**Abstract** Subduction-related volcanism in the northern part of the North Island of New Zealand shifted abruptly during the late Pliocene. This study focuses on the transition, in time and space, from the NNW-oriented Miocene–Pliocene Coromandel Volcanic Zone to the northeast-oriented active Taupo Volcanic Zone. The volcanic rocks marking this transition are exposed in the Tauranga Basin and adjacent Kaimai Range, and associated here with the recently defined Tauranga and Kaimai Volcanic Centres, respectively. New 40Ar/39Ar age determinations indicate that the transition occurred between 1.90 and 1.55 Ma, that is between the youngest age of silicic volcanism in the Tauranga-Kaimai area, and the age of the oldest silicic volcanism in the Taupo Volcanic Zone. This interpretation is generally consistent with recent plate models and with the initiation of the Kermadec Arc within the last 2 m.y.

**Keywords** 40Ar/39Ar; silicic volcanism; Tauranga Volcanic Centre; Kaimai Volcanic Centre; Pliocene; Pleistocene; Aongatete Ignimbrites; Waitariki Ignimbrite; Papamoa ignimbrites; Hauraki Fault

**INTRODUCTION**

Volcanic rocks exposed in the Tauranga area and Kaimai Range represent an important geographical and chronological transition from the Miocene–Pliocene Coromandel Volcanic Zone (CVZ) to the Pleistocene–Recent Taupo Volcanic Zone (TVZ) (Fig. 1). The Tauranga-Kaimai area contains at least 21 dacite-rhyolite domes or dome complexes together with 3 defined ignimbrite formations that are inferred to predate the onset of TVZ volcanism. The topographically higher domes and flows of the Tauranga-Kaimai area are in part covered and encircled by the younger, voluminous, and partially welded ignimbrites derived from the Rotorua Volcanic Centre in the TVZ to the south.

Previously available radiometric ages (see discussion below) from rocks of the Tauranga-Kaimai area were inadequate to constrain the transition between the two volcanic zones. Here, we present new 40Ar/39Ar age data on the volcanic rocks of the Tauranga-Kaimai area, in an attempt to date the transition between volcanism in the Coromandel Arc and that in the TVZ. We use these new data to constrain the timing of major displacement on the Hauraki Fault, which defines the eastern margin of the Hauraki Rift, and discuss some preliminary geochemical, mineralogical, and spatial relations of rhyolitic and dacitic volcanics within what are defined in Kear (1994) and here as the Tauranga and Kaimai Volcanic Centres.

**VOLCANIC GEOLOGY AND PREVIOUS CHRONOLOGY OF THE TAURANGA-KAIMAI AREA**

The Tauranga-Kaimai area includes a Pliocene subaerial volcanic sequence that forms the southern Kaimai Ranges, a series of late Pliocene rhyolite domes that form characteristic landforms, extensive tilted and gently dipping plateaus of late Pliocene–Quaternary ignimbrites, and Quaternary volcaniclastic sediments that largely infill the Tauranga Basin (Fig. 1) (Briggs et al. 1996). The central part of the modern Tauranga Basin is occupied by Tauranga Harbour, a large shallow estuary with a northern and southern entrance confined by tombolos and the barrier island complex of Matakania Island. The whole region is generally blanketed by a thick cover of late Pleistocene–Recent silicic tephras derived from the TVZ and Mayor Island Volcanic Centre. The sequence is summarised below in stratigraphic order (see Briggs et al. 1996).

**Late Miocene–Pliocene andesite–dacite cone volcanism**

The oldest rocks in the Tauranga-Kaimai area are andesitic to dacitic lavas, dikes, and volcaniclastic strata of the Kaimai Subgroup (defined by Houghton & Cuthbertson 1989), which are exposed in the uplifted Kaimai Range. Brathwaite & Christie (1996) presented 18 K-Ar ages (whole
rock, biotite, hornblende) from Kaimai Subgroup andesites
and dacites in the Waihi district north of the Tauranga area,
which range in age from 4.01 ± 0.09 to 5.56 ± 0.35 Ma.
The Kaimai Subgroup rocks are overlain by the Aongatete
Ignimbrites that crop out on the eastern side of the
Kaimai Range. The Aongatete Ignimbrites consist of
>290 m sequence of non-welded to densely welded
lenticular dacitic ignimbrites and tuffs, with evidence
(paleosols, intercalated sediments) for their having been
erupted over a prolonged period of time (Houghton &
Cuthbertson 1989).

Motiti Island is situated 12 km offshore from the Bay
of Plenty coast northeast of Tauranga (Fig. 2), and is a flat-
lying eroded remnant of a Pliocene andesitic composite
cone. It has been K-Ar (whole rock) dated by Itaya (in Henry
1991) at 4.32 ± 0.68 (Motiti Formation) and 3.42 ± 0.19 Ma
(Orongatea Formation), and has a similar composition to Kaimai Subgroup andesites.

The Otawa Volcanics are late Pliocene basaltic andesite, andesite, and dacite lavas and volcanic breccias (K-Ar ages from andesites of 2.95 ± 0.30 and 2.54 ± 0.05 Ma; Stipp 1968) that crop out in the Papamoa Range on the eastern side of the Tauranga Basin (Briggs et al. 1996). They form a series of eroded composite volcanoes that are locally hydrothermally altered and mineralised.

