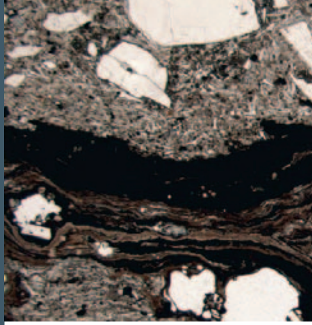

Supereruptions and Supervolcanoes: Processes and Products

Colin J.N. Wilson*

DOI: 10.2113/GSELEMENTS.4.1.29

Pyroclastic deposits and lava flows generated by supereruptions are similar to, but tens of times larger than, those observed in historic eruptions. Physical processes control eruption styles, which then dictate what products are available for sampling and how well the eruption sequence can be determined. These erupted products and their ordering in time permit reconstruction of the parental magma chamber. Supervolcanoes also have smaller eruptions that provide snapshots of magma chamber development in the lead-in to and aftermath of supereruptions. Many aspects of supereruption dynamics, although on a vast scale, can be understood from observations or inferences from smaller historic and prehistoric events. However, the great diversity in the timings of supereruptions and in the eruptive behaviour of supervolcanoes present continuing challenges for research.

KEYWORDS: explosive eruption, ignimbrite, fall deposit, Yellowstone, Long Valley, Taupo

INTRODUCTION

All interpretations of the lead-in to and consequences of supereruptions require data from eruption products to understand how supereruptions operate and to provide context for samples used to reconstruct the magma body. To illustrate this, I briefly summarize how silicic magmas are erupted and their products are emplaced (for a comprehensive treatment, see Sigurdsson et al. 2000). Then, I consider how the eruptive processes control what can be learned about the parental magma chamber. I use three case studies with which I am most familiar, from New Zealand and the United States, to illustrate three points. First, giant eruptions are not intrinsically different from smaller ones and much can be learned about supereruptions from smaller events. Second, supereruptions are highly variable, making the task of actually dealing with an imminent or ongoing one complex. Third, supereruptions occur interspersed with smaller eruptions, and understanding why one particular event should be so large is a continuing challenge.

STYLES OF ERUPTION

Effusive versus Explosive

Almost all controls on how viscous, silicic magmas erupt are related to how the volatile constituents (Bachmann and Bergantz 2008 this issue) dissolved in the magma behave on eruption. If the volatile content is low or decompression slow enough for gases to escape, then an effusive eruption occurs, forming a lava dome or flow (FIG. 1A). Although large lava flows (>100 km^3) are known (e.g. Yellowstone; Christiansen 2001), no purely lava-forming silicic supereruption has been documented. However, if the gases cannot easily escape as the magma decompresses, the growth of bubbles froths and tears apart the magma in a pyroclastic (explosive) eruption. Any addition of water near or at the Earth's surface serves to further increase the vigour of explosive activity as the water is flashed to steam by the hot magma. Pyroclastic activity dominates the evacuation of all supersized magma bodies, and rocks made up of shattered magma dominate supervolcano deposits.

Styles of Explosive Eruption: Fall versus Flow

During explosive eruptions, a mixture of fragmented magma and gases exits the vent at velocities of up to 600 m s^{-1} (>2000 km hr^{-1}). What then happens depends on how this mixture behaves as it enters the atmosphere, mixes with and heats air, and is slowed rapidly (Sparks et al. 1997). If enough air is incorporated, the erupted mixture becomes positively buoyant with respect to the surrounding atmosphere and rises like a giant hot-air balloon to heights where the mixture becomes neutrally buoyant and spreads sideways as an atmospheric plume. Fragments falling from the plume drape the land surface, forming a fall deposit (FIG. 1B). Even small plumes reach the tropopause (~10–17 km), while the largest observed plumes have reached a height of 45 km. There is a theoretical maximum height of ~55 km, which is controlled by the trade off between higher eruption rates (promoting higher plumes) and overloading of the plume by too much fragmental material, triggering collapse. Powerful plumes deposit material over thousands of square kilometres or more, to form what are termed plinian fall deposits.

In contrast, if upward momentum is lost before the eruptive plume becomes buoyant, then it will collapse back as a giant fountain. Material from this fountain will be forced outwards at the ground surface by the pressure gradient around the vent to form a laterally impelled pyroclastic flow (FIG. 1C). Documented examples of flows range between dilute suspensions (<1 vol% solids) and concentrated dispersions (tens of vol% solids). Pyroclastic flows are hot (hundreds of degrees Celsius) and exceedingly mobile, and may travel at speeds exceeding 100 m s^{-1}. Flows deposit material as they travel and/or where they come to rest, to form deposits termed ignimbrite or ash-flow tuff. Small flows can form by several mechanisms, but in supereruptions some form of fountaining is the only viable way of generating large pyroclastic flows.

