Journal of volcanology and geothermal research

ELSEVIER Journal of Volcanology and Geothermal Research 68 (1995) 1–28

Volcanic and structural evolution of Taupo Volcanic Zone, New Zealand: a review

C.J.N. Wilson [a,*], B.F. Houghton [a], M.O. McWilliams [b], M.A. Lanphere [c], S.D. Weaver [d], R.M. Briggs [e]

[a] *Institute of Geological and Nuclear Sciences, Wairakei Research Centre, Private Bag 2000, Taupo, New Zealand*
[b] *Department of Geophysics, Stanford University, Stanford, CA 94305, USA*
[c] *U.S. Geological Survey, MS 937, 345 Middlefield Road, Menlo Park, CA 94025, USA*
[d] *Department of Geological Sciences, University of Canterbury, Private Bag 4800, Christchurch, New Zealand*
[e] *Department of Earth Sciences, University of Waikato, Private Bag 3105, Hamilton, New Zealand*

Received 30 June 1994; accepted 19 September 1994

Abstract

The Taupo Volcanic Zone (TVZ) in the central North Island is the main focus of young volcanism in New Zealand. Andesitic activity started at c. 2 Ma, joined by voluminous rhyolitic (plus minor basaltic and dacitic) activity from c. 1.6 Ma. The TVZ is c. 300 km long (200 km on land) and up to 60 km wide, as defined by vent positions and caldera structural boundaries. The total volume of TVZ volcanic deposits is uncertain because a sub-volcanic basement has not been identified, but present data suggest bulk volumes of 15–20,000 km^3, and that faulted metasediments form most of the immediate subvolcanic basement. Rhyolite ($\geq 15,000$ km^3 bulk volume, typically 70–77% SiO_2) is the dominant magma erupted in the TVZ (mostly as caldera-forming ignimbrite eruptions), andesite is an order of magnitude less abundant, and basalt and dacite are minor in volume (< 100 km^3 each). The history of the TVZ is here divided into 'old TVZ' from 2.0 Ma to 0.34 Ma, and 'young TVZ' from 0.34 Ma onwards, separated by the Whakamaru eruptions, which obscured much of the evidence for older activity within the zone. The TVZ shows a pronounced segmentation into northeastern and southwestern andesite-dominated extremities with composite cones and no calderas, and a central 125-km-long rhyolite-dominated segment. Eight rhyolitic caldera centres have so far been identified in the central segment, of which two (Mangakino and Kapenga) are composite features, and more centres will probably be delineated as further data accumulate. These centres account for 34 inferred caldera-forming ignimbrite eruptions, in the c. 1.6-Ma lifetime of the central TVZ. The modern central TVZ is the most frequently active and productive silicic volcanic system on Earth, erupting rhyolite at c. 0.28 $m^3\ s^{-1}$, and available information suggests this has been so for at least the past 0.34 Ma. The rhyolites show no major compositional changes with time, though the extent of magma chamber zonation may have changed with the incoming of rifting and crustal extension in the past c. 0.9 Ma. Within the central TVZ, non-rhyolitic compositions have been erupted apparently irregularly in time and space; in particular there is no evidence for a geographic separation of basalts from andesites. Between 0.9 and 0.34 Ma, a major episode of uplift affected areas around the TVZ, while at the same time the main focus of activity may have migrated eastwards within the TVZ accompanying rifting along the axis of the zone. The modern TVZ is rifting at rates between 7 and 18 mm a^{-1} and restoration of the thin (15 km) 'crust' ($V_p \leq 6.1$ km s^{-1}) beneath the central TVZ to its pre-rifting thickness (25 km) implies that rifting at such rates may have begun only at c. 0.9 Ma. The TVZ is a rifted arc, but its longitudinally segmented nature, high thermal flux and voluminous rhyolitic volcanism make it unique on Earth.

* Corresponding author

SSDI 0377-0273(95)00006-2

1. Introduction

The Taupo Volcanic Zone (TVZ) is the dominant locus of late Pliocene to Quaternary volcanic activity in New Zealand resulting from subduction of the Pacific plate beneath the North Island (Figs. 1 and 2). The TVZ has produced at least 10,000 km^3 of magma or >90% of the late Pliocene to Quaternary eruptives known in New Zealand. The earliest TVZ activity was andesitic and began at c. 2 Ma. This was joined by intensive and volumetrically dominant rhyolitic activity from c. 1.6 Ma onwards, as the central part of the modern TVZ became the largest and most frequently active rhyolitic magmatic system on Earth (Houghton et al., 1995). Although numerous papers have been written on the structure, history and evolution of the TVZ, our understanding is still incomplete, and many interpretations require revision and continued investigation. This paper reviews aspects of the structure and evolution of the TVZ (especially the rhyolitic portion) in the light of existing and new data on the sources, ages and compositions of the eruptive products.

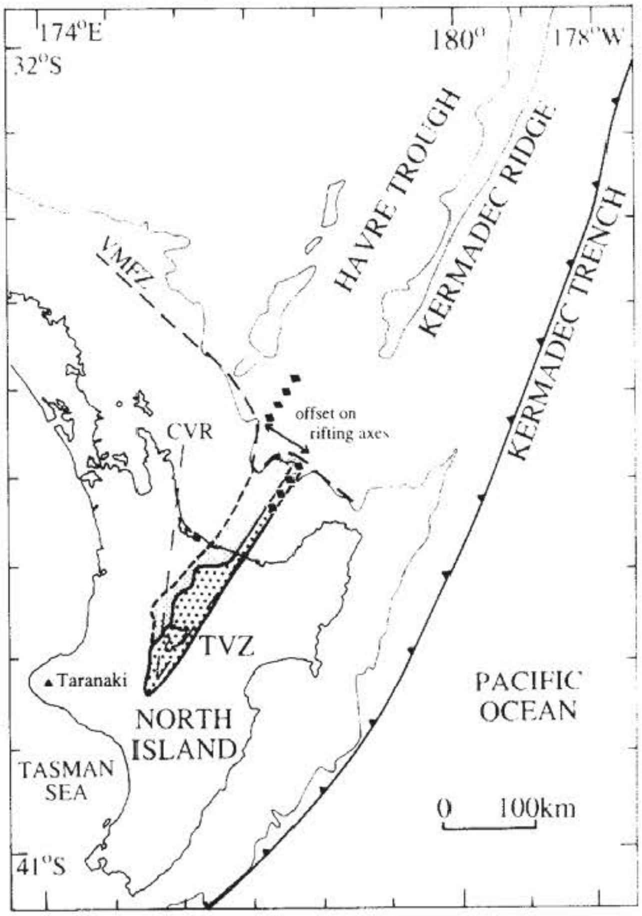

Fig. 1. Map to show the setting of the Taupo Volcanic Zone (*TVZ*) in the North Island of New Zealand, with respect to the Kermadec Trench and Ridge, and to the Havre Trough (outlined by the 2000-m isobath). The TVZ is stippled, coarser stippling represents young TVZ, active in the last 0.34 Ma (see text). The Central Volcanic Region (*CVR*) boundary is shown (from Stern, 1985, 1987) for comparison. The dashed line marked *VMFZ* forming the northeastern termination of the TVZ is the continental/oceanic crust boundary identified by Gamble et al. (1993a) parallel to or as a continuation of the Vening Meinesz Fracture Zone. Diamonds mark the lines of the modern rifting axes for the Havre Trough and the TVZ, to show their offset across the continental/oceanic crust boundary. Adapted from Gamble et al. (1993a).

The degree of detail established for the history of the TVZ diminishes exponentially with age, as older centres or their deposits have been buried during subsequent volcanism, destroyed by caldera collapse and regional subsidence, or obscured by erosion (Fig. 3). The most significant volcanic episode, which concealed much of the evidence of the earlier history of the TVZ, was the eruption and emplacement of the widespread Whakamaru-group ignimbrites with related fall deposits and associated caldera formation between 0.34 and 0.32 Ma (Wilson et al., 1986; Pringle et al., 1992; Houghton et al., 1994). Data concerning vents and eruptive centres (Fig. 5) during and after the Whakamaru-group events are much more complete than before, and so we use in this paper two unequal stages in the evolution of the TVZ:

old TVZ, that represented by activity from inception of the zone (c. 2 Ma) up to the onset of the Whakamaru-group eruptions (0.34 Ma; Fig. 4a), and

young TVZ, that represented by centres active during and after the Whakamaru-group eruptions (i.e. from 0.34 Ma to the present; Fig. 4b).

In addition, we refer to:

modern TVZ, active during and since the c. 65 ka Rotoiti eruption of Okataina volcanic centre, an interval considered by us to represent the current state of the zone (Fig. 4c), and

whole TVZ, to refer to all activity in the zone from 2 Ma to the present day.

In describing eruptive compositions in the zone, we use basalt, andesite, dacite and rhyolite, with boundaries at 53, 63 and 69% SiO_2 (following Ewart, 1982). Earlier papers (e.g., Cole, 1979, 1981) follow this convention except for using 67 or 68% SiO_2 as the divider between dacites and rhyolites, but 69% SiO_2 acts as a

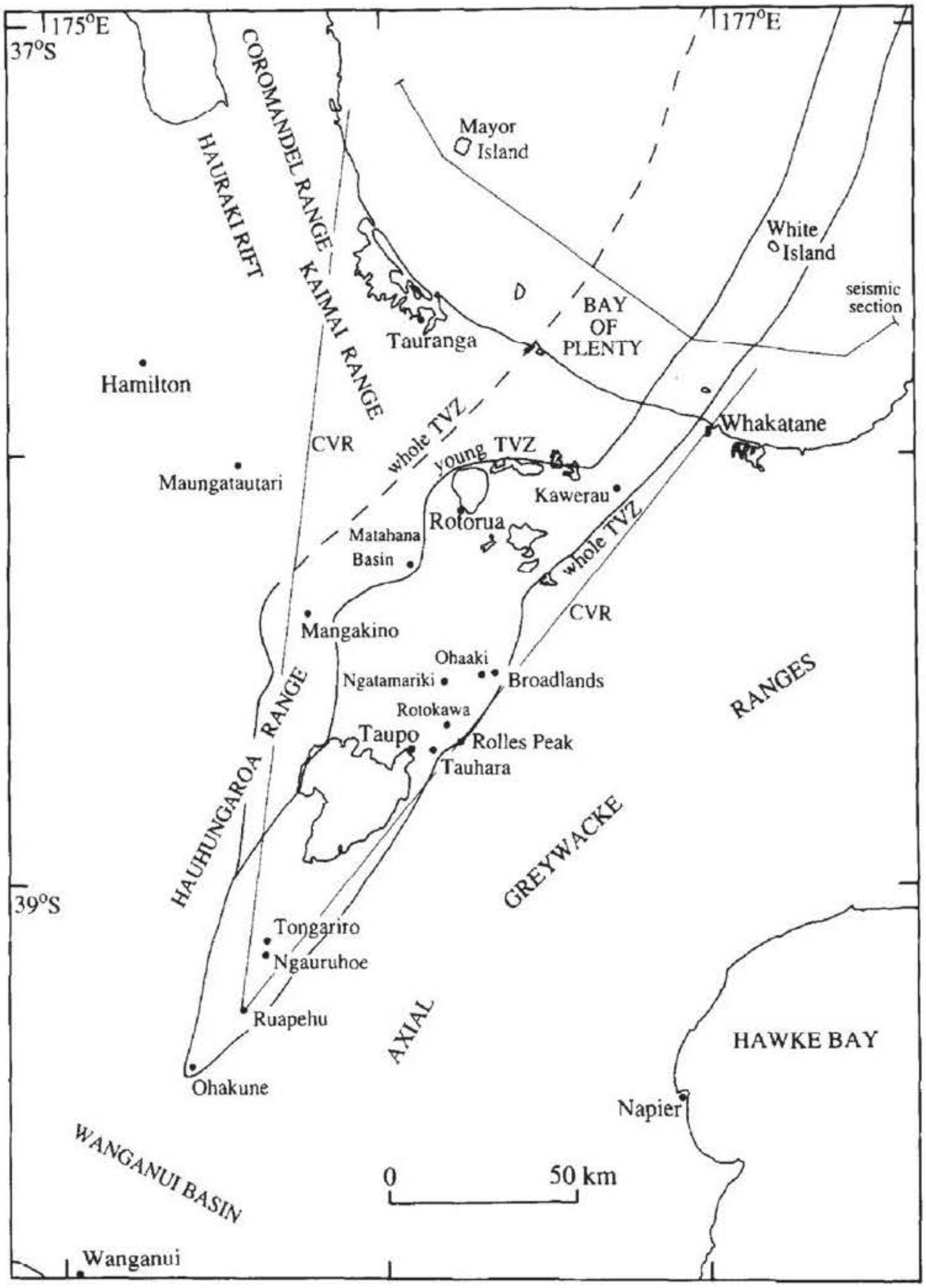

Fig. 2. Map showing localities and areas mentioned in the text. *Whole TVZ* and *young TVZ* boundaries are from Fig. 5 (onshore), and after Davey et al. (1995), Wright (1990) and Gamble et al. (1993a) (offshore). CVR boundary from Stern (1985, 1987). The 'seismic section' line marked in the Bay of Plenty is that of Davey et al. (1995).

more satisfactory divider between the smaller volume (<3 km^3), typically moderately to highly porphyritic and cone-forming dacites and the large volume (often >100 km^3) caldera-related rhyodacites to low-Si rhyolites.

2. Taupo Volcanic Zone

2.1. Definitions

The first use of the term 'Taupo Zone' was by Hochstetter (1864) to describe the zone of young eruptive

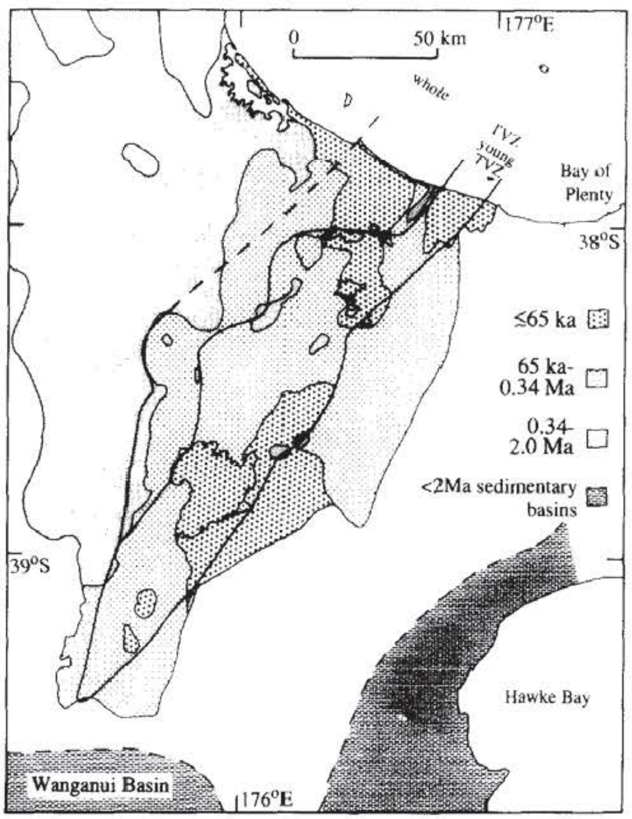

Fig. 3. Sketch map to illustrate areas in the TVZ and vicinity where surface outcrops yield information about units of the ages indicated. Note that most of the TVZ is buried beneath ignimbrites of 0.34 Ma or younger ages which obscure much of the detail for older volcanism and eruptive centres within the zone. Extensive outcrops of pre-0.34 Ma ignimbrites providing stratigraphic information occur NW of the TVZ, and these older units are also represented by primary and secondary-reworked pyroclastics in young sedimentary rocks in the Wanganui Basin and in areas around and south of Hawke Bay (e.g., Seward, 1974, 1976; Shane, 1991, 1994; Shane and Froggatt, 1991; Alloway et al., 1993; Pillans et al., 1994). Unornamented areas have little or no information about TVZ eruptives older than 65 ka due to erosion.

centres and associated geothermal systems extending from Ruapehu to White Island, and this definition is adhered to here. In general terms, the TVZ denotes a NNE-SSW-trending zone of late Pliocene to Quaternary arc volcanism from Ohakune to White Island and beyond to the edge of the continental shelf. Specifically the zone is defined as the area enclosed by an envelope drawn around all caldera structural margins and individual vent sites associated with the NNE–SSW-orientated Kermadec subduction system (Fig. 1) and active during the c. 2 Ma lifetime of the zone (Fig. 5; Wilson et al., 1984). As such, the TVZ is defined from vent positions (and structural elements such as caldera margins directly associated with volcanism), and does not include faulting parallel to but outside the envelope of vents.

In some papers the TVZ is treated as the active portion of a larger area termed the Central Volcanic Region (CVR; Figs. 1 and 2). The CVR was first defined as the geographic area, which included the TVZ and flanking plateaux of welded ignimbrite, within which TVZ-derived volcanic rocks formed mappable units (e.g., Thompson, 1964). However, the areas of ignimbrite used to define the CVR geologically are merely those where the relevant deposits have been preserved, and do not reflect either their original extent or include any defined vent areas. This use of the term CVR has disappeared in favour of a definition (e.g., Stern, 1985) as a wedge-shaped area, defined by gravity data and limited age data, which shares a common eastern boundary with the TVZ but which extends further to the northwest. We use the TVZ as the area for discussion in this paper, and suggest abandonment of the term CVR, for reasons which are discussed in Section 4.2, below.

