Counteracting the climate effects of volcanic eruptions using short-lived greenhouse gases

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Geophysical Research Letters

RESEARCH LETTER
10.1002/2014GL061886

Key Points:
- Deliberate emission of short-lived GHG could counteract volcanic cooling
- Climate model shows feasibility of counteracting a 3X Pinatubo eruption
- Shorter time scales make counteracting different from other climate engineering

Supporting Information:
- Readme
- Text S1
- Figure S1
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Citation:
Fuglestvedt, J. S., B. H. Samset, and K. P. Shine (2014), Counteracting the climate effects of volcanic eruptions using short-lived greenhouse gases, *Geophys. Res. Lett.*, *41*, doi:10.1002/2014GL061886.

Received 16 SEP 2014
Accepted 2 NOV 2014
Accepted article online 6 NOV 2014

Counteracting the climate effects of volcanic eruptions using short-lived greenhouse gases

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Abstract A large volcanic eruption might constitute a climate emergency, significantly altering global temperature and precipitation for several years. Major future eruptions will occur, but their size or timing cannot be predicted. We show, for the first time, that it may be possible to counteract these climate effects through deliberate emissions of short-lived greenhouse gases, dampening the abrupt impact of an eruption. We estimate an emission pathway countering a hypothetical eruption 3 times the size of Mount Pinatubo in 1991. We use a global climate model to evaluate global and regional responses to the eruption, with and without counteremissions. We then raise practical, financial, and ethical questions related to such a strategy. Unlike the more commonly discussed geoengineering to mitigate warming from long-lived greenhouse gases, designed emissions to counter temporary cooling would not have the disadvantage of needing to be sustained over long periods. Nevertheless, implementation would still face significant challenges.

1. Introduction

Large volcanic eruptions have caused substantial disruption to human civilization [*Self*, 2006]. This disruption may attain a global scale, through a transient cooling induced by sulphur dioxide injected into the stratosphere [*Robock*, 2000]. Although it cannot be predicted when, further disruptive eruptions will occur in the future [*Hyde and Crowley*, 2000; *Myhre et al*., 2013; *Pyle*, 1998; *Self*, 2006] and international contingency planning may be required. Geoengineering is considered as a potential, highly controversial, option for counteracting long-term man-made warming [*Caldeira et al*., 2013; *Crutzen*, 2006; *Royal Society*, 2009]. Volcanic eruptions, however, cause a short-term, natural cooling. Here we consider the extent to which a deliberate but temporary climate warming could be used to counteract such cooling, through emissions of potent but short-lived greenhouse gases. Using a global circulation model, we find that for a sulphate loading 3 times the size of the 1991 Mount Pinatubo eruption, we can substantially counterbalance the surface temperature response both globally and regionally. The precipitation response is more difficult to balance, but at the global average level, the effect of the eruption can be significantly dampened. Although counteracting volcanic eruptions does not suffer from some of the difficulties of long-term geoengineering, its implementation would pose enormous financial, practical, and ethical issues, which may rule it out as a credible strategy for preparing for such eruptions.

Volcanic sulphur dioxide injected into the stratosphere causes the formation of clouds of sulphuric acid droplets, reflecting sunlight back to space, and reducing surface temperature and precipitation in the years following an eruption. Volcanic eruptions cannot be predicted and thus represent a significant source of uncertainty in climate projections [*Intergovernmental Panel on Climate Change* (*IPCC*), 2013; *Self*, 2006]. Over the last millennium, records indicate four eruptions more than 3 times larger than Mount Pinatubo, in terms of amount of sulfate injected into the stratosphere [*Gao et al*., 2008].

Much attention is given to the societal impacts of warming from greenhouse gases (GHGs) due to human activity [*Intergovernmental Panel on Climate Change*, 2014a, 2014b]. Global or regional cooling due to a volcanic eruption could also lead to severe disruption. The largest eruptions are believed to have caused substantial impacts on civilization [*Robock*, 2000; *Self*, 2006]. However, regional climatic impacts are inherently less predictable than global changes, and in the case of volcanic eruptions, the spreading of the aerosol cloud can be strongly dependent on wind conditions at the time of eruption. A natural question is whether we could use current understanding of the climate system, if there was a perceived urgent need to mitigate such a volcanically forced cooling. Here we examine whether counteracting a volcanic cooling through deliberate emission of short-lived GHGs (SLGHGs) is worth a deeper, multidisciplinary study.

