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Volcano-meteorological tsunamis, the c. AD 200 Taupo eruption (New Zealand) and the possibility of a global tsunami

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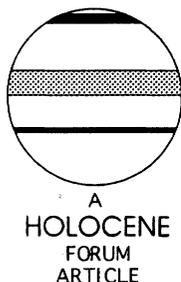
Abstract: Meteorological tsunamis are long-period waves that result from meteorologically driven disturbances. They are also generated by phase coupling with atmospheric gravity waves arising through powerful volcanic activity. The AD 1883 Krakatau eruption generated volcano-meteorological tsunamis that were recorded globally. Because of its extreme violence and energy release ($\gtrsim 150 \pm 50$ megatons explosive yield), and by analogy with the Krakatau event, it is highly possible that the ignimbrite-emplacement phase of the *c.* AD 200 Taupo eruption of North Island, New Zealand, generated a similar volcano-meteorological tsunami that may have reached coastal areas worldwide. Tsunami deposits of identical age to the Taupo eruption occur in central coastal New Zealand and probably relate to that event; definitive evidence elsewhere has not yet been found. In theory, volcano-meteorological tsunamis are likely to be produced during comparable eruptive events at other explosive volcanoes, and thus represent an additional volcanic hazard at coastal sites far from source. We suggest that evidence for such tsunamis, both for marine and lacustrine environments, may be preserved in geological records, and that further work searching for this evidence using a facies approach is timely.

Key words: Tsunami, meteorological tsunami, volcano-meteorological tsunami, palaeotsunami, atmospheric phase coupling, volcanic eruptions, seiches, natural hazards, ignimbrite, Taupo eruption, New Zealand.

Introduction

Tsunamis are long-period waves generated by seafloor disturbances in the ocean, ('Tsunami', as a collective noun, may be singular or plural (de Lange, 1998), but commonly 's' is appended to form the plural to avoid ambiguity.) Most disturbances are of seismic origin or result from volcanic processes (Table 1). However, a special type of tsunami is the 'meteorological tsunami' that results from the generation of waves by phase coupling with atmospheric compressional gravity or shock waves (Donn and Balachandran, 1969; Rabinovich and Monserat, 1996). Known by various local names (e.g., abiki, yota, rissaga, Seebär), these waves are not true tsunamis *sensu stricto* but are identical to 'normal' tsunamis in having long-period (*c.* 6–40 minutes) waves with speeds determined by water depth. They originate mainly through disturbances in the atmosphere, hence water surfaces, via the passage of strong cyclones (typhoons) or frontal squalls, or by atmospheric pressure jumps and trains of atmospheric gravity waves (Rabinovich and Monserrat, 1996). Such large long waves, like their seismogenic or volcanogenic counterparts, have caused loss of life and catastrophic destruction in coastal areas and thus represent a significant but previously underestimated natural hazard.

Meteorological tsunamis may also be generated through another

Table 1 Mechanisms for the generation of tsunami during volcanic eruptions (after Latter, 1981)[*]

Earthquakes accompanying eruptions
Submarine explosions
Pyroclastic flows entering the sea
Caldera collapse or subsidence
Landslides and avalanches of hot or cold rock or lahars entering the sea
Lava entering the sea
Base surges and associated shock waves
Phase coupling with atmospheric shock waves (this paper)

[*]Almost 25% of the deaths caused by volcanic eruptions have been attributed to accompanying volcanogenic tsunamis (Latter, 1981), and Tanguy *et al.* (1998) reported that 17% of the fatalities from eruptions since AD 1783 were the result of such tsunamis.

