the deposits of which are visibly mingled, and the coeval eruption of tephra lobe Ok_3, with distributions influenced by a prevailing westerly wind. The Ok_3 lobe (Fig. 8a; Donoghue et al. 1995b) is represented by lapilli units in sections east of Ruapehu. The wind direction then shifted to a prevailing southerly wind. Renewed activity saw the eruption and emplacement of relatively voluminous and coarse-grained Pinnacle Ridge tuff deposits in the proximal vent area and across the northern flanks of the volcano, and the coeval eruption of Ok_1 (Fig. 8b) and Ok_2 (Fig. 8a) tephra-fall deposits. These tephras similarly were dispersed northward, in two overlapping lobes suggested from their thickness trends.

The petrological similarities and close stratigraphic association of the Okupata and Pourahu members indicate that their eruption is intimately related, yet the exact chronostratigraphic and genetic relationships of these members (and Pinnacle Ridge tuff) are unfortunately not established because no one geological section preserves the complete stratigraphy. The tephrostratigraphy and chronology at source allows us only to constrain the time of the Okupata Member eruptions to within a ca. 2000-year period, between ca. 11,850 and 9800 years B.P., although the eruptive interval is undoubtedly much shorter than this. At the type section there is no time break or interval (e.g. a palaeosol or unconformity) between the two lapilli beds, indicating that they were erupted in relatively close succession.

Topping's (1973) isopachs are for the basal lapilli (Fig. 8b), which combines the thickness distributions of our units Ok_1 and Ok_2. His volume estimate of ca. 0.20 km^3 similarly represents the combined volume of these units, although he uses an earlier methodology (Cole and Stephenson 1972) to calculate volume.

Using Fierstein and Nathenson's (1992) methodology and Topping's (1973) isopach map, we have recalculated the total (combined) volume of Ok_1 and Ok_2 to be of the order of ca. 0.19 km^3. Part of this volume (ca. 0.06 km^3) is, however, attributed to the Pinnacle Ridge tuff deposits preserved within the 200-mm isopach (Hackett and Houghton 1989). Subtraction of this volume from the total leaves a combined volume of ca. 0.13 km^3 for Ok_1 and Ok_2. We calculate a volume of ca. 0.05 km^3 for Ok_2, indicating a probable volume (by difference) of ca. 0.08 km^3 for Ok_1. The volume of the northeastern lobe Ok_3 is ca. 0.1 km^3 (Fig. 8a; Donoghue et al. 1995b). Therefore, a reasonable estimate of the total minimum volume of Okupata Member is ca. 0.23 km^3.

The Pinnacle Ridge tuff deposit may have been at least 1 m thick over the entire northern slopes of Ruapehu, but most of the non-welded material has been completely eroded (Hackett and Houghton 1985, 1989). Based on extrapolation of Topping's (1973) Okupata Tephra isopachs back to source (in the vicinity of Pinnacle Ridge) Hackett and Houghton (1989) calculate that of a total volume of 0.06 km^3 within the 200-mm isopach (Fig. 8b), 54% (0.03 km^3) was deposited as primary distal tephra deposits on the ring plain and 30% (0.02 km^3) was eroded from the flanks of the cone and redeposited on the ring plain. Only 16% (0.01 km^3) is preserved as the proximal welded Pinnacle Ridge tuff which presently covers an area of 0.8 km^2.

The Taurewa eruptive episode

The eruptions of the Pourahu and Okupata Member deposits identify a brief ($\ll$2000 years) though significant period of volcanism at Ruapehu that we define as the Taurewa eruptive episode. This episode is significant because it not only represents a period of particularly explosive volcanism in Ruapehu's eruptive history but also marks a culmination in activity at this volcano ca. 10,000 years B.P.

Our eruptive model envisages that this episode, triggered by magma mingling (Donoghue et al. 1995a), was characterised by closely spaced plinian eruptions that led to contemporaneous emplacement of Okupata Member tephra deposits (Ok_1, Ok_2, Ok_3 and the Pinnacle Ridge tuff deposits) and Pourahu Member pyroclastic flow deposits. The pyroclastic flows were presumably generated through collapse of a non-buoyant eruption column (Sparks et al. 1997a) and therefore are associated with plinian tephra-fall deposits (Fig. 9).

We see no evidence on the ring plain of any other single eruption, or group of eruptions, that generated pyroclastic flows, and no evidence for other events of similar magnitude in the past ca. 22,500 years at Ruapehu.