Climate models have been shown to be suitable tools to capture the global temperature impact of the 1991 Mount Pinatubo eruption, the last individual eruption of significance for the climate system. A forecast soon after the eruption [*Hansen et al*., 1992] was subsequently shown to agree well with the observed responses [*Intergovernmental Panel on Climate Change*, 1994; *Robock*, 2000]. Here we use the National Center for Atmospheric Research (NCAR) Community Atmosphere Model version 4 (CAM4) [*Gent et al*., 2011], coupled to a slab ocean (which crudely represents the thermal capacity of the upper ocean but not the response of the ocean circulation) to estimate the climate response to a large volcanic eruption, with and without concurrent balancing emissions of a SLGHG. While the ability of present models to simulate the various climate impacts of volcanic eruptions is still being evaluated [see, e.g., *Canty et al*., 2013; *Driscoll et al*., 2012], such an experimental setup helps illustrate many aspects of the problem. After equilibration with year 2000 conditions of GHGs and aerosols, the model was forced by prescribing either changes in stratospheric sulphate aerosol, a short-lived but globally well-mixed GHG, or both, and run for 30 years after the start of the eruption. Stratospheric sulphate concentrations from the 1991 Pinatubo eruption were scaled by a factor of 3, roughly the size of the 1815 Tambora eruption in terms of amount of sulphur emitted.

2. Methods

Simulations were performed with the NCAR Community Earth System Model version 1.0.3 (CESM1.0.3), using the CAM4 with a resolution of 1.9°×2.5°, coupled to the Community Land Model version 4 and a slab ocean model (average depth of 54 m) [*Danabasoglu and Gent*, 2009; *Gent et al*., 2011; *Kiehl et al*., 2006].

For the baseline simulation, greenhouse gases and aerosol concentrations were held fixed at year 2000 values. After a 10 year spin-up, stratospheric sulphate aerosol optical depth (AOD) and/or concentrations of a globally well-mixed, hypothetical greenhouse gas were perturbed along prescribed pathways.

For volcanic aerosols, the temporal and regional responses estimated from the Pinatubo eruption were extracted from the inventory in *Ammann et al*. [2003]. For the 3X Pinatubo case, volcanic AOD was scaled by 3 throughout the eruption and subsequent response. See Figure S1 in the supporting information for the spatial and temporal evolution of volcanic aerosols.

For the hypothetical SLGHG case, we used a gas with the same radiative properties of CFC-11, as this gas was already implemented in the CESM radiation code. However, unlike CFC-11, it was neither long lived nor was it allowed to affect stratospheric ozone. Prescribed concentrations were [35.0, 12.9, 4.7, and 0.2] ppb for years 2–5 after the eruption, with linear interpolation such that concentrations gradually increase to 35 ppb during the first year. Thus, we assumed a rapid implementation of balancing emissions within the first year after the eruption and removal according to the adopted lifetime. In the Discussion section, we will discuss a more specific candidate for such a counteracting SLGHG.

The chosen concentration pathway corresponds to an *e*-folding lifetime of the SLGHG of approximately 1 year and was designed to balance the effects of the 3X Pinatubo simulation in CESM. Other *e*-folding times were tested. For the volcanic pathway used here, we found lifetimes of greater than 1.5 years to be impractical for designing a balancing GHG pathway

The 3X Pinatubo simulation reaches a peak top-of-atmosphere (TOA) shortwave radiative forcing from the direct aerosol effect in excess of $-15\,\mathrm{W\,m^{-2}}$ (Figure S2 in the supporting information). For a 12 month running mean, the peak is around $-12\,\mathrm{W\,m^{-2}}$. The total (shortwave + longwave) TOA flux perturbation, for a 12 month running mean, is $-5\,\mathrm{W\,m^{-2}}$. The requirement for the SLGHG pathway is to balance this forcing.