2.2. *Boundaries*

Onshore

For the old TVZ there is no clearly defined NW limit to the zone north and northeast of Mangakino caldera (Fig. 5; Wilson et al., 1984), partly because of a lack of reliable data on the ages and sources of youngest activity in the NNW–SSE-orientated Miocene–Pliocene volcanics of the southern Coromandel–Kaimai area (Skinner, 1986), and partly because of burial of evidence for any eruptive centres by younger ignimbrites (Fig. 3). Southwest of Mangakino, the western boundary is defined by andesitic centres along the Hauhungaroa Range which have typical arc-related compositions and 3 age determinations of between 2.02 ± 0.02 and 1.85 ± 0.02 Ma (Fig. 5; Stipp, 1968). The contemporaneous late Pliocene (1.8 Ma; Robertson, 1983) andesite-dacite centre of Maungatautari to the NW (Fig. 2) contains more K-rich compositions suggestive of a relationship to the old TVZ comparable to that of Taranaki to the modern TVZ (Fig. 1; Briggs, 1986). The eastern boundary is partly obscured by burial, but andesites older than 0.34 Ma occur close to or coincident with the eastern young TVZ margin (e.g.,

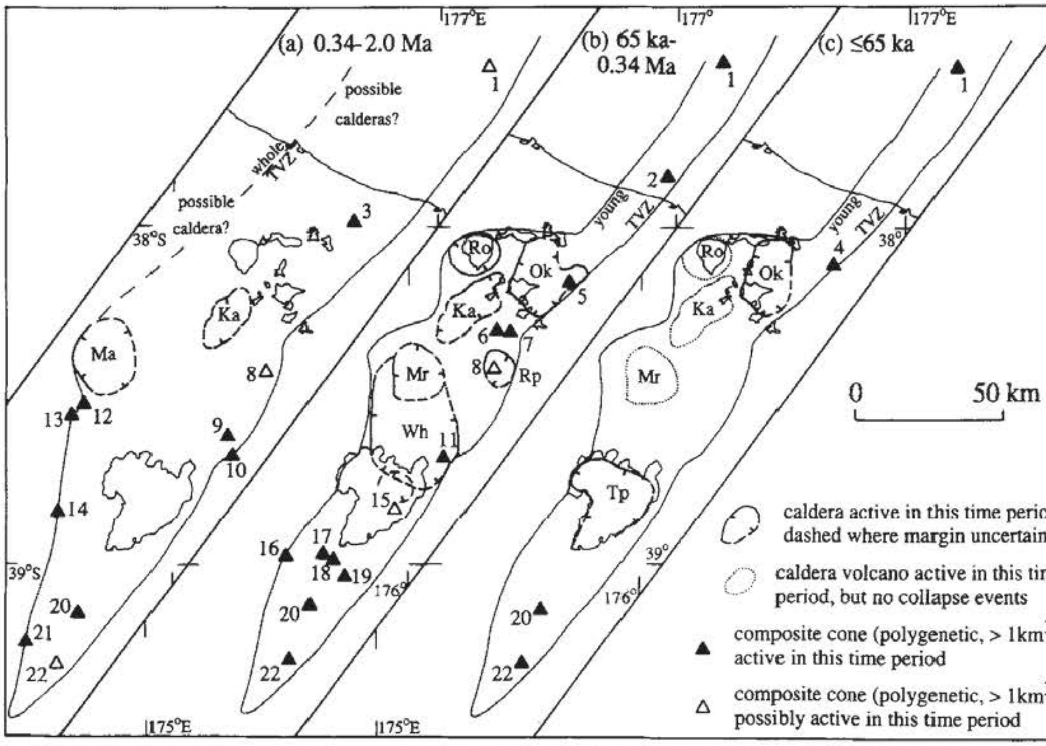

Fig. 4. Maps to illustrate eruptive centres active during (a) old-TVZ (2.0–0.34 Ma), (b) earlier young-TVZ (0.34 Ma to 65 ka), and (c) later young-TVZ (= modern TVZ) (< 65 ka); see text for details. Calderas (from Houghton et al., 1995) are: *Ka* = Kapenga; *Ma* = Mangakino; *Mr* = Maroa; *Ro* = Rotorua; *Rp* = Reporoa; *Tp* = Taupo; *Wh* = Whakamaru. Composite cones and dome complexes of non-rhyolitic composition larger than c. 1 km^3 are numbered thus: 1 = White Island (> 65 ka, Duncan, 1970); 2 = Motuhora/Whale Island (> 65 ka Ma; Duncan, 1970); 3 = Manawahe (0.43 Ma, Broughton, 1988); 4 = Edgecumbe (1.8–5.5 ^{14}C ka, Nairn, 1981); 5 = Puhipuhi (0.23–0.28 Ma from bracketing ignimbrites, Nairn, 1989 and Houghton et al., 1995); 6 = Maungaongaonga (0.18 Ma, B.F. Houghton et al., unpubl. data); 7 = Maungakakaramea (undated); 8 = postulated andesite cone engulfed during collapse of Reporoa caldera (> 0.23 Ma, Nairn et al., 1994); 9 = Rotokawa (> 0.34 Ma, Browne et al., 1992); 10 = Rolles Peak (0.71 Ma, B.F. Houghton et al., unpubl. data); 11 = Tauhara (0.19 Ma, B.F. Houghton et al., unpubl. data); 12 = Titiraupenga (1.85 Ma, Stipp, 1968); 13 = Pureora (undated); 14 = Hauhungaroa (2.0 Ma, Stipp, 1968; identification of source cone, C.J.N. Wilson, unpubl. data); 15 = postulated andesite cone engulfed during collapse of Taupo caldera (> 22.6 ^{14}C ka, Wilson et al., 1988, C.J.N. Wilson, unpubl. data); 16 = Maungakatote (< 0.32 Ma, Cole, 1978a); 17 = Kakaramea (≥ 0.22 Ma, Stipp, 1968); 18 = Tihia (undated); 19 = Pihanga (≥ 0.13 Ma, Stipp, 1968); 20 = Tongariro (onset > 0.34 Ma, Grindley, 1960); 21 = Hauhungatahi (> 0.34 Ma, Hackett, 1985); 22 = Ruapehu (oldest dated lava 0.22 Ma, Stipp, 1968).

Rolles Peak; 0.71 Ma, B.F. Houghton et al., unpubl. data).

For the young TVZ the onshore boundaries are unequivocally defined by caldera margins and vent positions on both the east and west sides (Wilson et al., 1984). The eastern margin appears to have remained stationary for the past 0.34 Ma and is essentially the same as that defined for the old TVZ, while the western margin is c. 15 km east of the old TVZ margin. Available geological and chronological evidence imply that the zone of vents defining the young TVZ has not in fact changed appreciably for at least the past 0.7 Ma.

Offshore

The offshore boundaries of the TVZ have only recently been explored in any detail. Earlier workers ended the TVZ at White Island, but new information (e.g., Wright, 1992, 1993a,b) show that young volcanism and accompanying extension continue north of White Island and off the edge of the continental crust, with concomitant changes in the compositions and styles of activity (e.g., Gamble et al., 1990, 1993a, b). A boundary between the TVZ and the Kermadec Ridge–Havre Trough system, its counterpart to the NNE, is drawn at the edge of the continental crust, and the inference made that the two systems are separated by a NW–SE-striking discontinuity (e.g., Lewis and Pantin, 1984; Wright, 1993b; Fig. 1). This discontinuity is variously interpreted as coincident with (e.g., Wright, 1993b) or parallel to but independent of (e.g., Lewis and Pantin, 1984) a structure termed the Vening Meinesz Fracture Zone. Nowadays, the inferred rifting axes of the young Havre Trough and the TVZ are offset

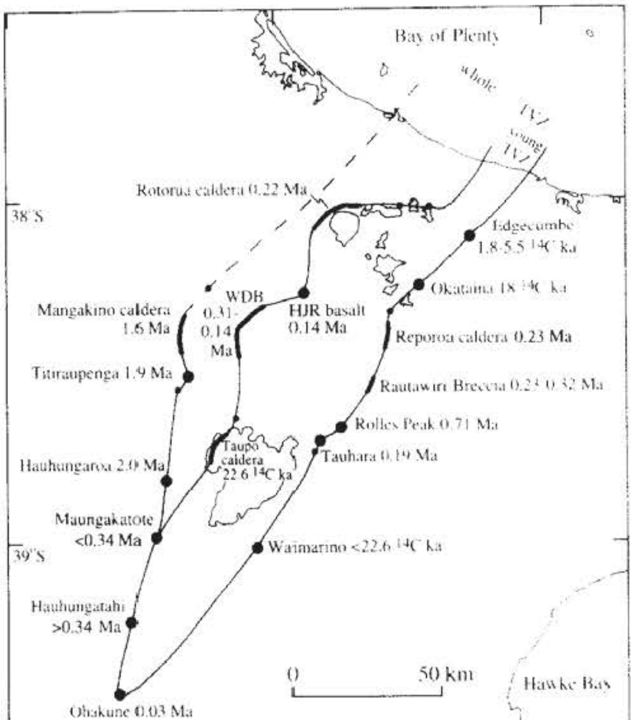

Fig. 5. Map to show the eruptive vents and caldera structural margins, with their ages, used here to define the boundaries of the onshore portion of the whole TVZ (2.0 Ma to present) and young-TVZ (0.34 Ma to present). Small filled circles mark vent sites used to constrain the TVZ boundaries but which are not dated. Data points and sources are: Rotorua caldera, Mangakino caldera, Houghton et al. (1995); Titiraupenga, Hauhungaroa, Stipp (1968); Maungakatote, Cole (1978a); Hauhungatahi, Ohakune, Waimarino, Hackett (1985); HJR (Harry Johnson Road) basalt, Tauhara, Rolles Peak, B.F. Houghton et al. (unpubl. data); Rautawiri Breccia, S.J.A. Brown, pers. commun., 1994, and bracketing ages from Houghton et al. (1995); Reporoa caldera, Nairn et al. (1994); Okataina, Edgecumbe, Nairn (1981, 1989); WDB (Western Dome Belt), Houghton et al. (1991, excluding one age of 0.397 Ma, now thought to be anomalous).

by c. 50 km (Fig. 1), and at present, rifting in the former exerts only a minor influence on volcanism on the New Zealand continental crust (Cole, 1978b; Gamble et al., 1993a).

Seismic data from the Bay of Plenty are interpreted by Davey et al. (1995) to suggest that volcaniclastic sedimentation inferred to be associated with late Pliocene to Pleistocene caldera-related eruptions once extended further NW than the boundaries of the young TVZ as inferred by us. The NW boundary of rifting, (and accompanying volcanism) inferred by Davey et al. (1995) to be 1.6 Ma or younger, is used by them and by us to broadly define the NW boundary of the whole TVZ (Fig. 1). However, more evidence of the age and nature of offshore and possible contemporaneous onshore volcanism in the southern Kaimai area (e.g., Houghton and Cuthbertson, 1989) is required to better define the TVZ boundary.

Subsurface

The thickness of volcanic deposits within the TVZ is controversial. To the east and west of the TVZ, pre-volcanic sediments ('greywacke'), largely of Mesozoic age, crop out at the surface in the axial and Hauhungaroa ranges (Fig. 2). Earlier workers (e.g., Grindley, 1960; Healy, 1962, 1964; Healy et al., 1964) inferred that the greywacke was essentially continuous underneath the zone at depths of a few km, though locally faulted and disrupted by feeder intrusions to the surface volcanism. In sharp contrast, proponents of the geophysically defined CVR (e.g., Calhaem, 1973; Stern, 1985, 1987; see section 4.2, below) propose that the entire sub-volcanic crust (c. 3–15 km) beneath the CVR has been replaced by intermediate to silicic intrusives. Estimates of the thickness of volcanic materials in the TVZ are essential for defining its size, yet the contact between volcanics and basement is only defined within the TVZ in a few geothermal drillholes (Fig. 2) at Rotokawa, Ohaaki and Kawerau (into greywacke) and possibly Ngatamariki (into a diorite intrusion) (Browne, 1978; Wood, 1983; Browne et al., 1992). Elsewhere, drillholes to c. 2.5 km depth do not penetrate the volcanic succession.

In the onshore TVZ, seismic basement with $V_p = 4.8–5.5\ km\ s^{-1}$ is usually inferred to lie below 2.0–2.2 km depth (e.g., Robinson et al., 1981; Stern and Davey, 1985) but the nature of this basement is ambiguous and it cannot be correlated with any particular lithology (Bibby et al., 1995). The *de facto* method generally used is to assume that the basement has an arbitrary density, or density gradient, and the thickness of 'volcanics' is calculated by modelling the residual gravity fields (e.g., Stern, 1979; Rogan, 1982; Hochstein et al., 1993). The thicknesses thus obtained and estimates of c. 12,000 km^3 of volcanics in the central TVZ alone (e.g., Healy, 1962; Cole, 1979) are minima, and the nature of the sub-volcanic basement is not uniquely defined. In the offshore Bay of Plenty region, the corresponding seismic basement is tentatively identified as faulted greywacke (Davey et al., 1995) which, if confirmed (e.g., greywacke is known to crop out further offshore; Gamble et al., 1993a), would strongly imply that greywacke (or its underlying high-grade metamorphic equivalent) rather than Quaternary igneous rocks was the dominant shallow basement lithology comprising the upper crust under the TVZ.

Velocity estimates from seismic studies are not detailed enough to distinguish between competing model lithologies for the deep crust below the TVZ, but imply that the base of rocks with $V_p \cong 6.1$ km s^{-1} lies at only 15 km depth under the TVZ, and that these are underlain by 'anomalous mantle' with $V_p = 7.4$–7.5 km s^{-1} (Stern and Davey, 1987). This crustal model, whilst seemingly consistent with concepts of crustal thinning by rifting or spreading in the TVZ (e.g., Stern, 1985; Cole et al., 1995), is clearly too simple. Models of continental arc magmatism in general (e.g., Hildreth, 1981) and for the TVZ in particular (e.g., Wilson et al., 1984; Hochstein et al., 1993; Graham et al., 1995) require the presence of intrusive rocks at depth in the TVZ, yet these have not been distinguished geophysically. Petrogenetic models for TVZ eruptives imply there should be major quantities ($>40 \times 10^4$ km^3) of mafic to intermediate cumulates as the complementary crystalline fractions to intermediate to silicic compositions generated by fractional crystallization (e.g., Blattner and Reid, 1982; McCulloch et al., 1994). V_p values reported elsewhere (e.g., Christensen, 1982) for some mafic gabbros and granulites at the relevant pressures and temperatures may reach the 7.4–7.5 km s^{-1} values inferred for the 'anomalous mantle' of Stern and Davey (1987). We thus concur with Hochstein et al. (1993) that the 15-km-deep horizon may not represent the crust–mantle boundary, but may be a demarcation between a quartzofeldspathic greywacke/granitoid/granulite zone above and mafic cumulates below, and that the crust–mantle petrological boundary may lie at greater depths (and has not yet been seismically defined).

3. TVZ volcanism in time and space

The chronology of the onshore TVZ has been established by different techniques on two time scales differing by an order of magnitude or more in duration and the detail with which the eruptive history is known. The first, more detailed time framework is back to and including the Rotoehu Tephra (65 ka), where radiocarbon techniques have been used to establish a detailed chronology for the modern TVZ, encompassing 3 caldera-forming events together with at least 52 smaller eruptions (Howorth, 1975; Vucetich and Howorth, 1976; Froggatt and Lowe, 1990; Wilson, 1993). The second time scale covers 2 Ma to 65 ka, for which at least 31 caldera-forming events are inferred (Houghton et al., 1995), but the stratigraphy and chronology of smaller events are poorly documented, as are chronologies for the composite volcanoes and eruptions of non-rhyolitic compositions in the central TVZ. In the sections below we discuss the dating techniques, then review the eruptive histories of the major compositional groupings in TVZ history, concentrating on the chronology of the pre-65 ka events.

3.1. Dating techniques

Four direct approaches have been taken to dating the pre-65 ka rocks in the TVZ; fission-track dating of glass or zircons (e.g., Kohn, 1973; Murphy and Seward, 1981; Alloway et al., 1993), whole-rock K/Ar techniques (e.g., Stipp, 1968; Houghton et al., 1991), K/Ar techniques on mineral separates (Soengkono et al., 1992), and Ar/Ar techniques on feldspar separates (e.g., Pringle et al., 1992; Houghton et al., 1995). The first has been used both within the zone to obtain ages of ignimbrites as well as for inferred correlative tephra horizons in marine sediments (e.g., Seward, 1974, 1975; Alloway et al., 1993), whereas the other techniques have only been used on primary eruptive products in or immediately around the TVZ itself. An additional, indirect technique also used is to relate eruptive units and their re-worked equivalents to the magnetostratigraphic time scale (e.g., Ninkovich, 1968; Watkins and Huang, 1977; Seward et al., 1986).

Earlier fission-track ages on glass or zircon from eruptive units used to construct previous chronologies for TVZ rhyolitic volcanism (e.g., Froggatt, 1983; Wilson et al., 1984; Wilson, 1986) are now known to be inaccurate (Alloway et al., 1993; Houghton et al., 1995), for various reasons which are not well understood. There appears to be reasonable agreement between age estimates from isothermal plateau fission track techniques (e.g., Alloway et al., 1993) and Ar-Ar techniques (e.g., Houghton et al., 1995), as judged by comparisons with the palaeomagnetic polarity time scale, but the two techniques have so far been applied to entirely different deposits in different areas (fall deposits and reworked tephras in the Wanganui Basin versus primary ignimbrites in the TVZ, respectively).

3.2. Basalts

Eruptive form and styles

Basaltic activity in the onshore TVZ contrasts strongly with that of other compositions in that all examples known represent relatively small, monogenetic events, erupting as:

(a) locally dispersed pyroclastics forming tuff rings or scoria cones in the central TVZ, occasionally also with associated lavas, having volumes of the order 0.001–0.1 km^3 (e.g., Houghton et al., 1987; Brown et al., 1994);

(b) widely dispersed fall deposits of volume 0.01–1.0 km^3 mostly associated in time and/or space with larger-scale rhyolitic activity (e.g., Nairn, 1981; Walker et al., 1984);

(c) minor volumes of lava and pyroclastics associated with the large andesitic composite cones (e.g., Graham and Hackett, 1987); or

(d) a mafic component mixed into rhyolitic eruptive units (e.g., Sutton et al., 1995).

