Energetic Constraints on the Time-Dependent Response of the ITCZ to Volcanic Eruptions

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ABSTRACT: Energetic constraints on the time-dependent response of the intertropical convergence zone (ITCZ) to volcanic eruptions are analyzed using the Community Earth System Model Last Millennium Ensemble project. The energetic constraints are found to vary during the first few years, governed by conjoined variations of the energy budgets of the stratosphere, troposphere, and oceans. Specifically, following eruptions, sulfate aerosols heat the stratosphere by longwave absorption and cool the surface by shortwave reflection, leading to contrasting energy transport anomalies in the stratosphere and troposphere, which are of comparable strength during the first year. Similar contrasting responses are also seen by the mean and eddy components of atmospheric energy transport (AET). Consequently, ocean energy transport (OET) dominates the anomalous total interhemispheric energy transport during the first year. However, wind-driven OET, generally assumed to constrain shifts of the ITCZ, has a negligible role in the transient ocean response. Consistent with theory, anomalous cross-equatorial tropospheric energy transport, dominated by the anomalous Hadley circulation, is strongly negatively correlated with ITCZ shifts. However, due to the strong anomalous stratospheric energy fluxes, the commonly used energy flux equator (derived from net AET) is a poor predictor of transient ITCZ shifts following eruptions. El Niño-like conditions typically appear during the second year after eruptions, and La Niña-like conditions appear after the third year. These variations modulate ITCZ shifts in a complex manner, via changes in surface conditions and in associated energy transport variations in the atmosphere and oceans.

SIGNIFICANCE STATEMENT: Volcanic eruptions can lead to strong climatic impacts, which include global cooling and shifts of the tropical rain belt, with major socioeconomic implications. The global impacts of volcanic events motivate potential future geo-engineering applications, based on mimicking the effect of volcanic eruptions by artificial aerosol loading of the stratosphere. Understanding the transient and long-term response to volcanic eruptions is therefore of prime importance. Here we analyze energetic constraints on time-dependent shifts of the tropical rain belt in response to volcanic eruptions. In contrast to previous findings, we show that due to contrasting stratospheric and tropospheric responses, net atmospheric energy transport is a poor predictor for ITCZ shifts during the first year following eruptions. Instead, we elucidate the roles of components of atmospheric and oceanic energy transport in setting the transient response. El Niño-like conditions typically appear during the second year after eruptions, and La Niña-like conditions appear after the third year. These variations affect the tropical rain belt in a complex manner, via changes in surface conditions and in associated energy transport variations in the atmosphere and oceans.

KEYWORDS: Intertropical convergence zone; Atmospheric circulation; Energy transport; Stratosphere–troposphere coupling

1. Introduction

Tropical and extratropical volcanic eruptions lead to shifts of the tropical rain belt (Robock 2000; Timmreck 2012; Iles et al. 2013; Iles and Hegerl 2014; Kremser et al. 2016), with major socioeconomic implications to tropical populations (Stothers 1984, 2000; Liu et al. 2016). These shifts modulate the ensuing interannual and decadal climate variability by coupled dynamic and thermodynamic ocean–atmosphere feedbacks, which are not well understood (Pausata et al. 2015; Stevenson

et al. 2016; Pausata et al. 2020). Large uncertainties in historical records make it difficult to characterize the direct climate response to volcanic eruptions. Therefore, studies of the response to volcanic eruptions often rely on simulations by comprehensive climate models (Schneider et al. 2009; Maher et al. 2015; Otto-Bliesner et al. 2016). Based on model simulations, previous works have shown a strong link between annually averaged variations in the atmospheric energy budget and shifts of the intertropical convergence zone (ITCZ) following volcanic eruptions (Schneider et al. 2009; Pausata et al. 2015; Colose et al. 2016; Colose and LeGrande 2016; Stevenson et al. 2016). Building on these studies, here we elucidate the energetic constraints on shifts of the ITCZ during the first few years following eruptions.

