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Climatic Impact of Volcanic Eruptions

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Volcanoes and Climate

Connections between volcanic eruptions and climatic change have been suggested on timescales from days to 10^8 years. On short timescales, one approach to the volcano/climate problem has been to compare the times of historic volcanic eruptions with changes in yearly, monthly or seasonal surface temperatures on regional, hemispheric and global scales. These studies involve either a direct comparison of the times of significant eruption years with temperature records (e.g., Rampino and Self, 1982, 1984; Stothers, 1984; Angell and Korshover, 1984; Mass and Portman, 1989) (Table 1) or a study of composited temperature records for a number of years (months, seasons) before and after a chosen set of eruptions (e.g., Mass and Schneider, 1977; Self et al., 1981, Taylor et al., 1980). Eruptions are usually chosen using various measures of volcanic intensity including the VEI (Volcanic Explosivity Index) of Newhall and Self, 1982, or the DVI (Dust Veil Index) after Lamb, 1970, or on the basis of the stratospheric aerosol loading determined directly by observations or indirectly from the acidity record of ice cores. The various studies give similar results—the composites show a Northern Hemisphere cooling of 0.2 to 0.3°C for 1 to 3 years after eruptions for a number of eruptions grouped together (Fig. 2), and individual volcanic events that produced significant aerosol clouds such as Krakatau, 1883 or Tambora, 1816 are followed by Northern Hemisphere coolings of 0.3 to 0.7°C for 1 to 3 years after the eruption (Table 1) (Baldwin et al., 1976; Rampino and Self, 1984; Stothers, 1984; Angell and Korshover, 1985). Zonally, the cooling is amplified at high latitudes. Regional records show more variability , especially meridionally.

Bradley (1988) has used monthly and seasonal temperature data, and finds that several of the larger eruptions of the past 100 years are followed by significant negative anomalies in summer and fall temperatures. Temperature decreases after major eruptions are found to be abrupt and short lived (1 to 3 months), with a recurrence of the cooling about 12 and 24 months after the eruptions. The maximum effect of about 0.4°C occurs in the summer and fall months immediately following the eruptions, and falls off in the same seasons over the next 2 years.

Mass and Portman (1989) recently suggested that the volcanic signal was present in temperature records, but smaller than previously thought. It was limited to those eruptions that created the densest aerosol clouds in the last 100 years, and was enhanced by subtracting out other sources of interannual variability, e.g. the El Nino/Southern Oscillation. Mass and Portman stressed that the volcanic "signal" of 0.1 to 0.2°C is of the same order as "background" temperature variations in non-volcanic years, and found no evidence of large coolings in the first few months after the eruptions. It may be, however, that stratospheric aerosol clouds have some effect on the ENSO phenomena, either triggering them, or intensifying already existing ENSO patterns (Handler, 1984). Handler (1986) has also suggested a connection between stratospheric aerosols and the strength of the yearly Indian monsoonal precipitation .

These studies have attempted to "isolate" the volcanic signal in noisy temperature data. Thyis assumes that it is possible to isolate a distinct volcanic signal in a record that may have a combination of forcings (ENSO, solar variability, random fluctuations, volcanism) that all interact. The key to discovering the greatest effects of volcanoes on short-term climate may be to concentrate on temperatures in regions where the effects of volcanic aerosol clouds may be amplified by perturbed atmospheric circulation patterns. This is especially true in sub-polar and mid-latitude areas affected by changes in the position of the polar front. Such climatic perturbations can be detected in surface temperatures and in proxy evidence such as decreases in tree-ring widths and frost rings, changes in the treeline,weather anomalies such as unusually cold summers, severity of sea-ice in polar and sub-polar regions, and poor grain yields and crop failures (for a review see Rampino et al., 1988). In low latitudes, sudden temperature drops have been correlated with the passage overhead of the volcanic dust cloud (Stothers, 1984). For some eruptions, such as Tambora, 1815, these kinds of proxy and anectdotal information have been summarized in great detail in a number of papers and books (e.g., Post, 1978; Stothers, 1984; Stommel and Stommel, 1986; C.R. Harrington, in press). These studies lead to the general conclusion that regional effects on climate, sometimes quite severe, may be the major impact of large historical volcanic aerosol clouds.