While it is widely (but not universally) inferred that mafic (specifically basaltic) magmatism must fuel almost all the other volcanism in the TVZ (cf. Hildreth, 1981), the observation that the basalts are erupted in monogenetic events implies that their appearance at the surface requires unusual circumstances. The volumes of even the largest individual basaltic events (c. 1 km^3 at Tarawera 1886 AD) seem to be typically one order of magnitude smaller than the largest andesite or dacite eruptions, and 2–3 orders of magnitude smaller than the largest rhyolite events. This would imply that the circumstances under which basaltic volcanism could occur either coincided with only small volumes of magma being available, or that some preferential filter acted to preclude large-scale basaltic activity.

Eruptive styles vary widely both within and between individual centres, with the greatest variations being due to differing degrees of interaction with external water (e.g., Houghton and Hackett, 1984; Houghton et al., 1987).

Distribution in time and space

A map of basaltic eruptive centres is given in Fig. 6. Note that all the basalts so far documented in the TVZ are synchronous with, or post-date the 0.34–0.32 Ma Whakamaru-group eruptions. We infer that this is caused by the poor preservation-potential of earlier basaltic products, rather than due to any change in the compositions of TVZ magmas with time, as basalt occurs as a minor lithic component in several of the large-volume pre-0.34 Ma rhyolitic eruptions. The limited sizes and distributions of the basaltic units means that they are particularly vulnerable to burial, or destruction by caldera collapse. A key feature of the geographic distribution of the basalt centres is that they are more or less evenly distributed across the zone (see Fig. 6 and section 4.3 below).

3.3. Andesites

Eruptive forms and styles

The greatest proportion of andesite has been erupted, as lavas or as locally to regionally dispersed pyroclastic deposits, from composite cones at the southern and northern extremities of the TVZ. These cones range in size from c. 10 up to > 200 km^3 and are all inferred or known to be polygenetic structures, with extended lifetimes that exceed 300 ka in the cases of Ruapehu and Tongariro. Individual eruptions range up to the order 1–10 km^3. The associated pyroclastics occur dominantly as fall deposits having dispersal characteristics from strombolian to plinian, with only minor pyroclastic flow deposits on the composite cones themselves (e.g., Nairn and Self, 1978; Hackett and Houghton, 1986, 1989).

Within the central TVZ, composite cones also occur both as surface and buried edifices (Figs. 4 and 6), though of lesser volume (≤ 10 km^3) than their counterparts in the andesite-dominated TVZ extremities. Andesitic volcanism is also represented by:

(a) 3 ignimbrites, in the Mangakino, Matahana Basin and Broadlands areas (e.g., Wilson, 1986);

(b) buried discrete lava flows or shallow sill complexes intersected in geothermal drillholes (e.g., Grindley, 1965); and

(c) mixing components associated with more-silicic eruptives (e.g., Blake et al., 1992).

In addition, a diorite plutonic body intersected 2.46 km below ground level in a single drillhole at Ngatamariki

(Browne et al., 1992) represents the only in-situ plutonic material known in the TVZ; it has no known extrusive equivalent.

Eruptive styles of andesites cover a wider spectrum than those of basalts, with flow deposits representing a significant proportion of the pyroclastic products. As with basalts, a wide range of magma–water ratios is inferred, from extremely wet (e.g., White Island; Houghton and Nairn, 1991) to dry fire-fountaining strombolian activity (e.g., Ngauruhoe; Gregg, 1956). The associated fall deposits are more voluminous and widely dispersed than the basaltic examples, with several late Quaternary fall deposits from Tongariro being of plinian dispersal (based on isopach data of Topping, 1973).

Distribution in time and space

A map of andesitic centres and individual eruptive sites is given in Fig. 6. Comparison with previous compilations (e.g., Cole, 1990) shows that many more andesites can be recognised, especially in the central TVZ. The resulting distribution of andesites (Fig. 6) shows that the geographic separation drawn by Cole between andesites and basalts in the central TVZ is not valid. In particular, for the andesites:

(1) Andesite (and associated dacite) landforms are conspicuous on the eastern margin of the TVZ only because synvolcanic tectonic subsidence is minimal here and thicknesses of rhyolitic eruptive products an order of magnitude less than along the axis of the TVZ. Andesite (to dacite) eruption products also occur as buried cones and domes, and as welded ignimbrites farther west in the TVZ, and andesite to dacite pyroclastic material is also documented as being mixed with, accompanying, or interfingering with, more voluminous rhyolitic products.

(2) During the time periods represented by the andesite (to dacite) centres used by Cole (1984, 1986, 1990) to define an eastern 'arc', numerous other vents of basalt, andesite, dacite and rhyolite composition

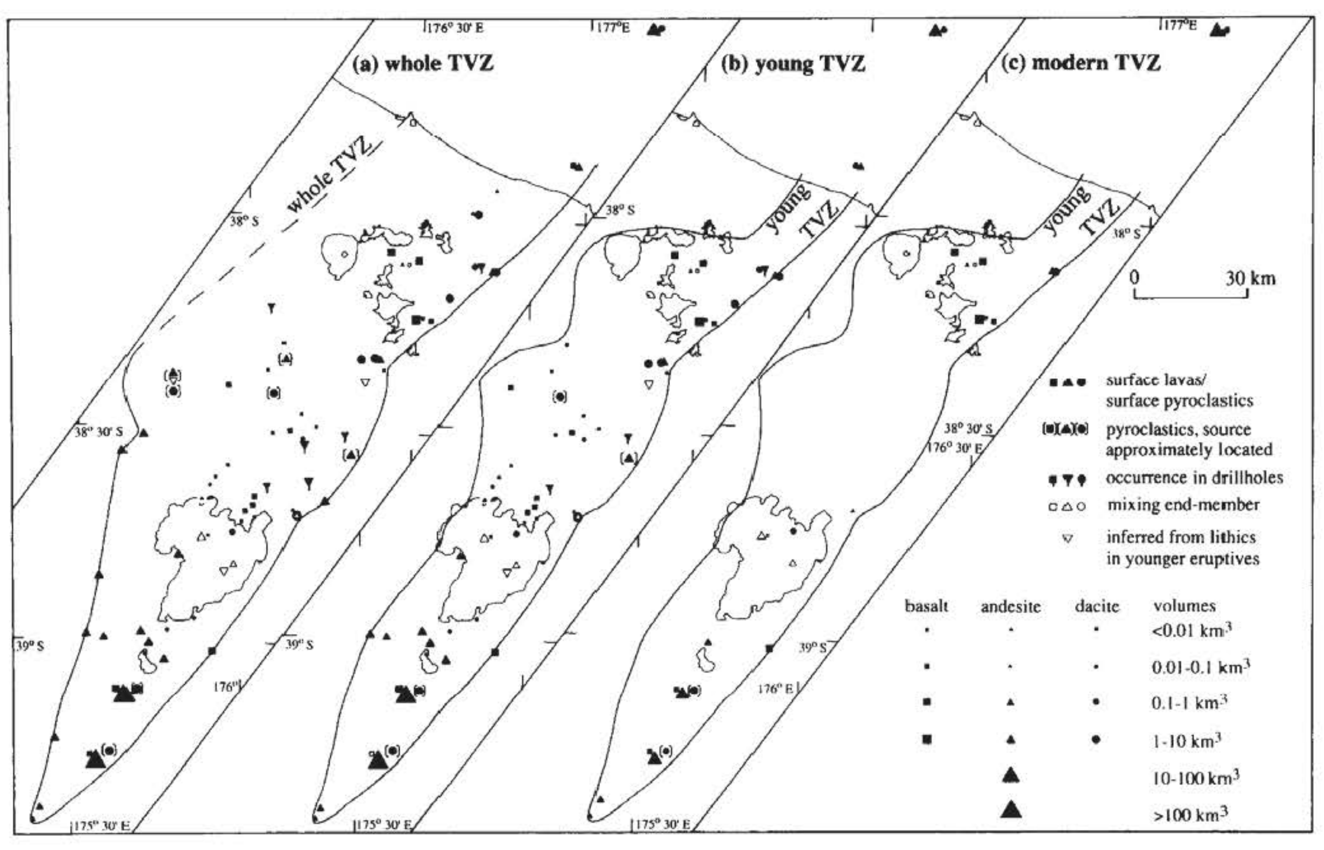

Fig. 6. Maps summarising all known occurrences of non-rhyolitic magma compositions in TVZ for: (a) whole TVZ, 2.0 Ma to present; (b) young-TVZ, 0.34 Ma to present; and (c) modern TVZ, 65 ka to present. Compiled from Cole (1978a), Browne (1978), Nairn (1981), Wood (1983), Houghton et al. (1987), Browne et al. (1992), Briggs et al. (1993), Sutton et al. (1995) and our unpublished data.

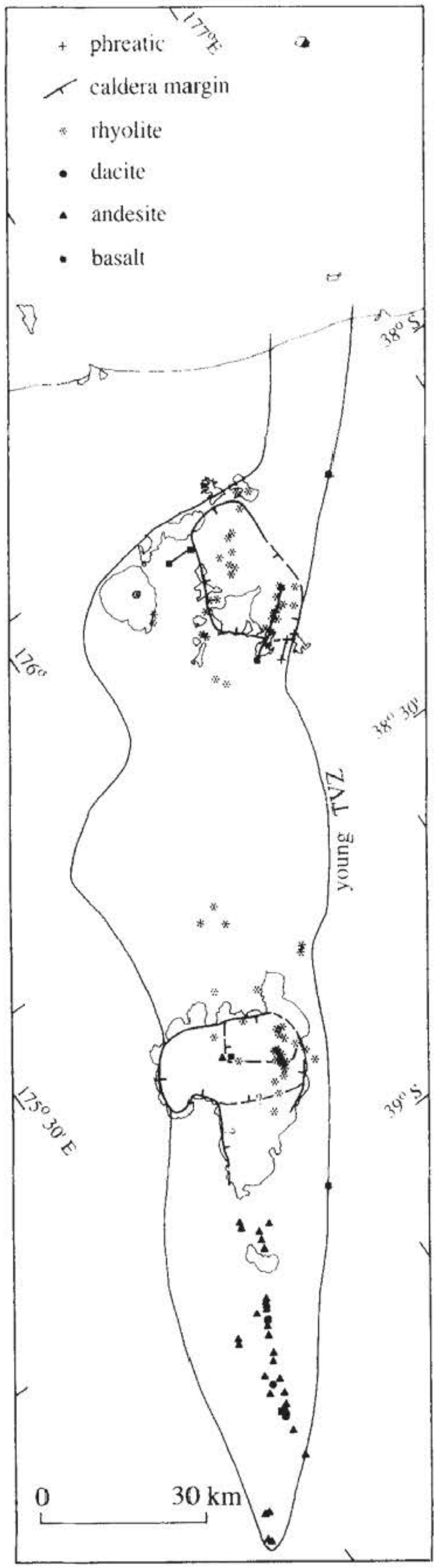

Fig. 7. Provisional compilation map of known or inferred vent sites and calderas formed during eruptions of the last 65 ka in modern TVZ (i.e. all eruptions including and following the Rotoiti eruption at Okataina). The following are omitted due to lack of information on the vent positions; the 38–26 ^{14}C ka fall deposits from within Okataina caldera (Howorth, 1975; G.P.L. Walker, pers. commun., 1994) and some of the c. 50 to c. 28 ka fall deposits from within Taupo caldera (Vucetich and Howorth, 1976; C.J.N. Wilson, unpubl. data). Caldera outlines from Houghton et al. (1995) and Davy (1993), vent sites compiled from Cole (1978a), Nairn (1981, 1989) Hackett (1985) and Wilson (1993).

have been active. These vents include other andesite-dacite centres west of the 'arc', and rhyolitic vents on the 'arc'. This is particularly the case for the modern TVZ, where compositions representative of all those known in the lifetime of the TVZ have been erupted from vents within a single 30-km-wide band (Fig. 7).

The compositions of andesitic eruptives vary widely within any particular time period, and there is no apparent geographical control on composition (e.g., Graham and Hackett, 1987). This, coupled with the volumes and lifetimes of the andesitic edifices, suggests that the processes generating andesite are not confined to any one area or segment of the TVZ, but that in the central TVZ there is a filter mechanism acting to partially block the passage of andesite to the surface. Andesite magmas have been generated sufficiently frequently and at high enough rates to maintain thermal conduits at least below the major centres at Ruapehu, Tongariro and White Island for a period of the order 10^5 years, but chronologies are not established for other centres.

3.4. Dacites

Eruptive forms and styles

Dacites are much less abundant than either andesites or rhyolites, and have a more restricted range of eruptive forms. Dacites are seen as:

(a) lavas with some subordinate pyroclastics that form either independent composite cones, or dome complexes, or portions of andesite-dacite cones (in no case exceeding c. 3 km^3; e.g., Reid and Cole, 1983; Graham and Worthington, 1988),

(b) hybrid or mixed pumices associated with dominantly rhyolitic ignimbrites (e.g., Sutton et al., 1995), or

(c) discrete fall deposits of sub-plinian to plinian dispersal both at andesite-dominated composite cones (e.g., Hackett and Houghton, 1986; Graham and Hackett, 1987) and rhyolitic calderas (e.g., Sutton et al., 1995).

The eruptive styles of the dacitic pyroclastics include a spectrum from dry to wet fall deposits (e.g., at Taupo; Wilson, 1993), as well as some discrete flow deposits (e.g., at Tauhara; Graham and Worthington, 1988).

Distribution in time and space

A map of dacitic centres and individual eruptive sites is given in Fig. 6. Note that almost all the surface dacite lavas known post-date the 0.34–0.32 Ma Whakamaru-group ignimbrites. As with basalts in the central TVZ, this is inferred to reflect burial of older outcrops by younger eruptives or destruction during caldera collapse. Compared with an earlier compilation (Reid and Cole, 1983) more examples of dacites are now known, in part because account is taken here of buried dacites drilled in geothermal fields, and also because of recognition of dacitic eruptions in the dominantly rhyolitic pyroclastic successions. Dacitic magmas appear to have been generated on timescales and/or in modest volumes such that they almost always form either single eruptions (or mixing compositions within single rhyolitic events), or cones with an apparently short lifetime (e.g., Edgecumbe; Duncan, 1970; Nairn, 1981), and have not formed long-lived volcanic systems.

3.5. Rhyolites

Eruptive forms and styles

Rhyolitic volcanism is manifested in 3 main forms.

(a) Large pyroclastic eruptions, of volumes between 30 and >300 km^3 (magma), generating ignimbrites, and with accompanying caldera formation.

(b) Pyroclastic eruptions, generating numerous but relatively small fall deposits and occasional ignimbrites, and involving magma volumes rarely >10 km^3. These events occur from vents within existing calderas, but are not accompanied by significant caldera collapse.

(c) Dome-building eruptions, with various quantities of locally to widely dispersed pyroclastic deposits. Few of these eruptions have involved >10 km^3 magma. They occur both from vents within pre-existing calderas as well as outside calderas (e.g., Wilson et al., 1986) and are not accompanied by caldera collapse.

Forms (b) and (c) are well displayed by the young activity at Okataina (Howorth, 1975; Nairn, 1981, 1989) and Taupo (Wilson, 1993), and available mapping and chronological evidence suggests similar styles and sizes of events are the norm at the rhyolitic volcanoes between episodes of caldera collapse. It is notable that the sizes of rhyolitic eruptions range over a wide spectrum, of the order 0.01 km^3 to 500 km^3 magma, and that the smallest known rhyolite events were probably comparable in size to many basaltic and andesitic events.

Eruptive styles range across the entire spectrum known for rhyolites, with the exceptions of rheomorphic ignimbrites or spatter-fed lavas. Pyroclastic fall deposits are known which span the range from wet (e.g., Self and Sparks, 1978; Walker, 1981a; Self, 1983) to dry (e.g., Walker, 1981b) and in dispersal from sub-plinian to plinian (e.g., Walker, 1980; Wilson, 1993). Ignimbrites range in volume from <1 to >300 km^3, and in emplacement temperature from non-welded to intensely welded. The most important controls on eruptive style in documented examples have been the inferred volume eruption rate and the timing and degrees of interaction with external water (e.g., Self, 1983; Wilson and Walker, 1985; Houghton and Wilson, 1989; Brooker et al., 1993), while the chemical compositions of the rhyolites (including their volatile contents) have seemingly played no vital role.

Distribution in time and space

A chronology for rhyolitic eruptions is largely established for the past 65 ka, although ongoing studies in the Maroa–Taupo (CJNW) and Okataina (G.P.L. Walker and Z. Jurado, pers. commun., 1994) areas are showing the presence of more eruptions from 22.5 ^{14}C ka to 65 ka than are accounted for in the literature. However, prior to 65 ka, chronologies are established for only the largest, caldera-forming eruptions (mostly from Ar/Ar dating, Houghton et al., 1995), and for a few domes (e.g., Houghton et al., 1991). Our observations imply that smaller-scale but more frequent activity of forms (b) and (c) above has always been present as a background between the major caldera-forming episodes.