* School of Geography, Geology and Environmental Science, and Institute of Earth Science and Engineering, University of Auckland PB92019, Auckland Mail Centre, Auckland 1142, New Zealand
E-mail: cjn.wilson@auckland.ac.nz

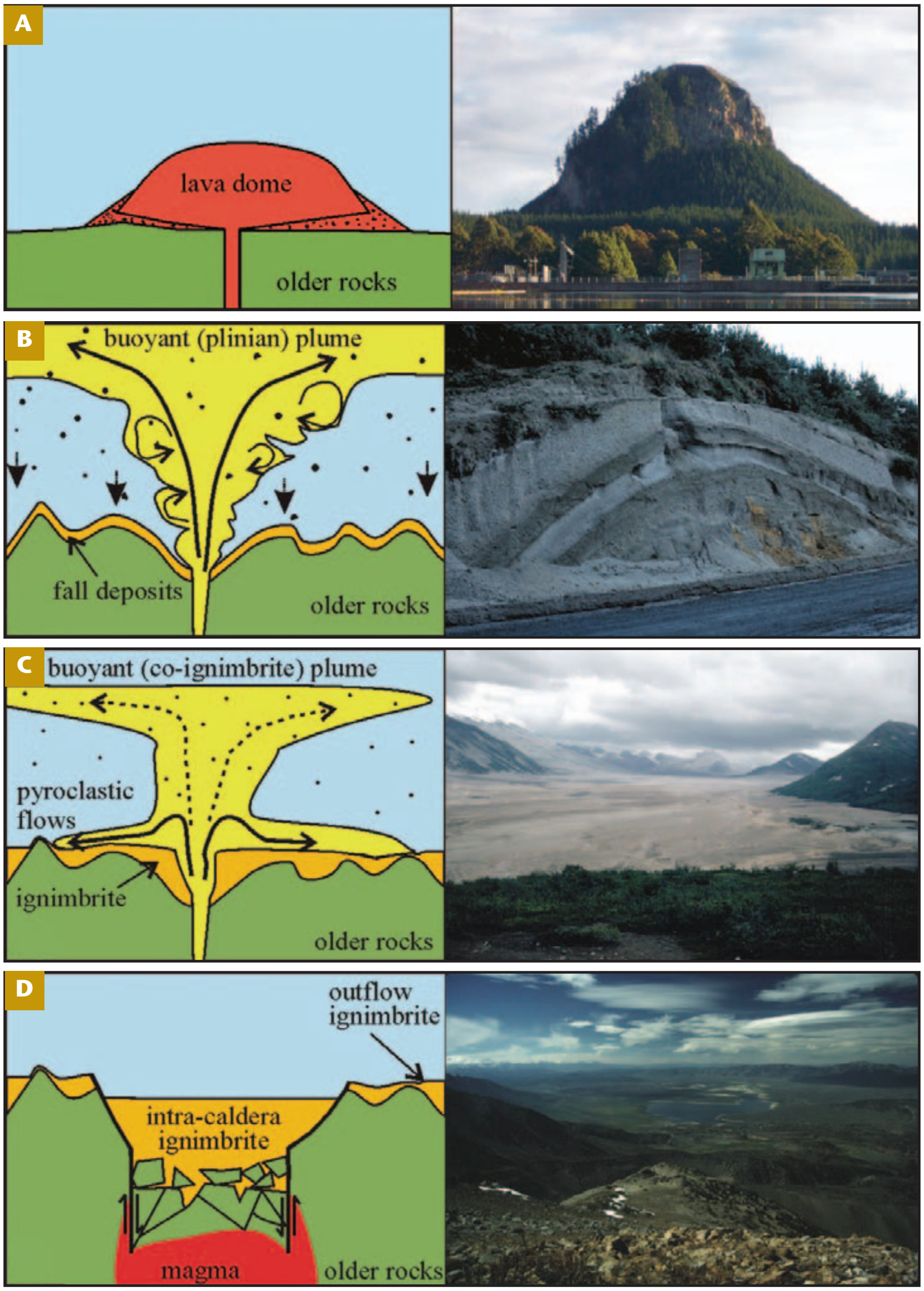

Figure 1 Sketches of the major styles of volcanic eruption discussed in the text. (**A**) Effusive activity yielding a lava dome (Pohaturoa, New Zealand). (**B**) Fall activity from a buoyant plume giving landscape-mantling fall deposits (1.8 ka Taupo eruption, New Zealand). (**C**) Pyroclastic flow activity from a fountaining plume (accompanied by a secondary fine-ash-rich buoyant plume) that yields landscape-filling ignimbrite (1912 eruption, Valley of Ten Thousand Smokes, Alaska). (**D**) Collapse of country rocks into the emptied magma chamber during or after eruption to form a caldera, often expressed at the surface by a topographic depression (0.76 Ma Long Valley eruption, California).

Early theoretical models of explosive eruption processes treated plumes as either buoyant or non-buoyant. Currently, we recognize that high, buoyant eruption plumes (generating plinian fall deposits) can occur simultaneously with large pyroclastic flows. In turn, pyroclastic flows are always accompanied by buoyant (co-ignimbrite) plumes that generate coeval fall deposits. The major differences are that co-ignimbrite plumes are dominated by fine ash and generated above moving flows, not at the primary vent.

