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PALEOHYDROLOGY OF BREAK-OUT FLOODS FROM VOLCANOGENIC LAKES IN THE TAUPO VOLCANIC ZONE, NEW ZEALAND

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INTRODUCTION

All floods result from the rapid release of water, stored either in the atmosphere or in terrestrial reservoirs (lakes, snow and ice), onto the landscape (O'Connor & Webb, 1988). The magnitude of the flood therefore reflects the characteristic volume and release rates of the water source and the physiographic properties of the landscape that receives it, while the combination of geologic, climatic and physiographic factors make some areas of the Earth more flood-prone than others. Large-scale terrestrial impoundments of surface water, i.e. lakes, are an effective source of major floods because they can release huge volumes of water directly into a drainage route should their margins be breached.

Conversely, a large meteorological flood requires that precipitation and run-off be integrated across a wide geographic area over a period of time. Temporary lake impoundments can develop behind barriers composed of ice, morainic material, landslides, fluvial deposits, and volcanic debris (Costa & Schuster, 1988). Globally, 16% of the 714 identified Holocene (last 10000 years) volcanoes host one or more crater lakes (Delmelle & Bernard, 2000), while 8% of naturally-dammed lakes have volcanic origins (Costa & Schuster, 1988), being impounded by debris avalanche (volcanic landslide) deposits, lava flows, pyroclastic flows, or lahar deposits.

THE TAUPO VOLCANIC ZONE

The Taupo Volcanic Zone (TVZ) in the central North Island of New Zealand is an area of intense Quaternary silicic and andesitic volcanism associated with extension and thinning of continental crust at the southern end of the Tonga-Kermadec volcanic arc. The region is characterized by frequent caldera-forming explosive eruptions from multiple short-lived, nested and/or overlapping volcanic centres (Houghton *et alii*, 1995; Wilson *et alii*, 1995). Individual eruption volumes have ranged up to 10^2-10^3 km^3, with primary fall deposition occurring over >10^6 km^2 and ignimbrites covering >10^4 km^2. The combined effects of volcanism and faulting in the central TVZ have led to the creation and destruction of numerous lakes spanning several orders of magnitude in scale and longevity developed in topographic lows formed by explosion craters, volcano-tectonic calderas, and depressions dammed by pyroclastic and lava flows, lahar deposits, and debris avalanches (Lowe & Green, 1992; Smith *et alii*, 1993; Manville, 2001b). Amongst these, historic break-out floods have been recorded from the summit Crater Lake of Mt. Ruapehu (O'Shea, 1954), and from intracaldera Lake Tarawera (White *et alii*, 1997). Prehistoric break-out floods have also been identified from the geological record at Lake Tarawera (Hodgson & Nairn, 2000) and intracaldera Lake Taupo (Manville & Wilson, 2004; Manville *et alii*, 1999), and inferred at Lake Rotorua. It is these events which form the subject of this paper.

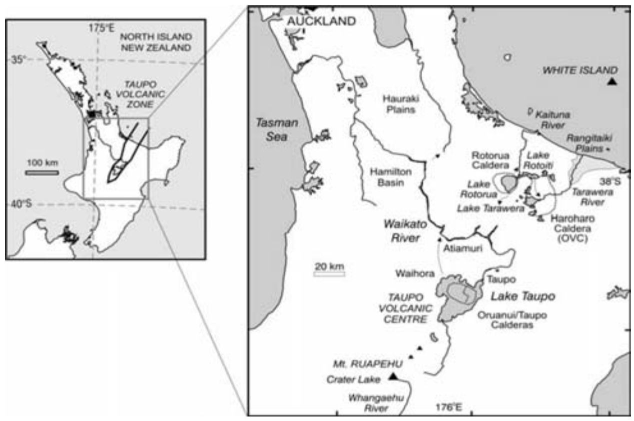

Figure 1 Location map of volcano-lake systems in the Taupo Volcanic Zone known to have produced breakout floods. Former drainage paths indicated by arrows

PALEOHYDRAULIC ANALYSIS

Typical paleohydraulic analyses consist of converting geologic stage-level indicators and channel geometry to estimates of peak discharge (O'Connor & Webb, 1988), with flow velocity and duration being additional important parameters. An additional factor in break-out floods is the outflow hydrograph from the impoundment, which is governed by such parameters as the geometry and growth rate of the breach , and reservoir hypsometry (Walder & O'Connor, 1997).

DAM-BREAK ANALYSIS

Estimation of the potential flood hydrograph from a dam-breach is complicated by its dependence on a number of inter-related factors: the volume and area of the lake; the geometry, size and composition of the dam; the breach dimensions and geometry; and the breach development time. Four main techniques (MANVILLE, 2001a), have been developed to assess the magnitude of potential dambreak floods from a range of natural (landslide, moraine, ice) and artificial dams (concrete, earthen): (i) empirical regression relationships between observed values of peak discharge and some dimensional characteristic of the lake, dam, or breach (e.g. COSTA, 1988; FROEHLICH, 1987; MACDONALD & LANGRIDGE-MONOPOLIS, 1984; WALDER & O'CONNOR, 1997); (ii) parametric models, based on timestepping calculations of flow through a developing breach using published broad-crested weir-flow equations (e.g. CHOW, 1959; PRICE *et alii*, 1977); (iii) dimensional analysis, a method of grouping parameters of unknown significance together in order to simplify relationships and compare their relative importance (e.g. SINGH & QUIROGA, 1988; WALDER & O'CONNOR, 1997; WEBBY & JENNINGS, 1994); and (iv) physically-based dam erosion models, often computer-implemented (e.g. COLEMAN & ANDREWS, 2000; FREAD, 1996; PONCE & TSIVOGLOU, 1981). Of these, (i)-(iii) require the minimum input data and are hence the most useful for paleohydraulic analyses where information is limited.

