Did the Toba Volcanic Eruption of ~74k BP Produce Widespread Glaciation?

Alan Robock[1], Caspar M. Ammann[2], Luke Oman[3], Drew Shindell[4],
Samuel Levis[2], and Georgiy Stenchikov[1]

[1]*Department of Environmental Sciences, Rutgers University, New Brunswick, New Jersey*
[2]*National Center for Atmospheric Research, Boulder, Colorado*
[3]*Department of Earth and Planetary Sciences, Johns Hopkins University, Baltimore, Maryland*
[4]*NASA Goddard Institute for Space Studies, New York City*

Submitted to *Journal of Geophysical Research*

December, 2008

Revised March, 2009

Corresponding Author:
Alan Robock
Department of Environmental Sciences
Rutgers University
14 College Farm Road
New Brunswick, NJ 08901
Phone: 732-932-9800, x6222
Fax: 732-932-8644
E-mail: robock@envsci.rutgers.edu

Abstract

It has been suggested that the Toba volcanic eruption, approximately 74 ka BP, was responsible for the extended cooling period and ice sheet advance immediately following it, but previous climate model simulations, using 100 times the amount of aerosols produced by the 1991 Mt. Pinatubo eruption, have been unable to produce such a prolonged climate response. Here we conduct six additional climate model simulations with two different climate models, the National Center for Atmospheric Research Community Climate System Model 3.0 (CCSM3.0) and National Aeronautics and Space Administration Goddard Institute for Space Studies ModelE, in two different versions, to investigate additional mechanisms that may have enhanced and extended the forcing and response from such a large supervolcanic eruption. With CCSM3.0 we include a dynamic vegetation model to explicitly calculate the feedback of vegetation death on surface fluxes in response to the large initial reduction in transmitted light, precipitation, and temperature. With ModelE we explicitly calculate the effects of an eruption on stratospheric water vapor and model stratospheric chemistry feedbacks that might delay the conversion of SO_2 into sulfate aerosols and prolong the lifetime and radiative forcing of the stratospheric aerosol cloud. To span the uncertainty in the amount of stratospheric injection of SO_2, with CCSM3.0 we used 100 times the Pinatubo injection, and with ModelE we used 33, 100, 300, and 900 times the Pinatubo injection without interactive chemistry, and 300 times Pinatubo with interactive chemistry. Starting from a roughly present day seasonal cycle of insolation, CO_2 concentration, and vegetation, or with 6k BP conditions for CCSM3.0, none of the runs initiates glaciation. The CCSM3.0 run produced a maximum global cooling of 10 K and ModelE runs produced 8-17 K of cooling within the first years of the simulation, depending on the injection, but in all cases, the climate recovers over a few decades. Nevertheless, the "volcanic winter" following a

supervolcano eruption of the size of Toba today would have devastating consequences for humanity and global ecosystems. These simulations support the theory that the Toba eruption indeed may have contributed to a genetic bottleneck.

1. Introduction

The eruption of the Toba volcano (2.5°N, 99°E) on the island of Sumatra in Indonesia (Figure 1) was the largest explosive event of at least the past 100,000 years [*Zielinski*, 1996; *Oppenheimer*, 2002; *Self and Blake*, 2008]. The erupted volume in dense rock equivalent material (~2800 km^3) was about three orders of magnitude larger than for the 1980 eruption of Mount St. Helens. *Zielinski et al.* [1996] estimate that the eruption was 71,000 years ago ± 5000 years, based on ice-core dating. *Oppenheimer* [2002] reviewed nine different estimates over a wide range, and concluded that the eruption was 74,000 ± 2000 years ago. For the purposes of this paper, the exact date does not matter, and we assume that Zielinski is looking at the Toba signal.

Zielinski et al. [1996] showed ice core evidence for a 1000-year cool period with low $\delta^{18}O$ and increased dust deposition immediately following the eruption. In the South China Sea, in the immediate vicinity of the eruption, *Huang et al.* [2001] found corroborating evidence for a tropical cooling that lasted for 1000 years following the eruption. However, *Schultz et al.* [2002] analyzed Indian Ocean sediments just before and just after the Toba deposits and find no evidence for large climate change in the Indian monsoon system on the 100-1000 yr time scale. *Oppenheimer* [2002] claimed that the cooling was only about 1 K, but based his analysis on simplistic *T*-forcing relationships. We know of no observational evidence for climate change on the annual to decadal time scale, so there certainly could have been a very large, short-lived response to the eruption.

Rampino and Self [1992, 1993] and *Zielinski et al.* [1996] suggested that the 1000-year cool period (stadial) following the eruption could have been caused by the eruption. For this to have happened, the forcing from the volcanic aerosols would have had to last for a long time, feedbacks must have amplified or lengthened the stratospheric forcing, or the climate system would have had to have been at a tipping point, and a "little" nudge from a short-term, but large, reduction in shortwave forcing would have been enough to push it into the stadial. An intermediate-complexity climate model simulation by *Li and Berger* [1997], however, did not produce an ice age following a Toba-like perturbation. In the only general circulation model (GCM) simulation of the response to Toba so far, *Jones et al.* [2005] calculated the climate response to a stratospheric loading 100 times that of the 1991 Mount Pinatubo, but with the same timing as that of the Pinatubo eruption. They found a very large climate system response, with a maximum global-average cooling of 10 K. But the climate system started to warm back up in a few years, and the cooling was only about 2 K after a decade with no indication of the initiation of an ice age. Here we examine three additional factors that may have served to produce conditions favorable for sustained glacial conditions, with three different GCM experiments. (*Harris* [2008] reported preliminary results for Toba experiments with the same GCM as *Jones et al.* [2005], but with a slab ocean. The results were similar.)

Typically, for smaller eruptions, the e-folding time for stratospheric residence of volcanic sulfate aerosols is about one year [*Robock*, 2000]. *Rampino and Self* [1992] suggested that the normal amount of water vapor in the stratosphere might limit the formation of sulfate aerosols after extreme injections, and although they understood that Toba would have injected a large quantity of water into the stratosphere, they assume that only 10% of the SO_2 injected by Toba would convert to particles. *Bekki et al.* [1996] used a simple model of this phenomenon to suggest that the forcing from Toba might have lasted for a decade. However, the volcanic sulfate deposition from the Toba eruption found by *Zielinski et al.* [1996] in the Greenland ice cap only lasted about six years. To help answer this question we examine the water vapor response in the stratosphere.

Another feedback avenue is that the cold and dark conditions after the Toba eruption might have killed much of the vegetation on the planet, which in most parts of the world would have produced a higher surface albedo, which could have then led to further cooling. We examine this hypothesis with a GCM containing a coupled dynamic vegetation model that includes this feedback.

It is the SO_2 injections into the stratosphere from volcanic eruptions that causes climate change, as the resulting sulfate aerosol particles remain for a few years, blocking out solar radiation and cooling the surface [*Robock*, 2000]. *Jones et al.* [2005] assumed that the SO_2 emission from Toba was 100 times that of Pinatubo, but other estimates put it closer to 300 times Pinatubo [*Bekki et al.*, 1996; *Oppenheimer*, 2002]. Would such a larger volcanic forcing have been more likely to produce an ice age? To answer this question, we ran a GCM with a range of forcings: 33, 100, 300, and 900 times the Pinatubo forcing.

