Influence of volcanic eruptions on tropical hydroclimate during the last millennium

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INFLUENCE OF VOLCANIC ERUPTIONS ON TROPICAL HYDROCLIMATE
DURING THE LAST MILLENNIUM

by

Christopher M. Colose

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Abstract

Volcanic eruptions exert the most important radiative forcing on Earth’s climate during the pre-industrial interval of the last millennium. In this thesis, I investigate the role of volcanic eruptions in altering tropical climate, including temperature and rainfall. I primarily use forced transient simulations of the last millennium as a tool to explore how explosive volcanic events project onto the hydrologic cycle, as well as the imprint of water isotopologues (H\textsubscript{2}\textsuperscript{18}O, H\textsubscript{2}\textsuperscript{16}O) associated with rainfall. Attention is given to the South American continent specifically (in chapter 2), and to the entire tropics (in chapter 3).

In Chapter 2, I show that volcanic eruptions cool the South American continent and alter rainfall, decreasing the intensity of the austral summer monsoon, and decreasing rainfall in the northern part of the continent during austral winter. These factors also conspire to influence the isotopic signal left behind, informing the detectability of volcanic excursions in the paleoclimate record and the anticipated hydroclimate response at the continental scale for future eruptions.

The results of chapter 2 emerge from a simple composite response to many of the largest volcanic eruptions during the last millennium and instrumental period. In chapter 3, I advocate for a more targeted approach in how volcanic eruptions are stratified when interpreting physical responses or comparing to past records; in particular, I highlight the role of the spatial structure in volcanic forcing in altering the mean intertropical convergence zone (ITCZ) position, and the associated response of different monsoon systems. The main finding in chapter 3 is that the ITCZ moves away from the preferentially forced hemisphere, which leads to unique ENSO behavior, river discharge anomalies, or patterns in isotopic anomalies, depending on the location of the eruption. In this chapter, I also make contact with recent advances in understanding ITCZ
migrations through the lens of the atmospheric energy budget. I discuss the significance of these findings for interpreting the paleoclimate record.

In chapter 4, I expand upon chapter 3 by quantifying individual feedbacks (including water vapor and clouds) that arise in response to different spatial structures of volcanic forcing. I demonstrate that cloud and water vapor distributions differ dramatically for aerosol loadings that are northern hemisphere focused, southern hemisphere focused, or fairly symmetric about the equator. Such feedback differences may amplify or dampen ITCZ movements or complicate inferences of how feedbacks are expected to behave in a warming world.
Acknowledgments

First and foremost, I offer deepest thanks to my adviser, Mathias Vuille. During my 5+ years at Albany, I have had tremendous freedom to carve out my own research path and pursue interests only loosely related to the grant topic that funded the large duration of my stay. Throughout my meandering journey, especially early on, he gently nudged me in the right direction by providing certain tasks and in helping to clarify the scope of my own investigation. He is also a meticulous editor and greatly improved the quality and readability of much of the content in this manuscript.

A special thanks to Allegra LeGrande at NASA’s Goddard Institute for Space Studies. Allegra effectively acted as a “second adviser” and was patient during the early portions of my PhD career, when I was still working out the kinks and fundamentals of doing basic research. Allegra was a host during two summer visits at NASA GISS, and has provided invaluable support.

I thank the remainder of my committee: Brian Rose, Aiguo Dai, and Oliver Elison Timm, for reading and commenting on my thesis material, and for developing questions in my qualification exam. All of my committee members have consistently kept “open doors” to questions and the exchange of ideas. I especially thank Brian Rose, who compels those he advises (formally or informally) to be a good philosopher of science, in addition to a good scientist. I’ve sincerely enjoyed my inspiring discussions on topics far removed from thesis content with him, and in sharing the thrill of thinking deeply about simple things.
I thank the entire support staff in the Department of Atmospheric & Environmental Science, which keep all the little things running smooth and without worry on the part of the graduate students. I thank my officemates and friends; during my later years, I’ve gotten to know Ted Letcher and Cameron Rencurrel well, and I’ve enjoyed my company with them during the good times (e.g., Graduate Climate Conference) and the bad (e.g., the 2016 election). They have each provided some humor during the otherwise dull sessions of coding.

My path into climate science came from an unusual direction. I would be remiss to not mention inspiration that came from contributors at http://www.realclimate.org/, a blog run by climate scientists, and a source that helped mold my own interest in the climate system. I especially thank Gavin Schmidt, who took the time to answer naïve questions I had as an undergraduate, and who offered his support upon entering graduate school. I am excited to come to NASA GISS, where he is now director, for at least a couple years to do a postdoctoral fellowship. I also want to mention the writings of Raymond Pierrehumbert, which I consider to be the most influential in shaping the way I think about the big picture of how a planet’s climate operates.

I’m grateful to my parents, Matt and Carol, for their love and guidance growing up, and continuous support even when they didn’t understand the subtle reasons behind taking another half decade out of life to complete one’s doctorate’s degree. Surely, they no longer want to hear the phrase, “I’m almost done,” and I am glad I can put that to rest. Finally, my girlfriend Melissa was of tremendous support during my time. Dating someone with the unparalleled amount of free time and endless financial resources of a
graduate student would simply be too overwhelming for many, but I am happy she was on my side during the challenge. Thank you.
Statement of Publication/Contribution of Authors

I, Christopher Colose, was the lead researcher for all material presented in this manuscript. Dr. Mathias Vuille at the University at Albany, and Dr. Allegra LeGrande at NASA’s Goddard Institute for Space Studies, contributed to Chapters 2 and 3 through discussions of methodology and results. Both acted in a supervisory capacity through the duration of my PhD track, and provided comments on all chapters. Dr. LeGrande also provided data used in Chapters 2 and 3.

Chapters 2 and 3 of this dissertation include the following published material, with only minor changes to ensure continuity in figure/section numbering throughout the thesis. Inclusion of these chapters is necessary to provide a coherent and appropriately sequenced investigation of the subject-matter.

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Results are shown for global (0°N-90°N minus 0°S-90°S), tropical (0°N-30°N minus 0°S-30°S), and extratropical (30°N-90°N minus 30°S-90°S) domains. All datapoints represent individual eruptions after averaging over the five ensemble members.
Chapter 1

Introduction

1.1. Opening

“During several of the Summer Months of the Year 1783, when the Effect of the Suns Rays to heat the Earth in these northern Regions should have been greatest, there existed a constant Fog over all Europe. This Fog was of a permanent Nature; it was dry, and the Rays of the Sun seem’d to have little Effect towards dissipating it, as they easily do a moist Fog arising from Water. They were indeed rendered so faint in passing thro’ it, that when collected in the Focus of a Burning Glass they would scarce kindle brown Paper; Of course their Summer Effect in heating the Earth was exceedingly diminished. Hence the Surface was early frozen...

Hence the first Snows remained on it unmelted, and received continual Additions.

The Cause of this Universal Fog is not yet ascertained. Whether it was adventitious to this Earth, and merely a Smoke proceeding from the Consumption by Fire of some of those great burning Balls or Globe which we happen to meet with in our rapid Course round the Sun, and which are sometimes seen to kindle and be destroy’d in passing our Atmosphere, and whose Smoke might be attracted and retain’d by our Earth: Or whether it was the vast Quantity of Smoke, long continuing to issue during the Summer from Hecla in Iceland, and that other Volcano which arose out of the Sea near that Island; which Smoke might be spread by various Winds over the northern Part of the World; is yet uncertain.”

-Benjamin Franklin

The above quote, written over 200 years ago by Benjamin Franklin (Franklin, 1784) was perhaps the first attempt to link volcanism and atmospheric phenomena, in particular when a strange “dry fog” (a sulfate haze resulting from tropospheric oxidation of volcanogenic sulfur gases) and unseasonably cold weather struck Europe. Franklin also suggests in the quote that a meteorite may have been responsible. However, there was indeed a major eruption in Iceland (albeit Laki, not Hekla, which is ~75 km away) in 1783-1784, one of the largest volcanic eruptions in recorded historical times. The Laki eruption produced ~14.7 km³ of basaltic lava
(Thordarson and Self, 1992), a volume sufficient to cover New York City (~789 km\textsuperscript{2} in area) with over 18 meters in basalt. In Iceland, the haze lead to the loss of most of the island's livestock (from eating fluorine contaminated grasses), crop and vegetation failure (due to acid rain), and because of famine and disease, the death of ~20% of the country's human residents (Thordarson and Self, 2003).

Volcanic eruptions are very important for society, and have strongly impacted past cities and civilizations (e.g., an eruption at Santorini in Greece at ~1620 B.C. produced a thick layer of pumice and ash that buried Bronze Age settlements and permanently altered the local topography, see Friedrich et al., 2006; or the 79 A.D. Vesuvius eruption that deposited large quantities of ash onto the buildings of the Roman city, Pompeii). In 1883, the eruption of Krakatoa in Indonesia destroyed hundreds of villages and killed ~33,000 people. In 1902, the Pelée eruption in the Caribbean killed ~29,000 people (e.g., Luong et al., 2003).

Historical paintings reflect changes in the color of the sky following large eruptions due to the optical effects of the volcanic cloud, and in fact the ambitious task of using red-to-green ratios in historical artwork to reconstruct paleo- aerosol optical depths has been undertaken (Zerefos et al., 2014). The optical effects caused by Tambora for several years following 1815 have been widely reported, including sunspots seen from the naked eye, dimming of the moon and stars in a clear-sky atmosphere, and prolonged sunsets and twilights observed near London, features that have been used to calculate visible optical depths (Stothers, 1984).

Volcanic eruptions also impact air travel, as with the Icelandic Eyjafjallajökull eruption in 2010. Since the residence time of ash can be as long as weeks, the ash does not have a strong climate impact but is able to paralyze air traffic far away from the source. Eyjafjallajökull caused
the cancellation of 108,000 flights, particularly in Europe, disrupted the travel plans of 10.5 million passengers, and cost the airline industry in excess of $1.7 billion (Budd et al., 2011).

Many eruptions of the sort described above are therefore of interest as a geologic hazard and to historians and archeologists. However, in many cases, the aerosol cloud is primarily restricted to the troposphere and the volcanic material (see next section) falls out on timescales of days to weeks, leaving minimal long-term impact extending beyond the recovery timescale of that which was affected by the eruption.

Explosive stratovolcanic events provide an extra dimension to the impact of volcanic eruption. The 1815 Tambora eruption (on the Indonesian island of Sumbawa), for instance, immediately killed tens of thousands of people, but the impacts of the volcano were felt worldwide due to significant global cooling (Oppenheimer, 2003; Raible et al., 2016), resulting in major changes to European and North American weather, as well as crop failures. The eruption allegedly inspired Mary Shelley’s “Frankenstein” while North America experienced the so-called “Year without a Summer” (anecdotally, summer frosts were pervasive in the northeast United States and snowfall was recorded in June in Albany, NY; Baron, 1992). Limited instrumental measurements suggest Northern Hemisphere (NH) temperatures dropped by ~1°C in 1816 (Stothers, 1984).

1.2. Volcanoes and Climate—A primer
This remainder of this thesis focuses on the climate impact of volcanic eruptions, especially on tropical hydroclimate and during the last millennium (LM). I also stress the relatively short-to-intermediate impact (~seasons to years) following volcanic eruptions, which approximately corresponds to the residence time of sulfate aerosols that form in the stratosphere. Indeed, it is now recognized that the metric most closely related to the volcanic projection onto climate is the sulfur dioxide (SO$_2$) emissions from volcanoes, not necessary the explosivity, volume of erupted magma, or ash injection (Pollack et al., 1976; Rampino and Self, 1984).

Volcanoes also emit Hydrogen Sulfide (H$_2$S) that is rapidly converted to SO$_2$, along with Halogens, CO$_2$, and water vapor. The oxidation of SO$_2$ by OH and subsequent reactions yield sulfuric acid vapor (H$_2$SO$_4$) that condenses onto particles to form sulfate aerosols, typically at sizes similar to a visible wavelength and where scattering of solar radiation is strongest.

It would be possible for eruptions to warm the surface if the sulfate aerosols were very small or large (<0.05μm or >2.2μm; see e.g., Coakley and Grams, 1976; Lacis et al., 1992). In the former case, scattering is very small; for the large particle limit, the thermal component begins to increase with particle size while scattering asymptotes to a constant value as particle size becomes larger than the contributing solar wavelengths (Lacis, 2015). Thus, for large particles, the shortwave and longwave contributions will be of comparable magnitude. The longwave component also depends on the height of the aerosol, since the greenhouse effect depends upon the temperature difference between the surface and emission layer. However, particles are only very small in their formative stages, and if very large, tend to fallout quickly, leaving intermediate size particles (where solar scattering dominates) as the expected players to

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1 This will refer to either the 850-1850 or 850-2005 C.E. interval, corresponding to the pre-industrial LM and historical simulation period for the experiments in CMIP5/PMIP3. The study in chapter 2 includes the historical extension period, but chapter 3 does not.
the perturbation in the stratospheric aerosol layer. Thus, large sulfur-rich volcanic excursions will almost always be expected to cool the terrestrial climate.

Sulfate aerosols typically heat the stratosphere via absorption of longwave radiation, but in principle could cool the stratosphere if injected in the upper stratosphere, because of the higher local temperatures and increased efficiency for cooling to space (Lacis, 2015) relative to the capacity for absorbing upwelling radiation from lower layers. Stratospheric temperatures are observed to increase in the lower and mid stratosphere following the El Chichón (April 1982) and Mt. Pinatubo (June 1991) eruptions (Randel et al., 2016). For large tropical eruptions, the stratosphere warming results in anomalous temperature gradients aloft between the equator and poles, and an enhancement of the polar vortex. This has been implicated in leading to warming over sectors of the northern mid-latitudes during boreal winter (e.g., Robock and Mao, 1992; Kirchner et al., 1999; Shindell et al., 2004; Stenchikov et al., 2004, 2006) via dynamical responses in the jet stream that overcome the direct radiative effects of shortwave scattering in the winter high-latitudes. This highlights a difference between tropical and high latitude eruptions. However, fewer studies have targeted the implications of hemispheric asymmetry in the aerosol loading until very recently, which I visit in chapter 3.

The fragmented magmatic material forming the ash of the eruption is comprised of larger (typically >10μm to mm) and settles out quickly, leaving behind the aforementioned sulfur gases that form the enduring aerosol layer of concentrated sulfuric acid. It is worth noting that explosive eruptions may have a relatively small sulfur injection. For example, Mt. St. Helens in 1980 was very explosive but did not put much sulfur into the stratosphere, thus resulting in minimal climate impact (Robock, 1981). Similarly, sulfur-rich eruptions that are not sufficiently explosive to protrude into the stratosphere will have little climate impact. In fact, annual
anthropogenic SO₂ emissions are larger than background eruption fluxes or even the input from a Pinatubo-sized eruption. However, the actual sulfur burden from volcanoes and human activity is comparable (Graf et al., 1997), due to the fact that the sulfur is often injected at higher elevations with volcanoes, even for background tropospheric eruptions. However, volcanoes dominate the variability in stratospheric loading where the residence time is much longer. Thus, the residence time for volcanic SO₂ is higher than for anthropogenic SO₂. Water vapor released in the eruption also affects atmospheric chemistry and the rate of sulfate formation (LeGrande et al., 2015), although a proper treatment of chemistry in models is in its infancy and not the way volcanic forcing is implemented in the Coupled Model Intercomparison Project Phase 5 (CMIP5)/Paleoclimate Model Intercomparison Project Phase 3 (PMIP3) generation of GCM’s.

CMIP5/PMIP3 is the most recent iteration of coordinated model experiments with multiple contributing groups (Braconnot et al., 2012; Taylor et al., 2012); the simulations covering the period 850-1850 C.E. as part of the past1000 initiative (with some groups contributing historical extensions to these runs) are a subset of the contributing target paleo time intervals (along with the Last Glacial Maximum and mid-Holocene). Notably, the past1000 runs are transient simulations using the same model versions as for future projections, and include sporadic volcanic activity in the input forcing, allowing for a greater sample of events to be probed than is possible using just the instrumental period.

The most recent climatically meaningful volcanic excursion was that of Mt. Pinatubo in 1991, which briefly interrupted the ongoing long-term global warming trend (Figure 1.1) and helped produce global cooling by up to ~0.5 °C despite the concurrent El Niño at the time (the effects of ENSO are linearly removed in Figure 1.1, which amplifies the Pinatubo influence). Such cooling also emerges in a composite sense when considering the largest eruptions during
the instrumental period (Hansen et al., 1996). Figure 1.2 shows the evolution of the greenhouse effect strength during the LM period as simulated in the CESM Last Millennium Ensemble (see chapter 3 for details). The longwave optical depth, although small, increases the opacity of the atmosphere immediately following large eruptions. However, the decrease in atmospheric temperature (due to scattering) leads to a reduction in atmospheric water vapor, the principle positive feedback that acts to amplify global cooling. Chapter 4 revisits this problem and sets the stage for future work in quantifying the diverse fast feedback responses to different volcanic eruptions.

Pinatubo injected about 20 teragrams \((1 \text{Tg} = 10^{12} \text{g})\) of \(\text{SO}_2\) into the stratosphere and elevated the stratospheric sulfate burden by a factor of \(~60\) (McCormick et al., 1995; Kremser et al., 2016), which is a relatively small to intermediate eruption in the context of the last millennium, but is the best observed that is climatically relevant. There is still notable uncertainty in the absolute size of volcanic eruptions, even during the instrumental era. For instance, the Stratospheric Aerosol and Gas Experiment II (SAGE II) satellite instrument, which operated between 1984 and 2005, was saturated following Mt. Pinatubo, leading to uncertainties in the early-stage aerosol vertical distribution and optical depth (Arfeuille et al., 2013).

The current standard for volcanic reconstructions during the 850-1850 C.E. period is based on networks of sulfate peaks in ice cores (in both the Arctic and Antarctica). Volcanic sulfate fluxes derived from ice cores are scaled directly to Aerosol Optical Depth (AOD, Crowley and Unterman, 2013, hereafter CU13) or to stratospheric sulfur injections and, combined with atmospheric modelling of aerosol distribution, atmospheric mass loading (Gao et al., 2008, hereafter G08). G08 provides separate injection and loading datasets; the link between surface sulfate deposition and injection is based on different scaling factors for tropical and extratropical
eruptions, based on nuclear bomb tests, observations for Mt. Pinatubo, and modeling studies. The sulfate mass time-series (as a function of latitude, height, and time) is then derived from a simple parameterized stratospheric aerosol transport model, and an interpolated vertical distribution of volcanic aerosol based on information from lidar measurements following Pinatubo. This simple model does not allow for cross-equatorial aerosol transport, so the horizontal spread for tropical eruptions is artificially based on multiple injections in two hemispheres (according to the injection estimates appropriate for the different polar cores). In CU13, AOD (at 550 nm) is scaled to ice core sulfate fluxes linearly for Pinatubo-and-smaller eruptions, based on satellite-derived AOD and the flux of sulfate to Antarctica at the time of the eruption. An aerosol effective radius dataset (important for the radiative properties) is provided based on an empirical function of AOD.

Since transfer functions linking the sulfate fluxes to loading, or to the effective radius of the aerosol particles, rely heavily on observations of the Mt. Pinatubo event, this essentially makes all volcanic events in the last millennium a scaled version to Pinatubo (LeGrande and Anchukaitis, 2015).

The implementation of volcanic forcing in the CMIP5/PMIP3 past1000 runs is somewhat crude. The pre-industrial volcanic history in the past1000 runs were based on either the G08 or CU13 reconstructions, of which groups could choose either (or both in the rare case where multiple simulations were performed, as the NASA GISS group did) and implementation of several radiative properties (including particle size) were left to the individual model groups.

Important differences in the datasets exist, which complicates proxy-model or model-model comparisons. For example, Icelandic volcanoes are problematic since much of the aerosol delivery may be tropospheric and still leave a large imprint on NH high latitude ice cores, and
different procedures for screening eruptions thought to be globally important may be employed. The Laki eruption has very little imprint in the CU13 dataset, but is larger in G08. Furthermore, the CU13 dataset scales down the AOD for larger-than-Pinatubo eruptions, due to the larger particle size thought to be associated with large eruptions (Timmreck et al., 2010) and the associated decrease in shortwave scattering. Thus, the implied forcing for large eruptions in G08 is larger than in CU13. Different model implementations of a given dataset further exasperate this problem. For instance, in the CSIRO-MK3L-1-2 (Australian model) contribution to the past1000 intercomparison, the aerosol forcing implied by CU13 is simply converted to an equivalent reduction in total solar irradiance at the top-of-atmosphere, and information concerning the meridional structure of the forcing is lost.

For the next iteration of modeling experiments in CMIP6/PMIP4, a new dataset that provides an improved history of the timing and magnitude of eruptions over the last 2500 years is available (Sigl et al., 2013, 2015). The results in each chapter, however, are composited according to the model forcing, so the timing of volcanic eruptions is not relevant.

Of notable relevance to this thesis (chapter 3) is the spatial structure of the volcanic forcing. Usually, signatures of simultaneous sulfate signature in both polar regions is used to infer an eruption of tropical origin, since the stratospheric circulation allows for global transport for a low latitude eruption, whereas a high latitude aerosol cloud remains confined to the hemisphere of injection. Information regarding the source of sulfur release from past eruptions can also be obtained from historical records or from the chemical fingerprinting of tephra (fragmental material produced by a volcanic eruption). The acidity peaks and tephra also act as chronostratigraphic markers that are used to correlate different ice cores and reduce age uncertainty (e.g., Coulter et al., 2012; Sigl et al., 2013).
Because of the more spatially restricted aerosol loading, it is therefore usually assumed that mid-to-high latitude eruptions have a much more local climate impact than tropical eruptions. However, chapter 3 illustrates the far-field impact of an asymmetric eruption by way of adjustments in the tropical circulation to the introduction of even small temperature gradients aloft, which are difficult to maintain dynamically in the tropics (Sobel et al., 2001).

Information concerning whether past eruption plumes reached the stratosphere can be obtained from sulfur isotope analysis (Savarino et al., 2003; Baroni et al., 2007). Usually, the geochemical processes that separate isotopes from each other depend on differences in isotopic mass. For mass-dependent fractionation of sulfur isotopes, sulfur isotope ratios (δ\textsuperscript{32}S, δ\textsuperscript{33}S, δ\textsuperscript{34}S, δ\textsuperscript{35}S) measured in a reaction product would be approximately proportional to the relative difference in mass between the lighter and the heavier isotope (e.g., δ\textsuperscript{33}S ≈ 0.515 δ\textsuperscript{34}S, see notation below). In oxygen, the \textsuperscript{17}O/\textsuperscript{16}O ratio often changes by approximately half as much as the \textsuperscript{18}O/\textsuperscript{16}O ratio, since the mass difference between \textsuperscript{17}O and \textsuperscript{16}O is roughly half as large as the mass difference between \textsuperscript{18}O and \textsuperscript{16}O.

