# Volcanic winter in the Garden of Eden: The Toba supereruption and the late Pleistocene human population crash

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# Volcanic winter in the Garden of Eden: The Toba supereruption and the late Pleistocene human population crash

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## ABSTRACT

The Toba eruption (ca. 73,500 B.P., Indonesia) was the largest explosive eruption of the last few hundred thousand years. Several lines of evidence suggest that Toba produced an estimated $10^{15}$–$10^{16}$ g of stratospheric dust and $H_2SO_4$ aerosols, and ice-core data and atmospheric modeling indicate a ca. 6-yr residence time for the dense global aerosol cloud. Such a stratospheric aerosol loading is predicted to have caused a 'volcanic winter' with possible abrupt regional coolings of up to 15°C (similar to nuclear winter scenarios), and global cooling of 3–5°C (and possibly greater) for several years. Ice-core data suggest that Toba may have contributed to the initial severe cooling of a millennium-long cold event, suggesting involvement of climate feedback responses such as ocean cooling, and increased sea ice and snow cover.

Botanical studies and model simulations suggest that the local and regional effects of the predicted post-Toba cooling would have been disastrous for vegetation. Cold-sensitive tropical vegetation would have been almost totally destroyed, and the predicted reduction of early growing-season temperatures by ≥10°C in higher latitudes could have killed most temperate and subarctic forests, with surviving vegetation severely damaged and recovery times taking decades. Global climate models predict that the Toba atmospheric perturbation would have caused severe drought in the tropical rainforest belt and in monsoonal regions. These results constitute a global ecological disaster, with expected reductions in standing crops of plants and animals especially in the tropics. Evidence for these abrupt environmental changes may be detectable in high-resolution palynological records, coral reefs, and ice cores.

Genetic studies indicate that sometime prior to ca. 60,000 yr ago humans suffered a severe population bottleneck (possibly only 3,000–10,000 individuals), followed eventually by rapid population increase, technological innovations, and migrations. The climatic effects of the paroxysmal Toba eruption could have caused the bottleneck, and the event might have been a catalyst for the technological innovations and migrations that followed. The present results as to the predicted environmental and ecological effects of the eruption lend support to a possible connection between the Toba event and the human population bottleneck, and suggest that similar bottlenecks among other organisms might be expected at about the same time. Some chimpanzee populations appear to have undergone such a bottleneck ca. 70,000–60,000 yr ago.

## INTRODUCTION

The Toba eruption, represented by the voluminous Younger Toba Tuff rhyolitic ignimbrite of Sumatra, Indonesia, and a very widespread ash fall in the Indian Ocean, is one of the greatest known volcanic events in the geologic record (Ninkovich et al., 1978; Rose and Chesner, 1990; Rampino and Self, 1993a). The exceptional size of the Toba eruption makes it an important test case of the maximum possible effects of explosive volcanism on the global atmosphere and climate (Rampino et al., 1988; Rampino and Self, 1992, 1993a). Recently, Ambrose (1998a), Gibbons (1993), and Rampino and Self (1993b) pointed out the possible similarity in timing of the Toba event and a putative marked reduction in the population of human ancestors based on genetic studies, and suggested a connection with the environmental aftereffects of the eruption. This 'bottleneck" in human population, estimated by some to have involved a reduction to less than 10,000 individuals for a period of up to 20,000 yr (Wains-coat et al., 1986; Jones and Rouhani, 1986; Harpending et al., 1993; Gibbons, 1993; Rogers and Jorde, 1995; Ayala and Escalante, 1996), was eventually followed by a population explosion and migrations of modern humans. New information on the atmospheric effects of the Toba eruption from ice-core studies (Zielinski et al., 1996a, b), on climate changes at the time of the eruption, and a review of the inferred environmental effects of large increases in global atmospheric opacity derived from nuclear winter models (e.g., Harwell and Hutchinson, 1985) allows a revised estimate of the climatic impact of Toba and possible effects on the environment and human population.

## TOBA ERUPTION AND ITS ATMOSPHERIC IMPACT

The Toba ignimbrite deposits have been dated by the K/Ar method at 73,500 ±3,500 B.P., and $^{40}Ar/^{39}Ar$ age determinations give 73,000 ±4,000 B.P. (Chesner et al., 1991). The Toba ash layer occurs in deep-sea cores from the Indian Ocean at the time of the Oxygen Isotope Stage 5a-4 transition, estimated at 73,910 ±2,590 B.P. The widespread co-ignimbrite ash layer has a dense rock equivalent volume (DRE) of about 800 km³ (Chesner et al., 1991). The ignimbrites on Sumatra, consisting of a few very thick pyroclastic flows, have a volume of ~2,000 km³ DRE (Chesner et al., 1991; Rose and Chesner, 1990), for a total eruption volume of ~2,800 km³ (DRE). Woods and Wohletz (1991) estimated Toba eruption cloud heights of 32 ±5 km, and the duration of continuous fallout of Toba ash over the Indian Ocean has been estimated at two weeks or less (Ledbetter and Sparks, 1979).

Release of sulfur volatiles is especially important for the climatic impact of an eruption, as these form sulfuric acid aerosols in the stratosphere. Although the intrinsic sulfur content of rhyolite magmas is generally low, the great volume erupted is sufficient to give an enormous volatile release. Based on studies of the sulfur content of the Toba deposits, Rose and Chesner (1990) estimated that ~3 × 10¹⁵ g of $H_2S/SO_2$ (equivalent to ~1 × 10¹⁶ g of $H_2SO_4$ aerosols) could have been released from the erupted magma. The amounts of fine ash and sulfuric acid aerosols that could have been generated by Toba was estimated independently

using data from smaller historical rhyolitic eruptions (Rampino and Self, 1992) as up to 2 × 10¹⁶ g of fine (<2 micron) dust and ~1.5 × 10¹⁵ g of sulfuric acid aerosols.

Optical depth is defined as the negative natural logarithm of the attenuation of incident sunlight, $\tau = -\ln(I/I_0)$; where $I_0$ and $I$ are the initial and final light intensity, respectively. In the case of volcanic $H_2SO_4$ aerosols, $\tau = 6.5 × 10^{-15} M$, where $M$ is the global aerosol loading in grams (Stothers, 1984a, b). Aerosol optical depths can be compared with the smoke optical depths estimated for nuclear winter scenarios; however, for volcanic aerosols and dust, $\tau$ is largely dependent on the single scattering albedo ($\omega$) of the cloud, whereas for soot (smoke) clouds, the absorption optical depth $\tau_a$ is primarily a function of the high absorptivity of the soot particles in the visible (fluffy soot is also highly absorptive in the infrared). For volcanic $H_2SO_4$ aerosols and dust, $\omega$ is ~1, and therefore, formally, $\tau_a$ is ~0.