CFC-11 has a radiative efficiency of about $0.25\,\mathrm{W\,m^{-2}\,ppb^{-1}}$ at sub-ppb concentrations [*Hodnebrog et al*., 2013], but because of the gradual saturation of the absorption bands at higher concentrations, more CFC-11 (35 ppbv) is required to generate the required $5\,\mathrm{W\,m^{-2}}$ peak forcing (Figure S2 in the supporting information) than might be anticipated from this radiative efficiency.

For the studies of regional and seasonal effects, we ran three independent ensemble members for the 3X Pinatubo and balanced cases. Perturbations were applied at different times in a long, stabilized baseline run; i.e., the starting points for both atmosphere and slab oceans differ between ensembles. Twelve additional ensemble members were constructed by comparing the perturbed cases to different, temporally independent periods of the baseline simulations, to yield a 15 member ensemble. No baseline drift was

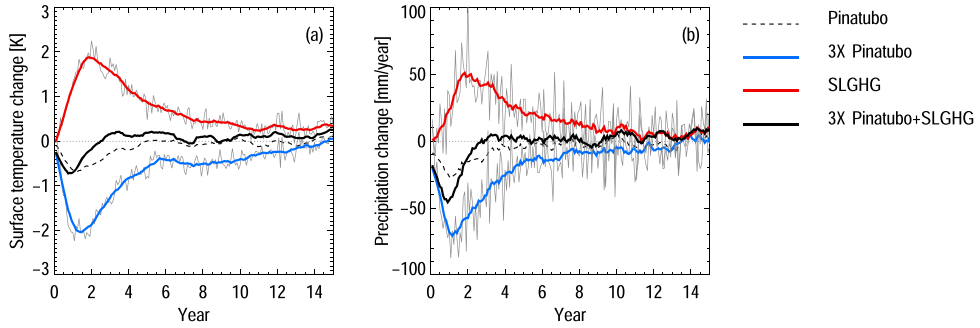

Figure 1. Climate response to volcanic and short-lived GHG (SLGHG) perturbations. (a) Simulated global mean surface temperature response to Pinatubo, 3X Pinatubo, SLGHG emissions, and a simultaneous 3X Pinatubo and SLGHG perturbation. (b) Global mean change in precipitation. For the 3X Pinatubo and 3X Pinatubo + SLGHG cases, the grey lines indicate month-to-month variability.

observed, in either temperature or precipitation, for the time periods used. As a measure of statistically significant deviation between the signal and baseline simulations, we performed Student's *t* test calculations, requiring $p < 0.05$.

The climate impact of an eruption depends on its latitude. Eruptions at high latitudes mostly affect the hemisphere where they occur since the *e*-folding lifetime of the stratospheric aerosol is less than a year. However, low-latitude volcanic eruptions generally lead to stronger, more global cooling effects since the sulphur aerosols reach higher altitudes where they have a longer lifetime allowing it to spread over larger areas [*Kirtman et al.*, 2013]. Here we assume the same location as Mount Pinatubo (which is located at 15°08′30″N, 120°21′00″E).

3. Results

A 3X Pinatubo eruption on its own would cause substantial climatic disruption over a period exceeding 5 years (Figure 1). After 1 year, global mean temperatures drop by 2 K, and global mean precipitation drops by 70 mm yr^{-1}. This is followed by a gradual recovery. The emission of a SLGHG, sustaining a concentration path designed to approximately match the volcanic forcing, counteracts the main global mean temperature and, partially, precipitation effects of the eruption. Residual modeled effects are similar in magnitude to those found from a Pinatubo-sized event.

Note that the global mean model response to a volcanic eruption depends on the climate sensitivity of the model and when a slab ocean is used, the assumed depth of that slab. Our modeled temperature response to a Pinatubo-sized eruption is stronger than some recent studies [e.g., *Gao et al.*, 2008; *Harris and Highwood*, 2011], and weaker than others [e.g., *Hansen et al.*, 1992; *Soden et al.*, 2002].