There has been a continued divergence of approaches to the rhyolite centres in the central TVZ. Cole and co-workers (e.g., Cole, 1990) have continued and developed the tradition established by Healy (1962, 1964) that there are only 4 rhyolitic volcanic centres in the central TVZ, viz. Okataina, Rotorua,

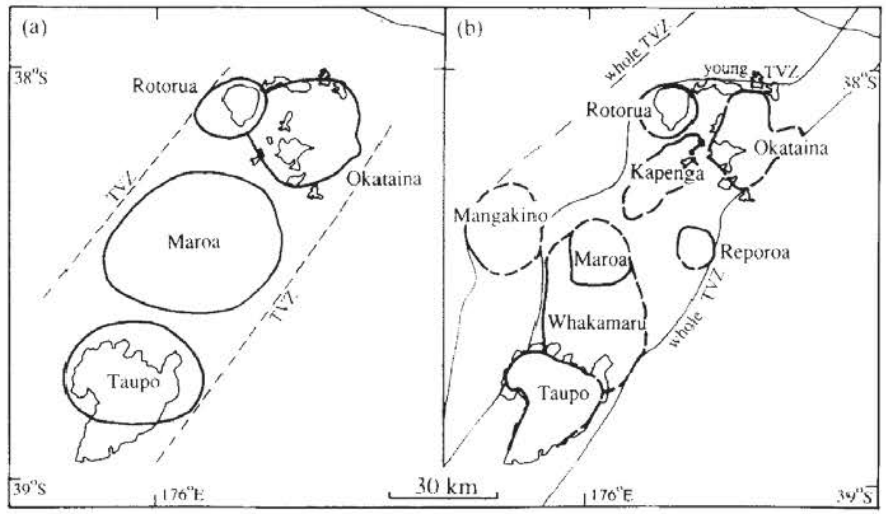

Fig. 8. Maps to show the outlines of (a) the 4 rhyolitic volcanic centres postulated to occupy the central TVZ by Cole (1990), versus (b) the 8 rhyolite caldera centres discussed in Wilson et al. (1984, 1986), Nairn et al. (1994) and Houghton et al. (1995). See text for discussion.

Maroa and Taupo (Fig. 8a), each of which represents a composite caldera together with surrounding vents. This view is tied in with an assertion that the maximum age of silicic volcanism in the central TVZ is 0.6 Ma, with opinion that the 'older centres to the west (e.g., Mangakino...) are not structurally part of the TVZ' (Cole, 1990, p. 451). We believe this view is untenable if the TVZ is defined on the basis of vent positions, as is implicit in the name. In contrast, we suggest that there are at least 8 recognisable caldera centres which have been developed in the central TVZ during its history (Fig. 8b), and it is likely that further centres remain to be delineated. Two centres encompass single caldera collapse episodes (Reporoa, Rotorua), four are compound features with multiple nested or overlapping collapses within a relatively short time period (Maroa, Okataina, Taupo, Whakamaru) and two are composite structures where collapse episodes related to independent periods of magmatism have coincided in geographic position (Mangakino, Kapenga). A summary of major activity at each centre is given in Fig. 9, and further details are available in Nairn (1981, 1989),

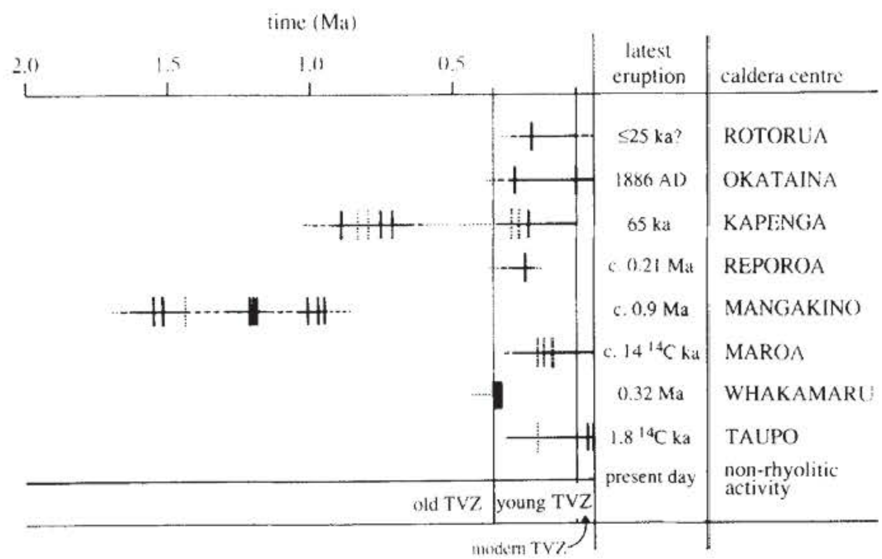

Fig. 9. Summary of the time spans occupied by eruptive activity at the central TVZ caldera centres. Vertical lines denote caldera collapse episodes; dotted lines denote that a collapse episode is inferred to have happened, but that the age is uncertain.

Wilson (1986), Wilson et al. (1984, 1986) and Nairn et al. (1994). These are commented on here where appropriate in the light of new evidence.

Not all silicic vents in the TVZ lie within or on the structural margins of recognised calderas. At Taupo, some of the extra-caldera eruptives are linked chemically to other units erupted from within the caldera (Sutton et al., 1995) and it is clear that 'leakage' of magma from a common source occurs beyond caldera margins (cf. Yellowstone, where the intra-and extra-caldera rhyolites are very different; Hildreth et al., 1991). However, in all cases where data are adequate, it is clear that the relative volumes of intracaldera products (including those associated with caldera formation) versus extra-caldera products are in a ratio typically of 5–50:1. The calderas formed at each rhyolite volcanic centre are thus separate in time and/or space, but eruptive products from the magma chambers responsible for caldera-forming eruptions can overlap in time and space thus requiring arbitrary boundaries between centres, for example, Maroa versus Taupo (Wilson et al., 1986; Sutton et al., 1995).

Rotorua

Formation of Rotorua caldera has been linked to the eruption of the Mamaku Ignimbrite (now dated at 0.22 ± 0.01 Ma; Houghton et al., 1995), and drillhole and geophysical information demonstrate substantial displacement and inferred thickening of the ignimbrite across the southern boundary of the caldera (Wood, 1992; Lamarche, 1992). In addition, an earlier ignimbrite (Pokai) whose age is bracketed by other ignimbrites with ages of 0.32 and 0.22 Ma is suggested as also coming from the Rotorua caldera (Wood, 1992). The youngest activity at Rotorua was the eruption of 3 lava domes at ≤ 25 ka and the magmatic system here possibly remains viable.

Okataina

The Okataina Volcanic Centre was previously linked by Nairn (1981, 1989) to eruption of 4 ignimbrites, but current models definitely attribute only the 0.28 ± 0.01 Ma Matahina Ignimbrite (Bailey and Carr, 1994; date from Houghton et al., 1995) and 65 ka Rotoiti eruption (Nairn, 1981; Davis, 1985; Wilson et al., 1992) to caldera collapses at this centre. Okataina, along with Taupo, remains highly active (average 0.08 $m^3 s^{-1}$ over the past 65 ka) and is among the most productive rhyolite volcanoes documented. Post-caldera activity at Okataina contrasts with Taupo in that there have been fewer but generally larger eruptions, often with substantial volumes of lava that have largely filled the younger caldera structure. The most recent activity was at Tarawera in 1886 AD (the largest known TVZ basalt eruption) and 700–900 years ago (the youngest TVZ rhyolite eruption).

Kapenga

New mapping and age determinations indicate that the area termed Kapenga volcanic centre by Rogan (1982) and Wilson et al. (1984) is a composite structure that developed during at least three periods of volcanism, two of them accompanied by inferred caldera-forming ignimbrites at 0.89 ± 0.04 to 0.68 ± 0.04, and 0.32 ± 0.02 to 0.22 ± 0.01 Ma, respectively (Houghton et al., 1995). The youngest activity in the Kapenga volcanic centre occurred in its NE portion, generating numerous lava domes and ending with the c. 65 ka Earthquake Flat ignimbrite eruption, which was not accompanied by caldera collapse. On the basis of our observations and unpublished mapping by E.F. Lloyd, I.A. Nairn and C.P. Wood, we consider that the Kapenga volcanic centre as presently defined consists of four temporally separate but geographically overlapping volcanic centres.

Reporoa

The Reporoa caldera had previously been identified as a structural basin due to lack of evidence for caldera-related volcanism (e.g., Rogan, 1982; Wilson et al., 1984). New mapping however has identified a lithic lag breccia facies within the 0.23 ± 0.01 Ma (Houghton et al., 1994) Kaingaroa Ignimbrites, previously attributed to Okataina, along the east side of the structural depression and it is now interpreted as a caldera (Nairn et al., 1995). The Reporoa caldera seems to reflect only this one collapse event, but it is notable among the calderas in the young TVZ for the volumes of andesite to dacite lavas erupted in and adjacent to the caldera (Nairn et al., 1994). Post collapse volcanic activity is apparently limited to 3 rhyolite domes on the southern margin, one of which has been K/Ar dated at 0.216 ± 0.019 Ma (B.F. Houghton et al., unpubl. data).

Mangakino

New Ar/Ar age data from ignimbrites inferred to be from the Mangakino volcanic centre show that activity there occurred in two distinct periods, from c. 1.6 to 1.53 ± 0.04 Ma and 1.21 ± 0.04 to 0.95 ± 0.03 Ma (Houghton et al., 1995), and it seems probable that the Mangakino caldera as originally envisaged is a composite structure. Data from a drillhole near Mangakino township (C.P. Wood, unpubl. drillhole log, 1987) show that there are 280 m of lacustrine sediments between the Whakamaru-group ignimbrites and an older unidentified ignimbrite, implying that the area formed a lacustrine depocentre up to 0.34 Ma. Burial by pre-0.34 Ma sediments and the Whakamaru-group ignimbrites has obscured any post-caldera eruptive activity, but lithic fragments of rhyolite lava in some of the Mangakino ignimbrites argue for at least some pre-caldera effusive activity.

Whakamaru

The Whakamaru caldera was proposed by Wilson et al. (1986) as a source for the Whakamaru-group ignimbrites, which have ages from Ar/Ar methods of between 0.34 and 0.32 Ma (Pringle et al., 1992; Houghton et al., 1995). Coarse lithic breccias have been reported from Whakamaru-group ignimbrites on the western margins of the caldera (S.J.A. Brown, pers. commun., 1994) and in sections exposed in the Paeroa Fault scarp some 10 km beyond the proposed caldera margin (Keall, 1988), the latter suggesting that structural collapse could have occurred beyond the caldera boundary of Wilson et al. (1986). On the basis of palaeomagnetic anisotropy measurements Lamarche and Froggatt (1993) proposed Whakamaru source vents which extended from southern Lake Taupo north for c. 60 km. However, these vents are inconsistent with the distribution of the deposits and lithic size data (Wilson et al., 1986; S.J.A. Brown, pers. commun., 1994), and the flow directions inferred can be shown or inferred largely to represent late-stage movement along palaeovalleys. All reliable data known to us are only consistent with the Whakamaru-group ignimbrites coming from sources north of Lake Taupo. Maroa and Taupo composite calderas are now considered to be entirely younger structures developed within and beyond the northern and southern margins, respectively, of the Whakamaru caldera.

Maroa

The definition of Maroa volcanic centre was modified by Wilson et al. (1986) so as to include only the composite Maroa caldera and its associated dome complex. Field stratigraphic evidence and new age data on rhyolite domes and basalts (B.F. Houghton et al., unpubl. data) imply that Maroa entirely postdates the Whakamaru caldera centre. Field data imply that activity at Maroa diminished from 60–30 ka as activity at Taupo increased in volume and intensity (Wilson, 1993), until at today the magmatic system at Maroa can be considered only feebly active.

Cole (1979, 1984, 1990) has used the term Maroa Volcanic Centre to refer to a larger structure (Fig. 8a) which was introduced by Healy (1964) as the Mokai Ring Complex, and inferred to be a caldera. However, all subsequent work has failed to support the concept of the Mokai Ring Complex, for 3 reasons. First, the 'ring' simply links a number of rhyolite and dacite domes and/or cones, which in themselves show evidence for vent alignments or structural controls along trends at various angles to the ring. Second, there is no evidence for displacement across the 'ring' except where it coincides with the western margin of the Whakamaru caldera (cf. Fig. 8a, b). Third, mapped regional faults and several older and younger caldera boundaries cut across the proposed 'ring' structure.

Taupo

Volcanism at Taupo is now interpreted to post-date entirely the Whakamaru-group ignimbrites, and the present-day Taupo volcanic centre lies across and south of the southern margin of the inferred Whakamaru caldera (Wilson et al., 1986). Although age data are sparse, field data (E.F. Lloyd and C.J.N. Wilson, unpubl. data) suggest that the early history (0.32 Ma to 65 ka) of Taupo is complex, involving at least 1 caldera-forming episode. The 22.6 ^{14}C ka Oruanui eruption is now interpreted to be the major caldera-forming event at Taupo (e.g., Wilson, 1991). The volcano has been more frequently active since the Oruanui event than previously supposed, with evidence for 28 eruptive events (Wilson, 1993). The distribution of repose intervals and eruptive volumes suggest that the volcano may represent a chaotic system, as there is apparently no relationship between eruption volumes and the repose periods either before or after each event.

Petrological studies (e.g., Sutton et al., 1995) show that many of the Taupo rhyolites can be grouped on chemical and isotopic evidence into magma batches which shared a common source or magma body, and that sizeable batches of magma appear to have been generated and erupted rapidly, on time scales of thousands to a few tens of thousands of years. Taupo has a mean eruption rate of c. 0.2 $m^3 s^{-1}$ over the past 65 ka, making it the most productive single rhyolitic volcano known.

Other possible calderas in the TVZ

Existing geological and geophysical data (e.g., coarse lithic breccias, seismically observed cross-cutting structures) imply that other caldera structures remain to be demarcated in the central TVZ, in addition to those which may be delineated within existing composite structures at Kapenga or Mangakino. There are two possible end-member models which encompass the possible relationships between the caldera volcanoes in the TVZ.

(1) Each large ignimbrite eruption is represented by one caldera; any overlap or nesting of calderas is coincidental. The central TVZ as a whole could thus be considered as one giant (125 × 60 km) composite caldera volcano.

(2) Caldera collapse is confined to restricted areas of the central TVZ which are currently labelled as 'caldera volcanic centres'. Other parts of the central TVZ never see substantial silicic volcanism, but are gradually buried by ejecta and downwarped by subsidence accompanying regional extension and volcanism elsewhere.

On present data it is not possible to distinguish between these two extreme views. For example, geophysical evidence at Taupo implies that the caldera formed at 1.8 ^{14}C ka was nested within the 22.6 ^{14}C ka Oruanui caldera (Northey, 1982; Davy, 1993), but petrological data show that the rhyolite erupted at 1.8 ^{14}C ka was distinctively different from that involved in the Oruanui event (Sutton et al., 1995) and, it could be argued, represents an entirely separate magma batch.

There are at least two other areas, not presently defined as calderas, known to us where available information suggests large-scale pyroclastic volcanism with accompanying caldera collapse may have occurred.

(1) In the Broadlands area, where an extensive, thick, lithic-rich pyroclastic unit (Rautawiri Breccia) is known from drillholes (e.g., Grindley and Browne, 1968; Wood, 1983). An andesitic welded ignimbrite of similar age and (from lithic sizes) derived from a nearby source crops out nearby on the Kaingaroa Fault scarp to the east (S.J.A. Brown, pers. commun., 1994), but is apparently not the same unit as recorded in the drillholes (C.P. Wood, pers. commun., 1994).

(2) In the southern Kaimai area, where coarse, lithic-rich breccias occur in the lower Pleistocene (not reliably dated) Waiteariki Ignimbrite (Houghton and Cuthbertson, 1989). Any possible source caldera seems most likely to be concealed beneath younger welded ignimbrites in the northern Mamaku Plateau area.

Note that any new calderas demonstrated in the Kaimai area (and also possibly present in the offshore Bay of Plenty; Davey et al., 1995) would demarcate the NW limits of the old TVZ and the whole TVZ much more clearly than is possible at present.

4. Observations and models of the TVZ

4.1. Limitations of data

Several models have been put forward to explain the present structure and/or evolution of the TVZ, and these are reviewed below. However, it is important to realise that all these models have several limitations.

(1) *The availability of suitable radiometric age data*. Until recently many key stratigraphic units had not been dated, and age data for others are of questionable accuracy, because of various limitations. For example, several of the older andesites in the Coromandel–Kaimai area used as evidence for the SE-ward migration of andesitic centres (section 4.2, below) are in areas of hydrothermal alteration and thus suspect (e.g., note the 16.2 to 2.6 Ma spread of ages for one andesite unit reported in Adams et al., 1974). Some other ages are fission-track determinations on glass which are now known to be anomalously young from comparisons between Ar/Ar and fission-track ages of other units in the TVZ (e.g., Alloway et al., 1993).

(2) *The nature of the exposures available*. The combined effects of episodic eruption of moderate to large ignimbrites, caldera collapse events, and burial by primary volcanic and secondary volcaniclastic deposits in areas of regional tectonic subsidence in the zone, means

that evidence available from surface outcrops for understanding the evolution of the zone is biased and incomplete (Fig. 3). Only the history of the TVZ for the last 65 ka or so can be considered to be reasonably complete for some areas. Clearly also, geological. geochemical and geophysical coverage of the submarine, offshore extension of the TVZ is presently limited in scope (e.g., Wright, 1992; Gamble et al., 1993a; Gamble and Wright, 1995; Davey et al., 1995).

(3) *The detail of geological studies.* Our existing understanding of the tectonic structures, source volcanoes, eruptive histories, and magma compositions in the TVZ is rapidly evolving. Recent examples of new information include recognition of the complex submarine structure of the northernmost TVZ (Wright, 1990, 1992, 1993a,b), identification of a new caldera centre in the Reporoa area (Nairn et al., 1994), the discovery that the number of late Quaternary eruptions at Taupo volcano has been seriously underestimated (Wilson, 1993), and the recognition of complex magmatic histories at young centres like Taupo (Sutton et al., 1995).