Physical and chemical processes in dense aerosol clouds may act in a 'self-limiting" manner, significantly reducing the amount of long-lived $H_2SO_4$ aerosols (Rampino and Self, 1982; Pinto et al., 1989). Using one-dimensional aerosol microphysical and photochemical models, Pinto et al. (1989) showed that for an aerosol cloud of ~10¹⁴ g of $SO_2$, condensation and coagulation are important in producing larger sized particles, which have a smaller optical depth per unit mass and settle out of the stratosphere faster than smaller particles. In the 10¹⁴ g $SO_2$ simulation of Pinto et al. (1989), aerosol optical depth ($\tau$) reaches a peak of 1.3 within five months, with a lifetime of about two years, in good agreement with observations for the Tambora aerosols (see the section 'Toba and Volcanic Winter"). However, the maximum sulfur volatile emission that Pinto et al. modeled was 2 × 10¹⁴ g of $SO_2$, and no data exist on the behavior of $H_2SO_4$ aerosols in >10 times denser clouds.

Another possible limitation on aerosol loading is the amount of water in the stratosphere available to convert $SO_2$ to $H_2SO_4$. Stothers et al. (1986) calculated that ~4 × 10¹⁵ g of water might be available in the ambient stratosphere, and injection into the stratosphere of up to 5.4 × 10¹⁷ g of $H_2O$ from Toba is possible (Rose and Chesner, 1990), more than enough water to convert the sulfur gases emitted by Toba into $H_2SO_4$ aerosols. In previous studies, we somewhat conservatively estimated that only ~10% of the total amount of aerosols possible would have actually formed in the Toba cloud, or ~1 × 10¹⁵ g of $H_2SO_4$ aerosols (Rampino and Self, 1992, 1993a).

### Toba and long-term climatic change?

Within the resolution provided by deep-sea cores, the Toba supereruption occurred during a major climatic event, the Oxygen Isotope Stage 5a-4 boundary (Ninkovich et al., 1978; Rampino and Self, 1992, 1993a) (Fig. 1). This boundary represents the episode of most rapid ice accumulation, and the main Northern Hemisphere warm-cold climatic transition, during the last glacial cycle. At the time, sea level is estimated to have dropped rapidly, and the first significant peak in ice-rafted detritus in the North Atlantic during the last glacial cycle occurred at the same time (Ruddiman, 1977; Heinrich, 1988). Summer and winter sea-surface
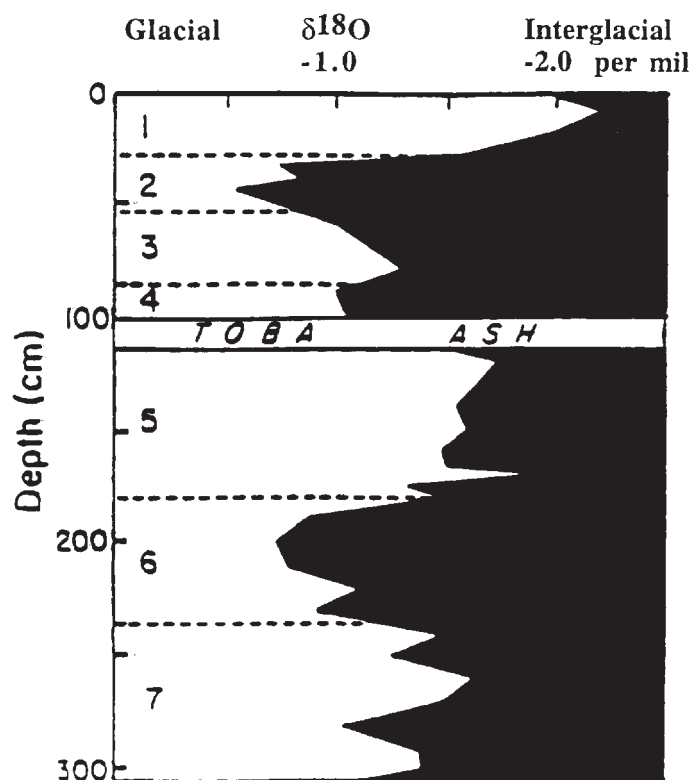
Figure 1. Piston core RC14-37, Indian Ocean, showing position of the 15-cm-thick Toba ash layer at the Oxygen Isotope Stage 5a-4 boundary. Vertical scale is depth in core (Ninkovich et al., 1978).

temperatures in the North Atlantic (Sancetta et al., 1972) decreased by about 10°C in less than 5,000 yr (but apparently lagging the ice accumulation), the greatest decrease of the last glacial cycle. European pollen records suggest that the cooling began very abruptly (Woillard and Mook, 1982; Guiot, 1990) at ~74,000 B.P., and abrupt changes in ocean bottom-water circulation in the Atlantic may account for a very rapid shift in benthic fauna (Streeter and Shackleton, 1979). Important climatic shifts are also seen in palynological studies from deep-sea cores and land areas in the tropics (e.g., van der Kaars, 1991; van der Kaars and Dam, 1995).

The unusual rapidity of the climatic cooling at the Stage 5a-4 boundary led to an examination of the possible role that volcanic triggering may have played in these long-term climatic changes (Rampino et al., 1979; Rampino and Self, 1992, 1993a). These studies concluded that the Toba event might have enhanced the long-term cooling trend, and that the cooling itself might have helped to trigger the eruption through stresses related to the initial changes in sea level, providing a kind of "volcano-climate" feedback effect (Rampino et al., 1979; Zielinski et al., 1994, 1996b).

## RESULTS OF ICE-CORE STUDIES

The exceptional magnitude of the Toba eruption makes it a natural target for studies of volcanic events preserved in polar ice cores. Recent work on the GISP2 (Greenland Ice Sheet Project 2)

ice core from Summit, Greenland, has revealed a ca. 6-yr-long period of enhanced volcanic sulfate dated at 71,100 ±5,000 B.P. that has been identified with the Toba eruption (Zielinski et al., 1996a, b) (Fig. 2). The magnitude of this $SO_4^{2-}$ signal is the largest in the entire 110,000 yr of the GISP2 record, and the high electroconductivity measurements (ECM) show that the peak has a volcanic origin, although no characteristic ash particles were found in filtered meltwater samples from the acidity peak (suggesting a distant source).

Zielinski et al. (1996a) estimate that the total atmospheric loading of $H_2SO_4$ for the ca. 6-yr period of the ice-core peak ranged from ~0.7 to $4.4 \times 10^{15}$ g, in general agreement with the above estimates for Toba derived from volcanological techniques and scaling from smaller eruptions (Rose and Chesner, 1990; Rampino and Self, 1992, 1993a). Estimates of aerosol loadings range from ~150 Mt to 1,000 Mt per year over the ca. 6-yr period of the ice-core peak. This would translate into aerosol optical depth perturbations between ~1 and 10.

The $SO_4^{2-}$ signal identified with Toba coincides with the beginning of a ca. 1,000-yr cooling event seen in the ice-core record between Interstadials 19 and 20, but is separated from the major ca. 9,000-yr glacial period of Isotope Stage 4 by the ca. 2,000-yr-long Interstadial 19 (Fig. 2). A similar cool pulse between interstadials is seen in the pollen record of the Grande Pile in northeastern France, dated by radiocarbon methods as ca. 70,000 B.P. (Woillard and Mook, 1982).