The primary driver of large-scale climatic response following eruptions is the injection of sulfur dioxide into the stratosphere. Once in the stratosphere, it forms sulfate aerosols that disperse globally, with a typical residence time of around 1–2 years, which is generally longer in the tropics and decreases in the extratropics due to subsidence by the stratospheric Brewer–Dobson

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circulation (Gerber et al. 2012). Eruptions that do not reach the stratosphere have a limited effect on climate that fades within weeks following the eruption, once the volcanic matter is washed out or settles to the ground (Robock and Mass 1982; Niemeier et al. 2009). The direct radiative forcing by the stratospheric sulfate aerosols cools the lower atmosphere and oceans by reflecting shortwave radiation, and warms the stratosphere by increased absorption of longwave radiation, with a net cooling effect (Angell 1997; Robock 2000). Additional processes, which evolve on multiple time scales, typically include: 3–6 months decrease in tropical precipitation by surface cooling; global cooling for 1–3 years by increased global-mean albedo; ozone depletion by heterogeneous chemical reactions involving the sulfate aerosols for 1–2 years; Arctic sea ice feedbacks which can extend albedo anomalies to over a decade (Schneider et al. 2009); and nonmonotonic adjustment of the Atlantic meridional overturning circulation (AMOC) over several decades (Stenchikov et al. 2009; Mignot et al. 2011; Swingedouw et al. 2015; Pausata et al. 2015). Multiple consecutive eruptions can result in persistent multidecadal global cooling. Such a chain of eruptions, for example, is associated with the onset of the Little Ice Age during the fifteenth to nineteenth centuries (Zhong et al. 2011; Miller et al. 2012), with enduring effects on the tropical rain belt (Sachs et al. 2009; Yan et al. 2015). Motivated by these long-term effects, artificial stratospheric aerosols are commonly considered in the context of future geo-engineering applications (Gerber et al. 2012; Clarke et al. 2015; Laakso et al. 2016; National Academies of Sciences and Engineering 2021).

Given the complexity of the response, energetic constraints can provide a powerful conceptual framework for understanding the mechanisms linking volcanic eruptions to climatic changes (Robock and Mao 1995; Fischer et al. 2007; DallaSanta et al. 2019). In particular, the position of the ITCZ is known to covary with the energy flux equator (EFE) of the atmosphere (the latitude where vertically integrated meridional energy transport in the tropics diverges and vanishes; Kang et al. 2008), which generally shifts toward the differentially heated hemisphere. This covariance is predicated on the assumption that inter-hemispheric energy fluxes are dominated by the Hadley circulation (i.e., the mean tropospheric tropical overturning circulation; Kang et al. 2008; Schneider et al. 2014). However, the strong stratospheric temperature anomalies associated with eruptions (e.g., Schneider et al. 2009) suggest stratospheric energy transport needs to be considered when applying energetic constraints on ITCZ shifts following eruptions.

Under ideal conditions, where energy transport is dominated by the Hadley circulation and atmospheric energy transport varies linearly near the equator, the position of the ITCZ (ϕ_I) is expected to be negatively correlated with cross-equatorial AET (AET_0) and inversely proportional to the meridional gradient of AET at the equator, which approximately equals equatorial atmospheric net energy input (NEI_0):

$$\phi_I = -\frac{1}{a} \frac{AET_0}{NEI_0}, \quad (1)$$

where a is Earth's radius (Bischoff and Schneider 2014; Schneider et al. 2014). Fractional changes in the position of the

ITCZ can therefore be decomposed into components resulting from variations in either AET_0 or NEI_0 (Bischoff and Schneider 2014; Shaw et al. 2015):

$$\frac{\delta\phi_I}{\phi_I} = \frac{\delta AET_0}{AET_0} - \frac{\delta NEI_0}{NEI_0}. \quad (2)$$