Following the Taurewa episode, explosive plinian activity at Ruapehu virtually ceased (Table 2), and Ruapehu entered what was to be a prolonged ca. 8000-year period of relative quiescence. The focus of activity shifted to neighbouring Tongariro volcano where, within a period of <1000 years, a series of closely spaced plinian eruptions deposited several metres of tephra (recorded by Pahoka Tephra and Mangamate Tephra; Table 2) on the Tongariro and Ruapehu ring plains (Topping 1973; Donoghue et al. 1995b; Nakagawa et al. 1998). The next major period of activity at Ruapehu occurred in the late Holocene (ca. 1850 years B.P. to present) and is represented by the Tufa Trig Formation tephras (Donoghue et al. 1997), which have erupted on average once every 100 years. The volume of tephra erupted in each case is, however, comparatively minor, with each eruption producing ca. <0.1 km^3 of tephra (Table 7; Donoghue 1991; Donoghue et al. 1995b). The volume of ash deposited during the most recent June 1996 Ruapehu eruptions (which contributed to a new Tufa Trig Formation member Tf19; Donoghue et al. 1997) is similarly small, and estimated to be 0.01 km^3 (Cronin et al. 1996).

Significance of the Taurewa eruptive episode

Our re-examination of the ca. 10,500 B.P. eruptive record demonstrates the importance of integrating the total volcanic record from all sectors of the ring plain and the flanks of a volcano when attempting to understand both the processes that characterise eruptions and the potential hazards at andesitic volcanoes. In this study,

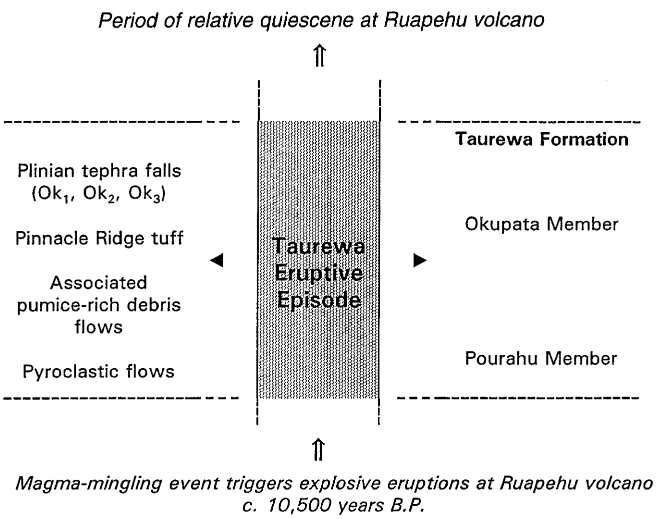

Fig. 9 Summary of events that constitute the Taurewa Eruptive Episode. The episode was triggered by a magma-mingling event and was followed by a period of relative quiescence (see text)

Table 7 Estimates and comparisons of volume (calculated to the nearest 10×10^6 m^3) for Okupata Member and younger late Holocene Tufa Trig Formation tephras

Formation	Member		Volume estimates (km^3)	
Tufa Trig Formation	Tf19		0.01[a]	
	Tf6		0.04[b,c]	
	Tf5		0.10[b,c]	
Taurewa Formation	Okupata Member	Ok_3	–	0.1[b,c]
		Ok_2	0.05	0.20[d,e]
		Ok_1	0.08	
	Pinnacle Ridge tuff			0.06[f]
	Pourahu Member		Not determined	

[a] Data from Cronin et al. (1996)
[b] Volumes calculated using the methodology of Fierstein and Nathenson (1992)
[c] Data from Donoghue et al. (1995b)
[d] Volumes calculated using the methodology of Cole and Stephenson (1972)
[e] Data from Topping (1973)
[f] Data from Hackett and Houghton (1989)

correlation of tephra fall and tephra flow deposits has identified a much larger eruption than was previously envisaged at Ruapehu and has extended our understanding of eruptive processes, event magnitudes, and potential hazards at this volcano.

We now recognise the potential for pyroclastic flows to be generated during eruptions at this volcano, although indications are that such events are likely to occur only infrequently, on time scales of several to tens of thousands of years.

Eruptions that produce in excess of $0.2\ km^3$ of ejecta can be expected to occur on similar time scales at Ruapehu. Although volumetrically small when compared with recent historic eruptions at other stratovolcanoes around the world (the 1980 eruption of Mount St. Helens and the 1982 eruption of El Chichón produced ca. 1.1 and $0.5\ km^3$ of ejecta, respectively; Punongbayan and Tilling 1989; Tilling 1982), the Taurewa Formation eruptives are significant in Ruapehu's eruptive history, and recognition of these events remains important to the overall assessment of future volcanic risk at this and other TgVC volcanoes.