Thus, the ice-core evidence suggests that although the Toba signal occurred during the transition from Isotope Stage 5a to Stage 4, it was preceded and followed by abrupt climate oscillations that preceded the start of the major early Wisconsinan glaciation (Zielinski et al., 1996a, b), and thus may not have had a direct effect on initiation of the long-lived glacial stage (Kerr, 1996). Zielinski et al. (1996a) noted, however, that the ca. 1,000-yr stadial period that immediately follows the Toba acid spike began with an intense pulse of cooling indicated by increased $Ca^{2+}$ deposition for ca. 200 yr (Fig. 2), and they suggest that this anomalous cooling pulse might have been triggered by the ocean-atmosphere feedbacks related to the unusually long-lived Toba aerosol perturbation.

## TOBA AND VOLCANIC WINTER

A stratospheric aerosol burden for Toba of ~$1 \times 10^{15}$ g yields a global aerosol optical depth ($\tau$) of ~10 (Stothers, 1984a, b). An injection into the stratosphere of 10% of the estimated fine dust could have generated a very short-term (a few months) dust-cloud also of optical depth of ~10. Since Toba is a low-latitude volcano, dust and volatiles could have been injected efficiently into both Northern and Southern Hemispheres (Rampino et al., 1988), although the season of the eruption is unknown. These estimated aerosol optical depths ($\tau$ = 10–20) are roughly equivalent in visible opacity to smoke-cloud absorption optical depths, $\tau_a$ of ~2 (Turco et al., 1983), which is within the range used in nuclear-winter scenarios of massive emissions of soot from burning urban and industrial areas in the aftermath of nuclear war.
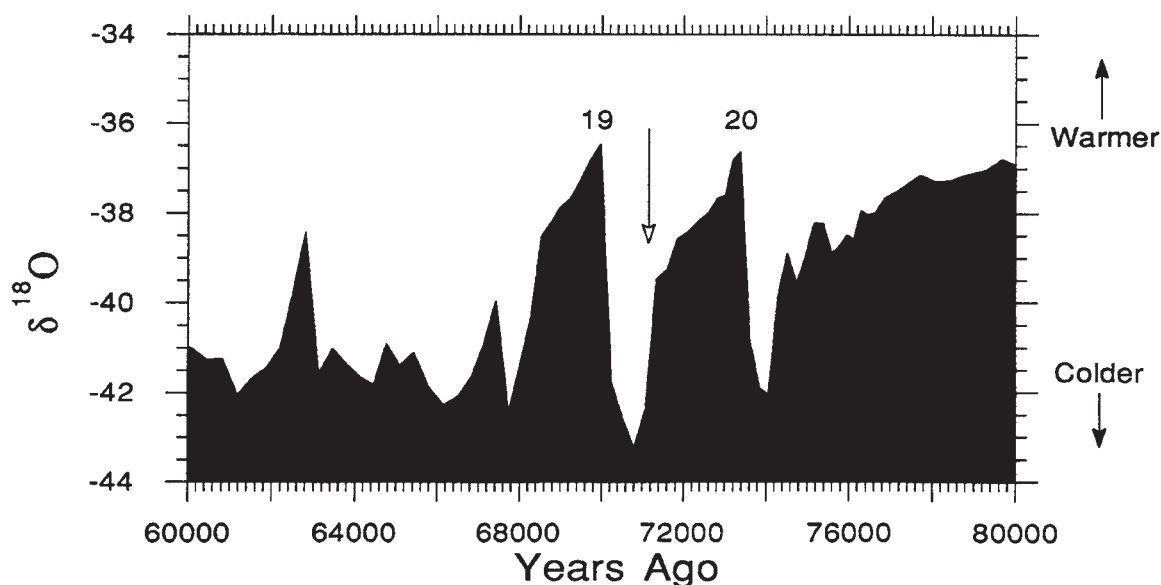
Figure 2. The δ18O record of the GISP2 ice core (Greenland) for the period between 60,000 and 80,000 yr ago showing the cool stadial between warmer interstadials 20 and 19. Arrow shows timing of Toba aerosol peak in the same core, during the initial rapid cooling in the stadial event (Zielinski et al., 1996b).

Although the climate conditions and duration of a nuclear winter have been much debated, simulations by Turco et al. (1990) predict average absorption optical depths ($\tau_a$) of up to 2–3 on hemispheric scales, for a full-scale nuclear exchange. For the initial acute phase (first 1–3 months), 3-D General Circulation Model (GCM) experiments, for a worst-case July smoke injection (absorption optical depths of 0.3–3), predict that land temperatures in the 30° to 70° N zone could range from ~5° C to ~15° C colder than normal, with freezing events in midlatitudes during the first months. At lower latitudes, GCM simulations suggest coolings of 10° C or more. Drastic decreases in precipitation in the first few months are also predicted. During the chronic phase (first 1–3 yr), with a possible ca. 1-yr residence time for the smoke, global average $\tau_a$ of 0.5 could persist, with significant cooling of 5° C or more in areas covered by the smoke cloud. Ocean-surface cooling of ~2–6°C might extend for several years, and persistence of significant soot optical depths for 1–3 yr might lead to longer term (decadal) climatic cooling, primarily through climate feedbacks including increased snow cover and sea ice, and perturbed sea-surface temperatures.

The injection of massive amounts of volcanic dust into the stratosphere by a supereruption such as Toba, producing dust optical depths of ~10, might be expected to lead to similar immediate surface cooling, creating a "volcanic winter" (Rampino et al., 1988; Rampino and Self, 1992). Volcanic dust probably has a shorter residence time in the atmosphere (3–6 months) than soot (Turco et al., 1990) and spreads from a point source, but volcanic dust is injected much higher into the stratosphere; hence, Toba ash would have had a wide global coverage despite its short lifetime. Evidence of the wide dispersal of the dust and ash from

Toba can be seen from lake deposits in India, where the reworked Toba ash forms a layer up to 3 m thick, and from the widespread ≥10-cm-thick ash layer in the Indian Ocean (Acharya and Basu, 1993; Shane et al., 1995).

Evidence for rapid and severe cooling from the direct effects of volcanic ash clouds comes from the aftermath of the 1815 Tambora eruption. Madras, India, experienced a dramatic cooling during the last week of April 1815, a time when the relatively fresh ash and aerosol cloud from Tambora (April 10–11) would have been overhead. Morning temperatures dropped from 11° C on Monday to -3° C on Friday (Stothers, 1984a). A similar, but much smaller effect occurred as the dust cloud from the 1980 Mount St. Helens eruption passed over downwind areas in the United States (Robock and Mass, 1982).