Empirical analysis

Regression relationships between observed values of peak discharge and some dimensional characteristic of the lake or dam, such as the total drop in lake level during the flood (often synonymous with the depth of the lake), and/or the volume of water released (also referred to as the excess lake volume) typically take a power-law form with empirically-derived coefficients (e.g. COSTA, 1988; WALDER & O'CONNOR, 1997). However, the actual peak discharge of a dam-break flood is rarely a simple function of the volume of lake water or dam height/lake depth or some combination of the two. Furthermore, the data on which the relationships are based are often collected in different ways, and the actual data points themselves are highly scattered. Therefore, regression lines can be fitted almost at whim, and different published equations can produce estimates of peak discharge that vary by an order of magnitude for the same inputs (COSTA, 1988).

An alternative set of empirical relationships has been developed based on the dimensional characteristics of the breach as well as the lake/dam system (e.g. JOHNSON & ILLES, 1976; MACDONALD & LANGRIDGE-MONOPOLIS, 1984; WEBBY & JENNINGS, 1994). However, these are difficult to apply unless the breach geometry is well-preserved.

Dimensional analysis

Dimensional analysis, as applied to dambreak phenomena is a technique for investigating the relative importance of breach growth rate and lake volume and lake shape on the peak discharge of a dambreak flood (WALDER & O'CONNOR, 1997), assuming that the fundamental physical mechanisms involved are the same for all dambreak flood events. Given that an ideal dam breach behaves as a broad-crested weir with critical flow through the outlet (i.e. Froude No., F =1), breach discharge Q_p and lake volume V_o can be recast in dimensionless terms:

dimensionless peak discharge $$Q_p^* = \frac{Q_p}{g^{0.5} d^{2.5}} \qquad (1)$$

dimensionless lake volume $$V_o^* = \frac{V_o}{d^3} \qquad (2)$$

where g is gravity and d the drop in water level or breach depth.

Analysis of a physically-based model of dam-breach formation shows that the dimensionless peak discharge is primarily a function of the dimensionless lake volume and the dimensionless breach erosion rate k^* $(= k/g^{0.5}d^{0.5})$ so that only three parameters are required to predict the full outflow hydrograph: d, V_o, and the vertical breach erosion rate k (typically 10-100 m/h). The last two factors are most critical as these govern whether or not substantial drawdown of the lake occurs before the breach has reached its maximum size. Other influences on outflow rate such as breach geometry (width-depth ratio) and lake hypsometry are relatively well constrained and can be excluded for simplicity. The breach hydrograph therefore depends on a dimensionless parameter given by WALDER & O'CONNOR (1997):

$$\eta = k^* V_o^* = \frac{kV_o}{g^{0.5} d^{3.5}} \qquad (3)$$

In practical terms, Q_p is only influenced by the breach erosion rate for $V_o^* < 10^4$, while $\eta > 1$ means that breaching of the dam is effectively instantaneous so that peak discharge is solely controlled by the dimensions and geometry of the outlet. In dimensional form the relationship becomes:

$$Q_p = C\, g^{0.5} d^{2.5} \qquad (4)$$

The value of the coefficient C depends on the breach geometry and lake hypsometry. Solution of the relevant equations for a number of general and special geometric/hypsometric cases produces a series of graphs which enable dimensionless peak discharge to be rapidly determined from η: for example, C has the value 1.94 for overtopping failures where $\eta \gg 1$ (WALDER & O'CONNOR, 1997).

Parametric methods

An alternative means of reconstructing the outflow hydrograph from a dam failure is to apply open-channel weir flow equations (e.g. FREAD, 1996; PRICE *et alii*, 1977) in a time-stepping numerical model which balances the flow through an enlarging breach with the rate of reservoir drawdown and hence hydraulic head reduction (MANVILLE, 2001a). Although this method tends to overestimate discharge because it neglects backwater and flow resistance effects and assumes level-pool routing in the lake, it offers a very rapid means of determining potential peak discharges for a range of combinations of lake size and breach geometry. The parameters most commonly required for this technique are the final dimensions of the breach (width, depth and sidewall slope), the breach development time, and a hypsometric description of the lake (surface area vs. elevation).

PALEOHYDRAULIC ANALYSIS

While paleohydraulic analysis of breach development can give information about the maximum *potential* peak discharge at the outlet, actual evidence of high-magnitude discharges is better obtained from further downstream. Two methods of obtaining information about the size of a paleoflood are (i) analysis of the maximum size of particles that were transported, and (ii) analysis of the maximum stage heights and water surface profile of the flood, from which flow velocity and discharge can be reconstructed.

Flow competence and velocity

Local discharges can be estimated using established relationships between hydraulic flow parameters (i.e. velocity, shear stress, stream power) and the maximum size of particles that can be transported (COSTA, 1983; KOMAR, 1989; O'CONNOR, 1993). In an analysis of the Bonneville flood, O'CONNOR (1993) compared boulder deposits with local flow conditions derived from stepbackwater models (see below) to obtain a regression relationship between particle size D_i (the mean intermediate diameter in cm of the 5 largest boulders at a site) and flow velocity v (m/s):

$$v = 0.29\ D_i^{0.61} \qquad (5)$$

In combination with estimates of flow cross-section from channel geometry and stage level markers, such velocity estimates can be used to derive discharge.