Also important is our recognition of magma-mingling events and their significance in triggering explosive eruptions at Ruapehu and other TgVC volcanoes. Studies (Donoghue et al. 1995a; Nakagawa et al. 1998) clearly indicate that the largest eruptive events at the TgVC volcanoes are associated with magma mixing events and that this process is possibly a prerequisite to explosive pyroclastic-flow-forming eruptions. This is consistent with the behaviour described at numerous other stratovolcanoes, where magma-mixing processes are identified as probable triggers to explosive Plinian eruptions, examples being the relatively recent climactic 15 June 1991 eruption of Mt. Pinatubo (Pallister et al. 1995) and the ca. 2200 years B.P. eruption of layer-C tephra, the largest Holocene eruptive from Mount Rainier (Venesky and Rutherford 1997).

Studies (Donoghue et al. 1995a; this paper) also indicate that climactic events at TgVC volcanoes are followed by lengthy periods of relative quiescence, i.e. approximately 8000 years at Ruapehu volcano and ca. 9000 years for Tongariro volcano. A period of renewed activity commenced at Ruapehu ca. 1850 years B.P. Since then, eruptions have been dominantly small phreatic and phreatomagmatic events, with intermittent strombolian activity and frequent lahar generation. This change in eruptive regime, from plinian to phreatomagmatic eruption styles, has been attributed to the formation of a semi-permanent summit crater lake ca. 1850 years B.P. (Donoghue et al. 1997), and we expect future activity at Ruapehu, over the next tens to hundreds of years to be much the same.

Acknowledgements Research funding for S.L.D. was provided through research grants awarded by The Croucher Foundation (Hong Kong), The Research Grants Council (Hong Kong) and The Committee on Research and Conference Grants (University of Hong Kong). Funding for E.A.M. and K.H. was provided by an ROPA grant. Funding for J.A.L., V.E.N. and A.S.P. was provided by a grant from the New Zealand Foundation for Research, Science and Technology. We express our sincere thanks to K. Scott, R. Hoblitt, and an anonymous reviewer for their particularly thoughtful and encouraging reviews of the paper; the Lake Rotoaira Trust for allowing access to field sites in restricted areas near Lake Rotoaira; the New Zealand Department of Conservation for permission to collect geological samples within Tongariro National Park; J. Patterson (Analytical Facility, Victoria University of Wellington, New Zealand) for his earnest assistance with microprobe operation; R. Price and J. Chapman (La Trobe University) for providing their unpublished data; and T.B. Wong (University of Hong Kong) for kindly draughting the maps. Full descriptions of tephra sections referred to in this text can be obtained from S.L.D. upon request.

References

Arguden AT, Rodolfo KS (1990) Sedimentologic and dynamic differences between hot and cold laharic debris flows of Mayon Volcano, Philippines. Geol Soc Am Bull 102:865–876

Blakemore LC, Searle PL, Daly BK (1987) Methods for chemical analysis of soils. N Z Soil Bureau Scientific Report 80

Buesch DC (1991) Changes in depositional environments resulting from emplacement of a large-volume ignimbrite. Soc Sediment Geol Spec Publ 45:139–153

Carey SN (1991) Transport and deposition of tephra by pyroclastic flows and surges. Soc Sediment Geol Spec Publ 45:39–57

Cole JW (1978) Andesites of Tongariro Volcanic Centre, North Island, New Zealand. J Volcanol Geotherm Res 3:121–153

Cole JW, Stephenson TM (1972) Calculation of the volume of a tephra deposit. Distribution of high-alumina basalts in Taupo Volcanic Zone. Department of Geology, Victoria University of Wellington, vol 1, pp 1–15

Crandell DR, Mullineaux DR (1973) Pine Creek volcanic assemblage at Mount St. Helens, Washington. US Geol Surv Bull 1383-A

Crandell DR, Booth B, Kazumadinata K, Shimozuru D, Walker GPL, Westercamp D (1984) Source book for volcanic hazards zonation. UNESCO, Paris

Cronin SJ, Neall VE, Palmer AS, Lecointre JA (1996) Report on the elemental deposition rates of the 17–19 June 1996 Ruapehu ash falls. Miscellaneous report, Massey University, New Zealand

Donoghue SL (1991) Late Quaternary volcanic stratigraphy of the southeastern sector of the Ruapehu ring plain, New Zealand. PhD thesis, Massey University, Palmerston North, New Zealand