A solitary outcrop of basalt lavas with pseudo-pillow structures occurs on a small island 30 m offshore on the harbour side of Matakana Island (Hollis 1995), and is the only occurrence of basalt in the Tauranga area. The Matakana Basalt has been K-Ar dated at 2.7 ± 0.1 Ma (Hollis 1995). It is normally magnetised (Briggs et al. 1996), and has been correlated with the 2.58–3.58 Ma Gauss Chron (cf. Berggren et al. 1995).

Pre-TVZ silicic dome complexes and ignimbrite

Minden Rhyolite domes are the most prominent landforms in the Tauranga area, typified by Mount Maunganui, a steep-sided flat-topped 252 m dome. Minden Rhyolite (Skinner 1986) includes all the rhyolites in the Tauranga and Bay of Plenty region northwest of the TVZ (Houghton & Cuthbertson 1989) that are mainly late Miocene–Pliocene in age (Adams et al. 1994).

The Kaimai dome is the largest lava dome and flow complex, situated in the southern Kaimai Range (Fig. 1). It is cut by the Hauraki Fault, and is onlapped by the Waiteariki Ignimbrite to the north and by the Waimakariri and Mamaku Ignimbrites to the south and east. The Kaimai and Kakahu domes are spatially separated to the southwest from those in the Tauranga area, and define the location of the Kaimai Volcanic Centre (Kear 1994), which represents
a major rhyolitic volcanic centre in the southern Kaimai Range. Circular or arcuate caldera structures or negative gravity lows that help to define younger calderas in the TVZ (Rogan 1982; Wilson et al. 1984; Bibby et al. 1995) and late Miocene–Pliocene calderas in the CVZ (Malengreau et al. 2000) have not been observed in the southern Kaimai Range, but nevertheless the surface exposures represent voluminous outpouring of rhyolitic lavas.

The Minden Rhyolite domes and flows of the Tauranga area collectively define the Tauranga Volcanic Centre and can be subdivided into four groups based on spatial association, phenocryst mineral assemblage, and chemistry, as discussed later (see Discussion). Gravity maps of the Tauranga area (Woodward & Ferry 1973) show a large negative residual Bouguer anomaly east of Katikati in northern Tauranga Harbour and may thus indicate a caldera structure filled with low density volcanic material, although there is no direct evidence for caldera collapse in this area. The gravity low does not coincide with the distribution of the rhyolite and dacite domes and flows that lie well to the southeast.

Rutherford (1978) measured fission-track ages on obsidian from Bowentown (2.29 ± 0.21 Ma) and Mount Maunganui (4.34 ± 0.38 Ma), but these ages have been considered less reliable due to the partial fading and annealing of fission tracks in volcanic glass, which can result in an age that is too young (Westgate 1989; Alloway et al. 1993). Takagi (1995) dated the Bowentown rhyolite dome by the K-Ar method, obtaining ages of 2.51 ± 0.25 Ma from a whole-rock sample and 2.77 ± 0.31 Ma from plagioclase. These ages have large relative errors but are comparable with a K-Ar age of 2.89 ± 0.07 Ma for the same dome by Brathwaite & Christie (1996). Takagi (1995) also determined K-Ar ages from the Mangatawa rhyolite dome (2.36 ± 0.08 Ma, whole rock; 2.28 ± 0.15 Ma, plagioclase), and Minden Peak dome at Te Puna Quarry (1.52 ± 0.23 Ma, whole rock).

Ignimbrite sheets are prominent landform features in the southern Kaimai and Tauranga areas, forming a series of extensive plates. The Whakamarama Plateau is underlain by the Waiteariki Ignimbrite that has been tilted 3–5º eastwards by uplift along the Hauraki Fault (Fig. 1). The Waiteariki Ignimbrite is a widespread, large-volume welded ignimbrite, locally >220 m thick, that extends northwards to underlie the Tauranga Harbour. It is encountered in drillholes throughout the Tauranga urban and harbour areas (Harmsworth 1983). The Waiteariki Ignimbrite has previously been interpreted as originating from a source to the southeast and the older Whakamarama area, forming a large-uplift volcanic margin environment.

The Ongatiti Ignimbrite is a major caldera-related welded ignimbrite (Wilson 1986) erupted from the Mangakino Volcanic Centre and previously dated by \(^{40}\text{Ar}/^{39}\text{Ar}\) at 1.21 ± 0.04 Ma (Houghton et al. 1995). Non-welded portions are known to extend to both the west (Lowe et al. 2001) and east (Shane et al. 1996) coasts of the North Island, and two small outcrops in the Tauranga Basin west of the Wairoa River record the northernmost extent of the welded portion of the unit.

The Te Puna and Te Ranga Ignimbrites are localised to the central Tauranga Basin, and may represent either locally derived eruptive units restricted to the Tauranga Basin, or, more likely, distal equivalents of widespread and as yet unidentified ignimbrites from TVZ (Whitbread-Edwards 1994; Hollis 1995). However, until this study, their ages were only known to be post-Ongatiti. The Te Puna Ignimbrite is a hornblende-bearing, pumice-rich, crystal-rich, non-welded to partially welded ignimbrite exposed in several outcrops around Tauranga Harbour. At Omokoroa the Te Puna Ignimbrite overlies lignites and fluvial pumiceous sands, suggesting that it flowed into an estuarine or swampy lake margin environment.