Caldera Formation

Calderas are an integral part of supervolcanoes and their eruptions (Lipman 1997; Miller and Wark 2008 this issue). The timing of caldera collapse during an eruption is important in causing a change from a single vent (favouring a buoyant plume and generation of plinian falls) to multiple vents around a ring fracture (favouring generation of pyroclastic flows; Hildreth and Mahood 1986). Additionally, the topographic basin generated by collapse provides a trap for eruption products. Thus in many supereruptions, as much as one-third to one-half of all the eruption products, often reaching kilometres in thickness, may be intra-caldera tuff (Fig. 1d) (Lipman 1997; Mason et al. 2004).

ERUPTION PRODUCTS AND THEIR VALUE FOR INTERPRETING SUPERERUPTIONS

The magma in supereruptions has certain intensive characteristics (e.g. temperature, volatile content, crystal content), all of which may vary with position in the magma chamber (Hildreth 1981; Bachmann and Bergantz 2008). The physical ways in which magma is erupted, however, strongly influence what can be learned about the parental chamber. Eruption products are thus raw materials for investigating processes that are essentially unobservable.

If the magma is erupted as coherent liquid lava, many of its characteristics are preserved. If the magma is disrupted in explosive activity, however, different components can be fractionated, such as crystals versus glass, or incorporated during transport, such as lithic (rock) fragments. Regardless of how the magma is erupted, the resulting products can also undergo changes that complicate physical or petrological studies. Ideally the magma is quenched to a glassy solid, such as in the obsidian carapace to many rhyolite lava domes or the pumices present in most fall deposits. However, non-ideal circumstances usually apply. Because the glassy quenched melt is unstable, prolonged exposure to high temperature will cause devitrification (conversion of glass to crystalline solids). Devitrification is ubiquitous in lava domes and most ignimbrites, affecting the chemical characteristics. Commonly, devitrification is also accompanied by alteration by escaping hot gases, which transport selected components and deposit them elsewhere. After the eruption products have cooled, further alteration of glass, especially in pumice, occurs through secondary hydration from percolating rainwater or groundwater.

Lava Domes

Silicic lavas preserve some intrinsic characteristics of the magma (particularly crystal abundances and species) but, except for obsidian in thin carapaces and pumice, are typically devitrified and/or altered by a gas phase. Domes form at the close of many pyroclastic eruptions from the magmatic dregs that are poorer in gas or are more slowly erupted.

Fall Deposits

Fall deposits originate from a high plume, mantling the landscape (Fig. 1b) and becoming thinner and finer grained away from the source. Most fall deposits are layered, reflecting changes in plume height or in the nature of the dispersing winds. In the absence of added water, the particles that make up fall deposits show a restricted range of grain sizes – that is, they are well sorted (Fig. 2a). Parameters that define fall deposits include their volume and dispersal characteristics, their maximum and average grain size, and the degree of sorting at a given distance from the vent (Walker 1980). These parameters are of great value in reconstructing eruption characteristics, such as plume heights and discharge rates (Carey and Sparks 1986). In supereruptions, fall deposits >10 cm thick cover >10^6 km^2 (Fig. 3a) and are typically unconsolidated because of rapid and strong cooling of the eruption products during transport (compare with ignimbrites, below). Proximal fall deposits may contain metre-scale fragments, while deposits more than a few hundred kilometres from source are dominated by fine glassy ash. In supereruptions, the biggest fall deposits are not generated at the vent but are derived from the buoyant ash clouds above pyroclastic flows. Such fall deposits may exceed 1000 km^3 (bulk volume) and extend for >2000 km (Huff et al. 1992; Ukstins Peate et al. 2003). In some Tertiary and many older examples, co-ignimbrite fall deposits alone provide the main evidence for a supereruption and the nature of its parental magma chamber (Perkins and Nash 2002).

Fall deposits have two major advantages for reconstructing magma chambers, after allowance for fractionation of components during eruption and transportation. First, they almost always are deposited as glassy, fresh material, and second, because they are laid down layer by layer, systematic vertical sampling provides a time sequence of compositional changes through an eruption.

Ignimbrites

Ignimbrites contrast markedly with fall deposits. Small ignimbrites pond in valleys and basins (Fig. 1c), whereas large examples bury entire pre-existing landscapes to produce extensive plateaus. Big ignimbrites can be hundreds of metres thick in outflow sheets that may cover >20,000 km^2 (Fig. 3) and accumulate to >2 km in thickness as intra-caldera material (Fig. 1d). Ignimbrites display a wide range of particle sizes, from ash to blocks (Fig. 2 b, c). Layering in ignimbrites may be due to superposition of deposits from multiple successive pyroclastic flows or, less often, may reflect processes during transport of a single flow. Individual pyroclastic flows, or closely spaced sequences of flows, need not always follow the same pathway, and so a single ignimbrite may be the product of numerous flows emplaced sectorially in overlapping successive or contemporaneous fans (Wilson and Hildreth 1997). Thus samples taken vertically through an ignimbrite at one site may cover only a fraction of the eruption sequence. Few supereruption-related ignimbrites have been studied in enough detail to reconstruct their depositional patterns, and samples from such deposits are not easily linked to an eruptive sequence.

