GEOSPHERE

GEOSPHERE; v. 14, no. 2

doi:10.1130/GES01513.1

7 figures; 3 tables

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CITATION: Newhall, C., Self, S., and Robock, A., 2018, Anticipating future Volcanic Explosivity Index (VEI) 7 eruptions and their chilling impacts: Geosphere, v. 14, no. 2, doi:10.1130/GES01513.1.

Science Editor: Shanaka de Silva
Guest Associate Editor: Robert Stern

Received 8 February 2017
Revision received 21 June 2017
Accepted 22 December 2017

Anticipating future Volcanic Explosivity Index (VEI) 7 eruptions and their chilling impacts

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ABSTRACT

Worst-case or high-end subduction-related earthquakes and tsunamis of 2004 and 2011 are painfully fresh in our memories. High-end subduction-related volcanic eruptions have not occurred in recent memory, so we review historical and geologic evidence about such eruptions that will surely recur within coming centuries. Specifically, we focus on Volcanic Explosivity Index (VEI) 7 eruptions, which occur 1–2 times per thousand years.

A variety of environmental changes followed the VEI 7 eruption of Rinjani (Samalas), Indonesia, in CE 1257 and several more eruptions of VEI 6 or 7 that occurred in the succeeding few centuries. The Rinjani eruption and its impacts are relatively well documented, modeled, and, for the purposes of attribution, uncomplicated by antecedent eruptions. It seems likely that the Rinjani eruption triggered the onset of the Little Ice Age, and subsequent large eruptions sustained it. Although climatic effects of eruptions like Pinatubo (Philippines) and even Tambora (Indonesia) lasted only a few years, it seems that coupling of oceans, sea ice, and atmosphere after larger eruptions can force decade- to century-long cooling, decreased precipitation, and major impacts on crops.

The next VEI 7 will affect a world very different from that of CE 1257. Today, populations within 100 km of candidate volcanoes range from fewer than 1000 people in remote areas to between 20 and 30 million people near several candidates in Indonesia and the Philippines. If a VEI 7 eruption occurs, those populations will be at dire risk, and eruptions in some locations could destabilize financial centers, national economies, and even peace between nations. Distal effects on air travel, the global positioning system, and climate will be felt by a high-technology, globally interdependent world.

We suggest and apply criteria to identify candidates for future VEI 7 eruptions, and discuss likely challenges for short-range forecasting of such events. Preparation for such low-probability but high-consequence events is difficult to imagine, yet some modest early measures can be considered. Volcanologists should refine geologic histories and ensure at least baseline monitoring of candidate volcanoes, and consider how they will judge the likelihood that an impending eruption will be of VEI 7. Forward-thinking governments and industries would be wise to consider how a proximal or distal VEI 7 eruption will affect their jurisdictions, and what responses make the most economic and sociopolitical sense.

INTRODUCTION

Explosive volcanic eruptions occur in a wide range of sizes, and the modern world has not yet seen an eruption of either Volcanic Explosivity Index (VEI) 7 or 8, the highest 2 orders of magnitude known from written and geologic history (Newhall and Self, 1982; Mason et al., 2004; Global Volcanism Program, 2013). Most, although not all, of these very large eruptions have occurred above subduction zones.

The latest VEI 7 event, at Tambora, Indonesia, in CE 1815, was at the low end of VEI 7 eruptions. Its global effects were widely reported (Stommel and Stommel, 1983; Stothers, 1984; Harington, 1992; de Jong Boers, 1995; Oppenheimer, 2003a; Raible et al., 2016). In this paper we list all known Holocene VEI 7 eruptions (Table 1) and discuss effects from that of Rinjani (Samalas) Volcano in Indonesia in CE 1257. We then discuss effects more generally, including how a VEI 7 today will have very different effects than those of Rinjani (CE 1257) or Tambora, on a world that now has much higher populations, higher technology, and more global interdependence than in those earlier years. Our emphasis on VEI 7 eruptions is to call attention to this class of eruptions that are considerably more likely than a VEI 8 supereruption, yet still have remarkable global impacts.

For long-range forecasting we suggest characteristics of volcanoes that might produce future VEI 7 eruptions. Experiences at Pinatubo (Philippines), Rabaul (Papua New Guinea), and Campi Flegrei (Italy) illustrate the challenges of short-range forecasting at those volcanoes. Neither forecasting nor mitigation will be easy, and there are special challenges for low-probability, high-consequence events. No volcano observatory or community will stay on high alert for events like VEI 7 eruptions that occur globally only once or twice per millennium and locally far less often. However, because the next VEI 7 eruption could occur at any time, and affect the entire world, we recommend some modest new monitoring and response planning.

SIZES OF EXPLOSIVE VOLCANIC ERUPTIONS

An imperfect but handy measure of explosive magnitude is the VEI (Newhall and Self, 1982), a broadly logarithmic scale similar to Richter magnitude for earthquakes and to a more precise but also more data-demanding magni-

TABLE 1. KNOWN OR PROBABLE VEI 7 ERUPTIONS OF PAST 10,000 YEARS

Volcano	Eruption year	Caldera diameter (km)	Volume erupted (bulk or DRE)*	VEI (Global Volcanism Program, 2013)	M† (LaMEVE database of Crosweller et al., 2012 and online updates)	References
Tambora, Indonesia	CE 1815	6	41 ± 4 km^3 DRE (Kandlbauer and Sparks, 2014).	7	7.0	Sigurdsson and Carey, 1989; Self et al., 2004; Kandlbauer and Sparks, 2014; Gertisser and Self, 2015; Raible et al., 2016
Kuwae, Vanuatu	CE 1450s	6 × 12?	Volume of caldera (and inferred DRE of magma erupted): 32–39 km^3 (Monzier et al., 1994) and 23–56 km^3, rounded to 30–60 km^3 DRE (Witter and Self, 2007). Németh et al. (2007) question the size.	n/a	6.8	Monzier et al., 1994; Robin et al., 1994; Witter and Self, 2007; Nemeth et al., 2007; Plummer et al., 2012
Rinjani (Samalas) Indonesia	CE 1257	6 × 8.5	33–40 km^3 DRE (Vidal et al., 2015)	7	7.2	Nasution et al., 2004b; Lavigne et al., 2013; Vidal et al., 2015
Baegdusan (Korea) = Changbaishan (China)	CE 946	5	96 ± 19 km^3 bulk volume	7	7.4	Yin et al., 2012; Xu et al., 2012, 2013; Wei et al., 2013; Oppenheimer et al., 2017
Dakataua, Papua New Guinea	ca. CE 605–675	10 × 13	75 km^3 collapse volume (Lowder and Carmichael, 1970)	6	7.4	Lowder and Carmichael, 1970; Torrence, 2012
Ilopango, El Salvador	ca. CE 260	8 × 11	70 km^3 bulk volume, for the fall deposit alone (Kutterolf et al., 2008).	6	6.7	Hart and Steen-McIntyre, 1983; Dull et al., 2001; Kutterolf et al., 2008
Taupo, New Zealand	ca. CE 233	~8 × 12	~100 km^3 bulk volume including coignimbrite ash	6	6.9	Wilson, 1985; Wilson and Walker, 1985; Davy and Caldwell, 1998; Potter et al., 2015
Ambrym, Vanuatu	ca. CE 50 ± 100	13	Based on caldera reconstruction by McCall et al. (1970), we estimate an approximate volume of collapse of 70 km^3; minimum estimate based on deposits (Robin et al., 1993): 70 ± 10 km^3 bulk volume and >20 km^3 DRE	6	6.8	McCall et al., 1970; Robin et al., 1993; Németh and Cronin, 2008; Sheehan, 2016
Iwo-Jima, Japan	ca. 730–850 BCE	10–12	Not known; inclusion here based on caldera diameter but off-island deposit not documented	—	—	Kaizuka, 1992; Newhall et al., 1998; Nagai and Kobayashi, 2015
Santorini (Minoan), Greece	ca. 1610 ± 14 BCE	6–7	~117–129 km^3 bulk (Johnston et al., 2014)	7	6.5	Druitt et al., 1999; Friedrich et al., 2006; Druitt, 2014; Johnston et al., 2014; Bruins, 2016
Aniakchak, Alaska, USA	1620–1430 BCE (Blackford et al., 2014); 1790–1640 BCE (Bacon et al., 2014)	10	>50 km^3 bulk (Miller and Smith, 1987)	6	6.9	Miller and Smith, 1987; Blackford et al., 2014; Bacon et al., 2014
Cerro Blanco-Robledo, Argentina	ca. 2300 ± 60 BCE	6 × 10	110 km^3 bulk	7	—	Fernández-Turiel et al., 2013; Baez et al., 2015
Kikai, Japan	ca. 5280 BCE	19	200 km^3 bulk	7	7.2	Maeno and Taniguchi, 2007
Crater Lake (Mazama), Oregon, USA	5677 ± 150 BCE (Zdanowicz et al., 1999)	7	150 km^3 bulk; 75 km^3 DRE	7	6.8	Bacon, 1983; Zdanowicz et al., 1999
Kurile Lake, Kamchatka, Russia	6437 ± 23 BCE	8 × 14	155 km^3 bulk	7	7.2	Braitseva et al., 1995, Ponomareva et al., 2004; Plechov et al., 2010
Fisher caldera	ca. 8090 BCE	11 × 18	~100 km^3 DRE based on caldera size	6	6.7	Miller and Smith, 1977; Stelling et al., 2005; Gardner et al., 2007
L'vinaya Past	ca. 8700 BCE	8	75–80 km^3 bulk	6	7.0	Melekestsev et al., 1988; Braitseva et al., 1995

Note: VEI—Volcanic Explosivity Index. Dashes under VEI and M indicate no data.

*Bulk volume—volume of deposit, including pore space between grains; DRE—dense rock equivalent, or roughly the volume of magma erupted (pore free). More eruptions of VEI 7 in the same period are also likely, judging from dimensions of other young calderas not included here, and the rapid rate with which deposits are eroded (>60% of the original 1991 Pinatubo deposit was reworked within a decade). Dates of eruptions are in calendar years; those originally from ^{14}C are calibrated.

†M—eruption magnitude, using scale of Pyle (1995, 2000).

tude scale for volcanic eruptions proposed by Pyle (1995). The main criterion for estimating VEI is bulk volume of pyroclastic deposits, but other criteria can be used when the volume is not well known. The VEI scale is shown graphically in Figure 1.

Explosive eruptions of VEI 3 and higher exhibit a power-law VEI frequency relation; eruptions of each higher value on the VEI scale (i.e., each step of 10× larger in size) are ~6–7× less frequent than those of the preceding lower VEI (Simkin, 1993; Deligne et al., 2010). Small eruptions occur every day somewhere in the world, but the largest eruptions have global return periods of 1000, 10,000, or even >100,000 yr. Since the mid-1800s, the largest we have seen anywhere on Earth are a few low-end VEI 6 eruptions, including Krakatau (Indonesia) in 1883, Katmai (Alaska) in 1912, and Pinatubo in 1991.

VEI 7 eruptions are of the next larger order of magnitude, producing at least 100 km^3 of ash and pumice deposit (bulk volume) and large masses of sulfur, carbon dioxide, and other volcanic gases that will circle Earth many times. Eruptions of VEI 7 of the past 10 k.y., and a few that are classed as VEI 6 based on minimum estimates of deposit volume and might later be upgraded

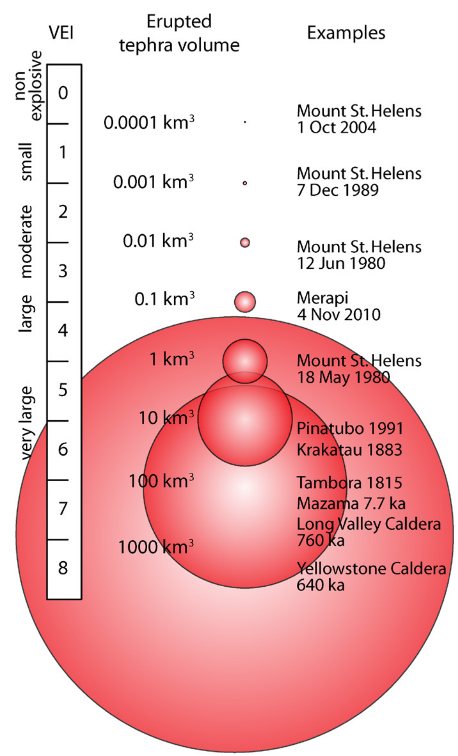

Figure 1. The Volcanic Explosivity Index (VEI) scale of explosive magnitude, based mainly on volumes (in km^3) of pyroclastic deposits (1 km^3 = 1,000,000,000 m^3). Figure courtesy of Wendy Stovall, U.S. Geological Survey.

to VEI 7, are shown in Table 1. Where products were eroded from land or deposited directly in the sea, volume estimates have high uncertainties and are probably too low.

VEI 7 eruptions of the Holocene include the low-end VEI 7 eruption of Tambora in 1815, a larger eruption of Rinjani in 1257, and even larger eruptions of Kikai (Japan; ca. 7.2 ka) and Crater Lake (Oregon, USA; ca. 7.7 ka). Table 1 and similar compilations by Decker (1990), Simkin (1993), Mason et al. (2004), Deligne et al. (2010), Brown et al. (2014), and more, all point to a minimum global recurrence frequency of ~1–2 times per millennium. If one corrects for incompleteness in historical and geologic records (Kiyosugi et al., 2015; Rougier et al., 2016), the correct value is probably closer to 2 per millennium. Recurrence frequencies at individual volcanoes are mostly between 1 per 10 k.y. and 1 per 100 k.y.

All eruptions in Table 1 occurred at arc/subduction zone volcanoes, and the next VEI 7 eruption will very likely come from such a setting. In a recent assessment, more than 90% of the world's volcanic risk is concentrated in 5 countries with subduction zone volcanoes (Loughlin et al., 2015).

In this paper we note VEI 8 eruptions only in passing, e.g., those of Taupo (New Zealand) 26.5 ka (Self, 1983; Wilson, 2001), Toba (Indonesia) ca. 74–72 ka (Chesner and Rose, 1991; Chesner, 2012), and Yellowstone (western USA) ca. 640 ka (Christiansen, 2001). Such eruptions have global return times of tens to hundreds of thousands of years (Decker, 1990; Mason et al., 2004). As suggested by Deligne et al. (2010) and others, processes of VEI 8 eruptions may differ significantly from those leading to the more common VEI 7 events, and their consequences are also in debate (cf. Williams, 2012, and references therein).