Deviations from this mass-dependent relation are referred to as mass-independent fractionation (MIF). Since the gas-phase oxidation of SO\textsubscript{2} into sulfuric acid by OH is mass-dependent, sulfate aerosol particles formed in the troposphere leave no MIF signals. However, measurements of the sulfur isotope composition have shown that sulfate formed in the stratosphere from the oxidation of volcanic SO\textsubscript{2} possesses a significant signature of MIF, while sulfate formed in the troposphere does not. This fact relies on reactions that occur in the presence of UV radiation, such as the photolysis of either SO\textsubscript{2} or SO\textsubscript{3}, the isotopic signature of which could then be transferred to the SO\textsubscript{4} aerosol product and used as a proxy for aerosols that were injected above the ozone layer. However, such information has not yet been implemented.
coherently into the height distribution of paleo-eruptions in GCM’s. Furthermore, there is debate on the robustness of this technique for high latitude eruptions such as Laki where the tropopause is relatively low, and so volaties could be injected into stratospheric altitudes and have a climate impact but still be low enough where MIF of sulfur isotopes is not an important photochemical process (Lanciki et al., 2012; A. Schmidt et al., 2012).

1.3. Water Isotopes

In chapters 2 and 3, I make use of the isotope-enabled NASA GISS ModelE2-R, which features interactive water isotopes (strictly, ‘isotopologues’ when referring to a molecule rather than an atom), including H$_2^{18}$O, H$_2^{16}$O, and HDO (where D is deuterium, or $^2$H). These are termed stable isotopes, since they do not undergo radioactive decay like $^3$H (tritium) or radioactive carbon. Isotopic analysis has formed the backbone of quantitative paleoclimatology; for example, the principle ocean proxy for deriving information about past ocean conditions is the oxygen-18 ratio measured in carbonate found in benthic or planktonic foraminifera (and corals), and also the isotopic content that derived from precipitation in ice cores and speleothems (cave deposits).

Ocean water contains approximately two $^{18}$O atoms for every thousand $^{16}$O atoms. Note that the ratios of the heavy isotopic water H$_2^{18}$O/H$_2^{16}$O or HDO/H$_2^{16}$O are typically expressed in parts per thousand (‰) deviations from some agreed upon standard (often the Vienna standard mean ocean water, or V-SMOW), which is done to help eliminate laboratory bias introduced during sample analysis. The delta notation is defined such that:
\[ \delta = \frac{R_{sa} - R_{st}}{R_{st}} \times 1000 \]  

(1)

where \( R_{sa} \) and \( R_{st} \) are the isotope ratios of the sample and standard, respectively. Thus, if a sample with \(^{18}\text{O}/^{16}\text{O}\) is \(2.000 \times 10^{-3}\) (Araguás-Araguás et al., 2010), relative to a value of \(2.005 \times 10^{-3}\) for V-SMOW, the sample will be said to be relatively depleted in the heavy isotope, with \(\delta^{18}\text{O} = -2.49 \text{%}\).

The recognition that the stable isotopic composition of precipitation and important climate parameters were linked, and the establishment of empirical relationships between the observed isotopic composition of meteoric water and environmental parameters gained ground in the 1950s and early 1960s (Dansgaard, 1953; 1964). Relationships between the isotopic composition of precipitate and surface air temperature, distance from the coast, height above sea level, and rainfall amount (traditionally described by many “effects”, including the temperature, continental, orographic, and amount effects, etc, see e.g., Galewsky et al., 2016 for a review.) spurred an enormous body of research, including the identification of large-scale isotope-temperature slopes (both in the spatial and temporal dimension) used, for instance, in glacial-interglacial reconstructions. Dansgaard explained successfully how and why in many regions of the Earth the isotopic composition of precipitation is linearly related to the local temperature at the precipitation site and why there is a breakdown of this effect in deep convective (tropical) environments, where the total rainfall is a better predictor of isotopic composition.

Evaporation and condensation are examples of mass-dependent processes, which influence how different isotopes of a substance change phase. Water molecules in liquid with heavy oxygen (or hydrogen) isotopes have greater binding energies and evaporate less readily than the light isotopologue. Similarly, the saturation vapor pressure for \(\text{H}_2^{18}\text{O}\) is lower than for
H$_2^{16}$O, hence the concentration of heavy isotopes in the liquid phase is higher than in the vapor phase equilibrated with this liquid. Indeed, a simple two-phase condensation model is perhaps the most basic starting point (Galewsky et al., 2016), particularly if we portray the global hydrologic cycle beginning in the tropics/subtropics, with transport and condensation in higher latitudes—in such a model, vapor with a prescribed temperature-dependent isotopic composition condenses under the assumption of an open (‘Rayleigh’) system, in which condensate is immediately removed after formation (in dynamic and isotopic equilibrium) with no further isotopic exchange with vapor. As cooling of some air mass proceeds, water vapor condenses and new portions of the precipitation are isotopically enriched in heavy isotopes relative to the remaining vapor (and resulting in the vapor being more depleted) and with further cooling the vapor and condensate become isotopically lighter due to the rainout of heavy isotopes as more condensation events occur. Similarly, intense rain in convective systems removes heavy isotopes from the air, much as it removes soluble gases like SO$_2$.

In general, however, the isotopic signal recorded at a site will integrate multiple climate processes, including mixing, evaporation, rainfall, temperature, history of saturation, etc. from source to site. Different atmospheric circulation states and different pathways of moisture masses to their precipitation site may be important during the Last Glacial Maximum, for example. This fact has compelled a parallel branch to emerge to inform the observational realm: modeling isotopes in GCM’s, which itself has a rich history (e.g., Sturm et al., 2010; Werner, 2010).

Isotope modeling is computationally expensive, due to the need to duplicate portions of the model code to simulate multiple water isotopologues, and to advect and mix isotopic tracers from different air masses while computing appropriate fractionation factors at phase change to allow a modification of the isotopic composition in different reservoirs. For this and legacy
reasons related to model development and prioritization, isotope simulations were not common across participating groups in PMIP3. A notable exception is the NASA GISS group; Chapters 2 and 3 make use of the NASA GISS isotope-equipped model (ModelE2-R) to map the isotopic signal following explosive volcanoes. This is a novel aspect to this thesis, as virtually no literature existed during the writing that attempted to simulate the isotopic signal following volcanic eruptions, which is important for paleoclimate attempts to understand the climate response following volcanic excursions (e.g., Ridley et al., 2015, albeit with carbon isotopes) during the LM. In chapter 3, for example, I highlight that isotopic signals following a large-scale ITCZ shift may be entirely different for a NH volcano like Laki, when compared to a SH volcano like Kuwae (in the 1450’s).

1.4. Last Millennium Climate

The pre-industrial interval of the LM (prior to the onset of the industrial era around 1850 C.E.) is a time interval only weakly affected by anthropogenic forcing, and provides an opportunity to test hypotheses due to its relatively long duration and because climate conditions are similar to the present day. Moreover, the LM contains the majority of Earth’s annually resolved proxy records (see Jones et al., 2009 for a review), and is thus valuable in model-proxy assessments of high-frequency forced or internal variability. Importantly, among the currently available collection of simulations, the CMIP5/PMIP3 projects have, for the first time, produced multiple LM (and other selected paleo-intervals), historical, and future climate simulations using the same model configurations and resolutions (Taylor et al. 2012).
Nonetheless, the LM is challenging when considering climate change or any climatic expression of forced signals because we typically encounter relatively low signal-to-noise ratios (compared to climate changes associated with doubled CO₂ or glacial-interglacial transitions, for example), thus demanding more complicated efforts in separating the forced response from internal variability (e.g., Schurer et al., 2013). This is especially true in discussions of regional and continental-scale phenomena such as North American mega-drought (e.g., Ault et al., 2013; Coats et al., 2015a, 2015b) although large volcanic eruptions are clearly discriminated by a cooling effect at a hemispheric/global scale.

The most prominent features in the secular evolution of LM hemispheric-scale climate are the so-called “Medieval Climate Anomaly” (MCA) — or in earlier literature often identified as the Medieval Warm Epoch (Lamb, 1965) or Medieval Warm Period — followed by the “Little Ice Age” (LIA), occurring from ~950-1250 C.E. and ~1400-1700 C.E., respectively (Mann et al., 2009). Historical reconstructions indicate that NH (where data is more abundant) surface air temperatures exhibited variability on decadal-to-centennial time scales superimposed on a secularly decreasing trend (Mann et al., 2009; PAGES2k, 2013), a feature broadly reproduced in model simulations (e.g., Otto-Bliesner et al., 2016, see Figure 1.3).

However, despite the expression of these periods in hemispheric-scale reconstructions (e.g., Mann et al., 2009) there is considerable spatio-temporal structure in the pattern of warming/cooling (PAGES2k, 2013). This limits the usefulness of these terms that have remained intact for legacy reasons but can be problematic for discussing particular climate events/phenomena. Indeed, it has become apparent that these periods were not characterized by uniformly warmer or cooler temperatures, but rather by a range of temperature, hydroclimate,
and marine changes with distinct temporal and regional expressions. During both intervals, distinct hydroclimatic anomalies have been identified on regional and global scales.

Our knowledge of LM climate comes predominately from different proxies such as tree-ring widths/densities, coral records, historical written evidence, ice cores, boreholes, cave deposits (speleothems) and lake sediments (Jones et al., 2009, PAGES2k, 2013). The other tool that can be employed for diagnosing mechanisms driving LM climate are GCMs, which must be driven by the best reconstructions of external forcing such as solar irradiance, greenhouse gases, orbital changes, mass of volcanic aerosol (or some measure of this impact, such as stratospheric AOD and land-use changes (Schmidt et al., 2011; 2012). These reconstructions often have considerable uncertainty, and this uncertainty-space has generally not been well sampled (the NASA GISS group have explored the most combinations of available forcing reconstructions in the CMIP5 generation of LM simulations, while the NCAR CESM Last Millennium Ensemble features many ensemble members with single or all forcings in order to probe the influence of individual forcings as well as internal variability, but not the forcing uncertainty). Forcing uncertainty is independent of model or proxy errors, but nonetheless complicates model-proxy comparisons and assessments.

1.5. Motivating Problems in Tropical Hydroclimate

Volcanic eruptions are the most important external radiative forcing at the global scale during the pre-industrial part of the LM (Atwood et al., 2016), and exert their impact on the late 20th century temperature evolution. Both observational and modeling studies have consistently found a decrease in global temperature and precipitation following large explosive eruptions
(Robock, 2000), and the main regions experiencing decreased precipitation tend to be the tropics (Robock and Liu, 1994; Trenberth and Dai, 2007; Schneider et al., 2009) and monsoon regions (Schneider et al., 2009; Joseph and Zeng, 2011), although the South American response in particular has so far been poorly probed (done in chapter 2).

Chapter 2 provides one of the first assessments of the tropical South American Summer Monsoon (SASM) response to volcanic forcing using the NASA GISS ModelE2-R. Variability of precipitation and isotope ratios in tropical South America is especially linked to Intertropical Convergence Zone (ITCZ) variability and precipitation, the El Niño–Southern Oscillation (ENSO), and SAMS intensity (e.g., Vuille et al., 2012) making it an especially relevant target for this thesis. Tropical South America is a region where hundreds of millions of people live and where the hydroclimate seasonality is a distinctive feature characterizing the continental climate. Unfortunately, the few large eruptions during the instrumental period offer little insight into the volcanic imprint over the region, since they typically occur quasi-simultaneously with El Niño events (as with Mt. Pinatubo and El Chichón), which dominate the South American response. Furthermore, the isotope observations are especially limited during these excursions.

Chapter 3 provides a larger tropical perspective. In the tropics, the large-scale circulation is dominated by Hadley cells; the tropics are characterized by an uneven distribution of diabatic heating/cooling (convective latent heating in convergence zones) and these forcing terms are nearly in balance with vertical motion (e.g., Rodwell and Hoskins 1996), since horizontal temperature gradients are small enough that horizontal motions bringing air in from somewhere else cannot help balance local radiative cooling, unlike in higher latitudes. In the tropics, convection is concentrated in a relative narrow area (e.g., the ITCZ) and more gradual subsidence elsewhere. The ITCZ is thus part of the rising branch of the Hadley cell and
characterized by a maximum in climatological precipitation (Figure 1.4). Understanding ITCZ dynamics in response to volcanic forcing is the focus of chapter 3 and 4.

In the last few years, an energetic framework has emerged in the literature linking the ITCZ to far-field forcings that are asymmetric about the equator (see chapter 3 and 4 for review). This framework begins with the recognition that the precipitation in the ITCZ is approximately co-located with the latitude where energy fluxes diverge, i.e., Hadley cells transport energy in the direction of their upper tropospheric flow and this flow is directed away from the ascending branch. The anomalous Hadley circulation in response to an asymmetric forcing transports energy toward the energetically deficient hemisphere, while the moisture (in the lower part of the circulation) is directed toward the relatively warmer hemisphere and toward the newly established latitude where meridional energy fluxes vanish (Bischoff and Schneider et al., 2014; see Figure 1.5). Chapter 3 provides a first attempt to link this energetic paradigm with volcanic forcing in the LM paleoclimate simulations, while highlighting an under-appreciated aspect of how high latitude volcanic forcing can project onto tropical hydroclimate signals, including rainfall and streamflow.

Global precipitation, and rainfall in wet tropical and monsoon regions decreases in several instrumental datasets and in CMIP5 models when compositing over several 20th century eruptions (Iles and Hegerl, 2014). Iles et al. (2013) examined the global precipitation response to large low-latitude volcanic eruptions using an ensemble of last millennium simulations from HadCM3. In the tropics, areas experiencing post-eruption drying coincide well with climatologically wet regions, while dry regions get wetter on average, but their changes are spatially heterogeneous. This pattern is of opposite sign to, but physically consistent with, projections under global warming.
However, a large body of literature also suggests that CMIP5 generation models do not do well in capturing dynamical responses to volcanic eruptions (e.g., Driscoll et al., 2012), including the previously mentioned warming in parts of the Northern Hemisphere during the winter following a large tropical volcanic eruption, or that the CMIP5/PMIP3 past1000 ensemble does not robustly exhibit drying over Meso-America following clusters of major volcanic eruptions that were detected in a recent speleothem record (Winter et al., 2015). Anchukaitis et al. (2010) argued that several climate models do not correctly simulate the post-volcanic hydroclimate anomalies over Asia following large eruptions.

A large motivation for chapter 3 was to provide a more targeted model analysis. This includes the use of many ensemble members (provided by the NCAR community for the LM following the IPCC AR5), which differ in initial conditions and constitute a much larger capacity to probe initial condition uncertainty than is possible with just a single realization. Surprisingly, the role of internal variability has been under-appreciated in model evaluation of volcanic response (Bittner et al., 2016; McGraw et al., 2016), especially when considering regional hydroclimate changes, or when comparing to observations (of which the instrumental period effectively constitutes just one ensemble member in a perfect model). Compositing over many eruptions helps reduce the noise for one ensemble, but may still be insufficient to gauge the diversity of possible responses to a given eruption. Furthermore, in some studies, eruptions are only composited based on a threshold forcing rather than also stratifying by spatial structure. As shown in chapter 3, compositing over different eruption classifications lead to very different monsoon responses, including in Asia. These results thus enforce the need for a careful consideration of internal variability and the method of how climatically important eruptions are sorted in a robust proxy-model comparison.
The problem of how volcanic asymmetries project onto tropical or global climate has been poorly studied in part because there have only been three climatically significant eruptions during the second half of the 20th century. The largest and most recent of these (Mt. Pinatubo) was not very asymmetric in forcing structure while the first (Mt. Agung in Indonesia) occurs before the satellite era. El Chichón (Chiapas, Mexico) and Mt. Pinatubo also occurred quasi-simultaneously with strong El Niño events, thus limiting the ability to confidently test hypotheses connecting volcanic forcing to large-scale tropical rainfall (Lehner et al., 2016).

Another outstanding problem has been the role of volcanic forcing in triggering ENSO. Some studies find no relation between ENSO and volcanic eruptions (Self et al., 1997; Ding et al., 2014), others find a tendency toward La Niña (McGregor and Timmermann, 2011), and still others suggest eruptions favor El Niño-like anomalies (Emile-Geay et al., 2008; Maher et al., 2015), and often due to different mechanisms. There has been promising idealized work at this front recently (Pausata et al., 2015) that explicitly links changes in the tropical Pacific to NH eruptions through a tapering off of the trade winds along the equator in the central Pacific (via an ITCZ shift), thereby favoring the development of El Niño in response to high latitude NH eruptions via the Bjerknes feedback. The El Niño response to NH eruptions was also demonstrated recently in the CESM Last Millennium Ensemble (Stevenson et al., 2016).

Chapter 3 provides a parallel study along these lines, noting that for SH volcanoes or hemispherically symmetric forcing (which Stevenson et al. call ‘tropical’ volcanoes, but I note that tropical volcanic eruptions can generate a hemispherically asymmetric aerosol cloud, as Mt. Agung and El Chichón did), the ENSO response is more equivocal, but often leads to a cooling of the tropical Pacific.
1.6. Asymmetries in Climate

Since chapters 3 and 4 focus on the meridional structure of the volcanic forcing, I briefly discuss one of the persistent characteristics of Earth’s recent climate history, which is the asymmetry between the NH and SH. The prevailing boundary conditions that result in such an asymmetry encompass a wide range of timescales, from the seasonal cycle to millions of years when e.g., sea ice extent, ocean circulation, orography, presence of continental ice masses, etc., are largely determined by continental drift or variations in the long-term carbon cycle. Several examples serve to highlight the evolution of an asymmetric climate:

1) The climate history during the Cenozoic (~ 65 millions years ago to the present) was highlighted by the glaciation of Antarctica ~36 million years ago (Ehrmann and Mackensen, 1992). While some ice existed in the NH during this interval, Greenland was not fully glaciated until at least 3 million years ago (e.g., Bierman et al., 2014).

2) During the last deglaciation (and probably previous glacial-interglacial transitions), there was substantial millennial-scale climate variability associated with changes in the strength of the global thermohaline circulation (e.g., Alley, 2007). In particular, there are numerous instances when deep-water formation in the northern Atlantic weakened, allowing fresh water to pool on the ocean surface and more expansive sea ice developed in the high latitudes of the NH (e.g., Isarin et al., 1998; Bradley and England, 2008 in the context of the Younger Dryas), and a reduction in the Atlantic meridional overturning circulation that led to an imposed source of asymmetry between the NH and SH (see e.g., Wang et al., 2006 for speleothem based evidence of this during the last glacial period). McGee et al. (2014) further review evidence for ITCZ movement during the Last Glacial Maximum, Heinrich Stadial 1, and mid-Holocene.
3) The modern climate is asymmetric. Nearly 70% of Earth’s landmass is in the NH, while the Southern Ocean is the only region on Earth to be land-free around an entire latitude belt. Presently, the SH receives slightly more annually averaged top-of-atmosphere (TOA) net radiation than the NH (Frierson et al., 2013; Stephens et al., 2015). Much of this slight imbalance seems to occur from the OLR asymmetry, since planetary albedo is impressively symmetric about the equator (Stephens et al., 2015) although this symmetry emerges from different contributions from the surface and atmosphere. In the NH, the Sahara desert is a net sink of radiation, owing to its high shortwave reflectance and strong emission of outgoing longwave radiation (OLR) to space, which appears to be a large contribution to the hemispheric asymmetry.

However, despite the net radiant energy input into the SH being somewhat greater than the NH, there is a much larger northward transport of energy by the ocean across the equator (on the order ~0.4 Petawatts, where 1 PW = 10^{15} W) that allows the annual-mean NH average temperature to be warmer than that of the SH (Feulner et al., 2013; Frierson et al., 2013; Kang et al., 2014; Marshall et al., 2014).

The present hemispheric asymmetry leads to interesting and potentially counter-intuitive responses when the planet is subjected to a radiative perturbation. For example, the Earth is currently relatively far from the Sun during boreal summer, yet Earth’s global-mean temperature is actually highest during July and exhibits an annual range of over 3°C, comparable to the annual mean increase associated with a doubling of CO₂. This range arises despite the out of-phase insolation signal due to land and ocean differences between hemispheres. Actually, during the writing of this thesis, the global-mean temperature anomaly in July 2016 surpassed all previous anomalies for that month by a rather large margin, suggesting this month was likely the
hottest month in absolute terms during the same interval (although August 2016 is statistically close).

4) Further identified or potential sources of hemispheric asymmetry include the response to greenhouse gases, Milankovitch variations (on orbital timescales), aerosols, or afforestation/deforestation (e.g., Chiang and Friedman, 2012; Friedman et al., 2013). The transient response to future CO$_2$ increase for example, results in the NH warming faster than the SH, due to the long timescale of the Southern Ocean to come into equilibrium with the new climate.

Stratospheric aerosols that originate from explosive volcanic activity are a potential source of hemispheric asymmetry. Although the scholarship has now reached a clear consensus that the ITCZ migrates away from the energetically deficient hemisphere, uncertainties remain in the regional expression of temperature and precipitation change (e.g., in regions that straddle the poleward extent of ITCZ migration, or the amount of rainfall in monsoon climates) and how these scale with the aerosol abundance (or the aerosol gradient). Furthermore, there is virtually no understanding as to whether actual paleo-volcanic activity engendered such a response that is potentially detectable isotopically. Chapter 3 aims to address all of these issues.

Finally, in chapter 4, I build off the results in chapter 3 with results calculating the radiative feedbacks that promote or dampen the hemispheric radiation asymmetry in response to explosive volcanism. For instance, do cloud shifts related to the ITCZ shift itself (or those in the extratropics) act to enhance the net energy input in one hemisphere relative to the other, thereby potentially amplifying or diminishing the aerosol forcing itself? It is now well appreciated that the meridional structure of forcing is critical in determining the realized climate response (Rose
et al., 2014), and this potentially extends to feedbacks on the ITCZ movement. Chapter 4 further explores how the distribution of water vapor and clouds change in response to different volcano classifications, which affect the atmospheric energy budget and the meridional energy fluxes.