The Te Ranga Ignimbrite is a non-welded, crystal-poor, variably pumice-poor to pumice-rich, sandy textured ignimbrite, notable for containing coarse glass shard textures, charcoal, and obsidian, and is restricted to low-lying areas of the central Tauranga Basin. In several sections it is intercalated with underlying lauristrine silts and overlying fluvial volcanioclastic sands, and may have been emplaced subaqueously.

The Waimakariri Ignimbrite overlies the Te Ranga Ignimbrite and is a voluminous partly welded ignimbrite derived from a source to the south, possibly the Rotorua Volcanic Centre. Its age is not known but is bracketed by age determinations on older and younger units to be between 0.32 and 0.22 Ma (Houghton et al. 1995). The overlying Mamaku Ignimbrite is the youngest major, landscape-forming volcanic unit in the Kaimai–Tauranga area. Dated by \(^{40}\text{Ar}/^{39}\text{Ar}\) at 0.22 ± 0.01 Ma (Houghton et al. 1995), its eruption is inferred to have been accompanied by the collapse of Rotorua caldera. The Waimakariri and Mamaku Ignimbrites together form the Mamaku Plateau that dips gently at 1–2º northwards towards Tauranga Basin and the Bay of Plenty coastline, and surrounds the upstanding rhyolite lava domes of Kaimai, Puwhenua, Otawainaku, and Mt Misery (Fig. 1). The Mamaku Plateau to the southeast and the older Whakamarama Plateau to the northwest are separated by the Wairoa River valley.

Intercaleted with all of the volcanic units that postdate the Waiteariki Ignimbrite are a variety of terrestrial and estuarine sedimentary deposits called the Matua Subgroup (Houghton & Cuthbertson 1989). These were derived from erosion, transportation, and redeposition of the consolidated and unconsolidated volcanic rocks and associated tephras. The Matua Subgroup sediments form terraces ranging to
80 m in height in the Tauranga Basin that are exposed in coastal cliffs around Tauranga Harbour. They consist of fluvial pumiceous silts, sands, and gravels, diatomaceous lacustrine and estuarine muds, lignites, and peats, intercalated with thin distal ignimbrites and fall deposits (Pahoia Tephra). A wide variety of sedimentary structures occurs including cross-bedding, planar stratified and massive units, and water escape post-depositional slumping and liquefaction structures.

The Tauranga area is generally blanketed by a thick cover of tephra, including the older deeply weathered Pahoia Tephra and Hamilton Ash (Briggs et al. 1996). These are overlain by the Rotoehu Ash and a sequence of younger tephra derived from the TVZ. Tauranga Harbour is a large barrier-enclosed estuary, and the tombolos of Bowentown and the Upuhue rhyolite tephra deposits (Briggs et al. 1996). These are overlain by the Rotoehu Ash and a sequence of younger tephra derived from the TVZ. Tauranga Harbour is a large barrier-enclosed estuary, and the tombolos of Bowentown and Mount Maunganui have been tied to the mainland by a system of progradational dune ridges formed during the Holocene.  

NEW 40Ar/39Ar AGES FOR TAURANGA RHYOLITIC VOLCANISM

Analytical methods

Our new 40Ar/39Ar ages from feldspar separates presented here in Table 1 were determined at Stanford using mineral separates prepared by crushing, panning, multiple magnetic separation, and hand picking at the University of Waikato. Mineral separates were packed in ultrapure Cu foil and irradiated at the Oregon State University TRIGA reactor using USGS sanidine standard 85G003 (Taylor Creek Rhyolite sanidine) as a flux monitor with an assumed age of 27.92 Ma. Irradiation parameter J values were corrected for neutron flux gradients in the reactor. Noble gases were extracted in 8 min heating cycles in a Staudacher-type double-vacuum resistance furnace with a Ta crucible and Mo liner, and purified by reaction with SAES Zr-Al getters for an additional 5 min. Isotopic measurements were made using a Mass Analyser Products 216 noble gas mass spectrometer. Isotopic abundances were calculated by linear peak extrapolation to inlet time during 6–10 serial scans of 40Ar, 39Ar, 38Ar, 37Ar, and 36Ar. The mass spectrometer data were corrected for extraction line blanks, decay of 39Ar and 37Ar since irradiation, instrumental isotopic mass fractionation, and interfering Ca and K-derived Ar isotopes produced on irradiation. Typical system blanks for 40Ar at 1200°C were in the order of 1 to 3 × 10–15 moles. The uncertainties reported include uncertainties in peak heights, blank values, neutron flux, decay rates of 37Ar, 39Ar and 40K, and flux monitor age. Isochron ages are used in all discussions, except where otherwise stated.