Tambora and 1816: The Test Case?

Instead of searching for the small climatic effects of small historic eruptions in climate data sets, it may be instructive to look in detail at the strongest volcanic perturbation in recent history, the Tambora eruption of April, 1815 and the events that followed it. Perhaps smaller eruptions (in terms of aerosols released) might have similar, but less extreme, effects on climate that can be detected if one knows what to look for. Stothers (1984) estimates an optical depth of about 1.0 following the Tambora eruption, and the presence of large acidity peaks in ice cores from Greenland and Antarctica argues for a global disperesal of the aerosol cloud of some 10^{14} g. In most studies of weather records or proxy data, the year 1816, following the Tambora eruption, displays the strongest signal of possible "volcano weather" in historical times. For example, in a key area in the Eastern Hudson Bay region in midsummer 1816, the reduction in mean daily temperature from the long-term average was about 5 to 6°. It has been suggested that a reduction in mean daily temperature in this range could lead to the formation of perennial snowcover in northern Canada (Wilson, 1985). The negative anomaly in July 1816 brought the absolute mean temperature to below 5°C, more than 2° lower than the modern record in 1965. The median height of the freezing level above eastern Hudson Bay in July 1965 (two years after the Agung eruption) was about 600 to 700 m lower than the modern 10 year value. This suggests that the -6°C anomaly in 1816 might have produced a 1000 m drop, to bring the freezing level to within 1500 m of the surface. Such conditions were maintained for only two years, 1816 and 1817, but one may wonder how many years of consecutive extreme seasonal weather would be required to establish an effective snowcover, and bring ice/albedo feedback into play. Frequent and extended snowfall also took place in the region in 1816-1817— in 1816, there were only five weeks without measurable snowfall (mid-July to mid-August). Furthermore, the atmospheric circulation pattern was dominated by a high pressure area over the surface of Hudson Bay, which remained ice covered through the summer of 1816 (Wilson, 1985).

Catchpole and Faurer (1983) show sea ice patterns in 1816 that are also consistent with a highly meridional atmospheric circulation pattern over eastern North America, showing strong north to northwest winds in July and August of 1816 (Catchpole and Faurer, 1983). Northwesterly winds brought cold air southward into the northeastern United States, bringing the "Year without a Summer". The meridional circulation pattern also brought cold weather to western Europe, and there is some evidence that, between these two waves, warmer weather dominated, with an opening of the usually ice-covered Greenland Sea between 74° and 80°N (Wilson, 1985).

It should be pointed out however, that the cold weather in the Hudson Bay area began before April 1815, with unusually cold years beginning in 1811/12, so that the Tambora eruption cannot be the only cause of the climate shift. Perhaps the coincidence of the

Tambora eruption with a time of low sunspot numbers (indicating lower solar activity) led to the unusually cold weather of the 1811-1817 period.

Frost rings in trees in the Western U.S. also show a correspondence with years of volcanic eruptions (LaMarche and Hirschboeck, 1984). These may indicate outbreaks of exceptionally cold weather during the summer months, again possibly the result of increased meridional circulation. Light rings in subarctic trees from northern Quebec also show a correspondence with volcanic eruptions; for example, damage is most widespread in 1816-1817 (Filion et al., 1986). Tree rings have also been used to create synoptic summer temperature patterns for Europe, which show significantly cooler summers from 1812 to 1816 (Briffa et al., 1988). The same kinds of data can be assembled for the other major volcanic aerosol clouds of the last 200 years, and similarities noted. If one can establish patterns that appear to be characteristic of "volcanic" perturbation, then it may be possible to detect such patterns even for the smaller eruptions, and to see how well climate models simulate such perturbations. It may be necessary not only to model the volcanic perturbation, but also to include the effects of ENSO events and possible solar variations in the model runs.