Ignimbrites can show a vast range of textures and structures, due to processes during and after emplacement (Ross and Smith 1961). A common feature is welding (Grunder and Russell 2005), which takes place when the ignimbrite is still so hot on emplacement (>550–600°C) that the glassy fragments deform and merge under the weight of overlying material (before devitrification), forming a dense rock that can be difficult to distinguish from lava in hand specimen. In the extreme case, the deposited mass may revert to a viscous liquid and flow like lava (rheomorphic) for some period before solidifying. Devitrified whole-rock ignimbrite samples may be unsuitable for any petrological purposes other than generalized geochemical labeling (e.g. rhyolite versus dacite) or crystal-specific studies. However, all but a tiny fraction of ignimbrites may consist largely of devitrified material. Some mineral species such as the Fe-Ti oxides are inevitably altered if the host rock is devitrified, but others such as zircon or feldspars may remain pristine and usable. Reconstructing the nature of a supersized magma body from such compromised material is an ongoing challenge.

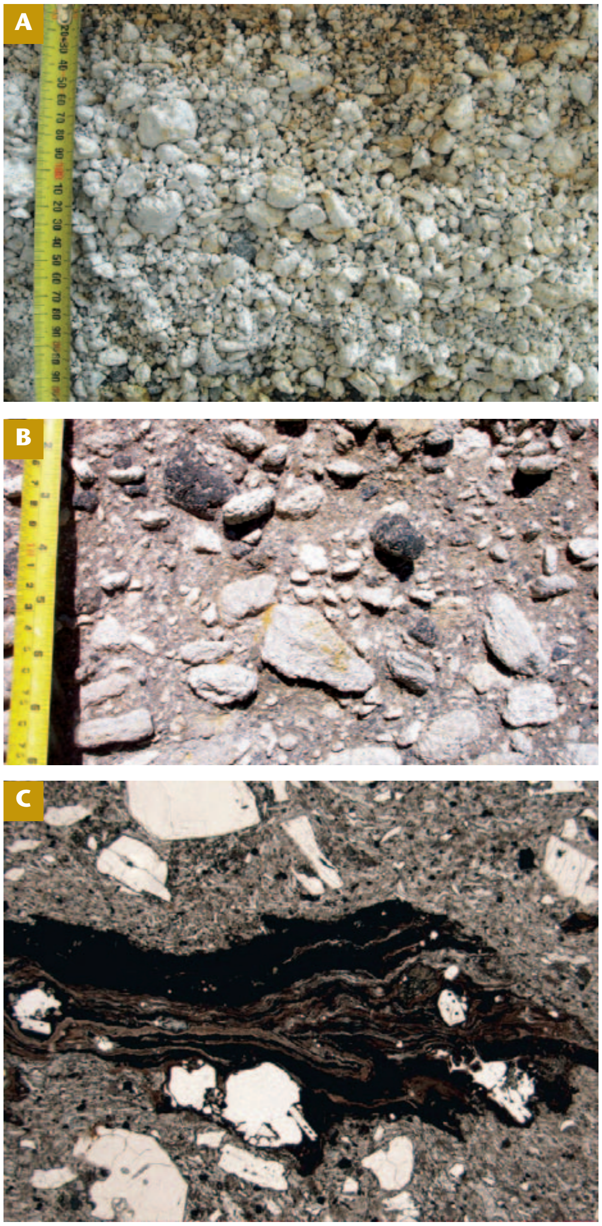

Figure 2 Textures of some pyroclastic deposits. (**A**) Typical proximal plinian fall deposit (Oruanui, New Zealand). Note the uniformity of particle sizes, the particle angularity and the absence of fine-ash matrix, the result of winnowing in the high eruption plume. Darker fragments are lithic material. (**B**) Typical ignimbrite (Whakamaru, New Zealand). Note the poor sorting, the presence of ashy matrix, and the rounding of pale pumice clasts due to abrasion in the pyroclastic flow. (**C**) Photomicrograph of welded ignimbrite (Huckleberry Ridge Tuff, Yellowstone). Field of view is about 9 mm wide. The dark-brown fragment in the centre is a flattened pumice clast, set in a matrix of shards, that is, shattered frothy magma. The streaking within the pumice represents mixing within the magma chamber, and the heterogeneous nature of the matrix is due to mixing of fragments during transport. White fragments are crystals of quartz and feldspar.

THE DYNAMICS OF SUPERERUPTIONS

Very few supereruptions have been investigated in enough detail to be able to reconstruct their dynamics and the original magma chamber. This is due to the many challenges in establishing (a) how such eruptions are initiated and finish, including determining the eruption duration; (b) the distributions and volumes of pyroclastic deposits and lava; (c) how rapidly such vast volumes of magma and gas are erupted; and (d) the links between eruption sequences and magma chamber evacuation. Three case studies with which I am most familiar suggest that there is wide diversity in supereruption timings and style.