mechanism, namely powerful volcanic activity, as shown by an analysis of the AD 1883 Krakatau eruption (Ewing and Press, 1955; Press and Harkrider, 1966; Simkin and Fiske, 1983). This eruption evidently generated volcanogenic meteorological tsunamis, hereby termed 'volcano-meteorological tsunamis'. The
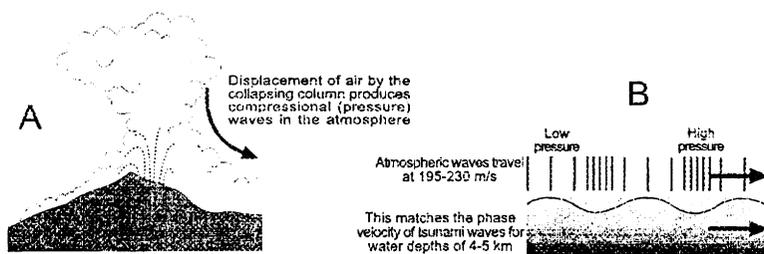
Figure 1 Model for generation of volcanically induced meteorological tsunamis by atmospheric phase coupling during a powerful volcanic eruption (>100 megaton explosive yield). (A) The eruption displaces a large volume of air laterally, generating compressional (pressure) gravity waves in the atmosphere. This is most efficiently achieved by column collapse or a lateral blast. (B) High and low pressures depress and raise the water surface, respectively. If the pressure disturbance travels at the same speed as a tsunami wave, the water resonates and a volcano-meteorological tsunami is formed.

strongest gravity waves, on the basis of detailed records of atmospheric pressure at the Batavia (now Jakarta) Gas Works near Krakatau, coincided with emplacement of the major pyroclastic flows during the eruption (Latter, 1981; Self and Rampino, 1981). The effects were detected worldwide. For example, waves were observed at major ports in the North Sea (Ewing and Press, 1955; Press and Harkrider, 1966; Simkin and Fiske, 1983) and along the New Zealand coast where heights varied from 0.3 to 1.8 m. They were also recorded within 600 km²-Lake Taupo in New Zealand where the maximum wave heights during seiching reached 0.5 m (de Lange and Healy, 1986; de Lange, 1998). The accompanying acoustic waves were sufficiently strong to be heard in central and western Australia, Sri Lanka, and Rodriguez Island in the Indian Ocean, the last being *c.* 4650 km from Krakatau (Latter, 1981; Simkin and Fiske, 1983).

Thus a meteorological tsunami may be generated both in marine and lacustrine environments through the emplacement of ignimbrite through the collapse of an eruption column, or by a very powerful lateral blast, as modelled in Figure 1. We consider here the possibility that the *c.* AD 200 Taupo eruption, New Zealand, also generated a volcano-meteorological tsunami. It is likely that this tsunami had major impacts locally and may have been felt on a global scale.

Comparison of the Taupo and Krakatau eruptions

The Taupo eruption, which occurred *c.* 1850 ^{14}C BP (Froggatt and Lowe, 1990), was the latest and most powerful of the eruptions from the rhyolitic Taupo caldera volcano in Taupo Volcanic Zone (Figure 2) since *c.* 22 600 ^{14}C BP (Walker, 1980; Wilson and Walker, 1985; Wilson, 1993). The eruption involved five phases of plinian (including 'ultraplinian') and phreatomagmatic fall activity, and a climactic sixth phase resulted in the extremely violent emplacement of the Taupo Ignimbrite ('Unit Y7' of Wilson, 1993) over an area of *c.* 20 000 km². The pyroclastic flow is considered to have been erupted at velocities exceeding 200–300 m s^{-1} (Wilson, 1985; 1993; Wilson and Walker, 1985). No tsunami would have been directly generated by this emplacement because nowhere did the flows enter the sea (Figure 2). However, we propose that the eruption and emplacement of the ignimbrite could have generated a volcano-meteorological tsunami, based on similarities between this event and the AD 1883 Krakatau ignimbrite-generating eruption.