The annual mean regional response is more complex (Figures 2a and 2b). To reduce the effects of natural, interannual variability on our analysis, ensembles were run and averaged as described above. Hatched areas indicate statistical significance from the baseline at $p < 0.05$. (For individual ensemble temperature and precipitation responses in the balanced case, see Figures S3 and S4 in the supporting information).

For 3X Pinatubo, cooling is most pronounced at northern latitudes, exceeding 4 K over the Arctic Ocean and North America, on average for the period 2–6 years after the eruption. In year 2 (not shown), we find a peak annual mean cooling of 8 K over Alaska. For the balanced simulation, the mean temperature change (of both signs) is limited to less than 1 K in all regions.

Precipitation has a larger natural variability than temperature, and we do not expect to find significant regional changes for a 3X Pinatubo perturbation (although we note that any lack of statistical significance in changes does not necessarily indicate that there is a lack of impact of those changes). Figures 2c and 2d show the annual mean regional response in precipitation. In the unbalanced case, we find the largest reductions around the equator. In the balanced case, this pattern remains but is dampened. Figures 2e and 2f show that absolute zonal mean precipitation changes are strongest just north of the equator; however, this is also the region with largest intrinsic natural variability. A reduction in precipitation is seen throughout

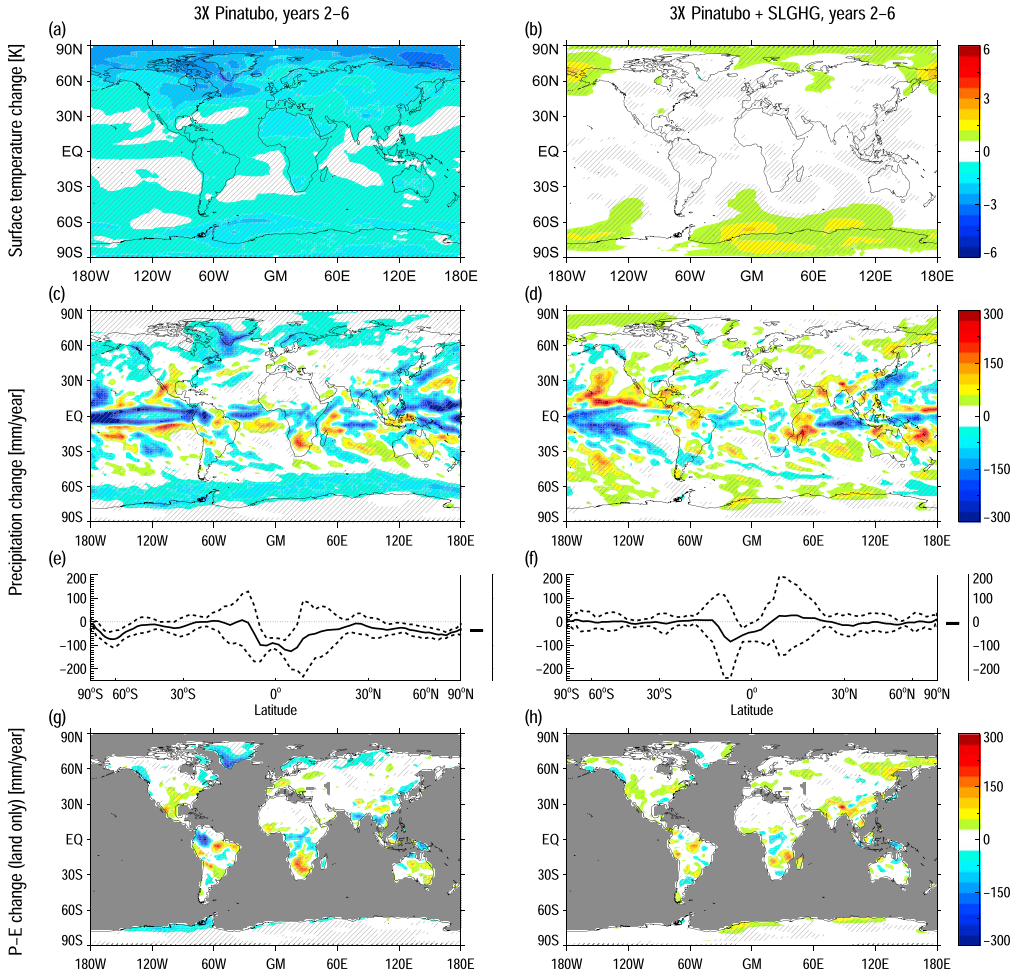

Figure 2. Annual mean regional surface temperature response (K) to (a) a 3X Pinatubo eruption and (b) 3X Pinatubo balanced by short-lived GHG (SLGHG) emissions, averaged over years 2–6 after the eruption and across 15 ensemble members. Hatching indicates statistically significant deviations from the baseline at $p < 0.05$. (c and d) Regional and (e and f) zonal mean precipitation changes