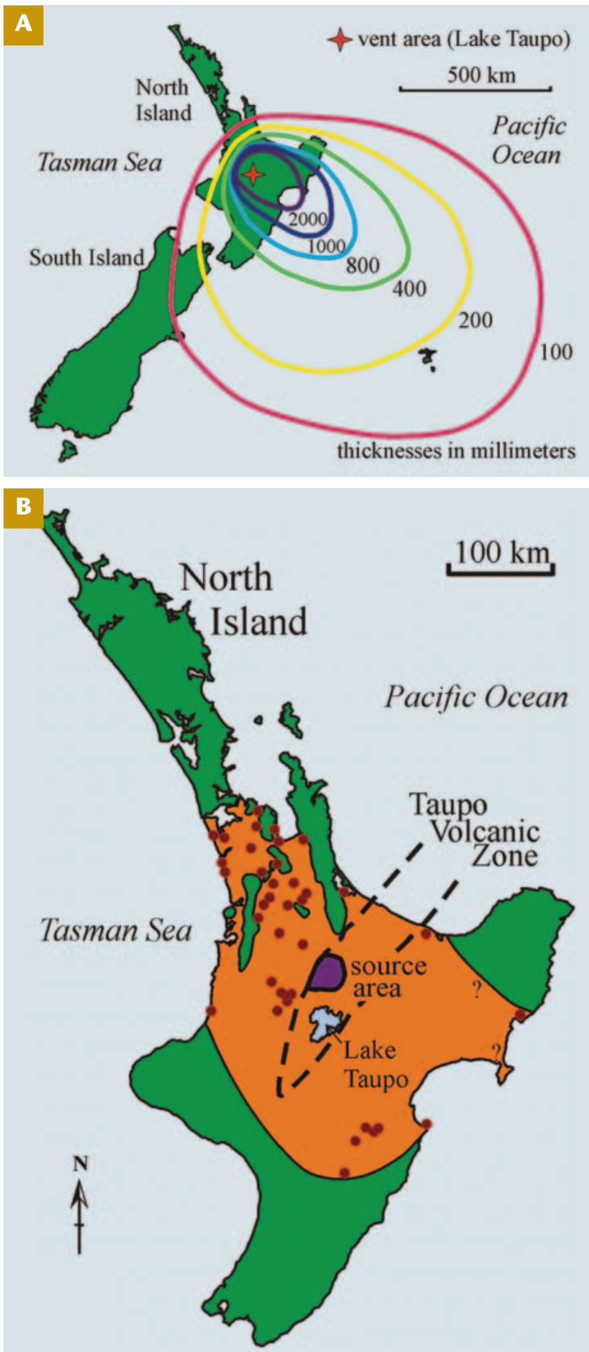

FIGURE 3 Illustrative maps showing the extent of widespread pyroclastic products from two New Zealand supereruptions. (**A**) Isopachs (lines of equal thickness) of fall deposits from the 26.5 ka Oruanui eruption. (**B**) Observed sites (red dots) and inferred original extent (orange) of the ignimbrite from the 1.0 Ma Kidnappers eruption.

The 26.5 ka Oruanui eruption, New Zealand (~530 km^3 of magma) shows evidence, in the form of erosion intervals and/or reworked horizons, for spasmodic activity, including a hiatus of weeks and other shorter breaks (Wilson 2001). The eruption was a series of large-scale outbreaks of increasing vigour, daisy-chained into a single geological event. In contrast, the 0.76 Ma Bishop Tuff eruption (~600 km^3 of magma) at Long Valley (California, United States) displays evidence for only one short time break, and most of the eruption volume may have been emplaced in only about six days (Wilson and Hildreth 1997). The 2.06 Ma Huckleberry Ridge Tuff eruption (~2500 km^3 of magma; Christiansen 2001) at Yellowstone shows evidence for time breaks, possibly of months or more, at several stages (Wilson unpublished data). In the lowest fall deposits, the presence of horizons of reworked material suggests that normal weather-related processes (including wind, hail deposition, rainfall and formation of shallow-water ponds) affected the deposit, implying breaks in deposition. Between the three major ignimbrite units, there is local evidence for cooling, including variations in welding intensity, which indicates additional time breaks.

All three eruptions show evidence suggesting that plinian fall deposits were partially to completely synchronous with pyroclastic flows. The Oruanui and Huckleberry Ridge eruptions are inferred to have lasted for months or years. The length of these eruptions would make it difficult to decide when a future eruption had actually finished and make complex the foreseeing of the climactic stage(s) that would have the greatest impact (see Self and Blake 2008 this issue). In the Oruanui and Bishop Tuff examples, the eruptions commenced at single vents. Caldera collapse and the development of multiple vents occurred part way through each eruption, so that about one-third of the eruption products was confined to the caldera.