Results

Two ages obtained from the base and top of the Aongatete Ignimbrites (3.94 ± 0.02 and 3.58 ± 0.02 Ma, respectively) confirm earlier geological interpretations that suggested a prolonged period for this eruption sequence (Houghton & Cuthbertson 1989). The Aongatete Ignimbrites are thus distinctly older than previously suggested (mid Pleistocene; see Houghton & Cuthbertson 1989 for a summary of previous age estimates) and are not considered to be related to the Kaimai and Tauranga Volcanic Centres. The 40Ar/39Ar ages from 10 rhyolite and dacite lava domes and flows of the Tauranga Volcanic Centre range from 2.69 ± 0.04 to 1.95 ± 0.02 Ma (Table 1). There is no obvious distinction in age between the dacites (Upuhue and Kopukairua) and the rhyolites. The Papamoa Ignimbrites are

<table>
<thead>
<tr>
<th>Lab no.</th>
<th>Location</th>
<th>Rock type</th>
<th>SiO2 wt%</th>
<th>Grid reference</th>
<th>Plateau age (Ma)</th>
<th>Isochron age (Ma)</th>
<th>40Ar/39Ar</th>
</tr>
</thead>
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<tr>
<td>411</td>
<td>Minden Peak</td>
<td>rhyolite</td>
<td>74.6–76.0</td>
<td>U14/788840</td>
<td>2.16 ± 0.02</td>
<td>2.16 ± 0.03</td>
<td>295 ± 4</td>
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<td>405</td>
<td>Kaimaikaroro</td>
<td>rhyolite</td>
<td>70.9</td>
<td>U14/739741</td>
<td>–</td>
<td>2.39 ± 0.06</td>
<td>401 ± 16</td>
</tr>
<tr>
<td>408</td>
<td>Mount Maunganui</td>
<td>rhyolite</td>
<td>76.7–77.3</td>
<td>U14/902919</td>
<td>2.35 ± 0.02</td>
<td>2.35 ± 0.06</td>
<td>302 ± 52</td>
</tr>
<tr>
<td>407</td>
<td>Mangatawa</td>
<td>rhyolite</td>
<td>72.7</td>
<td>U14/962842</td>
<td>2.46 ± 0.03</td>
<td>2.39 ± 0.13</td>
<td>319 ± 49</td>
</tr>
<tr>
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<td>Upuhue</td>
<td>dacite</td>
<td>68.5–69.9</td>
<td>U14/977828</td>
<td>2.63 ± 0.02</td>
<td>2.69 ± 0.04</td>
<td>260 ± 18</td>
</tr>
<tr>
<td>409</td>
<td>Papamoa dome</td>
<td>rhyolite</td>
<td>70.0</td>
<td>U14/901805</td>
<td>2.52 ± 0.02</td>
<td>2.50 ± 0.03</td>
<td>306 ± 6</td>
</tr>
<tr>
<td>433</td>
<td>Kopukairua</td>
<td>dacite</td>
<td>66.8–68.0</td>
<td>U14/955701</td>
<td>2.20</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>416</td>
<td>Otawaimaku</td>
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<td>72.6</td>
<td>U15/917610</td>
<td>1.95 ± 0.01</td>
<td>1.95 ± 0.02</td>
<td>296 ± 7</td>
</tr>
<tr>
<td>420</td>
<td>Puwhenua</td>
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<td>–</td>
<td>U15/843593</td>
<td>2.06 ± 0.02</td>
<td>2.14 ± 0.04</td>
<td>257 ± 18</td>
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<tr>
<td>431</td>
<td>Mount Misery</td>
<td>rhyolite</td>
<td>72.2</td>
<td>U14/908724</td>
<td>2.67 ± 0.02</td>
<td>2.69 ± 0.03</td>
<td>277 ± 20</td>
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<tr>
<td>439</td>
<td>Lower Papamoa Ignimbrite</td>
<td>dacite/rhyolite</td>
<td>63.8–70.8</td>
<td>U14/971810</td>
<td>2.40 ± 0.02</td>
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<td>295 ± 2</td>
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<tr>
<td>432</td>
<td>Upper Papamoa Ignimbrite</td>
<td>dacite/rhyolite</td>
<td>64.2–70.0</td>
<td>U14/978780</td>
<td>1.87 ± 0.01</td>
<td>1.90 ± 0.10</td>
<td>290 ± 20</td>
</tr>
</tbody>
</table>

Kaimai Volcanic Centre

435 Kaimai dome rhyolite – T15/680659 2.89 ± 0.02 2.86 ± 0.08 323 ± 29
415 Kakahu rhyolite – T15/678550 2.87 ± 0.01 2.87 ± 0.02 297 ± 5
507 Waiteariki Ignimbrite dacite/rhyolite 67.5–71.6 U14/783729 2.09 ± 0.03 – –

