Reports

Can Rapid Climatic Change Cause Volcanic Eruptions?

Abstract. *Many major volcanic eruptions coincide with cooling trends of decadal or longer duration that began significantly before the eruptions. Dust veils provide positive feedback for short-term (less than 10 year) global cooling, but seem unlikely to trigger glaciations or even minor climate fluctuations in the 10- to 100-year range. On the contrary, variations in climate lead to stress changes on the earth's crust—for instance, by loading and unloading of ice and water masses and by axial and spin-rate changes that might augment volcanic (and seismic) potential.*

A causal relationship has frequently been suggested between volcanic eruptions and periods of climate cooling on time scales ranging from a few years (*1–4*) to millions of years (*5–7*). For example, Lamb (*2*) correlated periods of low temperature during the last several hundred years with atmospheric volcanic dust loading, which was derived in part from climatic data. Thus, some of the proposed correlations (*2, 3, 8*) may be artifacts.

On time scales of 10^3 to 10^6 years, Bray (*7, 9, 10*) suggested a correlation between supposed volcanically active periods ("volcanic pulses") and glacial advances. However, according to Bray's data, within the past 30,000 years some pulses of glaciation began 300 to 700 years before the volcanic events that supposedly triggered them, and some other glacial advances began up to 1000 years after the volcanic pulse. During the past 2 million years, major ice-sheet expansions have lagged behind so-called volcanic triggering events by up to 10,000 years (*7, 11*).

Studies of climatic records and explosive volcanic eruptions over the past 200 years indicate that possible volcanically induced temperature deteriorations have only about the same range as background variations (< 1°C), even though the data cover a wide range of eruption magnitudes (*12–15*). Furthermore, volcanically induced cooling has a lifetime usually of 1 to 3 years and certainly of not more than 7 years (*15*). If Bray's interpretations are correct, evidence for glacial advances might well be expected after the large eruptions of the late 19th and early 20th century, but this period was marked rather generally by glacial retreats (*16*).

Table 1. Time distribution of known volcanic eruptions of large magnitude within the last 100,000 years. Large magnitude eruptions are loosely defined as those ejecting large volumes of usually silicic magma and pumice (bulk deposit volume ≥ 10 km^3 and often > 50 km^3) or an outstandingly large volume compared to those of other eruptions in a well-known sequence (for example, New Zealand and Japan). Several of the deposits listed are known from both deep-sea cores and land deposits.

Actual date (A.D.)	Age (years B.P.)	Dating method*	Volcano or eruption name	Region of source volcano	Bulk volume of deposit (km^3)	Reference
1912	38	Historical	Katmai	Alaska	12–15	(*40*)
1883	67	Historical	Krakatau	Indonesia	~18	(*41*)
1815	135	Historical	Tambora	Indonesia	≤150	(*17*)
	1,250	^{14}C	White River (east lobe)	Alaska	~20	(*42*)
260	1,690	^{14}C	Ilopango (Tierra Blanca)	Central America	20–30?	(*43*)
	1,820	^{14}C	Taupo Pumice Formation	New Zealand	>30	(*44*)
79	1,871	^{14}C; archeology	Vesuvius	Italy	≤10	(*45*)
	1,890	^{14}C	White River (north lobe)	Alaska	>25	(*42*)
	2,000	^{14}C	Sunda	Indonesia	>10?	(*46*)
	3,450†	^{14}C	Santorini (Minoan)	Aegean Sea	>40	(*47*)
	3,500†	^{14}C; archeology	Vesuvius (Z1 layer)	Italy	~40	(*45*)
	3,800†	^{14}C	Waimihia	New Zealand	⩾15	(*44*)
	6,900†	^{14}C	Akahoya	Japan	>50 (?150)	(*48*)
	7,340†	^{14}C	Mazama (Crater Lake)	United States	>42	(*49*)
	18,500	^{18}O-biostratigraphy	Santorini (Y2 ash)	Aegean Sea	>20	(*45*)
	20,500	^{14}C	Oruanui Formation	New Zealand	≈150	(*50*)
	21,000–22,000	^{14}C	Aira-Tn and Ito pyroclastic flow	Japan	>225	(*48*)
	24,500 ± 900	^{14}C	Ata	Japan	>100?	(*51*)
	28,400	^{14}C	Roseau (Dominica)	Caribbean	>40	(*52*)
	30,000–32,000	^{14}C	Shikotsu	Japan	>50	(*48*)
	30,000	^{14}C	Mangaeone	New Zealand	>20	(*53*)
	30,000	^{14}C	Daisen Kurayoshi	Japan	>50?	(*48*)
	30,000–35,000	^{14}C	Campanian Tuff-Ischia (Y5)	Mediterranean	>70	(*45*)
	41,700	^{14}C	Rotoehu-Rotoiti ash	New Zealand	>100	(*54*)
	70,000–90,000	Fission track	Ontake	Japan	?	(*48*)
	75,000	^{18}O-biostratigraphy	Toba (Sumatra)	Indonesia	~2000	(*30*)
	75,000	^{18}O-biostratigraphy	Y6 ash (Gulf of Mexico)	Caribbean	>50?	(*22*)
	85,000 ± 5,000	^{18}O-biostratigraphy	Los Chocoyos (Y8 ash, Gulf of Mexico) = D ash (Eastern Pacific)	Caribbean	>400?	(*29*)