The stratospheric injection of ≥$10^{15}$ g of sulfur volatiles, and the time required for the formation and spread of volcanic $H_2SO_4$ aerosols in the stratosphere should lead to an extended period of increased atmospheric opacity and surface cooling. In earlier studies, a maximum e-folding residence time for the Toba acid aerosols in the stratosphere of ca. 1 yr was inferred based on historical eruptions (Rampino and Self, 1992, 1993a), producing a sulfuric-acid aerosol optical depth of $\tau = $ ~10 decreasing to $\tau = 0.1$ (~$10^{13}$ g) over the course of several years. The ice-core record, however, indicates stratospheric loadings of $10^{14}$–$10^{15}$ g of $H_2SO_4$ aerosols for up to 6 yr after the eruption (Zielinski et al., 1996a).

This agrees with model calculations by Pope et al. (1994) that predict oxidation lifetimes (time required to convert a given mass of sulfur into $H_2SO_4$ aerosols) of between 4 yr and 17 yr, and diffusion lifetimes (time required to remove unoxidized $SO_2$ by diffusion to the troposphere) of between 4 yr and 7 yr for total

sulfur masses between $10^{15}$ g and $10^{16}$ g. For atmospheric injection in this range, the diffusion lifetime is the effective lifetime of the cloud, because the $SO_2$ reservoir is depleted before oxidation is completed.

In earlier studies, it was argued that if the relationship between Northern Hemisphere cooling and aerosol loading from large eruptions is approximately linear, then scaling up from the A.D. 1815 Tambora eruption would lead to an ~3.5° C hemispheric cooling after Toba (Rampino and Self, 1993a). Similarly, empirical relationships between $SO_2$ released and climate response (Palais and Sigurdsson, 1989) suggested a hemispheric surface-temperature decrease of about 4 ±1° C. The recent discovery that Toba aerosols had a much longer residence time in the atmosphere than those of much smaller historical eruptions raises the possibility of a considerably greater cooling from Toba. Model results of Pollack et al. (1993) suggest that because of the heat capacity of the upper mixed layer of the oceans, the e-folding time for cooling by steady-state volcanic forcing is about 15 yr. The eruption clouds of individual historic eruptions have been too short-lived to drive lower tropospheric temperatures to their steady-state values (Pollack et al., 1993), but the apparently long-lasting Toba aerosols may mean that the temperature changes in the troposphere attained a larger fraction of their steady-state values.

## POSSIBLE ENVIRONMENTAL AND ECOLOGICAL EFFECTS OF THE TOBA ERUPTION

The climatic and environmental impacts of the Toba super-eruption are potentially so much greater than that of recent historical eruptions (e.g., Hansen et al., 1992; Stothers, 1996) that instrumental records, anecdotal information, and climate-model studies of the effects of these eruptions may not be relevant in scaling up to the unique Toba event (Rampino et al., 1988; Rampino and Self, 1993a). Various studies on the effects of extremes of atmospheric opacity and climate cooling on the environment and life have been carried out, however, in connection with studies of nuclear winter and the effects of asteroid impacts on Earth (e.g., Harwell, 1984; Green et al., 1985; Tinus and Roddy, 1990), and some of these may be more relevant to the Toba situation.

Two major effects on plant life from high-atmospheric opacity are reduction of light levels and cold temperatures. For aerosol optical depths between ~1 and ~10, the reduction in light levels expected from the Toba eruption would range from dim-sun conditions (~75% sunlight transmitted), like those seen after the 1815 Tambora eruption, to that of an overcast day (~10% sunlight transmitted). Experiments with young grass plants have shown how net photosynthesis varies with light intensity. For a decrease to 10% of the noon value for a sunny summer day, photosynthesis was reduced by ~85% (van Keulen et al., 1975), and photosynthesis also drops with decreasing temperatures (Redman, 1974).

Resistance of plants to unusual cold conditions varies considerably. Conditions in the tropical zone are most relevant to possible impacts on early human populations in Africa. Tropical forests are very vulnerable to chilling, and Harwell et al. (1985) argue that

for freezing events in evergreen tropical forests, essentially all above-ground plant tissues would be killed rapidly (see also Taylor and Rowley, 1971; Sweeney and Hopkinson, 1975). Average surface temperatures in the tropics today range from ~16°C to 24°C. Nuclear winter scenarios predict prolonged temperature decreases of 3–7° C in equatorial Africa, and short-term temperature decreases of up to 10°C. Many tropical plants are severely damaged by chilling to below 10–15° C for a few days (Table 1) (Leavitt, 1980; Hutchinson et al., 1985; Greene et al., 1985). Harwell (1984) compiled data showing that the $LT_{50}$ (temperatures required to kill at least 50% of the plants after exposure to cold for 2 hr or more) for most tropical plants was in the range +5 to –2°C (Harwell, 1984). Seedlings and saplings are most vulnerable. Even more serious is the fact that most tropical forest plants have limited seed banks, and the seeds typically lack a dormant phase. Furthermore, regrowth tends to produce forests of limited diversity, capable of supporting much less biomass (Harwell et al., 1985).

Even for temperate forests, destruction could be very severe (Harwell, 1984; Harwell et al., 1985). In general, the ability of well-adapted trees to withstand low temperatures (cold-hardiness) is much greater than is needed at any single time of the year, but forests can be severely damaged by unusual or sustained low temperatures during certain times of the year. For example, Tinus and Roddy (1990) estimated the cold-hardiness of Rocky Mountain Douglas fir based on controlled growth experiments in which they derived the temperature decreases that would be necessary to kill at least 50% ($LT_{50}$) of the fir trees (Fig. 3). They found that a simulation of a 10°C decrease in temperatures during winter would have minimal effect on the cold-hardy and dormant trees, whereas a similar 10°C drop in temperature during