The largest AD 1883 Krakatau pyroclastic flow event produced about 1×10^{10} m³ of ignimbrite (Self and Rampino, 1981). Taking this as the minimum tephra volume, the tephra volume index k_{Vtd}
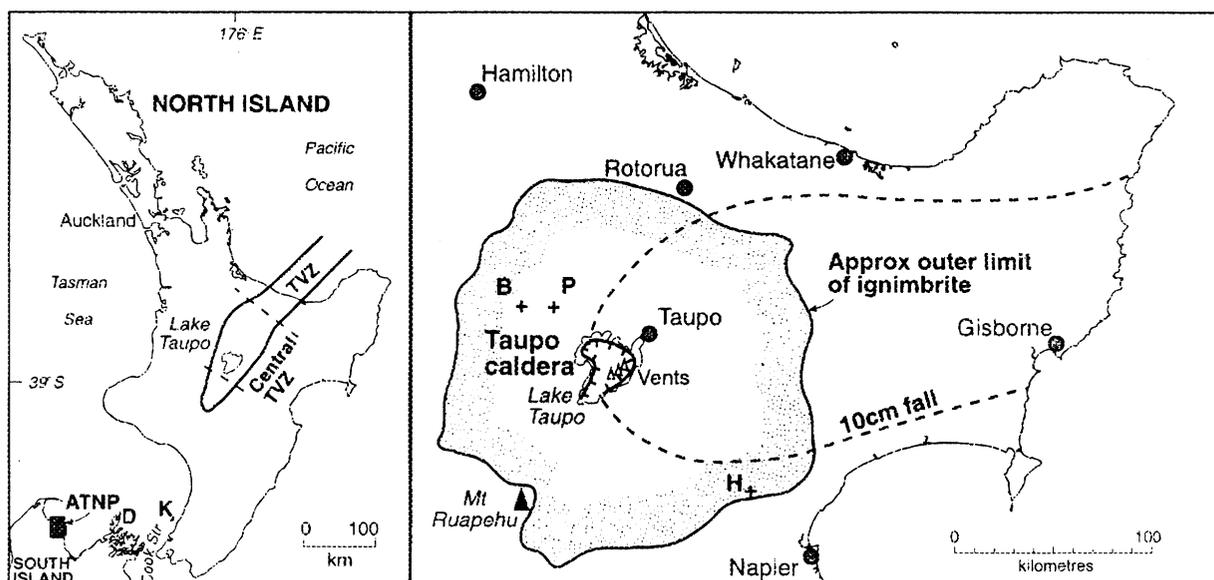
Figure 2 Distributions of Taupo Ignimbrite and 10 cm isopach for tephra-fall deposits (after Wilson and Walker, 1985) of the Taupo eruption from Taupo caldera volcano in the central Taupo Volcanic Zone (TVZ; Houghton *et al.*, 1995; Wilson *et al.*, 1995), New Zealand. Vent positions after Smith and Houghton (1995). B and P mark Benneydale and Pureora buried forests, respectively; H = Hukanui cave, Puketitiri. Location map: K = Kapiti Island; D = D'Urville Island; ATNP = Abel Tasman National Park.

is 6 and the eruption size index M_{VEI} is 6.0 (Pyle, 1995). The Taupo Ignimbrite deposit has a volume of about 3×10^{10} m^3 (Wilson, 1993) corresponding to $k_{VEI} = 6$ and $M_{VEI} = 6.4$. Both events are likely to have occurred within 300–400 seconds (Self and Rampino, 1981; Wilson, 1993). Thus, these two events may be considered to have a similar energy release, equivalent to 150 ± 50 megatons of TNT, based on Pyle (1995). However, other volcanological indicators for the emplacement of Taupo Ignimbrite, including the deposit's low-aspect ratio, its wide range of facies and lateral variations, the mountainous topography it surmounted, and its 'scalping' and overturning of substrates, all point to an extremely high eruption rate ($c. 3 \times 10^7$ m^3 s^{-1}) and an exceptional rate of release of kinetic energy (Wilson and Walker, 1981; 1985; Walker and Wilson, 1983; Wilson, 1985; 1993; Smith and Houghton, 1995), exceeding the Krakatau event both in magnitude and emplacement violence. The estimated energy release figure for Taupo must therefore be considered a minimum.

On this basis, it is highly probable that volcano-meteorological tsunamis were generated around the globe by the eruption of Taupo Ignimbrite. In the case of the Krakatau event, the antipodean waves were difficult to see without the aid of tide gauges. Nonetheless, they were observed. Further, the circum-Pacific and Indian Ocean meteorological tsunamis had maximum heights exceeding 0.25 m with periods of around 20 minutes. These were reported by casual observers (e.g., Simkin and Fiske, 1983). Therefore it is likely that noticeable disturbances would have occurred at locations around the Pacific and Indian oceans within 20 hours of the eruption of the Taupo Ignimbrite.

