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The climatic effects of the direct injection of water vapour into the stratosphere by large volcanic eruptions

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Abstract

We describe a novel mechanism that can significantly lower the amplitude of the climatic response to certain large volcanic eruptions. The proximity of oceans to some volcanoes can cause significant entrainment of water into coignimbrite clouds during the eruption. If sufficiently large amounts of this entrained water vapour enter the stratosphere, a climatically significant amount of water vapour can be left over in the lower stratosphere after the eruption, even after sulphate aerosol formation. This excess stratospheric humidity warms the climate, and acts to balance the climatic cooling induced by the volcanic aerosol, especially because the humidity anomaly lasts for a period that is longer than the residence time of aerosol in the stratosphere. In particular, Northern Hemisphere cooling is reduced in magnitude. We discuss this mechanism in the context of the discrepancy between the observed and modelled cooling following the Krakatau eruption in 1883.

1 Introduction

Volcanic eruptions have the capacity to dramatically alter global climate on long climatic timescales. Large explosive eruptions affect climate by injecting SO₂ into the stratosphere, which eventually combines with existing water vapour to form sulphate aerosols that scatter sunlight and cool the surface. The climatic effects of eruptions have been summarised by e.g., Robock (2000).

Here we describe a novel mechanism that might significantly alter the response of the climate to a large volcanic eruption: so much water vapour is directly injected into the stratosphere by certain eruptions that a climatically significant amount is left over after the formation of stratospheric sulphate aerosol. This stratospheric water vapour anomaly exerts a positive radiative forcing on the surface, offsetting the negative climate forcing from the volcanic stratospheric aerosol, leading to a surface temperature response that is smaller in amplitude (i.e. more positive) than if forced by the volcanic

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aerosol alone.

This mechanism is similar to that of Joshi and Shine (2003), except that they proposed that it was heating of the tropical tropopause layer by volcanic aerosol that allowed more water vapour into the stratosphere, rather than direct injection by the eruption itself. In this case, the injection of significant amounts of water into the stratosphere is achieved by entrainment of large amounts of water into coignimbrite clouds associated with an eruption (Darteville et al., 2002). Moreover, this mechanism will be significantly enhanced for those large volcanic eruptions that occur in close proximity to large bodies of water, such as volcanoes on small islands (Francis and Self, 1983).

Previous studies have suggested the possibility of injection of water directly into the stratosphere from volcanic eruptions; between 10 Mt (Pitari and Mancini, 2002) to 540 Mt (S. Self, private communication, 2007) from Pinatubo, ~2 Gt following Tambora in 1816 (Glaze et al., 1997) and ~27 Gt from the 76 kya Toba eruption (Bekki et al., 1996).

Stratospheric water vapour (hence SWV) can have significant climatic effects; the radiative forcing due to SWV increases of O (1) ppm in the 20th century have been modelled as being up to $+0.5 \text{ Wm}^{-2}$ (e.g., Forster and Shine, 2002). A water vapour anomaly of 0.95 ppmm, or ~1.5 ppmv in the stratosphere is equivalent to a total mass of water of 500 Mt, which is entirely consistent with the inventories described above. The partial cancellation of volcanic cooling by direct injection of water into the stratosphere is therefore entirely possible. Moreover, a humidity anomaly in the tropical lower stratosphere is distributed around the stratosphere by atmospheric motions, eventually dissipating on a timescale of 5–10 years (Hall and Waugh, 1997); in other words the water vapour anomaly outlasts the volcanic aerosol loading by some years.

In this work we shall examine the effects of direct injection using a parameterisation of injection on a state-of-the-art coupled ocean-atmosphere climate model which has the climatic effects of volcanic aerosol included in it. We then discuss the implications of the mechanism for certain specific eruptions, particularly Krakatau in 1883.