Other ignimbrites

421 Aongatete Ignimbrites dacite – T14/658877 3.58 ± 0.02 3.55 ± 0.05 307 ± 17
428 Aongatete Ignimbrites dacite – T14/658877 3.94 ± 0.02 3.93 ± 0.06 303 ± 33
402 Ongatiti Ignimbrite rhyolite 70–75 U14/783729 1.32 ± 0.01 1.34 ± 0.02 291 ± 5
403 Te Puna Ignimbrite rhyolite 72.6–75.7 U14/828855 0.929 ± 0.012 – –
438 Te Ranga Ignimbrite dacite/rhyolite 68.6–76.6 U14/845772 0.274 ± 0.016 – –
most likely associated with the effusion of one of these lavas, except the Upper Papamoa Ignimbrite has a slightly younger (but overlapping within 2 σ error) age of 1.90 ± 0.10 Ma. Thus, silicic activity including all lavas and ignimbrites at the Tauranga Volcanic Centre ranges in age from 2.69 to 1.90 Ma. The K-Ar age for the Matakana Basalt (2.7 ± 0.1 Ma) lies within 1 σ error of the age span of earliest silicic activity of the Tauranga Volcanic Centre. Its high-Al, calc-alkaline characteristics and similarity in composition to TVZ basalts (Hollis 1995) suggest it may have played a parallel role in the generation of silicic magmatism to basalt in the TVZ (e.g., Graham et al. 1995).

Age determinations from eruptives of the Kaimai Volcanic Centre range from 2.87 ± 0.02 to 2.09 ± 0.03 Ma. These include the rhyolite domes of Kaimai and Kakahu, and the Waiteariki Ignimbrite. This age range overlaps with Tauranga Volcanic Centre activity but is slightly older than it, and has a similar longevity of volcanic activity (0.78 Ma) to the Kapenga Volcanic Centre in TVZ (Wilson et al. 1995).

The 40Ar/39Ar ages obtained here (1.32 ± 0.01 Ma, plateau age; 1.34 ± 0.02 Ma; isochron age) for the Ongati Ignimbrite differ from the previously published plateau age of 1.21 ± 0.04 Ma (Houghton et al. 1995), but both ages (our isochron age and the previously published plateau age) overlap within 2σ error (using the method of Ward & Wilson 1978). Also, both suites of ages overlap within 1 σ error of the K-Ar age of 1.25 ± 0.09 Ma determined on hornblende separates by Soengkono et al. (1992).

The Te Puna Ignimbrite yielded a much younger plateau age (0.929 ± 0.012 Ma) than the rest of the Tauranga Volcanic Centre eruptives, which strongly suggests that it is likely to be a distal TVZ ignimbrite erupted during Period IIA (1.21–0.89 Ma) of Houghton et al. (1995). The Te Puna Ignimbrite is reversely magnetised and presumably was erupted during the early Matuyama Chron (Turner & Kamp pers. comm.). However, it cannot be correlated with the Rocky Hill Ignimbrite, which has similar mineralogical and geochemical affinities, because the Rocky Hill Ignimbrite has a slightly older age (1.00 ± 0.05 Ma determined by Houghton et al. 1995) and is normally magnetised and belongs to the Jaramillo Subchron (Soengkono et al. 1992; Tanaka et al. 1996).

Similarly, the Te Ranga Ignimbrite (0.274 ± 0.016 Ma) may be a distal correlative of an ignimbrite erupted during Period IIIB (0.28–c. 0.15 Ma) from the TVZ. The Te Ranga Ignimbrite has mineralogical similarities and stratigraphic relationships to the Waitho Ignimbrite (Fransen & Briggs 1981; Fransen 1982) exposed on the western side of the Mamaku Plateau, and the two ignimbrites may be equivalent. The age of the Waitho Ignimbrite is not determined, but it is bracketed by ignimbrites dated at 0.32 ± 0.02 and 0.22 ± 0.01 Ma (Houghton et al. 1995).

DISCUSSION AND GEOLOGICAL IMPLICATIONS

Dating the transition from Coromandel to Taupo volcanism

The Tauranga and southern Kaimai region occupies an important position in understanding the temporal and spatial transitions from CVZ to TVZ (Fig. 2). Andesitic volcanism in the CVZ commenced at c. 18 Ma (Adams et al. 1994) with the production of the Kuwotu Volcanic Complex in the north (Hayward et al. 2001), and according to the age data determined by Adams et al. (1994), volcanism evolved more or less continuously to 4 Ma in southern CVZ. Between c. 18 and 9 Ma, an andesitic arc was active, and the first rhyolitic eruptions occurred at c. 10 Ma (Adams et al. 1994). However, these onshore records of late Cenozoic volcanism in CVZ are incomplete because of poor exposures, hydrothermal alteration and/or deep weathering, subaerial and marine erosion, and burial by younger eruptions. A much more complete record of inferred CVZ volcanism has recently been established from cores collected from the Ocean Drilling Program (ODP) Leg 181, located offshore to the east and downwind from CVZ and TVZ (Carter R. M. et al. 1999; Carter L. et al. 2003). These cores provide a record of major rhyolitic eruptions from CVZ and TVZ, identified from 197 macroscopic (>10 mm thick) tephras, and are interpreted to indicate that rhyolitic volcanism began c. 12 Ma, that is, somewhat earlier than previously assumed (Carter et al. 2003). Furthermore, the frequency and thickness of tephas shows that volcanism was more or less continuous on a >10^5 yr basis, and that the tempo and intensity increased throughout the late Miocene and Pliocene and then into the Quaternary when the TVZ formed (Carter et al. 2003). Carter et al. (2003) also proposed that the transition from CVZ to TVZ, previously placed at c. 4–2 Ma (Adams et al. 1994; Houghton et al. 1995; Wilson et al. 1995), was apparently seamless, without obvious breaks or changes in tephra composition.