The Toba eruption certainly must have had significant impacts on stratospheric chemistry, which may have affected ozone and other gases, with a chemical feedback to prolong or enhance the forcing. To test this hypothesis, we also conducted an experiment with a model that includes these chemistry feedbacks, using the most likely scenario, 300 times Pinatubo.

All but one of our experiments were conducted from initial conditions representing current Milankovitch orbital configuration, current vegetation, and current greenhouse gas concentrations. *Rampino and Self* [1992, 1993] showed that the long-term trend of insolation at 65°N in July was declining during the period of the Toba eruption, indicating that solar forcing was pushing the Earth toward glacial conditions. CO_2 concentration was about 230 ppm, only 60% of current concentrations.

An additional mechanism, not considered here, is the increase in surface albedo that would be caused by the ash blanket from a supereruption that would cover the land near the eruption. *Jones et al.* [2007] used a GCM to show that while an ash blanket deposited over North America from a Yellowstone eruption would have large local effects, it would only have a 0.1 K global average cooling. Their experimental design, which ignored the potential moderating effect of a stratospheric aerosol cloud or vegetation growth on the ash, results in an upper bound on the ash effect. For Toba, much of the ash fell in the Indian Ocean, so we expect this effect to have also been small for Toba.

Clearly, a volcanic eruption is not required to produce a glaciation, so it is obvious that if the climate system was poised to cool dramatically anyway, a slight nudge could have sped it along. With lower CO_2 concentrations, different solar activity, or even different vegetation patterns producing a different planetary albedo [*Sagan et al.*, 1979], the sensitivity of the climate system to massive radiative forcing might have been higher and maybe more prone to switch. Such experiments remain to be done, so we interpret the experiments presented here as answering the question of whether a Toba-like eruption could produce an ice age today, and the answer is "no."

2. Climate models and experiments

We conducted six GCM experiments, using the National Aeronautics and Space Administration Goddard Institute for Space Studies ModelE, in two different configurations, and the National Center for Atmospheric Research Community Climate System Model 3.0

(CCSM3.0). We describe each of the models and the experiments here, and the results in the next section.

We used CCSM3.0 [*Collins et al.*, 2006; *Kiehl et al.*, 2006] at T31 resolution with atmospheric and land surface resolution of 2.8° by 2.8° for physics calculations, with the standard 26 levels in the vertical topping out at about 36 km. The ocean as well as the dynamic/thermodynamic sea ice model were run at a resolution of roughly 1° with 40 levels in the vertical in the ocean. The land component, the Community Land Model (CLM) 3.0 [*Dickinson et al.*, 2006], was coupled to a dynamic vegetation model to produce the Community Land Model-Dynamic Global Vegetation Model (CLM-DGVM) [*Levis et al.*, 2004], to evaluate whether the vegetation response to the large forcing acts as a feedback. The simulated vegetation represents a potential vegetation that is consistent with the climate. The CLM-DGVM grows plants and simulates vegetation cover and leaf area index to enable two-way coupling between the atmosphere and vegetation. Plant community composition and ecosystem structure are updated annually in response to the establishment of new plants, resource competition, growth, mortality, and fire, and therefore respond to the integrated conditions experienced during a year. Leaf phenology (the seasonal emergence and senescence of leaves) updates leaf area index daily in response to air temperature or soil water. The CCSM3.0 simulation was forced with the Pinatubo SO_2 history, but scaled up by a factor of 100, the same as *Jones et al.* [2005], by prescribing the aerosol amounts. We started the model with a balanced climate with forcing from 6000 years BP, but with current CO_2 for the vegetation. (This mismatch was not intentional, but should make no difference in the results, as the Toba forcing is so large. In this run, the Northern Hemisphere summer solar forcing was a little higher than present and CO_2 was lower.)

To investigate the effect of the size of the volcanic forcing on the broader climate system, we ran the stratospheric version of the ModelE climate model with 4° latitude by 5° longitude horizontal resolution and 23 levels in the vertical up to 80 km [*Schmidt et al.*, 2006]. It was fully coupled to a 4° latitude by 5° longitude dynamic ocean with 13 levels in the vertical [*Russell et al.*, 1995]. It is important to use a full dynamic ocean in these simulations, as was done by *Jones et al.* [2005] and in all our simulations, to obtain the most realistic climate response to such a large forcing. This model is connected to an aerosol module [*Koch et al.*, 2006] that accounts for SO_2 conversion to sulfate aerosols, though with fixed oxidant amounts, and transport and removal of the aerosols. The radiative forcing from the aerosols is fully interactive with the atmospheric circulation. This climate model has been tested extensively in global warming experiments [*Hansen et al.*, 2005; *Schmidt et al.*, 2006] and to examine the effects of volcanic eruptions on climate [*Shindell et al.*, 2004; *Oman et al.*, 2005, 2006a, 2006b], nuclear winter [*Robock et al.*, 2007a, 2007b], and geoengineering [*Robock et al.*, 2008]. The 1991 Pinatubo eruption put 20 Tg of SO_2 into the stratosphere, and we forced the model with a range of injection amounts, 33, 100, 300, and 900 times the Pinatubo amount, ran the model each time for 10 years, and compared the results to a long control run with no volcanic forcing.

For the 300 times Pinatubo run, we input 6 Gt of SO_2, based on *Bekki et al.* [1996], over a 7 day period. These were put into the model in the grid box containing 2°N, 100°E evenly over the altitude range 24-32 km. All the ModelE experiments were forced with the same time and space distributions of SO_2, but with different amounts. It is possible that the Toba injection was higher, but the higher the injection the lower the atmospheric density and the faster the fall speed of the particles, so, as long as the injection is into the lower stratosphere, the exact altitude range would only have a small impact on the cloud lifetime.

A second experiment was conducted with ModelE to investigate the effects of stratospheric chemistry. We used the same ModelE atmospheric configuration, but stratospheric interactive atmospheric chemistry was included [*Shindell et al.*, 2006] to see if this would substantially modify the results. The atmospheric chemistry model extends from the surface to the mesosphere, and is fully coupled to the aerosol model, so that oxidant changes affect sulfate and sulfate affects heterogeneous chemistry. The use of prognostic hydroxyl radical (OH), rather than fixed concentrations, allows it to be depleted, slowing the conversion of SO_2 to sulfate aerosol. Changes in stratospheric ozone will cause a radiative forcing directly, and will also alter the radiative flux reaching the troposphere with potential impacts on the lifetime of many gases there including methane, another potent greenhouse gas. This run used a different ocean GCM, the HYCOM model [*Sun and Bleck*, 2006]. The model was forced with 300 times the Pinatubo emission and, as above, the volcanic eruption is simulated by injecting sulfur dioxide into the model.

3. Results

The dynamic vegetation model results with the NCAR climate model are shown in Fig. 2. Using the scaled-up Pinatubo mass, the volcanic forcing lasted for less than four years, and the climate response was almost identical with *Jones et al.* [2005], with a maximum global-average cooling of 10 K and a cooling of about 2 K after a decade, followed by a relatively quick recovery with no indication of the initiation of large ice sheets. Global average precipitation declined by about 45% for several years and then recovered like the temperature. This appears to be a robust result, as it came from a completely different climate model, but with the same forcing. The coupled vegetation feedback appears to have had little large-scale effect, although we cannot rule out some influence (we have not performed a simulation with fixed vegetation).