TABLE 1. DAMAGE TO TROPICAL PLANTS AND CROPS OF TROPICAL ORIGIN BY CHILLING

| Species | Chilling treatment | Result |
|---|---|---|
| Rice (*Oryza sativa*) | 7 °C during flower formation | Flowers do not form |
| Rice and hybrid sorghum | 13–16 °C during pollen formation | Sterile pollen |
| Cotton (*Gossypium arboreum*) | Exposure of seeds to 30 minutes at 12 °C, 2–4 hr. after they start to absorb water | Seeds die or roots develop abnormally |
| Maize (*Zea mays*) sorghum-some hybrid varieties | Chill to 10 °C for 2.5 days and then restore to 25 °C | Permanently inactivates and photosynthesis in sorghum leaves and reduces the rate in maize to 33–67% of that of controls |
| Pasture legumes (similar to clovers) in Queensland, Australia | Grow plants at various temperatures down to the chilling range | Severe growth inhibition by chilling |
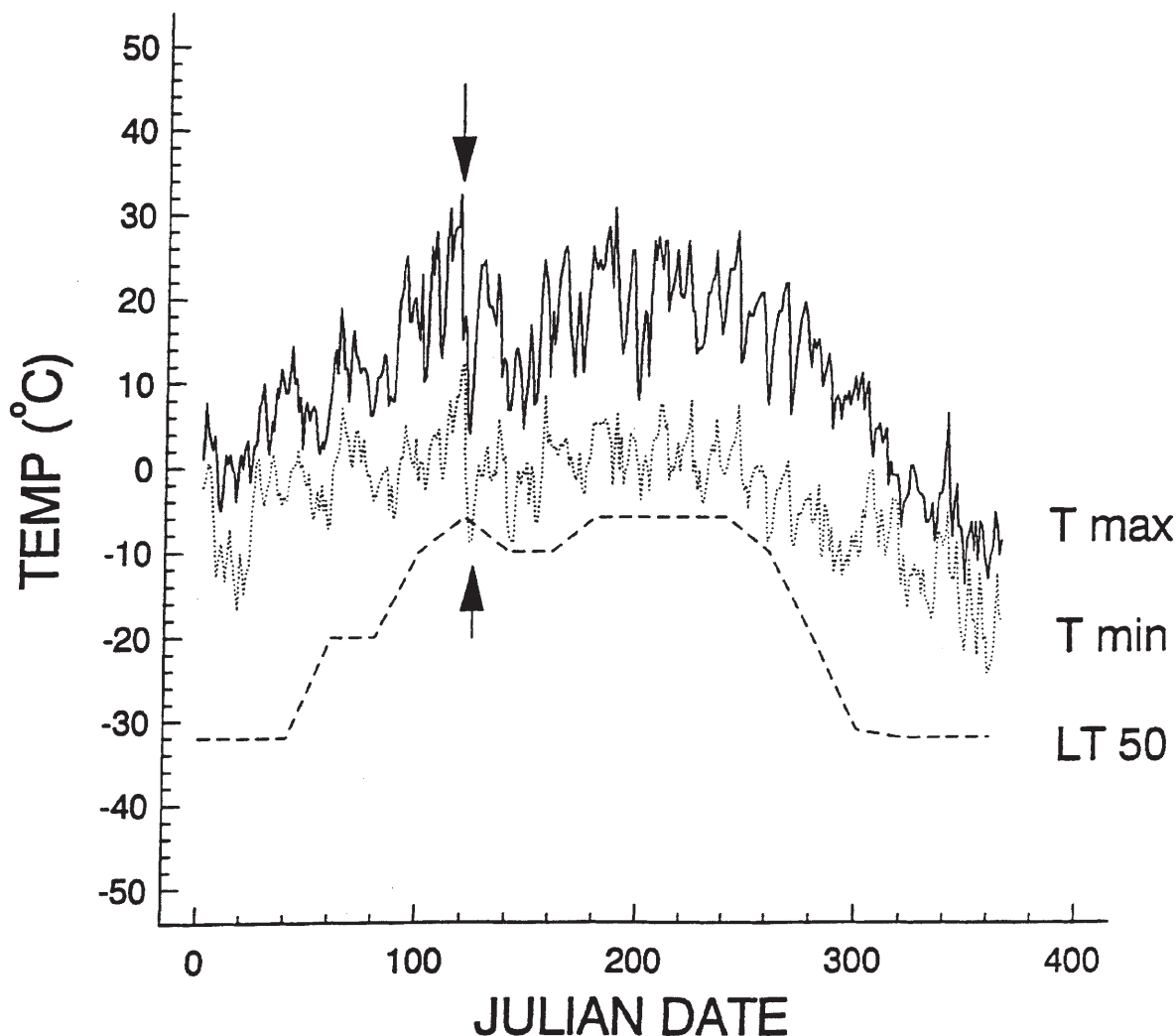
Figure 3. Simulation of the effects of cooling event during the growing season on forest ecosystems. Daily maximum (continuous line) and minimum (dotted line) temperatures for 1987 at Lucky Peak Nursery, Boise, Idaho, with the temperature curves offset to simulate a drop of 10°C after April 30 (top arrow). The calculated cold-hardiness lethal temperatures ($LT_{50}$) associated with Rocky Mountain Douglas fir trees (based on controlled growth experiments) is shown as the dashed line. Note that the minimum temperature on May 1 drops below the $LT_{50}$ temperature (bottom arrow), killing 50% of the trees and severely damaging the rest of the forest (after Tinus and Roddy, 1990).

the growing season (when cold-hardiness is decreased) leads to a 50% dieback and severe damage to surviving trees (including damage of new growth and invasion by pathogens), resulting in the loss of at least a year's growth.

The situation for deciduous forest trees would be even worse than that for the evergreens, as their entire foliage would be new and therefore lost. For example, Larcher and Bauer (1981) determined that cold limits of photosynthesis of various temperate zone plants range from -1.3°C to -3.9°C, approximately the same range as the tissue-freezing temperatures for these plants. Lacking adequate food reserves, most temperate forest trees would not be able to cold harden in a timely manner and would die or suffer additional damage during early freezes in the fall (Tinus and Roddy, 1990).

Harwell (1984) carried out simulations for a mixed conifer and hardwood temperate forest for a longer term reduction in annual average temperature of 3°C, 6°C, or 9°C for a period of 5 yr (assuming normal conditions of light and rainfall) (Fig. 4). For a 3°C annual temperature reduction for 5 yr, biomass was predicted to fall by 25%, with recovery in 30-40 yr, whereas for a 6°C cooling, biomass fell by 80% and returned to 50% of normal after 50 yr. For the extreme 9°C cooling, biomass fell by 90%, and recovered only to 33% of normal after 50 yr (Fig. 4). After recovery, species composition was different. For grassland ecosystems, simulations of a 5-yr period of temperature decrease of 3-9°C led to a reduction in net primary production ranging from 9% to 51% (Harwell, 1984), with recovery times of several years.
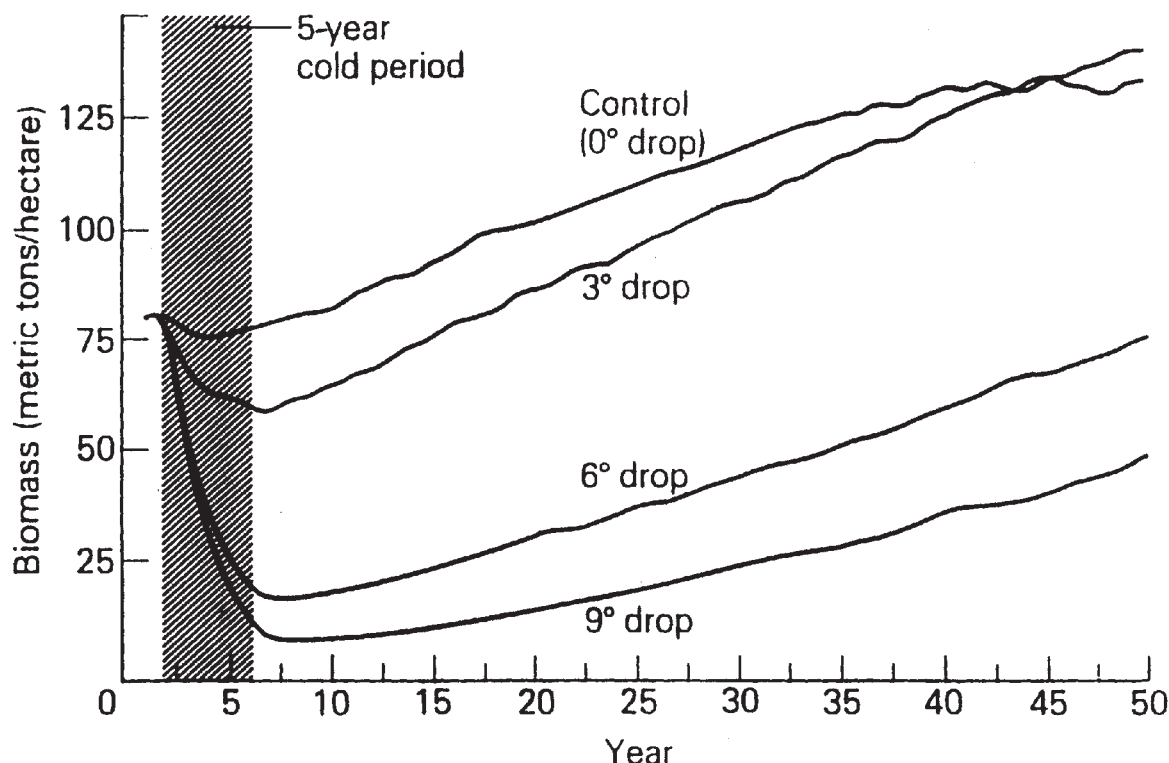
Figure 4. Simulation of reduction of growth and recovery for a mixed conifer and woodland forest (typical of eastern Tennessee, U.S.A.) as a result of reductions in annual average temperatures. Reductions in annual average temperatures of 3°C, 6°C, and 9°C for 5 yr (control run with 0°C temperature drop also shown) (see text for explanation of results) (after Harwell, 1984).