Step-backwater modelling

Step-backwater analysis is based on the principle of conservation of mass and energy in a steady, gradually varying flow. A series of flow cross-sections defined by channel geometry and paleostage height indicators can be used to calculate an energy-balanced water-surface profile that is a function of discharge, flow resistance, and channel geometry (O'CONNOR & WEBB, 1988). Calculated profiles can be iteratively fitted to the geologic data to estimate peak discharge at various points along the flow path. However, the method is very sensitive to channel geometry, including cross-section stability (i.e. no erosion or deposition during the flood), spacing, and areas of ineffective flow such as channel embayments and flood plains. Hydraulic jumps arising from rapidly varying flow conditions, for example along steep irregular channels, can cause numerical instabilities in the model, while the key assumption of steady, slowly varying flow is often untenable for dam-break floods, which are characterized by a rapid rising limb. The technique has however been applied to bedrock gorge-confined reaches of prolonged paleofloods such as the Missoula (O'CONNOR & BAKER, 1992) and Bonneville events (O'CONNOR, 1993).

LAKE TAUPO

Lake Taupo is the largest lake in the modern TVZ, with an area of 620 km^2 and a volume of c. 60 km^3 (LOWE & GREEN, 1992). It partially occupies a volcano-tectonic collapse structure whose present configuration largely reflects caldera collapses during the 26.5 kyr Oruanui and 1.8 kyr Taupo eruptions, and faulting and downwarping on regional structures (DAVY & CALDWELL, 1998; WILSON, 2001). The lake overflows with a mean discharge of c. 130 m^3/s (including a 60 m^3/s contribution from hydro-electric diversions) into the 320 km long Waikato River, the North Island's longest and largest. The river descends 340 m in the first 180 km downstream of the outlet, through a steep, narrow valley confined between late Pleistocene fluvial terraces and ignimbrite gorges, before crossing the lowlands of the Hamilton Basin. Extensive hydroelectric development has turned the first 180 km of the Waikato River into a chain of eight reservoirs, which, coupled with two thermal power plants further downstream, produced ~25 % of New Zealand's power. Meteorological floods in the Waikato basin are buffered by the storage capacity of Lake Taupo and the thick permeable pyroclastic sequences that blanket much of the central North Island.

POST-26.5 KYR ORUANUI FLOOD

The c. 530 km^3 (magma-equivalent) Oruanui eruption from the Taupo volcanic centre at 26.5 kyr (WILSON, 2001) had a profound impact on the geography of the central North Island, destroying a long-lived Pleistocene lake system, creating a new landscape around the vent site where unwelded pyroclastic flow deposits accumulated to thicknesses of hundreds of metres, and producing a major new closed topographic depression through caldera collapse and ignimbrite emplacement (MANVILLE & WILSON, 2004).

Intracaldera shoreline terraces

Post-eruption, Lake Taupo accumulated to eventually reach a highstand level of c. 500 m a.s.l., as marked by a poorly preserved shoreline terrace (GRANGE, 1937; MANVILLE & WILSON, 2003; MANVILLE & WILSON, 2004). Initial overspill occurred through the Waihora Bay area, 20 km west of the present outlet, resulting in a drop in lake level of c. 20 m to an elevation controlled by a resistant horizon of welded Whakamaru Ignimbrite. This stable stillstand likely persisted for decades to centuries, allowing formation of a better-developed shoreline terrace before the ignimbrite barrier was eventually compromised near the present outlet in the Taupo area (Figure 2).

Tributaries downstream of Lake Taupo are inferred to have re-established and sappedtheir way back to the lake, possibly aided by drainage of an ephemeral lake developed at c. 360-400 m a.s.l. in the Reporoa Basin, which lowered base levels and accelerated erosion. Breaching of the rim of Lake Taupo triggered a drop in its level from c. 480 to c. 405 m a.s.l. and established the course of the upper Waikato River in essentially its modern form. Significantly, post-Oruanui fluvial terraces along the upper Waikato River grade to this 405 m lowstand level, not the 480 m a.s.l. stillstand. Drawdown is inferred to have been rapid, based on the absence of any intermediate shoreline terraces or tributary deltas surrounding the lake, or any resistant horizons in the ignimbrite barrier. Downcutting below 405 m a.s.l. was prevented by a broad surface of indurated Huka Falls Formation sediments and the accumulation of a several metre-thick lag of dense lithic gravel at the base of the breach (MANVILLE & WILSON, 2004).

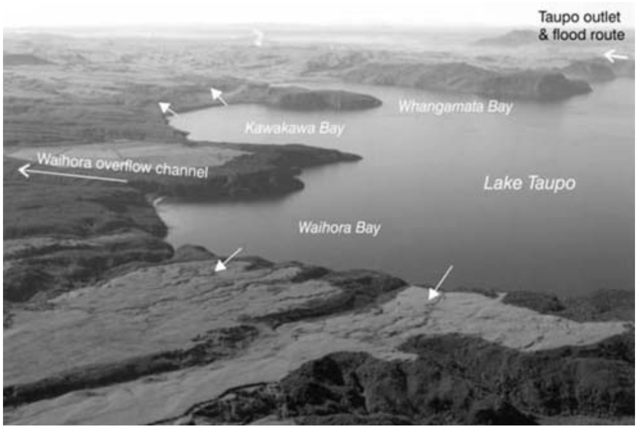

Figure 2 - Post-Oruanui shoreline terraces in the Western Bays of Lake Taupo (photo L.Homer, GNS)

Extracaldera flood features

A c. 80 m drop in the level of a water body of the magnitude of Lake Taupo corresponds to the release of c. 60 km^3 of water, enough to inundate the entire North Island of New Zealand to a depth of 0.5 m. Intracaldera evidence suggests that this release was relatively rapid, but the best indication of a catastrophic break-out flood is the Smythe's Quarry Boulder Member (THOMPSON, 1958), a massive bouldery unit (Figure 3).