The vegetation distribution did change dramatically as a result of the reduction of sunlight and the following cooling and precipitation reduction induced by the large aerosol cloud (Fig. 3). Broadleaf evergreen trees virtually disappear, as do tropical deciduous trees, with midlatitude deciduous trees also reduced. These trees gradually reappear with warming in the following decades. However, the effect of this vegetation change on climate seems minimal. In retrospect, this result is not that surprising. Amazon deforestation experiments have shown that while removing trees and replacing them with grasses does increase the surface albedo, which would cool the surface, it also increases the Bowen ratio, which would warm the surface, as more radiation goes into heating the surface than into evapotranspiration [e.g., *Zhang et al.*, 1996]. The response we found here is similar.

When we forced ModelE with 33, 100, 300, and 900 times the Pinatubo SO_2, we got a radiative forcing that lasted for 4-7 years (Figs. 4, 5). We consider 300 times Pinatubo (6 Gt SO_2) the most likely amount, and so analyze that case in more detail. With the SO_2 injected into the equatorial stratosphere, the resulting sulfate aerosols are transported poleward and then gradually removed from the system, but produce a huge optical depth (Fig. 5) and global average change in downward shortwave radiation stronger than -100 W m^{-2} for a couple years (Fig. 4). The atmospheric residence time of these aerosols, as calculated by our model, agrees very well with the length of time of the deposition observed in ice cores [*Zielinski et al.*, 1996]. Global cooling for the four runs ranges from 8 to 17 K, with the larger forcing also producing a slightly longer cooling. But the response is quite non-linear. After virtually all the sunlight is reflected, additional aerosols have a small effect. For the 300 times Pinatubo case, the maximum cooling is 15 K for three years after the eruption and then reduces to 5 K after a decade. No matter what the amount of SO_2, there is no evidence for ice age initiation. Although snow persists for several summers in the midlatitudes of the Northern Hemisphere, it melts as the aerosols leave the atmosphere and full insolation returns. Figure 4 also shows that global average precipitation is reduced by 30-60% for several years, and this, along with the dark conditions, would also have a strong impact on vegetation.

Because of the large stratospheric heating from the volcanic aerosols, mainly from absorption of upwelling longwave radiation from the troposphere, the tropical tropopause cold trap is warmed substantially, resulting in huge increases in stratospheric water vapor (Fig. 6). Thus the idea of *Rampino and Self* [1992] and *Bekki et al.* [1996] that the volcanic SO_2 would linger, having used up all the stratospheric water vapor initially, is not supported by this experiment that actually models the phenomenon with a climate model that can simulate the processes.

When we simulate 300 times the Pinatubo SO_2 with a model with full atmospheric chemistry, we find a larger and more long-lasting response than without the interactive chemistry, but still no evidence of ice age initiation (Fig. 7). The maximum cooling is about 18 K. The details of the conversion from SO_2 to sulfate show that inclusion of interactive chemistry causes the peak sulfate loading to occur about two years later then in the fixed OH simulation. The peak sulfate loading is less than the fixed OH simulation but its distribution in time is much broader (i.e., longer lasting) causing a slightly larger and longer lasting cooling. For example, cooling greater than 10 K persists for about 6 years in the fixed OH simulation (Figure 4) but lasts for ~11 years in the interactive chemistry experiment. Stratospheric ozone increases modestly following the eruption, as the reduction in reactive hydrogen oxides that can catalytically destroy ozone more than offsets additional losses following reactions on sulfate aerosol surfaces. Hence stratospheric ozone changes contribute a relative small positive forcing rather than amplifying the volcanic cooling. Tropospheric OH is reduced as the solar flux reaching the troposphere is dramatically decreased by the volcanic aerosol layer. This causes a substantial change to the oxidation rate of many species in the troposphere, including methane. However, the reduced oxidation capacity does not last long enough to lead to large changes in tropospheric methane, though it does increase modestly, providing an additional warming chemical response.

4. Discussion and conclusions

We have not been able to show that a Toba-like eruption could produce a glacial advance, given the current distribution of solar radiation and current CO_2 or similar radiation and lower CO_2. But a Toba-like eruption could certainly produce a decade-long volcanic winter, with serious effects on plant and animal life.

Joshi and Shine [2003] describe how volcanic eruptions warm the tropical tropopause, resulting in additional water vapor injected into the lower stratosphere. For eruptions the size of the 1991 Pinatubo eruption, they suggest that the additional water vapor produces a positive radiative forcing of about 0.1 W m^{-2}, slightly counteracting the cooling from the eruption. This feedback is incorporated in the simulations here, and in spite of the additional water vapor, a very large eruption would still produce very large cooling. In addition, we have shown that, with a model including chemistry, the stratospheric forcing and response is prolonged for a couple years for a 300 times Pinatubo injection, but water vapor is not a limiting factor.

Ambrose [1998] and *Rampino and Ambrose* [2000] suggested that the human race narrowly passed through a population bottleneck at some time around this eruption, possibly enhanced by the 1000-year cold period following the Toba eruption. While our results show that indeed the eruption could have produced great stress on humans and their environment, it would have been quite concentrated in the few very dark, cold, and dry years immediately following the eruption. We find no evidence that the 1000-year cold period seen in Greenland ice core records was directly generated by the Toba eruption, although the temporal inference shown by *Zielinski et al.* [1996] is certainly quite intuitive. Gradual global cooling of a few degrees would have provided enough time for population migrations, but the effects of a sudden winter would not. In the past our species survived many glacial advances and retreats, but it is probable that the sudden dark, cold, and dry conditions that followed the supereruption of Toba about 74,000 years ago could have largely destroyed the food supplies of humans and therefore caused a significant reduction in population sizes. Obviously, some human populations survived the climate changes, and *Petraglia et al.* [2007] found evidence in one location in India for human presence both before and after the Toba eruption. Louys [2007] actually found extinctions of several species at four different southeast Asia locations following the Toba eruption, but most species reappeared at those locations following the eruption.

Gathorne-Hardy and Harcourt-Smith [2003] found no evidence for a human bottleneck, but absence of evidence is not evidence of absence. Furthermore, they used estimates of climate change after the eruption smaller than found by us and *Jones et al.* [2005]. *Ambrose* [2003] convincingly refuted the arguments of *Gathorne-Hardy and Harcourt-Smith* [2003], and showed that a large "volcanic winter," followed by a 1000-year period of cool temperatures, could have produced the bottleneck. In fact the results of *Jones et al.* [2005] and the results presented here show that the volcanic winter would have been colder and longer-lasting than Ambrose assumed, which strengthens his argument.

Acknowledgments. We thank two anonymous reviewers and Stephen Self for valuable suggestions, and Greg Faluvegi for scientific programming support at GISS. This work is supported by NASA grant NNG05GB06G and NSF grants ATM-0313592, ATM-0351280, and ATM-0730452. Model development and computer time at GISS are supported by NASA climate modeling and atmospheric chemistry modeling grants. The National Center for Atmospheric Research is supported by the National Science Foundation.

References

Ambrose, S. H. (1998), Late Pleistocene human population bottlenecks, volcanic winter, and differentiation of modern humans, *J. Human Evolution*, *34*, 623-651.