The large amount of dead wood produced by the dead and damaged trees, exacerbated by drought conditions (see below) might lead to increased forest fires. Tinus and Roddy (1990) estimate that the combined forests of the Northern Temperate Zone have a total of $1 \times 10^{12}$ m$^3$ of standing wet, live biomass. An LT$_{50}$ event would thus produce conservatively ~$0.5 \times 10^{12}$ m$^3$ of dead forest material, or ~$0.2 \times 10^{12}$ t of dry biomass. Burning of this biomass would release large amounts of reactive species such as hydrocarbons, organic acids, and nitrogen compounds into the global atmosphere (Andreae et al., 1988). This perturbation should be recorded as elevated concentrations of $NH_4^+$ and organic acids ($HCOO^-$) in ice cores (e.g., Legrand et al., 1992; Whitlow et al., 1994). Tinus and Roddy (1990) estimated an output of ~$2 \times 10^{10}$ kg of soot from these forest fires. Total present annual emissions of $NH_4^+$ from Northern Hemisphere forest fires is $3 \times 10^9$ kg (Hegg et al., 1988).

The effect of the Toba eruption on the oceans is more difficult to estimate. Regionally, the effect on ocean biota of the fallout of ~4 g/cm$^2$ of Toba ash over an area of $5 \times 10^6$ km$^2$ in the Indian Ocean must have been considerable. Deposition rates of N, organic C, and CaCO$_3$ all rise sharply in the first few centimeters of the Toba ash layer, indicating that the ash fallout swept the water column of most of its particulate organic carbon and calcium carbonate (Gilmour et al., 1990). Smit et al. (1991) reported that the lysocline in the eastern Indian Ocean suddenly became shallower at the time of the Toba eruption.

Another possible effect of a dense aerosol cloud is decreased

ocean productivity. For example, satellite observations after the 1982 El Chichón eruption showed high aerosol concentrations over the Arabian Sea, and these values were associated with low surface productivity (as indicated by phytoplankton concentrations) from May through October of that year (Strong, 1993). Brock and McClain (1992) suggested that the low productivity was related to weaker than normal monsoon winds, and independent evidence suggests that the southwest monsoon in the area arrived later and withdrew earlier than usual, and that the wind-driven Somali current was anomalously weak. Conversely, Genin et al. (1995) reported enhanced vertical mixing of cooled surface waters in weakly stratified areas of the Red Sea following the Pinatubo eruption, which resulted in algal and phytoplankton blooms that precipitated widespread coral death.

Oxygen-isotope studies of corals following the 1991 Pinatubo eruption provide evidence that aerosol-induced cooling of the southwestern Pacific could lead to significant weakening of Hadley Cell circulation and rainfall, and might precipitate long-term El Niño–like anomalies with extensive drought in many tropical areas (Gagan and Chivas, 1995). Climate model simulations predict significant drought in tropical areas from weakening of the trade winds/Hadley circulation and from reduction in the strength of the summer monsoon (e.g., Pittock et al., 1986, 1989; Turco et al., 1990). For example, Pittock et al. (1989) presented GCM results that showed a 50% reduction in convective rainfall in the tropics and monsoonal regions. Detailed study of ~75,000-yr-old

corals with annual growth records from various parts of the world ocean could provide resolution of some of the short-term climatic effects of the Toba eruption.

## BOTTLENECK IN HUMAN POPULATION

Recent debate about the origin of modern humans has focused on two competing hypotheses: (1) the 'multiregional' hypothesis, in which the major subdivisions of our species evolved slowly and in situ, with gene flow accounting for the similarities now observed among groups (Wolpoff, 1989, 1996); and (2) the 'replacement' hypothesis, in which earlier populations were replaced between 100,000 and 30,000 yr ago by modern humans that originated in Africa (Stringer and Andrews, 1988; Lahr, 1996; Lahr and Foley, 1994; Rogers and Jorde, 1995). Recently, a number of genetic studies have been aimed at testing these two hypotheses. Studies of nuclear and mitochondrial DNA (mtDNA) from present human populations led to the conclusion that the modern populations originated in Africa and spread to the rest of the Old World

ca. 100,000 yr ago (Wainscoat et al., 1986; Jones and Rouhani, 1986; Harpending et al., 1993; Tischkoff et al., 1996). Harpending et al. (1993) recently proposed the 'weak Garden of Eden' version of the replacement model, in which modern humans dispersed from Africa ca. 100,000 yr ago and then went through a population bottleneck ca. 50,000 ± 20,000 yr ago, followed by a dramatic population increase (Harpending et al., 1993) (Fig. 5A).