This unit can be traced from 68 to 88 km downstream of Lake Taupo and rests unconformably on fluvially reworked gravels and sands that lie with erosive contact on primary Oruanui pyroclastics that fill a pre-eruptive paleovalley. The SQBM is buried by the main aggradational phase of post-Oruanui volcaniclastic sedimentation and was hence deposited relatively early in the post-eruptive chronology (MANVILLE & WILSON, 2004). The member comprises boulders between 1 and 10 m in intermediate diameter, often arranged in imbricate cluster bedforms and bar forms separated by anastomosing braid channels, in a matrix of finer gravels and coarse sands. Clast lithology varies downstream, reflecting local erosion and transport. Distinctive clasts of hydrothermally altered pumiceous breccia outcropping at Atiamuri have been transported at least 8 km from Ohakuri. This flood deposit is taken as marking the base of the Hinuera C Formation (MANVILLE & WILSON, 2004), the main phase of post-Oruanui volcaniclastic resedimentation coinciding with the peak of the Last Glacial Maximum (PILLANS *et alii*, 1993). Massive fluvial aggradation along the Waikato River eventually enabled the river to overtop a wind-gap and avulse into the Hamilton Basin, where it became confined by incision following climatic amelioration at c. 17.6 kyr (NEWNHAM *et alii*, 2003).

POST-1.8 kyr TAUPO FLOOD

The most recent explosive eruption from the Taupo volcanic centre occurred at 1.8 kyr (WILSON & WALKER, 1985). During a complex and multiphase eruption c. 35 km^3 of rhyolitic magma was expelled as a series of Plinian and phreatomagmatic fall deposits more than 0.1 m thick over an area of 30000 km^2 east of the vent, and as a climactic radially-distributed pyroclastic flow that devastated an area of 20000 km^2 (WILSON, 1985; WILSON & WALKER, 1985). The pyroclastic flow deposited two geomorphic variants: a thin veneer layer that drapes the pre-eruption landscape and a thicker, but still unwelded, valley-filling facies which infilled the Waikato River valley to depths of 40-70 m for 120 km downstream of the outlet. At the climax of the eruption, much of pre-eruption Lake Taupo was either expelled, evaporated, or drained into the sub-rectangular caldera collapse structure beneath the lake floor (DAVY & CALDWELL, 1998). The post-eruption highstand shoreline is marked by a semi-continuous, tectonically warped and offset, wave-cut bench and shoreline deposits at elevations of 28-43 m above the modern lake level of 357 m a.s.l. (RIGGS *et alii*, 2001; WILSON *et alii*, 1997). Refilling of Lake Taupo to the mean highstand level of +34 m is estimated to have taken 15-20 years, based on modern precipitation and run-off rates and basin hypsometry (SMITH, 1991; WILSON & WALKER, 1985), before the unconsolidated barrier of ignimbrite choking the outlet channel was breached by overtopping.

Failure is inferred to have been rapid, based on the absence of well-developed intermediate shoreline terraces between the +34 m highstand level and a semicontinuous wave-cut bench at +2-5 m, indicating rapid drawdown during a single phase that released c. 20 km^3 of water (MANVILLE *et alii*, 1999; WILSON *et alii*, 1997). A variety of paleohydraulic techniques have been used to constrain the magnitude of the post-1.8 kyr flood (MANVILLE *et alii*, 1999): estimates of peak discharge at the breach fall in the range of 17000 - 30000 m^3/s (Figure 4).

Field evidence for flooding

Evidence of a catastrophic release of 20 km^3 of water from Lake

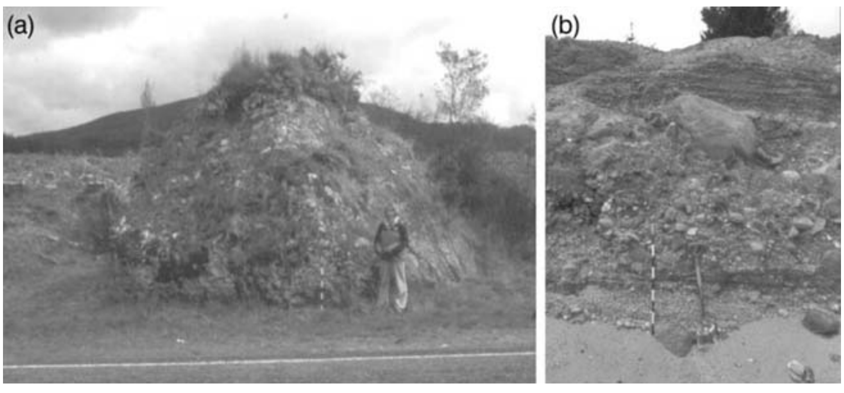

Figure 3. (a) Boulder transported by the post-Oruanui break-out flood from intracaldera Lake Taupo, 75 km downstream of the outlet. (b) Matrix facies of the Smythe's Quarry Boulder Bed, metre stick for scale

Taupo can be traced downstream along the Waikato River for over 220 km (MANVILLE, 2002; MANVILLE *et alii*, 1999). Features associated with a large magnitude flood include: (1) a 12 km long, vertical walled spillway immediately downstream of the outlet that is floored with thin downstream-dipping fans of coarse lithic gravel and boulder lags; (2) streamlined landforms sculpted from older deposits and exhumed river terraces; (3) bouldery fan deposits or expansion bars downstream of valley constrictions; (4) fine-grained slackwater deposits in off-channel areas; (5) valley-wide erosional unconformities; and (6) buried forests in distal areas. Many of these features are also recognised, albeit at a larger scale, in other terrestrial environments affected by major paleofloods (BAKER, 1985; O'CONNOR, 1993).