Ambrose, S. H. (2003), Did the super-eruption of Toba cause a human population bottleneck? Reply to Gathorne-Hardy and Harcourt-Smith, *J. Human Evolution*, *45*, 231-237.

Bekki, S., J. A. Pyle, W. Zhong, R. Toumi, J. D. Haigh, and D. M. Pyle (1996), The role of microphysical and chemical processes in prolonging the climate forcing of the Toba eruption, *Geophys. Res. Lett.*, *23*, 2669-2672.

Collins, W. D., et al. (2006), The Community Climate System Model version 3 (CCSM3), *J. Climate*, *19*, 2122-2143.

Dickinson, R. E., K. W. Oleson, G. B. Bonan, F. Hoffman, P. Thorton, M. Vertenstein, Z.-L. Yang, and X. Zeng (2006), The Community Land Model and its climate statistics as a component of the Community Climate System Model, *J. Climate*, *19*, 2302-2324.

Gathorne-Hardy F. J., and W. E. H. Harcourt-Smith (2003), The super-eruption of Toba, did it cause a human bottleneck?, *J. Human Evolution*, *45*, 227-230.

Hansen, J., et al. (2005), Efficacy of climate forcings, *J. Geophys. Res.*, *110*, D18104, doi: 10.1029/2005JD005776.

Harris, B. (2008), The potential impact of super-volcanic eruptions on the Earth's atmosphere, *Weather*, *63*(8), 221-225, doi:10.1002/wea.263.

Huang, C.-Y., M. Zhao, C.-C. Wang, and G. Wei (2001), Cooling of the South China Sea by the Toba eruption and correlation with other climate proxies ~71,000 years ago, *Geophys. Res. Lett.*, *28*, 3915-3918.

Jones, G. S., J. Gregory, P. Stott, S. Tett, and R. Thorp (2005), An AOGCM simulation of the climate response to a volcanic super-eruption. *Clim. Dyn.*, *25*, 725-738.

Jones, M, T., Sparks, R. S. J., and P. J. Valdes (2007), The climatic impact of supervolcanic ash blankets, *Clim. Dyn.*, *29*, 553-564, doi:10.1007/s00382-007-0248-7.

Joshi, M. M., and K. P. Shine (2003), A GCM study of volcanic eruptions as a cause of increased stratospheric water vapor, *J. Climate*, *16*, 3525-3534.

Kiehl, J. T., C. A. Shields, J. J. Hack, and W. D. Collins, (2006), The climate sensitivity of the Community Climate System Model Version 3 (CCSM3), *J. Climate*, *19*, 2584-2596.

Koch, D., G. A. Schmidt, and C. V. Field (2006), Sulfur, sea salt, and radionuclide aerosols in GISS ModelE, *J. Geophys. Res.*, *111*, D06206, doi:10.1029/2004JD005550.

Levis S., G. B. Bonan, M. Vertenstein, and K. W. Oleson (2004), *The Community Land Model's dynamic global vegetation model (CLM-DGVM): Technical description and user's guide*. NCAR Tech. Note TN-459+IA, National Center for Atmospheric Research, Boulder, Colorado, 50 pp.

Li, X. S., and A. Berger (1997), *Simulating climatic impact of the Toba eruption with the LLN 2-D climate model*, *Sci. Report 1997/14*, Institut d'Astronomie et de Géophysique G. Lemaitre, Université catholique de Louvain, Louvain-la-Neuve, Belgium.

Louys, J. (2007), Limited effect of the Quaternary's largest super-eruption (Toba) on land mammals from Southeast Asia, *Quaternary Sci. Rev.*, *26*, 3108-3117.

Oman, L., A. Robock, G. Stenchikov, G. A. Schmidt, and R. Ruedy (2005), Climatic response to high latitude volcanic eruptions, *J. Geophys. Res.*, *110*, D13103, doi:10.1029/2004JD005487.

Oman, L., A. Robock, G. L. Stenchikov, and T. Thordarson (2006a), High-latitude eruptions cast shadow over the African monsoon and the flow of the Nile, *Geophys. Res. Lett.*, *33*, L18711, doi:10.1029/2006GL027665.

Oman, L., A. Robock, G. Stenchikov, T. Thordarson, D. Koch, D. Shindell and C. Gao (2006b), Modeling the distribution of the volcanic aerosol cloud from the 1783-1784 Laki eruption, *J. Geophys. Res.*, *111*, D12209, doi:10.1029/2005JD006899.

Oppenheimer, C. (2002), Limited global change due to the largest known Quaternary eruption, Toba ≈74 kyr BP? *Quaternary Sci. Rev.*, *21*, 1593-1609.

Petraglia, M., et al. (2007), Middle Paleolithic assemblages from the Indian subcontinent before and after the Toba super-eruption, *Science*, *317*, 114-116, doi:10.1126/science.1141564.

Rampino, M. R., and S. H. Ambrose (2000), Volcanic winter in the Garden of Eden: The Toba super-eruption and the Late Pleistocene human population crash, in *Volcanic Hazards and Disasters in Human Antiquity*, F. W. McCoy and G. Heiken, Eds., Geological Society of America Special Paper 345, pp. 71-82.

Rampino, M. R., and S. Self (1992), Volcanic winter and accelerated glaciation following the Toba super-eruption, *Nature*, *359*, 50-52.

Rampino, M. R., and S. Self (1993), Climate-volcanism feedback and the Toba eruption of 74,000 years ago, *Quat. Res.*, *40*, 269-280.

Robock, A. (2000), Volcanic eruptions and climate, *Rev. Geophys.*, *38*, 191-219.

Robock, A., L. Oman, G. L. Stenchikov, O. B. Toon, C. Bardeen, and R. P. Turco (2007a), Climatic consequences of regional nuclear conflicts, *Atm. Chem. Phys.*, *7*, 2003-2012.

Robock, A., L. Oman, and G. L. Stenchikov (2007b), Nuclear winter revisited with a modern climate model and current nuclear arsenals: Still catastrophic consequences, *J. Geophys. Res.*, *112*, D13107, doi:10.1029/2006JD008235.

Robock, A., L. Oman, and G. Stenchikov (2008), Regional climate responses to geoengineering with tropical and Arctic SO_2 injections, *J. Geophys. Res.*, *113*, D16101, doi:10.1029/2008JD010050.

Russell, G. L., J. R. Miller, and D. Rind (1995), A coupled atmosphere-ocean model for transient climate change, *Atmos.-Ocean*, *33*, 683-730.

Sagan, C., O. B. Toon, and J. B. Pollack, (1979), Anthropogenic albedo changes and the Earth's climate, *Science*, *206*, 1363-1368, doi:10.1126/science.206.4425.1363.

Schmidt, G. A., et al. (2006), Present day atmospheric simulations using GISS ModelE: Comparison to in-situ, satellite and reanalysis data, *J. Climate*, *19*, 153-192.

Schulz H., K.-C. Emeis, H. Erlenkeuser, U. von Rad, and C. Rolf (2002), The Toba volcanic event and interstadial/stadial climates at the marine isotopic stage 5 to 4 transition in the Northern Indian Ocean, *Quat. Res.*, *57*, 22-31.

Self, S., and S. Blake (2008), Consequences of explosive supereruptions, *Elements*, *4*, 41-46, doi:10.2113/gselements.4.1.41.