NOTE REGARDING ERUPTION SCALES

Since the VEI was introduced (Newhall and Self, 1982), it has become widely used despite some admitted limitations. The following are three notable limitations.

1. The VEI scale combines the concepts of magnitude and intensity as initially defined and distinguished by Walker (1980) and later adopted by Pyle (1995). Prolonged dome-building eruptions can be practically nonexplosive yet produce large volumes of dome-collapse pyroclastic flows, thereby earning a misleadingly high VEI.
2. Below VEI 3, several orders of magnitude are combined, making the scale ordinal rather than cardinal. Generalizations about power-law behavior apply only to eruptions of VEI ≥ 3.
3. Only bulk volumes are considered. This has the advantage of not requiring bulk density data, but the disadvantage of not being strictly comparable from one eruption to another. Dense rock equivalency (DRE) is a useful, normalizing concept, but users should be aware of considerable uncertainty in its calculation.

Ideally, any description of an explosive eruption should describe both magnitude (expressed in mass) and intensity (expressed in mass ejection rate per unit time). The most popular alternative to VEI is the mass-based magnitude (M) scale of Pyle (1995, 2000, 2015). The M scale as originally proposed (Pyle, 1995) referred only to tephra (sensu latu, all pyroclastic material). With the abovementioned exception of dome-collapse materials, it can be expected to correlate well with VEI with the added advantage of higher accuracy and precision if the deposit volumes and bulk densities are well known. The LaMEVE database of Crosweller et al. (2012) also refers only to tephra, excluding lava. However, Pyle (2000, 2015) and Oppenheimer (2011; using the abbreviation M_e) combined masses of both tephra and lavas. Ultimately, a way forward to more consistent and quantitative description of eruptions might be to use both intensity and magnitude, but add subscripts to indicate the material measured, e.g., M_t for magnitude based on tephra, or M_l for magnitude based on lava, and the same for intensities I_t and I_l. Estimates based on a combination of tephra and lava could be indicated as M_{tl} or I_{tl}. This suggestion follows convention in seismology to indicate the basis on which the magnitude of an earthquake was estimated, e.g., M_w for moment magnitude.

We support use of the M and I in lieu of VEI when data are sufficient, but we also note that many eruptions are too poorly known for robust estimation. Of the 17 probable VEI 7 eruptions of the past 10 k.y. (Table 1), 8 or more are too poorly known for robust estimation of M, much less of I. In Table 1 we include M as reported in the LaMEVE database (Crosweller et al., 2012). Given currently available, generally sparse, data, and the fact that VEI captures both magnitude and intensity into a widely understood, order-of-magnitude shorthand, we retain VEI throughout the rest of this paper.

KNOWN CLIMATIC AND RELATED EFFECTS OF A VEI (OR M) 7 ERUPTION: RINJANI (SAMALAS) IN CE 1257

The largest sulfate signal in ice cores from the past several thousand years, ~2.5 times that of Tambora in 1815, dates to CE 1258; this signal has been associated with cold weather in the years following (Stothers, 1999, 2000; Oppenheimer, 2003b, 2011). A caldera-forming eruption from the Rinjani volcanic center on Lombok Island, Indonesia, was already known to have occurred between CE 1250 and 1300 (Nasution et al., 2004b). Lavigne et al. (2013), Vidal et al. (2015, 2016), and Guillet et al. (2017a) connected the Rinjani eruption to the strong sulfate signal in ice cores. Based on tephra dispersal and wind patterns, Lavigne et al. (2013) further suggested that this eruption occurred between May and October of CE 1257.

Alloway et al. (2017) described widespread distribution of the Rinjani ash across Indonesia and hypothesized widespread social disruption, but archaeological evidence to date is inconclusive. Recent abstracts describe a search for buried settlements on Lombok Island (Lavigne et al., 2017) and indicate that Guillet et al. (2017b) are continuing to explore global climatic effects of the eruption.

Timmreck et al. (2009) noted that proxy temperature reconstructions by Mann et al. (2008) show only a slight temperature drop after 1258, so they proposed (after Pinto et al., 1989) that large aerosol droplets in the Rinjani cloud would have dropped out quickly and limited the cooling effect. We believe that conclusion needs to be revisited.

What other changes were reported after this CE 1257 eruption? A dry fog and cool temperatures were noted in the northern summer and autumn of 1258 (Stothers, 2000). Heavy rains during the summer of 1258 ruined crops across Europe, bringing famine, and grain shortages in France led to inflation of food prices in 1258 (Stothers, 2000). England was hard hit and archaeologists have now discovered mass graves with more than 10,000 skeletons at Spitalfields in London, dating close to 1258 (Sidell et al., 2007; Connell et al., 2012; Walker, 2012; Campbell, 2017) (Fig. 2). Carly Hilts (in a capsule summary of Walker [2012] for the magazine Current Archaeology, http://www.archaeology.co.uk/articles/features/londons-volcanic-winter.htm) wrote:

> The year [1258] got off to a bad start, with the monk [Matthew Paris] recording *such unendurable cold, that it bound up the face of the earth, sorely afflicted the poor, suspended all cultivation, and killed the young of the cattle* [Italics added to indicate quoted material from Walker, 2012]. But that was just the beginning. As the year progressed the situation deteriorated and it was not long before Paris had cause to put pen to parchment once more. By midsummer he could write that:
>
> > … when April, May, and the principal part of June, had passed, and scarcely were there visible any of the small and rare plants, or any shooting buds of flowers; and, in consequence, but small hopes were entertained of the fruit crops. Owing to the scarcity of wheat, a very large number of poor people died; and dead bodies were found in all directions, swollen and livid, lying by fives and sixe's in pigsties, on dunghills, and in the muddy streets… When several corpses were found, large and spacious holes were dug in the cemeteries, and a great many bodies were laid in them together.

The Eurasian winter of 1257–1258 was cold, but that of 1258–1259 was quite mild, as predicted by models and observations of winter warming and summer cooling in northern Europe (Robock and Mao, 1992; Robock, 2000; Shindell et al., 2004; Fischer et al., 2007). Systematic Medieval Irish chronicles confirm the occurrences of mild winters in Ireland in 1258–1259 and 1259–1260 (Ludlow et al., 2013), followed by slightly colder than average episodes during several decades thereafter. Stothers (2000) reported that by 1260–1261 winters again turned severe. In Iceland, people were forced to slaughter many of their livestock and sea ice surrounded the island and the Ill River in Alsace uncharacteristically froze.

Guillet et al. (2017a) and Campbell (2017) agreed that the eruption of 1257 must have contributed to famines and high mortality of the time, particularly in England and Japan, but cautioned that because some of the famines started earlier, not all can be attributed to the eruption. With similar caution, Xoplaki et al. (2016) noted that in the eastern Mediterranean, several years of intense cold following the 1257 eruption might have hastened the decline of the eastern Roman (Byzantium) empire, which already was in process before the eruption. In Asia, the Mongol empire was also starting to decline following the

Figure 2. Mass grave at Spitalfields, London, ca. CE 1258. Photo courtesy of MOLA (Museum of London Archaeology). Many of the deaths are thought to be from crop failure and famine, greatly aggravated by the 1257 eruption of Rinjani (Samalas) Volcano.

death of Genghis Khan, and its dissolution was so rapid in ca. 1260 that Oppenheimer (2011), citing Jackson (1978), suggested that the Rinjani eruption might have been the last straw for the Mongol empire as well. However, Morgan (2009) proposed a more prosaic explanation of the Mongol empire decline, i.e., that it simply grew too large and was unable to keep itself together. Details of timing are crucial for attribution, and we judge that the eruption of Rinjani aggravated but cannot be blamed for all of these changes.

Cooling did not stop within a few years after the Rinjani eruption. Instead, it marked the start of the Little Ice Age (LIA) (Grove, 2001). Although many different dates have been suggested for the start of the LIA, we take as definitive Grove's (2001) conclusion that it began in the mid-thirteenth century and intensified in the fourteenth century. Within the precision of definition and dating, this is coincident with the Rinjani eruption. Thereafter, throughout the course of the LIA, global temperatures were ~0.5 K below those of the Medieval Warm Period, and Europe experienced many cold winters (Fig. 3). Although the LIA has been attributed to a decrease in solar warming of the Earth (e.g., Eddy, 1976; Bard et al., 2000), and there is enough association to consider solar variability as a contributor to the LIA (Schneider and Mass, 1975; Crowley, 2000), there is growing evidence that volcanic eruptions are the main culprit (Robock, 1979; Porter, 1986; Crowley et al., 2008; Miller et al., 2012; Plummer et al., 2012; Slawinska and Robock, 2018). Rinjani is the first and largest of several eruptions in succeeding decades; Gao et al. (2008) identified additional, moderate to large sulfate injections in CE 1228, 1268, 1275, and 1284, and concluded that the cumulative volcanic sulfate flux in the thirteenth century was 2–10 times larger than that in any other century within the last millennium. One eruption that occurred around the time of Rinjani, plus or minus ~50 yr, was from Quilotoa (Ecuador) and was about twice the size of Pinatubo in 1991 (Hall and Mothes, 2008a; Mothes and Hall, 2008).

Cooling of ~0.5 K, the average surface cooling during the Little Ice Age, might not seem like much, but it apparently had a significant global effect on crops and health. Atlantic Ocean pack ice began to grow around Iceland, and was a nuisance for roughly a century (Lamb, 1979). In Europe, some years lacked a warm summer for growing crops, and there was great famine and plague in 1315–1317. The prices of wheat in Europe doubled from ca. CE 1250 to 1300, and, after returning to pre-1250 prices by 1400, rose sharply again from 1550 to 1650. Lamb (1995) interpreted those changes to reflect 15%–20% shorter growing seasons. In Norway, many farms were abandoned and production and tax yields dropped by as much as 70% from 1300 to 1387. Cod fishing in Scotland declined as the cod moved south.

Black Death (caused by the *Yersinium pestis* bacterium) is thought to have spread from China to Europe, and, in only a few years (CE 1348–1350), between 75 and 200 million people in Europe were killed by this disease. We have not found any suggestion of a direct causal link between the plague bacterium and volcanic eruptions, but reduced crop yields, famine, and cold surely must have made populations more vulnerable to the plague than they would otherwise have been. Living conditions were generally poor and led to social unrest in several places. To be sure, not everything can be blamed on the cooler climate, but people simply cannot go hungry for very long without becoming restless, ill, or both. High food prices at the time are clear evidence that food was in short supply.

In the tropical western Pacific, Nunn (2000a, 2000b, 2007) inferred widespread environmental disruption ca. CE 1300 ± 50. On Pacific islands from Guam through Fiji to the Tuamoto Islands, many archaeological sites moved inland from coastal areas, or onto new islands. What might have caused these cultural changes? One possibility is a drop in sea-level change that exposed and killed coral reefs. Nunn (2000a, 2000b), citing many studies of coral microatolls and lagoon sediments, inferred one phase of ~75 cm of sea-level drop in the decades following CE 1250, and another of ~40 cm in decades following ca. CE 1450. Dickinson (2004) inferred crossover (when declining ambient high-tide level falls below the mid-Holocene highstand low-tide level) starting ca. CE 1250. Sea level in Israel may also have dropped at or about the time of the Rinjani eruption by ~0.5 ± 0.2 m, attributed to coincidence of a positive North Atlantic Oscillation and a negative Southern Oscillation Index (Toker et al., 2012). Lesser drops in sea level, no more than 25 cm, were inferred by Goodwin and Harvey (2008), Lambeck et al. (2010), and Gehrels et al. (2011). Church et al. (2005) documented a drop of slightly less than 1 cm after the Pinatubo eruption, and Grinsted et al. (2007) reported small, more variable changes after 5 eruptions of Pinatubo scale and smaller.

Might there have been alternate factors contributing to such cultural disruption reported by Nunn (2000b, 2007)? Droughts, crop failures, and widespread

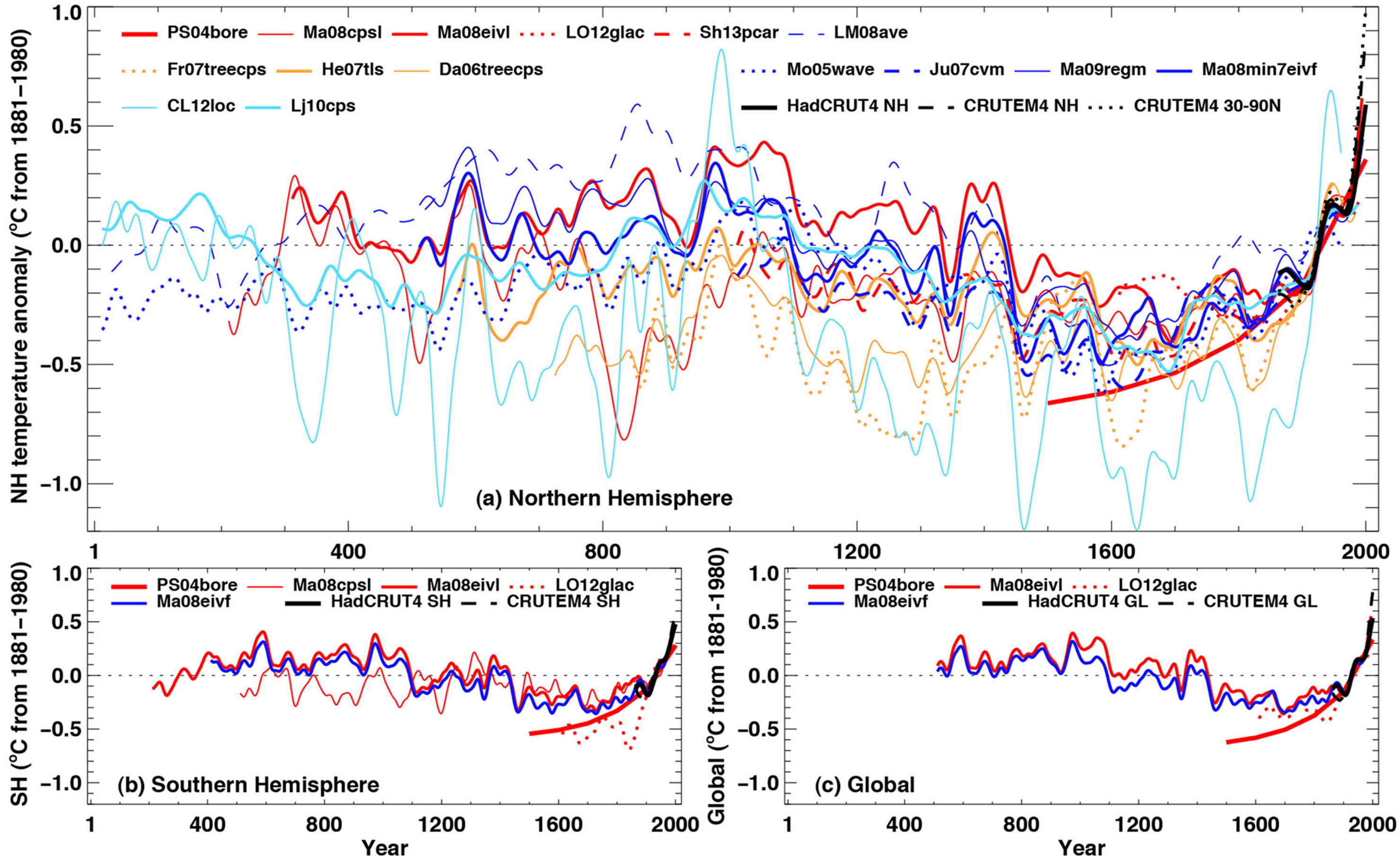

Figure 3. (A) Reconstructed Northern Hemisphere (NH) annual temperatures during the past 2000 yr. (B) Reconstructed Southern Hemisphere (SH) annual temperatures during the past 2000 yr. (C) Reconstructed global annual temperatures. Individual reconstructions are shown as indicated in the legends, grouped by color according to their spatial representation (red—land-only all latitudes; orange—land-only extratropical latitudes; light blue—land and sea extra-tropical latitudes; dark blue–land and sea all latitudes) and instrumental temperatures shown in black [Hadley Centre Climatic Research Unit (CRU) gridded surface temperature-4 data set (HadCRUT4) land and sea, and CRU Gridded Data set of Global Historical Near-Surface Air Temperature Anomalies Over Land version 4 (CRUTEM4) land only (Morice et al., 2012)]. All series represent anomalies (°C) from the 1881–1980 mean (horizontal dotted line) and have been smoothed with a filter that reduces variations on time scales <~50 yr (from Masson-Delmotte et al., 2013, figure 5.7 therein, with further information about each individual reconstruction).