4.2. Central Volcanic Region and migrating andesite arc

Several workers (e.g., Stern, 1985; Cole et al., 1995 and references therein) consider the TVZ to represent merely the post-2 Ma part of a continuously evolving, migrating arc, active for > 20 Ma, whose position is defined at any stage by andesite and dacite centres. In particular, the wedge-shaped Central Volcanic Region (CVR) is inferred to represent the area swept out by movement of this arc in the past 4 Ma (Fig. 1 and Fig. 10a). The CVR is primarily defined on the basis of gravity anomalies and anomalous geophysical properties of the mantle, and considered to be a back-arc basin generated by spreading occurring perpendicular to the length of the migrating arc. The concepts of the CVR linked with the migrating andesite arc both remain popular (e.g., Cole et al., 1994), yet neither are supported by available evidence.

The geophysical anomalies by which the CVR is defined, especially its western margin, cut across all observable tectonic and geophysical elements which

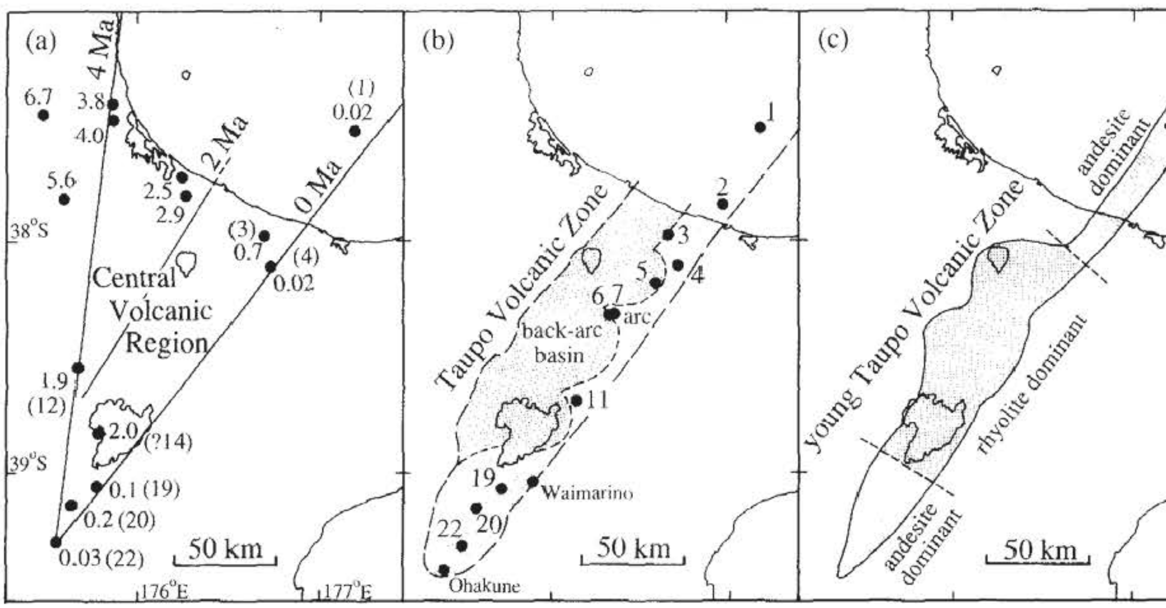

Fig. 10. Three maps to show the differing geometries of models or summaries of late Pliocene to Quaternary volcanism in the central North Island (see text for discussion). (a) The Central Volcanic Region (CVR) model of Stern (1987). Points marked are those used by Stern to define the envelope of the CVR and derive inferred time lines for the onset of andesitic volcanism; the numbers represent K/Ar ages used by Stern, and the bracketed numbers refer to those given to composite cones in Fig. 4 of this paper. (b) The TVZ structure of Cole (1984). Points marked are those centres variously used by Cole (1984, 1986, 1990) to define the 'arc'; numbers by each point are those in Fig. 4 in this paper (the un-numbered point is the Waimarino basalt (Hackett, 1985; Graham and Hackett, 1987). (c) The segmented structure of young TVZ as noted by Healy (1962) and Wilson et al. (1984).

are NNW–SSE-to N–S-orientated (e.g., Hauraki Rift; Hochstein et al., 1986). Several structural elements within the TVZ itself (e.g., parts of the margin to the Whakamaru caldera: Wilson et al., 1986) can be interpreted as reflecting the intersection of the roughly N–S 'Coromandel' and NNE–SSW 'TVZ' trends. This supports the notion that the TVZ trend has been superimposed on the earlier (Miocene–Pliocene) 'Coromandel' structures, rather than representing a continuous tectonic rotation from one orientation to the other (cf. Calhaem, 1973). The geochronological basis for delineating the CVR is not soundly based, as its area includes places where the (andesitic) volcanism goes back to at least the Miocene (e.g., Ballance, 1988). The surface geology map (New Zealand Geological Survey, 1972) used by various authors (e.g., Stern, 1985) to define the CVR utilises incorrect ages for some key units and allocates others to a Quaternary age without supporting radiometric age data. We thus suggest the term CVR be abandoned.

The concept of the migrating andesite arc was first quantified on the basis of limited K/Ar data by Stipp (1968), enlarged on by others (e.g., Brothers, 1986), and since has been promoted mostly by geophysical workers (e.g., Sissons, 1979; Stern, 1985, 1987; Smith et al., 1989; Cole et al., 1995) on the basis of the same K/Ar age data base. Support for this was derived from similarities between geodetically determined widening-rate estimates for the TVZ and the perceived migration rate for the onset of andesite volcanism. In the past 4 Ma, the arc migration is postulated to have incorporated rotation, with a hinge near Ruapehu, and to have been accompanied by asymmetric spreading of the CVR with growth of new 'granitic to andesitic' crust under the CVR (Stern, 1985, 1987). This migrating-arc concept is fundamentally flawed in many ways.

(1) The age data used to define the arc represent a selective suite of age determinations that do not date the *onset* of andesitic activity (as is required for validity of the migrating-arc model), but merely the material sampled at or near the present-day land surface. Onset ages are poorly defined in most places due to subsequent burial of the rocks concerned, but available radiometric and palaeontological age data refute the concept of an andesite arc which progressively migrates SE across the North Island (Ballance, 1988). Instead, it appears likely that at some stage between c. 4 Ma and 2 Ma there was a switch from Miocene–Pliocene arc volcanism associated with NNW–SSE trends (e.g., Coromandel; Skinner, 1986) to Pliocene–Quaternary arc volcanism on the NE–SW trend of the TVZ.

(2) The eastern margin of CVR (i.e. the eastern margin of the modern TVZ) is not a simple andesite arc as postulated by Stern (1985, et seq.) and Cole (e.g., 1990), as this area has been the locus of eruptions of all compositions from basalt to rhyolite for at least the last c. 0.71 Ma (see Fig. 4 and Fig. 6). No measurable migration of this eastern boundary has occurred in this time period, during which c. 15 km of migration and accompanying spreading should have occurred, according to estimates from the perceived migration rate of the andesitic volcanoes (Sissons, 1979; Stern, 1987).

(3) The asymmetric spreading postulated to occur as part of the migrating-arc model is not supported by the geological evidence. First, evidence is wholly lacking for systematic migration of andesitic vents; for example, along the eastern margin of the CVR and the TVZ eruptive activity of this composition ranges in age from 0.71 Ma at Rolles Peak (B.F. Houghton et al., unpubl. data) to 3.5 ka in the dominantly rhyolitic Waimihia eruption at Taupo volcanic centre (Blake et al., 1992). Second, basement greywacke has been reached by drilling at Rotokawa, Ohaaki and Kawerau geothermal fields (Browne, 1978; Wood, 1983; Browne et al., 1992), crops out on land near the Bay of Plenty coast (Healy et al., 1964), and is reported from the offshore Bay of Plenty (e.g., Gamble et al., 1993a). In all cases the greywacke crops out within the proposed zone of spreading, in areas where the model would require that greywacke should not occur. Seismic data from the offshore Bay of Plenty are used to infer that faulted greywacke underlies the volcanic succession there to a depth of c. 5 km, and count against wholesale replacement of the upper crust by igneous rocks (Davey et al., 1995).

(4) If the triangular wedge of the CVR represents opening up of the central North Island and formation of new crust by asymmetric spreading along the migrating arc, there are major problems in accommodating such a scenario within the tectonic framework of the North Island and adjacent areas. Pre-Quaternary sediments east of the CVR along the Bay of Plenty coast show no rotation relative to their counterparts west of the CVR (e.g., Walcott, 1987) which would be noticeable had the CVR opened up as a simple wedge. The

lack of rotation cannot be explained by the CVR opening being accommodated by compensatory dextral slip along faults in the North Island shear belt, because in turn there is no evidence for the amounts of displacement required where the relevant faults meet the eastern TVZ margin or cut the edge of the continental shelf (Wright, 1990, 1993b). Historical geodetic deformation data (e.g., Darby and Williams, 1991; Cole et al., 1995) demonstrate that widening rates of the TVZ are not systematically increasing along the axis of the zone from south to north as predicted by the spreading model.

4.3. *Andesite-dacite arc: bimodal back-arc basin*

This model proposes that the TVZ is represented by two distinct structures, an eastern arc defined by a line of andesite-dacite composite volcanoes, and a western marginal basin (the Taupo–Rotorua Depression) where rifting and subsidence is occurring and the volcanism is bimodal, with dominant rhyolites accompanied by minor amounts of high-Al basalt (e.g., Cole, 1979, 1984, 1990; Reid and Cole, 1983; Cole et al., 1995; Fig. 8b). However, this model is not supported by available data, in that the andesite-dacite centres (of ages from 0.71 Ma to <10 ka) used to define the eastern arc represent only a selection of the sites where andesite-dacite magmas have erupted in the TVZ during this time period (compare Fig. 4 and Fig. 8a). There is no simple spatial separation of high-Al basalt and andesite, and the patterns of basaltic and andesitic eruptives are complex and overlapping in time and space. Activity at the rhyolitic centres is not simply bimodal (rhyolite plus high-Al basalt), and a broad spectrum of non-rhyolitic compositions is found at all centres where adequate data are available (e.g., Sutton et al., 1995).

Studies of gas components (CO_2, Ar, He, N_2) in geothermal fluids (e.g., Giggenbach et al., 1993; Giggenbach, 1995) have been used to support separation of the TVZ into 'basaltic back-arc' (e.g., low CO_2/He ratios) and 'andesitic arc' (high CO_2/He ratios) areas, but again, the data are at odds with this. For example, ratios of CO_2/He reported by Giggenbach et al. (1993, their fig. 6) do not reflect either the compositions of associated historic eruptions, or the positions of sample points with respect to Cole's (1990) boundary between his 'arc' and 'back-arc' areas.

4.4. *Plastic deformation and crustal fusion below the TVZ*

It has been proposed (Hochstein and Regenauer-Lieb, 1989; Hochstein et al., 1993) that the unusually high rates of surface heat flux and rhyolitc volcanism in the TVZ are anomalous compared to other arcs. They consider these anomalies reflect processes whereby the TVZ acts as the hinge-line for rotation (at c. $6°\ Ma^{-1}$; Walcott, 1987) of the eastern side of the North Island with respect to the west (as in the spreading CVR model; section 4.2), and the resulting deformation causes heating in the ductile lower crust to generate the rhyolitc melts. This model, while addressing the problem of heat transfer in the TVZ crust, is contradicted in two key areas. First, long-term crustal deformation patterns measured by palaeomagnetic data show the postulated rotation of the eastern side with respect to the western side of the TVZ has not occurred (e.g., Walcott, 1987). Second, all available chemical data imply that the rhyolites cannot be the result of crustal fusion alone, but reflect a complex process of limited assimilation and prolonged fractionation from large amounts of mafic (mantle-derived) parents (e.g., Graham et al., 1995).

4.5. *Segmentation in the TVZ*

Some of the problems discussed above in relation to models of the TVZ arise because the models take no account of the variable nature of the TVZ along its length. Healy (1962) first drew attention to the segmented nature of the TVZ whereby rhyolitic activity in the onshore young TVZ is almost entirely confined to a central segment, where the fluxes of magma and heat transported by geothermal fluids are an order of magnitude or more higher (Fig. 10c; Wilson et al., 1984; Bibby et al., 1995). Hydrated low-Si rhyolite volcanics have recently been documented from the offshore part of the TVZ beyond White Island (Gamble et al., 1993a), but their volumes are uncertain, and do not negate the observation that rhyolitic vents are essentially absent between White Island and Okataina, and the total thermal flux in that sector much lower than to the south. No offshore calc-alkaline rhyolite centre or caldera on the scale of those in the central TVZ has been demonstrated to exist, and no Quaternary rhyolite pyroclastic deposit has been attributed to an offshore source.

The presence of the highly active, rhyolite-dominated, central 125-km-long segment in the TVZ is not easily explicable on present information. The dominance of rhyolite there has been attributed to the nature of crustal materials, specifically the inferred presence at depth of Miocene–Pliocene igneous rocks, forming a continuation of the Coromandel system (Cole, 1990), but this cannot in itself explain the accompanying extraordinary thermal flux. We speculate that superimposition of the NNE–SSW TVZ trend on the older NNW–SSE Coromandel trend may have resulted in additional heat sources being made available to fuel the high thermal flux below the central TVZ via higher generation rates of mafic magma below the crust.

5. Evolution of the TVZ

5.1. *Rhyolite eruptive forms and styles*

There is a widespread perception (e.g., Healy, 1971; Cole, 1984; Cole et al., 1995) that welded ignimbrites represent an earlier phase of activity at individual centres and within the zone as a whole, and that smaller, more frequent plinian or dome-building activity is now the norm. Several lines of evidence suggest this view of TVZ rhyolitic eruption styles is incomplete.

(1) It is clear that the welding states of the caldera-forming ignimbrites vary widely, and an increasing role is now recognised for caldera-forming phreatomagmatic eruptions which have generated deposits equally as big as the better known welded ignimbrites but which are poorly preserved and/or exposed and/or documented (e.g., Walker, 1979; Nairn, 1981; Self, 1983; Wilson, 1986, 1991). Major caldera-forming eruptions have in fact continued to the present, with the latest two (65 ka at Okataina and 22.6 ^{14}C ka at Taupo) happening to be phreatomagmatic and forming non-welded ignimbrites, together with voluminous and widespread fall deposits.

(2) The abundance of young post-collapse domes within and around caldera centres of the young TVZ leads to the misleading impression that dome-building activity only follows a cessation of caldera-forming eruptions. At all the young centres, age relationships of the domes, together with data from the lithic fractions of the caldera-related pyroclastics imply that dome-building events and their associated pyroclastics occur as volumetrically subordinate activity both before and after caldera formation.

(3) The apparent greater abundance of smaller volume pyroclastic deposits since 65 ka is in large part due simply to preservation. The majority of exposures which show contacts between the larger older deposits are in areas where the smaller post-65 ka deposits are also poorly preserved, either because of rapid erosion or because of an upwind position relative to the source areas. Where sections are available in favourable, little-eroded downwind areas outside the TVZ (e.g., Iso et al., 1982; Manning, 1994), the number of deposits preserved per unit time (calibrated against the dated ignimbrites) is comparable to that of ^{14}C dated units preserved at the same locality.

5.2. *Geochemistry of eruptive products*

Following the reconnaissance studies of Ewart (Ewart et al., 1975, and references therein), detailed petrological studies are becoming available for individual volcanoes (for example, Ruapehu: Graham and Hackett, 1987; White Island; Graham and Cole, 1991; Mangakino: Briggs et al., 1993; Taupo: Sutton et al., 1995). However, data are still sparse and rarely tied to precise chronology and volume relationships which permit evaluation of evolution with time or position within the individual volcanoes or the TVZ as a whole. In general though, no major overall changes in chemistry of the non-rhyolitic eruptives can be seen. Some of the earliest andesites in the TVZ at Pureora and Titiraupenga (Cole and Teoh, 1975; Froude and Cole, 1985) are similar to the lavas in centres younger than 0.25 Ma in the Tongariro–Ruapehu area (Cole, 1978a; Graham and Hackett, 1987). Earlier suggestions that olivine-bearing andesites have been erupted only in the last 50 ka (e.g., Cole, 1984) have not been borne out by subsequent work (e.g., Graham and Hackett, 1987).

For the caldera volcanoes, it has been proposed that there are differences in composition between earlier, lower SiO_2 (Mangakino) and later, higher SiO_2 (Taupo, Okataina) rhyolite eruptives (Hochstein et al., 1993). However, detailed or reconnaissance petrological studies available to us on the caldera-related eruptives (Fig. 11) suggest that any systematic differences are masked by the fact that the individual compositions

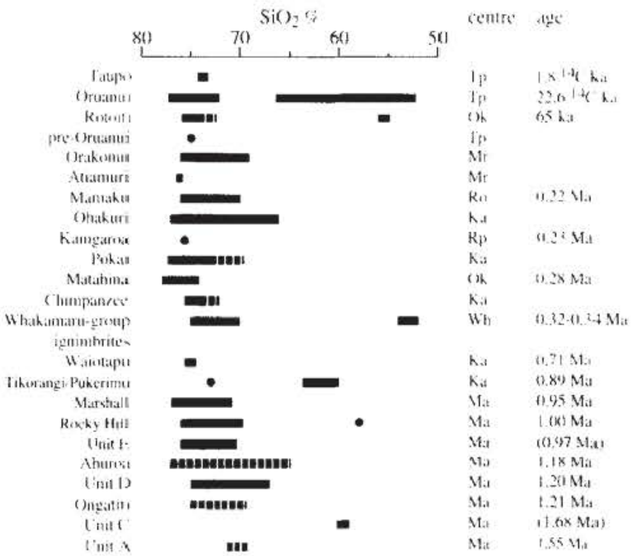

Fig. 11. Summary of the SiO_2 ranges (all calculated volatile-free, to total 100%) of the major, caldera-forming ignimbrite units in the central TVZ. Solid bars represent units where the variation in composition is thought to be continuous, dashed lines represent units where compositional discontinuities are thought to be present, circles represent single analyses. Centres are as labelled in Fig. 4. Sources of data are: Taupo, Oruanui, pre-Oruanui (Sutton et al., 1995); Rotoiti (Davis, 1985); Orakonui, Atiamuri, Mamaku, Ohakuri, Tikorangi/Pukerimu (S.D. Weaver and B.F. Houghton, unpubl. data); Kaingaroa (Nairn, 1981); Pokai, Chimpanzee (Karhunen, 1993); Matahina (Carr, 1984); Whakamaru-group ignimbrites (S.J.A. Brown, pers. commun., 1994); Waiotapu (R.M. Briggs, unpubl. data); Marshall, Rocky Hill, Unit E, Ahuroa, Unit D, Ongatiti, Unit C, Unit A (Briggs et al., 1993, R.M. Briggs, unpubl. data).

are not compared with their respective eruption volumes. Most large ignimbrites for which there are adequate data have compositions ranging from c. 69% SiO_2 up to high Si rhyolite (> 75% SiO_2) from at least 1.21 Ma onwards. On the other hand, there is evidence to suggest that some magma chambers which erupted to produce the larger welded ignimbrites from Mangakino (Briggs et al., 1993) and the earliest phase at Kapenga (S.D. Weaver and B.F. Houghton, unpubl. data) were zoned over a wider range of compositions than those of younger ignimbrites (e.g., Carr, 1984; Davis, 1985; Sutton et al., 1995). This is interpreted by Briggs et al. (1993) to suggest that a lack of crustal extension at Mangakino permitted development of compositionally zoned chambers (cf. Taupo; Sutton et al., 1995).