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2 Model setup

The model used here is the Hadley Centre's new climate model HadGEM1 (Johns et al., 2006). The version used as the control is the so-called "ALL" integration, which has all natural and anthropogenic forcings since 1860 (Stott et al., 2006). The primary effect of volcanoes on climate is the radiative heating effect of sulphate aerosol in the stratosphere produced from volcanic SO_2 (Robock, 2000). HadGEM1 uses a monthly mean climatology of stratospheric volcanic aerosol optical depth which is converted into a mass mixing ratio and spread evenly above the tropopause. For more details the reader is referred to Stott et al. (2006, 2768–2769).

We have chosen the Krakatau eruption of 1883 as the baseline eruption, as our mechanism may have special significance for it (see below). However, the parameterisation of stratospheric aerosol loading due to Krakatau's eruption displays many characteristics to other large tropical eruptions in climate models, such as spreading in the extratropics on a timescale of 1–2 years. It should therefore be quite general and applicable to many tropical eruptions.

The assumption in this study is that water vapour in the eruption plume rises to a height of ~40 km (Francis and Self, 1983), and that excess SWV left over after combination with SO_2 is evenly mixed in longitude. To simulate this in HadGEM1, we add 500 Mt of water vapour to the model with a constant mixing ratio between 0–40 km and 10 S–0 N over a period of 10 days. The rationale for the zonally averaged shape is that while we wish to simulate the effect of a relatively confined pulse in the tropics, the timescale for such a pulse to travel zonally round the equator is very small. The total time-integrated amount of water vapour artificially added to the model in this way in the stratosphere (above 100 hPa) is then 1.5 ppmv. The water vapour anomaly is advected around the GCM and the difference between these perturbation experiments and the control runs analysed.

Sensitivity experiments have been carried out using initial conditions from different control cases in order to assess the effects of interannual variability. There are two

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scenarios: volcano (denoted V), and volcano plus injected stratospheric water (denoted VSW), and each ensemble has 4 members.

3 Results

Figure 1 shows the evolution of water vapour q at the 50 hPa level, taken as representative for the model stratosphere. The top panel shows q in ensemble V. A wet anomaly can be seen just after the eruption at the equator, which is consistent with volcanic aerosol warming the tropical tropopause layer, and allowing more water vapour into the stratosphere (e.g., Joshi and Shine, 2003).

The middle panel of Fig. 1 shows q in the VSW ensemble. The q anomaly is transported upwards and polewards by the mean circulation and down into the midlatitudes by the stratospheric Brewer-Dobson circulation. A smaller fraction of water vapour is transported into the southern midlatitudes than the northern midlatitudes because of the seasonal asymmetry of the Brewer Dobson circulation. The anomalies decay over a timescale of 5–10 years, consistent with other studies of stratospheric transport (Hall and Waugh, 1997).

The bottom panel of Fig. 1 shows the difference in ensembles average between VSW and V. There is significantly more water vapour at 50 hPa in ensemble VSW than in V; immediately after the eruption there is an excess of 2 ppmv in the tropics, while even after 5 years the difference in the northern midlatitudes is O (1 ppmv), which is known to be enough to have climatically significant effects (Forster and Shine, 1998).

The change in energy balance is shown in Fig. 2. The radiative forcing has been calculated using a method (Forster and Taylor, 2006), which was found to agree well with diagnosed forcings. Changes in top of the atmosphere radiative fluxes, near surface temperature changes together with estimated climate feedback terms can be used to estimate the global mean climate forcing for a transient climate simulation. The climate feedback terms used for HadGEM1 are; $Y_{LW}=2.43 \text{ Wm}^{-2} \text{ K}^{-1}$ and $Y_{SW}=1.05 \text{ Wm}^{-2} \text{ K}^{-1}$ (Forster and Taylor, 2006); these were deduced from increasing CO_2 concentration

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simulations so there is an implicit assumption that the same values are valid for other forcing factors.