fighting are certainly possible, the latter indicated by construction of many fortifications. As noted here, global precipitation also can be reduced due to the effect of radiative changes wrought by volcanic aerosols, and the effect might be particularly acute in the western Pacific if El Niño conditions (cool and dry in the western Pacific) were induced by the eruptions, as discussed here. Anderson et al. (1992) and Rein et al. (2005), based on independent sedimentary records in Peru, showed a marked increase in the frequency of El Niño conditions starting in the mid-thirteenth century and persisting for most of the LIA. This would very likely have adversely affected food production in the western Pacific. The mechanism by which the El Niño conditions were sustained through the LIA might be the same as that which triggered an El Niño right after the 1257 eruption. Morrill et al. (2003) noted a major change in the Asian monsoon ca. CE 1300, and although they did not attribute it to volcanic activity, we suggest that it, too, might have begun with the 1257 eruption of Rinjani.

From the fifteenth century until ca. 1800, temperatures became even colder in Europe. Pack ice around Iceland expanded even more during the fifteenth to nineteenth centuries (Lamb, 1979; Ogilvie and Jónsson, 2001; Massé et al., 2008). The Kuwae (Vanuatu) eruption has been blamed by many for a notable temperature drop in the 1450s (e.g., Monzier et al., 1994; Robin et al., 1994; Witter and Self, 2007). However, Németh et al. (2007) argued that deposits of this age are too small for this eruption to have been responsible for a major global spike in SO_4, and Plummer et al. (2012) reported evidence for at least two large eruptions that decade, so perhaps there were other sources as well. There are so many potential volcanic sources that it is easy to speculate, but difficult to firmly associate, specific eruptions with sulfate spikes in ice cores. The global picture, however, is clear; i.e. that sulfate aerosol from all (volcanic) sources during the LIA was significantly higher than before or since (Gao et al., 2008). Other sulfate contributors probably included the eruptions of an unknown volcano in 1587 (possibly Billy Mitchell, Papua New Guinea), Huaynaputina (Peru) in 1600, Parker (Philippines) in 1641, Long Island (Papua New Guinea) ca. 1660, and an unidentified 1695 eruption (Sigl et al., 2015).

GENERAL HAZARDS AND RISKS OF VEI 7 ERUPTIONS

Generalizing beyond the Rinjani event, what hazards and risks should we expect from other VEI 7 eruptions? We distinguish between three zones of hazards and risk: proximal, distal direct, and distal indirect.

Proximal Hazards and Risks

A VEI 7 eruption occurring today near a large city today could erase the entire city. Pyroclastic flows from VEI 7 eruptions can travel 20, 30, or even 100 km in any or all directions from the volcano. Pyroclastic flows from Kikai traveled as far as 100 km, across water and land, those from the Taupo 1.8 ka eruption traveled at least 80 km, and those from the 39 ka Campanian eruption of Campi Flegrei reached distances of at least 70 km through what is now a densely populated urban area (Scarpati et al., 2015). Any town or city in the path would be destroyed, and death tolls could reach millions unless mass evacuations had been made. Pyroclastic flows that enter the sea trigger local tsunamis that can travel for a few hundred kilometers and wipe out coastal communities, as happened at Krakatau in 1883: the dilute ash-cloud surge facies of the Krakatau eruption flowed across the top of the sea and wreaked havoc at land fall, 40 km across the sea (Carey et al., 1996; Paris et al., 2014), and the eruption was only a low VEI 6.

The proximal zone can also receive a thick fall of pumice and ash, tens of centimeters or even meters thick. Figure 13 of Lowe (1988) shows expectable decreases in thickness with distance, using large New Zealand eruptions as examples. Falling ash and pumice is rarely lethal by itself, but it can kill by accumulating on roofs that then collapse (Blong, 1984; Tanguy et al., 1998; Witham, 2005; Auker et al., 2013). Farther away from the volcano, but still what we might consider proximal, it can damage crops and cause expensive damage in urban areas [e.g., with power lines, computers, water treatment facilities, centralized HVAC (heating, ventilation, air conditioning) systems, machinery, and vehicles] (Heiken et al., 1995; Wilson et al., 2012b; Wardman et al., 2012; Wilson et al., 2012a, 2014).

Most of these hazards can be illustrated from Pinatubo, where a low-end VEI 6 eruption occurred in 1991 (Figs. 4–6). This eruption was of the same type as most VEI 7 eruptions, albeit 10 times smaller. Many details can be found in Newhall and Punongbayan (1996a).

VEI 7 eruptions typically last only a few days; however, ash problems do not end with the eruption and primary deposition. There will likely be large volumes of pyroclastic flow deposits on the slopes of the volcano that will remain hot and unstable for years. If these are subjected to erosion, there will be secondary explosions of fine ash that can reach as high as commercial flight levels (Torres et al., 1996). One of the damaging ash-aircraft encounters at Pinatubo was with ash from a secondary explosion a year after the eruption had ended (Casadevall et al., 1996), and problems like this will be magnified in case of a VEI 7 eruption. Fortunately, volumes of these secondary explosions are small, so the geographic reach of secondary ash fall will be limited to a maximum of a few hundred kilometers.

Should a VEI 7 eruption occur through an existing caldera lake or onto a volcanic massif covered with snow and ice, large-volume lahars would form immediately (Pierson et al., 1990; Mastin and Witter, 2000). A lahar from even the small VEI 3 eruption of Nevado del Ruiz (Colombia) in 1985 swept the town

Figure 4. Pyroclastic flow deposit (cream colored) filling 200-m-deep Marella Valley, southwest side of Pinatubo (Philippines) ~10 km from the vent. Photo courtesy of R.P. Hoblitt, U.S. Geological Survey.

Figure 5. Ash-fall damage to rooms at Clark Air Base, ~25 km east of Pinatubo (Philippines). The photograph is in true color (courtesy of W.E. Scott, U.S. Geological Survey).

Figure 6. Town of Bacolor, 35 km southeast of Pinatubo (Philippines), buried by an average of 5 m of lahar deposits. In the foreground, roofs are visible; in the background, even roofs are buried. Photo courtesy of C.G. Newhall, U.S. Geological Survey.

of Armero clean to the foundations of buildings, with a death toll of ~25,000 residents. Much larger lahars in mountainous terrain can travel hundreds of kilometers, e.g., the 1877 Chillos Valley lahars from Cotopaxi volcano in Ecuador (Wolf, 1878; Mothes et al., 1998).

Like secondary explosions, lahars and problems of excess sediment do not end with the eruption. Rain-induced lahars increase dramatically after a large explosive eruption, because vegetation has been thinned or damaged, and fine ash on the ground surface reduces infiltration of rainfall so much that most rain runs off in flash floods. Such flash floods scour fresh, unconsolidated volcanic deposits, bulking themselves into lahars. At Pinatubo, lahars persisted for roughly a decade and the effect will be similar or longer for a VEI 7, depending on topography, any welding of the pyroclastic flow deposits, and the intensities of rainfall in the area. Even after lahars stop, the huge amounts of sediment in stream and river channels around the volcano will have reduced channel capacity so much that even normal rains can cause flooding. Without massive dredging, this effect can last for many decades (Major et al., 2000; Gran et al., 2011; Pierson and Major, 2014).

An excellent, if now somewhat dated, review of the effects of these proximal and distal hazards was given by Blong (1984). Recent updates on potential effects on health and infrastructure were given by Horwell and Baxter (2006) and Wilson et al. (2012b), respectively, and there are excellent Web sites on both topics at http://www.ivhhn/org and http://volcanoes.usgs.gov/ash.

Three very large changes make today's risks vastly higher than at previous VEI 7 eruptions. The first is population growth. For example, greater Naples, which at the time of the last eruption of Campi Flegrei had a population of ~100,000, today has ~4.4 million. Greater Manila, which was barely more than a few villages at the time of the Taal 5600 yr ago eruption, now has a population of ~25 million. The population of the world was ~400 million when Rinjani erupted in 1257, and slightly less than 1 billion when Tambora erupted in 1815; today it is more than 7 billion.

The second change is complexity of systems for the delivery of basic human needs of food, water, and shelter. At the time of previous VEI 7 eruptions, communities were more or less self-sufficient. In early times, communities could be severely threatened if ash or related hazard caused their crops to fail. Today, we have complex logistical chains that can bring in food and avert famine, but these logistical systems are themselves subject to disruption.

The third change is an explosion of technology. An examination of likely effects on just a few modern systems can suffice to show how the next VEI 7 eruption will have very different impacts than previous ones. For example, today's electrical power grids are vulnerable to disruption because ash on lines and on transformers causes arcing or flashovers that will shut down the system. Wet ash is particularly effective in causing flashovers, and is also heavy enough to collapse either lines or tree branches overhanging residential power lines (Heiken et al., 1995; Wilson et al., 2012b; Wardman et al., 2012). The damage for the utilities is significant, but is dwarfed by the costs of widespread, prolonged interruptions of power to our entire society. Nearly everything we do depends on having steady electrical supply.

The effects of volcanic ash on computers and other electronics with semiconductor chips are unclear. Certainly, semiconductor manufacturers go to great expense to avoid contamination by dust. Yet, during the only direct tests with which we are familiar, Gordon et al. (2005) and Wilson et al. (2012a) found that computers were more resilient than expected, unless the ash is also moist or there was prolonged exposure to corrosive volcanic gases. Fresh ash, perhaps fresher than that used in the experiments, can scavenge acidic aerosols from the ash cloud (Rose, 1977) and thus be potentially more troublesome.

Vehicles and machinery will also be affected unless well protected. Ash is slippery when wet and increases the rate of motor vehicle accidents. If dry, it quickly clogs vehicle air filters. If the ash gets into the moving parts of machinery, abrasion becomes a serious problem. If the machinery is used to power infrastructure, e.g., pumps for irrigation, domestic, or industrial water supply, or turbines for generating electricity, that infrastructure may need to be taken offline frequently for maintenance. Jet aircraft are especially vulnerable and are discussed below.

The global positioning system (GPS) is used for a multitude of purposes, including civilian and military navigation, precise synchronization of clocks, and monitoring for volcanic and earthquake studies. Scientists have known for several years that atmospheric moisture and volcanic ash introduce phase delays in the GPS signals, so that one's apparent location (using precise, dual-frequency receivers) is changed by as much as 30 cm or more (Houlié et al., 2005; Grapenthin et al., 2013). Presumably, clocks would also be incorrect, because the GPS time signal would also be delayed by a small amount. Small shifts would not cause problems except where very high precision or synchronization is required; e.g., most financial transactions around the globe rely on GPS time signals. Other satellite remote-sensing returns, e.g., for weather forecasts and military surveillance, would also be degraded until the dust settles.

Of potentially greater concern, bloggers during the Eyjafjallajökull (Iceland) ash crisis reported that GPS units took longer than normal (many minutes instead of seconds) to lock into enough GPS satellites to navigate, and some were unable to lock in at all. Another report said that their GPS clocks reset themselves back to 1200:00:00. Larson (2013) analyzed the signal-to-noise ratio (SNR, analogous to bars of cell phone reception) for GPS stations under relatively small ash plumes from Alaskan volcanoes, and found substantial degradation lasting for ~30 min. The effect will surely be stronger and longer lasting in a VEI 7 eruption.

Reported effects of ash on other communications systems are remarkably few. The only two noted by Wilson et al. (2012b) are possible interference in radio waves from lightning, and overloading of networks because everyone is trying to use the network at once. Additional effects, anticipated in Wilson et al. (2009b), include overheating of equipment in case of air-conditioning breakdown, damage to dishes and towers due to weight of ash, failure of local exchanges due to corrosion of connectors, and more.

Distal Hazards, Direct and Indirect

Unlike most other rapid-onset natural hazards, volcanic eruptions can have global effects, as reviewed in Robock (2000), Self (2006, 2015), and Timmreck (2013). The two main avenues for global effects are via impacts on air traffic and on climate, the first a direct hazard, the latter indirect.

Effects on Air Traffic

Ash that is still high in the atmosphere and still relatively concentrated can pose a severe hazard to jet aircraft, and several planes within 500 km of an erupting volcano have lost all power and nearly crashed before they got one or more engines restarted (Guffanti et al., 2010; Clarkson et al., 2016). Tropospheric winds can blow ash for a few thousand kilometers; stratospheric winds can blow fine-grained ash around the world several times. Eyjafjallajökull ash (tropospheric only) spread thinly across Europe in 2010 and caused massive disruption of air traffic, business, and many other activities. Ash from Hudson Volcano in 1991 and Puyehue-Cordon Caulle in 2011 (both in Chile) disrupted air traffic in Argentina, Australia, and New Zealand.