Donoghue SL, Gamble JA, Palmer AS, Stewart RB (1995a) Magma mingling in an andesite pyroclastic flow of the Pourahu Member, Ruapehu Volcano, New Zealand. J Volcanol Geotherm Res 68:177–191

Donoghue SL, Neall VE, Palmer AS (1995b) Stratigraphy and chronology of late Quaternary andesitic tephra deposits, Tongariro Volcanic Centre, New Zealand. J R Soc N Z 25:112–206

Donoghue SL, Neall VE, Palmer AS, Stewart RB (1997) The volcanic history of Ruapehu during the last 2 millennia based on the record of the Tufa Trig Tephras. Bull Volcanol 59:136–146

Fergusson GJ, Rafter TA (1959) New Zealand ^{14}C age measurements–4. N Z J Geol Geophys 2:208–241

Fierstein J, Nathenson M (1992) Another look at the calculation of fallout tephra volumes. Bull Volcanol 54:156–167

Fisher RV, Schmincke H-U (1984) Pyroclastic rocks. Springer, Berlin Heidelberg New York

Flemming CA (1953) The geology of Wanganui Subdivision. N Z Geol Surv Bull 52

Froggatt PC, Lowe DJ (1990) A review of late Quaternary silicic and some other tephra formations from New Zealand: their stratigraphy, nomenclature, distribution, volume, and age. N Z J Geol Geophys 33:89–109

Gourgaud A, Fichaut M, Joron J-L (1989) Magmatology of Mt. Pelée (Martinique, French West Indes). I. Magma mixing and triggering of the 1902 and 1929 Pelean nuées ardentes. J Volcanol Geotherm Res 38:143–169

Hackett WR, Houghton BF (1985) Pinnacle Ridge member, Whakapapa Formation: a welded airfall deposit from Ruapehu Volcano, Taupo Volcanic Zone. N Z Geol Surv Rec 8:24–29

Hackett WR, Houghton BF (1989) A facies model for a Quaternary andesitic composite volcano: Ruapehu, New Zealand. Bull Volcanol 51:51–68

Hobden BJ, Houghton BF, Lanphere MA, Nairn IA (1996) Growth of the Tongariro volcanic complex: new evidence from K–Ar age determinations (Note). N Z J Geol Geophys 39:151–154

Hoblitt RP, Kellog KS (1979) Emplacement temperature of unsorted and unstratified deposits of volcanic rock debris as determined by paleomagnetic techniques. Geol Soc Am Bull 90:633–642

Le Bas MJ, Le Maitre RW, Streckeisen A, Zanettin B (1986) A chemical classification of volcanic rocks based on the total alkali–silica diagram. J Petrol 27:745–750

Le Maitre RW (1984) A proposal by the IUGS Subcommission on the Systematics of Igneous Rocks for a chemical classification of volcanic rocks based on the total alkali silica (TAS) diagram. Aust J Earth Sci 31:243–255

Lowe DJ, Hogg AG (1986) Tephrostratigraphy and chronology of the Kaipo Lagoon, an 11,500-year-old montane peat bog in Urewera National Park, New Zealand. J R Soc N Z 16:25–41

McClelland EA (1996) Theory of CRM acquired by grain growth, and its implications for TRM discrimination and paleointensity determination in igneous rocks. Geophys J Int 126:271–280

McClelland EA, Druitt TH (1989) Paleomagnetic estimates of emplacement temperatures of pyroclastic deposits on Santorini, Greece. Bull Volcanol 51:16–27

Miller CD (1978) Holocene pyroclastic-flow deposits from Shastina and Black Butte, West of Mount Shasta, California. J Res US Geol Surv 6:611–624

Mothes PA, Hall ML, Janda RJ (1998) The enormous Chillos Valley Lahar: an ash-flow generated debris flow from Cotopaxi Volcano, Equador. Bull Volcanol 59:233–244

Nakagawa M, Nairn IA, Kobayashi T (1998) The ~10 ka multiple vent pyroclastic eruption sequence at Tongariro Volcanic Centre, Taupo Volcanic Zone, New Zealand. J Volcanol Geotherm Res 86:45–65

Norrish K, Chappell BW (1977) X-ray fluorescence spectroscopy. In: Zussman J (ed) Physical methods in determinative mineralogy. Academic Press, London, pp 201–272

Norrish K, Hutton JT (1969) An accurate X-ray spectrographic method for the analysis of a wide range of geologic samples. Geochim Cosmochim Acta 33:431–451