Shindell, D. T., G. Faluvegi, N. Unger, E. Aguilar, G. A. Schmidt, D. Koch, S. E. Bauer, and R. L. Miller (2006), Simulations of preindustrial, present-day, and 2100 conditions in the NASA GISS composition and climate model G-PUCCINI, *Atmos. Chem. Phys.*, *6*, 4427-4459.

Shindell, D. T., G. A. Schmidt, M. E. Mann, and G. Faluvegi (2004), Dynamic winter climate response to large tropical volcanic eruptions since 1600, *J. Geophys. Res.*, *109*, D05104, doi:10.1029/2003JD004151.

Sun, S., and R. Bleck (2006), Multi-century simulations with the coupled GISS-HYCOM climate model: Control experiments, *Clim. Dyn.*, *26*, 407-428.

Zhang, H., A. Henderson-Sellers, and K. McGuffie (1996), Impacts of tropical deforestation. Part I: Process analysis of local climatic change, *J. Climate*, *9*, 1497-1517.

Zielinski, G. A., P. A. Mayewski, L. D. Meeker, S. Whitlow, M. S. Twickler, and K. Taylor (1996), Potential atmospheric impact of the Toba mega-eruption ~71,000 years ago, *Geophys. Res. Lett.*, *23*, 837-840.

Figure Captions

Figure 1. The 1000 rupiah banknote from Indonesia showing Danau Toba (Lake Toba), the lake-filled caldera left from the largest eruption of the past 100,000 years. (Some sophisticated countries have volcanoes right on the money.) Also shown is a Google Earth image of the caldera on the island of Sumatra. The lake is about 86 km long and 30 km wide, and has a large island inside, the resurgent block of the caldera.

Figure 2. Monthly-average global-mean temperature (red), and precipitation (blue) anomalies (top), and downward shortwave radiation anomalies at the surface (bottom), for the NCAR CLM3.0 climate model forced with 100 times the amount of aerosol as the 1991 Mt. Pinatubo eruption in the Philippines put into the stratosphere.

Figure 3. Top: Percent coverage of four major vegetation types for the year before the eruption. Bottom: As above, but averaged for the four years just after the eruption. Broadleaf evergreen trees virtually disappear, as do tropical deciduous trees.

Figure 4. Changes in temperature (red), and precipitation (black) (top), and monthly-average global mean downward shortwave radiation at the surface (bottom), for the ModelE simulations without interactive chemistry. The 0.67 Gt SO_2 case is 33 times Pinatubo, 2 Gt SO_2 is 100 times Pinatubo, 6 Gt SO_2 is 300 times Pinatubo, and 18 Gt SO_2 is 900 times Pinatubo.

Figure 5. Latitudinal and seasonal distribution of optical depth in the mid-visible for the 300 times Pinatubo forcing used for the ModelE simulation without interactive chemistry.

Figure 6. Changes in monthly-average global mean specific humidity for the ModelE 300 times Pinatubo simulation without interactive chemistry, showing large reductions of water vapor in the cold troposphere, but huge increases in the stratosphere, as the tropical tropopause cold trap is warmed by the volcanic aerosols in the lower stratosphere.

Figure 7. Climate and chemistry results from the 300 times Pinatubo ModelE simulation with interactive chemistry. Top: Changes in global average surface air temperature and global average planetary albedo. Middle: Atmospheric burdens of sulfur dioxide and sulfate. Bottom: Percentage change (relative to mean prior to eruption) of OH in the lower stratosphere, OH in the troposphere, and ozone in the lower stratosphere. Dates are arbitrary, and the eruption is simulated in year 0.

Figure 1. The 1000 rupiah banknote from Indonesia showing Danau Toba (Lake Toba), the lake-filled caldera left from the largest eruption of the past 100,000 years. (Some sophisticated countries have volcanoes right on the money.) Also shown is a Google Earth image of the caldera on the island of Sumatra. The lake is about 86 km long and 30 km wide, and has a large island inside, the resurgent block of the caldera.

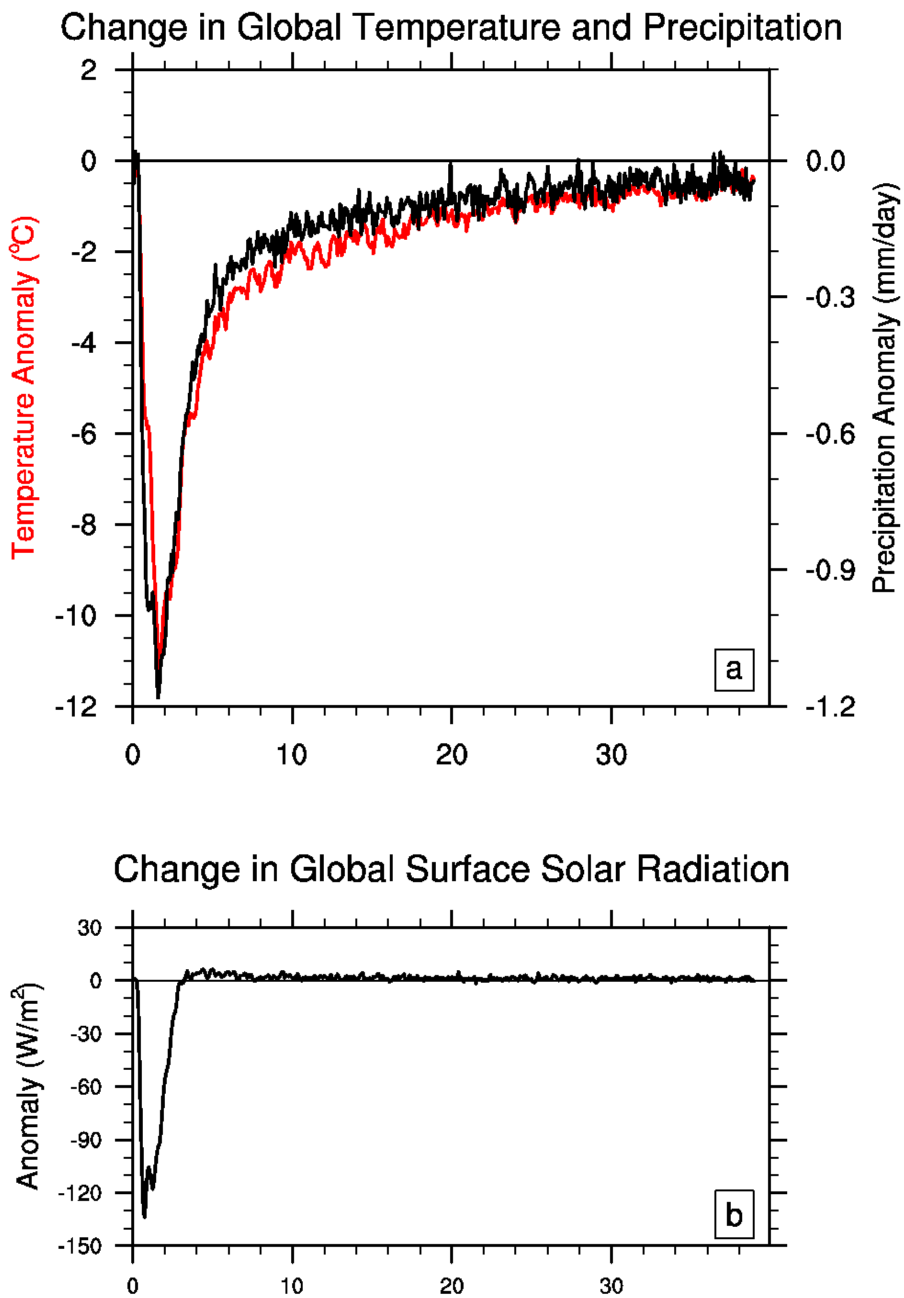

Figure 2. Monthly-average global-mean temperature (red), and precipitation (blue) anomalies (top), and downward shortwave radiation anomalies at the surface (bottom), for the NCAR CLM3.0 climate model forced with 100 times the amount of aerosol as the 1991 Mt. Pinatubo eruption in the Philippines put into the stratosphere.