An analysis by Oxford Economics (2010) showed that the economic impacts of the 2010 air traffic disruption in Europe were considerable, U.S. $5 billion over 1 month. The Eyafjallajökull eruption was relatively small but lasted for several weeks, produced fine-grained ash, and occurred during winds that blew the ash across all of Europe. The largest impact, estimated to be U.S. $2.2 billion, was on the airlines and other parts of the aviation sector, from deferred business and leisure travel. Next largest was impact on the hospitality sector, where lower visitor spending meant U.S. $1.6 billion in net losses worldwide. Productivity losses from stranded workers were estimated at U.S. $490 million, and losses to businesses that could not export or import fresh fruits, vegetables, and flowers were in the hundreds of millions of dollars. Losses in just-in-time supply chains of small, lightweight, high value items like electronics and pharmaceuticals must also have occurred.

Costs are incurred with every encounter of aircraft with ash. A crash would cost many lives, the cost of the aircraft (as much as several hundred million dollars), and potentially large liability settlements depending on the airline and the site of the crash. Luckily, to date no crashes or fatalities have resulted from aircraft-ash encounters. However, encounters even without crashes will damage engines and other components of the aircraft. Repair to a relatively new Boeing 747 in Alaska in 1989 was estimated as U.S. $80 million, comparable to the new cost (at that time) of roughly U.S. $120–150 million. Costs today are much higher. Engines may need to be replaced and repaired, at a minimum cost of several million dollars plus the cost of taking the plane out of service briefly. Maintenance and replacement cycles would also need to be shortened.

As a result of near disasters in Indonesia in 1982 and Alaska in 1989, and highlighted by further encounters with Pinatubo ash, all stakeholders in the ash-aviation problem met in 1991 to work toward a way to keep planes away from ash. This led to a network of 9 regional Volcanic Ash Advisory Centers (VAACs) being established through the 1990s that now operates continually to warn pilots of drifting ash clouds. While safety has improved greatly over the pre-VAAC days, VAACs still face significant challenges and there are still a few inadvertent ash-aircraft encounters.

Because modern society depends so strongly on air transport and is at the same time unwilling to tolerate high risk in the air, air traffic control authorities in each country are forced to balance safety precautions with requirements or socioeconomic pressures to keep planes flying. Most countries follow guidelines from the International Civil Aviation Organization (ICAO), a United Nations body. An ICAO task force convened after the 2010 disruption in Europe recently proposed loosening the guidelines, thus putting more onus for safety onto the carriers (Civil Aviation Authority, 2017). One positive effect of this change is that for future eruptions, the new, less conservative guidelines will lead to less disruption of air traffic than would have occurred under the old guidelines.

Ash from the low VEI 6 Pinatubo eruption was a regional event, reaching across much of Southeast Asia. Fortunately, none of the planes that inadvertently flew into it lost power, but some required engine replacements and other costly repairs. Guidelines for aviation safety from ash were not in effect at the time of the Pinatubo eruption, but today's standards (even the loosened standards) would have precluded air travel in most of Asia for several days. Ash from a VEI 7 eruption, even under the new ICAO guidelines, is likely to cause a major disruption in global air travel.

Ash from a VEI 7 eruption will cover a much larger area for weeks or longer, potentially girdling Earth at whatever latitude the eruption occurs. Ash from eruptions spreads poleward, and that from eruptions in the tropics will spread poleward in both hemispheres. Ash from a high-latitude eruption may have strong effects over a more limited area. The ash cloud from a VEI 7 eruption will include at least 10 times more ash than that of Pinatubo, and much of it will be very fine grained and able to spread over much larger areas than did the ash from Pinatubo. The exact effects will depend on many factors, including the duration of the eruption, percentage of fine ash, wind patterns, and availability of alternate airports. As we saw in the European ash-aviation crisis of 2010, the amount of economic disruption is increased greatly if ash covers broad areas for weeks at a time.

Effects on Global Climate

The last VEI 7 eruption, Tambora, occurred ~200 yr ago. Ideally, we would like to have had data from many such past eruptions to assess natural variability of climatic impacts, and to identify the most robust signals from such large eruptions, but we have had only a few to assess within historical time. Can we use more frequent and recent smaller VEI 5 and 6 eruptions, which were better observed to extrapolate impacts from VEI 7 eruptions? And can general circulation models of the climate system help to predict the effects of the next VEI 7 eruption?

Early students of the relationship between volcanism and climate hypothesized that fine-grained volcanic ash (dust) was a major contributor to cooling after eruptions (Lamb, 1970, placed in historical context by Kelly et al., 1998). The VEI in its earliest incarnation was envisioned as a tool to refine this approach with better volcanological information. But contemporary research including that by Pollack et al. (1976), Turco et al. (1982), and Toon and Pollack (1982) pointed to sulfate aerosol, not silicate dust, as the principal contributor to cooling, a conclusion that has been reaffirmed by many (e.g., Robock, 2000; Langmann, 2014). While fine ash can last for a few months (Vernier et al., 2016), sulfate aerosol from tropical eruptions persists with an e-folding time scale of 1 yr, and thus still remains in the stratosphere and exerts a cooling effect for 2–3 yr. The climate response is determined by the amount of aerosol and its size distribution, with smaller particles being more effective per unit mass at scattering sunlight and cooling Earth. While the 1963 Agung (Indonesia) and 1982 El Chichón (Mexico) eruptions, both VEI 5, had measurable climatic impacts, the 1980 Mount St. Helens (Washington, USA) eruption, also VEI 5, did not (Lipman and Mullineaux, 1981; Robock, 1981), because it injected very little sulfur into the stratosphere, and did not create a sulfate aerosol cloud. Without explicit information about sulfate, the VEI is an imperfect proxy for climate effect. Fortunately, sulfur in the stratosphere falls onto the Greenland and Antarctica ice sheets and is preserved there, so ice cores can give us an excellent record of the stratospheric aerosol loading, the amount of the material that actually caused the climate response. Nonstratospheric sulfate from nearby sources like Icelandic and Antarctic volcanoes can be filtered out in bihemisphere correlation between ice cores. Recent compilations and analyses of this record by Gao et al. (2008) and Crowley and Unterman (2013) were used to produce records of past eruptions and force climate model simulations, and a recent comprehensive record by Sigl et al. (2015) is now being used to produce forcing data sets for future simulations. If the stratospheric input is >~5 Tg SO_2, then volcanic eruptions produce noticeable climate impacts, as observed and modeled from recent such eruptions.

Table 2 lists the various climatic impacts of recent VEI 5 and 6 eruptions with the mechanism and timing. Most of these were described in Robock (2000). Here we describe recent research that provides context for some of these impacts, and whether the impact can be scaled up to VEI 7 eruptions.

Volcanic eruptions have long been known to induce global cooling, particularly over land (e.g., Stommel and Stommel, 1983; Robock and Free, 1995; Harington, 1992; Robock, 2000; Cole-Dai, 2010). This is because tiny sulfate aerosol droplets in the eruption cloud spread over the globe in the lower stratosphere, reflecting away incoming solar energy before it can warm Earth.

The most recent large eruption, Pinatubo in 1991, pumped ~20 × 10^6 t of SO_2 into the stratosphere (Bluth et al., 1992), lowering average global surface temperature by 0.5 °C for 2 yr (McCormick et al., 1995; Parker et al., 1996), with summer cooling and winter warming in the Northern Hemisphere (Robock and Mao, 1992; Robock, 2000). Some large tropical eruptions, like Krakatau in 1883 and Pinatubo in 1991, have global impacts, as the stratospheric aerosols are spread poleward in both hemispheres by the Brewer-Dobson circulation, but the aerosols from the slightly smaller 1963 Agung stayed confined mostly to the Southern Hemisphere while the aerosols from the smaller 1982 El Chichón were confined to the Northern Hemisphere. The atmospheric circulation at the time of the eruption and interactions with the aerosols are important.

High-latitude eruptions, like Laki (Iceland) in 1783 and Katmai in 1912, spread their aerosols to the subtropics of the same hemisphere, but not to the tropics or the other hemisphere. Because the mean stratospheric circula-

TABLE 2. CLIMATIC IMPACTS OF RECENT ERUPTIONS, MECHANISMS, ONSET TIMES, AND DURATIONS

Impact	Mechanism	Begins	Duration
Reduction of diurnal cycle	Blockage of shortwave radiation and emission of longwave radiation	Immediately	1–4 days
Summer cooling of Northern Hemisphere tropics, subtropics	Blockage of shortwave radiation	Immediately	1–2 yr
Increased CO_2 sink	More diffuse radiation	Immediately	1–2 yr
Global cooling	Blockage of shortwave radiation	Immediately	1–3 yr
Centennial cooling	Reduced heat flow to Arctic Ocean	Immediately	10–100 yr
Reduced tropical precipitation	Blockage of shortwave radiation, reduced evaporation	Immediately	~1 yr
Reduction of Asian and African summer monsoon	Continental cooling, reduced ocean evaporation, reduction of land-sea temperature contrast	0.5–1 yr	1 summer
Stratospheric warming	Stratospheric absorption of shortwave and longwave radiation	Immediately	1–2 yr
Winter warming of Northern Hemisphere continents	Stratospheric absorption of shortwave and longwave radiation, dynamics	0.5–1.5 yr	1 or 2 winters
Increased probability of El Niño in 1 yr, then La Niña	Tropospheric absorption of shortwave and longwave radiation, dynamics	1-2 weeks	1–2 yr
Ozone depletion, enhanced ultraviolet	Dilution, heterogeneous chemistry on aerosols	1 day	1–2 yr

tion is poleward, aerosols from high-latitude eruptions have shorter e-folding lifetimes of only 3–5 months, depending on the time of year, and hence, per unit mass of aerosol, a shorter lived and lower magnitude climate impact (Oman et al., 2005, 2006; Schneider et al., 2009; Kravitz and Robock, 2011). However, cooling of Earth in only one hemisphere may shift the Intertropical Convergence Zone toward the other hemisphere, causing global scale precipitation changes (Frierson and Hwang, 2012; Haywood et al., 2013). The impact of high-latitude eruptions on climate depends on the time of year, with little impact in the fall and winter when there is little insolation (Kravitz and Robock, 2011).

Through modeling and review of historical records, Oman et al. (2005, 2006) suggested that strong cooling of northern continents after eruptions in the high northern latitudes substantially weakened the Asian and African monsoons and associated rainfall and river discharge. Effects on rainfall reach beyond just monsoons. Trenberth and Dai (2007) analyzed precipitation and river discharge globally after the 1991 Pinatubo eruption, and found ~10% decreases in precipitation and river runoff for 2 yr. Cooling following large eruptions includes the oceans. Lower sea-surface temperatures reduce evaporation, atmospheric moisture, and precipitation, but the main effect is more cooling over land than over the oceans, weakening the dynamical driver of summer monsoon circulation and reducing the precipitation over land. Zambri and Robock (2016) confirmed this in climate model simulations for the past 150 yr for the largest recent tropical eruptions, Krakatau in 1883 and Pinatubo in 1991.

Looking at longer time scales, Baldini et al. (2015) correlated millennia-long glaciations with some known VEI 7 eruptions and the VEI 8 eruption of Toba. While it may be that century-scale effects might be extended to millennia with repeated volcanic eruptions, the identification and understanding of those multiple eruptions have not yet been made.

Robock (2000) reported no evidence for changed El Niño frequency following large volcanic eruptions, but Adams et al. (2003) found statistical evidence for a volcanic effect. Emile-Geay et al. (2008) used a simple climate model to show that El Niño events tend to occur in the year subsequent to major tropical eruptions, including Tambora in 1815 and Krakatau in 1883. Although McGregor and Timmermann (2011) concluded that eruptions would induce La Niña conditions, new analysis of more complex general circulation models (GCMs) by Maher et al. (2015) found the same relationship, but only following an initial increase in El Niño likelihood, with La Niña event probability peaking in the third year after eruption. These impacts are rather weak and reflect changes in the statistical distribution of El Niño and La Niña events. In recent history, a strong El Niño had already started at the time of the 1982 El Chichón eruption and a weak El Niño had started before the 1991 Pinatubo eruption, just by chance. Khodri et al. (2017), however, have recently shown that an El Niño is more likely in the boreal winter following large tropical eruptions, and have explained the mechanism for this as the cooling of tropical Africa and subsequent atmospheric and oceanic circulation responses.

Sulfate aerosols in the stratosphere not only reflect sunlight, but also absorb incoming near-infrared and outgoing longwave radiation. For tropical volcanic eruptions, this absorption causes local heating in the equatorial lower stratosphere, creating greater than normal temperature and density gradients between the equator and poles. These anomalous gradients result in a strengthened stratospheric polar vortex, with strengthened zonal winds leading to positive temperature anomalies over northern Eurasia and sometimes parts of North America, along with significant cooling in the Middle East (Robock and Mao, 1992; Perlwitz and Graf, 1995; Robock, 2000). A winter warming pattern (Shindell et al., 2004; Oman et al., 2005; Fischer et al., 2007) is the positive phase of the North Atlantic Oscillation, an index of the wintertime variability of north-south Northern Hemisphere sea-level pressure gradients between 110°W and 70°E (Hurrell, 1995; Christiansen, 2008), or the Arctic Oscillation (AO), the first empirical orthogonal function of Northern Hemisphere winter monthly sea-level pressure anomalies (Thompson and Wallace, 1998).

A positive AO corresponds to anomalously low pressure over the pole, and anomalously high pressure at mid-latitudes, with the anomalies changing signs in the negative phase. The Zambri and Robock (2016) climate model simulations projected such a winter warming in the first winter following the large 1883 Krakatau and 1991 Pinatubo eruptions, as observed.

When multiple large sulfur-rich eruptions occur within years or decades of each other, their influence on climate seems to be additive, as seen in the prolonged LIA after the Rinjani eruption. Each successive eruption prolongs the volcanic cooling, and both empirical evidence and modeling suggest that they can cause a shift in the North Atlantic Oscillation. Shifts in oceanic currents last decades or centuries, much longer than the few years of impact from individual eruptions (Otterå et al., 2010; Zhong et al., 2011; Miller et al., 2012; Booth et al., 2012; Zanchettin et al. 2012, 2013). The long duration of impacts, some with decade-scale lags, reflect coupling of the atmosphere, the sea, and sea ice in which cold atmospheric temperatures also lower sea-surface temperatures and cause significant changes in ocean circulation (Mignot et al., 2011; Metzner et al., 2012, 2014; Lehner et al., 2013; Schleussner and Feulner, 2013; Ding et al., 2014).