for 3X Pinatubo and 3X Pinatubo balanced by SLGHG emissions. In Figures 2e and 2f, the dashed lines show the 2σ envelope for the 15 ensemble members. The lines to the right of the plots indicate the global mean precipitation change (variability across ensemble members is within the line width). (g and h) Precipitation minus evaporation with ocean regions masked to highlight changes over land.

the Northern Hemisphere (NH), but it may not be discernible above natural variations. In the balanced simulation, the ensemble mean is consistent with no changes. Since the surface water balance is important for impacts of climate change, we have also calculated the change in difference between precipitation and evaporation over land (Figures 2g and 2h). This picture is more complex, and the changes are statistically significant over relatively few areas. Nevertheless, in some regions, the changes are smaller in the balanced case (e.g., South America, central Africa, and Greenland), while in other regions, they become more pronounced (e.g., northeastern Asia). Figures 2g and 2h have the ocean masked out to emphasize differences over land. For an unmasked plot, see Figure S5 in the supporting information.

Figure 1b shows a significant global mean precipitation response in the first year, even in the balanced case. There are a number of possible reasons for this. The forcing paths of our aerosols and SLGHGs, while similar, are not identical. There is a strong residual negative forcing in the first year of the balanced simulations (see Figure S2 in the supporting information). The volcanic aerosols are rapidly emitted from the eruption, while the GHG level builds up more slowly over the first year. A better fit to the negative volcanic forcing in the model might have reduced the precipitation response in the balanced case, but in any case, in practice achieving a balanced forcing in the early months following an eruption might be impossible. In the model,