5.3. *Rifting in the TVZ*

Abundant geological and geodetic evidence suggests that at the present day the TVZ is actively widening (e.g., Sissons, 1979; Grindley and Hull, 1986; Walcott, 1987; Nairn and Beanland, 1989; Darby and Williams, 1991). This widening is generally inferred to be primarily rifting (i.e. thinning by faulting in the upper crust and ductile flow in the lower crust) rather than spreading (i.e. wholesale replacement of crust by igneous intrusions) because the balance of evidence is that the sub-volcanic, shallow crust under the TVZ is faulted greywacke rather than igneous intrusives (section 4.2, above). Key questions are for how long this rifting has been going on, and are modern (last c. 100 years), geodetically determined deformation rates, with WNW–ESE widening of the TVZ at 7–18 mm a^{-1} (if no horizontal contraction perpendicular to this direction is assumed), representative of the long-term rates? Three lines of evidence are important here.

(1) Elsewhere the process of rifting is invariably accompanied by normal faulting and tilting of originally sub-horizontal strata, with the amount of tilting increasing with time (e.g., Jackson and White, 1989). In the TVZ, ignimbrites older than c. 0.9 Ma (i.e. Tikorangi, Marshall and earlier units; Houghton et al., 1995) crop out immediately west of the young TVZ margin (as defined from vent sites) and are not significantly tilted or faulted. Ignimbrites back to 0.34 Ma immediately east of the young TVZ are not tilted (except possibly gently to the NNE along the axis of the zone), and are only faulted adjacent to and in the axial ranges in the North Island shear belt (Grindley, 1960). In contrast, over a broad area within the young TVZ, ignimbrites are tilted and intensively faulted (Fig. 12). These observations imply that since c. 0.9 Ma, rifting has largely or entirely been confined to the young TVZ, and this is supported by the distribution of geodetic strain rates (Sissons, 1979; see discussion in Darby and Williams, 1991). Evidence for any pre-0.9 Ma rifting in the young TVZ is obscured by deep burial of the relevant marker ignimbrites.

(2) There is evidence for a major episode of tectonic activity over wide areas of the central North Island at some stage between c. 1 Ma and 0.34–0.32 Ma. In the King Country west of the TVZ, major uplift and erosion occurred in the interval between the Rocky Hill (1.00 ± 0.05 Ma) and Whakamaru-group (0.34–0.32 Ma) ignimbrites, and the latter units occupy palaeovalleys in the same position, and cut to roughly the same depths, as the modern valleys (Wilson, 1986; revised ages from Houghton et al., 1995). To the south

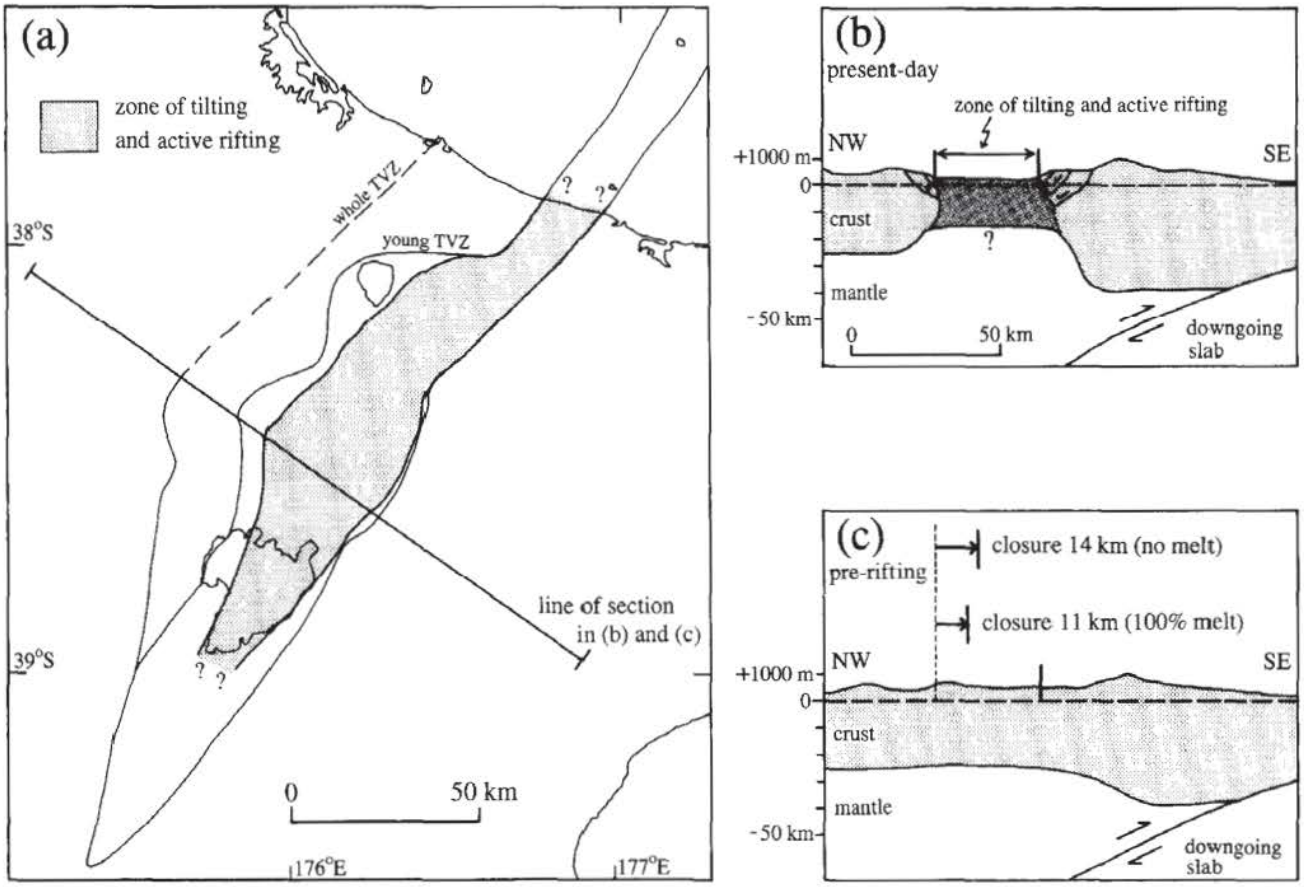

Fig. 12. (a) Map to show the area (stippled) within the young-TVZ where ignimbrites are significantly tilted and faulted, and hence inferred to be an area of active extension and rifting. The line of the cross section in (b) is from Smith et al. (1989) (b) Cross-section along the line marked in (a) to show the present-day crustal geometry presented by Smith et al. (1989). (c) Schematic cross section along the same line as in (b), but with closure by the amounts needed to restore the 'crust' (i.e. material with $V_p \leq 6.1$ km s^{-1}) to the thickness (25 km) reported for NW of the TVZ. The two figures for closure represent the two absolute extremes: (1) 'no melt' assumes crustal thinning is entirely through tectonic stretching; and (2) '100% melt' assumes all rhyolite magma represents crustal fusion products, and that all the rhyolite is dispersed outside the TVZ and does not contribute to filling the TVZ depression.

and east of the TVZ, distribution of a pyroclastic deposit recorded as the Potaka tephra (Shane, 1994) and Kidnapper Tuff (Seward, 1975) with a fission-track age of 1.05 ± 0.05 Ma (Alloway et al., 1993) is taken to suggest this unit predates the latest major uplift of the axial ranges (Shane, 1994). In contrast, in the axial ranges themselves, one of the Whakamaru-group ignimbrites (Te Whaiti) occupies a series of palaeo-valleys which represent a rugged palaeo-relief, exceeding 700 m, which is closely comparable with that present today (Wilson et al., 1986). The implication of these observations is that major uplift and dissection west, south and east of the TVZ occurred at some stage between c. 1 Ma and c. 0.34 Ma, as has been proposed by other workers (e.g., Yoshikawa, 1993, with allowance for revised ages of the pyroclastic marker horizons).

(3) Seismic profiles in the offshore Bay of Plenty are interpreted to show that the axis of rifting, tilting and subsidence migrated gradually eastwards (relative to the line of the seismic section) to its present position over the past c. 1.6 Ma (Davey et al., 1995). In contrast, the differing degrees of faulting and tilting within and outside the young TVZ imply that any movement of the rifting axis (if anywhere present onshore prior to 0.9 Ma), or the initiation of rifting, was abrupt.

Given the crustal cross section presented by Stern and Davey (1985, 1987; see Fig. 12), one crude technique to examine the timing and extent of rifting is to restore the thinned crustal material with $V_p \leq 6.1$

km s^{-1} beneath the TVZ to the same 25 km thickness proposed for the North Island NW of the TVZ, and estimate the amount of widening thus accounted for (Fig. 12). If this is done, then the c. 35 km average maximum width of the post-0.9 Ma TVZ (average width of the central young TVZ) would have originally been c. 21 km, and the concomitant widening rate c. 16 mm a^{-1}. If it assumed that crustal material has also been lost by assimilation and eruption as rhyolitic pyroclastic deposits beyond the TVZ margins (and that all TVZ rhyolites were generated by assimilation), then the original width would have been c. 24 km at most, and the corresponding widening rate 12 mm a^{-1}. These figures are remarkably similar to the geodetically determined widening rates (18 mm a^{-1} along a section close to that in Fig. 12; Darby and Williams, 1991). We thus suggest that either the present regime of rifting has only been extant at its modern rate for the last c. 0.9 Ma in the onshore TVZ or, if rifting has been going on for > 0.9 Ma, the widening rate must have increased with time.

6. Discussion and conclusions

The TVZ is an exceptionally active area of young volcanism and tectonism with a total extrusive magma flux of roughly 0.3 $m^3 s^{-1}$ (of which ≥ 90% is rhyolitic; Healy, 1962; Wilson et al., 1984). In addition, there is a modern geothermal heat flux of 4200 ± 400 MW (Bibby et al., 1995), which is interpreted as equivalent to another c. 1.2 $m^3 s^{-1}$ of magma intruded into the crust (Wilson et al., 1984). The central TVZ is the most frequently active and productive Quaternary rhyolitic system on Earth (Houghton et al., 1995), but differs from other arc-related silicic systems in the unusually high proportion of erupted rhyolite (Hochstein et al., 1993), and from non-arc-related silicic systems in the high productivity (Crisp, 1984) and wide spectrum of compositions extant at any one time (cf. Yellowstone; Hildreth et al., 1991). The volumes of basalt, andesite, dacite and rhyolite eruptives over the past 0.34 Ma are, respectively, c. 5, 300, 20, 3000 km^3 (based on our volume estimates), and the geographic distribution of rhyolite vents and erupted volumes of both rhyolites and andesites are notably uneven.

The reasons for the production of such huge volumes of rhyolite relative to other compositions, and the restricted geographic distribution of the rhyolite vents are unsolved major questions. The rhyolites are variously attributed to the following (see Graham et al., 1995, for further discussion):

(1) partial melting of some sub-crustal lithology (downgoing slab or mantle wedge), or fractional crystallization of melts therefrom (e.g., Blattner and Reid, 1982; Conrad et al., 1988);

(2) crustal fusion of greywacke or its lower crustal equivalent (e.g., Ewart and Stipp, 1968; Cole, 1981; Reid, 1983), or earlier intrusives (e.g., Cole, 1990; Graham et al., 1992) by intrusion of mafic magmas;

(3) crustal fusion (of unspecified lithologies) induced by plastic crustal deformation (e.g., Hochstein et al., 1993); or

(4) some combination of assimilation and fractional crystallization (e.g., McCulloch et al., 1994).

What does seem clear is that magmas generated in or below the mantle wedge are largely if not wholly basaltic (see references above), and that density filtering, and a complex and poorly understood series of assimilation and fractional crystallization processes take place to yield a dominantly rhyolitic assemblage at the surface (e.g., McCulloch et al., 1994; Graham et al., 1995). The major stumbling block to understanding the volumetric and petrogenetic relationships between the magmas, and the structure and hydrothermal circulation beneath the volcanic systems, is that the nature of the crust between 2.8 km and > 15 km depth is very poorly understood. Only limited data from certain areas are available from xenoliths (Ewart and Cole, 1967; Graham, 1987; Graham et al., 1990) which imply that, in addition to metasediments (equivalents of the surface greywackes), granitoids, meta-igneous granulites and mafic cumulates (some cognate) are present in the crust beneath the TVZ.

The distribution of vents, and eruptive compositions of composite cones and calderas, serve to emphasise the segmented nature of the TVZ, particularly over the last 0.34 Ma (Fig. 4; Wilson et al., 1984). Models that neglect this feature (Section 4, above) inherently cannot explain why young onshore rhyolitic volcanism is so rare north of Okataina or south of Taupo, or why the total thermal flux in the central TVZ is so high, both relative to other parts of the TVZ and on an absolute scale. In contrast to rhyolites, basaltic, andesitic and dacitic magmatism occurs along the entire length and width of the zone, with common models proposed for their petrogenesis throughout the zone (e.g., Graham and Hackett, 1987; Gamble et al., 1993b). In particular, we interpret any differences in trends between the various basalts (as divided on major elements) and their respective derivative andesites throughout the TVZ as reflecting crustal process such as fractionation acting under differing values of P_{H_2O} (e.g., Gaetani et al., 1993), rather than to any spatial variation in the major-element composition of the parental melts (Gamble et al., 1990, 1993b). Significant changes in petrogenesis of the non-rhyolitic magmas occur over the transition from oceanic to continental crust, and reflect the different nature of the mantle sources and influences of continental crust as a density filter and/or contaminant (e.g., Gamble et al., 1993a, b).

The distribution of vents and their magma compositions within the young and modern TVZ are not consistent with the presence of an intact eastern andesite-dacite arc and a rifted western marginal basin (Section 4.3). In overall terms, a better descriptor for the zone is a rifted arc (i.e. an arc disrupted by rifting), where the rate of widening varies along strike from essentially zero to 18 mm a^{-1}. Most of the present-day areas of highly active volcanism (Fig. 7) and high geothermal heat flux (Bibby et al., 1995) fall along a c. 15-km-wide corridor from Ruapehu to White Island, and we propose that this corridor represents the line of the subduction-related thermal locus. Rifting, at the current geodetically determined rates, could have generated the central TVZ with its thinned layer of material with $V_p \leq 6.1$ km s^{-1} within the last c. 1 Ma, and there is evidence from areas surrounding the TVZ for major tectonic activity after 0.9–0.95 Ma, possibly reflecting the onset or acceleration of rifting (Section 5.3). Data from seismic studies in the offshore Bay of Plenty are interpreted (Davey et al., 1995) to suggest that rifting may have commenced earlier there, and migrated south and east since c. 1 Ma, but evidence to confirm this onshore is presently lacking. There is some limited evidence from petrological studies (Section 5.2) to suggest a difference in magma chamber zonation between the oldest system at Mangakino and the active systems like Taupo, which may be attributable indirectly to the rifted nature of the crust at the present day. One unusual aspect of the central TVZ is that rhyolitic volcanism is currently concentrated at its northern (Okataina) and southern (Taupo) extremities (Figs. 4c, 6c, 7); does this mean that the rhyolitic-dominated segment of the TVZ is migrating, or merely that present activity fortuitously happens to be occurring there?

Even with the amount of interest and research in the Taupo Volcanic Zone, there are many aspects of its volcanic and structural evolution that remain obscure. Major aspects that are currently being addressed or need to be considered in the future include the following.

(1) *Chronology*. A skeletal chronology has been established for the major caldera-forming events, while limited data is available from smaller events, mostly in the last 0.34 Ma, and a detailed chronology established for the last 65 ka. However, the data are not yet adequate to define the ages and histories of the andesite composite cones and most of the offshore volcanism is not dated at all.

(2) *Eruptive styles and dynamics*. The various ways in which the different magma types erupt give valuable information on eruptive styles and dynamics, but also play an important role in biasing the picture of past activity. Those eruptions producing unconsolidated pyroclastics will have a lower preservation potential in the short to medium term (10^3 to 10^6 years) than those which produce welded ignimbrites or lavas, but in turn are more likely to leave a record as reworked volcaniclastic sediments in adjacent sedimentary basins where long-term preservation (10^6 to 10^{8+} years) is more likely. At present, our records of eruptions are only adequate for the last few tens of thousands of years at best (e.g., Taupo, Okataina), and for the historic era at worst (e.g., White Island). It is thus clear that our understanding of activity in the TVZ is controlled more by what is preserved than by what has actually occurred, and proxy sources of information need to be developed (see note 3, below).