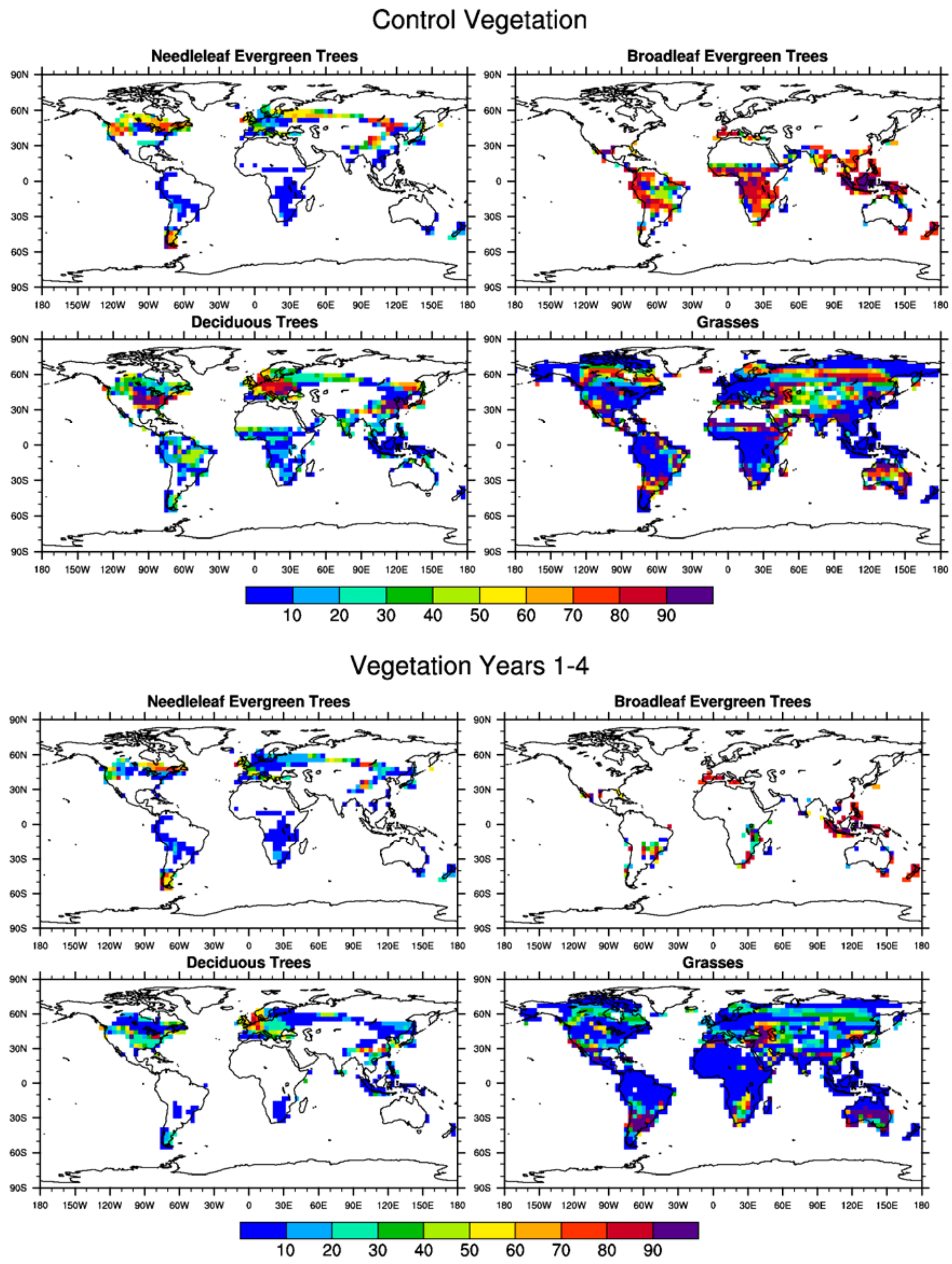

Figure 3. Top: Percent coverage of four major vegetation types for the year before the eruption. Bottom: As above, but averaged for the four years just after the eruption. Broadleaf evergreen trees virtually disappear, as do tropical deciduous trees.