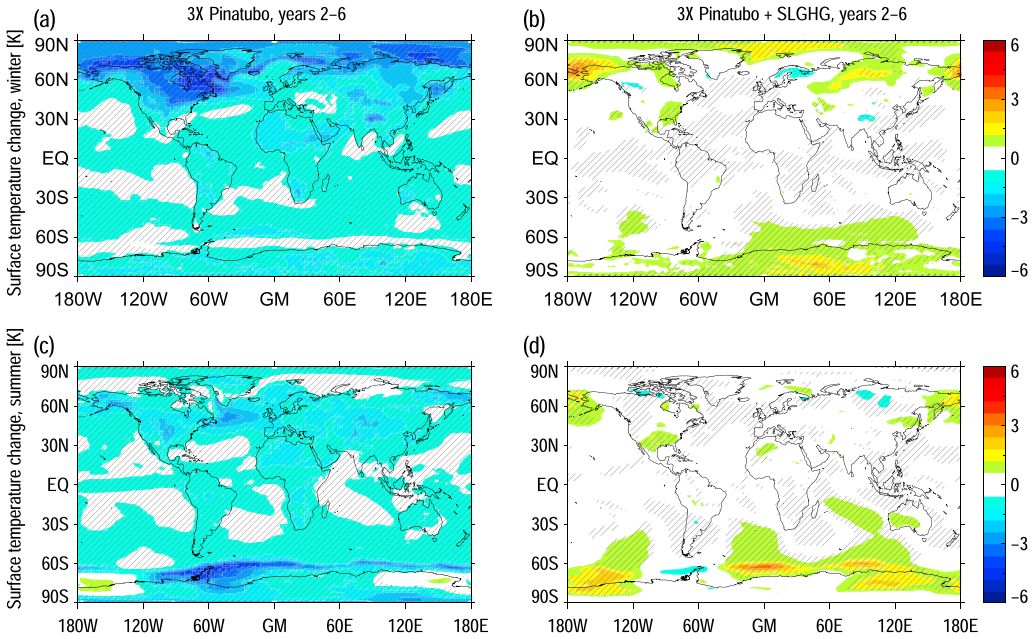

Figure 3. Seasonal response of surface temperature (K) to (a and c) a 3X Pinatubo eruption and (b and d) 3X Pinatubo balanced by SLGHG emissions, averaged over years 2–6 after the eruption and across 15 ensemble members. Hatching indicates statistically significant deviations from the baseline at $p < 0.05$. (top row) NH winter response (December-January-February). (bottom row) NH summer response (June-July-August).

we have also assumed instantaneous global mixing of the GHG, while the volcanic aerosols are first spread around the equator then gradually transported poleward (see Figure S1 in the supporting information).

Another factor is the different response of precipitation to GHG and aerosol forcing. The total precipitation response is made up of a fast response that is related to the nature of the radiative forcing, in particular the surface-atmosphere partitioning of this forcing, and a slow response that is more related to surface temperature changes [*Andrews et al*., 2010]. For example, *Ferraro et al*. [2014] discuss the modeled response to a (geoengineered) stratospheric aerosol perturbation, the impact of this aerosol on heating rates in the upper troposphere, and the subsequent effect of this on precipitation. This component of the tropical precipitation response is found to be nearly instantaneous when the forcing appears. The fast precipitation response to SLGHGs, due to their direct influence on atmospheric heating rates, will differ to the fast response to the volcanic aerosol, which mainly originates in the stratosphere. Hence, it is difficult to balance the initial precipitation change, since the aerosol and SLGHGs influence atmospheric heating in different ways. Balancing becomes more straightforward once the slow (surface temperature) response becomes a more important driver of the precipitation change.

A further concern is the *seasonal* temperature response to the volcanic and SLGHG forcing (Figure 3). Again, we use the mean of three model ensemble members. For 3X Pinatubo, the cooling is strongly seasonal, with the NH experiencing winter mean surface temperature reductions in excess of 6 K over large areas for years 2–6 after the eruption. In the NH summer, we still see a cooling, but the largest effects are to be found in the Antarctic. For the balanced simulation, we find residual seasonality in the temperature response. We find a pronounced winter cooling over northern Europe, balanced by an Arctic warming. In NH summer, the Antarctic exhibits regional changes between +4 K and −4 K. Such regional changes will depend on the details of the atmospheric and ocean circulation during and after the eruption. The results would also vary between climate models.

4. Discussion

Our simulations indicate that SLGHG emissions could in principle be used to counteract parts of the climate impact of a large eruption. However, there are numerous practical, financial, scientific, philosophical, and ethical issues.

The chosen GHG needs to have a short atmospheric lifetime to allow for dynamic concentration adjustments as knowledge of the volcanic forcing evolves. However, the lifetime should be long enough to give independence of emission location (in the case of one emission site). Our simulations indicate that for an event with a peak top-of-atmosphere flux perturbation (longwave + shortwave) of $-5\,\mathrm{W\,m^{-2}}$ (see Figure S2 in the supporting information), a GHG lifetime of >1.5 years makes it difficult to design an effective counterforcing path. Instead of only one emission site globally, there could be one or several emission points in each hemisphere, which would allow better control of development of atmospheric levels of the countergas.

We have focused on comparing years 2–6 in our study of cancellation. However, in years 0–2, there is a significant response in precipitation even in the case with counteremissions (Figure 1b, black line, see discussion above). Even with better fit to the natural forcing path, the world would probably still experience perturbed conditions during the first couple of years. It would be more serious if these conditions lasted over longer periods (up to a decade). Our perspective in this work has been to study whether it would be possible to cancel effects on this time scale.