SMALLER ERUPTIONS AND SUPERERUPTIONS

Supereruptions almost never occur in isolation in young (Quaternary) volcanoes but are bracketed by eruptions of smaller volume. In Tertiary and older supereruptions, the presence of precursor or subsequent smaller eruptions has only rarely been established (e.g. Bachmann et al. 2000). At present it is not known whether this contrast reflects different behaviour or is due to poor preservation of eruption products. The three volcanoes considered in the previous section show a variety of behaviours prior to and following their supereruptions. This reflects differences in features such as the continuity of precursor and subsequent eruptive activity, the compositional relations between precursor eruptions and the supereruption products, and the rates at which the supersized bodies of rhyolite accumulated in the crust.

At Yellowstone, only one small lava eruption occurred shortly before the Huckleberry Ridge Tuff was deposited, and no eruptions occurred afterwards for 200 ky or so (Christiansen 2001). A younger supereruption at Yellowstone (the 0.64 Ma Lava Creek Tuff) was also shortly preceded by lava eruptions and followed by a gap in activity of ~120 ky. No small-scale pyroclastic eruptions are recorded either before or after these supereruptions, but the chances of their preservation are remote. The Yellowstone system is interpreted to have generated its supereruptions as part of large-scale cycles of activity, mostly involving eruptions of 1–100 km^3 in volume with repose periods in the 10^4 to 10^5-year range, but available data do not permit growth rates of pre-eruptive magma bodies for the supereruptions to be assessed. Thus an ongoing challenge in monitoring Yellowstone today (Lowenstern and Hurwitz 2008 this issue) is determining whether its magma system could produce a supereruption now.

At Long Valley, the Bishop Tuff supereruption at 0.76 Ma represented the climax of only one (the second) of six distinct magmatic systems active since ~3.5 Ma (Hildreth 2004). The unified magma body which fed the Bishop Tuff eruption began growing about 160 ky before the climactic event (Simon and Reid 2005; Reid 2008 this issue). The latest precursor activity (of lava) occurred ~30 ky before the main eruption, and subsequent eruptions commenced at a date that is identical to the Bishop Tuff event within error. The precursor and immediately subsequent rhyolites show similar compositional characteristics to those of the Bishop Tuff itself. Assuming a 160 ky incubation period for the Bishop Tuff magma, the erupted magma body grew at an average rate of ~4 km^3 ky^{-1} (Hildreth and Wilson 2007).

Taupo shows marked contrasts to Long Valley, despite similar volumes for the climactic supereruptions. At or near Taupo volcano, 11 eruptions are recorded in the ~35 ky preceding the Oruanui eruption (Fig. 4a). Unlike at Long Valley, the precursor eruption compositions are diverse and include at least two distinct rhyolites that erupt in alternation (Wilson et al. 2006). Thus shortly prior to the Oruanui event, at least two magmatic systems were active contemporaneously, but the vent sites were geographically separated (Fig. 4b). The Oruanui magma type is recorded for ≤40 ky prior to the climactic eruption. After the Oruanui event, there was a break of about 6 ky, followed by a sequence of 28 eruptions up to the present day. None of these eruptions show any compositional linkage to the Oruanui magma (Sutton et al. 2000), despite being erupted from geographic areas of the volcano that overlap with the footprint of the Oruanui magma body (Fig. 4c). Rhyolite magma bodies at Taupo have grown at rates as high as 10–40 km^3 ky^{-1} during several periods over the past 60 ky.

CONCLUSIONS

Can we answer the fundamental question: do supereruptions represent 'normal' volcanic processes acting on bloated volumes, or are they special? In terms of physical processes, what is currently known about supereruptions represents a logical extension of activity observed and inferred from smaller eruptions of the same composition: only the volume of magma released in single eruptions of supervolcanoes is exceptional. At present, there are no grounds, theoretical or field-based, for supposing that supereruptions will involve novel processes that do not also apply to (and may be discovered from) less-than-supersized eruptions. For example, the understanding of the behaviour of large plumes from eruptions generating pyroclastic flows was widely advanced by satellite images of the 1991 Pinatubo eruption (e.g. Koyaguchi and Tokuno 1993), and concepts being developed for supereruption plumes stem from that work (Baines and Sparks 2005).

Two other conclusions about supereruptions are equally clear. First, knowledge based on field-focused studies of the physical processes operating in supereruptions and their host volcanoes lags substantially behind the understanding of historic eruptions and small to moderate-sized prehistoric events. There are several reasons for this, including (1) the scale of the subject, (2) the age and consequent state of erosion of many supereruption deposits, and (3) the difficulty in quantifying huge masses of devitrified and altered rock. Second, there is great diversity in the natures of individual supereruptions and their host volcanoes, particularly in the overall timing of such eruptions, and in the frequency and causes of associated smaller-scale activity. The latter point is a rich subject for future work; for example, why should one small eruption from a mature chamber be followed by an eruption three orders of magnitude larger, as in the Bishop Tuff and Oruanui events? Supereruptions and their host volcanoes represent a fertile area for field-focused studies for many decades to come.