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**Figure 1.1:** Time-series of global-mean near surface (or SST) temperature anomalies (°C) relative to 1880-1909. Data is from the Land-Ocean Temperature Index provided by NASA GISS, from 1880-November 2016. Monthly and annual raw data shown by grey dots and red line, respectively. ENSO-removed curve from 1951-present (annual-mean, blue line) is based on lagged linear regression at each grid point. Temperature was allowed to lag between the Niño 3 index from 0-6 months at each point, with the lag decided by the highest correlation between the ENSO state and local detrended temperature with
two years after major volcanic eruptions removed in the regression. Vertical dashed line prior to the Mt. Pinatubo eruption.
**Figure 1.2:** Time-series of the global and annual mean greenhouse effect strength (W/m²) as simulated in the CESM Last Millennium Ensemble (all forcing simulations, see chapter 3 for details) from 850-2005 C.E., calculated as the difference between upwelling surface longwave radiation and the outgoing longwave radiation (OLR). Dark grey shows the ensemble spread (10 members) and the red line is the ensemble mean. Global and annual mean precipitable water (calculated from a mass-weighted vertical integral of specific humidity, in kg/m²) shown in aqua.
Figure 1.3: NH, tropical (30°S-30°N), and global-mean temperature anomalies for three different periods to broadly represent the Little Ice Age (legend in bottom panel), relative to the Medieval Climate Period (950-1200 C.E.) in the CESM Last Millennium Ensemble all and individual forcing experiments. Number of ensembles used for each forcing experiment shown above the forcing name.
Figure 1.4. Annual-mean precipitation (fill, mm/day) from GPCPv2.2 from 1981-2010. Wind (near-surface) vectors (length based on magnitude, in m/s) from NCAR/NCEP reanalysis over the same period.
Figure 1.5. Schematic of meridional energy transport (PW) by the atmosphere. The black curve is taken from a control simulation in a symmetric aquaplane (CESM 1.2.1). The red curve is an idealized representation of how this transport could change for a warm perturbation in the Southern Hemisphere.
Chapter 2

The influence of volcanic eruptions on the climate of tropical South America during the last millennium in an isotope-enabled GCM

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Abstract

Currently, little is known on how volcanic eruptions impact large-scale climate phenomena such as South American paleo-ITCZ position and summer monsoon behavior. In this paper, an analysis of observations and model simulations is employed to assess the influence of large volcanic eruptions on the climate of tropical South America. This problem is first considered for historically recent volcanic episodes for which more observations are available, but where fewer events exist and the confounding effects of ENSO lead to inconclusive interpretation of the impact of volcanic eruptions at the continental scale. Therefore, we also examine a greater number of reconstructed volcanic events for the period 850 C.E. to present that are incorporated into the NASA GISS ModelE2-R simulation of the Last Millennium.

An advantage of this model is its ability to explicitly track water isotopologues throughout the hydrologic cycle and simulating the isotopic imprint following a large eruption. This effectively removes a degree of uncertainty associated with error-prone conversion of isotopic signals into climate variables, and allows for a direct comparison between GISS simulations and paleoclimate proxy records.

Our analysis reveals that both precipitation and oxygen isotope variability respond with a distinct seasonal and spatial structure across tropical South America following an eruption. During austral winter, the heavy oxygen isotope in precipitation is enriched, likely due to reduced moisture convergence in the ITCZ domain and reduced rainfall over northern South America. During austral summer, however, more negative values of the precipitation isotopic composition are simulated over Amazonia, despite reductions in rainfall, suggesting that the
isotopic response is not a simple function of the ‘amount effect.’ During the South American monsoon season, the amplitude of the temperature response to volcanic forcing is larger than the rather weak and spatially less coherent precipitation signal, complicating the isotopic response to changes in the hydrologic cycle.
2.1 Introduction

2.1.1 Volcanic Forcing on Climate

Plinian (large, explosive) volcanic eruptions are a dominant driver of naturally forced climate variability during the Last Millennium (LM, taken here to be 850 C.E. to present; e.g., Stothers and Rampino, 1983; Hansen et al., 1992; Crowley et al., 2000; Robock et al., 2000; Robock, 2003; Goosse et al., 2005; Yoshimori et al., 2005; Emile-Geay et al., 2008; Cole-Dai, 2010; Timmreck, 2012; Iles et al., 2013; Schurer et al., 2014). In addition to their importance for 20th century climate, they are the largest magnitude external forcing during last 1000 years of the pre-industrial period, the most recent key interval identified by the Paleoclimate Modelling Intercomparison Project Phase III (PMIP3). As such, these eruptions serve as a natural testbed to assess the skill of climate models in simulating how climate responds to external perturbations. Although the most significant climate impacts of eruptions are realized over just a few years following the eruption, they provide the source of the largest amplitude perturbations to Earth’s energy budget during the LM. For example, the eruption of Mt. Pinatubo in June 1991, although transitory, exerted a radiative forcing comparable to an instantaneous halving of atmospheric CO$_2$ [Hansen et al., 1992; Minnis et al., 1993; see also Driscoll et al. (2012) for models in the Coupled Model Intercomparison Project Phase 5 (CMIP5)]; several paleo-eruptions during the LM likely had an even larger global impact (Figure 2.1).

The principle climate impact from volcanic eruptions results from the liberation of sub-surface sulfur-containing gases such as sulfur dioxide, which are injected into the stratosphere and react with water to form sulfate aerosols (e.g., Harshvardhan and Cess, 1976; Coakley and
Grams, 1976; Pollack et al., 1976, 1981; Lacis et al., 1992). The most pronounced impact of large tropical eruptions includes a radiatively cooled troposphere and heated stratosphere (e.g., Lacis et al., 1992; Robock and Mao, 1995; Stenchikov et al., 1998). Sulfate aerosols from the Mt. Pinatubo eruption grew from a background effective radius of ~0.2 μm up to ~0.8 μm, strongly scattering incoming solar radiation. For sulfate aerosols in this size range, this shortwave scattering is 5-10x larger than the increase in infrared opacity from the aerosols, and results in a warming stratosphere and cooling of Earth’s surface (Turco et al., 1982; Lacis et al., 1992).

Studies on the impacts of volcanic eruptions have generally focused on global or Northern Hemisphere metrics (e.g., Lucht et al., 2002; Gillett et al., 2004; Shindell et al., 2004; Oman et al., 2005; Oman et al., 2006; Anchukaitis et al., 2010; Peng et al., 2010; Evan et al., 2012; Zhang et al., 2013; Man et al., 2014; Stoffel et al., 2015), for instance in examining responses to the East Asian monsoon or the Arctic Oscillation (e.g., Ortega et al., 2015). Comparatively little attention has been given to the Southern Hemisphere, or to South America specifically (although see Joseph and Zeng, 2011, and Wilmes et al., 2012). Some previous work has focused on the Southern Annular Mode in the ERA-40 and NCEP/NCAR reanalysis, in addition to a previous version of NASA Goddard Institute for Space Studies (GISS) Model-E (Robock et al., 2007) and in a subset of CMIP3 models (Karpechko et al., 2010) or in CMIP5 (Gillett and Fyfe, 2013).

How volcanic forcing is expressed over South America remains an important target question for several reasons. First, recognition of the South American monsoon system (SAMS) as an actual monsoon system is less than two decades old (Zhou and Lau, 1998), and thus study of SAMS dynamics is still relatively young (section 2.1.3) and very little work has been done
specifically focused on volcanic eruptions. For instance, should we expect to see a reduction in austral summer rainfall? Secondly, the largest volcanic eruptions during the late 20th century (e.g., Mt. Agung, 1963, Indonesia; El Chichón, 1982, Mexico; Mt. Pinatubo, 1991, Island of Luzon in the Philippines- hereafter, these three events are referred to as L20 eruptions) occur quasi-simultaneously with an anomalous El Niño-Southern Oscillation (ENSO) state, and in general represent a small sample size in a noisy system. This limits the prospect of robust hypothesis testing and guidance for what impacts ought to be expected following large eruptions at the continental scale. Finally, South America offers promise for a comparatively dense network of high-resolution proxy locations relative to other tropical regions (see below), offering the potential to detect whether South American hydroclimate signals to large eruptions are borne out paleoclimatically.

In this study, we explore the post-volcanic response of South American climate operating through the vehicle of unique model simulations (spanning the LM) using the recently developed GISS ModelE2-R (LeGrande et al., 2016, in prep; Schmidt et al., 2014a), which allows for the sampling of a greater number of events than is possible over the instrumental period. Emphasis is placed on temperature and precipitation, but a novel part of this study extends to the response of water isotopologues (e.g., H$_2^{18}$O) [colloquially referred to hereafter as ‘isotopes’ and expressed as $\delta^{18}$O in units per mil (‰) vs. Vienna Standard Mean Ocean Water]. The isotopic composition of precipitation ($\delta^{18}$O$_p$) is a key variable that is directly derived from proxy data used in tropical paleoclimate reconstructions.

The aim of this paper is to create a potentially falsifiable prediction for the isotopic imprint that a volcanic eruption should tend to produce across the South American continent. The ability to explicitly model the isotopic response allows for a less ambiguous comparison of
simulations and paleoclimate records and for hypothesis testing. It is unclear whether or not the current proxy archives are suitable to test such a prediction with high confidence, given dating uncertainties (in both proxies and in the actual timing of eruptions), or the level of noise in proxy data and the real world. Additionally, the prevailing high-resolution archives in South America only feature a few tropical records (Vimeux et al., 2009; Neukom and Gergis, 2012; Vuille et al., 2012). Nonetheless, the growing number of high-resolution records offers hope that testing the modeled response to high-frequency volcanic signals will be an avenue for future research. This can also better inform debate centered on the inverse problem in interpreting isotopic signals (i.e., what do observed changes in proxy data imply about past climate changes?), which remains contentious (section 2.1.4).

The structure of this article is as follows: in the remaining part of section 2.1, we summarize previous literature on the impact of large volcanic eruptions on paleoclimate, in addition to a discussion of South American climate. Section 2.2 presents data and methodology, including how volcanic forcing is implemented in ModelE2-R. Section 2.3 discusses our results and we end with conclusions in section 2.4.

2.1.2. Volcanic forcing during the Last Millennium

Volcanic forcing has had a very large influence on the climate of the LM (Crowley, 2000; Hegerl et al., 2003; Shindell et al., 2004; Mann et al., 2005; Hegerl et al., 2006; Fischer et al., 2007; D’Arrigo et al., 2009; Timmreck, 2012; Esper et al., 2013; Ludlow et al., 2013; Schurer et al., 2014). Several studies (Miller et al., 2012; Schurer et al., 2014; Atwood et al., 2016; McGregor et al., 2015) collectively provide a compelling case that volcanic forcing may be substantially more important than solar forcing on a hemispheric-to-global scale during the LM,
in addition to driving a large portion of the inter-annual to multi-decadal variability in LM simulations (Schmidt et al., 2014b).

Two volcanic forcing datasets (Gao et al., 2008; Crowley and Unterman, 2013) relying on ice core reconstructions of volcanism are used as input in the LM ModelE2-R simulations (and are the CMIP5/PMIP3 LM standard), as discussed in Section 2.2.

2.1.3. Tropical South American Climate

South America is home to nearly 390 million people. The continent spans a vast meridional extent (from ~10 °N to 55 °S), contains the world’s largest rainforest (the Amazon), in addition to one of the driest locations on Earth (the Atacama desert). The continent has diverse orography, spanning the high Andes along the Pacific to Laguna del Carbón in Argentina, the lowest point in the Southern Hemisphere. Because of this, South America hosts a rich diversity of climate zones and biodiversity, all of which may respond in unique ways to external forcing.

The most prominent climatic feature of tropical and subtropical South America is the South American monsoon system (Zhou and Lau, 1998; Marengo et al., 2001; Vera et al., 2006; Garreaud et al., 2009; Marengo et al., 2012). Much of South America is in a monsoon regime, with tropical/subtropical rainfall over the continent exhibiting a pronounced seasonal cycle. Unlike other monsoon systems such as that in Asia, low-level easterly winds prevail during the entire year in tropical South America, although the wind anomalies do change direction when the annual mean wind field is removed from winter and summer composites (Zhou and Lau, 1998).

During austral winter, the maximum in continental precipitation is largely restricted to north of the equator, in a band-like pattern associated with the oceanic Inter-Tropical Convergence Zone (ITCZ). During austral summer, convection is displaced from northwestern
South America, and a band of heavy precipitation covers much of the continent, from the southern Amazon Basin to central Brazil and northern Argentina. A distinctive feature of the SAMS is the South Atlantic Convergence Zone (SACZ), a band of cloudiness and precipitation sourced primarily from the tropical Atlantic that extends diagonally (southeastward) from the Amazon towards southeastern Brazil (Figure 2.2).

The SAMS onset occurs around the end of October and the demise between the end of March and April (e.g., Nogués-Paegle et al., 2002; Vera et al., 2006; Silva and Carvalho, 2007). The dominant mode of intraseasonal precipitation variability over South America during summer exhibits a dipole pattern (Nogués-Paegle and Mo, 1997), seesawing between the SACZ region and Southeastern South America, the latter including the densely populated La Plata basin with local economies strongly dependent on agricultural activities.

The SAMS is strongly modulated by ENSO behavior on inter-annual timescales (Vuille and Werner, 2005; Garreaud et al., 2009). In general, SAMS-affected regions of tropical South America tend to experience drier than normal conditions during El Niño, while conditions in subtropical latitudes are anomalously humid, including the southeastern part of the continent. Surface air temperatures tend to be anomalously warm in tropical and subtropical South America during El Niño events. These relationships depend somewhat on the time of year, and during La Niña events, the pattern is essentially reversed.

2.1.4. Recent South American Monsoon reconstructions from isotopic proxies

SAMS variability spanning most of the Holocene has been diagnosed from speleothem records in the Peruvian Andes (Kanner et al., 2013) and a review focused on the last 1,000-2,000 years was given in Bird et al. (2011) and Vuille et al. (2012). In all cases, a critical piece of
information that is required to properly diagnose paleo-SAMS variability is the ability to translate oxygen isotope variability from natural recorders into a physical climate signal of interest.

Early work on isotopes in ice core records from the tropical Andes detected a Little Ice Age (LIA) signal in the oxygen isotope composition of the ice, with results initially interpreted to reflect variations in local temperature due to their resemblance to ice core records from Greenland (e.g., Thompson et al., 1995, 1998) and due to their isotopic enrichment over the past 150 years, in parallel with rising global mean temperatures (Thompson et al., 2006). A temperature-dependence to oxygen isotope variability has been long known and is particularly important in mid-to-high latitudes (Dansgaard, 1964) and is most directly related to the ratio of initial and final water vapor content of a parcel that is transported horizontally, rather than the temperature-dependence of fractionation itself (Hoffman and Heimann, 1997).

This interpretation in the tropics has been challenged through a number of observational and modeling efforts (Hardy et al., 2003; Vuille and Werner 2005; Vimeux et al., 2005, 2009; Kanner et al., 2012) which suggest that the isotopic signal is more closely related to the degree of rainout upstream in regions of intense convection (in the case of South America, over the Amazon basin). Additionally, since sea surface temperatures (SST) in the Pacific have a large influence on SAMS intensity on inter-annual timescales in the present, oxygen isotope variability over much of tropical South America is linked to the state of the equatorial Pacific (Bradley et al., 2003; Vuille et al., 2003a,b).

In regimes that are highly convective in nature as in tropical South America, empirical evidence shows that the amount of precipitation (the so-called “amount effect”, Dansgaard, 1964) rather than the condensation temperature correlates most strongly with δ18Op variability, at
least on seasonal to inter-annual time scales. In reality, however, the rainout most relevant for the oxygen isotope signal may be at a significant distance from the site where the proxy is derived, potentially complicating the use of local calibrations to climatology as a guide for $\delta^{18}O_p$ interpretations (Schmidt et al., 2007). Isotopic concentrations are explainable as being a function of isotopic concentration of evaporative fluxes, rainout along the moisture transport path, and mixing.

The influence of precipitation amount on $\delta^{18}O_p$, in addition to changes in the partitioning of precipitation sources, has also been identified on decadal to orbital timescales through speleothem records and lake sediments (Cruz et al., 2005; Van Breukelen et al., 2008; Bird et al., 2011; Kanner et al., 2012). These studies have also highlighted the role of latitudinal displacements of the ITCZ, which is ultimately the main moisture conduit for precipitation over the South American continent. Furthermore, many records collected throughout South America now provide evidence for enriched $\delta^{18}O_p$ values during the Medieval Climate Anomaly, which is indicative of weakened SAMS convection and rainout, followed by depleted $\delta^{18}O_p$ values, suggesting heavier rainfall during the LIA in tropical South America (Bird et al., 2011; Apaestegui et al., 2014) with an opposite response in Northeast Brazil (Novello et al., 2012). This, in turn, has been interpreted in terms of North Atlantic SST anomalies (Vuille et al., 2012; Ledru et al., 2013) and the position of the Atlantic ITCZ.

Nonetheless, oxygen isotopes respond in unique ways depending on the climate forcing of interest. Indeed, a unique, quantitative local relationship between an isotope record and any particular climate variable of interest is unlikely to hold for all timescales and prospective forcing agents (Schmidt et al., 2007) thus motivating the use of forward modeling to work in
conjunction with proxy-based field data. For the remainder of this paper, we focus specifically on the volcanic forcing response.

2.2 Methodology

2.2.1 Data

The primary tool used in this study is the water isotope-enabled GISS ModelE2-R. ModelE2-R is a fully coupled atmosphere-ocean GCM (LeGrande et al., 2016, in prep; Schmidt et al., 2014a) that explicitly tracks stable water isotopes. The version used here is the same as the non-interactive atmospheric composition (NINT) physics version used in the CMIP5 experiments (Miller et al., 2014). The current model features 2° latitude x 2.5° longitude horizontal resolution and 40 vertical levels in the atmosphere up to 0.1 hPa, and is coupled to the Russell Ocean that conserves heat, water mass, and salt (Russell et al., 1995) at 1° x 1.25° resolution with 32 vertical levels. ModelE2-R includes stratospheric dynamics and prescribed ozone and aerosol species.

Due to uncertainties in past radiative forcing, a suite of LM simulations using ModelE2-R have been run with different combinations of plausible solar, volcanic, and anthropogenic land use histories (Schmidt et al., 2011, 2012) but with identical greenhouse gas and orbital evolution. These simulations span the period 850-2005 C.E. There are two reconstructions of past volcanic activity (Gao et al., 2008; Crowley and Unterman, 2013) that are used in six combinations of the ModelE2-R LM simulations (see the ‘past1000’ experimental design at http://data.giss.nasa.gov/modelE/ar5/, and below).
For the LM, three forcing combinations are available in the GISS ModelE2-R simulations that use the Crowley reconstruction for volcanic perturbations. These include Pongratz et al. (2008) [land]/Krivova et al. (2007) [solar], Kaplan et al (2010) [land]/Krivova et al. (2007) [solar], and Pongratz et al. (2008) [land]/Steinhilber et al. (2009) [solar] (see Schmidt et al., 2011, 2012). We focus only on results from the Crowley reconstruction prior to 1850 CE due to a mis-scaling of the Gao forcing in the model that roughly doubled the appropriate radiative forcing. For the historical period (1850-present), the volcanic forcing history is based on Sato et al. (1993) and is equivalent among the different (six) simulation members.

Crowley and Unterman (2013) discuss the details behind the LM Aerosol Optical Depth (AOD) reconstruction that defines the volcanic forcing time-series in ModelE2-R (Figure 2.1). This estimate is derived from sulfate peaks in ice cores, which are relatively well dated and referenced to the historical record during the satellite era. Crowley and Unterman (2013) provide an AOD history over 4 latitude bands (from 0-30° and 30-90° in both hemispheres). ModelE2-R uses a cubic spline to interpolate this forcing dataset over 24 latitude bands. The choice of volcanic eruptions used for the LM analysis (section 2.2 below) is based on the AOD dataset from this 24-latitude grid.

We note that there are more recent volcanic reconstructions available (e.g., Sigl et al., 2015) suggesting modifications to the timing or magnitude of LM eruptions, as well as developments of datasets focusing on sulfur injection and microphysics-based evolution of the aerosol forcing (e.g., Arfeuille et al., 2014). In this contribution, we are agnostic concerning the veracity of the forcing datasets that were standard for CMIP5/PMIP3, but stress that timing of eruptions is irrelevant in our modeling context and that the model results should be interpreted as a self-consistent response to the imposed AOD and particle size.
Water isotope tracers are incorporated into the model’s atmosphere, land surface, sea ice, and ocean. These isotopes are advected and tracked through every stage of the hydrologic cycle. At each phase change (including precipitation, evaporation, ice formation or melting) an appropriate fractionation factor is applied (Schmidt et al., 2005) and all freshwater fluxes are tagged isotopically. Stable isotope results from the lineage of GISS models have a long history of being tested against observations and proxy records (e.g., Schmidt et al., 2007; LeGrande and Schmidt, 2008, 2009; Lewis et al., 2010, 2013, 2014; Field et al., 2014).

In addition to the model, we briefly explore the observed instrumental record to assess responses to eruptions occurring during the 20th century. To do this, we take advantage of the NASA GISS Surface Temperature analysis (GISTEMP) land-ocean index (Hansen et al., 1999), and Global Precipitation Climatology Centre (GPCC) v6, a monthly precipitation dataset over land (Schneider et al., 2011). For Figures 2.2 and 2.3 where ocean climatological data is shown, we use the Global Precipitation Climatology Project (GPCP) version 2.2 (Adler et al., 2003), a combined land station and satellite product available since 1979. These datasets are called upon to gauge the tropical climate response following the three L20 eruptions. We use the 2.5° resolution GPCC dataset, as that is comparable to the GISS model and what is justified by the station coverage in this part of the world. The GPCC product offers considerably better global and South American coverage than other precipitation datasets, although observational density for rainfall is still considerably more problematic over South America than for many other regions of the globe. There is a sharp drop-off in the number of rain gauge stations used earlier in the 20th century over much of the South American continent. Figure S1 shows the station density at the time of each L20 eruption, as well as the total number of land stations over South America with time.
Finally, in section 2.3.1 we present data from the Global Network of Isotopes in Precipitation (GNIP) accessible from the International Atomic Energy Agency (IAEA) for $\delta^{18}O_p$, as a test of the model’s ability to track the seasonal hydrologic cycle in the form of its isotopic response over South America before discussing the Last Millennium results. Unfortunately, there is considerable spatial and temporal heterogeneity in the GNIP data over South America. In fact, only a few stations have data overlap with one or two eruptions and with a sufficient number of $\delta^{18}O_p$ data points to establish reasonable seasonal or annual statistics. Additionally, the post-volcanic (L20) anomalous isotope field over South America strongly resembles the ENSO expression on the isotope field (Vuille et al., 2003a) and with large spread between events (not shown). This suggests that internal variability (ENSO) dominates the forced (volcanic) response in this very small historical sample size, thereby leaving little hope that the prevailing network of observations is suitable for hypothesis testing and model validation in our context.

2.2.2. Super-posed Epoch and Composite Analysis

We present the spatial pattern of observed and simulated response for temperature and precipitation over land for two L20 eruptions (El Chichón and Mt. Pinatubo). Results are shown for annual-means in 1983 and 1992. We choose only two for brevity, as our argument that assessing the signal in any specific region is difficult in a small sample of eruptions is unaffected. Because of the dominant influence of unforced variability on tropical South American climate (Garreaud et al., 2009) overriding the volcanic signal during the L20 eruptions, we instead present a superposed epoch anomaly composite of the tropical-mean temperature anomaly, zonally averaged from 30°S to 30°N. Results are shown for years -3 to +5, with zero defining the time of the eruption. This composite is formed for all three L20 eruptions.
In all cases, the five years prior to the eruption were subtracted from the superposed composite. Other sensible choices for the non-eruption reference period do not significantly change the results.