(3) *Correlations*. At present, variable amounts of detail are available for primary eruption products in the TVZ itself and immediately surrounding areas, and for distal primary eruption and secondary reworked products in sedimentary basins on land and on the ocean floor. However, correlations between the two sources of information are still at an early stage. It is clear that a far more detailed record of TVZ activity may be preserved in the sedimentary record, but our lack of understanding of the role of eruption styles and volumes in controlling what material is delivered to the sedimentary depocentres is at present a major handicap to building a more detailed picture of TVZ activity.

(4) *Structure*. The surface traces of faults are mapped in detail throughout the TVZ, but surprisingly little is known about the displacements across the faults, their geometries at depth, and how much extension can be accounted for by faulting at shallow levels as opposed to deformation of weakly consolidated near-surface volcanic deposits. In addition, the relationships between the caldera-bounding fractures and regional faults have not been explored in any detail, with regard to such problems as the partition of displacements during caldera collapse, the relationships between regional deformation and caldera collapse events, and any potential for regional faults to act as conduits for magmatism. In addition, the complementary roles of rifting and magmatism in controlling the structure and heat flux in the TVZ have not been clearly demarcated.

(5) *The nature of the crust below the TVZ*. The very nature of the shallow lithologies in the TVZ makes acquisition of geophysical data difficult or renders the results ambiguous (e.g., Stern and Davey, 1985, 1987; Davy, 1993; Bibby et al., 1995). New approaches will need to be found to explore the crust between c. 2 and > 15 km depth if any real progress is to be made in delineating the lithologies at depth (particularly whether mafic cumulates or intrusions in the deep crust have been called 'anomalous mantle'), exploring the ways in which the crust is accommodating the rifting and magmatic intrusion, and locating and imaging magma bodies below the volcanoes or elsewhere.

(6) *The source of the surface heat flow*. The TVZ contains one of the highest arc-related surface heat fluxes per unit area documented (Stern, 1985, 1987; Hochstein et al., 1993; Bibby et al., 1995), and the deep source of this flux is generally accepted (*pace* Hochstein and Regenauer-Lieb, 1989; Hochstein et al., 1993) to represent magma being emplaced into the crust. Information is lacking about the method(s) by which c. 80% of this thermal flux is transmitted to the surface by geothermal fluids, versus c. 20% by volcanism (Wilson et al., 1984). By understanding the nature of the sub-TVZ crust (note 5, above), a better picture can be gained of the thermal partitioning and transmittal processes.

Acknowledgements

We thank the New Zealand Foundation for Research Science and Technology, the Royal Society of London and the UK Natural Environment Research Council for financial support. We acknowledge discussions with Hugh Bibby, Grant Caldwell and Peter Wood, and thank John Gamble, Jim Cole, Peter Wood and especially Wes Hildreth for reviews.

References

Adams, C.J.D., Wodzicki, A. and Weissberg, B.G., 1974. K-Ar dating of hydrothermal alteration at the Tui Mine, Te Aroha, New Zealand. N.Z. J. Sci., 17: 193–199.

Alloway, B.V., Pillans, B.J., Sandhu, A.S. and Westgate, J.A., 1993. Revision of the marine chronology in the Wanganui Basin, New Zealand, based on the isothermal plateau fission-track dating of tephra horizons. Sediment. Geol., 82: 299–310.

Bailey, R.A. and Carr, R.G., 1994. Matahina Ignimbrite Part I: physical geology and eruptive history. N.Z. J. Geol. Geophys., 37: 319–344.

Ballance, P.F., 1988. Late Cenozoic time-lines and calc-alkaline volcanic arcs in northern New Zealand —further discussion. J.R. Soc. N.Z., 18: 347–358.

Bibby, H.M., Caldwell, T.G., Davey, F.J. and Webb, T.H., 1995. Geophysical evidence on the structure of the Taupo Volcanic Zone and its hydrothermal circulation. J. Volcanol. Geotherm. Res., 68: 29–58.

Blake, S., Wilson, C.J.N., Smith, I.E.M. and Walker, G.P.L., 1992. Petrology and dynamics of the Waimihia mixed magma eruption, Taupo Volcano, New Zealand. J. Geol. Soc. London, 149: 193–207.

Blattner, P. and Reid, F.W., 1982. The origin of lavas and ignimbrites of the Taupo Volcanic Zone, New Zealand, in the light of oxygen isotope data. Geochim. Cosmochim. Acta, 46: 1417–1429.

Briggs, R.M., 1986. Petrology and geochemistry of Maungatautari, a medium-K andesite-dacite volcano. N.Z. J. Geol. Geophys., 29: 273–289.

Briggs, R.M., Gifford, M.G., Moyle, A.R., Taylor, S.R., Norman, M.D., Houghton, B.F. and Wilson, C.J.N., 1993. Geochemical zoning and eruptive mixing in ignimbrites from Mangakino volcano, Taupo Volcanic Zone, New Zealand. J. Volcanol. Geotherm. Res., 56: 175–203.

Brooker, M.R., Houghton, B.F., Wilson, C.J.N. and Gamble, J.A., 1993. Pyroclastic phases of a rhyolitic dome-building eruption: Puketarata tuff ring, Taupo Volcanic Zone, New Zealand. Bull. Volcanol., 55: 395–406.

Brothers, R.N., 1986. Upper Tertiary and Quaternary volcanism and subduction zone regression, North Island, New Zealand. J.R. Soc. N.Z., 16: 275–298.

Broughton, A.K., 1988. The geology of the Manawahe volcano, Bay of Plenty. B.Sc.(Hons) thesis, Victoria University, Wellington, New Zealand, 84 pp. (unpubl.).

Brown, S.J.A., Smith, R.T., Cole, J.W. and Houghton, B.F., 1994. Compositional and textural characteristics of the strombolian and surtseyan K-Trig basalts, Taupo Volcanic Centre, New Zealand: implications for eruption dynamics. N.Z. J. Geol. Geophys., 37: 113–126.

Browne, P.R.L., 1978. Petrological logs of drillholes, Kawerau geothermal field. N.Z. Geol. Surv. Rep., 84: 1–57.

Browne, P.R.L., Graham, I.J., Parker. R.J. and Wood, C.P., 1992. Subsurface andesite lavas and plutonic rocks in the Rotokawa and Ngatamariki geothermal systems, Taupo Volcanic Zone, New Zealand. J. Volcanol. Geotherm. Res., 51: 199–215.

Calhaem, I.M., 1973. Heat flow measurements under some lakes in the North Island, New Zealand. Ph.D. thesis, Victoria University, Wellington, New Zealand (unpubl).

Carr, R.G., 1984. The Matahina Ignimbrite: its evolution including its eruption and post-depositional changes. Ph.D. thesis, Auckland University, Auckland, New Zealand (unpubl.).

Christensen, N.I., 1982. Seismic velocities. In: R.S. Carmichael (Editor), Handbook of Physical Properties of Rocks, Volume II. CRC Press, Boca Raton, FL, pp. 1–228.

Cole, J.W., 1978a. Andesites of the Tongariro Volcanic Centre, North Island, New Zealand. J. Volcanol. Geotherm. Res., 3: 121–153.

Cole, J.W., 1978b. Tectonic setting of Mayor Island. N.Z. J. Geol. Geophys., 21: 645–647.

Cole, J.W., 1979. Structure, petrology, and genesis of Cenozoic volcanism, Taupo Volcanic Zone, New Zealand —a review. N.Z. J. Geol. Geophys., 22: 631–657.

Cole, J.W., 1981. Genesis of lavas of the Taupo Volcanic Zone, North Island, New Zealand. J. Volcanol. Geotherm. Res., 10: 317–337.

Cole, J.W., 1984. Taupo–Rotorua Depression: an ensialic marginal basin of North Island, New Zealand. In: B.P. Kokelaar and M.F. Howells (Editors), Marginal Basin Geology. Geol. Soc. London, Spec. Publ., 16: 109–120.

Cole, J.W., 1986. Distribution and tectonic setting of late Cenozoic volcanism in New Zealand. In: I.E.M. Smith (Editor), Late Cenozoic Volcanism in New Zealand. R. Soc. N.Z. Bull., 23: 7–20.

Cole, J.W., 1990. Structural control and origin of volcanism in the Taupo volcanic zone, New Zealand. Bull. Volcanol., 52: 445–459.

Cole, J.W. and Teoh, L.H., 1975. Petrography, mineralogy, and chemistry of Pureora andesite volcano, North Island, New Zealand. N.Z. J. Geol. Geophys., 18: 259–272.

Cole, J.W., Darby, D.J. and Stern, T.A., 1995. Taupo Volcanic Zone and Central Volcanic Region —back arc structures of North Island, New Zealand. In: B. Taylor (Editor), Back-arc Basins: Tectonism and Magmatism. Plenum Publishing, New York, NY, pp. 1–28.

Conrad, W.K., Nicholls, I.A. and Wall, V.J., 1988. Water-saturated and -undersaturated melting of metaluminous crustal compositions at 10 kb: evidence for the origin of silicic magmas in the Taupo Volcanic Zone, New Zealand, and other occurrences. J. Petrol., 29: 765–803.

Crisp, J.A., 1984. Rates of magma emplacement and volcanic output. J. Volcanol. Geotherm. Res., 20: 177–211.

Darby, D.J. and Williams, R.O., 1991. A new geodetic estimate of deformation in the Central Volcanic Region of the North Island, New Zealand. N.Z. J. Geol. Geophys., 34: 127–136.

Davey, F.J., Henrys, S.A. and Lodolo, E., 1995. Asymmetric rifting in a continental back-arc environment, North Island, New Zealand. J. Volcanol. Geotherm. Res., (this volume).

Davis, W.J., 1985. Geochemistry and petrology of the Rotoiti and Earthquake Flat pyroclastic deposits. M.Sc. thesis, University of Auckland, Auckland, New Zealand (unpubl.).

Davy, B.W., 1993. Seismic reflection profiling of the Taupo caldera, New Zealand. Explor. Geophys., 24: 443–454.

Duncan, A.R., 1970. The petrology and petrochemistry of andesite and dacite volcanoes in eastern Bay of Plenty, New Zealand. Ph.D. thesis, Victoria University, Wellington, New Zealand (unpubl.).

Ewart, A., 1982. The mineralogy and petrology of Tertiary–Recent orogenic volcanic rocks: with special reference to the andesitic-basaltic compositional range. In: R.S. Thorpe (Editor) Andesites. John Wiley and Sons, Chichester, pp. 25–95.

Ewart, A. and Cole, J.W., 1967. Textural and mineralogical significance of the granitic xenoliths from the Central Volcanic Region, North Island, New Zealand. N.Z. J. Geol. Geophys., 10: 31–54.

Ewart, A. and Stipp, J.J., 1968. Petrogenesis of the volcanic rocks of the central North Island, New Zealand, as indicated by a study of Sr^{87}/Sr^{86} ratios, and Sr, Rb, K, U and Th abundances. Geochim. Cosmochim. Acta, 32: 699–735.

Ewart, A., Hildreth, W. and Carmichael, I.S.E., 1975. Quaternary acid magma in New Zealand. Contrib. Mineral. Petrol., 51: 1–27.

Froggatt, P.C., 1983. Towards a comprehensive upper Quaternary tephra and ignimbrite stratigraphy in New Zealand using electron microprobe analysis of glass shards. Quat. Res., 19: 188–200.

Froggatt, P.C. and Lowe, D.J., 1990. A review of late Quaternary silicic and some other tephra formations in New Zealand: their stratigraphy, nomenclature, distribution, volume, and age. N.Z. J. Geol. Geophys., 33: 89–109.

Froude, D.O. and Cole, J.W., 1985. Petrography, mineralogy and chemistry of Titiraupenga volcano, North Island, New Zealand. N.Z. J. Geol. Geophys., 28: 487–496.

Gaetani, G.A., Grove, T.L. and Bryan, W.B., 1993. The influence of water on the petrogenesis of subduction-related igneous rocks. Nature, 365: 332–334.

Gamble, J.A. and Wright, I.C., 1995. The southern Havre Trough: geological structure and magma petrogenesis of an active back-arc rift complex. In: B. Taylor (Editor), Back-arc Basins: Tectonism and Magmatism. Plenum Publishing, New York, NY, pp. 29–62.

Gamble, J.A., Smith, I.E.M., Graham, I.J., Kokelaar, B.P., Cole, J.W., Houghton, B.F. and Wilson, C.J.N., 1990. The petrology, phase relations and tectonic setting of basalts from the Taupo Volcanic Zone, New Zealand, and the Kermadec Island Arc–Havre Trough, S.W. Pacific. J. Volcanol. Geotherm. Res., 43: 253–270.

Gamble, J.A., Wright, I.C. and Baker, J.A., 1993a. Seafloor geology and petrology in the oceanic to continental transition zone of the Kermadec–Havre–Taupo Volcanic Zone arc system. N.Z. J. Geol. Geophys., 36: 417–435.

Gamble, J.A., Smith, I.E.M., McCulloch, M.T., Graham, I.J. and Kokelaar, B.P., 1993b. The geochemistry and petrogenesis of basalts from the Taupo Volcanic Zone and Kermadec Island Arc, S.W. Pacific. J. Volcanol. Geotherm. Res., 54: 265–290.

Giggenbach, W.F., 1995. Variations in the chemical and isotopic composition of fluids discharged over the Taupo Volcanic Zone. J. Volcanol. Geotherm. Res., 68: 89–115.

Giggenbach, W.F., Sano, Y. and Wakita, H., 1993. Isotopic composition of helium, and CO_2 and CH_4 contents in gases produced along the New Zealand part of a convergent plate boundary. Geochim. Cosmochim. Acta, 57: 3427–3455.

Graham, I.J., 1987. Petrography and origin of metasedimentary xenoliths in lavas from Tongariro Volcanic Centre. N.Z. J. Geol. Geophys., 30: 139–157.

Graham, I.J. and Cole, J.W., 1991. Petrogenesis of andesites and dacites of White Island volcano, Bay of Plenty, New Zealand, in the light of new geochemical and isotopic data. N.Z. J. Geol. Geophys., 34: 303–315.

Graham, I.J. and Hackett, W.R., 1987. Petrology of calc-alkaline lavas from Ruapehu volcano and related vents, Taupo Volcanic Zone, New Zealand. J. Petrol., 28: 531–567.

Graham, I.J. and Worthington, T.J., 1988. Petrogenesis of Tauhara dacite (Taupo Volcanic Zone, New Zealand) — evidence for magma mixing between high-alumina andesite and rhyolite. J. Volcanol. Geotherm. Res., 35: 279–294.

Graham, I.J., Blattner, P. and McCulloch, M.T., 1990. Meta-igneous granulite xenoliths from Mount Ruapehu, New Zealand: fragments of altered oceanic crust? Contrib. Mineral. Petrol., 105: 650–661.

Graham, I.J., Gulson, B.L., Hedenquist, J.W. and Mizon, K., 1992. Petrogenesis of Late Cenozoic volcanic rocks from the Taupo Volcanic Zone, New Zealand, in the light of new lead isotope data. Geochim. Cosmochim. Acta, 56: 2797–2819.

Graham, I.J., Cole, J.W., Briggs, R.M., Gamble, J.A. and Smith, I.E.M., 1995. Petrology and petrogenesis of volcanic rocks from Taupo Volcanic Zone: a review. J. Volcanol. Geotherm. Res., 68: 59–87.

Gregg, D.R., 1956. Eruption of Ngauruhoe 1954–1955. N.Z. J. Sci. Tech., B37: 675–688.

Grindley, G.W., 1960. Sheet 8 Taupo; Geological Map of New Zealand 1:250,000. Dept. Sci. Indus. Res., Wellington, New Zealand.

Grindley, G.W., 1965. The geology, structure and exploitation of the Wairakei geothermal field, Taupo, New Zealand. N.Z. Geol. Surv. Bull., 75.

Grindley, G.W. and Browne, P.R.L., 1968. Geology of the Broadlands geothermal field. N.Z. Dept. Sci. Indus. Res. Geotherm. Rep., 5: 5–43.

Grindley, G.W. and Hull, A.G., 1986. Historical Taupo earthquakes and earth deformation. R. Soc. N.Z. Bull., 24: 173–186.

Hackett, W.R., 1985. Geology and petrology of Ruapehu volcano and related vents. Ph.D. thesis, Victoria University, Wellington, New Zealand (unpubl.).

Hackett, W.R. and Houghton, B.F., 1986. Guide to Excursion C4: Active Composite Volcanoes of Taupo Volcanic Zone. N.Z. Geol. Surv. Rec., 11: 61–114.

Hackett, W.R. and Houghton, B.F., 1989. A facies model for a Quaternary andesite composite volcano: Ruapehu, New Zealand. Bull. Volcanol., 51: 51–68.

Healy, J., 1962. Structure and volcanism in the Taupo Volcanic Zone, New Zealand. In: G.A. Macdonald and H. Kuno (Editors), The Crust of the Pacific Basin. Am. Geophys. Union, Geophys. Monogr., 6: 151–157.

Healy, J., 1964. Volcanic mechanisms in the Taupo Volcanic Zone, New Zealand. N.Z. J. Geol. Geophys., 7: 6–23.

Healy, J., 1971. The tempo of rhyolitic volcanism in New Zealand. Acta 1st Int. Scientific Congress on the Volcano of Thera. Archaeological Services of Greece, Athens, pp. 64–72.

Healy, J., Schofield, J.C. and Thompson, B.N., 1964. Sheet 5 Rotorua; Geological Map of New Zealand 1:250,000. Dept. Sci. Indus. Res., Wellington, New Zealand.

Hildreth, W., 1981. Gradients in silicic magma chambers: implications for lithospheric magmatism. J. Geophys. Res., 86: 10153–10192.

Hildreth, W., Halliday, A.N. and Christiansen, R.L., 1991. Isotopic evidence concerning the genesis and contamination of basaltic and rhyolitic magma beneath the Yellowstone Plateau volcanic field. J. Petrol., 32: 63–138.