With much current interest in greenhouse gases, including CO_2, can large explosive eruptions add significantly to other sources? Compared to CO_2 from all sources, the amount of CO_2 added to the atmosphere by VEI 7 and larger eruptions is small. The annual average of all volcanic CO_2 emissions is estimated to be ~100 times less than the sum of anthropogenic emissions (Gerlach, 2011). During one rare high-end VEI 7 eruption, e.g., that of the Bishop Tuff of the Long Valley Caldera (California, USA), CO_2 emission was at least 0.45 Gt (after Anderson et al., 1989), slightly more than that of all volcanoes annually. During the last deglaciation, volcanic eruptions and volcanic CO_2 may have increased several fold (Huybers and Langmuir, 2009), but would still be far less than the current anthropogenic emission of CO_2, so it is unlikely that volcanic CO_2 will ever be a major addition to the larger anthropogenic greenhouse gas budget.

It is interesting that direct measurement of atmospheric CO_2 after volcanic eruptions shows slight slowing of the rate of ongoing, anthropogenic-dominated increases. After the 1963 eruption of Agung, 1974 eruption of Fuego (Guatemala), 1982 eruption of El Chichón, and 1991 eruption of Pinatubo, the increase of CO_2 concentrations measured at Mauna Loa (Hawaii, USA) actually slowed for a year. In what is now known as the Pinatubo effect, it appears that these eruptions stimulated biologic activity. One possible mechanism for this is that iron in the ash fertilized the ocean and increased algal growth (Frogner et al., 2001; Cather et al., 2009; Achterberg et al., 2013). Another possible mechanism that applies to vegetation on land is that sulfate aerosol renders incoming sunlight more diffuse, and plant photosynthesis is known to be more efficient in diffuse light than direct sunlight (Gu et al., 2003; Mercado et al., 2009). Gu et al. (2003) found that the effect in forests lasted 3–4 yr after the Pinatubo eruption. So, it appears that despite small additions of CO_2 from eruptions, net transfer of CO_2 from the atmosphere into biomass temporarily lowers surface temperatures below the already noted effect of sulfate aerosol (Angert et al., 2004; Kandlbauer et al., 2013).

Volcanic aerosol clouds in the stratosphere serve as locales that can favor heterogeneous chemical reactions between chlorine, nitrate, and chlorine gas (reservoir species from anthropogenic chlorofluorocarbons), release chlorine, and destroy ozone, in the same way the ozone hole is formed every spring over Antarctica. However, the ozone hole depends on polar stratospheric clouds, which only form in the cold polar night, while volcanic aerosols can serve this purpose at other times and places. After the eruptions of El Chichón in 1982 and Pinatubo in 1991, there was notable additional destruction of ozone (Solomon, 1999). Even the recent small 2015 Calbuco eruption in Chile, which emitted <1 Tg SO_2 into the stratosphere, caused Antarctic ozone depletion (Solomon et al., 2016). This impact of relatively small volcanic eruptions is enhanced now because of anthropogenic chlorine in the stratosphere, but as the effects of the Montreal Protocol of 1987 and its later amendments, adopted to prevent further emissions, continue to lower stratospheric chlorine over the next several decades, VEI 4 and smaller eruptions will have diminishing impact on ozone. In contrast, VEI 7 or larger eruptions will continue to have major impacts on ozone. Petrologic measurements to constrain halogen release, combined with modeling, suggest major ozone destruction after Plinian eruptions in Central America (Krüger et al., 2015) and after the VEI 7 Minoan eruption of Santorini (Greece; Cadoux et al., 2015). Bekki et al. (1996) and Bindeman (2006) estimated as much as 60% loss of ozone after large caldera-forming eruptions, gradually recovering over 1–2 decades.

Can climatic impacts of VEI 7 or larger eruptions be scaled up from measured effects of Pinatubo, El Chichón, and Agung? Ice cores show a strong increase in sulfate for ~6 yr, inferred by Zielinski et al. (1996) to be the Toba signal. Jones et al. (2005) and Robock et al. (2009) simulated the climate response to the Toba eruption, assuming that the stratospheric loading was a multiple of the 1991 Pinatubo eruption from 33 to 900 times the amount of aerosols, and calculated large climate responses with global average cooling of 7 K to 19 K. But even with strong sulfate signals in ice cores, there is some question as to whether effects will scale linearly with mass of sulfate. Larger injections, in theory, would produce larger aerosol droplets, which per unit mass would be less effective at reflecting sunlight. Furthermore, larger droplets would fall faster, resulting in shorter lifetimes (Pinto et al., 1989). Following this reasoning, Timmreck et al. (2010) inferred that climate effects of Toba would have been far less than the size of the eruption might suggest.

Is there empirical evidence for effects of the Toba eruption? Rampino and Self (1992) listed cases of abrupt cooling and sea-level drop that might have been amplified by the Toba eruption. From Ca^{2+} values in Greenland ice, Zielinski et al. (1996) inferred that the Toba eruption may have been responsible for as much as 200 yr of cooler temperatures, but did not cause full-blown glaciation that began ~2000 yr later. Proxy data from Lake Malawi (Africa; Lane et al., 2013) show no notable change in climate immediately above cryptotephra inferred to be Toba ash, but proxy data from elsewhere are ambiguous, owing in part to imprecision in dating of distal cores and archaeological sites (Williams, 2012). Eventually, more proxy data tied stratigraphically to the Toba ash will probably resolve current debates.

Socioeconomic Effects

A complete account of the economic and social impacts of a VEI 7 eruption, both proximal and distal, is beyond our expertise, but we present some points in the following.

During pre-eruption unrest and continuing throughout and beyond any period of evacuation, the host country will be severely challenged. The complexity of evacuating a large city was impressed on us by Hurricane Katrina (cf. Gheytanchi et al., 2007; Kendra et al., 2008), and evacuations in the face of an impending eruption will be much longer and with much higher uncertainty.

If a VEI 7 eruption occurs, little will survive within areas swept by pyroclastic flows. Beyond the zone of pyroclastic flows, ash fall could easily blanket a major city (e.g., Mexico City, Rome, Tokyo, Seoul, Manila, Singapore, or Jakarta), and bring it to a standstill by closing roadways, railways, and airways. Given the incredibly complex logistics of food, water, health care, and other supplies in an urban area, imagine the logistical nightmare if transport within any large city were stopped even for a week or two. Millions of hungry people do not stay quiet for long. National leadership may be challenged and replaced.

Farther afield, the effects of ash and climatic change will still pose unprecedented challenges. Agriculture is today a global enterprise, with much interdependence across national borders. Other goods are moved as well, with some key components moved by air. Supply chains can be very fragile, as shown during floods in Thailand in 2010. Financial systems are very sensitive to economic and political disturbances in the world, both of which might be expected in a VEI 7 eruption. Those who evacuate and survive a VEI 7 eruption will be refugees in need of food and shelter for an extended period (months to years), yet countries of the world are already resisting large numbers of refugees.

Last but not least, what if food or water shortages lasted for years or decades? For a while, countries with excess food and water can sell to those with less (Baum et al., 2015). Farmers, like any others, will try first to keep growing but, after a time, may lose either the will or the capacity to keep on. Eventually, major adjustments may be needed in economies. If desperation sets in, social unrest will grow and people now have many modern weapons with which to wage war on neighbors who have more food. Several have already drawn analogies between the effects of giant eruptions and those of nuclear war (Rampino et al., 1988; Schneider and Thompson, 1988; Robock and Toon, 2010).

LONG-RANGE FORECASTING OF VEI 7 ERUPTIONS

A number of conceptual and numerical models have been suggested to explain how large volumes of magma accumulate and erupt. Among these models, we note those of Jellinek and DePaolo (2003), Aizawa et al. (2006), Bachmann and Bergantz (2008), Gregg et al. (2012, 2015), de Silva and Gregg (2014), Caricchi et al. (2014a), Cashman and Giordano (2014), and Cashman et al. (2017). All raise interesting discussions, and the latter two papers call special attention to a growing number of cases of magma plumbing systems with several levels and complexity that will challenge any numerical model. There is, as well, a suggestion raised by Deligne et al. (2010), de Silva and Gregg (2014), Caricchi et al. (2014a), and Jellinek (2014) that different models may be needed for different sizes of magma systems, particularly for so-called supereruptions. We refer interested readers to the papers cited above.

It is obvious that a large volume of eruptible magma is required for a VEI 7 eruption. That volume will not be directly measurable until after an eruption, but can sometimes be inferred beforehand from petrologic and geophysical studies and eruptive histories. A triggering mechanism is also required, e.g., gradual buildup of overpressure, magma mixing, or tectonically induced convection or filter pressing, none directly observable but often inferable from the petrology of past products and/or from modern monitoring.

Beyond these rather general requirements, what other observable or inferable pre-eruption features or processes might indicate potential in a volcano for VEI 7 eruptions? Because no VEI 7 eruption has yet occurred in the era of modern volcano studies, hard evidence is sparse, but we can make some educated guesses. We draw from our experience in looking at caldera unrest around the world (Newhall and Dzurisin, 1988; Acocella et al., 2015), from the recent database of Large Magnitude Explosive Volcanic Eruptions (LaMEVE, http://www.bgs.ac.uk/vogripa/; Crosweller et al., 2012; Brown et al., 2014), from experience with the VEI 6 eruption of Pinatubo (Newhall and Punongbayan, 1996a), and from various recent papers.

With admitted uncertainty, we suggest and use six observable or inferable criteria (a–f) in drawing up a list of potential candidate volcanoes for future VEI 7 eruptions, and present in Table 3 a list of volcanoes that we know satisfy at least one of these criteria. Please debate whether these observables are the best ones, and what others might be used.

(a) The first criterion is volcanoes that have already produced at least one VEI 7 or larger eruption and have since been relatively quiet for thousands of years or longer. We include a few whose largest eruptions were high-end VEI 6 events. These have not already depleted their available magma. Considerable magma remains after most large eruptions, and what is erupted is usually replaced. Many VEI 7 eruptions are repetitions of previous VEI 7 eruptions from the same center (Newhall and Dzurisin, 1988; Crosweller et al., 2012; Brown et al., 2014), after a recharge period of at least a few thousand years. Even where the eruptive history is not well known, volcanoes with calderas commensurate with VEI 7 eruptions (~10 km in diameter or larger) are good prospects for future VEI 7 eruptions. Volcanoes within this category that have also already met or exceeded their known repose for VEI ≥ 7 eruptions might be considered candidates if they exhibit regularity in recharge and eruption. Table 3 also includes high-end VEI 6 eruptions to show the larger inter-VEI 7 eruptions. We exclude most volcanoes where the latest VEI 7 eruption occurred more than one million years ago since nearly all known intervals between multiple VEI 7 or larger eruptions at single volcanoes are shorter (Table 3; Fig. 7). A few exceptions to this 1 Ma older limit, e.g., the Kulshan caldera (Cascades), Acoculco (Mexico), and Nevado de Chachani (Peru), are included because they also show evidence of continuing, much younger activity.

TABLE 3. VOLCANOES THAT MAY BE CANDIDATES TO PRODUCE A VOLCANIC EXPLOSIVITY INDEX 7 ERUPTION IN FUTURE MILLENNIA*

1. Name	2. Global Volcanism Program Volcano Number	3. Lat	4. Long	5. Large eruption age (cal ka BP) or year (CE)†	6. Deposit volume (bulk, km³)	7. Repose (k.y.) since previous >50 km³	8. Edifice volume, km³ (most from Grosse et al., 2014)	9. Criteria for inclusion (see text for key to letters)	10. Pop30 km (in million people)	11. Pop100 km (in million people)	12. At risk	13. References
Region: Eurasia												
Vulsini-Latera (Italy)	211003	42.60	11.93	ca. 200	few			a	0.2	4.6	Tuscany	1, 2, 3
Vulsini-Bolsena (Italy)				590–490	few							1, 2, 3
Sabatini (Italy)	211809	42.17	12.22	ca. 400–375	>200[4]			a			Rome	1, 2, 4
Colli Albani (Italy)	211004	41.73	12.7	350	80			a	2.9	5.3	Rome	1, 2
				407	21							
				457	34							
Campi Flegrei (Italy)	211010	40.83	14.14	14.9	79	24.4		a, f	3, 0	6	Pozzuoli, Naples	1, 2
				39.3	320	20.7						
				50	100							
Santorini (Greece)	212040	36.4	25.4	3.7	117–129[5]			a, f	0.01	0.07	Santorini Island	1, 2, 3, 5
Kos (Greece)	212805	36.83	27.26	161	110			b	0.04	0.43	Nisyros, Kos, Yali	1, 2
Acigöl-Derinkuyu (Turkey)	213004	38.57	34.52	110	15	70	870[6]	a, b	0.22	2.3	Nevsehir, Turkey	1, 2, 6
				180	13							
Erciyes Dagi (Turkey)	213010	38.53	35.45	n/a	n/a		114	c, e	0.8	1.8	Kayseri, Turkey	1
Nemrut Dagi (Turkey)	213020	38.65	42.23	30	65	160	37 (maximum, 136)	a	0.23	1.6	Lake Van area, Turkey	1, 2
				190	>69	305						
				495	caldera formation							
Ararat (Turkey)	213040	39.7	44.28	n/a	n/a		226	c	0.14	2.8	Iğdir, Turkey	1
Kazbek (Georgia)	214020	42.7	44.5	n/a	n/a		506	c	0.008	1.7	Tblisi, Georgia	1
Aragats (Armenia)	214060	40.55	44.2	n/a	n/a		465	c	0.15	3.1	Yerevan, Armenia	1
Damavand (Iran)	232010	35.95	52.11	n/a	n/a		56 (or 400[7])	c, e	0.06	11.9	Tehran, Iran	1, 7
Region: Africa												
Corbetti/Awasa (Ethiopia)	221290	7.18	38.43	200	~100	470		b	1?	8?	Awassa, Ethiopia	1, 2, 8
				670	1000[2, 8]	230						
				900		400						
				1300								
O'a (Ethiopia)	221280	7.47	38.58	240	276			a	2?	8?	Ethiopia	1, 2
Gademota (Ethiopia)	221818	7.88	38.6	180–280[9]	n/a			a		8?	Ethiopia	1, 9
Gedemsa (Ethiopia)	221230	8.35	39.18	ca. 200[10]	10 km caldera			a	0.74	8?	Ethiopia	1, 10
Menengai (Kenya)	222060	–0.2	36.07	>29 (^{14}C); 36 Ar-Ar[11]	8 × 12 km caldera			a	0.9	4.3	Nakuru, Nairobi	1, 2, 11
Region: Oceania												
Rotorua (New Zealand)	241816	–38.08	176.27	240	~145 DRE			a				1, 2, 12
Okataina (New Zealand)	241050	–38.12	176.5	61	~100 DRE	~240		a, d	0.08	0.36	North Island, New Zealand	1, 2, 12, 13
				280[12] or ca. 320[13]	150 DRE	~250						
				550[13]	~90 DRE							
Kapenga (New Zealand)	241817	–38.27	176.27	61	16			a	~0.05	~0.3	North Island, New Zealand	1, 2, 12, 14
				275	100 DRE	~55						
				330	115	350						
(Matahana A)				680[14]	150	35						14
				710–720	~530	90						
(Matahana B)				805	150	85						14
				890[12]	~70							