Age of the Taupo eruption

In Table 2 and Figure 3 we document all published dates obtained for the Taupo eruption using radiocarbon dating, dendrochronology, ice-core dating and interpretations of historical Chinese and Roman records. New radiocarbon ages obtained on short-lived seeds and leaves from trees killed by the emplacement of Taupo Ignimbrite are listed in Table 3.

Calibrated dates estimated from the radiocarbon measurements, spanning an interval of nearly 200 cal. years at the 2-sigma level ($c.$ AD 130 to AD 320) (Table 2), are imprecise because the calibration curves in this period fluctuate and have relatively large errors (Stuiver et al., 1998). The two dendrochronology-based age estimates are some 50 cal. years apart, that most recently obtained being AD 232 ± 15 (Sparks et al., 1995). However, this latter date does not agree with a date of AD 181 ± 2 derived from the Greenland GISP2 ice core (Zielinski et al., 1994). Consequently, the precise year of the eruption is uncertain, which is why we use $c.$ AD 200 as an approximate estimate for now. It is known from botanical and dendrochronological data that the eruption took place during the austral late summer or early autumn (Table 2). Based on the data in Table 2, we targeted $c.$ AD 130 to AD 320 as the time interval for documented tsunamis in both historical and geological records that might relate to the Taupo eruption.

Evidence for tsunami deposits associated with the Taupo eruption

Finding definitive evidence for prehistoric, and even historic, tsunamis is difficult (e.g., Dawson et al., 1991; Chagué-Goff and Goff, 1999a; 1999b; Goff and Chagué-Goff, 1999). Volcano-meteorological tsunamis will tend to have small waves (<1 m) except close to source, or where resonance amplifies the waves. The AD 1883 Krakatau eruption produced waves in the River Thames, England, that were evident to casual observers (Press and Hark-

rider, 1966), even though they were much less than 1 m in height. The same eruption resulted in waves as high as 1.8 m in Auckland Harbour (Tamaki Estuary), New Zealand, the considerably larger size here being the result of resonance and amplification. These are the largest tsunami-like waves recorded at Auckland in the last 180 years (de Lange and Healy, 1986; de Lange, 1998; de Lange and Fraser, 1999), and they were larger than the Krakatau-generated waves observed on the open coast around New Zealand. Therefore, although we consider that the likely global response is small, locally there may be a much larger (and potentially damaging) response (e.g., Gomis et al., 1993).

Small waves may go unnoticed by coastal inhabitants or, if noticed and recorded, they may not be obvious as a tsunami event to the researcher (Cox, 1979; de Lange, 1998). In the case of meteorological tsunamis, waves may also form in lakes, where they may be more likely to be reported as unusual. It is possible that seiching and 'overspilling' of lakes for unknown physical reasons, as reported in some historical records (e.g., Baillie, 1999), may be volcano-meteorological in origin, and the Taupo eruption clearly is a contender for any such event occurring in the target period. However, seiching in a lake is not normally linked to tsunami activity at the coast. Therefore it is likely that any tsunami associated with the Taupo eruption will not be recorded as a tsunami event in any accessible data base. This is true of the international World-wide Tsunami Database (NGDC, 1997): only two historical tsunami events have been recorded for the Pacific during the target period. The first, reported in China AD 28 June 173 ('Po Hai' tsunami), was probably associated with a local earthquake (Iida et al., 1967; Soloviev and Go, 1974). The other event, documented in southern Mexico AD 17 March 186, is now known to be an erroneous data base entry. In the Mediterranean region several possible contenders are recorded, both at Rhodes, in AD 142 and AD 148, and another event is recorded for the Dead Sea in AD 315 (NGDC, 1997). However, the origin of these tsunamis is uncertain and they may relate to earthquakes or other causes.