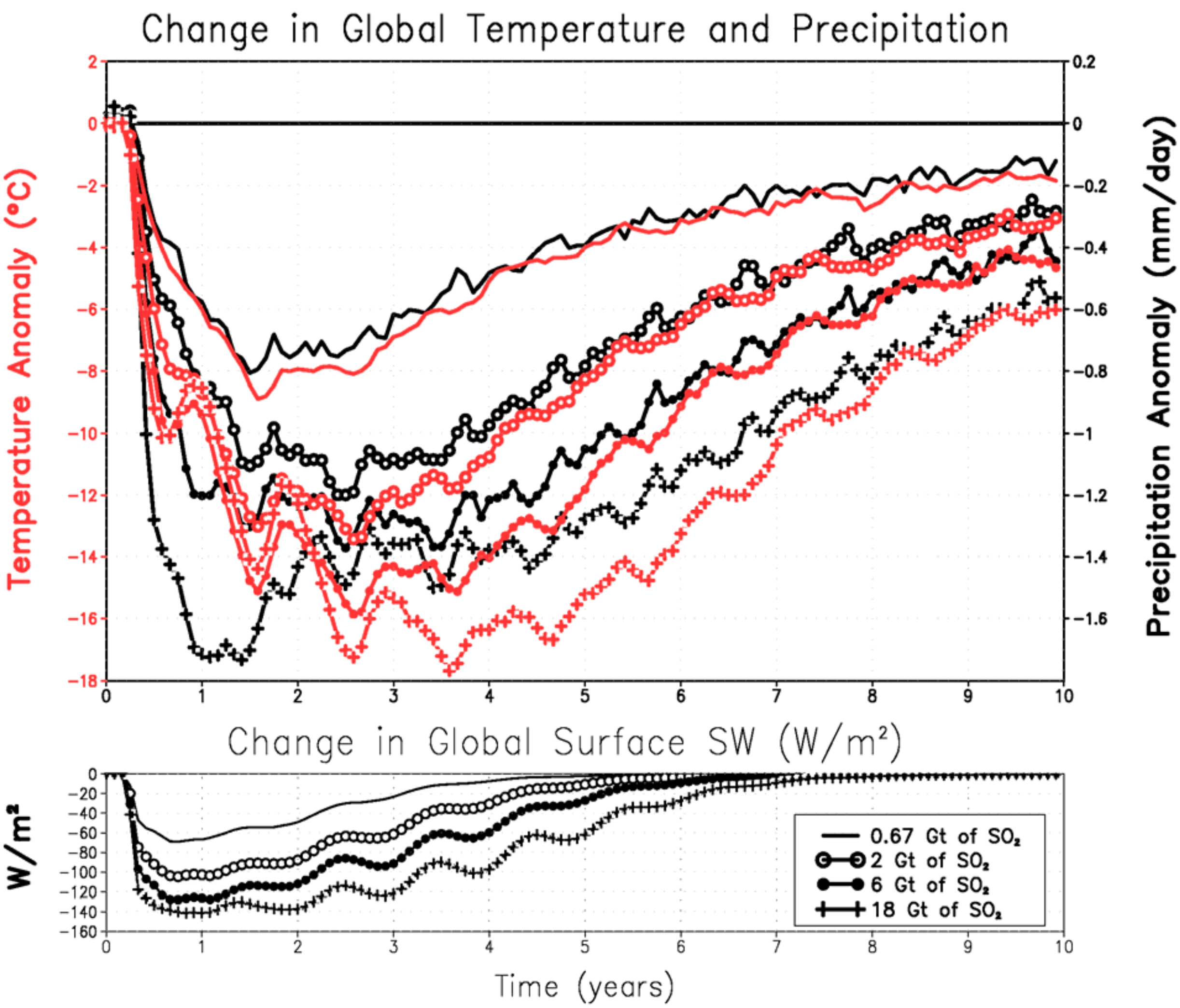

Figure 4. Changes in temperature (red), and precipitation (black) (top), and monthly-average global mean downward shortwave radiation at the surface (bottom), for the ModelE simulations without interactive chemistry. The 0.67 Gt SO_2 case is 33 times Pinatubo, 2 Gt SO_2 is 100 times Pinatubo, 6 Gt SO_2 is 300 times Pinatubo, and 18 Gt SO_2 is 900 times Pinatubo.

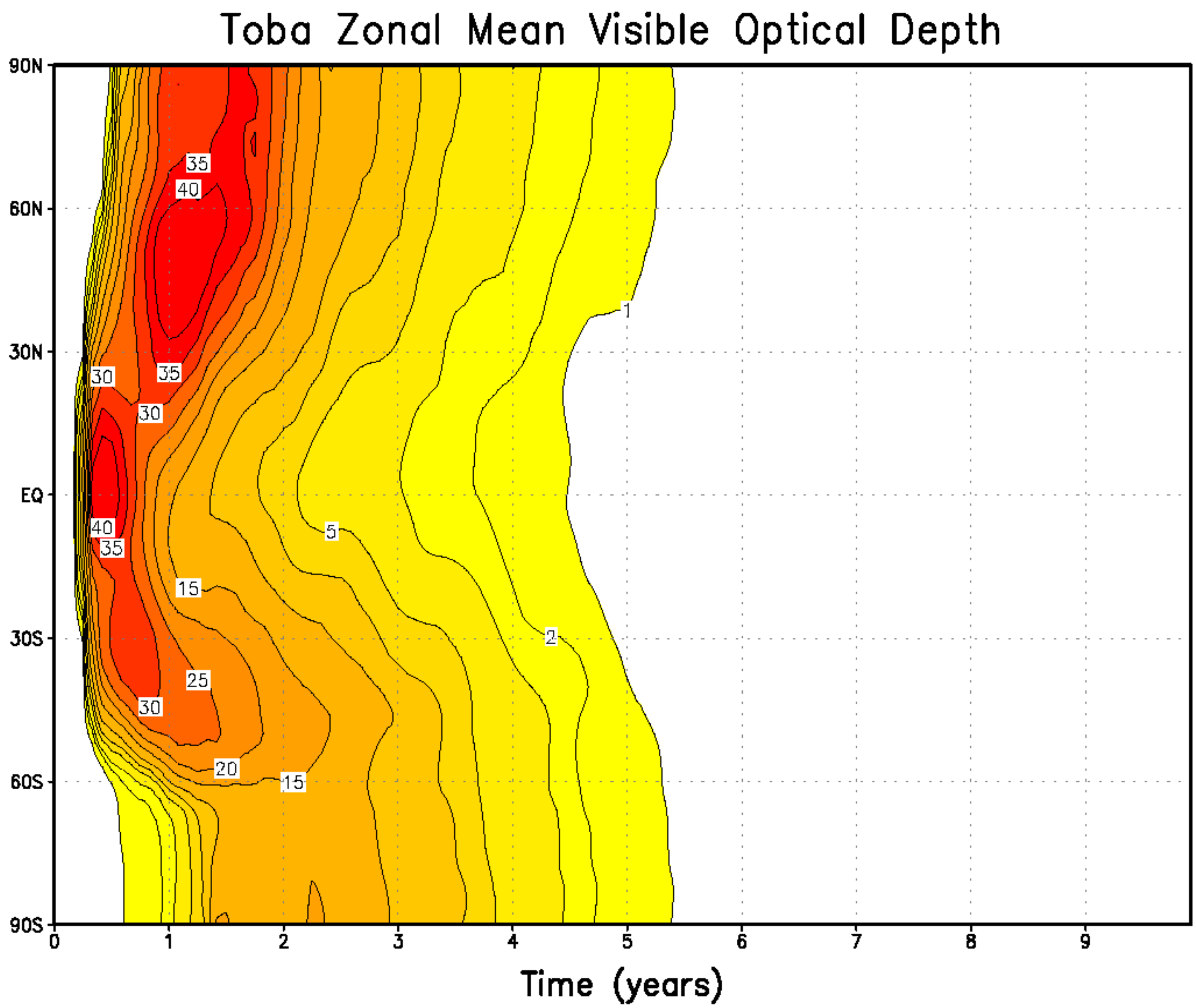

Figure 5. Latitudinal and seasonal distribution of optical depth in the mid-visible for the 300 times Pinatubo forcing used for the ModelE simulation without interactive chemistry.