(*continued*)

TABLE 3. VOLCANOES THAT MAY BE CANDIDATES TO PRODUCE A VOLCANIC EXPLOSIVITY INDEX 7 ERUPTION IN FUTURE MILLENNIA* (*continued*)

1. Name	2. Global Volcanism Program Volcano Number	3. Lat	4. Long	5. Large eruption age (cal ka BP) or year (CE)†	6. Deposit volume (bulk, km^3)	7. Repose (k.y.) since previous >50 km^3	8. Edifice volume, km^3 (most from Grosse et al., 2014)	9. Criteria for inclusion (see text for key to letters)	10. Pop30 km (in million people)	11. Pop100 km (in million people)	12. At risk	13. References
Region: Oceania (*continued*)												
Reporoa (New Zealand)	241060	–38.42	176.33	230	230			a				1, 2, 12
Ohakuri (New Zealand)		–38.37	176.05	240	150			a			Central North Island	15
Mangakino (New Zealand)	241818	–38.35	175.75	950–970	415	45		a	0.02	0.5	North Island, New Zealand	1, 2, 12
(Kauketea, Rocky Hill and Kidnappers combined)				1000–1010	5260	185						
(Ahuroa, Unit D)				1180–1200	460	40						
(Ongatiti, Mangatewaiiti)				1210–1240	1920	175						
(Unit C)				1400	115	140						
(Units A, B)				1530–1550	690	90						
(Pakihikura)				1630[16]	1000							16
Whakamaru (New Zealand)	241061	–38.42	176.08	325	1150	10		a	0.028	0.48	North Island, New Zealand	1, 2, 12
including Maroa				335	1000							
Taupo caldera (New Zealand)	241070	–38.82	176.00	1.8	105[17]	25.3		a	0.03	0.16	North Island, New Zealand	1, 2, 17
				26.5	1170							
Iamalele (Papua New Guinea)	253050	–9.52	150.53	n/a	n/a			b	0.008	0.114	Australasia air routes	1
Rabaul (Papua New Guinea)	252140	–4.27	152.20	1.4	24			a, f	0.2	0.25	Rabaul, air routes	1, 2
Hargy (Papua New Guinea)	252100	–5.33	151.10	11	10 x 12 km caldera		2.2	a	0.014	0.085	air routes	1
Sulu Range (Papua New Guinea)	252090	–5.5	150.94	n/a	n/a		5.1	b, f	0.015	0.14	air routes	1
Dakataua (Papua New Guinea)	252040	–5.06	150.11	0.65[18]	10 x 13 km caldera			a	0.004	0.11	air routes	1, 2, 18
Talasea/Garua (Papua New Guinea)	252060	–5.27	150.09	n/a	n/a			b	0.04	0.12	air routes	1
Long Island (Papua New Guinea)	251050	–5.36	147.12	CE 1661 ±10	30		96	a	0.004	0.08	Australasia air routes	1, 2
				19	100							
St. Andrew Strait (Papua New Guinea)	250010	–2.38	147.35	n/a	n/a			b	0.003	0.048	Australasia air routes	1
Ambrym (Vanuatu)	257040	–16.25	168.12	ca. 2 ka	>70[19, disputed by 20]			a	0.01	0.06	air routes	19, 20
Kuwae (Vanuatu)	257070	–16.83	168.54	CE 1450s	89[21, disputed by 22]			a	0.006	0.05	air routes	1, 2, 21, 22
Region: Southeast Asia												
Toba (Indonesia)	261090	2.58	98.83	74	2800	426		a	0.17	3.4	Medan	1, 2
				500	60	340						
				840	820	370						
				1200	80							
Maninjau (Indonesia)	261817	–0.35	100.2	52[23]	220–250[24]			a	0.71	3.5	Bukittinggi	1, 2, 23, 24
Ranau (Indonesia)	261251	–4.83	103.92	550 ± 150	100			a	0.021	0.94M	South Sumatra	1, 2
Rajabasa	261290	–5.78	105.63	n/a	20 km diameter caldera[25]			a	0.4	8	Sunda Strait, Lampung	1, 25

(*continued*)

TABLE 3. VOLCANOES THAT MAY BE CANDIDATES TO PRODUCE A VOLCANIC EXPLOSIVITY INDEX 7 ERUPTION IN FUTURE MILLENNIA* (*continued*)

1. Name	2. Global Volcanism Program Volcano Number	3. Lat	4. Long	5. Large eruption age (cal ka BP) or year (CE)†	6. Deposit volume (bulk, km³)	7. Repose (k.y.) since previous >50 km³	8. Edifice volume, km³ (most from Grosse et al., 2014)	9. Criteria for inclusion (see text for key to letters)	10. Pop30 km (in million people)	11. Pop100 km (in million people)	12. At risk	13. References
Region: Southeast Asia (*continued*)												
Awibengkok (Indonesia)	263040	–6.73	106.7	n/a	n/a			b	2.33	36.6	Jakarta (including Bogor)	26
Sunda (Indonesia)	263090	–6.77	107.6	ca. 50[27]	6 × 8 km caldera			a	5.7	32.8	Bandung	1, 2, 27
South Bandung cluster[28], from Patuha to Guntur	263070, 263080, 263805	–7.15	107.5	n/a	n/a			b	3.4	24.5	Bandung	28
Wilis (Indonesia)	263270	–7.81	111.76	n/a	n/a		319	c	2.2	21.5	Kediri	1
Tengger (Indonesia)	263310	–7.94	112.95	>38[29]	16 km diameter caldera			c	1.4	22.6	Malang, Lumajang	1, 2, 29
Iyang-Argapura (Indonesia)	263330	–7.97	113.57	n/a	n/a		780	c	1.6	10.5	East Java	1
Ijen (Indonesia)	263350	–8.06	114.24	>50	80[30]			a	0.67	6.8	Easternmost Java	1, 2, 30
Batur (Indonesia)	264010	–8.24	115.38	24 34	19[31] 84[31]			a	0.94	4.3	Denpasar, Bali	1, 2, 31
Buyan-Bratan (Indonesia)	264001	–8.28	115.13	310[32]	8 × 11 km caldera			a	0.95	4.35	Denpasar, Bali	1, 2, 32
Rinjani (Indonesia)	264030	–8.42	116.47	CE 1257[33]	60–80[34]		263	a	1.37	3.55	Mataram, Lombok Island	1, 2, 33, 34
Tambora (Indonesia)	264040	–8.25	118.00	CE 1815 43	~110–160[35] early caldera	43	640	a	0.09	1	Sumbawa Island	1, 2, 35
Tondano (Indonesia)	266070	1.23	124.83	ca. 700[36] 1300–2000	<50[37] >200[37]			a, b	0.9	1.56	Manado	1, 36, 37
Kitanglad-Kalatungan (Philippines)	271061	7.95	124.8	n/a	n/a		>200	c	0.36	5.3	central Mindanao	1
Malindang (Philippines)	271071	8.22	123.63	n/a	n/a		229	c	0.73	4.0	Ozamis, Oroquieta	1
Cuernos de Negros (Philippines)	272010	9.25	123.17	n/a	n/a		227	c	0.53	3.0	Dumaguete	1
Irosin (Philippines)	273010	12.77	124.05	41[38]	70[39]			a	0.52	3.4	Sorsogon province	1, 38, 39
Taal (Philippines)	273070	14.00	120.99	5.6[40] <140[41]	>50[41] >50			a	2.4	24.8	greater Manila	1, 40, 41
Laguna de Bay (Philippines)	273080	14.42	121.27	27–29 or older[41, 42]	500?[43]			a	7	26.3	greater Manila	1, 41, 42, 43
Region: East Asia–northwest Pacific												
Kikai (Japan)	282060	30.78	130.28	7.3 ca. 95[44] 630[44] 730[44]	200 150 n/a n/a	85		a	--	1	Kagoshima	1, 2, 44
Ata (Japan)	282070	31.2	130.55	108 240	350 150	132		a	0.17	1.9	Southern Kyushu	1, 2
Aira (Japan)	282080	31.58	130.67	28 456 500	456 100 100	428 44		a	0.9	2.6	Kagoshima, S Kyushu	1, 2
Kakuto (Japan)	282090	32.03	130.82	12 65 110 335 540 650	100 100 50 100 100 100	53 45 225 205 110	104	a	0.41	4	Southern Kyushu	1, 2

(*continued*)

TABLE 3. VOLCANOES THAT MAY BE CANDIDATES TO PRODUCE A VOLCANIC EXPLOSIVITY INDEX 7 ERUPTION IN FUTURE MILLENNIA* (*continued*)

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Region: East Asia–northwest Pacific (*continued*)												
Aso (Japan)	282110	32.88	131.10	87	600	29		a	0.23	6.5	Kumamoto	1, 2
				115	150	35						
				150	50	115						
				263–273	250							
Yufu-Tsurumi (Japan)	282130	33.28	131.43	600–615	63			a, c	0.71	7.8	Beppu, Oita	1, 2
Shishimuta/Kuju (Yabakei flow)	282120	33.17	131.25	870	100	130	25.9	a	0.18	7.8	Beppu, Oita	1, 2, 45
				1000	>110[45]	200						
				1200	100							
Kamitakara (Japan)	283837	36.20	137.50	500	100	110		a	0.11	5.1	Kanazawa, Toyama	1, 2
				610	50	40						
				650	100							
Momisawa-dake (Japan)	283839	36.36	137.61	336v390	~290			a	~0.1	~5	Kanazawa, Nagano	1, 2
Towada (Japan)	283271	40.47	140.92	14–15	100 (combined)	15		a	0.1	1.2	Sendai	1, 2
				30	50							
Hakkoda (Japan)	283280	40.65	140.88	470	100	310		a	0.48	1.9	Aomori	1, 2
				770–790	350	60						
				840–870	200							
Iwo-Jima/Ioto (Japan)	284120	24.75	141.33	2.6?[46]	~10 × 10 km caldera			a, f	0.001	0.001	military base, air routes	1, 2, 46
Toya (Japan)	285030	42.53	140.83	105	170			a	0.21	3.2	Muroran, Sapporo	1, 2
Shikotsu (Japan)	285040	42.68	141.38	25.4	125	11.6		a	0.32	3.1	Sapporo	1, 2
				37	200							
Akan (Japan)	285070	43.38	144.02	570	100			a, c	0.01	0.9		1, 2
Kutcharo (Japan)	285080	43.55	144.43	40	114	47		a	0.026	0.63	Northeastern Hokkaido	1, 2
				86.9	100							
Lvinaya Past (Russia)	290041	44.61	144.99	10.6	75–80			a	0.0006	0.006	North Pacific air routes	1, 2
Medvezhia (Russia)	290100	45.38	148.83	410	90			a	0.0006	0.004	North Pacific air routes	1, 2
Zavaritski (Russia)	290180	46.925	151.95	ca. 20	10-km-diameter caldera			a	0.00006	0.0001	North Pacific air routes	1, 2
Nemo (Russia)	290320	49.57	154.81	28	10			a	—	0.0006	North Pacific air routes	1, 2
				40	31							
				45	115							
Pauzhetka (Russia)		51.43	156.93									1, 2
Kurile Lake	300023			8.5	155	38		a	0.0004	0.004	North Pacific air routes	1, 2
				46	caldera formation	397						
Diky Greben	300022			443[47]	~300[47]							1, 2, 47
Opala (Russia)	300080	52.54	157.34	22	90	23	17.1	a	—	—	North Pacific air routes	1, 2, 48
				45	250[48]							
Gorely (Russia)	300070	52.56	158.03	39	120		13	a	—	—	North Pacific air routes	1, 2
Akademia Nauk/Karymsky (Russia)	300125	53.98	159.45	33	?			a	—	—	North Pacific air routes	1, 2, 47, 49

(*continued*)

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Region: East Asia–northwest Pacific (*continued*)												
	300130	54.05	159.44									
Odnoboky				69	12.5							
				95	?							
				177	minimum 50[49]							
Polivinka				432	>42							
				1130	100							
Bolshoi Semyachik (Russia)	300150	54.32	160.02	ca. 530	~100			a	0.0005	0.007	North Pacific air routes	1, 2
Uzon-Geyzir (Russia)	300170	54.5	159.97	42 78	150 46			a	0.0003	0.009	North Pacific air routes	1, 2
Krasheninnikov	300190	54.59	160.27	35–38[50]	9 × 11 caldera			a			North Pacific air routes	1, 50
Khangar (Russia)	300272	54.85	157.53	6.9 39 ca. 400[47]	<10 6–8 km diameter 12 × 16 km caldera			a	0.000004	0.001	North Pacific air routes	1, 2, 47
Alney-Chashakondzha (Russia)	300450	56.7	159.65	n/a	n/a		207	b, c, f[51]	—	0.011	Sredinny Range	1, 51
Changbaishan (China–North Korea)	305060	41.98	128.08	CE ca. 942 448	100 100	447	76	a, f	0.03	1.67		1, 2
Region: Americas and Caribbean												
Okmok		53.42	–168.13					a	—	—	North Pacific air routes	1, 2, 52
Okmok II				2	>>15 DRE[52]							
Okmok I				12	>>30 DRE							
Fisher (USA)	311350	54.67	–164.35	5.9 10	14 >100[53]			a	—	—	North Pacific air routes	1, 2, 53
Emmons Lake (USA)	312020	55.33	–162.07	25.8 96 234	>50[54] >50[54] >50[54]	70 140		a	0.0007	0.001	North Pacific air routes	1, 2, 54
Aniakchak (USA)	312090	56.88	–158.17	3.4 7.2	75 30			a?	—	0.0005	North Pacific air routes	1, 2
Katmai Volcanic Cluster (USA)	312140 to 312190	58.27	–155.16	CE 1912	28		>37	b, c[55]	—	—	North Pacific air routes	1, 2, 55
Baker-Kulshan-Hannegan (USA)	321010	48.78	–121.81	1150[56] 3700	117			a	0.009	1.99	Bellingham, nearby	1, 2, 56
Mazama (USA)	312160	42.93	–122.12	7.6	150			a, b	0.0004	0.27	Medford, eastern Oregon	1, 2
Newberry (USA)	322110	43.72	–121.23	75 ca. 300 ca. 500	50–70[57]		120 or 490[57]	a, b, c	0.016	0.18		1, 57
Medicine Lake (USA)	323020	41.53	–121.53	ca. 180	7 × 12 km caldera		148	b	0.0003	0.12	Northeast California	58
Lassen (USA)	323080	40.49	–121.51	609	>75			a	0.0023	0.44	Northern California	1, 2, 59
Long Valley, Mammoth Mountain (USA)	323150, 323822	37.7	–118.87	705 760	172 1380	55		a, f	0.008	0.08	Mammoth Lakes, air routes	1, 2