Hochstein, M.P. and Regenauer-Lieb, K., 1989. Heat transfer in the Taupo Volcanic Zone (NZ): role of volcanism and heating by plastic deformation. Proc. N.Z. Geotherm. Workshop, 11: 219–223.

Hochstein, M.P., Tearney, K., Rawson, S., Davey, F.J., Davidge, S., Henrys, S. and Backshall, D., 1986. Structure of the Hauraki Rift (New Zealand). Bull. R. Soc. N.Z., 24: 333–348.

Hochstein, M.P., Smith, I.E.M., Regenauer-Lieb, K. and Ehara, S., 1993. Geochemistry and heat transfer processes in Quaternary rhyolitic systems of the Taupo Volcanic Zone, New Zealand. Tectonophysics, 223: 213–235.

Hochstetter, F. von, 1864. Geologie von Neu Seeland. Beiträge zur Geologie der Provinzen Auckland und Nelson. (Novara Expedition, Geological Part 1(1)): 274 pp. (English translation by C.A. Fleming, Government Printer, Wellington, 1959).

Houghton, B.F. and Cuthbertson, A.S., 1989. Sheet T14BD Kaimai; Geological Map of New Zealand 1:50,000. Dept. Sci. Indus. Res., Wellington, New Zealand.

Houghton, B.F. and Hackett, W.R., 1984. Strombolian and phreatomagmatic deposits of Ohakune Craters, Ruapehu, New Zealand: a complex interaction between external water and rising basaltic magma. J. Volcanol. Geotherm. Res., 21: 207–231.

Houghton, B.F. and Nairn, I.A., 1991. The 1976–1982 strombolian and phreatomagmatic eruptions of White Island, New Zealand: eruptive and depositional mechanisms at a 'wet' volcano. Bull. Volcanol., 54: 25–49.

Houghton, B.F. and Wilson, C.J.N., 1989. A vesicularity index for pyroclastic deposits. Bull. Volcanol., 51: 451–462.

Houghton, B.F., Wilson, C.J.N., Lloyd, E.F., Gamble, J.A. and Kokelaar, B.P., 1987. A catalogue of basaltic deposits within the central Taupo Volcanic Zone. N.Z. Geol. Surv. Rec., 18: 95–101.

Houghton, B.F., Lloyd, E.F., Wilson, C.J.N. and Lanphere, M.A., 1991. K-Ar ages from the Western Dome Belt and associated rhyolitic lavas in the Maroa–Taupo area, Taupo Volcanic Zone, New Zealand. N.Z. J. Geol. Geophys., 34: 99–101.

Houghton, B.F., Wilson, C.J.N., McWilliams, M., Lanphere, M.A., Weaver, S.D., Briggs, R.M. and Pringle, M.S., 1995. Chronology and dynamics of a large silicic magmatic system: central Taupo Volcanic Zone, New Zealand. Geology 23: 13–16.

Howorth, R., 1975. New formations of late Pleistocene tephras from the Okataina Volcanic Centre, New Zealand. N.Z. J. Geol. Geophys., 18: 683–712.

Iso, N., Okada, A., Ota, Y. and Yoshikawa, T., 1982. Fission-track ages of late Pleistocene tephra on the Bay of Plenty coast, North

Island, New Zealand. N.Z. J. Geol. Geophys., 25: 295–303.

Jackson, J.A. and White, N.J., 1989. Normal faulting in the upper continental crust: observations from regions of active extension. J. Struct. Geol., 11: 15–36.

Karhunen, R.A., 1993. The Pokai and Chimp Ignimbrites of the NW Taupo Volcanic Zone. Ph.D. thesis, Canterbury University, Christchurch, New Zealand (unpubl.).

Keall, J.M., 1988. Volcanology and ignimbrite stratigraphy along the Paeroa Fault, Taupo Volcanic Zone. M.Sc. thesis, Victoria University, Wellington, New Zealand (unpubl.).

Kohn, B.P., 1973. Some studies of New Zealand Quaternary pyroclastic rocks. Ph.D. thesis, Victoria University, Wellington, New Zealand (unpubl.).

Lamarche, G., 1992. Seismic reflection survey in the geothermal field of the Rotorua caldera, New Zealand. Geothermics, 21: 109–119.

Lamarche, G. and Froggatt, P.C., 1993. New eruptive vents for the Whakamaru Ignimbrite (Taupo Volcanic Zone) identified from magnetic fabric study. N.Z. J. Geol. Geophys., 36: 213–222.

Lewis, K.B. and Pantin, H.M., 1984. Intersection of a marginal basin with a continent: structure and sediments of the Bay of Plenty, New Zealand. In: B.P. Kokelaar and M.F. Howells (Editors), Marginal Basin Geology. Geol. Soc. London, Spec. Publ., 16: 121–135.

Manning, D.A., 1994. Late Pleistocene tephrostratigraphy of the eastern Bay of Plenty region. In: D.J. Lowe (Compiler), Programme and Abstracts, International Inter-INQUA Field Conference and Workshop on Tephrochronology, Loess, and Paleopedology. University of Waikato, Hamilton, New Zealand, p. 42.

McCulloch, M.T., Kyser, T.K., Woodhead, J. and Kinsley, L., 1994 Pb-Sr-Nd-O isotopic constraints on the origin of rhyolites from the Taupo Volcanic Zone of New Zealand: evidence for assimilation followed by fractionation from basalt. Contrib. Mineral. Petrol., 115: 303–312.

Murphy, R.P. and Seward, D., 1981. Stratigraphy, lithology, palaeomagnetism, and fission track ages of some ignimbrite formations in the Matahana Basin, New Zealand. N.Z. J. Geol. Geophys., 24: 325–331.

Nairn, I.A., 1981. Some studies of the geology, volcanic history, and geothermal resources of the Okataina volcanic centre, Taupo Volcanic Zone, New Zealand. Ph.D. thesis, Victoria University, Wellington, New Zealand (unpubl.).

Nairn, I.A., 1989. Sheet V16AC Tarawera; Geological Map of New Zealand 1:50,000. Dept. Sci. Indus. Res., Wellington, New Zealand.

Nairn, I.A. and Beanland, S., 1989. Geological setting of the 1987 Edgecumbe earthquake, New Zealand. N.Z. J. Geol. Geophys., 32: 1–13.

Nairn, I.A. and Self, S., 1978. Explosive eruptions and pyroclastic avalanches from Ngauruhoe in February 1975. J. Volcanol. Geotherm. Res., 3: 39–60.

Nairn, I.A., Wood, C.P. and Bailey, R.A., 1994. The Reporoa caldera, Taupo Volcanic Zone: source of the Kaingaroa Ignimbrites. Bull. Volcanol., 56: 529–537.

New Zealand Geological Survey, 1972. North Island (1st Edition); Geological Map of New Zealand 1:1,000,000. Dept. Sci. Indus. Res., Wellington, New Zealand.

Ninkovich, D., 1968. Pleistocene volcanic eruptions in New Zealand recorded in deep-sea sediments. Earth Planet. Sci. Lett., 4: 89–102.

Northey, D.J., 1982. Seismic studies of the structure beneath Lake Taupo. Ph.D. thesis, Victoria University, Wellington, New Zealand (unpubl.).

Pillans, B.J., Roberts, A.P., Wilson, G.S., Abbott, S.T. and Alloway, B.V., 1994. Magnetostratigraphic, lithostratigraphic and tephrostratigraphic constraints on Lower and Middle Pleistocene sea-level changes, Wanganui Basin, New Zealand. Earth Planet. Sci. Lett., 121: 81–98.

Pringle, M.S., McWilliams, M.O., Houghton, B.F., Lanphere, M.A. and Wilson, C.J.N., 1992. $^{40}Ar/^{39}Ar$ dating of Quaternary feldspar: examples from the Taupo Volcanic Zone, New Zealand. Geology, 20: 531–534.

Reid, F.W., 1983. Origin of the rhyolitic rocks of the Taupo Volcanic Zone, New Zealand. J. Volcanol. Geotherm. Res., 15: 315–338.

Reid, F.W. and Cole, J.W., 1983. Origin of dacites of Taupo Volcanic Zone, New Zealand. J. Volcanol. Geotherm. Res., 18: 191–214.

Robertson, D.J., 1983. Paleomagnetism and geochronology of volcanics in the northern North Island. Ph.D. thesis, University of Auckland, Auckland, New Zealand (unpubl.).

Robinson, R., Smith, E.G.C. and Latter, J.H., 1981. Seismic studies of the crust under the hydrothermal areas of the Taupo Volcanic Zone, New Zealand. J. Volcanol. Geotherm. Res., 9: 253–267.

Rogan, A.M., 1982. A geophysical study of the Taupo Volcanic Zone, New Zealand. J. Geophys. Res., 87: 4073–4088.

Self, S., 1983. Large scale phreatomagmatic silicic volcanism: a case study from New Zealand. J. Volcanol. Geotherm. Res., 17: 433–469.

Self, S. and Sparks, R.S.J., 1978. Characteristics of widespread pyroclastic deposits formed by the interaction of silicic magma and water. Bull. Volcanol., 41: 196–212.

Seward, D., 1974. Ages of New Zealand Pleistocene sub-stages by fission-track dating of glass shards from tephra horizons. Earth Planet. Sci. Lett., 24: 242–248.

Seward, D., 1975. Fission-track ages of some tephras from Cape Kidnappers, Hawke's Bay, New Zealand. N.Z. J. Geol. Geophys., 18: 507–510.

Seward, D., 1976. Tephrostratigraphy of the marine sediments in the Wanganui Basin, New Z Zealand. N.Z. J. Geol. Geophys., 19: 9–20.

Seward, D., Christoffel, D.A. and Lienert, B., 1986. Magnetic polarity stratigraphy of a Plio-Pleistocene marine sequence of North Island, New Zealand. Earth Planet. Sci. Lett., 80: 353–360.

Shane, P.A.R., 1991. Remobilised silicic tuffs in middle Pleistocene fluvial sediments, southern North Island, New Zealand. N.Z. J. Geol. Geophys., 34: 489–499.

Shane, P.A.R., 1994. A widespread, early Pleistocene tephra (Potaka tephra, 1 Ma) in New Zealand: character, distribution, and implications. N.Z. J. Geol. Geophys., 37: 25–35.

Shane, P.A.R. and Froggatt, P.C., 1991. Glass chemistry, paleomagnetism, and correlation of middle Pleistocene tuffs in southern North Island, New Zealand, and Western Pacific. N.Z. J. Geol. Geophys., 34: 203–211.

Sissons, B.A., 1979. The horizontal kinematics of the North Island of New Zealand. Ph.D. thesis, Victoria University of Wellington, New Zealand (unpubl.).

Skinner, D.N.B., 1986. Neogene volcanism of the Hauraki Volcanic Region. In: I.E.M. Smith (Editor), Late Cenozoic Volcanism in New Zealand. R. Soc. N.Z. Bull., 23: 21–47.

Smith, E.G.C., Stern, T. and Reyners, M., 1989. Subduction and back-arc activity at the Hikurangi convergent margin, New Zealand. Pure Appl. Geophys., 129: 203–231.

Soengkono, S., Hochstein, M.P., Smith, I.E.M. and Itaya, T., 1992. Geophysical evidence for widespread reversely magnetised pyroclastics in the western Taupo Volcanic Zone (New Zealand). N.Z. J. Geol. Geophys., 35: 47–55.

Stern, T.A., 1979. Regional and residual gravity fields, central North Island, New Zealand. N.Z. J. Geol. Geophys., 21: 479–485.

Stern, T.A., 1985. A back-arc basin formed within continental lithosphere: the Central Volcanic Region of New Zealand. Tectonophysics, 112: 385–409.

Stern, T.A., 1987. Asymmetric back-arc spreading, heat flux and structure associated with the Central Volcanic Region of New Zealand. Earth Planet. Sci. Lett., 85: 265–276.

Stern, T.A. and Davey, F.J., 1985. Crustal structure studies within the central North Island: the Central Volcanic Region. Rep. 207, Geophys. Div., Dept. Sci. Indus. Res., Wellington, New Zealand, 47 pp.

Stern, T.A. and Davey, F.J., 1987. A seismic investigation of the crust and upper mantle structure within the Central Volcanic Region of New Zealand. N.Z. J. Geol. Geophys., 30: 217–231.

Stipp, J.J., 1968. The geochronology and petrogenesis of the Cenozoic volcanics of the North Island, New Zealand. Ph.D. thesis, Australian National University, Canberra, Australia (unpubl.). (Ages quoted here are those modified from this thesis in an unpublished 1968 manuscript by Stipp held on IGNS files at Wairakei.)

Sutton, A.N., Blake, S. and Wilson, C.J.N., 1995. An outline geochemistry of rhyolite eruptives from Taupo volcanic centre, New Zealand. J. Volcanol. Geotherm. Res. 68: 153–175.

Thompson, B.N., 1964. Quaternary volcanism of the Central Volcanic Region. N.Z. J. Geol. Geophys., 7: 45–66.

Topping, W.W., 1973. Tephrostratigraphy and chronology of late Quaternary eruptives from the Tongariro volcanic centre, New Zealand. N.Z. J. Geol. Geophys., 16: 397–423.

Vucetich, C.G. and Howorth, R., 1976. Late Pleistocene tephrostratigraphy in the Taupo district, New Zealand. N.Z. J. Geol. Geophys., 19: 51–69.

Walcott, R.I., 1987. Geodetic strain and the deformational history of the North Island of New Zealand during the late Cainozoic. Philos. Trans. R. Soc. London, Ser. A, 321: 163–181.

Walker, G.P.L., 1979. A volcanic ash generated by explosions where ignimbrite entered the sea. Nature, 281: 642–646.

Walker, G.P.L., 1980. The Taupo pumice: product of the most powerful known (ultraplinian) eruption? J. Volcanol. Geotherm. Res., 8: 69–94.

Walker, G.P.L., 1981a. Characteristics of two phreatoplinian ashes and their water-flushed origin. J. Volcanol. Geotherm. Res., 9: 395–407.

Walker, G.P.L., 1981b. The Waimihia and Hatepe plinian deposits from the rhyolitic Taupo volcanic centre. N.Z. J. Geol. Geophys., 24: 305–324.

Walker, G.P.L., Self, S. and Wilson, L., 1984. Tarawera 1886, New Zealand —a basaltic plinian fissure eruption. J. Volcanol. Geotherm. Res., 21: 61–78.

Watkins, N.D. and Huang, T.C., 1977. Tephras in abyssal sediments east of the North Island, New Zealand: chronology, paleowind velocity, and paleoexplosivity. N.Z. J. Geol. Geophys., 20: 179–198.

Wilson, C.J.N., 1986. Reconnaissance stratigraphy and volcanology of ignimbrites from Mangakino volcano. In: I.E.M. Smith (Editor), Late Cenozoic Volcanism in New Zealand. R. Soc. N.Z. Bull., 23: 179–194.

Wilson, C.J.N., 1991. Ignimbrite morphology and the effects of erosion: a New Zealand case study. Bull. Volcanol., 53: 635–644.

Wilson, C.J.N., 1993. Stratigraphy, chronology, styles and dynamics of late Quaternary eruptions from Taupo volcano, New Zealand. Philos. Trans. R. Soc. London, Ser. A, 343: 205–306.

Wilson, C.J.N. and Walker, G.P.L., 1985. The Taupo eruption, New Zealand. I. General aspects. Philos. Trans. R. Soc. London, Ser. A, 314: 199–228.

Wilson, C.J.N., Rogan, A.M., Smith, I.E.M., Northey, D.J., Nairn, I.A. and Houghton, B.F., 1984. Caldera volcanoes of the Taupo Volcanic Zone, New Zealand. J. Geophys. Res., 89: 8463–8484.

Wilson, C.J.N., Houghton, B.F. and Lloyd, E.F., 1986. Volcanic history and evolution of the Maroa–Taupo area, central North Island. In: I.E.M. Smith (Editor), Late Cenozoic Volcanism in New Zealand. R. Soc. N.Z. Bull., 23: 194–223.

Wilson, C.J.N., Switsur, V.R. and Ward, A.P., 1988. A new ^{14}C age for the Oruanui (Wairakei) eruption, New Zealand. Geol. Mag., 125: 297–300.

Wilson, C.J.N., Houghton, B.F., Lanphere, M.A. and Weaver, S.D., 1992. A new radiometric age estimate for the Rotoehu Ash from Mayor Island, New Zealand. N.Z. J. Geol. Geophys., 35: 371–374.

Wood, C.P., 1983. Petrological logs of drillholes BR26 to BR40: Broadlands geothermal field. N.Z. Geol. Surv. Rept., 108: 1–49.

Wood, C.P., 1992. Geology of the Rotorua geothermal system. Geothermics, 21: 25–41.

Wright, I.C., 1990. Late Quaternary faulting of the offshore Whakatane Graben, Taupo Volcanic Zone, New Zealand. N.Z. J. Geol. Geophys., 33: 245–256.

Wright, I.C., 1992. Shallow structure and active tectonism of an offshore continental spreading system: the Taupo Volcanic Zone, New Zealand. Mar. Geol., 103: 287–309.

Wright, I.C., 1993a. Pre-spread rifting and heterogeneous volcanism in the southern Havre Trough back-arc basin. Mar. Geol., 113: 179–200.

Wright, I.C., 1993b. Southern Havre Trough–Bay of Plenty (New Zealand): structure and seismic stratigraphy of an active back-arc basin complex. In: P.F. Ballance (Editor), Sedimentary Basins of the World, 2: South Pacific Sedimentary Basins. Elsevier, Amsterdam, pp. 195–211.

Yoshikawa, T., 1993. Geomorphic development of axial ranges in eastern North Island, New Zealand, correlated with ignimbrite volcanism in the Taupo Volcanic Zone. Bull. Dept. Geogr., Univ. Tokyo, 25: 1–27.