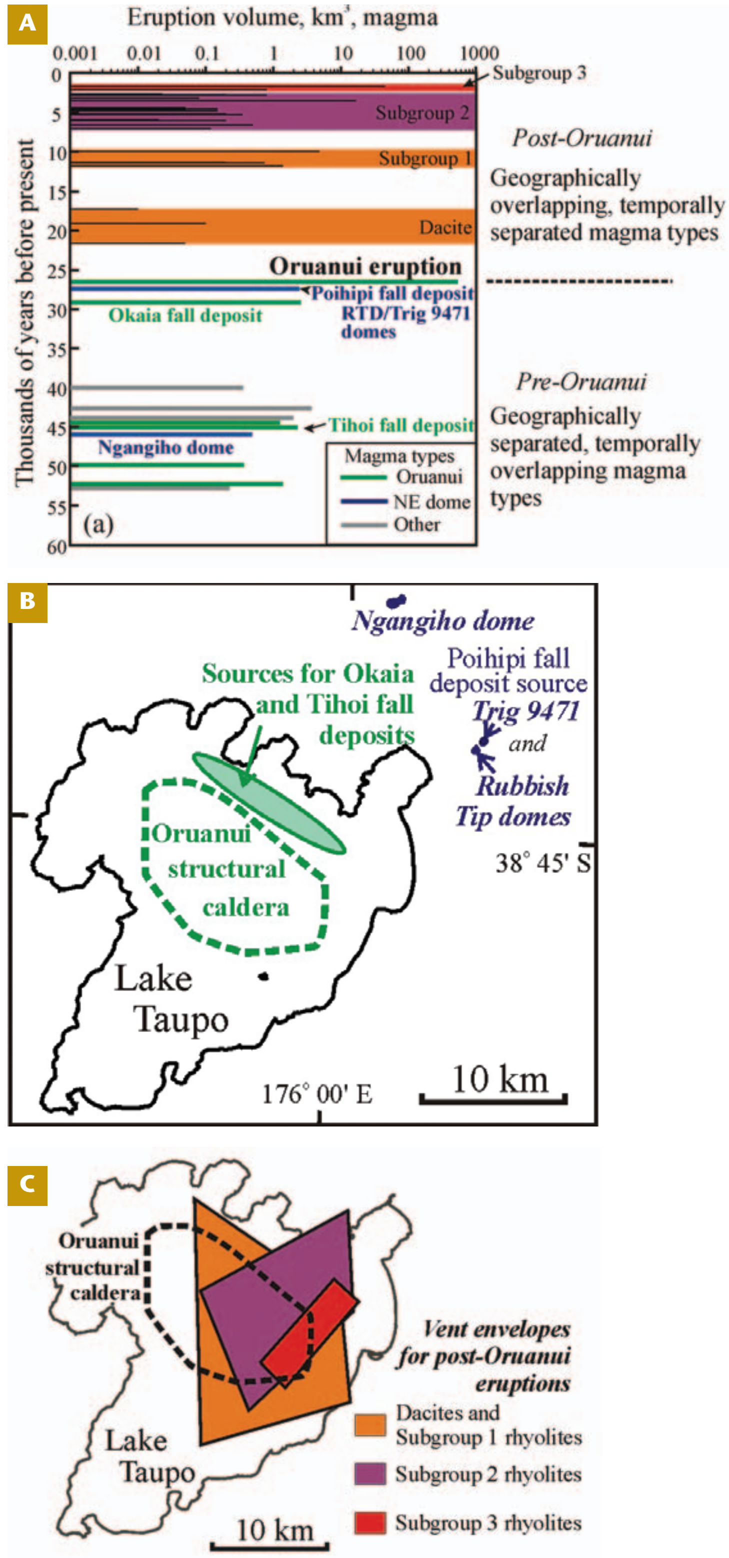

Figure 4 The evolving magmatic system at a young supervolcano: Taupo, New Zealand (for location of Lake Taupo see Fig. 3b). (**A**) Summary time–volume plot for young eruptions from Taupo (Sutton et al. 2000; Wilson et al. 2006). The Oruanui supereruption accompanied a change in the geographical and temporal positions of compositionally distinctive magma groups at the volcano. Colours in this panel are keyed to the maps in panels (**B**) and (**C**). Prior to the Oruanui eruption, contrasting magma groups were present at the same time, as evidenced by intercalation of their eruption products, but were separated geographically (**B**). In contrast, post-Oruanui eruptions vented within overlapping geographic areas, but magmas with similar compositions (subgroups) were separated in time so that compositions change step-wise through the eruption sequence (**C**).

ACKNOWLEDGMENTS

This work was supported by the Royal Society of London, UK NERC, the New Zealand FRST and the Marsden Fund of the Royal Society of New Zealand. For discussions and collaborations in this work I thank Fred Anderson, Steve Blake, Bruce Charlier, Bob Christiansen, Darren Gravley, Wes Hildreth, Bruce Houghton, Yang Liu, Steve Self, Steve Sparks and the late George Walker. Comments from Peter Lipman, Ian Parsons, Mary Roden-Tice, Ashley Bromley, Tenley Banik and the guest editors are appreciated.