LAKE TARAWERA

Lake Tarawera covers an area of 41 km^2 within the 64 ka Haroharo caldera in the Okataina Volcanic Centre (NAIRN, 2002), bounded by the western rim of the caldera and the resurgent lava dome complexes of Tarawera and Haroharo to the east. It represents a remnant of a Haroharo intracaldera water body modified by caldera resurgence and dome development (NAIRN, 2002).

KAHAROA FLOOD

The 5 km^3 (magma-equivalent) 700a Kaharoa rhyolite eruption formed much of the Tarawera dome complex and deposited plinian tephra falls over much of the North Island (NAIRN *et alii*, 2001). Primary block-and-ash flows and remobilized pyroclastic debris blocked the outlet to the lake, infilling a narrow channel cut through an older fan structure (Figure 5). Lake Tarawera rose c. 30 m above its pre-eruption level (298 m a.s.l.) before the dam was overtopped and failed catastrophically. Approximately 1.7 km^3 of water was released into the head of the Tarawera valley as the lake level fell by > 40 m, excavating a 300 m wide and 3 km long spillway before overtopping the 70 m high Tarawera Falls (HODGSON & NAIRN, 2000). Flood deposits, including boulders up to 13 m in diameter and giant bars, extend up to 40 km from the lake. Approximately 700 km^2 of the Rangitaiki Plains was resurfaced, and the shoreline advanced by c. 2 km (PULLAR & SELBY, 1971). Peak discharge at the outlet was estimated at c. 1.5 x 10^5 m^3/s, assuming instantaneous breach development, while boulder flow-competence relations further downstream indicate flows in the range of 10^4-10^5 m^3/s (HODGSON & NAIM, 2000).

1904 FLOOD

In the aftermath of the 1886 basaltic Plinian eruption of Tarawera volcano the level of Lake Tarawera rose by 12.8 m, before a rain-triggered break-out in November 1904 dropped it back by 3.3 m, generating a flood that was estimated to peak at c. 780 m^3/s 24 km downstream (WHITE *et alii*, 1997). The post-eruptive rise has been attributed to construction of a small alluvial fan across the outlet channel by flash-floods that remobilised 1886 AD pyroclastic and older material in the Tapahoro gully (HODGSON & NAIM, 2000).

CRATER LAKE, MT. RUAPEHU

The summit of Mt. Ruapehu, New Zealand's largest and most active onshore andesitic stratovolcano, normally hosts a hot acidic Crater Lake at an elevation of c. 2530 m a.s.l., with a volume of c. 9 x 10^6 m^3 (CHRISTENSON & WOOD, 1993). The lake has existed for c. 2000 years, based on the Holocene phreatomagmatic tephra and lahar records preserved on the eastern ring plain (CRONIN & NEALL, 1997; DONOGHUE *et alii*, 1997; LECOINTRE *et alii*, 2004), although it only became known to European settlers in 1879. Primary lahars have accompanied all large historic eruptions due to the explosive ejection

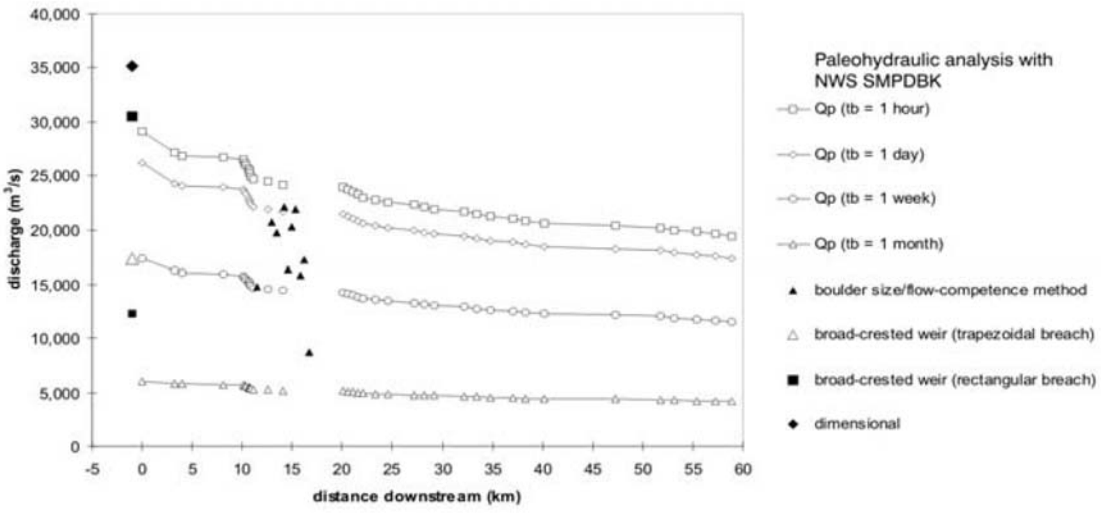

Figure 4 - Summary of palaeohydraulic analyses of the post-1.8 kyr Taupo flood. Breach peak discharge estimates from dimensional and parametric methods, downstream estimates from boulder flow competence data (equation 5), and hydraulic flow routing using US NWS programme SMPDBK (WETMORE & FREAD, 1984)