*The clustering of ^{14}C dates around 30,000 years before present (B.P.) may be an artifact of the method. †Corrected to sidereal years.

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As a counterargument, it could be claimed that recent eruptions have only been of small to moderate size (*1*) or occurred at the wrong time or in the wrong location to induce large-scale cooling (*7*). Much larger eruptions are known to have occurred during the Pleistocene and are claimed to have been adequate to initiate massive climatic deterioration (*7*). However, the eruption of Tambora in 1815, one of the largest eruptions during the past few thousand years (*17*), is associated with a hemispheric temperature decrease of only 0.5° to 1°C for 2 to 3 years (*13, 15, 18*). In this case, average global temperatures had already been decreasing since 1810 (*13, 19*) and then rose again in the 1820's. The year after the Tambora event, 1816, was the "year without summer" in the northeastern United States and northwestern Europe (*2, 20*). It is true that minor glacial advances occurred about this time, but they began before the great eruption (*16*, p. 462). Generally ignored, however, is the fact that the decade 1810 to 1820 coincided with the most pronounced low in the mean sunspot record of the last 250 years, indicating a lower solar ultraviolet output; it was also a time of low magnetic intensity. Thus there are other possible explanations for this climatic cooling (*13, 21*).

A record of the largest Quaternary eruptions is found in ash layers preserved in deep-sea cores. Correlation of these ash horizons with glacial epochs has been proposed (*6, 7, 22*) but has been contested by Ninkovich and Donn (*23*).

To test for a correlation between major explosive eruptions and climatic changes more systematically, we compiled a list of the largest known eruptions of the past 10^5 years (Table 1). We compared this chronology (*24*) with well-dated records of climate on three different time scales (Fig. 1): (i) 10^5 years, compared with major changes in North Atlantic Ocean temperatures from deep-sea cores (*25*); (ii) 10^4 years, compared with sedimentologic indicators of climatic variations as compiled from selected deep-sea cores with high rates of deposition (*26*); and (iii) 10^2 years, compared with instrumental temperature data (*27*).

We note an apparent coincidence, on all three time scales, of numerous large explosive volcanic events with climate cooling; however, many of the cooling trends were apparently initiated before the explosive volcanic eruptions. Many cooling intervals show no association with the large explosive eruptions listed in Table 1. We wish to make it clear that uncertainties in the geological dating of many volcanic and climatic events make any such correlation tentative. Previous workers have suggested that explosive volcanic events precede and cause climate change. We suggest that errors in dating and poor resolution make it difficult to determine which phenomenon leads the other, and we believe that a case can equally be made for the converse relationship—that is, climatic change as a trigger for some explosive volcanic eruptions.

Large explosive eruptions, defined as those that eject > 10 km³ of volcanic ash, may occur as close together as 29 years (for example, 1883 to 1912 in Table 1), but very large events (> 50 km³) occur globally only once every 1000 to 5000 years. They are usually of the plinian and ignimbrite-forming types (*28*) that inject