Our new data contrast with two elements of Carter et al.’s (2003) interpretations. (1) Our data show that silicic volcanism in the southern CVZ continued after 4 Ma. The bracketing ages for the Aongatete Ignimbrites (3.93–3.55 Ma; Table 1) suggest continuing mid-Pliocene silicic volcanism, though the source(s) for these units has not been located. In addition, coeval silicic explosive volcanism was occurring farther north in the Waihi Basin with the eruption of the Ohinemuri Subgroup ignimbrites. Hoskin et al. (1998) determined a U-Pb age from zircon of 3.69 ± 0.06 Ma for the Owahoroa Ignimbrite, and the other stratigraphically associated ignimbrites of the Waihi Basin (i.e., Corbett Ignimbrite and Waikino Ignimbrite), probably have similar ages. (Here, we omit the much younger FT/glass ages of 2.89 ± 0.38 Ma for the Owahoroa Ignimbrite and 1.5 ± 0.23 Ma for the Waikino Ignimbrite (Kohn 1973) because of the partial fading and annealing of fission tracks in volcanic glass.)

(2) There is a gap of 350 000 yr in the onshore record in the age range for silicic volcanism between 1.90 Ma at Tauranga (the youngest age of rhyolitic volcanism in southern CVZ) and 1.55 Ma in the TVZ (the earliest eruptive, Ignimbrite A, from the Mangakino Volcanic Centre; Houghton et al. 1995). (Note that while the 1.68 ± 0.07 Ma age determination for Ignimbrite C at Mangakino suggests it predates Ignimbrite A, field stratigraphic relationships show that the former overlies the latter, and hence must be younger). For large-scale ignimbrite eruptions, the gap is 540 000 yr between the Waitemaorongi Ignimbrite (2.09 ± 0.03 Ma plateau age; Kaimai) and Ignimbrite A (1.55 ± 0.05 Ma; Mangakino). Existing mapping does not resolve whether these gaps are real, but the stratigraphic record of tephas from the offshore marine record of Carter et al. (2003, fig. 3) also shows a lack of macroscopic tephra between c. 1.97 and 1.67 Ma. The onshore gap between major ignimbrites is comparable in length to the longest gap inferred in TVZ history, that between the Waiohau (0.71 Ma) and Whakamaru-group ignimbrites (0.34–0.32 Ma; Houghton et al. 1995). However, in the case of Tauranga-Kaimai versus Mangakino, there was a substantial shift in vent positions during this gap in volcanism, although the geographical
separation from Tauranga to Mangakino is similar to that between Okataina and Taupo centres in TVZ.

Implications for plate tectonic reconstructions

Initial arc volcanism (c. 18 Ma) in CVZ was oriented NNW (Skinner 1986; Herzer 1995), in response to southwest-dipping Hikurangi subduction (King 2000). The north-trending Colville Arc was also active at this time, but progressively migrated eastwards in conjunction with the northern segment of the subduction zone, and concurrently with opening of the South Fiji Basin (King 2000). By 10 Ma, spreading in the South Fiji Basin had probably ceased (King 2000), and volcanism continued in the Coromandel and Colville Arcs. Extension may have begun in central North Island c. 4–5 Ma (Stern 1987; Stern & Davey 1989) as volcanism ceased on the Colville Arc and the Havre Trough back-arc basin started to open, and further clockwise rotation of the Hikurangi margin resulted in the inception of the presently active Kermadec Arc at c. 2 Ma (Ballance et al. 1999). There has been a general eastward migration, younging, and widening of the Havre Trough since c. 4 Ma from the line of the Colville–Coromandel Volcanic Zone to the present-day Kermadec-Taupo Volcanic Zone (Wright 1993), and our new data indicate that at the very southern segment of the CVZ, the transition from CVZ to TVZ occurred between 1.90 and 1.55 Ma.

The time of transition from CVZ to TVZ appears to mark a significant period in the volcano-tectonic history of the New Zealand area, which includes the following: commencement of volcanism of the Kermadec Arc, pronounced and continued movement of the North Island Shear Belt in eastern North Island (King 2000), development of the Hauraki Rift (see below) and extensional block faulting in western North Island, and a pronounced increase in the frequency and volume of silicic volcanic activity in the TVZ. CVZ is characterised by lower magma output rates, smaller caldera sizes (c. 8 km diam., compared with c. 20 km diam. for TVZ), smaller eruptive volumes (where the modern degrees of exposure allow: <30 km³ for CVZ, 30 to >300 km³ for TVZ), and greater proportions of mafic (basalt to dacite) versus silicic (rhyodacite to rhyolite) magmas (40% mafic:60% silicic for CVZ; 5% mafic:95% silicic for TVZ). We relate these contrasts to increased rates of subduction and crustal extension in the TVZ that have led to a thinner crust and higher rates of silicic magma generation.