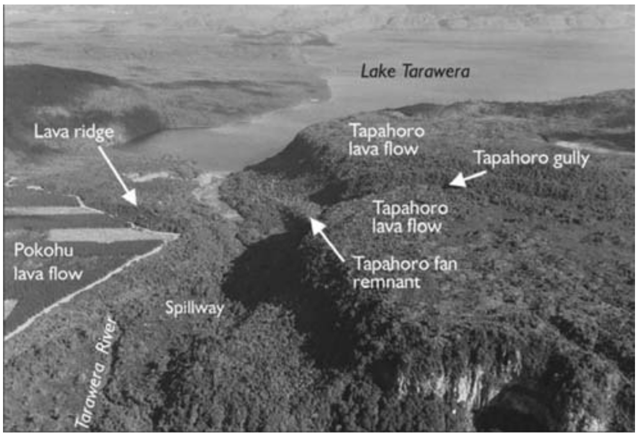

Figure 5 - Oblique aerial view of the outlet area of Lake Tarawera showing geomorphic elements of the pyro/volcaniclastic fan dam emplaced by the 700a Kaharoa eruption and the subsequent break-out flood (Photo DL Homer/GNS)

of Crater Lake water over the rim and volumetric displacement (CRONIN *et alii*, 1997; HEALY *et alii*, 1978; NAIRN *et alii*, 1979), with secondary lahars being triggered by heavy rain on ash deposits (CRONIN *et alii*, 1997; HODGSON & MANVILLE, 1999).

Lahars caused by break-outs from the Crater Lake include the 1953 Tangiwai disaster detailed below, and a flood in the Whangaehu River in January 1925 that was estimated at 430 m^3/s at Tangiwai (STILWELL *et alii*, 1954) and accompanied by an unexplained drop in the level of Crater Lake (O'SHEA, 1954). The cause of the 1861 lahar (CRAWFORD, 1870; TAYLOR, 1861), the largest historical event at Ruapehu, is unknown, but may have been a lake break-out as no eruption was reported.

THE 1953 TANGIWAI LAHAR

On Christmas Eve 1953 the summit Crater Lake of Mount Ruapehu breached an unstable barrier composed of volcanic material deposited during the 1945 eruption sequence (O'SHEA, 1954), and buttressed by the Crater Basin Glacier. Approximately 1.8 million cubic metres of water was released into the headwaters of the Whangaehu River, where it rapidly entrained snow, ice, and volcanic debris to form a lahar that reached the Tangiwai railway bridge c. 39 km downstream in a little over 2 hours (O'SHEA, 1954; STILWELL *et alii*, 1954). The flow critically damaged the bridge moments before the Wellington-Auckland express arrived: unable to stop in time the train fell into the lahar-swollen river with the loss of 151 lives, making it New Zealand's worst volcanic disaster.

Paleohydraulic analysis of the 1953 event indicates that the former summit glaciers did not impede the outflow of hot Crater Lake water, enlargement of an ice tunnel beneath the glaciers keeping pace with growth of a trapezoidal breach at the outlet to generate a peak clear-water discharge in the 300-400 m^3/s range. By c. 10 km downstream, entrainment of particulate material along the Whangaehu Gorge had bulked up the flow to c. 2000 m^3/s. Rapid attenuation on the Whangaehu Fan, calibrated against the 1995 lahar sequence (CRONIN *et alii*, 1997), then reduced the peak flow by c. 60% at 23 km downstream. The crest of the lahar reached Tangiwai railway bridge 39.0 km downstream 2.1-2.3 hours after the onset of breaching as a debris- to hyperconcentrated flow with a peak discharge of c. 590-647 m^3/s (STILWELL *et alii*, 1954). Paleohydraulic analysis has highlighted both the dependence of bulking factors on lahar magnitude and the role of infiltration losses on the Whangaehu Fan, and how these interact to produce a lahar attenuation curve for the Whangaehu River.

LAKE ROTORUA

The 20 km-diameter Rotorua caldera, formed at 220 kyr by eruption of the voluminous Mamaku Ignimbrite (MILNER *et alii*, 2002), is currently partially occupied by the shallow intracaldera Lake Rotorua. Extensive suites of highstand lacustrine terraces and shoreline benches fringe much of the Rotorua basin (GRANGE, 1937; KENNEDY *et alii*, 1978), evidence of a complex history of lake level oscillations influenced by volcanic activity at the adjacent Okataina Volcanic Centre (NAIRN, 2002).

The highest lacustrine terrace (387-414 m a.s.l.) within the Rotorua caldera corresponds to a post-220 ka highstand level associated with filling of the newly created basin in the immediate aftermath of the Mamaku Ignimbrite eruption. Considerable uncertainty surrounds the direction of overflow of this level, but the lake may have breached southwards through the Hemo Gorge, a v-shaped notch in the topographic rim of the caldera, through the Ngakuru Graben into the Waikato River drainage. At some time, however, a northeasterly outlet became established at a lower level. Eruption of the Rotoiti Breccia at 64 kyr from the adjacent Okataina Volcanic Centre (OVC) blocked this drainage route, causing the lake to rise by c. 90 m above its modern level. Lake level stabilised at c. 370-380 m a.s.l. for long enough for well-developed highstand lacustrine terraces to form around much of the basin, before a northerly overspill was superseded by a north-easterly break-out through the Haroharo caldera (formed by the Rotoiti eruption and probably also temporarily host to a post-eruption intracaldera lake) into the Rangitaiki Plains. A deeply incised channel buried beneath the floor of Lake Rotoiti is inferred to represent the path of this flood. Resumption of volcanic activity in the Haroharo caldera during eruption of the Mangaone Sub-group (NAIRN, 2002) again blocked this drainage path at c. 36 kyr, causing Lake Rotorua to rise to 348-353 m a.s.l. (MARX *et alii*, 2003) before breaking out again. Growth of the Haroharo resurgent dome complex at 9-7.5 kyr ultimately blocked this north-easterly route out of the Rotorua caldera, forcing Lake Rotorua to rise to c. 285-290 m a.s.l. before capture of the headwaters of the Kaituna River established the current outlet channel to the north (NAIRN, 2002).