Harpending et al. (1993) estimate the bottleneck reduced the human population to ~500 breeding females, or a total population as small as 4,000 for ca. 20,000 yr, or as small as 40 individuals for two centuries. Sherry et al. (1994) estimate from 40 to 600 females, which translates into a population size of <3,000. Because such a small population could not have occupied all of the temperate parts of the Old World where remains of *Homo erectus* have been found, the demographic implications of the genetic diversity data seem to contradict the multiregional hypothesis of a widespread transformation of *Homo erectus* into modern *Homo sapiens*. Furthermore, all non-African populations carry a small subset of African alleles. If the idea of regional continuity is
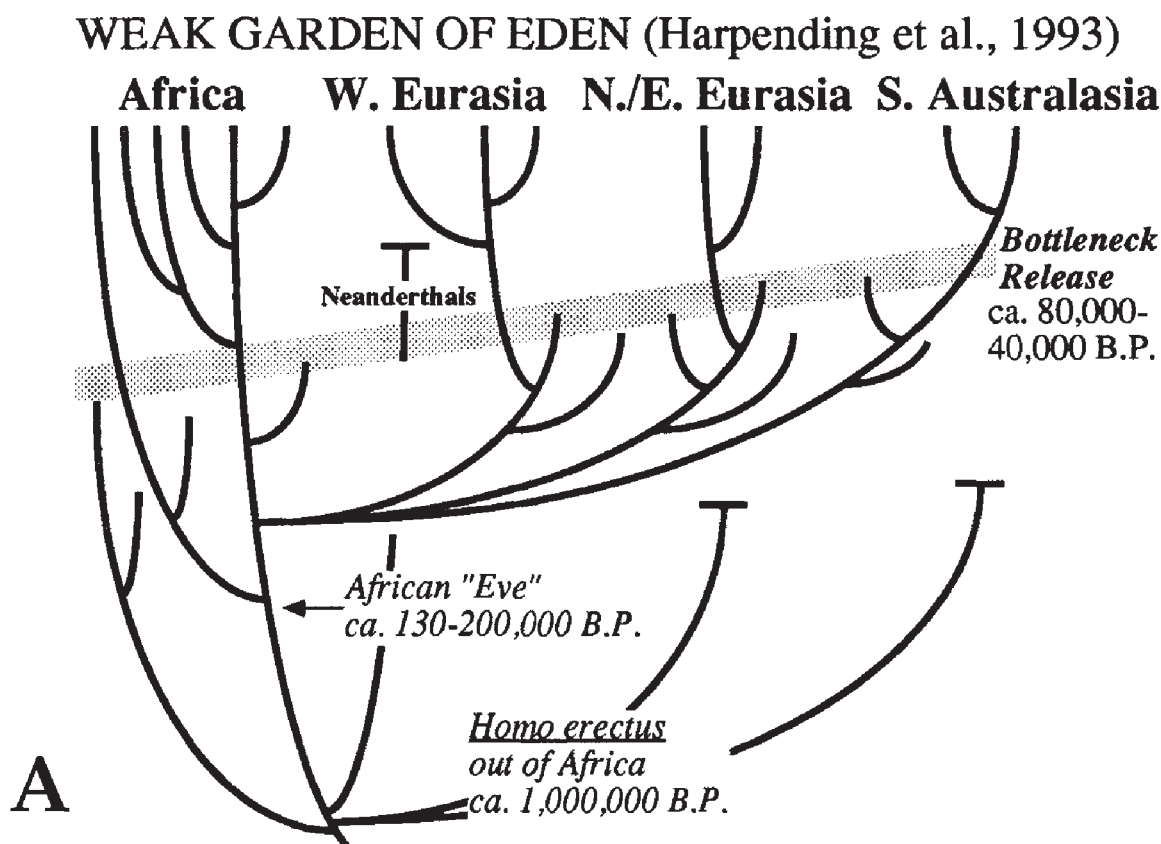


Figure 5 (on this and opposite page). The 'weak Garden of Eden' model (a) of Harpending et al. (1993), and (b) revisions to the model proposed by Ambrose (1998a). The original model does not explain the cause of the population bottleneck, and posits a time-transgressive release from the area of the origin of modern human technology in East Africa. The revised 'Garden of Eden/Volcanic Winter' model (b) suggests that volcanic winter following the Toba super-eruption caused the bottleneck, and that the release was globally synchronous, ca. 1,000 yr later, at the advent of Interstadial 19. Africa's apparent earlier release in the original model is the result of a larger population through the bottleneck.

correct, one would expect at least a few alleles that cannot be traced to recent African populations, especially if regional populations were established as early as 1.8 m.y. ago. However, no alleles have yet been found (Stoneking, 1993; Tischkoff et al., 1996a, b).

Hasegawa et al. (1993) inferred from the pattern of the human mtDNA tree that at 89,000 ±69,000 yr ago the human population expanded to ~150,000 breeding females or ~300,000 breeding individuals. Sherry et al. (1994) estimated mean population expansion times ranging from ca. 65,000 to 30,000 yr ago, with the African expansion possibly being the earliest. Harpending et al. (1993) also noted the earlier release in Africa (Fig. 5A) and proposed that it occurred in response to the invention and spread of improved technology, known as the Later Stone Age in subsaharan Africa and the Upper Paleolithic in Europe. This transition marks the appearance of more efficient patterns of resource exploitation (Ambrose and Lorenz, 1990; Klein, 1989, 1992, 1995) supporting higher population densities.

The chronology of this transition is now being resolved. The earliest Later Stone Age sites appear to date to more than 50,000 yr ago and are found in highland Kenya and Tanzania (Ambrose, 1998b). In Israel and the Sinai Peninsula, on the doorstep of Africa, the transition from the Middle to the Upper Paleolithic occurs about 46,000 yr ago (Bar-Yosef et al., 1996), and in Europe it occurs ca. 42,000 yr ago (Bischoff et al., 1989, 1994;

Cabrera-Valdes and Bischoff, 1989; Mercier et al., 1995). This pattern is consistent with a postulated East African origin for modern humans and modern human technology (Klein, 1992, 1995). However, given the generally earlier estimates for the date of the population recovery following the bottleneck, it is uncertain if this can be directly linked to the sudden great leap forward in human creativity and culture at ca. 40,000–50,000 yr ago.

The older date of the release from the bottleneck in Africa is also open to question. The amount of genetic diversity in a population is usually considered to reflect time because a constant rate of neutral mutations is assumed in a population of constant size. Larger populations passing through genetic bottlenecks, however, should have greater diversity than smaller ones (Relethford and Harpending, 1995). The largest landmass in the warm tropics, Africa, should have maintained the largest population through the late Quaternary bottleneck, and thus should have had the highest genetic diversity. This is supported by recently developed techniques of integrating craniometric and genetic estimates of diversity, which suggest that past population size in Africa was three times higher than elsewhere (Relethford and Harpending, 1994).