For the full LM spatial composites, we use only eruptions where vertically integrated (15 to 35 km) stratospheric AOD averaged from 30°N to 30°S exceeds 0.1 for at least 12 consecutive months in the simulation (top panel in Figure 2.1). For the LM composites, we focus only on seasonal (DJF and JJA) composites, and a given season will enter the composite if at least 2/3 months meet the AOD threshold; this criterion yields 15 eruptions since 850 C.E. The selection of events used in the LM composite is very weakly sensitive to this choice of latitude band. Mt. Pinatubo is the only L20 eruption in this composite, and is actually one of the smallest eruptions in this selection based on the maximum AOD encountered near the time of the eruption (see Table 1 for dates of each event). We believe sampling a larger number of events with greater forcing is a better way to understand the volcanic response in this model, rather than increasing the ensemble size for the L20 events.

For the LM “non-eruption” fields, we use 15 years prior to the eruption as a reference period to calculate the anomaly for each event, unless another event occurs during that time (overlap occurs only once for eruptions in 1809 and 1815) in which case the pre-1809 climatology is used twice. The exception is for Mt. Pinatubo, which again uses the previous five years to calculate the anomaly. When constructing seasonal averages of \( \delta^{18}O_p \), the oxygen isotope value for each month is weighted by the precipitation amount during that month, at each grid cell.

Since each post-eruption difference field is computed using the immediate response minus a local 15-year climatology, time is not relevant in this analysis and so we use all three
members with the Crowley forcing (representing over 3,000 years of simulation time) to generate a composite that features 45 volcanic “events” (15 eruptions in each of the three members). In the historical (post-1850) extension of these runs, the coding error that resulted in a misimplementation of the Gao forcing is not an issue, and so we use six ensemble members each (three volcanic events in six ensemble members) for the L20 results.

The ensemble-mean composite results displayed for the LM eruptions include contributions from three members that differ not just in the internal variability, but also in their solar and land-use forcing. Similarly, the L20 results are from model runs that also include other transient historical forcings occurring at the time of the eruption, including greenhouse gas increases throughout the duration of the event (although these forcings are the same among all ensemble members). However, in all cases we focus only on the immediate years after the eruption. Since the primary signal of interests is expected to be large compared to the impact of more slowly varying and smaller-amplitude forcings, the ensemble spread for a given eruption can be interpreted as a sampling of the model internal variability coincident with the event. We have tested our composite results using the same dates as our volcanic events in simulations with other varying forcings with no volcanoes (there are no volcano-only runs with this model version for the LM), and the results are indistinguishable from noise (not shown). The LM composite results are discussed in section 2.3.2.

Finally, it is now well appreciated that any climate response under investigation will be shackled to the spatial structure of the forcing imposed on a model. For example, preferential heating/cooling of one hemisphere will induce different tropical precipitation responses than a well-mixed gas that behaves CO2-like (Kang et al., 2008, 2009; Frierson and Hwang, 2012; Haywood et al., 2012). Figures S2 and S3 show the latitudinal AOD distribution structure for all
eruptions used in the generation of the LM composites within ModelE2-R. The mean of all events is rather symmetric between hemispheres (though somewhat skewed toward the Southern Hemisphere tropics). Thus, the resulting climate responses outlined in this paper ought to be viewed as a response consistent with a forcing that is relatively symmetric about the equator.

2.2.3. Influence of ENSO on the Late 20th Century (L20) eruptions

For the L20 volcanic events, El Niño events are occurring quasi-simultaneously with the eruption. This introduces a pervasive issue when attempting to isolate the volcanic signal (e.g., Robock, 2003; Trenberth and Dai, 2007; Joseph and Zeng, 2011) and is particularly important over South America (e.g. Garreaud et al., 2009).

In order to remove the effects of ENSO from the superposed epoch and spatial composite analyses described above in the GISTEMP and GPCC data, we first perform a multiple regression with the variable of interest over the period 1951-2005 using a linear time trend and the Niño 3 index as predictors (5°N-5°S, 150°W-90°W, data from http://www.cpc.ncep.noaa.gov/data/indices/) over the same period, excluding two years of data after each L20 eruption. At each grid cell, the Niño 3 index is lagged from 0-6 months and the correlation coefficient with the maximum absolute value (since a positive index can induce a negative anomaly in the variable of interest) is found. This is similar to the approach used in Joseph and Zeng (2011), allowing the maximum ENSO influence to be removed at each grid point at the expense of contemporaneous relationships. The lagged Niño index is then regressed against the time series of each variable and the residual from this regression is retained. This approach assumes a linear relationship between ENSO and the climate response over South
America, an assumption that appears justified on inter-annual to decadal time scales (Garreaud et al., 2009).

For each of the six ensemble members used in the model L20 composite, a similar procedure is performed in which the Niño 3 index (consistent with the realization of the Niño 3 domain SSTs in that model simulation) is calculated and regressed out in the same manner. For the full LM computations, the number of larger-amplitude events in the three-ensemble member composite should help average out the influence of Pacific SST variability, and no ENSO removal procedure is applied.

2.3. Results and Discussion

2.3.1. L20

Figure 2.3 illustrates that ModelE2-R reproduces the seasonal cycle of climatological rainfall (comparing Figure 2.3a with 2.3b) and oxygen isotope distribution (comparing Figure 2.3c with 2.3d) with some fidelity over South America. This includes a meridional migration of the ITCZ toward the summer hemisphere and an intensification of the South American monsoon during DJF. Where data permit (Figure 2.3c) there is good agreement between model and observations, both displaying oxygen isotope DJF enrichment relative to JJA in the tropics north of the equator and the higher latitudes south of 30°S, and depletion in the continental interior south of the equator associated with the monsoon wet season. ModelE2-R (Figure 2.3b) tends to produce too much precipitation over northeastern Brazil although the gross features of the seasonal migration in rainfall are well captured. This ability to accurately simulate the seasonality of δ¹⁸O_p over the tropical Americas has also been noted in two atmospheric GCMs
with no coupled ocean (NASA-GISS II and ECHAM-4, see Vuille et al., 2003a). The model also provides a skillful representation of δ^{18}O_p in response to ENSO (not shown), including increased δ^{18}O_p values over tropical South America in response to El Niño (Vuille and Werner, 2005).

Figure 2.4 shows the ENSO-removed superposed epoch analysis for tropical temperature associated with the recent three L20 eruptions. There is good agreement between the observed and modeled temperature response, both in amplitude and recovery timescale. The tropical-mean cooling is on the order of several tenths of a degree, and larger after Mt. Pinatubo (not shown individually).

The spatial structure of the post- El Chichón and Pinatubo events in land observations and the individual model realizations are shown in Figures 2.5 and 2.6, respectively. Observations exhibit cooling over much of the globe, especially after Mt. Pinatubo that is largely reproduced by the model. However, there is considerable spread among the individual ensemble members and between the two events, indicating a large role for internal variability in dictating the observed spatial pattern following these events. This is also true over South America. In GISTEMP, the high-latitudes of South America cool more than the tropical region of the continent after Mt. Pinatubo. There is still a residual signal from ENSO in tropical South America following both L20 eruptions that is not reproduced by the model. This is not unexpected, since ENSO events comparable to the magnitude of the historic realizations due not occur coincident with the volcanic forcing in the individual ensemble members. The magnitude of this signal is sensitive to the Niño index used in the regression method described above. Without ENSO removal, tropical South America warms following the two eruptions (not shown). The influence of ENSO appears minimal over the higher latitude sectors of the continent.
The precipitation pattern following the L20 eruptions exhibits substantial variability in space and across eruptions, with a general drying pattern over land in tropical latitudes. South America experiences less precipitation near the equator after Mt. Pinatubo (see also Trenberth and Dai, 2007), a pattern reproduced in some of the ensemble realizations. It should be noted that model-observation comparison is hindered not just by internal variability, but also by the specified historical volcanic forcing in the model. In fact, the Stratospheric Aerosol and Gas Experiment (or SAGE) II satellite sensor was saturated by the aerosol cloud after Mt. Pinatubo; subsequent work (Santer et al., 2014; Schmidt et al., 2014c) suggests that the forcing following Pinatubo is too large in the CMIP5 generation of models.

Because of the considerable variability seen in observations (following historical eruptions) and also across ensemble members, it is evident that a larger signal-to-noise ratio than is available from the L20 eruptions alone is required to help isolate any volcanic signal. ModelE2-R is the laboratory from which we proceed to sample a larger number of events, some of which contain larger amplitude than the L20 eruptions.

2.3.2. Last Millennium Composites

2.3.2.1. Temperature and Precipitation

Figure 2.7 shows the LM post-volcanic temperature composite for all 45 events. During both seasons, cooling is statistically significant over virtually the entire continent (stippling indicates significance at the 90% level, t-test). The temperature response is strongest in the interior of the continent, particularly during the austral winter. The enhanced high-latitude cooling exhibited in the observations after Mt. Pinatubo does not emerge in the model composite.
The precipitation anomalies for the LM composite are shown in Figure 2.8. As expected, there is a distinct seasonal structure in the response, with the largest anomaly concentrated in a narrow region north of the equator during austral winter, coincident with the location of climatological rainfall maxima in the region. During JJA, precipitation increases in the North Atlantic region following volcanic eruptions, while very strong and statistically significant precipitation reductions occur just north of the equator (including over northern Brazil, Ecuador, Venezuela, Colombia, and Guyana) and encompassing the northern Amazon Basin. This signal is consistent with a weakening of the moisture flux owing to the decrease in saturation vapor pressure due to cooling that is demanded by Clausius-Clapeyron (Held and Soden, 2006). During this season, the precipitation response is significant virtually everywhere in northern South America. Supplementary Figure (S5) further illustrates that the JJA precipitation response is remarkably robust to all eruptions that enter into the composite.

Figure 2.9b illustrates the relationship between area-averaged precipitation from 20°S to 0°, 77.5°W to 45°W (for DJF) and 0° to 10°N, 77.5°W to 52.5°W (for JJA) and the maximum AOD encountered for each eruption. These two regions were selected to reflect the seasonal migration of rainfall (Figure 2.2). 15 eruptions are displayed with the three-member ensemble spread given for each. Precipitation only increases north of the equator during austral winter in a few model realizations. Moreover, the magnitude of the precipitation response during JJA scales with the size of the eruption, particularly for very large eruptions (e.g., comparing five eruptions with AOD > 0.3 vs. those with smaller perturbations, although the spread amongst the ensemble members is large). The spatial composite for each individual eruption (each averaged over the three ensemble members) is shown in Figure S5.
The precipitation response during austral summer is more difficult to interpret (Figure 2.8a). During this season, the zonally oriented Atlantic ITCZ migrates southward and the SACZ becomes more intense as it is connected with the area of convection over the central and southeastern part of the continent. It is noteworthy that the land cools substantially more than the surrounding ocean (Figure 2.7), which one could expect to weaken the monsoon-sourced precipitation during DJF. While precipitation is indeed reduced over the tropical continent, the response is weaker than in JJA and less spatially coherent, with many areas failing to meet statistical significance. An analysis of the individual responses reveals that the signal is more eruption-dependent during DJF than during JJA (see Figure S4), with a few events actually exhibiting modest increases in precipitation. Nonetheless, there is a clear tendency for reduced DJF precipitation within the SAMS region, although there is little to no dependence of the mean rainfall anomaly on the magnitude of the AOD perturbation, at least above the 0.1 threshold used in this study (Figure 2.9b), unlike for equatorial South America during JJA. Conversely, the temperature response (Figure 2.9a) depends on the size of the eruption in both seasons, as is expected given its dependence on the size of the radiative forcing.

2.3.2.2. Tropical Hydroclimate Response

Since the South American climate is intimately linked to large-scale tropical dynamics, the global precipitation composite is shown in Figure S6 to better inform the model response. The most robust signal is characterized by a reduction in tropically averaged precipitation and the tendency for wet regions to become drier, and dry regions to become wetter (see also Iles et al., 2013; Iles and Hegerl, 2014), in contrast to the anticipated hydrologic response in a future, higher-CO$_2$ world (Held and Soden, 2006).
This pattern is a thermodynamic effect linked to reduced moisture convergence within the convergence zones and to reduced moisture divergence in the descending zones of the Hadley cell, which reduces the contrast in values of precipitation minus evaporation (P-E) between moisture convergence and divergence regions (Chou et al., 2009). The complete hydrologic response of the ΔP-E field (not shown) has the same spatial structure as the ΔP field, since evaporation is decreasing nearly everywhere in the tropics. Because both P and E are decreasing on the equator-ward flank of the ITCZ the ΔP-E signal is rather weak in the deep tropics, while ΔP-E increases more rapidly than ΔP in the subtropics.

The tendency for modest precipitation anomalies over the continent during DJF appears to be part of a pattern that spans a broad swath of longitudes across the entire deep tropics in association with the seasonal cycle. Nonetheless, the response during DJF is weaker over land.

2.3.2.3. Oxygen Isotope Anomalies

In order to relate the responses discussed in the previous sections back to a potentially observable paleoclimate metric, we show the composite Δδ¹⁸O_p field for the DJF and JJA seasons in South America (Figure 2.10). It should be cautioned that much of the isotopic variability that can be observed in proxies within the continental interior or high-elevation glacier sites will likely be seasonally biased toward the wet season months (Hardy et al., 2003).

During the JJA season, there is a strong enrichment of the δ¹⁸O_p pattern that is zonally extended over equatorial South America. In addition, there is a corresponding δ¹⁸O_p depletion in the adjacent North Atlantic sector. This response is inextricably coincident with the strong change in precipitation in the ITCZ domain that was assessed in Figure 2.8, and is broadly
consistent with a “rainfall amount” control on the isotopic imprint (Dansgaard, 1964). South of approximately 15°S, the sign of the anomaly reverses to a depletion of the heavy isotope. During the austral summer, volcanic eruptions lead to a clear negative excursion in $\delta^{18}O_p$ over virtually the entire SAMS region, including the Amazon basin, tropical Andes, and eastern Brazil. The statistical significance of the resulting isotopic anomaly extends throughout most of the landmass within the tropics and in the North Atlantic. There are small but non-significant exceptions (positive $\delta^{18}O_p$ excursions) such as in eastern Brazil. The negative excursions also include regions outside of the SAMS belt in the subtropics and mid-high latitudes of South America.

The austral summer $\delta^{18}O_p$ depletion is the opposite sign from what one would expect if the reduced precipitation were driving the isotopic response. Thus, it may well be that the strong temperature response to volcanic eruptions dominates the continent-wide oxygen isotope depletion during the DJF season and in the extratropics during JJA over the relatively weak precipitation response. Precipitation on the other hand appears to be the primary control knob of $\Delta\delta^{18}O_p$ during JJA within the ITCZ region.

The correlation between $\Delta\delta^{18}O_p$ and temperature or precipitation, based on a regression using all 45 volcanic events, is reported in Figure 2.9, using the same domains for DJF and JJA described in section 2.2.1. In the case of volcanic forcing it appears that the amplitude of the temperature response to volcanic eruptions over tropical South America is larger than the rather weak and spatially incoherent precipitation signal, although both the temperature and precipitation coefficients must be considered to characterize the isotopic variability during this season (Figure S7). This may explain why the DJF isotopic signal related to volcanic eruptions seems to respond to atmospheric cooling, even in the tropics, where isotopic variability is usually
more closely associated with changes in the hydrologic cycle. During JJA, the isotopic enrichment is much more closely associated with precipitation reduction north of the equator, whereas the JJA Δδ\textsuperscript{18}O\textsubscript{p}-temperature relationship is weak and non-significant.

Taken together, these results suggest that the primary controls on oxygen isotope variability may vary by forcing agent, rather than being determined inherently by the latitude of interest (e.g., “precipitation driven” in the tropics and “temperature driven” in the extratropics). This conclusion is compelled by the fact that the precipitation production and distribution in proxy records are the result of an interaction between multiple scales of motion in the atmosphere, the temperature of air in which the condensate was embedded, and exchange processes operating from source to sink of the parcel deposited at a site. Thus, a consistent description of how to interpret oxygen isotopes into a useful climate signal cannot be given without considering all of these processes and the target process of interest.

To further complement the spatial analysis, a composite Hovmöller diagram is utilized (Figure 2.11) in order to illustrate the time-evolution of the temperature, precipitation, and oxygen isotope response. For this plot, the start of each eruption is defined as the closest January to the first month in which AOD reaches 0.1 in order to illustrate the seasonal evolution (rather than compositing by “month from each eruption” as in Figure 2.4). Therefore, for all 45 events in the composite, the local AOD may reach this threshold within five months (before or after) of the January baseline point (eruptions in June are rounded up to the following January). The Hovmöller composites are plotted for ten years (beginning January three years prior to the eruption). The closest January point to the start of each eruption occurs in the 37\textsuperscript{th} month of the Hovmöller (solid black line in Figure 2.11a,b,d). Results are zonally averaged from 77.5°W to 45°W.
Figure 2.11a demonstrates a substantial temperature anomaly that peaks south of 10°S (compare also to Figure 2.7). The cooling lasts for several years following the eruption, and decays gradually until most of the signal is lost (~4 years after the eruption in the South American sector), but remains 0.1-0.2°C colder than the pre-eruption climatology. The zonally averaged peak reductions in South American precipitation anomalies occur over the tropical latitudes and last for a comparable period of time as the maximum temperature response. The precipitation anomaly itself migrates synchronously with the seasonal cycle (red line in Figure 2.11c maps out the latitude of maximum climatological precipitation averaged over all 15 year climatologies of each 45-member event, as a function of time of year). Figure 2.11b indicates that the largest precipitation response is confined to the equatorial and northern regions during JJA, with weak protrusion into higher tropical latitudes only 1-2 years after the eruption. The JJA isotopic enrichment in northern South America lasts for two seasons in our composite, while there is sustained isotopic depletion during DJF in the SASM region for about three years (Figure 2.11d).

Figure 2.12 provides additional statistical insight into the magnitude of the excursions described in this section. Here, we sampled 100 random 45-event composites in a control simulation with no external forcing (each “event”, two seasons in length, is defined as an anomaly expressed relative to a pre-eruption climatology as done previously). The anomalies were averaged over the same areas as in Figure 2.9, with different domains for DJF and JJA. Notably, for both seasons and for all three variables examined, the single 45-event post-volcanic composite (purple square) lies outside the distribution of all sampled 45-event composites constructed with no external forcing. Nonetheless, the distribution for a smaller sample of events
(black circles denote the data for each (15) eruption, each averaged over the three ensemble members) shows considerable spread.

The $\delta^{18}O$ anomalies discussed above result from changes in the isotopic content of precipitation, which may be due to changes in precipitation amount or to other changes in the isotopic composition of the water vapor that condensed to form the precipitate. The changes are not determined by changes in the seasonality of the precipitation. To illustrate this (Figure S8), we decomposed the $\Delta\delta^{18}O$ field (see Liu and Battisti, 2015) by weighting the monthly oxygen isotope field by the pre-eruption precipitation values. The results are indistinguishable from the total $\Delta\delta^{18}O$ field, suggesting that any changes in monsoon seasonality are negligible in contributing to the isotopic signal, unlike the orbital case considered in Liu and Battisti (2015).

2.4. Conclusions

In this study, we have analyzed the response of temperature, precipitation, and $\delta^{18}O$ over South America to volcanic forcing associated with large tropical eruptions during the Last Millennium. It is now well known that volcanic eruptions lead to large-scale cooling throughout the tropics, and this result extends to most of the South American continent as well, except in regions that may be simultaneously affected by opposing ENSO behavior. In general, the precipitation response has been more enigmatic, though our results are in broad agreement with numerous other studies showing that there is a substantial decline in tropical-mean precipitation.

However, the immediate post-volcanic impact over South America has a complex seasonal and spatial structure. During the austral winter, the precipitation response over the continent is slaved to the response of the large-scale circulation, including a weakening of
rainfall intensity within the ITCZ that is migrating northward. In the extratropics, the continent cools and exhibits slight precipitation declines nearly everywhere. Our results suggest the seasonal monsoon precipitation (during DJF) in ModelE2-R exhibits a fairly weak response that is scattered across the continent. It appears that volcanic forcing preconditions the tropical rainfall over the continent to decline during the wet season, but that this response is likely to be eruption-dependent and may be overwhelmed by internal variability.

A unique aspect of this study was to probe the $\delta^{18}$O$_p$ response to volcanic eruptions. During JJA, the precipitation isotopic composition is less negative in northern South America as convective activity produces substantially less precipitation. No such relation was found during the monsoon season, even within the tropics, where the large cooling appears to lead to more depleted $\delta^{18}$O$_p$, despite a weakened hydrologic cycle and reduced monsoon precipitation. In the extratropics, it appears that the temperature decline is driving isotopes toward more depleted values.

Unfortunately, validation of our model results is hindered by the paucity of observational stable isotope data and by the coincidence of volcanic eruptions with ENSO events over the 20th century. Nonetheless our results may provide some guidance in the search of volcanic signals in high-resolution isotopic or other temperature- and precipitation-sensitive proxy data from South America. Given the importance of volcanic forcing for climate variability over the past millennium, and in particular the LIA period, which has been identified as a period of significant climatic perturbation in isotopic proxies from South America, a better understanding of the climatic response to volcanic forcing over this region is urgently needed.

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**Supplementary information:** A link to the supplementary information for this chapter may be found at: [http://www.clim-past.net/12/961/2016/cp-12-961-2016-supplement.pdf](http://www.clim-past.net/12/961/2016/cp-12-961-2016-supplement.pdf)
References


Bradley, R.S., Vuille, M., Hardy, D.R., and Thompson, L.G.: Low latitude ice cores record


Table 1: Time of Eruptions and Global Aerosol Optical Depth (AOD) from Crowley and Unterman (2013). List of eruptions used in study.

Table 1. List of LM and L20 Eruptions

<table>
<thead>
<tr>
<th>Start Date of Eruption(a)</th>
<th>Seasons in LM Composite</th>
<th>Max AOD(b)</th>
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<td>972</td>
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<td>Apr 1963(d)</td>
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<tr>
<td>Apr 1982(d)</td>
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<tr>
<td>Jun 1991</td>
<td>1992</td>
<td>1992</td>
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\(a\)Start of Eruption dates based on when they can be identified in the Crowley /Sato time-series averaged over the latitude band from 30°S to 30°N. May be slightly different than actual eruption date.

\(b\)Maximum AOD over the 30°S to 30°N latitude band encountered in monthly time-series during the duration of each event.

\(c\)December in year prior to listed date.