VOLCANISM AND LONG-TERM CLIMATE CHANGE

One problem with studying historical eruptions is that they are quite small compared to those in the geologic record. The largest explosive eruption in historic times was probably the Tambora, 1815 explosion, with about 50 km^3 of erupted magma. This can be compared with large ignimbrite forming eruptions such as the Toba event of 75,000 yr BP, which erupted more than 2,800 km^3 of magma (Rose and Chesner, 1987). For effusive eruptions, the largest historic event was Laki, 1783, which produced 12 km^3 of basalt, compared with great flood basalt eruptions like the Roza flow in the Columbia River Group (14 Myr BP), composed of 700 km^3 of basaltic lava in a single flow (Devine et al., 1984; Rampino et al., 1988). Both of these eruptions may have released about 10^{16} g of sulfurous gases. The effects of these and other pre-historic eruptions on climate are not known. Both the Roza and Toba events appear to coincide with climatic coolings, but no cause and effect relationship has been established.

If historic eruptions can cause small changes in climate, then perhaps larger eruptions or groups of eruptions can cause major climate change. A number of authors have claimed correlation between volcanic eruptions and glacial fluctuations on 10^3 to 10^5 year time scales. For example, Bradley and England (1978) and Porter (1981,1986) proposed that second-order "Little Ice Age" glacial advances during the Holocene could have been driven by eruptions, since the glacial advances correlate with peaks of acidity in polar ice cores. Porter (1981) estimates that a global cooling of 1°K could lead to a snowline depression sufficient to cause glacier advances equivlent to those of the last several centuries, but this seems to be greater than the measured volcanic perturbations. Bray (1977, 1979a, b) proposed that large eruptions preceded glacial periods, with the glaciations triggered by atmospheric aerosol clouds. However, in this case, because of the inaccuracies in dating, it is often difficult to determine which came first, the eruptions or the glaciation.

On even longer timescales, studies by Kennett et al (1977) have delineated pulses of explosive volcanic activity in the Circum-Pacific region for the last 30 million years from studies of age determinations (mostly K/Ar dates) on igneous rocks, and counts of ash layers in deep-sea sediment cores. They find significant pulses of volcanism in the Plio-Pleistocene (about 2 Myr BP), latest Miocene to Early Pliocene (6 to 3 Myr BP), Late Miocene (11 to 8 Myr BP) and Middle Miocene (16-14 Myr BP). Combined with other data (e.g., Hein et al., 1978), it seems that pulses of widespread (perhaps global) explosive volcanism took place near 0.5, 2.5, 5, 10, 15, 20 and 40 Myr BP (Kennett et al., 1985). Some of these spurts in volcanism correlate with times of apparent pulses of plate motion and sea level changes (Masuda, 1986; Rampino and Stothers, 1987), so the three may be linked. These volcanic pulses can be also be correlated with times of global cooling and ice formation as seen in oxygen-isotope data for the Cenozoic. Flood basalt episodes at 65, 35, and 17 Myr BP also correlate with global coolings (Devine et al., 1984). This apparent correlation of volcanic episodes with times of global cooling may be coincidental, however, and there is no direct cause and effect mechanism to suggest beyond possible ice/albedo feedback.

Recommendations

1. Look at the strongest aerosol perturbations, e.g. Tambora, Krakatau, Agung, El Chichon, etc. Don't average together the effects of small and large eruptions.
2. Use regional and seasonal climate data sets where the effects of volcanism on climate may be amplified.
3. Study the spread of aerosol clouds for volcanic eruptions in different parts of the world at different times of the year. Compare with tracer models.
4. Look for changes in atmospheric circulation patterns that could be studied with GCMs. Observed temperature series and tree-ring studies could provide synoptic weather patterns.

5. Establish a new volcanic index based on VEI and data from ice cores, petrologic studies, optical depth measurements, etc.

6. Look at climate records that have both climate and volcanic signals, e.g. ice cores.

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