POTENTIAL FUTURE VOLCANIC LAKE BREAK-OUT FLOODS IN THE TVZ

The potential for future dam-break floods from the volcanic lakes in the central North Island depends in large part upon the role of future volcanic activity on their outlets. Lake Tarawera is particularly vulnerable to any resumption of activity at Tarawera, while the lake remains c. 9.5 higher than before the 1886 eruption due to only partial breaching of the volcaniclastic fan dam during the 1904 flood. Aside from this, two crater lakes, at Mt. Ruapehu and White Island, are actively filling following eruptive activity during the past decade.

CRATER LAKE, MT. RUAPEHU

The 1995-96 eruption sequence (BRYAN *et alii*, 1996), the largest eruptive episode since the 1945 eruptions, once again expelled the summit Crater Lake and deposited c. 7 m of unconsolidated tephra on the stable rock sill outlet to the lake basin (Figure 6). Subsequent refilling of the lake has raised the possibility of a lake break-out, with the potential to release up to 1.45 million m^3 of water (HANCOX *et alii*, 2001) in a repeat of the 1953 Tangiwai flood event.

The Department of Conservation, who manage the area as a national park, has installed an automated warning system using, United States Geological Survey acoustic flow monitor technology (LAHUSEN, 1996) to detect lahars in the upper Whangaehu valley. At present (October 2004), Crater Lake is c. 2.7 m below the stable rock sill level of the former

outlet: filling rates are a function of both climate cycles, that govern precipitation and freezing levels, and heating cycles in the lake that control evaporation. Crater Lake is forecast to reach levels where a hazardous-sized lahar, triggered by failure of the tephra dam, is probable sometime between January 2005 and January 2006.

CRATER LAKE, WHITE ISLAND

White Island is the summit of a large, frequently active, andesitic stratovolcano that lies 48 km offshore at the northern end of the TVZ. Intermittent volcanic activity, mostly in the form of small steam and ash eruptions, was first recorded in 1826 and has continued to the present day (HOUGHTON & NAIRN, 1991). The most recent eruption sequence formed a composite explosion crater/collapse pit with a floor below sea level and bordered by a low tuff ring c. 10 m high composed of unconsolidated pyroclastic ejecta. Since

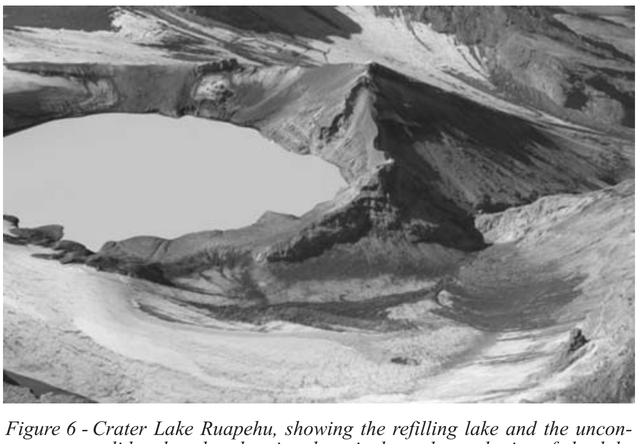

Figure 6 - Crater Lake Ruapehu, showing the refilling lake and the unconsolidated tephra barrier deposited on the rock rim of the lake basin by the 1995-96 eruptions (photo GT Hancox, GNS)

mid-2003 this crater has been filling with precipitation run-off and quenched fumarolic discharges, rising over 20 m between August 2003 and September 2004 to contain a total of c. 1.8 million m^3 of water at 49 °C and with a pH <1. The lake is expected to reach overflow between November 2004 and February 2005, spilling across the floor of an adjacent breached collapse crater filled with unconsolidated volcaniclastic material. Initial overflow is likely to enter preexisting gullies incised in this fill, causing headward erosion until the rim of the main lake is breached. Modelling suggests that rapid enlargement of the breach and outflow channel through the crater and tuff ring could generate an outbreak flood peaking at 10-100 m^3/s, while drawdown of 5 m would release 350000 m^3 of water (SCOTT *et alii*, 2004). Flows at the higher end of the predicted range are likely to inundate much of the Main Crater floor, presenting a serious hazard to any tourist parties on the island at the time.

DISCUSSION AND CONCLUSIONS

The Taupo Volcanic Zone hosts multiple indicators of dam-break flooding (Table 1), including: (i) sources of water, such as a lakes that have experienced a temporary highstand; (ii) dam remnants; and (iii) corroborating evidence for the rapid release of large volumes of water, both from within the lake basin in the form of markers of rapid

drawdown, and along the outlet channel in the form of indicators of large magnitude flows, above what would be expected from a purely meteorological event.