\(d\)Mt. Agung and El Chichón included in L20 but not LM composites.
Figure 2.1. Aerosol Optical Depth (AOD) used to force the NASA GISS ModelE2-R over the Last Millennium and (bottom) zoomed in on the period 1950-1999 (Crowley+Sato) as discussed in text. AOD is the vertically integrated (15-35 km) and latitudinal average from 30°S to 30°N. Note difference in vertical scale between graphs. Orange dashed line marks the AOD threshold for defining a LM eruption in the present study. Eruption events defined in text must sustain the threshold AOD for at least one year, so not all events above the orange dashed line are used in the composites.
Figure 2.2. (Top) Observed Climatological Precipitation for DJF (shading, in mm day\(^{-1}\)). SAMS box is drawn over the domain from 20°S to 0°, 77.5°W to 45°W and used for Figure 9 and 12. Data from the
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(averaged over the same domain) in the entire control simulation with no external forcing. In both panels, the correlation coefficient and p-value are reported for a) temperature and b) precipitation vs. $\Delta \delta^{18}O_p$ in each season and over the same domain. The regression uses all 45 volcanic events.
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Figure 2.11. Last Millennium Hovmöller diagram (10 years, time moving forward going upward, with year number labeled next to each month) for a) temperature anomaly (°C) b) precipitation anomaly (mm day$^{-1}$) using procedure described in text. Solid black lines mark closest January to start of each eruption used in composite. c) Same as panel b, except zoomed in on 10 °S to 10 °N and over 3 years of time beginning with the January closest to each eruption. Red line in panel c shows latitude of maximum climatological precipitation as a function of time of year. All results zonally averaged in model from 77.5°W to 45°W. d) Last Millennium Hovmöller diagram for oxygen isotopes in precipitation (per mil).
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Chapter 3

Hemispherically asymmetric volcanic forcing of tropical hydroclimate during the last millennium

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Abstract

Volcanic aerosols exert the most important natural radiative forcing of the last millennium. State-of-the-art paleoclimate simulations of this interval are typically forced with diverse spatial patterns of volcanic forcing, leading to different responses in tropical hydroclimate. Recently, theoretical considerations relating the intertropical convergence zone (ITCZ) position to the demands of global energy balance have emerged in the literature, allowing for a connection to be made between the paleoclimate simulations and recent developments in the understanding of ITCZ dynamics. These energetic considerations aid in explaining the well-known historical, paleoclimatic, and modeling evidence that the ITCZ migrates away from the hemisphere that is energetically deficient in response to asymmetric forcing.

Here we use two separate general circulation model (GCM) suites of experiments for the Last Millennium to relate the ITCZ position to asymmetries in prescribed volcanic sulfate aerosols in the stratosphere and related asymmetric radiative forcing. We discuss the ITCZ shift in the context of atmospheric energetics, and discuss the ramifications of transient ITCZ migrations for other sensitive indicators of changes in the tropical hydrologic cycle, including global streamflow. For the first time, we also offer insight into the large-scale fingerprint of water isotopologues in precipitation ($\delta^{18}O_p$) in response to asymmetries in radiative forcing.

The ITCZ shifts away from the hemisphere with greater volcanic forcing. Since the isotopic composition of precipitation in the ITCZ is relatively depleted compared to areas outside this zone, this meridional precipitation migration results in a large-scale enrichment (depletion) in the isotopic composition of tropical precipitation in regions the ITCZ moves away from (toward). Our results highlight the need for careful consideration of the spatial structure of
volcanic forcing for interpreting volcanic signals in proxy records, and therefore in evaluating the skill of Common Era climate model output.
3.1. Introduction

The ITCZ is the narrow belt of deep convective clouds and strong precipitation that develops in the rising branch of the Hadley circulation. Migrations in the position of the ITCZ have important consequences for local rainfall availability, drought and river discharge, and the distribution of water isotopologues (e.g., $\delta^{18}$O and $\delta$D, hereafter simply referred to as water isotopes, with notation developed in section 3.3) that are used to derive inferences of past climate change in the tropics.

Meridional displacements of the ITCZ are constrained by requirements of reaching a consistent energy balance on both sides of the ascending branch of the Hadley circulation (e.g., Kang et al., 2008, 2009; Schneider et al., 2014). Although the ITCZ is a convergence zone in near-surface meridional mass flux, it is a divergence zone energetically. The stratification of the tropical atmosphere is such that moist static energy (MSE) is greater aloft than near the surface, compelling Hadley cells to transport energy in the direction of their upper tropospheric flow (Neelin and Held, 1987). If the system is perturbed with preferred heating or cooling in one hemisphere, the anomalous circulation that develops resists the resulting asymmetry by transporting energy from the heated to the cooled hemisphere. Conversely, meridional moisture transport in the Hadley circulation is primarily confined to the low-level equatorward flow, so the response of the tropical circulation to asymmetric heating demands an ITCZ migration away from the hemisphere that is energetically deficient. Since the mean circulation dominates the atmospheric energy transport (AET) in the vicinity of the equator, the recognition that the ITCZ is approximately co-located with the latitude where meridional column-integrated energy fluxes vanish has provided a basis for relating the mean ITCZ position to AET. We note that this
perspective focused on atmospheric energetics is distinct from one that emphasizes sea surface
temperature gradients across the tropics (Maroon et al., 2016).

This energetic framework has emerged as a central paradigm of climate change problems,
providing high explanatory and predictive power for ITCZ migrations across timescales and
forcing mechanisms (Donohoe et al., 2013; McGee et al., 2014; Schneider et al., 2014). It is also
a compelling basis for understanding why the climatological annual-mean ITCZ resides in the
northern hemisphere (NH); it has been shown that this is associated with ocean heat transport,
which in the prevailing climate is directed northward across the equator (Frierson et al., 2013;
Marshall et al., 2014). The energetic paradigm also predicts an ITCZ response for asymmetric
perturbations that arise from remote extratropical forcing. This phenomenon is exhibited in many
numerical experiments, is borne out paleoclimatically, and has gradually matured in its
theoretical articulation (Chiang and Bitz, 2005; Broccoli et al., 2006; Kang et al., 2008, 2009;
Yoshimori and Broccoli, 2008, 2009; Chiang and Friedman, 2012; Frierson and Hwang, 2012;
Bischoff and Schneider, 2014; Adam et al., 2016).

Thus far, however, little or only very recent attention has been given to the relation
between transient ITCZ migrations and explosive volcanism (although see Iles et al., 2014; Liu
et al., 2016, section 2). This connection has received recent consideration using carbon isotopes
in paleo-records (Ridley et al., 2015) or in the context of volcanic and anthropogenic aerosol
forcing in the 20th century (Friedman et al., 2013; Hwang et al., 2013; Allen et al., 2015;
Haywood et al., 2015). The purpose of this paper is to use the energetic paradigm as our vehicle
for interpreting the climate response in paleoclimate simulations featuring explosive volcanism
of varying spatial structure.
Much of the existing literature highlighting the importance of spatial structure in volcanic forcing focuses on the problem of tropical vs. high-latitude eruptions and dynamical ramifications of changing pole-to-equator temperature gradients (e.g., Robock, 2000; Stenchikov et al., 2002; Shindell et al., 2004; Oman et al., 2005, 2006; Kravitz and Robock, 2011), which is a distinct problem from one focused on inter-hemispheric asymmetries in the volcanic forcing. Furthermore, episodes with preferentially higher aerosol loading in the southern hemisphere (SH) have received comparatively little attention, probably due to the greater propensity for both natural or anthropogenic aerosol forcing to be skewed toward the NH.

Here we show that it matters greatly over which hemisphere the aerosol loading is concentrated and that this asymmetry in aerosol forcing has a first-order impact on changes in the tropical hydrologic cycle, atmospheric energetics, and the distribution of the isotopic composition of precipitation.

3.2. Methods

To illuminate how the spatial structure of volcanic forcing expresses itself in the climate system, we call upon two state-of-the-art models that were run over the pre-industrial part of the last millennium, nominally 850-1850 C.E. (hereafter, LM), the most recent key interval identified by the Paleoclimate Model Intercomparison Project Phase 3 (PMIP3). An analysis of this time period is motivated by the fact that volcanic forcing is the most important radiative perturbation during the LM (LeGrande and Anchukaitis, 2015; Atwood et al., 2016). Furthermore, the available input data that defines volcanic forcing in CMIP5/PMIP3 feature a greater sample of events, larger radiative excursions, and richer diversity in their spatial structure.
than is available over the historical period. This allows for a robust composite analysis to be performed over this interval.

The two general circulation models (GCMs) that we use as our laboratory are NASA GISS ModelE2-R (hereafter, GISS-E2) and the National Center for Atmospheric Research (NCAR) Community Earth System Model (version 1.1) Last Millennium Ensemble (hereafter, just CESM to describe this set of simulations). The GISS-E2 version used here is the same as the noninteractive atmospheric composition physics version used in the CMIP5 initiative (called “NINT” in Miller et al., 2014). CESM is a community resource that became available in 2015 (Otto-Bliesner et al., 2016) and consists of several component models each representing different aspects of the Earth system; the atmospheric component is the Community Atmosphere Model version 5 (CAM5, see Hurrell et al., 2013), which in CESM features a 1.9° latitude x 2.5° longitude horizontal resolution with 30 vertical levels up to ~2 hPa. The GISS-E2 model is run at a comparable horizontal resolution (2° x 2.5°) and with 40 vertical levels up to 0.1 hPa.

Both GISS-E2 and CESM feature multiple ensemble members that include volcanic forcing. There are only a small number of volcanic eruptions in our different forcing classifications (see below) in each 1000 year realization of the LM, motivating an ensemble approach to sample multiple realizations of each eruption. There are currently 18 members in CESM, including 13 with all transient forcings during the LM and five volcano-only simulations. This number is much higher than the number of ensembles used for participating LM simulations in CMIP5/PMIP3. The volcanic reconstruction is based on Gao et al., (2008, hereafter, G08) and the ensemble spread is generated from round off differences in the initial atmospheric state (~10⁻¹⁴ °C changes in the temperature field). Sampling many realizations of internal variability is
critical in the context of volcanic eruptions given the different trajectories that can arise in the atmosphere-ocean system in response to a similar forcing (Deser et al., 2012).

For GISS-E2, there exist six available members that include a transient volcanic forcing history. Here, however we use only the three simulations that utilize the G08 reconstruction. This was done in order to composite over the same dates as the CESM events, and because the other volcanic forcing dataset that NASA explored in their suite of simulations (Crowley and Unterman, 2013) only provides data over four latitude bands, complicating inferences concerning hemispheric asymmetry. Taken together, there are 21,000 years of simulation time in which to explore the post-volcanic response while probing both initial condition sensitivity and the structural uncertainty between two different models. The three GISS-E2 members also differ in the combination of transient solar/land-use histories employed, but since our analysis focuses only on the immediate post-volcanic imprint, the impact of these smaller amplitude and slowly varying forcings is very small. We tested this using the composite methodology developed below on no-volcano simulations with other single forcing runs (in CESM) or with combined forcings (in GISS-E2) and found the results to be indistinguishable from that of a control run (not shown).

In both GISS-E2 and CESM, the model response is a slave to the spatial distribution of the imposed radiative forcing, which was based on the aerosol transport model of G08, rather than the coupled model stratospheric wind field, thus losing potential insight into the seasonal dependence of the response that may arise in the real world. For our purpose, however, this is a more appropriate experimental setup, since the spatial structure of the forcing is implicitly known (Figure 3.1).

The original G08 dataset provides sulfate aerosol loading from 9 km to 30 km (at 0.5 km resolution) for each 10° latitude belt. This reconstruction is based on sulfate peaks in ice cores
and a model of transport that determines the latitudinal, height, and time distribution of the stratospheric aerosol. In CESM, aerosols are treated as a fixed size distribution in three levels of the stratosphere, which provide a radiative effect, including shortwave scattering and longwave absorption. The GISS-E2 model is forced with prescribed Aerosol Optical Depth (AOD) from 15-35 km, based on a linear scaling with the G08-derived column volcanic aerosol mass (Stothers, 1984; Schmidt et al., 2011), with a size distribution as a function of AOD as in Sato et al (1993) – thus altering the relative long wave and shortwave forcing (Lacis et al, 1992; Lacis, 2015).

We note that the GISS-E2 runs forced with the G08 reconstruction in CMIP5/PMIP3 were mis-scaled to give approximately twice the appropriate AOD forcing, although the spatial structure of forcing in the model is still coherent with G08. For this reason, we emphasize the CESM results in this study. However, we still choose to examine the results from the GISS-E2 model for two reasons. First, we view this error as an opportunity to explore the climate response to a wider range of hemispheric forcing gradients, even though it comes at the expense of not being able to relate the results to actual events during the LM. Secondly, the GISS-E2 LM runs were equipped with interactive water isotopes (section 3.3). A self-consistent simulation of the isotope field in a GCM is important, since it removes a degree of uncertainty in the error-prone conversion of isotopic signals into more fundamental climate variables. To our knowledge, an explicit simulation of the isotopic distribution following asymmetries in volcanic forcing has not previously been reported.

In our analysis, we classify volcanic events as “symmetric” (SYMM), and “asymmetric” (ASYMMx), where the subscript X refers to a preferred forcing in the Northern Hemisphere (NH) or Southern Hemisphere (SH). Composites are formed from all events within each of the
three classifications in order to isolate the volcanic signal. All events must have a global aerosol loading $> 8$ Tg ($1$ teragram = $10^{12}$ g) averaged over at least one five-month period to qualify as an eruption and enter the composite. For comparison, the 1991 Mt. Pinatubo eruption remains elevated at $\sim 20$-30 Tg sulfate aerosol in the G08 dataset for about a year, and drops off to $< 1$ Tg after 4-5 years.

Events fall into the SYMM category if they have less than a 25% difference in aerosol loading between hemispheres, while the ASYMM$_{NH}$ events have an at least 25% higher loading in the NH relative to the SH. The opposite applies to events falling into the ASYMM$_{SH}$ category. The dates for which these thresholds are satisfied are taken from the original G08 dataset (Table 1), and thus the CESM and GISS-E2 composites are formed using the same events despite the GISS-E2 mis-scaling and other differences in model implementation.

Results are reported for the boreal warm season (averaged over the MJJAS months) and cold season (NDJFM), except for annual-mean results in Figures 8-9, showing the progression of signals at monthly resolution (Figure S6, S9-S12 in the Supplement). For each eruption, we identify the post-volcanic response by averaging the number of consecutive seasons during which the above criteria are met, typically 1-3 years. All seasons for an eruption lasting longer than 1 year are first averaged together to avoid over-weighting its influence in the composite. Anomalies are with respect to the corresponding time of year during the five years prior to the eruption. For overlapping eruptions, the five years prior to the first eruption are used instead. This relatively short reference period allows creating composites that are unaffected by changes in the mean background state due to low-frequency climate change during the LM. Composites for the SYMM, ASYMM$_{NH}$, and ASYMM$_{SH}$ cases are then obtained for each season and model by averaging over all anomaly fields within the appropriate classification, including all ensemble
members. A two-sided Student’s t-test was applied to all composites in order to identify regions where the anomalous signal is significantly different (p < 0.05) from the mean background conditions.

In no case does the classification of a given eruption change over the duration of the event, with the exception of the largest eruption (Samalas, 1258 C.E.), which straddles the 25% asymmetry criterion (SYMM and ASYMM$\text{NH}_{\text{H}}$) throughout the years following the event. This eruption would project itself most strongly onto the symmetric composite but may reasonably be classified as ASYMM$\text{NH}_{\text{H}}$ due to the greater absolute aerosol loadings in the NH. Due to this ambiguity, we omit the Samalas event from our main results. We note that there are far more asymmetric eruptions during the LM based on our criteria than SYMM cases, most of which easily meet the two thresholds outlined above. Because of this, the classification assigned to each event is quite robust to slightly different criteria in defining the ratio (or differences) in hemispheric aerosol loading. Since the asymmetric composites are formed from a relatively large number of events, our results are insensitive to the addition or removal of individual eruptions that may be more ambiguous in their degree of asymmetry. However, the SYMM composites are formed from only a few events, and are therefore more sensitive to each of the individual eruptions that are included.

We stress that in this study we are agnostic concerning the actual location of individual LM eruptions. Although aerosols from high-latitude eruptions tend to be confined to the hemisphere in which the eruption occurs, tropical eruptions may also lead to an asymmetric aerosol forcing, as happened during the eruptions of El Chichón and Mt. Agung during the historical period. The timing, magnitude, and spatial footprint of LM eruptions are important
topics of research (see e.g., an updated reconstruction from Sigl et al., 2015), and our composite should strictly be interpreted as a self-consistent response to the imposed forcing in the model.

Similar approaches of stratifying volcanic events during the LM have only begun to emerge in the literature (e.g., Liu et al., 2016). Iles and Hegerl (2015) showed the CMIP5 multi-model mean precipitation response to a few post-1850 eruptions, emphasizing the spatial structure of the aerosols (see their supplementary Figure S14) but noted that it would be desirable for a greater sample of events in order to group by the location of the aerosol cloud. The LM provides an appropriate setting for this. Additionally, we add to these results by presenting a simulation of the water isotope distribution following different volcanic excursions. We emphasize that we are screening events by spatial structure and since different magnitude eruptions enter into the different composites, a quantitative comparison of the different event classifications (or the two models) is not our primary objective and would require a more controlled experiment. Instead, we are reporting on the different composite responses as they exist in current LM simulations, and highlight the emergent structure that arises from different choices in how eruptions are sorted, much of which is shown to be scalable to different eruption sizes and robust to choices of model implementation.

3.3 Results

3.3.1. Temperature, Precipitation and ENSO response

Figure 3.2 illustrates the composite temperature anomaly for each classification and season in the CESM model. In both the ASYMM$$_{NH}$$ and ASYMM$$_{SH}$$ cases, the hemisphere that is subjected to the strongest forcing is preferentially cooled. In the ASYMM$$_{NH}$$ results, the cooling
peaks over the Eurasian and North American continents. As expected, there tends to be a much larger response over land, as well as evidence of NH winter warming in the mid-to-high latitudes, a phenomenon previously highlighted in the literature and often associated with increased (decreased) pole-to-equator stratospheric (mid-tropospheric) temperature gradients (Figure S1) and a positive mode of the Arctic/North Atlantic Oscillation (Robock and Mao, 1992, 1995; Stenchikov et al., 2002; Shindell et al., 2004; Ortega et al., 2015). This effect is weak in the ASYMM$_{NH}$ composite, likely because the maximal radiative forcing is located in the NH, offsetting any dynamical response, but is present in the SYMM and ASYMM$_{SH}$ composites in both models (see Figure S2 for the GISS-E2 composite).

In the SH, cooling is muted by larger heat capacity associated with smaller land fraction, with weak responses over the Southern Ocean while still exhibiting statistically significant cooling in South America, South Africa, and Australia in all cases. In fact, the cooling in the ASYMM$_{SH}$ composites is largely confined to the tropics, in contrast to the polar amplified pattern that is common to most climate change experiments. The cooling in all categories is communicated vertically (Figure S1) and across the free tropical troposphere, suggesting AET toward the forced hemisphere (section 3.4) for asymmetric forcing.

The cooling in the GISS-E2 model (Figure S2), displays a very similar spatial structure to CESM in all categories but with much greater amplitude due to the larger forcing. We note that the composite-mean forcing is similar between the four asymmetric panels, but larger in the symmetric cases. In Figure 3.3, we show the hemispheric and global average temperature response for both models after normalizing each event by a common global aerosol mass excursion, thereby accounting for differences in the average forcing among the different eruptions. This is done to highlight spread associated with internal variability and model
differences, and assumes the response pattern scales linearly to global forcing, which is unlikely to be true across all events and for the two models. Nonetheless, the gross features of the hemispheric contrast and reduction in global-mean temperature are shared between both models.

The CESM precipitation response is shown in Figure 3.4 (Figure S3 for GISS-E2). For both the ASYMM$_{NH}$ and ASYMM$_{SH}$ cases, the ITCZ shows a robust displacement away from the forced hemisphere. The precipitation reduction in the SYMM composites is much less zonally coherent, instead featuring tropical-mean reductions in precipitation and a slight increase toward the subtropics (see also Iles et al., 2013; Iles and Hegerl, 2014). Despite global cooling and reduced global evaporation (not shown), the ITCZ shift in ASYMM$_{NH}$ and ASYMM$_{SH}$ tends to result in precipitation increases in the hemisphere that is least forced (Figure 3.5), since the hemispheric-mean precipitation signal is largely influenced by the ITCZ migration itself.

The ensemble spread in precipitation for a selected eruption (1762 C.E., NDJFM) is shown in Figure S4, corresponding to the Icelandic Laki aerosol loading (a large ASYMM$_{NH}$ event). We note that the Laki eruption in Iceland actually occurred in 1783 C.E., but is earlier in our composite due to an alignment error in the first version of the G08 dataset. Results are shown for the 1763 C.E. boreal winter only (the full composite also includes 1762, see Table 1; Figure S4 also reports the winter 1763 Niño 3.4 anomaly in surface temperature for each ensemble member, and therefore we restrict the anomalous precipitation field to the same season). The ITCZ shift away from the NH is fairly robust across the ensemble members, particularly in the Atlantic basin, although internal variability still leads to large differences in the spatial pattern of precipitation, notably in the central and eastern Pacific.

The monthly time-evolution of the composite temperature and precipitation responses for the ASYMM$_{NH}$ and ASYMM$_{SH}$ cases can be viewed in an animation (Figures S9-S12). The
global and hemispheric difference in aerosol loadings is also shown for each timestep (at monthly resolution) in the animations. When averaged over the individual eruptions within each classification, the global aerosol mass loading remains elevated above 8 Tg for nearly two years, coincident with the peak temperature and precipitation response that begins to dampen out gradually and relaxes back to pre-eruption noise levels after ~4-5 years. The seasonal migration of anomalous precipitation in the ITCZ domain occurs in nearly the same way as the meridional movement of climatological rainfall, highlighting important connections between the timing of the eruption relative to the seasonal cycle of rainfall at a given location.

In both CESM and GISS-E2, the ITCZ shift is approximately scalable to eruption size. For both models, we define a precipitation asymmetry index, PAi (Hwang and Frierson, 2013) in each season as the area-weighted NH tropical precipitation minus SH tropical precipitation (extending to 20° latitude) normalized by the model tropical-mean precipitation, i.e.,

\[
PAi = \frac{P_{\text{EQ}-20^\circ N} - P_{20^\circ S-\text{EQ}}}{P_{20^\circ S-20^\circ N}} \quad (1)
\]

Supplementary Figure S5 illustrates the relationship between PAi and the AOD gradient between hemispheres (AOD is inferred for the CESM model by dividing the aerosol loading by 75 Tg in each hemisphere, an approximate conversion factor to compare the results with GISS-E2). The mis-scaling in GISS-E2 results in a wider range of AOD gradients than occurs in CESM. Both models feature more tropical precipitation in the NH (SH) during boreal summer (winter) in their climatology, with more asymmetry in CESM during boreal summer. Interestingly, the most asymmetric events in GISS-E2 (those that result in equatorward precipitation movements) can be sufficient to produce more precipitation in the tropical winter
hemisphere, thus competing with the seasonal insolation cycle in determining the seasonal precipitation distribution.