Whether counteracting measures should be implemented at a given time should be seen in relation to ongoing background natural variations. Good understanding of likely natural variations on a short and medium time scale, such as El Niño–Southern Oscillation variability and other periodic changes to oceanic circulation and heat uptake, would obviously strengthen risk assessments performed in the event of a major volcanic eruption.

It is useful to consider possible candidates for the SLGHG used to counteract the eruption and how much of those gases might be needed. We first note that anthropogenic CO_2 is not suitable because of the very long response time of the CO_2 perturbation [*Joos et al*., 2013; *Solomon et al*., 2009] and acidification of the oceans [*Rhein et al*., 2013].

One possibility is a hydrofluorocarbon, hydrofluorocarbon (HFC)-152a, which has a lifetime of 1.5 years (similar to the lifetime of the volcanic aerosol) and has an ozone depletion potential of zero as it contains no ozone-depleting atoms. HFC-152a has a radiative efficiency of $0.098\,\mathrm{Wm^{-2}\,ppb^{-1}}$ [*Hodnebrog et al*., 2013] at sub-ppb concentrations, but this is muted by about 30% (on the basis of detailed narrow band radiation code calculations) as a result of the saturation of the absorption bands when concentrations are on the order of 50–100 ppb. Hence, if HFC-152a emissions were used to construct a radiative forcing path similar to the one indicated in Figure S2 in the supporting information (peaking at $5\,\mathrm{W\,m^{-2}}$ at the end of the first year), it would require emissions of about 1.25 Gt during that year; atmospheric concentrations would then reach a peak of about 78 ppb at the end of the first year. Note that these concentrations are greater than the amount of CFC-11 used in the global climate model (GCM) experiments, as HFC-152a has a significantly lower radiative efficiency than CFC-11.

The requirement for the availability of gigatonnes of such gas presents formidable obstacles. To place this amount in context, current emissions of the most abundant HFC in the atmosphere, HFC-134a, are on the order of $150\,\mathrm{kt\,yr^{-1}}$ [*IPCC*, 2013]; hence, it would require a massive expansion in global production capacity to generate the required quantities of a gas such as HFC-152a, even assuming that sufficient raw materials are available. It would appear unfeasible to produce such quantities of gas after the occurrence of an eruption. Rather, it would have to be produced in advance and stored. Assuming that about 7 t of CO_2 equivalents are emitted for the production of 1 t of HFC [*McCulloch and Lindley*, 2003], the production would constitute around a quarter of the annual present-day emissions of fossil fuel CO_2 [*Ciais et al*., 2013], assuming fossil fuel energy sources. Finally, the present-day price of HFCs is on the order of a few thousand dollars per tonne; the production of 1 Gt constitutes a cost on the order of a trillion dollars, comparable to earlier estimates of costs of common geoengineering options, or mitigation of anthropogenic warming [*Royal Society*, 2009]. These costs have to be balanced against the potential damage resulting from the climate effects of the eruption, although the production and storage costs would be incurred irrespective of whether an eruption actually occurs. To our knowledge, assessments and quantification of cost to the society of the climatic effect of volcanic eruptions are not available but would be necessary in the assessment of possible counteracting regimes in response to volcanic eruptions. An alternative, or supplementary, approach to counteracting climate impacts of volcanic eruptions could be changing agriculture to become less sensitive to cold climates, e.g., by developing cold-resistant crops [*Engvild*, 2003]. Thus, counteracting the climate effects might not necessarily mean counteracting the climate change itself.

HFCs in present-day production have been selected to meet specific uses and partially on the basis of having a relatively low climate effect. A gas used for counteracting would not have such limitations. The choice could be designed more around minimizing the cost of counteracting, and the gas need not necessarily be a HFC. Furthermore, there may be advantages in using a suite of different GHGs as this avoids, to some extent, the radiative saturation mentioned in section 2. The same forcing could then potentially be achieved for lower concentrations, lower emissions, and lower costs.