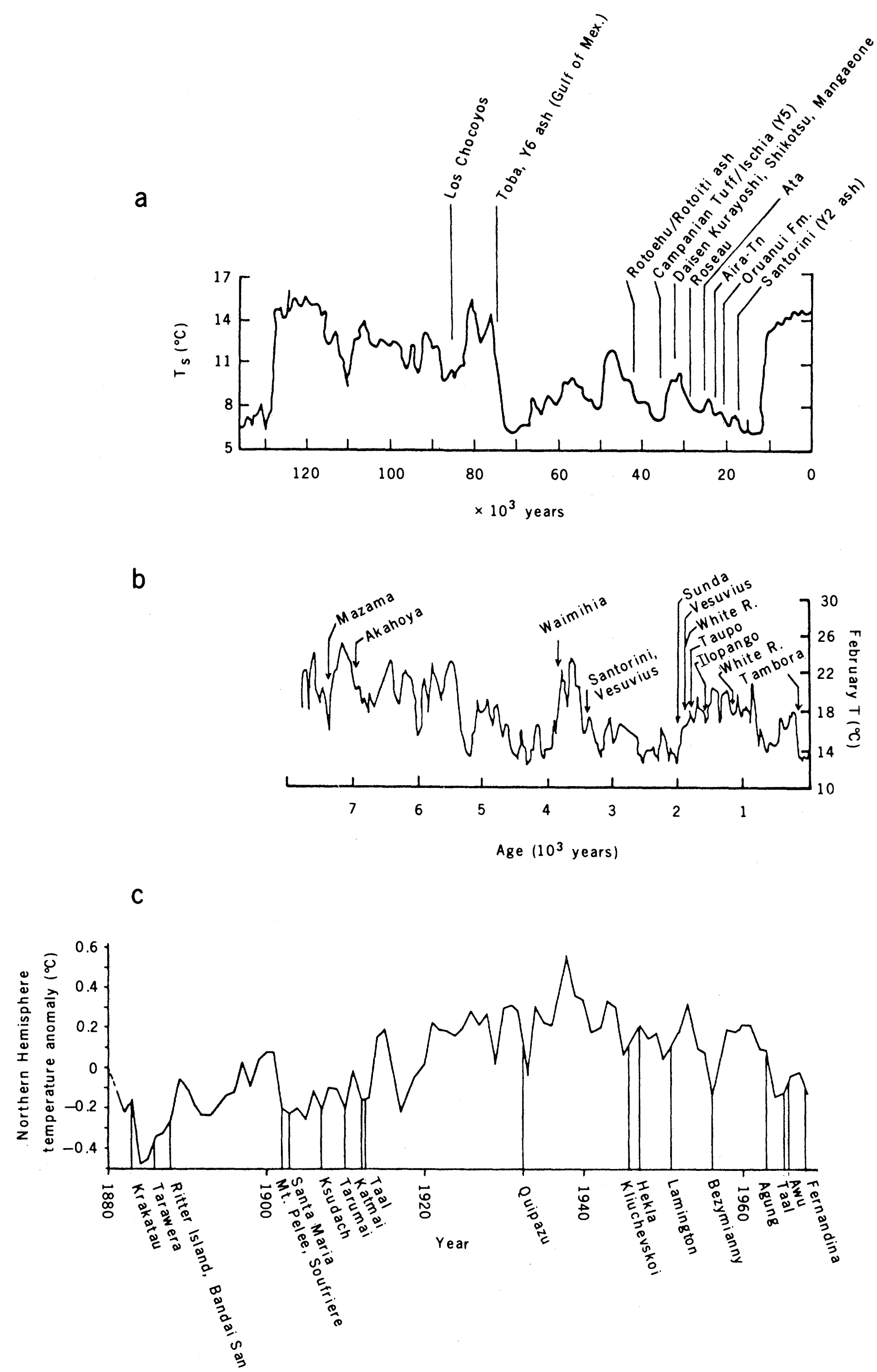

Fig. 1. (a) Faunal index for estimated summer sea-surface temperature in the North Atlantic during the past 130,000 years (*25*). (b) Paleotemperatures for the month of February in the Santa Barbara Basin during the past 8000 years (*26*). (c) Northern Hemisphere temperature anomalies and timing of the largest eruptions from 1880 to 1970. The temperature curve is from Budkyo (*27*), the eruption list from Newhall and Self (*55*).

large quantities of fine dust and gas into the stratosphere.