The meridional ITCZ shift leads to a number of important tropical climate responses. For example, an intriguing feature of the temperature pattern in Figure 3.2 is the El Niño response that is unique to the ASYMM$_{NH}$ composites. This is unlikely to be a residual feature of unforced variability, since there are 288 events in the ASYMM$_{NH}$ composites (16 eruptions in Table 1, multiplied by 18 ensemble members), significantly more than in the other categories. The GISS-E2 temperature composite (Fig. S2) also features a relatively weak cooling for ASYMM$_{NH}$, despite the very large radiative forcing. The relationship between ENSO and volcanic eruptions has, historically, been quite complicated due to the problem of separating natural variability from the forced response, and due to a limited sample of historical eruptions where ENSO events were already underway prior to the eruption. Older studies have suggested that El Niño events may be more likely 1 to 2 years following a large eruption (e.g., Adams et al., 2003; Mann et al., 2005; Emile-Geay et al., 2008). Our findings are also consistent with recent results (Pausata et al., 2015) that found an El Niño tendency to arise from a Laki-like forcing (in that study, a sequence of aerosol pulses in the high latitudes that was confined to the NH extratropics), and the El Niño response in CESM LME to different expressions of volcanic forcing was recently explored in Stevenson et al. (2016). Pausata et al. (2015) attributed the El Niño development directly to a southward ITCZ displacement. Since low-level converging winds are weak in the vicinity of the ITCZ, a southward ITCZ displacement leads to weaker easterly winds (a westerly anomaly) across the central equatorial Pacific. This was shown for a different model (NorESM1-M) and experimental setup, but also emerges in the ASYMM$_{NH}$ composite results for CESM. Indeed, a composite anomaly of $\sim 0.5^\circ$C emerges over the Niño 3.4 domain, lasting up to two years.
(Figure S6) with peak anomalies in the first two boreal winters after an eruption. Consistent with the SST anomalies, a relaxation of the zonal winds and re-distribution of water mass across the Pacific Ocean can be observed in the ASYMM$_{NH}$ composite response (Figure S7).

Since the ITCZ shift is a consequence of differential aerosol loading, we argue that the El Niño tendency in CESM is a forced response in ASYMM$_{NH}$ but otherwise depends on the state of internal variability concurrent with a given eruption, as no such ENSO response is associated with the composite SYMM or ASYMM$_{SH}$ composites, although we note that El Niño does tend to develop in response to the Samalas eruption that was removed from our composite, and would strongly influence the interpretation of the SYMM results due to the few events sampled (not shown, though see Stevenson et al., 2016). However, we also caution that this version of CESM exhibits ENSO amplitudes much larger than observations, and also features strong El Niño events with amplitudes that are ~2 times larger than strong La Niña events even in non-eruption years. Therefore, we choose not to further explore the dependence of our results on ENSO phasing.

Because the ITCZ responds differently to the three eruption classifications, there are implications for best practices in assessing the skill of climate model output against proxy evidence. For example, Anchukaitis et al. (2010) noted discrepancies between well-validated tree-ring proxy reconstructions of eruption-induced drought in the Asian monsoon sector and the precipitation response following volcanic eruptions derived from the NCAR Climate System Model (CSM) 1.4 millennial simulation. However, we note that monsoonal rainfall responds differently to ASYMM$_{NH}$, ASYMM$_{SH}$, or SYMM events in both GISS-E2 and CESM. Figure S8 shows a histogram of boreal summer (MJJAS) Asian-Pacific rainfall anomalies for all events in both models. ASYMM$_{NH}$ and SYMM eruptions generally lead to reductions in rainfall over the
broad region averaged from 65°-150°E, 10°-40°N (see also the spatial patterns in Figure 3.4 for CESM and Figure S3 for GISS E2-R). Because of the southward ITCZ shift in ASYMM\textsubscript{NH}, the most pronounced precipitation reductions occur for events within this category. In contrast, for ASYMM\textsubscript{SH} events, the northward ITCZ shift and associated monsoon developments are such that precipitation changes are relatively muted, and often the anomalies are positive.

3.3.2. River outflow

An ITCZ shift away from the forced hemisphere will manifest itself in several other components of the tropical hydroclimate system that are important to consider from the standpoint of both impacts as well as the development of testable predictions. One such important component of the hydrologic cycle is global streamflow, a variable that is related to excessive or deficient precipitation over a catchment. Rivers are important for ecosystem integrity, agriculture, industry, power generation, and human consumption. Streamflow anomalies associated with volcanic forcing in observations and models have previously been documented for the historical period (Trenberth and Dai, 2007; Iles and Hegerl, 2015). Here, we discuss this variable in the context of our symmetric and asymmetric composites.

The hydrology module of the land-component of CESM simulates surface and subsurface fluxes of water, which serve as input into the CESM River Transport Model (RTM). The RTM was developed to route river runoff downstream to the ocean or marginal seas and enable closure of the hydrologic cycle (Oleson et al., 2010). The RTM is run on a finer grid (0.5° x 0.5°) than the atmospheric component of CESM.

Figure 3.6 shows the river discharge anomalies in our different forcing categories. The southward ITCZ shift in ASYMM\textsubscript{NH} results in enhanced discharge in central and southern South
America, especially in the southern Amazon and Parana River networks. These territories of South America, along with southern Africa and Australia are the primary regions where land precipitation increases in the tropics for \(\text{ASYMM}_{\text{NH}}\), and the river flow in these areas tends to increase. Our results are also consistent with Oman et al. (2006), who argue for a reduced Nile River level (northeastern Africa) following several large high northern latitude eruptions, including Laki and the Katmai (1912 C.E.) eruption. Their results were viewed through the lens of weakened African and Indian monsoons associated with reduced land-ocean temperature differences; our composite results suggest that regional precipitation reductions may also be part of a zonally coherent precipitation shift.

In \(\text{ASYMM}_{\text{SH}}\), the ITCZ moves northward, resulting in reduced river flux in the Amazon sector and increases (reduction) in the Niger of central/western Africa during boreal summer (boreal winter). Interestingly, the Nile flow is also reduced in this case, although to a lesser extent, despite very modest precipitation increases during MJJAS for a southern hemisphere biased aerosol forcing. There are also modest discharge increases in southern Asia. However, there is simply very little land in regions where northward ITCZ shifts result in enhanced precipitation, suggesting less opportunity for increases in discharge to a SH biased eruption. For the \(\text{SYMM}\) eruptions, river discharge is reduced nearly everywhere in the tropics, consistent with the precipitation reductions that occur (Figure 3.3). The response is weaker or even reversed in the subtropics, such as in southern South America, where precipitation tends to increase (Iles and Hegerl, 2015).

3.3.3. Water isotopic variability
Another important variable that integrates several aspects of the tropical climate system is the isotopic composition of precipitation. Here, we focus on the relative abundance of $^{1}H_2^{18}O$ versus the more abundant $^{1}H_2^{16}O$, commonly expressed as $\delta^{18}O$, such that:

$$
\delta^{18}O_p \equiv \left\{ VSMOW^{-1} \frac{O_{mp}^{18}}{O_{mp}^{16}} - 1 \right\} \times 1000
$$

where $O_{mp}^{18}$ and $O_{mp}^{16}$ are the moles of oxygen isotope in a sample, in our case precipitation (denoted by the subscript mp). Delta values are with respect to the isotopic ratio in a standard sample, the Vienna Standard Mean Ocean Water (VSMOW = \(2.005 \times 10^{-3}\)).

$\delta^{18}O_p$ is a variable that is directly obtained from many paleoclimate proxy records. Therefore, rather than relying on a conversion of the local isotope signal to some climate variable, the explicit simulation of isotopic variability is preferred for generating potentially falsifiable predictions concerning the imprint associated with asymmetric volcanic eruptions.

Indeed, $\delta^{18}O_p$ variability is the result of an interaction between multiple scales of motion in the atmosphere, the temperature of air in which the condensate was embedded, and exchange processes operating from source to sink of the parcel deposited at a site.

Water isotope tracers have been incorporated into the GISS-E2 model’s atmosphere, land surface, sea ice and ocean, and are advected and tracked through every stage of the hydrologic cycle. A fractionation factor is applied at each phase change and all freshwater fluxes are tagged isotopically. Stable isotope results from the lineage of GISS-E2 models have a long history of being tested against observations and proxy records (e.g., Vuille et al., 2003; Schmidt et al., 2007; LeGrande and Schmidt, 2008, 2009).
Figure 3.7 shows the $\delta^{18}O_p$ response in the GISS-E2 model. Seasonal calculations are weighted by the precipitation amount for each month, although changes in the seasonality of precipitation are not important in driving our results (not shown). The literature on mechanistic explanations for isotope variability has a rich history of being described by several “effects” such as a precipitation amount effect in deep convective regions or a temperature effect at high latitudes (Dansgaard, 1964; Araguás-Araguás et al., 2000), so named as to reflect the most important climatic driver of isotopic variability at a site or climate regime. Notably, $\delta^{18}O_p$ tends to be negatively correlated with precipitation amount in the deep tropics and positively correlated with temperature at high latitudes (see e.g., Hoffman and Heimann, 1997 for a review of mechanisms). However, isotope-climate relations are generally complex. In our experiments, the $\delta^{18}O_p$ spatial pattern in the tropics (Figure 3.7) exhibits a similar pattern to precipitation changes induced by the ITCZ shift (Figure S5 for GISS-E2), particularly over the ocean. The meridional movement of the ITCZ leads to an isotopic signal that is more positive (enriched in heavy isotopes) in the preferentially forced hemisphere. The hemisphere toward which the ITCZ is displaced on the other hand experiences increased tropical rainfall and a relative depletion of the heavy isotope (more negative $\delta^{18}O_p$). Thus, the paleoclimatic fingerprint of asymmetric volcanic eruptions is characterized by a tropical dipole pattern, with more positive (negative) $\delta^{18}O_p$ associated with reduced (increased) rainfall.

Over land, South America stands out as exhibiting a palette of isotopic patterns depending on forcing category and season. The South American monsoon system peaks in austral summer, and the largest precipitation reductions occur in ASYMM$_{SH}$ when the ITCZ moves northward. There is a dipole pattern, characterized by isotopic enrichment (depletion) in $^{18}O$ in the northern (southern) tropics of South America in ASYMM$_{NH}$ during NDJFM, while the opposite pattern
emerges in $\text{ASYMM}_{\text{SH}}$, both associated with Atlantic and east Pacific ITCZ displacements. During the austral winter, climatological South American precipitation peaks in the northern part of the continent, and precipitation in this region is reduced in both the SYMM and $\text{ASYMM}_{\text{SH}}$ composites, leading to a large increase in $\delta^{18}O_p$. This is consistent with recent results in Colose et al. (2016), who used the isotope-enabled GISS-E2 model to form a composite of all large (AOD > 0.1) LM tropical volcanic events based on the Crowley and Unterman (2013) dataset. The eruptions analyzed in that study were smaller in amplitude due to differences in the scaling during implementation, as well as the fact that G08 tends to have larger volcanic events in the original dataset to begin with. In regions where tropical South American precipitation does not exhibit very large changes, such as in the NDJFM SYMM composites, temperature may explain much of the isotopic response, again consistent with findings in Colose et al. (2016).

3.3.4. Atmospheric Energetics

The overarching purpose of this work was to consider the influence of asymmetric volcanic forcing on the energetic paradigm outlined in section 3.1. This framework of analyzing ITCZ shifts in the context of asymmetric forcing predicts a net AET anomaly toward the hemisphere that is preferentially forced by explosive volcanism, with anti-correlated dry and latent energy fluxes both contributing to drive the ITCZ away from the forced hemisphere. To examine this relationship in CESM, we first write a zonal-mean energy budget for the atmosphere (Trenberth, 1997; Donohoe and Battisti, 2013):
\[
\frac{1}{2\pi a^2 \cos \phi} \frac{\partial AET}{\partial \phi}
= ASR_{TOA} - OLR_{TOA} + SW_{sfc}^\uparrow - SW_{sfc}^\downarrow + LW_{sfc}^\uparrow - LW_{sfc}^\downarrow + LH_{sfc} + SH_{sfc}
+ L_f S_n - \frac{1}{g} \int_0^p \frac{\partial (c_p T + L_v q + k)}{\partial t} dp
\]

(3)

where \( ASR_{TOA} \) is the absorbed solar radiation, \( OLR_{TOA} \) is outgoing longwave radiation at the top of the atmosphere (TOA), \( SW_{sfc}^\uparrow \) is reflected surface shortwave radiation, \( SW_{sfc}^\downarrow \) is shortwave received by the surface (sfc), \( LW_{sfc}^\uparrow \) is longwave radiation emitted (or reflected) by the surface, \( LW_{sfc}^\downarrow \) is longwave radiation received by the surface, \( LH \) is the latent heat flux, \( SH \) is the sensible heat flux, \( S_n \) is snowfall rate, \( q \) is specific humidity, \( k \) is kinetic energy, \( \phi \) is latitude, \( a \) is the radius of the Earth, \( T \) is temperature, \( c_p \) is specific heat capacity, \( L_v \) and \( L_f \) are the latent heats of vaporization and fusion, \( p \) is pressure (\( p=p_s \) at the surface), and \( g \) is the acceleration due to gravity. All terms are defined positive into the atmosphere, and the subscripts denote top-of-atmosphere (TOA) or surface flux (sfc) diagnostics. Equation 3 effectively calculates MSE transport (section 3.1) as a residual of energy fluxes in the model.

The last term \( \frac{\partial}{\partial t} \) on the right side of equation 3 is the time-tendency term, representing storage of energy in the atmosphere (hereafter, \( \text{STOR}_L \) and \( \text{STOR}_D \) for latent and dry energy, respectively. The time-derivative is calculated using finite differencing of the monthly-mean fields. The term in the parentheses is the moist enthalpy, or MSE minus geopotential energy. The kinetic energy is calculated in this study but is several orders of magnitude smaller than other terms, and hereafter is folded into the definition of \( \text{STOR}_D \). The tendency term must vanish on timescales of several years or longer, but is important in our context. We explicitly write out the
snowfall term since CESM (and any CMIP5 model) does not include surface energy changes associated with snow melt over the ice-free ocean as part of the latent heat diagnostic, and must be calculated to close the model energy budget.

Integrating yields an expression for the atmospheric heat transport across a latitude circle:

\[
AET(\phi) = 2\pi a^2 \int_{\frac{\pi}{2}}^{\phi} (R_{TOA} + F_{sfc} - STOR_L - STOR_D) \cos \phi \, d\phi \quad (4)
\]

where we have combined the TOA terms into \(R_{TOA}\) and the snowfall and surface diagnostics have collapsed into a single variable \(F_{sfc}\). Similarly, the latent heat flux \(H_L\) across a latitude circle is:

\[
H_L(\phi) = 2\pi a^2 \int_{\frac{\pi}{2}}^{\phi} (LH_{sfc} - L_v P - STOR_L) \cos \phi \, d\phi \quad (5)
\]

where \(P\) is precipitation in kg m\(^{-2}\) s\(^{-1}\). We note that transport calculations are presented for CESM and were done for only 17 ensemble members, since there are missing output files for the requisite diagnostics in one run.

Figure 3.8a shows the annual-mean climatological northward heat transport in CESM, as performed by the atmosphere, in addition to the dry and moisture-related components of AET. The total CESM climatological poleward transport is in good agreement with observational estimates (e.g., Trenberth and Caron, 2001; Wunsch, 2005; Fasullo and Trenberth, 2008), peaking at ~5.0 PW and ~5.2 PW in the SH and NH subtropics, respectively (1 petawatt = 10\(^{15}\) W). In CESM, the SH receives slightly more net TOA solar radiation than the NH (by ~1.3 W m\(^{-2}\) in the annual-mean), and the NH loses slightly more net TOA longwave radiation to space (by ~0.89 W m\(^{-2}\)). However, the CESM annual ocean heat transport is northward across the equator.
(not shown), keeping the NH warmer than the SH by ~0.98 °C. As a consequence, AET is directed southward across the equator (red line). Moisture makes it more difficult for the tropical circulation to transport energy poleward, and the transport of moisture in the low-level equatorward flow is directed northward across the equator and associated with an annual-mean ITCZ approximately co-located with the atmospheric energy flux equator (EFE), the latitude where AET vanishes. This arrangement of the tropical climate is consistent with satellite and reanalysis results for the present climate (Frierson et al., 2013; Kang et al., 2014).

In response to asymmetric volcanic forcing, anomalous AET is directed toward the preferentially forced hemisphere (Figure 3.8b,c), along the imposed temperature gradient. Results are shown for the annual-mean AET anomaly in ASYMM$_{NH}$ and ASYMM$_{SH}$ for one year beginning with the January after each eruption, although averaging the first 2-3 years yields similar results with slightly smaller amplitudes. The equatorial AET (AET$_{eq}$) anomaly averaged over all events and ensemble members for ASYMM$_{NH}$ (ASYMM$_{SH}$) is approximately 0.08 (-0.06) PW, defined positive northward, with much larger near-compensating dry and latent components. The anomalous moisture convergence drives the ITCZ shift away from the forced hemisphere. Anomalies in AET$_{eq}$ when considering each unique volcanic event (after averaging over the 17 ensemble members) are strongly anti-correlated with changes in the energy flux equator ($r = -0.97$, not shown), the latitude where AET vanishes.

The change in cross-equatorial energy transport for the SYMM ensemble/eruption mean (not shown) does not exhibit the coherence of the asymmetric cases for either AET or the individual dry and moist components, and in all cases does not emerge from background internal variability.
Quantifying the ITCZ shift is non-trivial, since the precipitation field is less sharply defined than the EFE, and climate models (including the two discussed here) exhibit a bimodal tropical precipitation distribution (often called a “double-ITCZ”), often with one mode of higher amplitude in the NH (centered at 8°-9°N in CESM). However, despite pervasive biases that still exist in the climatology of tropical precipitation in CMIP5 (e.g., Oueslati and Bellon, 2015), the anomalous precipitation response is still characterized by a well-defined ITCZ shift (or a shift in the bimodal precipitation distribution, e.g., Figure 9 in Stevenson et al., 2016) and the gross features presented here are in agreement with theoretical considerations. In our analysis, a movement in the latitude of maximum precipitation is not found to be a persuasive indicator of our ITCZ shift. In fact, the meridional shift is better described as a movement in the center of mass of the precipitation distribution, including changes in the relative amplitude of the two modes (e.g., a heightening of the SH mode for a southward ITCZ shift). Different metrics to describe the shift in the center of mass have been presented in the literature (e.g., Frierson and Hwang, 2012; Donohoe et al., 2013; Adam et al., 2016).

Here, we first adopt the precipitation median \( \phi_{\text{med}} \) definition (e.g., Frierson and Hwang, 2012) defined as the latitude where area-weighted precipitation from 20°S to \( \phi_{\text{med}} \) equals the precipitation amount from \( \phi_{\text{med}} \) to 20°N, i.e., where the following is satisfied:

\[
\int_{20^\circ\text{S}}^{\phi_{\text{med}}} P \cos(\phi) \, d\phi = \int_{\phi_{\text{med}}}^{20^\circ\text{N}} P \cos(\phi) \, d\phi
\]  

(6)

When considering the spread across eruption size (regressing the different events in all three categories together after averaging over ensemble members) we find a movement of \( \sim -8.9^\circ \)
shift in ITCZ latitude per 1 PW of anomalous AET$_{eq}$ (Figure 3.9). The sign of this relationship is a robust property of the present climate system, although it is higher than other estimates (Donohoe et al., 2013) that analyzed the ITCZ scaling with AET$_{eq}$ to a number of other time periods and forcing mechanisms (not volcanic), including the seasonal cycle, CO$_2$ doubling, Last Glacial Maximum, and mid-Holocene. It was argued in that paper that the ITCZ is “stiff” in the sense that a large AET$_{eq}$ is required to move the ITCZ. However, the sensitivity of this relationship may vary considerably depending on ITCZ metric considered (Figure 3.9 presents a scaling with different indices), based on the following equation (Adam et al., 2016):

$$\phi_{ITCZ} = \frac{\int_{20^\circ S}^{20^\circ N} \phi (P \cos(\phi))^N d\phi}{\int_{20^\circ S}^{20^\circ N} (P \cos(\phi))^N d\phi}$$  \hspace{1cm} (7)

Here, N controls the weighting given to the modes in the precipitation distribution. Typically $\phi_{ITCZ}$ moves toward the precipitation maximum as $N$ increases, but importantly, the sensitivity of a $\phi_{ITCZ}$ migration to a given anomaly in AET$_{eq}$ also changes. Figure 3.9 shows the regression of anomalous $\phi_{med}$ and $\phi_{ITCZ}$ ($N = 5$) against anomalous AET$_{eq}$ ($r = -0.94$). $\phi_{ITCZ}$ ($N = 3$) yields a high correlation ($r = -0.95$) and best follows a 1:1 line with the EFE (Figure 3.9, bottomleft). The slope of the relationship between ITCZ location and AET$_{eq}$ may vary by a factor of 4-5 depending on the relationship used. For example, there is approximately a $-11.7^\circ$ shift in ITCZ latitude per 1 PW of anomalous AET$_{eq}$ using $\phi_{ITCZ}$ ($N = 3$). Thus, we interpret our results as suggesting that energetically, it is not necessarily difficult to move the ITCZ, and urge caution in characterizing past ITCZ shifts as being difficult to reconcile with paleo-forcing estimates (Donohoe et al., 2013). Indeed, as many studies have used a
“precipitation centroid” or a similar variant to quantify tropical precipitation migrations, we recommend exploring the sensitivity of ITCZ shifts to different ways of characterizing the movement in precipitation mass unless the community can agree upon a well-defined “N” that suitably characterizes the precipitation distribution in both climate models and observations.

3.4. Conclusions

In this work, we have examined two models, NASA GISS ModelE2-R and the recently completed CESM Last Millennium Ensemble, and stratified volcanic events by their degree of asymmetry between hemispheres. We find a robust ITCZ shift away from the preferentially forced hemisphere, as a consequence of adjustments in the Hadley circulation that transports anomalous energy into the cooled hemisphere.

An important component of our work was using the GISS-E2 model to explicitly simulate the oxygen isotopic imprint following major volcanic eruptions with asymmetric aerosol forcing. The ITCZ shift following asymmetric forcing leads to a more positive isotopic signal in the tropical regions the ITCZ migrates away from, and a relative depletion in heavy isotopes in regions the ITCZ migrates to. These results provide a framework for the search of asymmetric volcanic signals in high-resolution isotopic or other temperature and precipitation sensitive proxy data from the tropics.