Indeed, a strong link between the position of the ITCZ and cross-equatorial AET is seen on seasonal or longer time scales, in both observations and simulations (Frierson and Hwang 2012; Donohoe et al. 2013, 2014; Adam et al. 2016a). Similarly, equatorial NEI has been shown to modulate the position of the ITCZ during the seasonal cycle (Adam et al. 2016a), during El Niño–Southern Oscillation (ENSO) episodes (Adam et al. 2016a; Lintner and Boos 2019), and in response to orbital variations (Bischoff et al. 2017; Adam et al. 2019; Molnar and Rajagopalan 2020). Following sufficiently strong volcanic eruptions, consistent with Eq. (1), the ITCZ is known to move away from the hemisphere of the eruption (i.e., away from the cooling hemisphere; Haywood et al. 2013; Iles et al. 2013; Iles and Hegerl 2014; Liu et al. 2016). However, the time-dependent relation of these shifts to the relevant energetic quantities has not been explored.

The tropical response to volcanic eruptions is also known to include ENSO-like variations during the first few years (McGregor and Timmermann 2011; Li et al. 2013; Stevenson et al. 2016; Liu et al. 2018; Predybaylo et al. 2020; Pausata et al. 2020). While the nature of these variations is not well understood, several mechanisms link ENSO variations to shifts of the ITCZ (Adam et al. 2016a; Lintner and Boos 2019; Pausata et al. 2020). Moreover, recent works have shown that since the mean wind-driven ocean energy transport is in the same direction as atmospheric energy transport (AET), it tends to moderate shifts of the ITCZ in response to extratropical heating (Green and Marshall 2017; Green et al. 2019; Kang 2020), but can also amplify ITCZ shifts due to the coupled response of atmospheric eddies (Roberts et al. 2017; Xiang et al. 2018; Afargan-Gerstman and Adam 2020). Therefore, consideration of coupled atmospheric and oceanic energy transport is required following volcanic eruptions.

Here we study energetic constraints on the transient response of the ITCZ to volcanic eruptions, exploring the roles of atmospheric and oceanic energy transport. Specifically, we analyze time-dependent shifts of the ITCZ following tropical and extratropical eruptions, simulated in the Community Earth System Model (CESM) Last Millennium Ensemble (LME) project (Otto-Bliesner et al. 2016). In contrast to previous works that focused on the role of AET, we show that the time-dependent shifts of the ITCZ are governed by conjoined variations in the energy budgets of the stratosphere, troposphere, and oceans. The data and methods are presented in section 2. An analysis of the time-dependent response is presented in section 3, followed by a discussion and conclusions in section 4.

2. Methods

a. Data

The CESM LME project includes fully coupled global simulations of Earth's climate during the period 850–1850 AD,

based on the CESM-CAM5-CN version of the model (Otto-Bliesner et al. 2016). The simulation period includes a total of 71 volcanic eruptions that injected sulfate into the stratosphere, thereby creating a lasting perturbation on Earth's climate. We use 17 ensemble members, of which 12 include all available forcings (volcanic, solar, orbital, greenhouse gases, ozone, aerosols, land use, and land cover) and 5 include volcanic forcing only. Our analysis is based on monthly data with spatial resolution of 1.9° latitude $\times$ 2.5° longitude, and 30 vertical levels.

Volcanic forcing in the CESM LME project is implemented as prescribed latitude, altitude and time-dependent stratospheric SO_4 loading, taken from Gao et al. (2008)—a reconstruction based on ice cores. Data are available for 43 vertical levels and 18 latitude belts, from which we calculate the global and hemispheric sulfate loading.

b. ITCZ position

The tropical mean precipitation in the CESM model is generally lower than observed near the equator, and higher than observed south of the equator, in particular over the Pacific. This problem, common to most modern climate models and generally known as the double-ITCZ bias (Meechoso et al. 1995; Lin 2007; Adam et al. 2018a), leads to an excessively doubly peaked precipitation distribution in the tropics, with a pronounced minimum near the equator (Fig. 1a). Due to the doubly peaked precipitation distribution, the latitude of maximal precipitation jumps between hemispheres during the seasonal cycle, or in response to strong hemispherically asymmetric precipitation variations. To avoid this problem, and to reduce grid dependence, the position of the ITCZ is calculated as