Palmer K (1990) XRF analyses of granitoids and associated rocks, St John's Range, South Victoria Land, Antarctica. Ann Contrib 13. Victoria University of Wellington, 23 pp

Palmer BA, Neall VE (1989) The Murimotu Formation: 9,500-year-old deposits of a debris avalanche and associated lahars, Mount Ruapehu, North Island, New Zealand. N Z J Geol Geophys 32:477–486

Pallister JS, Hoblitt RP, Meeker GP, Knight RJ, Siems DF (1995) Magma mixing at Mount Pinatubo: petrographic and chemical evidence from the 1991 deposits. In: Newhall G, Punongbayan R (eds) Fire and mud. University of Washington Press, Seattle, pp 686–732

Patterson DB, Graham IJ (1988) Petrogenesis of andesitic lavas from Mangatepopo Valley and Upper Tama Lake, Tongariro Volcanic Centre, New Zealand. J Volcanol Geotherm Res 35:17–29

Paulick H, Franz G (1997) The color of pumice: case study on a trachytic fall deposit, Meidob volcanic field, Sudan. Bull Volcanol 59:171–185

Pérez-Torrado FJ, Martí J, Mangas J, Dat S (1997) Ignimbrites of the Roque Nublo group, Gran Canaria, Canary Islands. Bull Volcanol 58:647–654

Price RCP, McCulloch MT, Smith IEM, Stewart RB (1992) Pd–Nd–Sr isotopic compositions from Egmont Volcano and comparisons with basalts and andesites from the Taupo Volcanic Zone, New Zealand. Geochim Cosmochim Acta 56:941–953

Punongbayan RS, Tilling RI (1989) Recent case histories. In: Tilling R (ed) Volcanic hazards. American Geophysical Union, Washington DC, pp 81–102

Scott KM, Vallance JW, Pringle PT (1995) Sedimentology, behaviour, and hazards of debris flows at Mount Rainier, Washington. US Geol Surv Prof Pap 1547

Smith GA, Lowe DR (1991) Lahars: volcano–hydrologic events and deposition in the debris flow–hyperconcentrated flow continuum. Soc Sediment Geol Spec Publ 45:359–370

Soil Survey Staff (1996) Key to soil taxonomy. SMSS Tech Monogr 19. Virginia Polytechnic Institute and State University, Blacksburg, Virginia

Sparks RS (1976) Grain size variations in ignimbrites and implications for the transport of pyroclastic flows. Sedimentology 3:147–188

Sparks RSJ, Sigurdsson H, Wilson L (1977) Magma mixing: a mechanism for triggering acid explosive eruptions. Nature 267:315–318

Sparks RSJ, Bursik MI, Carey SN, Gilbert JS, Glaze LS, Sigurdsson H, Woods AW (1997a) Volcanic plumes. Wiley, Chichester

Sparks RSJ, Gardeweg MC, Calder ES, Matthews SJ (1997b) Erosion by pyroclastic flows on Lascar Volcano, Chile. Bull Volcanol 58:557–565

Stewart RB, Price RC, Smith IEM (1996) Evolution of high-K arc magma, Egmont Volcano, Taranaki, New Zealand: evidence from mineral chemistry. J Volcanol Geotherm Res 74:275–295

Tilling RI (1982) Volcanic cloud may alter earth's climate. In: Aldana G, Garrett K (eds) The disaster of El Chichón. National Geographic 162:672–675

Topping WW (1973) Tephrostratigraphy and chronology of late Quaternary eruptives from the Tongariro Volcanic Centre. N Z J Geol Geophys 16:397–423

Valet JP, Kidane T, Soler V, Brassart J, Courtillot V, Meynadier L (1998) Remagnetization in lava flows recording pretransitional directions. J Geophys Res 103:9755–9775

Venesky DY, Rutherford MJ (1997) Pre-eruption conditions and timing of dacite–andesite magma mixing in the 2.2 ka eruption at Mount Rainier. J Geophys Res 102:69–86

Wilson CJN (1993) Stratigraphy, chronology, styles and dynamics of late Quaternary eruptions from Taupo Volcano, New Zealand. Phil Trans R Soc Lond A343:205–306

Wilson CJN, Switsur VR, Ward AP (1988) A new 14C age for the Oruanui (Wairakei) eruption, New Zealand. Geol Mag 125:297–300

Zlotniki J, Pozzi JP, Boudon G, Moreau MG (1984) A new method for the determination of the setting temperature of pyroclastic deposits (example of Guadeloupe, French West Indies). J Volcanol Geotherm Res 21:297–312