Closer to the New Zealand region there are no historical records that extend sufficiently far back to cover the Taupo eruption. Therefore it is necessary to search the geological record. However, it may be possible to identify only large events. Small tsunamis are difficult to recognize in the geological record, mainly because the features of tsunami deposits are similar to deposits produced by other processes. At present there is no single feature or 'signature' characteristic of a tsunami deposit and a facies approach with multiple diagnostic criteria is required. Tsunami facies are currently poorly defined but recent studies using various criteria have shown some success in identifying prehistoric tsunamis (palaeotsunamis) (e.g., Dawson, 1996; Hemphill-Haley, 1996; Chagué-Goff and Goff, 1999a; 1999b; Clague et al., 1999; Goff and Chagué-Goff, 1999).

Few sites in New Zealand have been examined for evidence of palaeotsunamis. The best-documented sites are in Abel Tasman National Park, South Island, on the southwestern margin of Cook Strait (Goff and Chagué-Goff, 1999; Figure 2). These sites have been little affected by historic tsunamis (de Lange and Healy, 1986; Fraser, 1998), with the shallow waters of the strait and adjacent bays attenuating tsunami energy. Only two historic events are known to have produced appreciable tsunami waves in the area, both caused by local earthquakes in AD 1855 and AD 1868. The first event produced waves 5–10 m in height in Cook Strait; the second generated waves <1 m high. Analysis of cores taken from wetlands in the park showed evidence for several tsunami events, including a deposit produced by the AD 1855 event. No trace of the AD 1868 event is evident (Goff & Chagué-Goff, 1999). This suggests that tsunami wave heights exceeding $c.$ 5 m in Cook Strait are probably needed to leave a recognizable deposit. The AD 1855 deposit is also the least substantial (thinnest)

Table 2 Estimates of the date of the Taupo eruption

Basis of estimate	Reference*	Radiocarbon age (BP)§	Calibrated or calendar age
(A) Radiocarbon dating			
Multiple ages on charcoal, wood, peat, lake sediment, seeds (*n* = 41)	1	1850 ± 10	1σ AD 135–199 (0.67)ᵃ AD 209–237 (0.33) 2σ AD 132–240 (1.00)
High-precision and AMS ages on leaves and seedsᵇ (*n* = 7)	2, 3 (Table 3)	1845 ± 19	1σ AD 130–240 (1.00)ᵃ 2σ AD 130–320 (1.00)
(B) Dendrochronology			
Cross-matching of *Phyllocladus trichomanoides* (tanekaha) chronology with tree-ring sequence of *Prumnopitys taxifolia* (matai)	4		AD 232 ± 15
Curve-matching of floating tree-ring sequence of *Phyllocladus trichomanoides*ᶜ	1, 5		*c*. AD 177 (166–195)ᵈ
(C) Greenland ice-core record			
Layer of sulphuric acid in GISP2 ice coreᵉ	6		AD 181 ± 2
(Adjusted by +40 yr because of postulated lag compared with tree-ring records)ᶠ	7		(?*c*. AD 221 ± 2)
(D) Ancient historical records			
Association with unusual atmospheric phenomena reported in Chinese and Roman literatureᵍ	8		*c*. AD 186
(E) Time of year of eruption			
Botanical and dendrochronological data from buried forest at Pureora (Figure 2)	5, 9		Austral late summer or early autumn (approx. March–early April)

*(1) Froggatt and Lowe (1990); (2) Lowe (1993); (3) Holdaway and Beavan (1999); (4) Sparks *et al.* (1995); (5) Palmer *et al.* (1988); (6) Zielinski *et al.* (1994); (7) Baillie (1996); (8) Wilson *et al.* (1980); (9) Clarkson *et al.* (1988).

§Conventional radiocarbon years before present based on the old (Libby) half-life. Reported as error-weighted mean ± 1 standard deviation.

ᵃBased on INTCAL98 terrestrial calibration curves of Stuiver *et al.* (1998) after subtraction of 27 years for the Southern Hemisphere offset (McCormac *et al.*, 1998a; 1998b) and using OxCal 3.0 (Bronk Ramsey, 1995). Values in parentheses are probabilities.