Age of the Hauraki Fault

The Hauraki Rift (Fig. 2) is an active continental rift that extends over 300 km from north of Little Barrier Island to TVZ in the south, where it is buried under late Pleistocene ignimbrites (Hochstein & Nixon 1979; Hochstein et al. 1986). The Hauraki Fault defines the eastern margin of the rift and has a maximum throw of c. 2.5–4 km in the central segment (Hochstein et al. 1986). The Kerepehi Fault forms a median fault in the rift, and both the Hauraki and Kerepehi Faults are seismically active (Houghton & Cuthbertson 1989; de Lange & Lowe 1990; Chick 1999). To the south, the Hauraki Rift and the Hauraki Fault are partially covered by the Mamaku Ignimbrite (0.22 Ma), which shows no evidence of displacement (Houghton & Cuthbertson 1989). The age of initiation of the Hauraki Rift is uncertain but is considered to be late Miocene or c. 7 Ma (Hochstein & Ballance 1993).

In the southern segment of the Hauraki Rift, the rift floor has subsided c. 1 km (Hochstein et al. 1986), indicating that the Hauraki Fault has a throw of c. 1 km in this region of the Kaimai Ranges. Exposures in the fault scarp of the Hauraki Fault show that the Waiteariki Ignimbrite is displaced by at least 400 m (Houghton & Cuthbertson 1989). However, the Ongatiti Ignimbrite has flowed from the south from the Mangakino Volcanic Centre into an already formed depression through the Hinuera Gap, at least as far as Morrinsville (Bowling 1989). We thus consider that a large proportion (>400 m) of the throw of the Hauraki Fault and the most rapid period of rift development, at least in the southern segment, occurred between the ages of the Waiteariki and Ongatiti

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Table 2  Rhyolite and dacite lava dome and ignimbrite groups in the Tauranga Volcanic Centre based on spatial relations and mineralogical and geochemical characteristics (data from Hughes 1993, Hall 1994, Whitbread-Edwards 1994).

<table>
<thead>
<tr>
<th>Group/age</th>
<th>Lava domes/ignimbrites</th>
<th>Phenocryst mineralogy (+ quartz + plagioclase)</th>
<th>Geochemistry</th>
</tr>
</thead>
<tbody>
<tr>
<td>Minden Peak</td>
<td>Minden Peak Manawata Kaikaiaroro</td>
<td>Crystal-rich hornblende rhyolites ± biotite ± orthopyroxene</td>
<td>High SiO₂ (70.9–76.2 wt%) High K₂O (2.8–3.8 wt%) Intermediate Zr (132–193 ppm)</td>
</tr>
<tr>
<td>2.16–2.39 Ma</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mount Maunganui</td>
<td>Mount Maunganui Mount Drury Moturiki Island Motuotau Island</td>
<td>Crystal-poor biotite rhyolites ± hornblende ± orthopyroxene</td>
<td>Very high SiO₂ (76.7–77.3 wt%) Very high K₂O (3.8–4.2 wt%) Low Zr (94–110 ppm)</td>
</tr>
<tr>
<td>2.35 Ma</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mangatawa</td>
<td>Mangatawa Upuhue Papamoa Kopukairua Waitao Waikite Otawhainuku Puwhenua Lower Papamoa Ignimbrite Upper Papamoa Ignimbrite</td>
<td>Crystal-rich hornblende – orthopyroxene rhyolites and dacites ± biotite</td>
<td>Low SiO₂ (64.0–73.0 wt%) High K₂O (2.6–3.1 wt%) Intermediate Zr (102–187 ppm)</td>
</tr>
<tr>
<td>1.90–2.69 Ma</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mount Misery</td>
<td>Mount Misery Greenpark Pukunui</td>
<td>Crystal-rich hornblende – orthopyroxene rhyolites</td>
<td>Low SiO₂ (72 wt%) High K₂O (2.8–3.1 wt%) High Zr (207–285 ppm)</td>
</tr>
<tr>
<td>2.69 Ma</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Ignimbrites, that is, between 2.09 ± 0.03 and 1.34–1.21 Ma. This age span coincides with the eruption of the younger dacite lavas of the Haowhenua Formation on Little Barrier Island in the northern segment of the Hauraki Rift at 1.6–1.2 Ma (Lindsay et al. 1999).

Spatial, mineralogical, and geochemical relations of silicic dome lavas in the Tauranga Volcanic Centre

The silicic dome lavas in the Tauranga area can be divided into four groups (Minden Peak, Mount Maunganui, Mangatawa, Mount Misery) based on their spatial association and mineralogical and geochemical characteristics, although their ages generally overlap. The four groups are summarised in Table 2 and their distribution is shown in Fig. 3. The mineralogical and geochemical data are from Hughes (1993), Hall (1994), and Whitbread-Edwards (1994).

The Minden Peak group forms the hornblende rhyolite domes northwest of the Wairoa River, and are characterised by high SiO$_2$, high K$_2$O, and intermediate Zr contents.

The Mount Maunganui group form a tight spatial cluster and have the most evolved compositions with the highest SiO$_2$ and K$_2$O contents and lowest Zr. The biotite rhyolite lava remnants of Motuotau and Moturiki Islands and Mount Drury are all genetically related to the lava dome of Mount Maunganui (Hall 1994).