The left panel shows the long-wave or LW change, the middle panel the SW change, and the sum is shown on the right panel. A clear SW cooling can be seen in V (in red), which is partially compensated for by LW heating, as volcanic aerosol does display some LW effects (see e.g., Ramachandran et al., 2000). The water vapour anomaly in VSW (shown in blue) exerts a globally averaged LW radiative forcing of $+0.33 \pm 0.09 \text{ Wm}^{-2}$ for the average of the two years following the eruption, which is consistent with the humidity anomaly shown in Fig. 1 (bottom) and previous work (Forster and Shine, 1998). The positive LW forcing also appears to last longer in VSW than V, consistent with the lifetime of the stratospheric water vapour anomaly being longer than the lifetime of the volcanic aerosol. There is very little SW effect, again consistent with stratospheric water vapour anomaly effects being felt mainly in the LW.

The near-surface response with respect to SWV amount is shown in Fig. 3, which shows the evolution of 1.5 m temperature in each ensemble member. The globally averaged change shows that the effect of the water vapour anomaly in VSW is to warm the climate relative to V. The 2 year average following the eruption is 0.10 K warmer in VSW; a T -test gives a p value of 0.06 that the model mean of VSW is not warmer than V. In the Northern Hemisphere, VSW is warmer by 0.10 K and in the Southern Hemisphere by 0.12 K for the average of the 1884–1885 period. A T -test gives p values of 0.16 and 0.03 that the water anomaly does not warm the model mean for the Northern and Southern Hemispheres, respectively. Again, ensemble VSW appears to stay warmer than ensemble V for 5–10 years after the eruption, consistent with the lifetime of the stratospheric vapour anomaly shown in Fig. 1.

The zonally averaged temperature response is shown in Fig. 4. The exaggerated cooling in ensemble V is reduced in ensemble VSW, especially in the northern midlatitudes and polar regions- this is apparent from examining the difference in cooling patterns (Fig. 4c).

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4 Discussion and implications for Krakatau

The water content of the magma, entrainment of water into the eruptive column, and even the season of the eruption significantly affect the amount and distribution of water vapour in the stratosphere 1–2 years after the eruption itself, and hence change the amount of surface cooling compared to the effect of stratospheric aerosol alone. It is also unclear how the amount of water vapour injected into the stratosphere might scale with other factors such as eruptive column height and amount of coignimbrite clouds. Such nonlinearities might prove to be significant, and will be investigated in future.

This work has special significance when applied to the eruption of Krakatau in 1883, which was one of the largest eruptions of modern times (Robock, 2000). The amount of magma produced by the eruption is thought to be 2–3 times as large as the amount produced by the eruption of Mt. Pinatubo in 1991 (Self and Rampino, 1981; Scaillet et al., 2003). However, general circulation models (GCMs) consistently overestimate the globally averaged climatic cooling due to this eruption (Knutson et al., 2006; Miles et al., 2004; Stenchikov et al., 2006); such an overestimation is puzzling in the light of the fidelity with which climate models simulate the climatic cooling following the Mt. Pinatubo eruption in 1991 (Randall, 2007).

Figure 5 shows observed near surface temperatures following Krakatau's eruption, and their associated uncertainties (Brohan et al., 2006; Rayner et al., 2006), together with simulated temperatures from the World Climate Research Programme's (WCRP's) Coupled Model Intercomparison Project phase 3 (CMIP3) multi-model dataset. Only those models that simulated volcanic eruptions by changing the aerosol amounts in the stratosphere are included. The CMIP3 models consistently cool more than was observed. The HadGEM1 (V) simulated temperatures are also plotted in Fig. 5, and similarly show a larger cooling (-0.35 K) than observed (-0.12 ± 0.21 K) for the global 2 year mean following the eruption.