ᵇWell-preserved macrofossil samples from trees killed by emplacement of Taupo Ignimbrite at Pureora and Bennydale buried forests, western Taupo (see Clarkson *et al.*, 1988; 1992; 1995), and carbonized leaves within Taupo Ignimbrite at Hukanui cave, Puketitiri, inland Hawke's Bay (Holdaway and Beavan, 1999).

ᶜFrom logs killed by emplacement of Taupo Ignimbrite at Pureora.

ᵈ1-sigma range in parentheses.

ᵉAssumes that the layer of sulphuric acid at AD 181 ± 2 in GISP2 core represents deposition of aerosols specifically from the Taupo eruption. This assumed correlation with the Taupo events has yet to be validated (e.g., using microanalysis of glass as in Zielinski and Germani, 1998a; 1998b) and it is conceivable that the layer may represent fallout from an alternative eruption, as yet unidentified – e.g., the northern lobe of the White River Ash, erupted from Mt Bona-Churchill, Alaska, and with a tephra volume of *c*. 10¹⁰ m³ (Simkin and Siebert, 1994), has an error-weighted mean radiocarbon age (*n* = 11) of 1887 ± 28 ¹⁴C BP (Lerbekmo *et al.*, 1975), corresponding to a 2-sigma calibrated range of AD 63–222 (*P* = 1.00) based on INTCAL98 (Stuiver *et al.*, 1998) and using OxCal 3.0 (Bronk Ramsey, 1995). Similarly, Ilopango caldera volcano, El Salvador, generated *c*. 10¹⁰ m³ of silicic material in an eruption *c*. AD 260 ± 150 (Simkin and Siebert, 1994; Zielinski *et al.*, 1994).

ᶠBaillie (1996) suggested that part of the GISP2 ice-core record (second millennium BC) is 'too old' by *c*. 40 yr (*c*. 2350 BC; based on tree-ring records). If this lag is confirmed and if it is applicable to the period relevant to the Taupo eruption, then the date of AD 181 ± 2 would shift by +40 yr to *c*. AD 221 ± 2. However, the conclusions of Baillie (1996) are disputed by Buckland *et al.* (1997) and Zielinski and Germani (1998a), and the purported lag may not necessarily apply to the later period.

ᵍThe postulated link between the atmospheric effects and the Taupo eruption is controversial (Froggatt, 1981; Wilson *et al.*, 1981; Froggatt and Lowe, 1990). Stothers and Rampino (1983) purported that there were errors in the translation of the Roman text used by Wilson *et al.* (1980) and suggested that the literary reference was to a supernova.
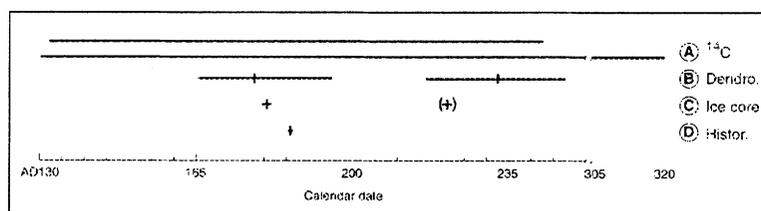
Figure 3 Timeline comparing estimates of the date of Taupo eruption (from Table 2). (A) 2 σ ranges for calibrated radiocarbon ages (interhemispheric offset-corrected, derived by OxCal): upper line 1823 ± 10 ¹⁴C BP, lower line 1818 ± 19 ¹⁴C BP; (B) dendrochronological estimates; (C) Greenland GISP2 ice-core data (with postulated +40 years-offset from dendrochronological data); (D) Chinese and Roman records of atmospheric effects.