The Mangatawa group of hornblende-orthopyroxene rhyolites and dacites may be aligned along northeast-trending faults in the basement that border the eastern segment of the Tauranga Volcanic Centre. They include the dacite domes of Upuhue and Kopukairua. The faults are not exposed at the surface, but numerous deeper older faults occur in the basement, shown in seismic reflection profiles offshore of Tauranga and the Bay of Plenty (Davey et al. 1995). The Mangatawa group lavas have low SiO$_2$, high K$_2$O, and intermediate Zr abundances, and are thus distinguished from the Mount Misery group, occurring in the same geographic area, which has the highest Zr contents of >200 ppm.

The silicic rocks of the Tauranga area have calc-alkaline compositions and major and trace elements and isotopic compositions similar to the rhyolites of TVZ. Tauranga rhyolites and dacites also have close affinities in their mineralogical compositions with TVZ rocks, and contain variable proportions of plagioclase, quartz, calcic hornblende, orthopyroxene, biotite, titanomagnetite, ilmenite, zircon, and apatite. The clusters of spatially associated rhyolites and dacites may represent genetically related silicic magmas that were independently and broadly synchronously derived. Their concordance with TVZ silicic volcanics suggest that they have a similar genesis, involving fractionation and partial melting of the crust (e.g., Graham et al. 1995), although there is evidence that other mafic magmatic components were also involved.

The Papamoa Ignimbrites span a wide age range and vary in composition from basaltic andesite to rhyolite, and have closest mineralogical and geochemical affinities with the Mangatawa group. The range in geochemical composition in terms of SiO$_2$, K$_2$O, and Zr contents is not included in Table 2, but the composition of the rhyolitic pumice in the Papamoa Ignimbrites falls in the range of those in the Mangatawa group, and they lack the high Zr contents of the Mount Misery group.

Nature and definition of volcanic centres in the Tauranga-Kaimai area, in comparison with TVZ

Onshore TVZ is a northeast-trending volcano-tectonic depression that extends c. 200 km southwest of the Bay of Plenty coastline, is 60 km wide, and contains andesitic and dacitic composite volcanoes, and eight rhyolitic calderas restricted to a central segment (Houghton et al. 1995; Wilson et al. 1995). Offshore TVZ extends 150 km to the northeast to the oceanic Kermadec Arc and Havre Trough/continental crustal transition (Gamble et al. 1993). The onshore rhyolitic calderas have been recognised from the location of lava dome complexes, negative Bouguer gravity anomalies associated with silicic vents, distribution and thicknesses of ignimbrites, location of lithic lag breccia facies in proximal ignimbrites, and structural evidence for collapse (e.g., Rogan 1982; Nairn et al. 1994; Wilson et al. 1995). Some calderas are nested or partially overlap, and more than 34 caldera-forming eruptions have occurred from these 8 volcanic centres (Houghton et al. 1995).

Earlier rhyolitic volcanic centres and caldera structures may also be present within the central TVZ, but these cannot be recognised because of structural downfaulting, or burial by younger voluminous ignimbrites and other eruptives (cf. Wilson et al. 1995). For example, there are some voluminous ignimbrites such as Waimakariri Ignimbrite whose sources are uncertain and that cannot be definitively attributed to any
of the presently known caldera volcanic centres in the TVZ (Milner 2001). In most cases, the rhyolite domes and lavas in the TVZ lie within or on the rims of calderas, or else are situated just outside the caldera rims. The only exceptions to this are some extra-caldera domes and associated subordinate pyroclastics southwest and southeast of the Taupo caldera that are linked geochemically to some intracaldera eruptives (Sutton et al. 1995).

The silicic rocks of the Tauranga-Kaimai area occur in spatially associated groups or volcanic centres, but there is no evidence for collapse structures or calderas. The negative Bouguer gravity anomaly in the northwestern area of the Tauranga Basin between Katikati and Matakania Island, described by Woodward & Ferry (1973), might indicate a caldera or collapse structure. However, it does not coincide with the distribution of the rhyolite and dacite domes and flows of the Tauranga-Kaimai area that lie well to the south and southeast. Lacustrine diatomaceous siltsstones are found in many places in the Tauranga area (e.g., Matakania Island; Hollis 1995), but there is no evidence that these siltsstones were deposited in caldera lakes. Circular or arcuate structures that may indicate caldera structures have been observed by remote sensing farther north in the CVZ (Skinner 1986; Belliss & Christie 1994; Christie et al. 1994; Krippner 2000), but have not been seen in the Tauranga-Kaimai area. Harmsworth (1983) recorded numerous buried rhyolitic lavas intersected by drill cores in the Tauranga Basin, and hence if Pliocene domes and flows existed in the Tauranga-Kaimai area, they are now buried and obscured by younger volcanic and sedimentary infill. Alternatively, these calderas may never have formed, because the ignimbrite eruptions were not voluminous enough to be caldera forming.

Because there is no evidence for calderas or collapse structures in the Tauranga-Kaimai area, the groups of silicic domes and flows and associated ignimbrites are designated as silicic volcanic centres and not as calderas, and this terminology and distinction from TVZ calderas is adopted as silicic volcanic centres and not as calderas, and this terminology and distinction from TVZ calderas is adopted.

REFERENCES