There is still considerably uncertainty in the timing and magnitude of LM eruptions. Improvements in particle size representation have been identified as critical target for improved modeling and comparisons to proxy data (e.g., G. Mann et al., 2015). Here, we argue that the inter-hemispheric asymmetry of the aerosol forcing also emerges as being of first-order
importance for the expected volcanic response. Future developments in model-proxy comparisons should probe the uncertainty space not just in the global-mean radiative forcing and coincident internal variability at the time of the eruption, but also the spatial structure of the aerosol cloud. For example, simulations that represent volcanic forcing simply as an equivalent reduction in total solar irradiance at the TOA are unrealistic and cannot be expected to be faithful to tropical climate proxy records.

We hope this contribution will help motivate the connection between the spatial structure of volcanic episodes and the expression on tropical hydroclimate as an urgent paleoclimate target in future studies and model intercomparisons. Such investigation also calls for high-resolution and accurately dated tropical proxy networks that reach across hemispheres. Developments in seasonally and annually resolved volcanic reconstructions from both hemispheres (Sigl et al., 2015) are of considerable importance in such assessments. Future modeling efforts that are forced with the explicit injection of volcanic species, while also probing multiple realizations of internal variability that will dictate the spatio-temporal evolution of the volcanic aerosol, are also urgently required as a tool for understanding both past and future volcanic impacts.

Acknowledgments

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Supplementary information: A link to the supplementary information for this chapter may be found at: http://www.earth-syst-dynam.net/7/681/2016/esd-7-681-2016-supplement.pdf
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**Figure 3.1.** Global Aerosol Loading (Tg) from Gao et al. (2008) in red line. \( \text{ASYMM}_{\text{NH}} \) (green circles), \( \text{ASYMM}_{\text{SH}} \) (blue circles), and SYMM (black circles) events that are used in composites are shown. Note that Samalas is omitted, as discussed in text. The time-series is at seasonal (five-month) resolution and thus multiple points may be associated with a single eruption. The hemispheric contrast (NH minus SH) clear-sky net solar radiation (FSNTC− in W/m\(^2\)) in CESM LME is shown in orange (offset to have zero mean).
Figure 3.2. CESM spatial composite of surface temperature anomaly (°C) for (top row) $\text{ASYMM}_{\text{NH}}$, (middle row) $\text{ASYMM}_{\text{SH}}$, and (bottom row) SYMM events, each in (left column) NDJFM and (right column) MJIAS. Stippling indicates statistical significance using a two-sided student’s t-test ($p < 0.05$).
**Figure 3.3.** Box-and-whisker diagrams showing the (red fill) global mean, (green fill) NH mean, and (blue fill) SH mean temperature anomaly in the ASYMM$_{NH}$, ASYMM$_{SH}$, and SYMM eruption cases on vertical axis. All events are normalized by a 20 Tg global loading size. For GISS-E2, loadings were multiplied by a factor of two to approximately account for the over-inflated forcing prior to analysis. Results are shown for the CESM and GISS-E2 model and for NDJFM and MJJAS, as labeled. Black solid line indicates the median, box width spans the 25-75% quartiles, and tails span the full interval for all cases. N=the number of events used in each category (consistent with the number of listed events in Table 1, multiplied by 18 ensemble members for CESM and 3 ensemble members for GISS-E2). Bottom panels (CTRL) show the spread of 100 randomly selected and non-overlapping events averaged over two seasons (relative to the previous five seasons) in a control run.
Figure 3.4. As in Figure 3.2, except for precipitation (mm/day).
Figure 3.5. As in Figure 3.3, except for precipitation (mm/day, normalized to 20 Tg in the forced simulations; mm/day in the control). N (not shown) is the same as in Figure 3.3.
Figure 3.6. As in Figures 3.2 and 3.4, except for river discharge (m$^3$/s, or 10$^6$ Sverdrups).
Figure 3.7. GISS-E2 spatial composite of the oxygen isotope anomaly (per mil) in (top row) ASYMM$_{NH}$, (middle row) ASYMM$_{SH}$, and (bottom row) SYMM events in (left column) NDJFM and (right column) MJJAS.
Figure 3.8. a) CESM climatology of atmospheric energy transport (PW, black), dry (red), and latent (dark blue) transports. b) Composite mean anomaly in atmospheric heat transport for ASYMM\textsubscript{NH} eruptions in total (black), dry (red), and latent (blue) components. Lighter (orange and aqua) lines represent individual eruptions, each averaged over 17 ensemble members. c) As in (b), except for ASYMM\textsubscript{SH} eruptions. Grey envelope corresponds to the total AET anomaly vs. latitude in a control simulation using 50 realizations of a 17-event composite (17 “events” with no external forcing, corresponding to the size of the ensemble). Vertical bars correspond to the range of (aqua) latent and (orange) dry components of cross-equatorial energy transport (AET\textsubscript{eq}) in the control composite.
Figure 3.9. Annual-mean ITCZ shift represented by changes in (topleft) $\phi_{med}$ and (topright) $\phi_{ITCZ}$ ($N = 5$) vs. change in AET$_{eq}$. Changes in $\phi_{ITCZ}$ ($N = 3$) vs. change in EFE (bottomleft). See text for definitions. Total AET vs. latitude for a small band centered around the equator for all volcanic events in (green) ASYMM$_{NH}$, (blue) ASYMM$_{SH}$, and (black) SYMM cases (bottomright). Black dashed line indicates climatological or pre-eruption AET values (different choices are indistinguishable). Colored arrows represent the direction of anomalous AET$_{eq}$. 
<table>
<thead>
<tr>
<th>Eruption Category</th>
<th>Seasons in LM Composite (MJJAS)</th>
<th>Seasons in LM Composite (NDJFM)</th>
</tr>
</thead>
<tbody>
<tr>
<td>ASYMM&lt;sub&gt;NH&lt;/sub&gt;</td>
<td>870, 901, 933/934, 1081, 1176/1177, 1213/1214, 1328, 1459, 1476, 1584, 1600/1601, 1641/1642, 1719/1720, 1762/1763, 1831, 1835/1836</td>
<td>871, 902, 934, 1082, 1177, 1214/1215, 1329, 1460, 1585, 1601, 1641/1642, 1720, 1730, 1762/1763, 1832, 1835/1836</td>
</tr>
<tr>
<td>ASYMM&lt;sub&gt;SH&lt;/sub&gt;</td>
<td>929, 961, 1158.5/1159.5, 1232, 1268, 1275/1276, 1341/1342, 1452/1453, 1593, 1673, 1693/1694</td>
<td>962, 1159, 1233, 1269, 1276/1277, 1285, 1342, 1453/1454, 1674, 1694</td>
</tr>
<tr>
<td>SYMM</td>
<td>854, 1001, 1284/1285, 1416, 1809/1810, 1815/1816</td>
<td>855, 1002, 1810, 1816/1817</td>
</tr>
</tbody>
</table>

2) Dates of Eruption events used in composite results, based on reconstructed stratospheric sulfate loadings from Gao et al. (2008).

3) Combined dates with a “/” indicate a multi-season event where every inclusive month is first averaged prior to entering the multi-eruption composite.
Chapter 4

The Role of Top-of-Atmosphere Radiative Responses on

ITCZ migrations for Hemispherically Symmetric and

Asymmetric Volcanic Forcing
Abstract

Results from the three volcanic classifications in chapter 3, in which volcanic forcing is preferentially focused in the northern hemisphere (NH), southern hemisphere (SH), or hemispherically symmetric, are further investigated. Here, I use the Approximate Partial Radiative Perturbation and radiative kernel technique to decompose the top-of-atmosphere (TOA) flux changes into contributions from the imposed aerosol forcing and that due to emergent feedbacks in the climate system. The hemispherically asymmetric component of the feedbacks is interpreted to reinforce or dampen the ITCZ migration. I show that in the CESM Last Millennium Ensemble, water vapor and temperature feedbacks reinforce and dampen (respectively) the TOA energy asymmetry of roughly equal magnitude, while clouds dampen the ITCZ shift by enhancing shortwave (SW) reflection away from the hemisphere forced more strongly by aerosols, and enhancing SW absorption in the hemisphere forced more strongly. This effect is stronger than the oppositely signed longwave (LW) cloud contribution in the extratropics, suggesting that in this model the net cloud effect resists the imposed aerosol forcing.

4.1. Introduction

The factors controlling the tropical precipitation distribution have been a subject of study for many decades. Historically, older work highlighted the role of tropical SST gradients and land surface properties; notably, when SST is specified as a boundary condition, the distribution of SST has been shown to be skillful in reconstructing the pressure field, low-level convergence, and associated vertical motion in the tropics (Lindzen and Nigam, 1987; Back and Bretherton,
Theories for the monsoon location often follow from this perspective and predict that the ITCZ lies close to the maximum near-surface moist static energy (Privé and Plumb, 2007), which is close to the maximum in SST over the oceans.

As highlighted in chapter 3, however, recognition that remote high-latitude thermal forcings could result in substantial shifts in the ITCZ has laid the foundation for an alternative (but not necessarily incompatible) perspective, one that focuses on the vertically integrated atmospheric energy budget (Kang et al., 2008; 2009); in this paradigm, heating of one hemisphere’s atmosphere relative to the other affects the ITCZ position by compelling the atmosphere to transport some fraction of the heating imbalance across the equator. A corollary is that ITCZ shifts are simply a consequence of the anomalous mass flux in the Hadley cell domain and associated meridional energy fluxes that are required in order to balance remote forcings. Within this framework, all factors that affect atmospheric energy transports, including surface albedo, water vapor, and cloud feedbacks, are important factors determining the zonal-mean precipitation migration. In fact, no direct information concerning SST changes are required to predict the precipitation response.

The energetic paradigm has origins in idealized models (e.g., aquaplanet configurations) that make use of a slab ocean as the lower boundary condition (Kang et al., 2008) ensuring closure of the surface energy budget. Since the net surface fluxes are prescribed in such setups, the atmospheric energy budget and tropical precipitation response becomes anchored to the TOA radiative fluxes and associated atmospheric energy transports, which has been found to be especially sensitive to cloud feedbacks (see also Voigt et al., 2014).

In this chapter, I will build upon the results of chapter 3 using the energetic paradigm as the vehicle in which I explore the response of a fully coupled model when subjected to volcanic
forcing of different spatial structures. The goal is to undertake an investigation of the radiative forcing and feedback patterns underpinning the results of the previous chapter. It is now well recognized that the magnitude of surface temperature change may depend greatly on the meridional structure of a climate forcing (e.g., Rose et al., 2014). Usually, diagnosing feedbacks is done in the context of exploring the climate sensitivity problem, although here I focus on how the pattern of heating or cooling may alter the dynamics of the tropical atmosphere and the resulting hydroclimate pattern.

In chapter 3, it was demonstrated that the response of the ITCZ differs dramatically for aerosol loading centered in the Northern vs. Southern Hemispheres, with the ITCZ shifting away from the hemisphere with the greater concentration of aerosols. These results are summarized in Figure 4.1 (all results shown from NCAR’s CESM Last Millennium Ensemble, as discussed in next section). Notably, for the asymmetric eruptions, the anomalous Hadley cell transports mass toward the energetically deficient hemisphere (Figure 4.1a and 4.1b). Consequently, there is anomalous atmospheric energy transport across the equator (AHT$_{eq}$) that acts to resist the asymmetric forcing and the corresponding ITCZ shift increases (decreases) precipitation away from (toward) the preferential aerosol loading. I note that the ratio of energy flux to mass flux throughout the Hadley circulation domain (sometimes called the ‘gross moist stability’, e.g., Neelin and Held, 1987; Hill et al., 2015), effectively a measure of the efficiency of meridional energy transport by the Hadley cells, exhibits minimal change throughout all experiments. In principle, the energetic paradigm for ITCZ migrations could break down if changes in the efficiency of the circulation at transporting energy produce a change in energy transports independently of the change in circulation, or vice versa, although this does not appear to be an
important consideration in this work. As shown in Figure 4.1d, the ITCZ shift is closely related to $\Delta AHT_{eq} (r = -0.93$ in the two year mean for each quantity).

The asymmetric response in precipitation (Figure 4.1c) suggests that low-level moisture convergence/divergence is more important than diminished large-scale evaporation (not shown) in determining the precipitation pattern, as is typically the case in climatology across the tropics (e.g., Neelin and Held, 1987). For the symmetric eruption case, there is virtually no ITCZ shift and a more uniform reduction in precipitation that arises from a reduction in net TOA energy input, and the demand for precipitation to closely balance evaporation globally.

A natural extension in further understanding this phenomenon is to also understand the meridional structure of the resulting feedback pattern that develops as a response to the symmetric/asymmetric aerosol forcing. For example, a shift in clouds away from the volcanically forced hemisphere may enhance SW reflection in the opposite hemisphere (away from the largest aerosol forcing), suggesting that the ITCZ migration might act as a negative feedback on its own movement. Furthermore, the induced hemispheric asymmetry by the aerosol forcing might be compensated locally in each hemisphere by changes in the other energy budget components, including locally decreased LW emission to space in the more strongly cooled hemisphere. However, pure local compensation is difficult to achieve due to meridional energy transport and the fact that the free tropical troposphere strongly resists large horizontal temperature gradients (Sobel et al., 2001).

In this work, I report the differences in feedback structure to northern, southern, and symmetric volcanic classifications. This seeks to understand how different patterns of forcing, temperature change, and the adjustments to the tropical circulation conspire to alter the emergent feedback structure.
4.2. Methods

The model used in this analysis is the CESM Last Millennium Ensemble (hereafter, LME), the details of which are described in chapter 3. I restrict my analysis to the five volcano-only ensemble members, to ensure that no other radiative perturbations are present in the calculations. Results are reported for the ASYMM$_{NH}$, ASYMM$_{SH}$, and SYMM volcanic classifications, adopting the same criteria as in chapter 3 (with the exception that one volcano was removed in the SYMM composite since it straddled the criteria to be a southern volcano, and the results here are averaged over a different time interval resulting in a change in classification). In chapter 3, every volcanic event must be associated with at least an 8 Tg aerosol loading from the Gao et al. (2008) dataset. Additionally, asymmetric events are ones in which there is at least a 25% aerosol mass difference between hemispheres. Otherwise, the event is classified as symmetric. Within a given simulation, there are 17, 12, and 5 volcanic excursions included in ASYMM$_{NH}$, ASYMM$_{SH}$, and SYMM, respectively (hence, given five ensemble members, there are 85, 60, and 25 events in these classifications). For each event, I define the post-volcanic response as the climate state averaged over two years beginning from the month of peak aerosol loading. The pre-volcanic climate is the five years prior to any substantial aerosol mass excursion. For overlapping eruptions, the five years prior to the first event are used. For each event, a difference is calculated from the post-eruption and pre-eruption intervals, and a composite is formed for the three volcanic classifications.

To diagnose the SW cloud feedback and aerosol forcing, I use the Approximate Partial Radiative Perturbation (APRP) method (Taylor et al., 2007). The APRP method makes use of a
simple shortwave layer model of the atmosphere and aims to decompose the net TOA solar radiation change (between an unperturbed and perturbed state) into individual contributions from surface albedo, clouds, noncloud scattering, and noncloud shortwave absorption. In this work, the change in clear-sky scattering is assumed to be equivalent to the volcanic shortwave forcing.

The change in the net absorbed solar radiation (ASR) at the TOA [\(\text{ASR} = \text{S}(1-\text{A})\), where \(\text{S}\) and \(\text{A}\) are the incoming solar radiation and planetary albedo evaluated at each grid cell] is \(\delta\text{ASR} \approx \delta\text{S}(1-\text{A}) - \text{S}\delta\text{A}\). The first term on the right side is zero, since the simulations analyzed here feature fixed incoming solar radiation. The last term on the right hand side represents all the forcings and feedbacks within the climate system that can affect planetary albedo. It is negative since an increase in albedo reduces the ASR.

In the APRP method, the planetary albedo is described as being a function of several parameters.

\[
A = A(c, \alpha_{s,clr}, \alpha_{s,oc}, \mu_{clr}, \mu_{cld}, \gamma_{clr}, \gamma_{cld}) \quad (1)
\]

where \(c\) is the fraction of a region occupied by clouds, \(\alpha_s\) is the surface albedo (\(\alpha_{s,clr}\) for clear-sky and \(\alpha_{s,oc}\) overcast conditions), and \(1-\mu_{clr}\) and \(1-\mu_{cld}\) represent the absorptivity of the clear-sky and cloudy atmosphere, respectively. \(\gamma_{clr}\) and \(\gamma_{cld}\) are atmospheric scattering coefficients that are associated with the albedo of the clear-sky and cloudy atmosphere (the subscript cld refers to absorption or scattering from the cloud itself). Resolving a radiative flux \(R\) into clear-sky and overcast components (i.e., obtaining overcast values for \(\alpha, \mu,\) and \(\gamma\)) is part of the APRP calculation, and is obtained by the relation, \(R = cR_{oc} + (1-c)R_{clr}\), i.e., from the total and clear-sky fluxes, both of which are saved as part of the standard output of the LME simulations.

The seven parameters can be obtained using monthly-mean output from any target GCM (in this case, the LME). The APRP model simply requires diagnostics for total (vertically-
integrated) cloud cover, as well as all-sky and clear-sky solar fluxes at both the TOA and surface. The relatively small number of requisite variables makes the APRP method ideal for routine academic application, given that other methods [such as using offline radiative transfer calculations to substitute the perturbed values of a given variable, such as water vapor, while holding all other variables constant (from an unperturbed control and perturbed experiment)] can be computationally demanding. Taylor et al. (2007) showed that APRP is quite accurate when compared to more sophisticated methods, often with errors of only a few percent, although the error is higher over bright surfaces and depends on the forcing agent in consideration.

In the APRP method, the surface albedo, atmospheric scattering coefficient, and atmospheric absorption coefficient are tuned to ensure that the TOA and surface shortwave fluxes are consistent with those in the GCM being analyzed. These single layer model parameters are calculated for both the post-volcanic and pre-volcanic time periods. The parameters are then individually perturbed in the surrogate model to obtain individual contributions to the albedo change. The SW forcing or feedbacks by the clear-sky atmosphere and cloud are calculated by multiplying the downward SW radiation at the TOA by $-\Delta A_{\text{clr}}$ and $-\Delta A_{\text{cld}}$, respectively, and dividing by the globally averaged surface temperature anomaly to obtain a temperature-normalized forcing, in the case of aerosols, or a feedback, in the case of clouds (each in units W m$^{-2}$ K$^{-1}$).

Changes in atmospheric SW absorption (which is partially due to water vapor, as well as aerosols) as well as the surface albedo feedback are obtainable from the APRP method, although I calculate these terms with a different method (see below) due to simplifications in the APRP setup. For example, SW absorption only occurs on the first downward pass in the atmosphere and is assumed to occur above the level of atmospheric reflection, which is not appropriate for a
stratospheric imposed aerosol. I note that the APRP method cannot be applied to LW feedbacks, since they depend significantly on (for example) the vertical profiles of water vapor and temperature, and a one-layer model of the atmosphere is not appropriate for representing these terms. Finally, I note that the SW cloud feedback inferred from the APRP method does not discriminate between temperature-related responses and any temperature-independent rapid cloud adjustments in the troposphere.

To calculate the LW feedbacks, as well as the SW albedo and SW water vapor terms, I make use of the radiative kernel technique (e.g., Shell et al., 2008; Soden et al., 2008). The power of the radiative kernel technique is that for a given feedback \( x \) in response to a climate transition \( S_1 \) to \( S_2 \), instead of substituting \( x_{S_1} \) for \( x_{S_2} \) individually (holding all other forcing or feedback terms constant) and performing offline radiative transfer calculations to isolate effect of the term of interest, it has been recognized that the change in TOA flux in response to a standard differential change \( \Delta x \) does not typically exhibit large variations among the radiative transfer routines of different GCMs. For example, in 2xCO\(_2\) experiments, the variation in specific humidity anomalies in different GCM’s is usually larger than the radiative flux change in response to a given change in specific humidity. Therefore, it is useful to separate feedbacks into the product of the change in climate components in response to an imposed forcing and the radiative kernel, the effect that climate changes have on the TOA radiative budget. Mathematically,

\[
dG \approx F + \sum_j \frac{\partial G}{\partial x_j(T)} \frac{\partial x_j(T)}{\partial T} dT = F + \sum_j K_j \frac{\partial x_j(T)}{\partial T} dT \tag{2}
\]
where, \( G \) is the net TOA energy flux (ASR minus OLR) which is zero in equilibrium, \( \mathcal{F} \) is the original forcing applied to the climate system, \( T \) is temperature, and \( x_j \) is the \( j \)'th climate variable that itself is a function of \( T \), which could be surface albedo or water vapor changes, for example. Here, \( K \) is the radiative kernel, which can be computed once and then provided to the community for a wide array of climate change applications.

The kernels used in this analysis are from Shell et al. (2008) and are based on the CAM3 present day climate state. Available 2D kernels include a surface skin temperature kernel (LW) and surface albedo (SW) kernel (each a function of latitude, longitude, and month of year); the available 3D kernels (that additionally include a pressure dimension) include atmospheric temperature (LW) and water vapor (both SW and LW) kernels. All kernels are provided for both all-sky and clear-sky conditions. I note that the LME uses the CAM5 atmospheric model, but CAM5 kernels are not yet publicly available. Small differences in the base climate state between CAM3 and CAM5 (especially associated with cloud masking) introduce uncertainty in this analysis. Furthermore, the LW component of aerosol forcing is not accounted for.

The method used to derive the kernels was to run the CAM3 radiation code off-line with small perturbations in surface and atmospheric temperature (1 K at each level), specific humidity (moistening that would occur in response to a 1 K warming everywhere, keeping relative humidity fixed), and a change in surface albedo of 1%. These are the differential changes \( \partial x_j' \) used to derive the radiative kernel output. The variable used for calculating the standard anomaly for water vapor is the change in logarithm of specific humidity, \( \ln(q) \), which is calculated for the LME at each point using the Clausius-Clapeyron equation assuming fixed relative humidity.

The implied contribution of the TOA flux change from a given feedback is obtained from the convolution of the kernel with the actual climate response from the LME output (equation 2).
Calculating feedbacks for the 3D fields requires vertical integration of the product at each pressure level from the surface to the tropopause, which is assumed to vary linearly from 100 hPa in the tropics to 300 hPa near the poles (Soden et al., 2008). Feedbacks (represented by $\lambda$) are obtained by normalizing by the global mean surface temperature anomaly between the post-eruption and pre-eruption state, as is often done in practice (alternatively, one could normalize by the local temperature change). When presenting normalized results (in units of W m$^{-2}$ K$^{-1}$) only, volcanic events in which the global temperature does not change by at least 0.1°C between the pre-eruption and post-eruption interval are omitted from the composite. This is done to exclude spuriously large feedbacks from the mean results. In these results, the total number of events is reduced to 60, 44, and 21 events for the ASYMM$_{NH}$, ASYMM$_{SH}$, and SYMM groups, respectively.