Can such posteruption stratospheric dust veils really cause sufficient climatic cooling to trigger glaciation? Although an eruption as large as Tambora (possibly > 100 km^3) probably has an effect on global climate, it is only for a few years. No geological evidence has been found to suggest that events of such magnitude ever occur frequently enough (less than ~ 10 years apart) to have a cumulative effect in depressing global temperatures. It is thus difficult to see how even very large eruptions could lower temperatures long enough to trigger glaciation, although certain climatic regimes may be more sensitive than at present to the slight perturbing effects of volcanic aerosol input into the upper atmosphere (*7*).

Although some of the geological dates might well be questioned, the instrumental temperature records for the past 200 years (Fig. 1) show that the major historical eruptions associated with coolings actually occurred after decadal length temperature decreases had been initiated; examples are Tambora in 1815, Krakatau in 1883, the 1902 series of eruptions (including Santa Maria and Mount Pelée), or Katmai in 1912 (*13*).

On the 10^5-year time scale, widespread Pleistocene ash layers in cores from the Gulf of Mexico [probably derived from Central American volcanoes (*29*)] are consistently younger than the initiation of sharp temperature decreases indicated by biostratigraphic studies in the same cores (*22*). These particular volcanic eruptions, therefore, could not have triggered the coolings.

Probably one of the greatest single explosive events of the last 2 million years was the eruption of Toba in Sumatra; its vast ash shower extended from Malaysia to India (*30*). Dated at 75,000 years ago, it apparently followed the initiation of the cooling that began with the end of the Brørup (Saint Pierre) interstadial by several thousand years.

Recognizing that some large volcanic eruptions do occur at times of global cooling, it seems reasonable to ask if the coolings and glaciations in some way trigger the explosive eruptions, or if some third factor affects both climate and volcanism.

Several possible mechanisms for climatic triggering of volcanic eruptions have been proposed. For example, the redistribution of water that accompanies glaciation and deglaciation gives rise to both hydroisostatic and glacioisostatic readjustments. Asymmetric mass loading, as on Greenland, requires that the global spin axis adjust to the new symmetry of mass. In this century, the mean pole movement is 6 cm (0.002 arc second) annually along the meridian 77°W (*31*). The rate of ice loading is far more rapid than that of isostatic readjustment. Realignment of the geoid will then lead to worldwide stress that might trigger eruptions (*32*). Crustal adjustments will be most active along the plate margins and major lineament-fault intersections. Volcanic eruptions may be expected along active subduction zones in island arcs; most of the eruptions listed in Table 1 are, in fact, from island arc volcanoes. Chappell (*33*) noted that the stress gradients associated with mantle flow beneath continental margins in response to glacial loading and unloading are 10^5 times larger than the stress gradients associated with earth tides, which are known to affect the timing of some earthquakes and volcanic eruptions (*34*).

According to the mechanism proposed by Matthews (*35*), hydrostatic unloading of the ocean basins during glaciation favors upward movement of basaltic magma and thus might lead to volcanism. During deglaciation the sea floor is depressed, with corresponding upwarp and shift of stress to the continents and islands.

Although isostatic effects need not create volcanism, they may well trigger subcritical magma bodies by reactivation of faults and perhaps through the mechanism of magma mixing (*36*). Injection of basaltic magma into rhyolitic magma chambers would favor rapid and possibly turbulent convection. Some large-volume tephra sheets apparently record an early stage of mixing of these basaltic and rhyolitic magmas.

Since it is now rather well established that the timing of major glaciations is correlated with orbital variations (*37*), it may also be postulated that the body tides set up by planetary motions may be triggers for some large explosive volcanic eruptions. Thus, both pulses of volcanism and climate change could be related ultimately to extraterrestrial mechanisms.

We realize that a possible link between intervals of climatic cooling and glaciation and great volcanic eruptions on such different time scales as decades to tens of thousands of years must have multiple and probably overlapping causal mechanisms. Variations in the output of solar radiation (especially ultraviolet and corpuscular radiation) have been suggested as a possible cause of climate fluctuations on the order of decades (*21*, *38*). If these short-term variations of solar activity are related to cycles of planetary tides on the sun, as some have suggested (*39*), it may be that comparable planetary tidal stresses on the earth provide the triggering mechanism for the explosive volcanic events. Changes in atmospheric circulation caused by the solar variations induce additional crustal stresses. In such cases, a coincidence of cool climate and some large explosive eruptions might not be related in a direct manner, but could be a result of common underlying causes.