(*continued*)

TABLE 3. VOLCANOES THAT MAY BE CANDIDATES TO PRODUCE A VOLCANIC EXPLOSIVITY INDEX 7 ERUPTION IN FUTURE MILLENNIA* (*continued*)

1. Name	2. Global Volcanism Program Volcano Number	3. Lat	4. Long	5. Large eruption age (cal ka BP) or year (CE)†	6. Deposit volume (bulk, km^3)	7. Repose (k.y.) since previous >50 km^3	8. Edifice volume, km^3 (most from Grosse et al., 2014)	9. Criteria for inclusion (see text for key to letters)	10. Pop30 km (in million people)	11. Pop100 km (in million people)	12. At risk	13. References
Region: Americas and Caribbean (*continued*)												
Valles (USA)	327817	35.87	–106.57	55	41			a	0.016	0.99	Albuquerque	1, 2
				1223	690	390						
				1613	690	167						
				1780	100							
Yellowstone (USA)	325010	44.43	–110.67	143	23			a, e, f	—	0.03?	Wyoming, Montana	1, 2
				173	115	466						
				639	1000	631						
				1270	300	843						
				2113	290	20						
				2133	2160							
Acatlán (Mexico)	341822	20.45	–103.57	650	150			a	~4	~5	Guadalajara	1, 2
La Primavera (Mexico)	341820	20.62	–103.52	95	87			a, b	~4	~5	Guadalajara	1, 2
Acoculco (Mexico)	341827	19.97	–98.2	1400[60]	18 km caldera	~1600		a	0.49	5.2	Tulancingo, Tlaxcala	1, 2, 60
				ca. 3000	32 km caldera							
Los Humeros (Mexico)	341093	19.68	–97.45	100	30			a	0.57	5.7	Teziutlan, Xalapa	1, 2
				240	40							
				460	230							
Amatitlan/Pacaya (Guatemala)	342110	14.38	–90.60	45	45			a	2.5	7	Guatemala City	1, 2
				119	34							
				191	75							
Atitlan (Guatemala)	342060	14.59	–91.18	84	420			a	0.67	8	Guatemala City	1, 2
				158	18							
Apaneca (El Salvador)	343010	13.89	–89.79	525 ± 245	63			a	1.1	6.2	El Salvador, Honduras	1, 2
Coatepeque (El Salvador)	343041	13.87	–89.55	60 ± 10	~70			a, e	1.2	6.5	San Salvador	1, 2
Ilopango (El Salvador)	343060	13.67	–89.05	CE ca. 450	>70 (fall alone = 70)			a	3	6.7	San Salvador	1, 2
				36	36							
				73–84	large							
Rincón de la Vieja (Costa Rica)	345020	10.83	85.33						0.08	0.66	Northwest Costa Rica	1, 61
(Guachipelin)				1310	15 × 18 km caldera[61]							
(Cañas Dulces)				1430	200 DRE[61]							
Miravalles (Guayabo) (Costa Rica)	345030	10.75	–85.15	ca. 600	multiple[62]		62	a	0.06	0.76	Northwest Costa Rica	1, 2, 62
				1450–1500	multiple							
Barva (Costa Rica)	345050	10.14	–84.1	8	?		308	a, c	2.0	3.4	Heredia, San Jose	1, 2, 63
				322	>25[63]							
Irazu-Turrialba (Costa Rica)	345060, 345070	9.98, 10.03	83.85W 83.77W	n/a	n/a		590[64]	c, e, f			San Jose	1, 64
Chacana (Ecuador)	352022	–0.38	–78.25	170	?			?	0.35	3.3	Ecuador, near Quito	1, 2, 65
				220	?							
				2700–1800	670[65]			caldera formed				
Chalupas (Ecuador)	352826	–0.80	–78.39	211	90			a, c	0.15	3.6	Ecuador, near Quito	1, 2
Nevado Chachani (Peru)	354007	–16.19	–71.53	1030	small		>190	a	0.89	1.0	Arequipa	1, 2
				1650	22.5							
				2420	50							

(*continued*)

TABLE 3. VOLCANOES THAT MAY BE CANDIDATES TO PRODUCE A VOLCANIC EXPLOSIVITY INDEX 7 ERUPTION IN FUTURE MILLENNIA* (*continued*)

1. Name	2. Global Volcanism Program Volcano Number	3. Lat	4. Long	5. Large eruption age (cal ka BP) or year (CE)[†]	6. Deposit volume (bulk, km^3)	7. Repose (k.y.) since previous >50 km^3	8. Edifice volume, km^3 (most from Grosse et al., 2014)	9. Criteria for inclusion (see text for key to letters)	10. Pop30 km (in million people)	11. Pop100 km (in million people)	12. At risk	13. References
Region: Americas and Caribbean (*continued*)												
Uturuncu (Bolivia)	355838	–22.27	–67.22	n/a	n/a			c, f			Bolivia	1, 66
Maipo-Diamante (Chile)	357021	–34.16	–69.83	450	450			a	0.008	1.7	Rancagua, Santiago, air routes	1, 2
Calabozos (Chile)	357042	–35.56	–70.50	150 300 800	225 225 225	150 500			0.001	0.31	Argentina air routes	1, 2
Maule, Laguna del (Chile)	357061	–36.02	–70.58	n/a	n/a			b, f	0.002	0.169	Argentina air routes	1, 2, 67
Copahue (Chile)	357090	–37.85	–71.17	500	100			a	0.003	0.14	Chile-Argentina border	1, 2
Puyehue-Cordon Caulle (Chile)	357150	–40.52	–72.20	121	20?		291	b, c	0.007	0.52	Southern Lakes, Argentina air routes	1, 2
Dominica	360080 to 360100	15.5	–61.4	36	58			a	0.06	0.58	Island of Dominica	1, 2, 68

Note: Numbered footnotes in columns 1–12 indicate the specific information drawn from that numbered reference, e.g., 117–1293 in column 6 (deposit volume) for Santorini indicates that volume is from reference (3) Johnston et al., 2014. Abbreviations: lat, long are in decimal degrees. Pop30 and Pop100 are human populations within 30 and 100 km radii, respectively, of the volcano center, as estimated in 2011 using LANDSCAN and reported in VOTW4 (reference 1). n/a is not available. Key to criteria for inclusion (column 9): (a) prior VEI 7 eruption; (b) leaks of silicic magma within past 100 k.y., especially if in arcuate cluster; (c) mature, large stratovolcanoes, mostly >200 km^3 in volume; (d) volcanic complexes thought to be underlain by large volume of magma, and showing recent unrest; (e) extensional tectonics, high rate of magma supply; (f) probable accumulation of excess gas. See text for details.

The following numbers (as both footnotes and in references column) indicate general sources (1 and 2), and added references, many not yet in (1) and (2). Full citations are in the reference list of the paper. 1, VOTW4 of Global Volcanism Program, 2013, and sources cited therein (online at http://volcano.si.edu); 2, LaMEVE of Crosweller et al., 2012, and sources cited therein (online at http://www.bgs.ac.uk/vogripa); 3, Acocella et al., 2012; Palladino et al., 2014; 4, Palladino et al., 2014; 5, Johnston et al., 2014; Bruins, 2016; 6, Le Pennec et al., 1994, 2005; 7, Davidson et al., 2004 ; 8, Rapprich et al., 2016; 9, Morgan and Renne, 2008; 10, Peccerillo et al., 2003; 11, Blegen et al., 2016; 12, Wilson et al., 2009a; 13, Cole et al., 2010, 2014; 14, Houghton et al., 1995; 15, Gravley et al., 2007, reporting volume of 100 km3 DRE; 16, Pillans et al., 2005; 17, Wilson and Walker, 1985; 18, Torrence, 2012; 19, Robin et al., 1993; Picard et al., 1994; Sheehan, 2016; 20, Nemeth and Cronin, 2008; 21, Monzier et al., 1994; Robin et al., 1994; Witter and Self, 2007; 22, Németh et al., 2007; 23, Alloway et al., 2004; 24, Purbo-Hadiwidjoyo et al., 1979; 25, Bronto et al., 2012; 26, Stimac et al., 2008; 27, radiocarbon age adopted from Dam, 1994, though Kartadinata et al., 2002, and Nasution et al., 2004a, prefer a K-Ar age of 180–205 ka; 28, Bronto et al., 2006, and papers on individual geothermal centers; 29, per Mulyadi, 1992, age uncertain; 30, van Bergen et al., 2000; 31, Sutawidjaja, 2009; 32, Ryu et al., 2013; 33, Nasution et al., 2004b; Lavigne et al., 2013; Alloway et al., 2017; 34, Vidal et al., 2015; 35, lower volume estimate from Self et al., 2004; higher from Kandlbauer and Sparks, 2014; 36, Utami et al., 2004; 37, Lécuyer, 1998, using caldera diameters; 38, Mirabueno et al., 2011; 39, Kobayashi et al., 2014; 40, Martinez and Williams, 1999; 41, Listanco, 1993; 42, Catane et al., 2005; 43, Wolfe, 1982; 44, Ito et al., 2017; 45, Kamata et al., 1989; 46, Newhall and Dzurisin, 1988; Newhall et al., 1998; Nagai and Kobayashi, 2015; 47, Bindeman et al., 2010; 48, Melekestsev, 2016; 49 Ponomareva et al., 2013; 50, Ponomareva et al., 2007; 51, Global Volcanism Program, 2013, ref 1, with specific mention of domes arranged on ring fault; 52, Larsen et al., 2007; 53, Stelling et al., 2005 and Gardner et al., 2007; 54, Mangan et al., 2003; 55, Kubota and Berg, 1967; Matumoto, 1971, Ward et al., 1991, Jolly et al., 2007; Murphy et al., 2014; Hildreth and Fierstein, 2000, 2012; 56, Hildreth, 1996; 57, Kuehn and Foit, 2006; Jensen et al., 2009; 58, Donnelly-Nolan et al., 2008, 2016; 59, Muffler and Clynne, 2015; 60, López-Hernández et al., 2009; Roy et al., 2012; 61, Deering et al., 2007; Molina et al., 2014; 62, Chiesa et al., 1992; Gillot et al., 1994; 63, Pérez et al., 2006; 64, Stoiber and Carr, 1973; Alvarado-Induni, 2005; 65, Hall and Mothes, 2008b; 66, Sparks et al., 2008, Michelfelder et al., 2014; 67, Hildreth et al., 2009; Singer et al., 2014; 68, Smith et al., 2013.

*Selected on the basis of six criteria relating to past eruptive history, unrest, tectonic setting, and magma composition and degassing (see text). We offer this admittedly speculative list to stimulate discussion. Readers are invited to suggest additional or alternate criteria and candidates.

[†]All cases of CE are prefaced by CE. If no prefacing CE, it's ka (either in calibrated ^{14}C ka BP, or simply ka for those that are beyond the range of ^{14}C).

(b) Criterion b is volcanoes that have leaked silicic magma (typically rhyolite or rhyodacite) from multiple vents in the past 100 k.y. Bacon (1985) argued that arcuate or circular clusters of silicic vents are more indicative of a large silicic magma body and more likely to be a precursor of a large eruption than linear, fault-controlled silicic vents. We agree, and include circular clusters in Table 3, but generally exclude linear clusters of silicic vents.

(c) Mature, large stratovolcanoes or stratovolcano clusters, many with silicic domes and/or past production of welded and unwelded tuffs, but which have been relatively quiet in recent millennia compose the third criterion. Many calderas, especially in arcs, have truncated remnants of one or more large stratovolcanoes on their rims, e.g., at Rinjani (Samalas), as noted by Lavigne et al. (2013). Accordingly, we reordered the spreadsheet of Grosse et al. (2014) according to volume of edifice (using the more conservative volume, column U in his spreadsheet) rather than maximum volume. From this list, we avoided shield volcanoes and selected instead stratovolcanoes with a history of explosive eruptions, many also with a history of silicic magma and domes. We include every volcano satisfying these criteria and having an edifice volume of ≥200 km^3. Below 200 km^3, we included a few more that seem abnormally quiet to us, but skipped those that produce multiple VEI 6 events (e.g., Pinatubo, Krakatau, Witori, Papua New Guinea) because the latter seem not to store large volumes of eruptible magma beyond reposes of a few hundred to a few thousand years. In some cases, we combine nearby volcanoes that others treat separately, e.g., those of the Katmai group in Alaska, and a cluster of volcanoes just south of Bandung, Indonesia.

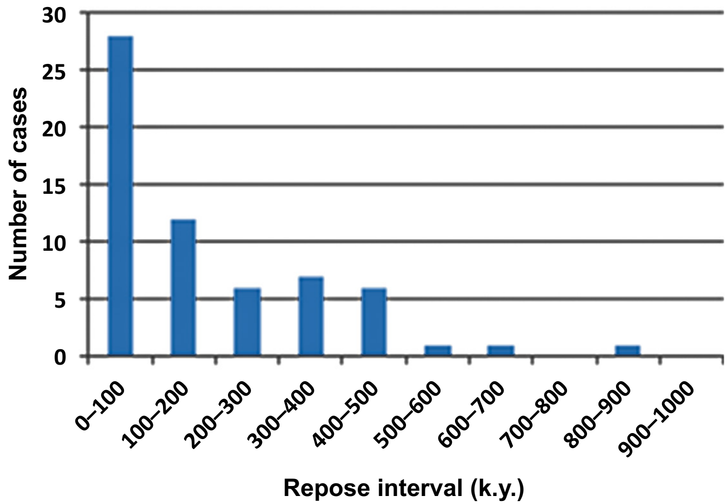

Figure 7. Repose intervals between eruptions of ≥50 km³ bulk volume at individual volcanic centers (in ka), shown in 100-k.y.-long bins. Data are from Table 3.