It would be important to minimize harmful compounds that are produced from production or atmospheric destruction of the chosen gas. If very large quantities of the countergas were needed, the oxidation capacity of the atmosphere might be suppressed. Reduced levels of the main tropospheric oxidant, the OH radical, would reduce the removal rate of methane [e.g., *Isaksen et al*., 2009]. This indirect effect would prolong the warming effect of the countergas and potentially be a problematic side effect. The eruption itself would, however, also disturb OH levels (and thereby CH_4) via several mechanisms [*Banda et al*., 2013; *Robock*, 2000; *von Glasow*, 2010], most importantly via changes in the UV fluxes that control OH formation. Further possible indirect effects of volcanic sulphate emissions on tropospheric clouds are not accounted for here [*Schmidt et al*., 2012].

In traditional geoengineering, which has been suggested to counter man-made warming caused by long-lived GHGs, a key concern is the termination problem; i.e., the expected rapid increase in temperature should there be a termination of geoengineering activities [*Royal Society*, 2009]. Counteracting volcanic cooling through SLGHGs would not have this problem, since the undesired forcing itself has a natural termination.

In assessing strategies for counterbalancing a volcanic eruption, a further question would be whether the aim is to balance global or regional effects (see, e.g., [*Ban-Weiss and Caldeira*, 2010] although that study discusses balancing a warming). Even if global mean temperature and precipitation changes are balanced, there are several other factors that will likely not be, for example, changes in photosynthetic activity and net primary productivity [*Haywood et al*., 2013]. Furthermore, the use of HFCs as countergases might lead to some amplification of stratospheric warming caused by the volcanic sulphate cloud [*Forster and Joshi*, 2005] (Figure S6 in the supporting information), with possible impacts on the larger-scale circulation.

Thus, there are many difficult outstanding issues reaching beyond atmospheric sciences. In addition to questions related to costs and risks, a counteremission would create challenges for international collaboration and governance. Even if global mean temperatures of precipitation changes are balanced, the unbalanced effects discussed above would lead to a redistribution of risks between different biophysical parts of the climate system as well as between regions and nations. Due to the absence of the termination problem, the redistribution of risks between generations would be less than for traditional geoengineering as long as a short-lived GHG is used. While traditional geoengineering is considered highly contentious due to ethical aspects and technological and scientific uncertainties [*Royal Society*, 2009], attitudes to counteracting natural cooling may be different.

Whether counteracting a future large volcanic eruption with GHG emissions should be seriously considered is dependent on the size of the eruption and the likely consequences of action versus inaction. The Pinatubo eruption did not warrant a counterbalancing. While several studies have assessed the geophysical effects of larger volcanic events [*Brovkin et al*., 2010; *Foley et al*., 2014; *Harris and Highwood*, 2011], their impacts on today's society have not been well explored [*Self*, 2006]. The ability of the present generations of climate models to accurately reproduce observed effects after recent eruptions is also still under evaluation [*Canty et al*., 2013; *Driscoll et al*., 2012]. Any future implementation of counteremissions would need to build on solid knowledge, including more detailed and better validated Earth system models and specially designed emission strategies. This requires extensive development and studies by several modeling groups and observational studies of active volcanoes. It is important to emphasize that the risk of future eruptions cannot be used to justify the past and current GHG emissions.

Volcanic eruptions will sooner or later significantly perturb the climate. In preparation for this, we need to consider whether deeper multidisciplinary studies are needed to build knowledge to aid decision making and to consider what size of eruption might trigger consideration of counteremissions. Even if scientific and ethical issues could be agreed upon, the financial costs and practical difficulties in producing and storing sufficient GHGs may be deemed prohibitive and exclude further consideration. In this case, alternative strategies for mitigating and adapting to such eruptions might be needed.

Acknowledgments

All simulations and other data used for the figures and analysis in the present paper are available upon request to the corresponding author. The authors would like to thank several colleagues for their work and constructive feedback. Part of the work was supported by the Research Council of Norway, through the SLAC project. Simulations were performed on the Abel supercomputer at the University of Oslo.

The Editor thanks two anonymous reviewers for their assistance in evaluating this paper.

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