Table 3 Radiocarbon ages on leaves and seeds from trees killed by emplacement of Taupo Ignimbrite at Benneydale and Pureora, and at Hukanui cave, Puketitiri

Lab. no.*	Age (^{14}C BP)	δ^{13}C (‰)	Sample type§	Location¶
Wk-2707a	1870 ± 45	−28.0	Rimu leaves (*Dacrydium cupressinum*)	Ohirea Rd, Benneydale S17/161963
Wk-2707b	1860 ± 45	−28.0	Rimu leaves (*D. cupressinum*)	Plains Rd, Pureora T17/347974
Wk-2708a	1850 ± 45	−23.1	Miro (*Prumnopitys ferruginea*) and matai (*P. taxifolia*) seeds	Ohirea Rd, Benneydale S17/161963
Wk-2708b	1850 ± 50	−23.1	Miro (*P. ferruginea*) and matai (*P. taxifolia*) seeds	Ohirea Rd, Benneydale S17/161963
Wk-2709	1840 ± 50	−27.2	Toro leaves (*Myrsine salicina*) (from tree crown)	Ohirea Rd, Benneydale S17/161963
NZA-7532	1715 ± 69	−27.9	Carbonized lacebark(?) leaf remains (*Hoheria sexstylosa?*)	Hukanui Pool (cave), Puketitiri V20/149115
NZA-8226	1871 ± 67	−28.6	Carbonized leaf (unidentified)	Hukanui Pool (cave), Puketitiri V20/149115

*Wk- prefixes refer to University of Waikato Radiocarbon Dating Laboratory (high-precision liquid scintillation spectrometry); NZA-prefixes refer to Rafter (formerly New Zealand) Radiocarbon Laboratory, Institute of Geological and Nuclear Sciences (accelerator mass spectrometry).
§Wk- samples collected and identified by B.R. Clarkson in December 1984 (Wk-2709) and August 1986 (Wk-2707, Wk-2708) (Lowe, 1993; this study); NZA- samples collected by R.N. Holdaway (Holdaway and Beavan, 1999).
¶See Figure 2; grid reference applies to the metric 1:50 000 topographical map series NZMS 260.

of those identified, suggesting that the other tsunamigenic events were possibly larger.

Based on radiocarbon dating, all but one of the palaeotsunami deposits in Abel Tasman National Park can be correlated with large, local earthquakes identified previously from palaeoseismic work (Goff and Chagué-Goff, 1999). Because the faults involved in these earthquakes are all closer to the park than that which generated the AD 1855 event, it seems likely that differences in deposit thickness are related to proximity to source and thus tsunami size.

The oldest tsunami event identified by Goff and Chagué-Goff (1999) occurs at the base of the cores. The deposit, thicker than that associated with the AD 1855 event, has been dated by radiocarbon (on charcoal) at 1774 ± 73 ^{14}C BP (NZA-6230), which calibrates to a calendrical date range of AD 125–431 ($P = 0.99$) and AD 87–101 ($P = 0.01$) based on Stuiver *et al.* (1998) and Bronk Ramsey (1995) after interhemispheric offset correction (McCormac *et al.*, 1998a; 1998b). This age range overlaps those for the Taupo eruption (Figure 4). There are no known local earthquakes of sufficient magnitude to generate a tsunami for this time period. Therefore we consider this deposit may represent the local volcano-meteorological tsunami generated by the Taupo eruption.

Probable tsunami deposits at Kapiti Island, southwestern North Island (Figure 2), of late-Holocene age may similarly be related to the Taupo eruption (Goff *et al.*, 2000). Pumice clasts preserved in peat overlying an uplifted shore platform at a site near the

tsunami deposits on Kapiti are likely to be Taupo-derived, and may be coeval with the tsunami event (J.R. Goff, personal communication, 1999). We note also that additional pumice deposits, derived primarily from plinian fallout and possibly also fluvially transported ejecta from the earlier phases of the Taupo eruption, are widely reported at coastal sites along North Island's east coast, on D'Urville Island near Abel Tasman National Park (Figure 2), and on Chatham Island, which is 900 km east of South Island (Wellman, 1962; Pullar *et al.*, 1977; McFadgen, 1985; 1989; 1994a). Although commonly attributed to sea-rafting processes, it is conceivable that some of these reworked, stranded pumice deposits, which can occur well inland from the modern shoreline, originated from the postulated Taupo-generated meteorological tsunami. One such stranded shoreline ridge, 4 m wide and 0.7 m thick, occurs 160 m inland of, and 3.5 m above, high water mark at Waitangi West Beach, Chatham Island (McFadgen, 1994b).