A number of theories have been put forward to explain this discrepancy. Natural variability such as the El Niño Southern Oscillation (ENSO) in both the observations

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and the GCMs has been put forward as a reason why the GCMs might actually only appear to overestimate surface cooling. Reconstructions of Niño-3 index (sea surface temperature or SST anomaly between 210° E– 270° E, 5° S– 5° N) do suggest an ENSO peaking in 1885 (Angell, 1988; Mann et al., 2000). However, the time-averaged Niño-3 index in 1884/85 was only approximately 0.6 K, suggesting a weak ENSO event with accompanying global air temperature anomalies of about 0.1 K. Additionally, globally-averaged air temperature anomalies tend to lag the SST anomaly by a few months. It is therefore doubtful that an ENSO event can explain the discrepancy between the models and observations in the two years following the eruption.

Recent work has suggested that GCMs underestimate Northern Eurasian warming following volcanic eruptions because they underestimate the increase in Arctic Oscillation (AO) index associated with stratospheric volcanic aerosol (Stenchikov et al., 2006). However, given the limited spatial and seasonal scale of the anomalies, this effect cannot account for the globally and annually averaged temperature discrepancy.

A recent study describing the GISS modelE simulation suggested a number of possible contributing factors to the apparent discrepancy (Hansen et al., 2007), although they do conclude that there is a reasonable agreement with their model. Starting a model simulation from an initial state that has cooled from previous eruptions, which is not done in many of the CMIP3 simulations, may reduce the simulated response to Krakatau by 10%. However HadGEM1 was initialised from a control simulation that incorporated a mean volcanic aerosol, which may explain why V cools slightly less than the average of the CMIP3 models (Fig. 5). There is an uncertainty on the level of the direct radiative forcing from the volcanic aerosols, which the GISS study estimates as $\pm 50\%$. Using a different observational data set than used here, they suggest that as the observed cooling over land and over SSTs following the 1991 Pinatubo eruption agree and that there may be an issue of the accuracy of the SSTs in the 1880s. However the estimates of observational uncertainty used here suggest that SST global means are more accurately known than land global mean temperatures (Brohan et al., 2006).

We hypothesise that the Krakatau eruption plume was different to that of Pinatubo

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because of the proximity of the former volcano to the sea, and the entrainment of significantly larger amounts of water into the coignimbrite clouds associated with the eruption, as stated above. An injection of 500 Mt of water into the stratosphere is a reasonable number for Krakatau given the different estimates for water entering the stratosphere after different eruptions in previous studies (see above).

Figure 6 shows the observed temperature changes with both the V and VSW simulations for the globe, NH and SH. For the V simulation the uncertainties on the global mean and NH temperatures do not overlap with the observed temperatures for the two years following the eruption, but they do for the VSW simulation. For the SH the VSW simulation is in better agreement than the V simulation, but the uncertainties are large enough that both simulations are consistent with the observations. The VSW simulation shows less cooling than the V simulation and is nearer the observed values.

Inclusion of the climatic effects of water vapour injection has consequences for changes in heat content and sea level in the world's oceans during the 20th century. Previous studies have suggested that the oceans were affected by Krakatau's cooling for some decades (e.g., Glecker et al., 2006). If the cooling associated with Krakatau is indeed less than the models suggest because of the mechanism explained above, model predictions of 20th century heat content and sea level change may be biased.

5 Conclusions

We have shown, using plausible volcanic inventories, that large volcanic eruptions can deposit a climatically significant amount of water in the stratosphere, which itself has a warming effect, i.e. it is a negative feedback to the cooling associated with the stratospheric volcanic aerosol. We hypothesise that the proximity of a given volcano to the sea, and the increased entrainment of water vapour into the eruption plume is the main cause behind the amplitude of the feedback. This mechanism can account for some of the difference between the observed and modelled temperature changes following the eruption of Krakatau in 1883. We stress that this explanation is one of a num-

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ber of explanations; however, our hypothesis is a plausible explanation given what is known about eruptions near large bodies of water. Additionally, it is interesting that some circumstantial evidence does exist for increased amounts of water in the middle atmosphere following the Krakatau eruption in the form of noctilucent clouds (Schroder, 1999).