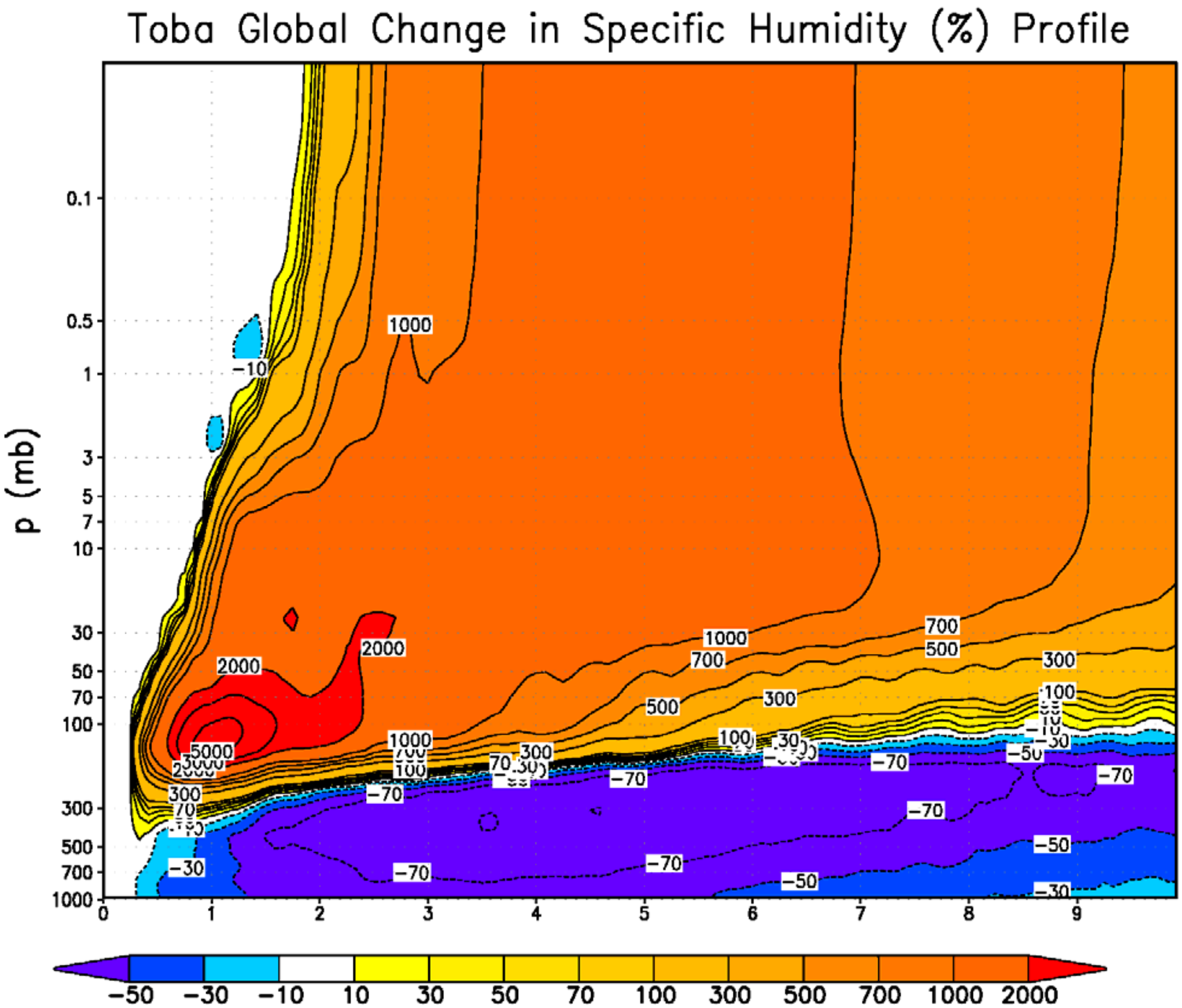

Figure 6. Changes in monthly-average global mean specific humidity for the ModelE 300 times Pinatubo simulation without interactive chemistry, showing large reductions of water vapor in the cold troposphere, but huge increases in the stratosphere, as the tropical tropopause cold trap is warmed by the volcanic aerosols in the lower stratosphere.

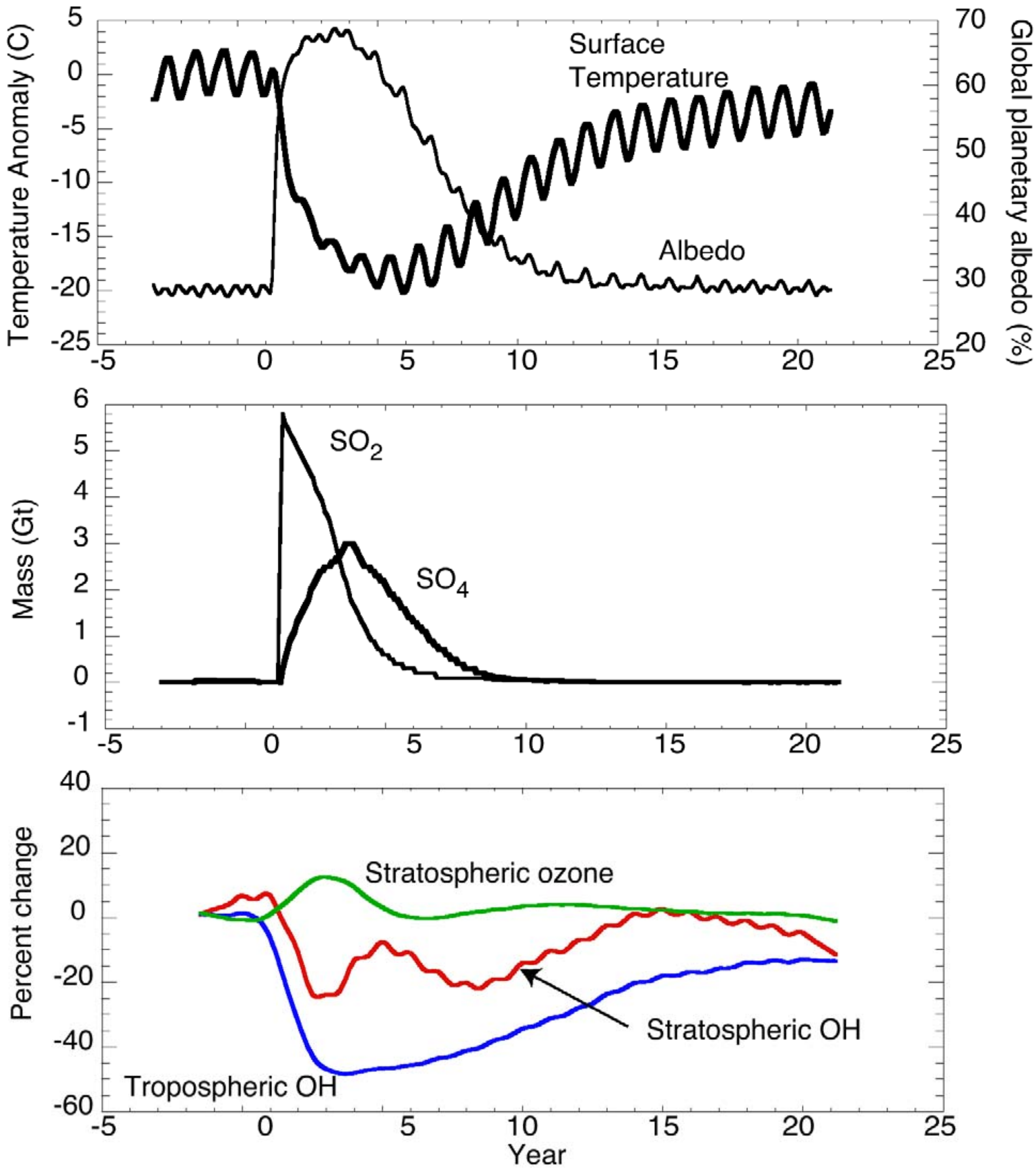

Figure 7. Climate and chemistry results from the 300 times Pinatubo ModelE simulation with interactive chemistry. Top: Changes in global average surface air temperature and global average planetary albedo. Middle: Atmospheric burdens of sulfur dioxide and sulfate. Bottom: Percentage change (relative to mean prior to eruption) of OH in the lower stratosphere, OH in the troposphere, and ozone in the lower stratosphere. Dates are arbitrary, and the eruption is simulated in year 0.