In addition to the 1953 Tangiwai event in New Zealand, historic crater lake rim failures have occurred at Agua in Venezuela in 1541 (NEALL, 1996), Kelut in Indonesia in 1875 (SIMKIN & SIEBERT, 1994), and Pinatubo in the Philippines in 2002 (BORNAS *et alii*, 2003; NEWHALL *et alii*, 2003). Notable historic break-out floods from other classes of volcanic impoundment have followed damming of rivers by pyroclastic flows at El Chichon in 1982 (MACIAS *et alii*, 2004); pyroclastic flows and lahars at Pinatubo in 1991 and subsequently (SCOTT *et alii*, 1996; UMBAL & RODOLFO, 1996); and by a debris avalanche at Mount St. Helens in 19 (MEYER *et alii*, 1986). Such initially clear-water break-outs frequently entrain abundant pyroclastic material from the dam and channel, rapidly increasing in discharge to generate major lahars (volcano-hydrologic mass flows and floods) in downstream catchments, often resulting in major loss of life (NEALL, 1996). Outside of New Zealand prehistoric volcanogenic floods have been recognised from the breaching of debris avalanche deposits at Nevado de Colima (CAPRA & MACIAS, 2002), lava flow dams on the Colorado River (FENTON *et alii*, 2002), pyroclastic flow valley dams on the Rhine (BAALES *et alii*, 2002); PARK & SCHMINCKE, 1997), and failure of the rim of the Aniakchak caldera (WAYTHOMAS *et alii*, 1996).

A plot of peak discharge versus lake volume discharge (Figure 7) shows that break-outs from caldera lakes, in New Zealand and worldwide, are amongst the largest known floods on Earth, only being exceeded by Late Pleistocene glacial lake outbursts and continental-

Lake	Date	Dam Type	Dam height/ lake drop H_o (m)	Excess lake volume V_o (x 10^6 m^3)	Peak discharge Q_p (m^3/s)
Rotorua	Post-220 kyr	Caldera rim	27	4500	23,000*
	Post-64 kyr	Pyroclastic flow	90	9600	50000 – 470000*
	Post-36 kyr	Pyroclastic flow	75	8100	50000 – 300000*
Taupo	Post-26.5 kyr	Caldera rim	80	60000	350000*
	Post-1.8 kyr	Pyroclastic flow	32	20000	35000* 25000 #
Tarawera	Syn-700 yr	Pyroclastic flow/lahar	40	1700	19000- 60000* 75000 #
	1904	Lahar	3.3	139	780**
Crater Lake, Ruapehu	1953	Pyroclastic fall & lava	7.9	1.8	110-1100* 400$

*peak discharge calculated using dimensional method. Range is for breach erosion rates of 1-100 m/h.
#peak discharge calculated using broad-crested weir equations
**gauged 24 km downstream of outlet
$calculated from BREACH model, broad-crested weir equations, and coupled thermal/hydrodynamic model.

Table I - Paleohydraulic data for volcanic lake break-out flood in the Taupo Volcanic Zone

scale meteorological events (BAKER, 2002; O'CONNOR *et alii*, 2002). Break-outs from other volcanogenic impoundments are comparable with floods caused by the breaching of other types of natural dam such as landslides and moraines (COSTA & SCHUSTER, 1988; SCHUSTER, 2000). Jokulhlaups (e.g. BJÖRNSSON, 1975; TOMASSON, 2002), although influenced by sub-glacial volcanism as a mechanism for melting ice, are functionally a class of glacier-lake breakout, as the outflow hydrograph is controlled by the glacier, not the volcano (CLARKE, 2003; WALDER & COSTA, 1996).

Crater lakes appear to be particularly ephemeral, being vulnerable to explosive ejection or drainage through breaching or collapse of the crater rim, while volcanic dams in general appear to predominantly fail by overtopping and surface erosion rather than piping or gravitational slope failure. This is apparently a function of the geometric, granulometric, and mechanical properties of the volcanic dam and channel. Unwelded pyroclastic dams appear more to be more prone to failure than debris avalanche deposits, possibly because they are typically composed of unconsolidated, fine-grained, non-cohesive, and low density (vesiculated) material. The frequency of events in the Taupo Volcanic Zone appears to relate to the intense geological scrutiny of this area rather than any special

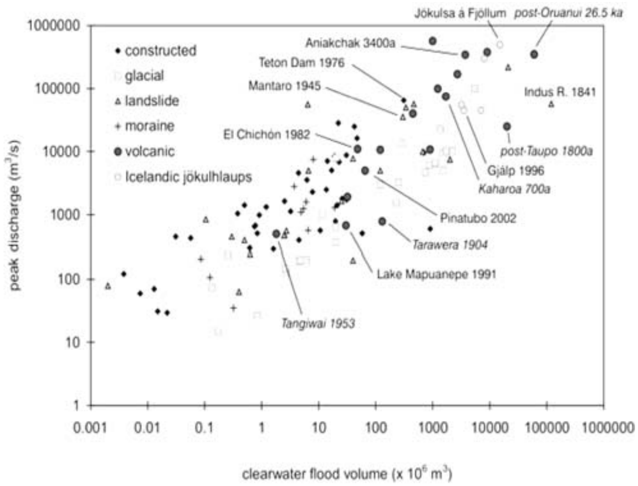

Figure 7 - Comparison plot of break-out floods from volcanic dams, versus landslide and artificial dam breaches. Data compiled from (COSTA, 1988; O'CONNOR et alii, 1993; SCHUSTER, 2000; WALDER & O'CONNOR, 1997)

feature of the actual geological environment. Similar detailed research in other volcanic areas is expected to show that volcanic lake break-out floods are a ubiquitous feature, and hence represent a significant new class of hazard that should be incorporated into hazard assessments.

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