What then could have caused these population bottlenecks and the subsequent expansion of modern humans? If the genetic information has been correctly interpreted, then humans were clearly an endangered species only a few tens of thousands of



WEAK GARDEN OF EDEN / VOLCANIC WINTER
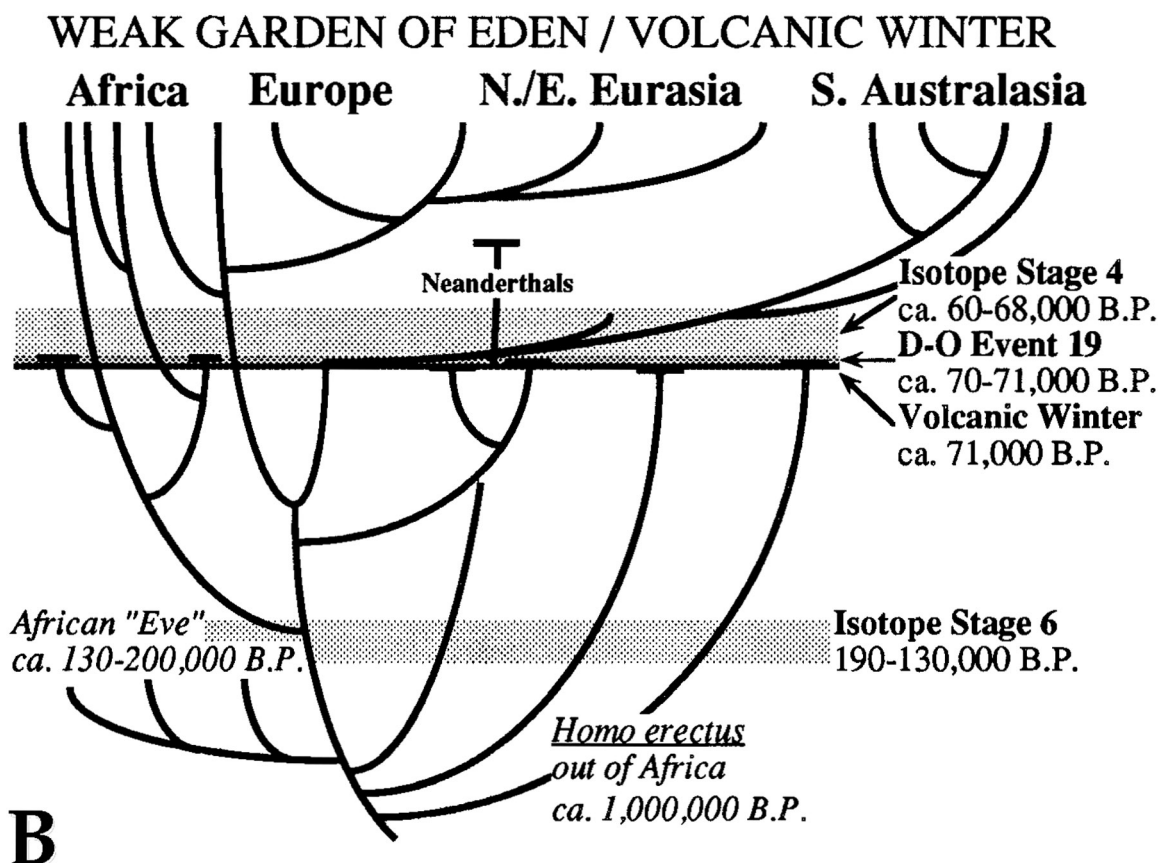
years ago (Cann et al., 1987; Gibbons, 1993). The timing of the Toba supereruption roughly matches the inferred timing of the bottleneck and release, and the predicted environmental after-effects of the Toba eruption might have been so severe as to lead to a precipitous decline in the population of human ancestors (Ambrose, 1998a; Rampino and Self, 1993b) (Fig. 5B).

One prediction of the idea that the bottleneck was caused by an environmental catastrophe is that other species should have been similarly affected and should show similar evidence in their DNA. Thus, it may be significant that recent analysis of mtDNA of the Eastern Chimpanzee (*Pan troglodytes schweinfurthii* ) (Morin et al., 1994) shows a pattern similar to human DNA, suggesting a severe reduction in population at about the same time as that in the human population (see Rogers and Jorde, 1995). Additional data on the genetic structure of eastern African chimpanzee mtDNA show that different subpopulations went through one or more bottlenecks at different times in the later Pleistocene (Goldberg, 1996). Several chimpanzee populations that reside within a postulated Pleistocene refugia in the montane region centered on western Uganda and eastern Zaïre (Hamilton, 1976) apparently experienced a bottleneck ca. 67,000 yr ago. Populations farther from the core refugium seem to have passed through a bottleneck much more recently, possibly during the Last Glacial Maximum (LGM, Oxygen Isotope Stage 2) ca. 20,000 yr ago (Goldberg, 1996). Chimpanzee populations in the core refugium area were apparently less affected by the cold climate of the LGM because it is only weakly reflected in their population genetic structure. These results are highly encouraging for the hypothesis that Toba's volcanic winter had a significant impact on the population of tropical animals. The genetic makeup of other organisms should be searched for similar patterns.

The apparently large magnitude of physical differences among our modern human races seems to support the multiregional model of an ancient (1–2 m.y.) origin for modern human races and argues against the recent African replacement model. The late Pleistocene population bottleneck, however, may account for the seeming paradox presented by the genetic data (Ambrose, 1998a): If we are genetically so similar, why don't we all resemble more closely our African ancestors? The answer may be that the small, isolated populations from the late Quaternary African diaspora that did survive the volcanic winter would have experienced the equivalent of the founder effect (Mayr, 1970). When populations were reduced to very small sizes, a random subset of preexisting genetic diversity would be retained in each isolated population. If population size remained small for many generations, then genetic drift could reduce genetic diversity even further. If the Toba volcanic winter and hypercold millennium between Interstadials 19 and 20 kept human populations small, then the founder effect could have operated for about 40 generations (assuming 25 yr per generation), contributing to the diversity seen in modern human populations.

## CONCLUSIONS

Evidence from volcanology, ice-core studies, and atmospheric modeling suggests that the Toba eruption of ca. 73,500 B.P. pro-

duced a widespread dust cloud, and a dense global $H_2SO_4$ aerosol cloud that persisted for up to 7.5 yr. The estimated dust and aerosol loadings of $10^{15}$–$10^{16}$ g would have produced regional to global aerosol optical depths between 1 and 10, similar to those predicted in scenarios of nuclear winter. The Toba aerosol cloud is predicted to have caused severe cooling, with immediate temperature decreases to near or below freezing in the tropics, hard freezes at midlatitudes, and an extended cooling of 3–5° C or more on a global basis. Drought in the tropics from weakening of the Hadley Cell circulation and monsoons is also possible.

The aftermath of the Toba eruption would have constituted a global environmental disaster, with especially severe effects in the tropics, where vegetation lacks cold hardiness. Even in temperate areas, botanical studies predict that forest and grassland ecosystems could have suffered widespread destruction, with recovery times as long as several decades. Furthermore, longer term cooling might have been induced or enhanced by positive feedbacks (e.g., ocean cooling, increased snow cover and sea ice) to the long-lived Toba aerosols. Ice-core evidence from Greenland shows that the Toba eruption coincided with a 200-yr period of sharp cooling that initiated a ca. 1,000-yr stadial event.

The Toba supereruption occurred during a window of time in which the early human population suffered an extreme bottleneck, with some estimates of as few as 3,000 individuals, followed by the expansion of modern humans. Botanical studies of the expected damage to natural ecosystems from severe coolings and drought such as expected in the aftermath of Toba predict a global environmental disaster that could have contributed to population crashes of various organisms. The recent discovery of genetic evidence that Eastern Chimpanzees suffered a population bottleneck at about the same time lends credence to the disaster theory.

## ACKNOWLEDGMENTS

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Michael R. Rampino and Stanley H. Ambrose

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