This mechanism has implications for the climatic perturbation resulting from the eruption of other volcanoes near large bodies of water, such as Agung in 1963, or even the eruption of Toba 70 kyr ago. Injections of large amounts of water vapour into the stratosphere might significantly cancel the cooling effect associated with such volcanoes. Future research will involve investigating these effects.

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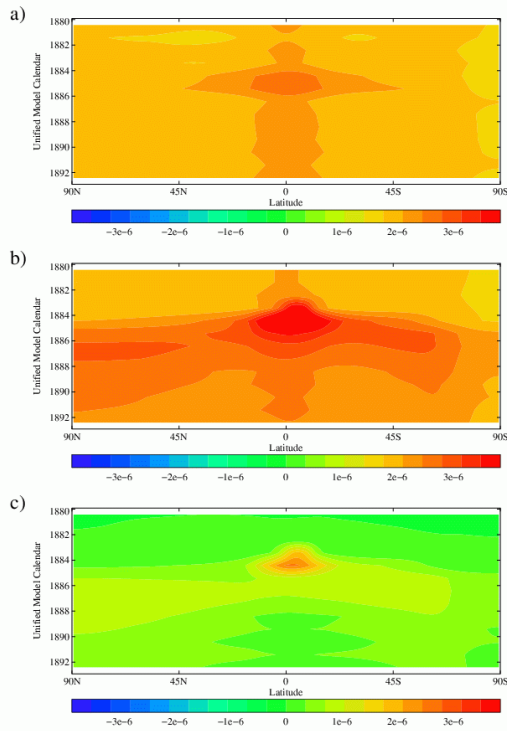


Fig. 1. Zonal cross sections of evolution of stratospheric humidity (ppm) at 50 hPa in ensemble V (top), VSW (middle), and VSW minus V (bottom). The time of the eruption is given as 1883 in order to aid comparison with Krakatau.

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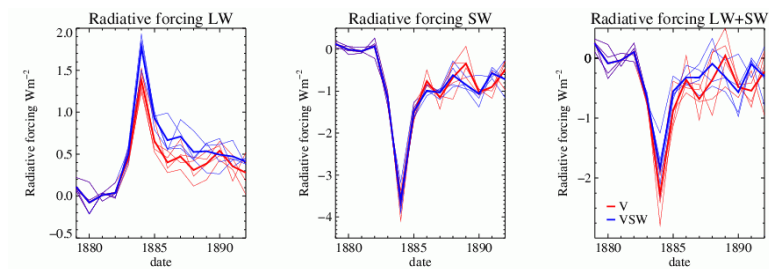


Fig. 2. Left: evolution of the LW radiative balance using the Forster and Taylor method in ensemble V (red) and VSW (blue); middle: for the SW radiative balance; right: LW+SW. Individual ensemble members, thin lines and the ensemble averages thick lines.

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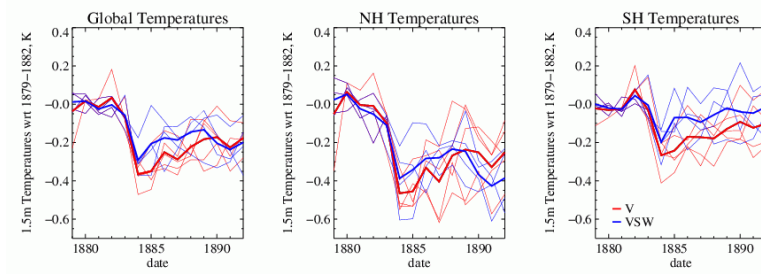


Fig. 3. As in Fig. 2 but the evolution of the globally averaged 1.5 m temperatures (left) NH temperatures (middle); and SH temperatures (right). Temperatures plotted with respect to 1879–1882 mean.