REFERENCES

Bachmann O, Bergantz GW (2008) The magma reservoirs that feed supereruptions. Elements 4: 17-21

Bachmann O, Dungan MA, Lipman PW (2000) Voluminous lava-like precursor to a major ash-flow tuff: low-column pyroclastic eruption of the Pagosa Peak Dacite, San Juan volcanic field, Colorado. Journal of Volcanology and Geothermal Research 98: 153-171

Baines PG, Sparks RSJ (2005) Dynamics of giant volcanic ash clouds from supervolcanic eruptions. Geophysical Research Letters 32 (24): L24808, doi: 10.129/2005GL024597

Carey S, Sparks RSJ (1986) Quantitative models of the fallout and dispersal of tephra from volcanic eruption columns. Bulletin of Volcanology 48: 109-125

Christiansen RL (2001) The Quaternary and Pliocene Yellowstone Plateau Volcanic Field of Wyoming, Idaho, and Montana. USGS Professional Paper 729-G: 120 pp

Grunder A, Russell JK (eds) (2005) Welding Processes in Volcanology. Journal of Volcanology and Geothermal Research 142: 1-191

Hildreth W (1981) Gradients in silicic magma chambers: implications for lithospheric magmatism. Journal of Geophysical Research 86: 10153-10192

Hildreth W (2004) Volcanological perspectives on Long Valley, Mammoth Mountain, and Mono Craters: several contiguous but discrete systems. Journal of Volcanology and Geothermal Research 136: 169-198

Hildreth W, Mahood GA (1986) Ring-fracture eruption of the Bishop Tuff. Geological Society of America Bulletin 97: 396-403

Hildreth W, Wilson CJN (2007) Compositional zoning of the Bishop Tuff. Journal of Petrology 48: 951-999

Huff WD, Bergstrom SM, Kolata DR (1992) Gigantic Ordovician volcanic ash fall in North America and Europe: biological, tectonomagmatic and event-stratigraphic significance. Geology 20: 875-878

Koyaguchi T, Tokuno M (1993) Origin of the giant eruption cloud of Pinatubo, June 15, 1991. Journal of Volcanology and Geothermal Research 55: 85-96

Lipman PW (1997) Subsidence of ash-flow calderas: relation to caldera size and magma-chamber geometry. Bulletin of Volcanology 59: 198-218

Lowenstern JB, Hurwitz S (2008) Monitoring a supervolcano in repose: Heat and volatile flux at the Yellowstone Caldera. Elements 4: 35-40

Mason BG, Pyle DM, Oppenheimer C (2004) The size and frequency of the largest explosive eruptions on Earth. Bulletin of Volcanology 66: 735-748

Miller CF, Wark DA (2008) Supervolcanoes and their supereruptions. Elements 4: 11-16

Perkins ME, Nash BP (2002) Explosive silicic volcanism of the Yellowstone hotspot: The ash fall tuff record. Geological Society of America Bulletin 114: 367-381

Reid MR (2008) How long does it take to supersize an eruption? Elements 4: 23-28

Ross CS, Smith RL (1961) Ash-Flow Tuffs: Their Origin, Geologic Relations and Identification. USGS Professional Paper 366, 81 pp

Self S, Blake S (2008) Consequences of explosive supereruptions. Elements 4: 41-46

Sigurdsson H, Houghton BF, McNutt SR, Rymer H, Stix J (eds) (2000) Encyclopedia of Volcanoes. Academic Press, San Diego, 1456 pp

Simon JI, Reid MR (2005) The pace of rhyolite differentiation and storage in an 'archetypical' silicic magma system, Long Valley, California. Earth and Planetary Science Letters 235: 123-140

Sparks RSJ, Bursik MI, Carey SN, Gilbert JS, Glaze LS, Sigurdsson H, Woods AW (1997) Volcanic Plumes. John Wiley & Sons, Chichester, UK, 590 pp

Sutton AN, Blake S, Wilson CJN, Charlier BLA (2000) Late Quaternary evolution of a hyperactive rhyolite magmatic system: Taupo volcanic centre, New Zealand. Journal of the Geological Society 157: 537-552

Ukstins Peate I, Baker JA, Kent AJR, Al-Kadasi M, Al-Subbary A, Ayalew D, Menzies M (2003) Correlation of Indian Ocean tephra to individual Oligocene silicic eruptions from Afro-Arabian flood volcanism. Earth and Planetary Science Letters 211: 311-327

Walker GPL (1980) The Taupo Pumice: product of the most powerful known (ultraplinian) eruption? Journal of Volcanology and Geothermal Research 8: 69-94

Wilson CJN (2001) The 26.5 ka Oruanui eruption, New Zealand: an introduction and overview. Journal of Volcanology and Geothermal Research 112: 133-174

Wilson CJN, Hildreth W (1997) The Bishop Tuff: New insights from eruptive stratigraphy. Journal of Geology 105: 407-439

Wilson CJN, Blake S, Charlier BLA, Sutton AN (2006) The 26.5 ka Oruanui eruption, Taupo volcano, New Zealand: Development, characteristics and evacuation of a large rhyolitic magma body. Journal of Petrology 47: 35-69