In this analysis, I report separately the Planck feedback, which is that due to a vertically uniform change in temperature equivalent to the surface value, and the lapse rate feedback, which is due to deviations from vertically uniform temperature change. The sum of these is the total atmospheric temperature feedback, which provides a fundamental stabilizing tendency for Earth’s climate. Alternative decompositions are also valid.

Finally, the LW cloud feedback still must be accounted for. Cloud kernels are typically not provided due to strong non-linearities associated with cloud masking effects. The simplest approach to calculating cloud feedbacks is to calculate the change in LW cloud radiative forcing ($\Delta$CRF$_{LW}$, where CRF$_{LW}$ is the difference between the all-sky and clear-sky net TOA LW flux at each grid point). $\Delta$CRF$_{LW}$ (evaluated as the difference from an unperturbed and perturbed climate state), therefore, may act as a useful proxy for the role of changing clouds in altering LW energy flows. However, even in the limit of unchanged cloud cover and cloud properties,
$\Delta CRF_{LW}$ will generally be non-zero, since changes in clear-sky parameters affect the calculation. Soden et al. (2008) offered a corrected CRF method that acts as a better indicator of the cloud feedback ($\lambda_{LW}$):

$$\lambda_{LW} = \frac{\Delta CRF_{LW}}{\Delta T_s} - \left[ \left( \lambda_T - \lambda_{T,cltr} \right) + \left( \lambda_{ln(q),LW} - \lambda_{ln(q),cltr,LW} \right) \right] \quad (3)$$

where the all-sky and clear-sky temperature feedbacks include both the Planck (atmosphere and surface) and lapse rate component. Equation 3 corrects for clear-sky contributions that make $\Delta CRF_{LW}$ a biased cloud feedback estimator.

The above method would also be applicable for estimating the SW cloud feedback (using $CRF_{SW}$ and appropriate shortwave feedbacks for the correction term) in the absence of a significant SW forcing term. For reference, I show in Figure 4.2 the meridional pattern of climate feedbacks in a 2xCO$_2$ experiment\textsuperscript{2}, where the SW forcing is minimal, calculated with the kernel method. The SW cloud feedback is shown using both the APRP and corrected $CRF_{SW}$ approach, to demonstrate the coherence of the two methods. However, only the APRP method is used for evaluating the SW cloud feedback in the volcanic experiments.

4.3. Results

4.3.1. Feedback Structure

\textsuperscript{2} The results in Figure 4.2 use a doubled CO$_2$ and control experiment from the low-resolution NCAR CCSM3 model (Yeager et al., 2006). Although strictly a different model than the LME, the primary purpose of this plot is illustrative. Furthermore, the zonal-mean structure of the 2xCO$_2$ feedbacks is similar amongst different models; this includes the rather uncertain cloud responses, which for the SW clouds is typically positive in the subtropics and negative toward the poles (Ceppi et al., 2016) and the LW component of the cloud feedback is positive in the tropics (Zelinka and Hartmann, 2010). See also Soden et al. (2008) for results from a GFDL model.
Figure 4.3 shows the meridional structure of the volcanic forcing and individual feedbacks, in the LME volcanic experiments. The spatial pattern of the volcanic forcing simply reflects the prescribed stratospheric forcing in the LME (Gao et al., 2008). It is notable that although the two year composite-mean, global-mean forcing differs among the different volcanic classifications (-1.7 W m\(^{-2}\), -2.1 W m\(^{-2}\), and -2.6 W m\(^{-2}\) for the ASYMM\(_{NH}\), ASYMM\(_{SH}\), and SYMM categories, respectively) this is simply a consequence of relative differences in the global aerosol mass excursion on the LME (which also features a single particle size); however, the regression slope relating the forcing to the aerosol mass is similar in all cases (the global mean forcing changing by about -0.12-0.13 Wm\(^{-2}\) per additional teragram of global aerosol mass), despite very different distributions of the aerosol. This slope is consistent with scalings provided from previous work (e.g., Hansen et al., 2002; Schmidt et al., 2011). For example, the last millennium volcanic forcing time-series provided to the community by Schmidt et al. (2011) is derived from the equation \( F_{volc} \sim -20 \frac{M_{volc}}{150} \), where \( M_{volc} \) is the global mass of the aerosol loading in teragrams. Although the particle size is prescribed in the LME and the above scaling is expected to be valid in the visible range, it is not self-evident that the slope relating the forcing to the prescribed aerosol loading would be nearly identical when aerosols are concentrated in different latitude bands, due to the decline in incoming SW energy moving poleward.

The zonal-mean feedback response for non-cloud variables in the SYMM category is similar to the 2xCO\(_2\) pattern. The most pronounced positive feedback that amplifies the temperature change (warming for 2xCO\(_2\), cooling for volcanoes) is the water vapor feedback, which peaks in the tropics and is rather symmetric about the equator. In both climate change examples, the lapse rate feedback partly compensates the water vapor feedback due to the tropical troposphere staying close to a moist adiabatic temperature profile. In the higher
latitudes, the lapse rate feedback tends to be positive since the surface exhibits a more pronounced temperature anomaly than the upper troposphere. Indeed, the polar amplified pattern in 2xCO$_2$ is not well communicated vertically, and the pole-to-equator temperature gradient increases in the upper troposphere, unlike near the surface.

The Planck feedback increasing in magnitude poleward in the 2xCO$_2$ experiment is caused by the polar amplified warming pattern, rather than by the spatial structure of the temperature radiative kernel (note that normalizing feedbacks by local temperature rather than global-mean temperature would smooth part of the structure in the purple line of Figure 4.2). In the SYMM composite, the surface temperature anomalies are not strongly amplified at high latitudes relative to the tropics, but the NH cools more strongly than the SH, resulting in asymmetry of the SYMM Planck feedback.

The Planck feedback is not expected to be strongly asymmetric about the equator in the tropics, due to homogenization of free tropospheric tropical temperatures by wave propagation (Sobel et al., 2001) and the anomalous vertical motion in the Hadley cell. Indeed, even for asymmetric volcanic classifications, the hemispheric contrast in temperature anomalies is small in the deep tropics (Figure 4.4). Strictly speaking, the low latitude surface temperature field may be asymmetric, leading to asymmetry in the tropical component of the Planck feedback; this is especially the case since the surface radiates like a blackbody in the infrared (unlike most individual atmospheric levels) and thus contributes a significant fraction to the total TOA flux despite being located at a low emitting height. Nonetheless, I find that the resulting asymmetry in the tropical Planck feedback is small in all classifications. However, extratropical differences in temperature anomalies are strong (Figure 4.4), with the hemisphere of greater aerosol loading cooling significantly. For the ASYMM$_{SH}$ volcanoes, the temperature anomalies are tropically
amplified (see Figure 3.2 in Chapter 3), which was interpreted to be a consequence of low aerosol loadings in the NH and high heat capacity in the SH high latitudes. This pattern leads to a tropically amplified pattern in the Planck feedback, unlike for the 2xCO₂ or ASYMM₉H cases.

The spatial structures of the water vapor anomalies differ substantially for the different classifications (Figure 4.5) leading to different patterns in the water vapor feedback (Figure 4.3), with the strongest water vapor anomalies shifted toward the hemisphere of stronger cooling. Due to the tropically amplified ASYMM₉H pattern, the water vapor anomalies are strong in the deep tropics for SH volcanoes. This is especially pronounced in the east Pacific (Figure 4.6 shows the horizontal spatial pattern of all forcings and feedbacks). In contrast, El Niño events are favored in the ASYMM₉H composite, a state that the ITCZ shift itself may help bring about (as argued in chapter 3; see also Pausata et al., 2016) due to the relaxation of zonal winds in the tropical Pacific that accompanies a southward ITCZ migration. This results in relatively little temperature change (or warming) in the equatorial Pacific relative to ASYMM₉H events.

The global water vapor feedback is strongest in the ASYMM₉H and SYMM composite (2.04 W m⁻² K⁻¹ and 1.96 W m⁻² K⁻¹, respectively) and smaller in the ASYMM₉H composite (1.64 W m⁻² K⁻¹). The vertically-integrated precipitable water anomalies are also somewhat smaller than is typical of global Clausius-Clapeyron scaling (Held and Soden, 2006) at ~5 % K⁻¹ in the ASYMM₉H case, however exhibits substantial variability among the individual volcanic events within a given classification. The statistical robustness of this result and a detailed consideration of the factors determining the different vapor distributions is beyond the scope of this work (see e.g., Rose and Rencurrel, 2016 for a discussion of the subtleties in temperature-humidity scalings), but the different patterns of humidity changes is potentially relevant for the sensitivity of the system to different volcanic forcing, especially since NH volcanoes in this composite exert
relatively little influence on tropical temperature change and the local ‘activation’ of feedbacks
there. However, I do emphasize that the water vapor feedback is notably asymmetric for
ASYMM\textsubscript{NH} and ASYMM\textsubscript{SH}, promoting cooling in the hemisphere forced more strongly by the
aerosols.

The surface albedo feedback is weak in all cases, but strongest in the ASYMM\textsubscript{NH}
composite where large northern eruptions result in small increases in sea ice and snow cover (not
shown). The SH surface albedo is resilient to change for all three eruption types.

The SW (LW) cloud feedback for SYMM eruptions is negative (positive) at most latitudes
(Figure 4.3) due to decreases in vertically-integrated cloud cover (not shown), especially low
clouds. For the asymmetric eruptions, the cloud feedback largely follows the pattern of the ITCZ
shift itself in the tropics. SW reflection increases, moving in parallel with the ITCZ cloud cover
(a positive feedback on volcanic cooling) but allows less LW emission to space (a negative
feedback on volcanic cooling). Figure 4.3 and 4.6 also illustrates that the SW component of the
cloud response acts as a negative feedback in the hemisphere forced more strongly by aerosols
(in ASYMM\textsubscript{NH} and ASYMM\textsubscript{SH}) by allowing more SW absorption locally, including in the
extratropics.

4.3.2. Role of feedbacks on the ITCZ shift

Operating under the assumption that TOA feedback asymmetries promote or dampen the
magnitude of ITCZ migration, we can assess the relative importance of hemispheric asymmetries
in the climate response for migrations in tropical precipitation. A consideration of the role of
tropical cloud feedbacks in determining the magnitude of ITCZ shift was done in an idealized
framework by Voigt et al. (2014), operating within a TOA budget framework. In that study, the
authors are directly able to answer the question of whether clouds (in the particular models analyzed) promote or dampen the magnitude of ITCZ shift, by a method in which they store the radiation-relevant cloud fields at each call of the radiation model in a unperturbed reference simulation and then prescribe them in the perturbed simulation. In this way, they decouple the cloud radiative influence from the actual circulation changes in their perturbed experiments. They found that the asymmetric cloud radiative response (including both SW and LW contributions) at the TOA was consistent with the role of clouds in acting as a positive or negative feedback on the ITCZ migration. Since the LME model simulations used here have already been run, I proceed with the assumption that hemispheric asymmetries in the feedbacks are a proxy for their relative leverage on the anomalous AHT$_{eq}$, and therefore on the movement of the ITCZ. Breakdowns in this assumption constitute an important source of uncertainty for the following conclusions.

Figure 4.7 shows the regression of the anomalous AHT$_{eq}$ against the hemispheric asymmetry (hereafter, $\Delta G_h$) in all forcings and feedbacks individually. Negative values of $\Delta G_h$ are associated with a larger energy sink in the NH relative to the SH (following the volcanic excursion), while positive values indicate a larger energy sink in the SH relative to the NH. A situation in which a larger energy sink in the NH (SH) accompanies northward (southward) $\Delta$AHT$_{eq}$ (i.e., a negative slope in the diagrams of Figure 4.7) is therefore interpreted as promoting an ITCZ shift. In contrast, if preferential energy loss in the NH is associated with southward total $\Delta$AHT$_{eq}$ (positive slope in the diagrams), then that individual feedback is interpreted as dampening the ITCZ shift.

As expected, the aerosols themselves promote an ITCZ shift, and explain most of the hemispheric TOA energy budget asymmetry. The correlation between $\Delta$AHT$_{eq}$ and the $\Delta G_h$ is of
marginally greater magnitude (r = -0.8) than that between ΔAHT_{eq} and the aerosol only component of ΔG_{r}(r= -0.84). Asymmetries in water vapor are the most important non-aerosol and non-temperature related feedback, and also promotes an ITCZ shift by reinforcing the energy sink in the preferentially forced hemisphere. The lapse rate and surface albedo feedbacks contribute minimally to the TOA asymmetry. The Planck feedback dampens the TOA asymmetry due to the locally decreased LW emission (an energy source) in the preferentially forced hemisphere. It was argued previously that the asymmetric component of this response must be dominated by the extratropics. The Planck asymmetry also tends to be stronger for ASYMM_{SH} eruptions than for ASYMM_{SH} eruptions, since the asymmetric temperature response tends to be larger for a NH eruption (Figure 4.4).

As hypothesized in the introduction, the SW component of clouds do indeed compensate for the aerosol-induced asymmetry, by shifting clouds into the darker hemisphere following asymmetric volcanic forcing. For ASYMM_{SH} events, the hemispheric asymmetry in the SW cloud negative feedback is even stronger than that of the Planck response. Much of this compensation is achieved in the tropics (Figure 4.8), due to the ITCZ shift itself. Indeed, ITCZ migrations are expected to buffer large asymmetries in hemispheric albedo on longer timescales (Stephens et al., 2015). However, in the LME simulations, the low cloud reduction in the extratropics (of the preferentially forced hemisphere) also dampens the hemispheric asymmetry and ITCZ shift. Note that in Figure 4.8, as clear-sky upwelling SW radiation preferentially increases in one hemisphere (positive values on the horizontal axis, a stronger energy sink for that hemisphere), the SW cloud response is such that reflection increases more strongly in the opposite hemisphere (positive values on the vertical axis indicate a stronger energy sink in the SH). This is true for both the tropical-only and extratropical-only domains.
Although the SW cloud compensation is large in the tropics, it is mostly offset by the LW cloud feedback. The sum of the cloud feedbacks acts to slightly diminish the ITCZ shift (Figure 4.7c), although this asymmetry actually arises from the extratropical contribution. This may be because the low cloud reductions in mid-latitudes exert greater leverage on the SW energy budget than on the LW, but the hemispheric asymmetry in net tropical cloud feedback is nearly zero.

Taken together, the non-cloud feedbacks exert very little influence on the asymmetry in the TOA energy budget (Figure 4.7j), largely due to cancellation of the Planck and water vapor components. Consequently, the role of all feedbacks combined follows that of the cloud contribution, and acts to diminish the TOA energy asymmetry and ITCZ shift.

4.4. Conclusions

In this study, the contribution of individual feedbacks (following large volcanic eruptions) to the asymmetry in TOA energy budget was quantified, with results compared across multiple eruption categories that differ in the spatial structure of aerosol loading. It is argued that non-temperature and non-cloud feedbacks promote asymmetry, largely due to the water vapor feedback, which enhances the energy sink in the hemisphere forced more strongly by aerosol scattering. However, temperature feedbacks largely compensate for this, leaving a residual cloud feedback, dominated by the SW component in the extratropics, that determines the net feedback asymmetry. In the CESM LME, this feedback acts to reduce the hemispheric asymmetry caused by the aerosols themselves, which was interpreted to be a negative feedback on the ITCZ migration.
As with most questions of cloud feedbacks, a natural follow-up question for future study is to assess whether these results are robust to multiple models, or to different radiative kernel data or methods to calculate the feedbacks. Even in an idealized framework, it has previously been found that the radiative impact of clouds on the ITCZ shift differs in sign and magnitude across models subjected to surface albedo perturbations, and is responsible for a large amount of model spread in the ITCZ shift (Voigt et al., 2014), although the transient response may be different to a rapid and large volcanic forcing.

In a fully-coupled model framework, changes in the ocean circulation may also act as a feedback on the ITCZ shift, since the atmosphere’s hemispheric energy balance is determined not just by TOA fluxes but also turbulent surface energy fluxes. Indeed, the ocean circulation cools the tropics more than the atmosphere in the long term mean (Trenberth and Caron, 2001) and wind-driven subtropical cells (driven by the surface stress from the trade winds) should be expected to be coupled to shifts in the ITCZ.

All results presented in this study represent a response consistent with the two-year average following volcanic eruptions. A further topic of inquiry is to assess the time evolution of the feedbacks and their collective role in determining the sensitivity of ITCZ displacements to external forcing. The magnitude of ITCZ shift and the temporal evolution of the associated precipitation anomalies may have large implications for regions that receive ITCZ-related rainfall, especially areas such as the Sahel that straddle the poleward edge of boreal summer ITCZ migration. Furthermore, since the movement of the ITCZ may leave paleoclimate imprints in water isotopes (as reported in chapter 3), it is imperative to further understand its susceptibility to change across forcing mechanisms and timescales in order to properly interpret paleoclimate data.
4.5. Appendix

The following elaborates on calculations in Figure 4.1.

The mass streamfunction is calculated as:

$$\psi_m(p) = \frac{2\pi a}{g} \int_0^p v^*(p) \cos \phi dp$$  \hspace{1cm} (A1)

where $a$ is the planetary radius, $g$ is acceleration due to gravity, $p$ is pressure, $\phi$ is latitude, and $v^*$ is an adjusted meridional velocity. The monthly CESM model output is provided with pressure as the vertical dimension, which has been interpolated from the native $\sigma$-p coordinate in the vertical, and a result is a spurious slight non-conservation of mass in the post-processed model files. Accordingly, a spurious mass transport is calculated as (Yang et al., 2014):

$$v^+ = \frac{\int_0^{ps} v \cos \phi \, dp}{\int_0^{ps} \cos \phi \, dp}$$  \hspace{1cm} (A2)

and $v^*(p) = v(p) - v^+$.

In figure 4.1d, the ITCZ latitude for both the pre-eruption and post-eruption state is calculated as a precipitation-weighted expected latitude (e.g., Adam et al., 2016):

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\[ \phi_{itcz} = \frac{\int_{20^\circ S}^{20^\circ N} \phi P \cos \phi \, d\phi}{\int_{20^\circ S}^{20^\circ N} P \cos \phi \, d\phi} \] (A3)

Other commonly used functions to calculate \( \phi_{itcz} \) (see e.g., Adam et al., 2016), such as a center of mass in which an equal amount of area-weighted precipitation falls between 20°S and \( \phi_{itcz} \) as it does from \( \phi_{itcz} \) and 20°N yield qualitatively similar results, but as discussed in chapter 3, the slope relating anomalies in cross-equatorial atmospheric energy transport to the magnitude of the ITCZ shift may vary considerably. For other indices considered, however, variations were well correlated with that of equation A3.
References


Figure 4.1. Mass streamfunction anomaly (Tg s$^{-1}$, or $10^9$ kg s$^{-1}$) for the (a) ASYMM$_{NH}$ and (b) ASYMM$_{SH}$ composite. Values defined positive (negative) for clockwise (counterclockwise) circulation in the pressure-latitude plane. (c) Zonal-mean precipitation anomalies for ASYMM$_{NH}$ (green line), ASYMM$_{SH}$ (blue line) and SYMM (black line) classifications. (d) Change in ITCZ latitude (see appendix) vs. anomaly in atmospheric energy transport across the equator ($\text{AHT}_{eq}$ in 0.1 PW, where 1 PW = $10^{15}$ Watts). Positive (negative) $\Delta \text{AHT}_{eq}$ represents anomalous northward (southward) energy transport. Colored points represent the individual eruptions in ASYMM$_{NH}$ (green), ASYMM$_{SH}$ (blue), and SYMM (black) after averaging over the five ensemble members. The cloud of grey points represents each realization of all eruptions. Regression line is fit through the colored data points. All anomalies are derived from the two-year average following peak eruption, relative to the five-year average prior to the eruption.
**Figure 4.2.** Zonal-mean climate forcing or feedbacks (W m$^{-2}$ K$^{-1}$) in a doubled CO$_2$ run using the radiative kernel method from the NCAR CCSM3 model (see text), except for the dashed aqua line (using the APRP method). Responses defined positive (negative) when the response amplifies warming (cooling) at that latitude. Responses are derived from data averaged over the final 10 years from the available control and perturbed monthly output.
Figure 4.3. Zonal-mean climate forcing or feedbacks (W m$^{-2}$ K$^{-1}$) for ASYMM$_{NH}$ (top), ASYMM$_{SH}$ (middle) and SYMM (bottom) from the CESM Last Millennium composite (see text for details), as labeled in the top panel. Responses defined positive (negative) when the response amplifies cooling (warming) at that latitude.
Figure 4.4. Hemispheric contrast in temperature anomalies (°C) for ASYMM\textsubscript{NH} (top) and ASYMM\textsubscript{SH} (bottom). Results shown for the composite mean in the pressure vs. latitude plane (defined by x°N minus x°S, where x is the labeled latitude).
Figure 4.5. Vertically integrated zonal-mean precipitable water anomalies expressed as a difference (solid lines, kg m$^{-2}$) or % change (dashed lines) normalized to the global-mean surface temperature anomaly. Results shown for the composite mean of all eruption categories.
Figure 4.6. Spatial pattern of all forcing or feedbacks (W m$^{-2}$ K$^{-1}$) in the composite mean for ASYMM$_{NH}$ (left column), ASYMM$_{SH}$ (middle column), and SYMM (right column), with the individual terms labeled for each row. Note the different scales for each variable. The cloud LW and cloud SW feedbacks have the same colorbar (shown once). Note the temperature feedback is the sum of the Planck and lapse rate components.
Figure 4.7. AHT$_{eq}$ anomaly (0.1 PW) vs. Hemispheric contrast (NH minus SH) in TOA radiative response (W m$^{-2}$) for (a) all forcings and feedbacks (b) aerosol only forcing (c) net cloud feedback (d) Planck feedback (e) Lapse Rate (f) Water Vapor (g) LW clouds (h) SW clouds (i) Surface Albedo (j) all non-cloud feedbacks, and (k) all feedbacks. Negative values on the x-axis represent a stronger anomalous
energy sink in the NH relative to the SH, and a positive (negative) slope through the data corresponds to an agent that dampens (promotes) asymmetry in the TOA energy budget. Note that the feedbacks are not normalized by temperature, and the different x-axis scales for panels a-b versus c-k.
Figure 8: Hemispheric Contrast (NH minus SH) in the SW cloud radiative response (W m⁻²) versus hemispheric contrast in upwelling clear-sky SW radiation. Negative values for the cloud response indicate an energy sink, so positive values on the vertical axis indicate increases in SW reflection by clouds in the SH relative to the NH. Positive values for upwelling clear-sky SW radiation indicate an energy loss term for the atmosphere, so positive values on the horizontal axis indicate preferential clear-sky SW reflection and a stronger sink of energy in the NH. Results are shown for global (0°N-90°N minus 0°S-90°S), tropical (0°N-30°N minus 0°S-30°S), and extratropical (30°N-90°N minus 30°S-90°S) domains. All datapoints represent individual eruptions after averaging over the five ensemble members.