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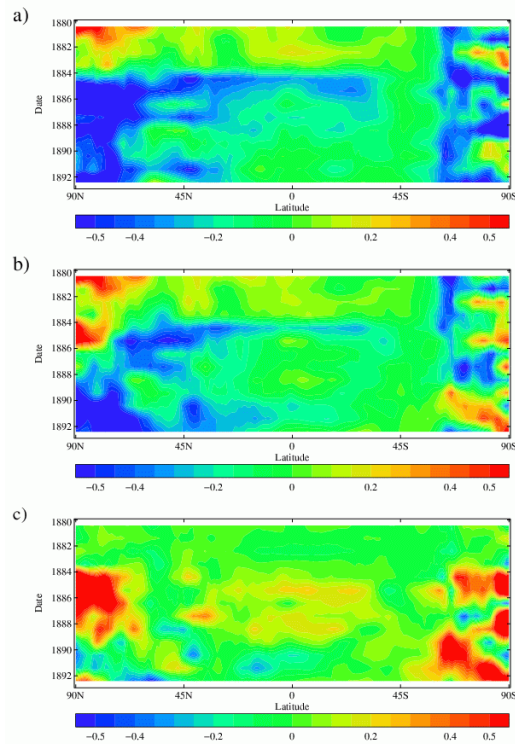


Fig. 4. Zonal cross sections of 1.5 m temperature anomaly minus the 1860–1880 average in ensemble V (top), ensemble VSW (middle), and VSW minus V (bottom) and M 1000 (diamonds). The model is sampled in the same way as the observations.

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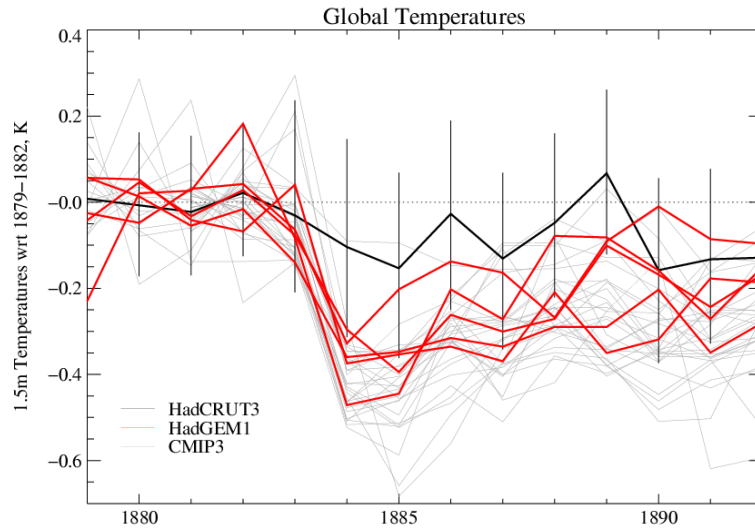


Fig. 5. Global mean temperatures with respect to 1879–1882 mean. Observed temperatures (black) shown with estimate of observational uncertainty (95% range) as vertical bars. CMIP3 “20th century simulations” shown in grey. Only those models that simulated volcanic eruptions by changing the aerosol amounts in the stratosphere are included. HadGEM1 V ensemble members shown in red.

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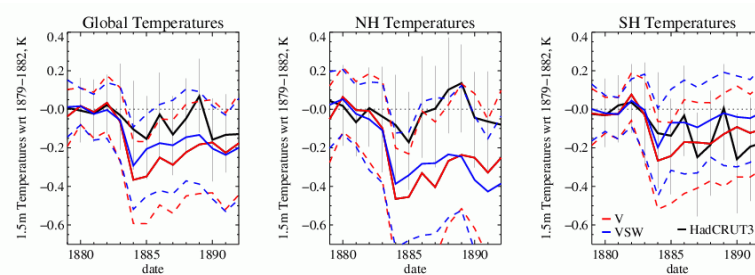


Fig. 6. Global mean, Northern Hemisphere and Southern Hemisphere observed temperatures with respect to 1879–1882 mean. HadGEM1 V ensemble average shown in red with estimate of its uncertainties (95% range) (red dashed). HadGEM1 VSW is as HadGEM1 V but shown in blue.

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