Conclusions and implications

(1) Meteorological tsunamis are produced by phase coupling with atmospheric gravity or shock waves, the latter usually being associated with meteorological forces. Volcanogenic meteorological tsunamis may be generated through powerful volcanic activity, including emplacement of ignimbrite or lateral blasts. The AD 1883 Krakatau eruption generated volcano-meteorological tsunamis that were recorded globally both in marine and lacustrine environments.

(2) Because of its extreme violence and high energy release (equivalent to $\geq 150 \pm 50$ megatons of TNT), and by analogy with the Krakatau event, it is highly probable on theroretical grounds that the ignimbrite-emplacement phase of the Taupo eruption *c.* AD 200 generated a similar, possibly larger, volcano-meteorological tsunami with worldwide coastal impacts.

(3) Tsunami deposits of identical age to the Taupo eruption, and emplaced by waves $\geq c.$ 5 m in height in Cook Strait, occur in central New Zealand. In the absence of large palaeoseismic events at this time, we contend that these deposits represent the local volcano-meteorological tsunami generated by the Taupo eruption. Strandlines of reworked Taupo-derived pumice found along New Zealand coastlines, usually attributed to sea-rafting processes, may have originated in part from the Taupo-generated tsunami.

(4) Tsunamis that coincide with the timing of the Taupo eruption are not yet known in the historical record, although possible contenders are described in the Mediterranean area. Finding definitive evidence for early historic or prehistoric tsunamis is difficult, however, because historical records are incomplete, tsunami
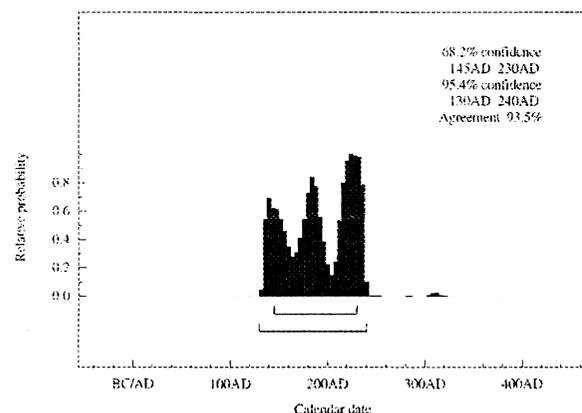
Figure 4 Comparison of radiocarbon ages for the Taupo eruption (Table 2A) with an age of 1774 ± 73 ^{14}C BP on possible coeval tsunami deposits in Abel Tasman National Park (Goff and Chagué-Goff, 1999). The OxCal combined probability method (Bronk Ramsey, 1995) shows a high level of agreement between the calibrated ages (93.5%).

'signatures' are not well defined, and volcano-meteorological tsunamis will tend to have small waves (<1 m) except close to source or where resonance amplifies the waves. If a probable link with an historic tsunami (marine or lacustrine) could be established, it would provide an additional dating method to refine the history of the Taupo volcano. In addition, the widely dispersed products of the Taupo eruption thus dated would provide a chronostratigraphic isochronous marker horizon of likely global extent.

(5) We suggest that a global search for geological evidence of tsunami deposits associated with the Taupo eruption, probably using a facies approach, would be worthwhile. Volcano-meteorological tsunamis generated from other highly energetic and voluminous eruptions (e.g., Kawakawa eruption *c.* 22 600 ^{14}C BP, Taupo volcano; Wilson, 1991) may also be preserved in the geological record.

(6) Volcano-meteorological tsunamis represent a volcanic hazard at coastal sites both proximal and far from source volcanoes. The hazards resulting from a powerful eruption at Taupo, the world's most productive single rhyolite volcano (Wilson, 1993), or its near neighbour at Okataina, would not be confined to New Zealand. Comparable eruptive events from ignimbrite-generating volcanoes elsewhere would also probably have global coastal impacts both in marine and lacustrine environments though the seriousness of such impacts would depend on local circumstances.

Acknowledgements

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