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B1-B2 Transition in Calcium Oxide from Shock-Wave and Diamond-Cell Experiments

Abstract. *Volume and structural data obtained by shock-wave and diamond-cell techniques demonstrate that calcium oxide transforms from the B1 (sodium chloride type) to the B2 (cesium chloride type) structure at 60 to 70 gigapascals (0.6 to 0.7 megabar) with a volume decrease of 11 percent. The agreement between the shock-wave and diamond-cell results independently confirms the ruby-fluorescence pressure scale to about 65 gigapascals. The shock-wave data agree closely with ultrasonic measurements on the B1 phase and also agree satisfactorily with equations of state derived from ab initio calculations. The discovery of this B1-B2 transition is significant in that it allows considerable enrichment of calcium components in the earth's lower mantle, which is consistent with inhomogeneous accretion theories.*

Calcium oxide, initially in the B1 (NaCl type) structure, is expected to transform to the B2 (CsCl type) structure at high pressure, by analogy with the B1-B2 transitions found in alkali halides (*1*, *2*). There is considerable interest in such transitions both in theoretical studies of oxide structures (*3*–*5*) and because of their possible occurrence in the earth's lower mantle. We have carried out shock-wave experiments on CaO to determine its equation of state (Hugoniot) (*6*), as well as x-ray diffraction, under static high pressures, through a diamond cell (*7*); both techniques demonstrate a B1-B2 transition in CaO at 60 to 70 GPa (0.6 to 0.7 Mbar). To our knowledge this is the first documentation of the B2 structure in an oxide of direct geophysical interest.

Our new data are given in Table 1 and Fig. 1 along with previous results for CaO (*8*). For comparison, theoretical Hugoniots based on finite strain theory and ab initio [modified electron gas (*4*)] calculations for the B1 and B2 phases are also given (*9*). The shock data provide an accurate dynamic compression curve for CaO and the x-ray data from the diamond cell confirm the nature of the structural transition. The diamond-cell experiments indicate that the transition begins at 60 (±2) GPa on the ruby-fluorescence pressure scale (*7*) and at room temperature (295 K), whereas the shock-wave data indicate a slightly higher transition pressure: about 63 to 70 GPa, but at approximately 1350 K (*10*). A volume decrease of 11 ± 1 percent is found at the transition in both sets of experiments, which is in agreement with simple systematics among the data for B1-B2 transitions in halides (*11*) and provides additional support for applying such systematics to oxides. The consistency of the shock-wave and diamond-cell results for the transition pressure in CaO provides an approximate but independent confirmation of the ruby-fluorescence scale calibration at about 65 (±5) GPa. Thus the B1-B2 transition in CaO may be a convenient and readily reversible pressure-calibration point for ultrahigh-pressure static experiments.

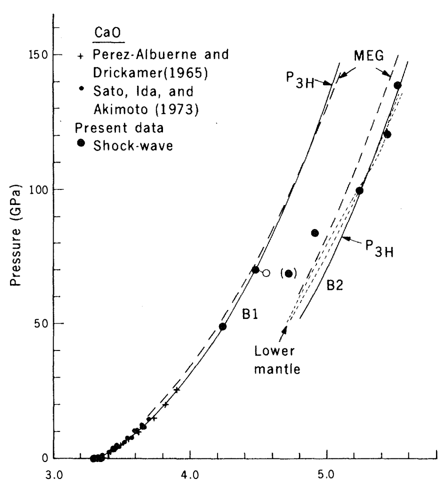

Fig. 1. New shock-wave data for CaO compared with previous, static compression data (*8*) and theoretical Hugoniots based on finite-strain (P_{3H}) and ab initio [modified electron gas (*MEG*)] calculations (*9*). Also shown are two seismological models for the lower mantle (*15*). Error bars for the shock-wave data are approximately the size of the symbols (or smaller); the datum in parentheses is considerably less certain. The open symbol represents an alternative interpretation of the 70-GPa result (*6*).

The shock-wave data corresponding to the B1 phase are in excellent agreement (Fig. 1) with the theoretical Hugoniot calculated from recent ultrasonic data for CaO (*9*) and compare favorably with the theoretical Hugoniot based on the modified electron gas theory (*4*). Hence, the equation of state of CaO in the B1 structure appears to be very well constrained both experimentally and theoretically. Although the ab initio results underestimate somewhat the density of the B2 phase, they predict its compressional behavior quite well. Our Hugoniot data are consistent with essentially iden-