(d) Criterion d includes volcanoes at which present-day gas release is low despite evidence for periodic resupply of magma into the systems. If small increments of fresh magma and volatiles are added every few years, decades, or centuries into the base of large volcanic systems, much more frequently than eruptions themselves, as suggested by the frequent occurrence of volcanic unrest without eruptions and/or by crystal stratigraphy, then systems that are plugged will accumulate volatiles. Eventually, volatiles will exceed saturation and continue to accumulate in exsolved bubbles. Such exsolved volatiles were discovered after massive stratospheric injections of SO_2 from El Chichón (Luhr et al., 1984; Luhr, 1990) and Pinatubo (Gerlach et al., 1996), well in excess of what could have been dissolved in the erupted magma. Wallace (2001) and Shinohara (2008) documented many more cases. Winson (2016) concluded from a Bayesian belief network that accumulation of excess volatiles, such as occurred at El Chichón and Pinatubo, is the most consistent prerequisite in producing large Plinian eruptions. We suspect that such an excess-volatile phase is necessary to sustain very high mass discharge rates and allow an eruption to grow to VEI 7 scale.

(e) Criterion e includes volcanoes with unusually high rates of magma supply from the lower to the upper crust (Shaw, 1985; Caricchi et al., 2014 a, 2014b). Magma in arcs can be generated and supplied at unusually high rates where two slab segments overlie each other. Where there are slab tears or windows, asthenosphere of the mantle wedge can also rise, which accelerates melting and magma supply or resupply. Silicic centers of continental rifts (e.g., Ethiopia) or hotspots combined with extension (e.g., Yellowstone) are also inferred to be sites of relatively high magma supply from depth. Initial magma generation is often mafic, with secondary generation of intermediate or silicic magmas.

Slab segmentation and tears are inferable though offsets of a volcanic front, corresponding transverse faults and shallow seismicity, and discontinuities in the Wadati-Benioff zone, sometimes accompanied by calderas at the volcanic front and mafic monogenetic volcanoes behind (Stoiber and Carr, 1973; Carr et al., 1974). Unusually high rates of magma supply are inferable by detailed geologic mapping and age dating, and heat flow calculations (Shaw, 1985).

Hughes and Mahood (2008) noted that calderas are more consistently associated with local extension than with high convergence rates or obliquity of subduction. In some cases, magma rises in arc-parallel extensional zones, e.g., the Taupo volcanic zone or the longitudinal graben of El Salvador.

(f) Criterion f includes volcanoes that have shown notable unrest in recent years, in addition to some other qualifying criterion, e.g., Laguna del Maule in Chile, because it is remarkably restless and has a long history of silicic domes in an arcuate pattern (Hildreth et al., 2009; Singer et al., 2014). Unrest adds to the long-range probability of eruption as estimated by other criteria.

The criteria (a–f) for inclusion on the list are intentionally generous, erring on the side of inclusion. Given global recurrence frequencies, a few of these volcanoes, perhaps joined by others, will produce VEI 7 eruptions in the current millennium and in each successive millennium.

We emphasize that most of the volcanoes in Table 3 will not produce a VEI ≥ 7 eruption within the next millennium; likewise, we are emphatically not forecasting a VEI ≥ 7 eruption from any specific volcano on this list anytime soon.

Table 3 is just one way to identify cities at risk from major explosive eruptions, organized by volcano and potential specifically for a VEI 7 eruption. Another way is to organize a list by city and then ask about nearby volcanoes and the various volcanic hazards they face. Heiken (2014, chapter 9 therein) presented the latter. We invite readers to consider both lists, as both approaches are valuable.

SHORT-RANGE (IMMEDIATE) FORECASTING

Long-range forecasts of infrequent events are useful for making response plans that can be used whenever the next VEI 7 eruption threatens. However, in reality, implementation of that plan may be exceedingly difficult, especially in proximal areas where evacuations of millions of people might be required. The sheer size and complexity of magma plumbing systems that can produce VEI 7 eruptions (Cashman and Giordano, 2014; Cashman et al., 2017) means that magma intrusions from depth can easily pause or even stall en route to the surface.

Most unrest at calderas dies out rather than culminates in eruption (Newhall and Dzurisin, 1988; Acocella et al., 2015). Large magmatic-hydrothermal systems buffer incoming magma intrusions, and then show unrest in repeated episodes before they finally erupt again. Petrologic evidence for repeated intrusions before the VEI 7 Minoan eruption of Santorini (Druitt et al., 2012), before smaller recent eruptions at Rabaul (Bouvet de Maisonneuve et al., 2015), and instrumentally monitored unrest at Rabaul and Campi Flegrei (Acocella et al., 2015; McKee et al., 2017) reflect such buffering.

Experience from unrest and smaller eruptions at Rabaul (1979–1994; McKee et al., 2017), Campi Flegrei (1500s and ca. 1950–present; Acocella et al., 2015; Chiodini et al., 2016; Kilburn et al., 2017), and Pinatubo (1990–1991; Newhall and Punongbayan, 1996a, and papers therein) also showed that despite prolonged early unrest, the final runup to eruption was just hours to days. That an eruption was imminent did not become clear until hours to days before those eruptions.

Typically, short-range forecasts of an eruption will be a series of forecasts that are increasingly specific about whether an eruption will occur, a time window within which it might occur, and, where possible, how large and explosive it might be. Forecasting size or explosive magnitude is particularly challenging. Most eruptions of volcanoes that we suggest are candidates for VEI 7 will be orders of magnitude smaller (e.g., VEI 2–5). Of the eruptions at Rabaul in 1994, Campi Flegrei in 1538, and Pinatubo in 1991, only the last reached VEI 6, and that just barely. Precursors to even the largest of these, Pinatubo, were unremarkable straight through the onset of VEI 3 eruptions, until just a day before the VEI 6 climactic phase. On the basis of this admittedly small sample, we doubt that scientists will know with confidence the size of an impending eruption until very close to its onset.

Because uncertainties will be high at all but the final steps to eruption, scientists will need to set aside their usual conservatism, and their usual high standards for certainty, and simply make the best calls that they can in close coordination with civil defense officials who will manage mitigative measures. Risks of false alarm may be high, but risks of not taking precautions will be even higher.

POSSIBLE RESPONSES AND PREPARATION

In proximal areas (broadly, within 100 km or striking range of pyroclastic flows), the essential tools of risk mitigation will be short-range forecasts and evacuations. Evacuations are undeniably disruptive and evacuations of hundreds of thousands to millions of people will be enormously complicated (e.g., during evacuation of New Orleans for Hurricane Katrina, as described by Gheytanchi et al., 2007, and Kendra et al., 2008).

The nightmare of volcanologists is the challenge of forecasting a potentially VEI 7 eruption in an urban area and convincing officials to move millions out of harm's way, especially in the periods months and weeks before a possible eruption, while uncertainties are still high (Newhall and Punongbayan, 1996b). Skepticism among officials will be high, and there may be a strong reluctance to accept false alarms, though in our experience scientists are more concerned about false alarms than officials and citizens. During pre-eruption unrest at a volcano thought to be capable of a VEI 7 eruption, decisions must be made about the timing and geographic extent of evacuations. In the history of volcanology, tens of thousands of people have been evacuated for months, and hundreds of thousands for a few days, but never millions for months.

The dilemma is illustrated at Vesuvius and nearby Campi Flegrei, where eruptions of VEI 4 and 5 are used for planning purposes, and where it is hoped that the volcanoes will give at least several days and preferably a week or more of clear warning on which major evacuation can be justified (Dipartimento di Protezione Civile, 2015). Evacuations will be automatically scaled to the planning basis eruption, but scientists and civil defense leaders together must still judge the best timing (Marzocchi and Woo, 2009). Chances are very high that an evacuation decision will need to be made while uncertainty is still high because, as we have seen earlier, the sure signs of an imminent eruption may start just hours or days before the eruption.

At a volcano capable of a VEI 7, the dilemma becomes more acute as scientists and officials much decide not only when to start an evacuation but also how large an area to evacuate. Absent any clear indication that an impending eruption will be of VEI 7 scale, officials are likely to evacuate for a smaller, more probable eruption, perhaps for a VEI 5 or 6. This is an extraordinarily heavy responsibility for scientists and officials, and the public will be affected by whatever decision is made. The larger and less likely a potential eruption, and the higher the stakes, the more difficult will be the evacuation decision. No scientist or official would knowingly put millions of people at risk, but the dilemma is that a decision to evacuate millions may need to be made even while uncertainty is disconcertingly high, with a high risk of false alarm.

Concurrently or before evacuations, other steps can also be taken. Public education is inexpensive and effective. Some may choose voluntary (self) evacuation. Outside the zone subject to pyroclastic flows, and in the event of tephra fall, populations can avoid buildings with large-span roofs that might collapse. In the event of lahars, populations can often move to safety on nearby high ground.

Relatively little can be done to save immobile property (buildings, infrastructure) from pyroclastic flows and lahars, but measures such as filters and temporary shutdowns can minimize damage from ash ingestion into HVAC systems. Mobile property, in contrast, can be moved to safety if scientists can give the right amount of advance warning, neither too much nor too little. Aircraft, trains, and other transport can be put on alert or repositioned, supply chains adjusted, business travel plans rearranged, nuclear power plants put into precautionary safe shutdown, food stockpiles increased, and other contingency plans invoked.

Far from a threatening volcano, responses might wait until after an eruption has actually occurred and its magnitude and likely effects are known with more certainty.

What responses will be needed for ash fall in particular? There is an excellent body of literature about the effects of distal ash on human health, infrastructure, aviation, and other aspects of modern society. Reviews by Horwell and Baxter (2006), Wilson et al. (2012b), G. Wilson et al. (2014), and Craig et al. (2016) provided tips for protective responses. Other tips were provided in a set of posters produced by the Volcanic Impacts Study Group in New Zealand

(2001–2017; http://www.aelg.org.nz/document-library/volcanic-ash-impacts/ and https://www.gns.cri.nz/Home/Learning/Science-Topics/Volcanoes/Eruption-What-to-do/Ash-Impact-Posters), a partner Web site (http://volcanoes.usgs.gov/ash; Volcanic Ash Impacts Working Group, 2017), and pamphlets of the International Volcanic Health Hazards Network (2007; http://www.ivhhn.org). Other experts in agronomy, GPS applications, respiratory function, and myriad other effects, can add details of their experiences and mitigation efforts, both successful and not.

There is a similarly excellent body of literature on adaptation strategies for climate change in general. One can start in the adaptation sections in Intergovernmental Panel on Climate Change (2014), recognizing that the most common effect will be cooling, opposite to current anthropogenic trends. Changes can be abrupt, geographically variable, and, it appears, also last for decades or centuries if effects of multiple eruptions are compounded. Uncertainty challenges any mitigation strategy, but the combination of empirical experience and models is already enough to plan for the most likely scenarios.

CONCLUDING REMARKS

We noted here that the recurrence frequency of a VEI 7 eruption somewhere in the world is between 1 and 2 per thousand years, probably closer to 2 per thousand years. We also noted that pyroclastic flows from VEI 7 could erase an entire city or region, and ash fall and aerosols could have global impacts on air travel, supply chains, and climate. The worst death tolls in historical eruptions are in the tens of thousands (Tanguy et al., 1998; Witham, 2005; Auker et al., 2013), whereas a VEI 7 eruption near an urban area today potentially could kill millions. Some might say that an annual probability of ~2×10^{-3} yr^{-1} is too small, or too daunting, to consider. We disagree. Our societies plan and mitigate for risks lower than this; let's do what can be done, within reason, to plan and prepare for a VEI 7 event.

Volcanologists know how to start, with geologic studies and monitoring of candidate volcanoes. A long-quiet candidate for a VEI 7 might not merit intensive study or monitoring, but at least the basic baseline studies should be done. Climate scientists are already running models of various scenarios, and detailed chronologies of various measures of climate and environmental change are needed to test those models.

Civil defense officials, planners, industries, and others can consult the growing body of information about effects of eruptions on our modern society, and can organize subject experts in interdisciplinary planning, such as is being done in New Zealand in the event of either an eruption in the Auckland Volcanic Field or a large silicic eruption affecting Auckland (Daly and Johnston, 2015; Deligne et al., 2017). The Auckland Volcanic Field is unlikely to produce a VEI 7, but Auckland could easily receive ash from a VEI 7 in the Taupo Volcanic Zone (Lowe, 1988; Newnham et al., 1999). Experts in all fields (health, agriculture, transport, high technology, and more) should review papers cited here and within the pertinent literature, and try to visualize how a VEI 7 eruption as described herein will affect their own sector. For a VEI 6 eruption, regional review may suffice. However, for a VEI 7, reviews need to be both regional and global.

If the next VEI 7 does not occur for another century or more while the pace of technological change continues to quicken, there will many new technologies, and many new impacts, not described here or yet imagined. Although we might not anticipate every effect of the next VEI 7 eruption, we already know enough about the physical characteristics of volcanic phenomena that those who are familiar with new technologies can try to anticipate most new vulnerabilities and how they might be mitigated.

A VEI 7 eruption is not unthinkable. A two in one thousand per year probability (2×10^{-3} yr^{-1}) for an event with major global consequences demands our attention. Forward-thinking governments and industries will be wise to consider how a nearby VEI 7 eruption will affect their jurisdictions, and equally the effects of a VEI 7 that occurs halfway around the world. What responses will make the most economic and sociopolitical sense?

ACKNOWLEDGMENTS

We thank reviewers Steve Sparks, Jake Lowenstern, and Bob Tilling and editor Shan de Silva for many excellent suggestions. Chris Newhall thanks Russell Blong and Aon-Benfield for an invitation that inspired this paper. Alan Robock is supported by National Science Foundation grant AGS-1430051. Alan and Chris met each other and visited volcanoes of the Philippines during their U.S. Peace Corps service in 1970–1972. Stephen Self joined in National Aeronautics and Space Administration–funded research at Dartmouth College that we thought would address volcano-climate links directly, but that turned out to be more useful as shorthand for describing eruptions. Collaboration on this paper brought us back to the early days of our careers; in our 40+ year careers to date, the three of us have had the privilege to witness the Volcanic Explosivity Index (VEI) 5 eruption of Mount St. Helens and the VEI 6 eruption of Pinatubo, and to work on products of VEI 7s and VEI 8s. We hope that the experiences and ideas of this paper will help those who eventually cope with the next VEI 7.

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