$$\phi_n = \frac{\int_{20^\circ\text{S}}^{20^\circ\text{N}} \phi [P \cos(\phi)]^n d\phi}{\int_{20^\circ\text{S}}^{20^\circ\text{N}} [P \cos(\phi)]^n d\phi}, \quad (3)$$

where ϕ is latitude, P is zonal-mean precipitation, and n acts as a smoothing parameter (Adam et al. 2018b). For $n = 1$, Eq. (3) yields the precipitation centroid; for large n , ϕ_n converges to the exact latitude of maximal area-weighted precipitation. Our results are not sensitive to the value of n for $1 < n \leq 10$. Colose and LeGrande (2016), who analyzed the same CESM LME data, used $n = 3$, which they found to yield the ratio closest to unity between EFE and ϕ_n variations. Here we define the position of the ITCZ (ϕ_I) as the precipitation centroid, which is a more commonly used metric, known to be well correlated with the EFE across a range of observational and modeling studies (Frierson and Hwang 2012; Donohoe et al. 2014).

c. Energy-flux framework

The column-integrated, zonal-mean energy balance of the atmosphere can be written as

$$\partial_t e + \partial_y \text{AET} = \text{NEI}, \quad (4)$$

stating that net energy input into the atmosphere (NEI) is balanced by energy storage in the atmospheric column and

FIG. 1. Zonal-mean precipitation for (a) ensemble-mean climatology: January–February–March (JFM; red) and July–August–September (JAS; yellow); and (b) annual mean for the first year after Northern (NH; green) and Southern (SH; blue) Hemisphere events. The zonally averaged annual mean precipitation (ANN; black) is shown for reference in both panels. Vertical lines show the latitude of the precipitation centroid for the corresponding averages. The gray line in (a) shows the observed zonally averaged annual mean precipitation (Global Precipitation Climatology Project for 1979–2014; Adler et al. 2003).

divergence of atmospheric energy transport (AET). (Cartesian notation is used for simplicity; all calculations are done in spherical coordinates.) Here e is the column-integrated sum of moist enthalpy (the sum of thermal and latent energy) and kinetic energy, and AET consists of the column-integrated meridional transport of moist static energy (MSE, the sum of moist enthalpy and geopotential energy) and kinetic energy¹ (for more details, see Adam et al. 2016a). Atmospheric NEI is calculated as the net radiative input into the atmospheric column at the top of the atmosphere and at the surface, and upward latent and sensible heat fluxes at the surface.²

From Eq. (4), net meridional AET can be calculated by integrating $\text{NEI} - \partial_t e$ from the poles to some latitude:

$$\text{AET}(\phi) = 2\pi a^2 \int_{-\pi/2}^{\phi} (\text{NEI} - \partial_t e) \cos(\phi') d\phi'. \quad (5)$$

To minimize integration errors, AET is calculated as the average of the integrals from the south and north poles. Interhemispheric total energy transport is similarly calculated by integrating the radiative balance at the top of the atmosphere (TOA; Fig. S2 in the online supplemental material). The EFE is calculated as the latitude where $\text{AET}(\phi)$ both diverges and changes sign near the equator. Our confidence in the ability of CESM to capture the transient dynamic and

¹ Kinetic energy is included in the calculation for completeness, but has a negligible contribution to the column-integrated budget.

² To close the atmospheric energy budget, here NEI also includes a snowfall rate term, which accounts for latent heat fluxes associated with snowmelt in the CESM model (Colose and LeGrande 2016).

energetic variations following volcanic eruptions is bolstered by its skill in simulating the observed seasonal cycles of the EFE, the latitude separating the northern and southern Hadley cells, and the position of the ITCZ (ϕ_I) across all ensemble members (Fig. S1). As in observations, a lag of about six weeks exists between variations of the EFE and ϕ_I (Adam et al. 2016a).

Ocean energy transport (OET) is calculated by integrating the surface energy balance (Fig. S3). We note, however, that as defined here, OET includes contributions from both meridional energy advection and ocean energy storage (OES) (Donohoe et al. 2020). (Similarly, total energy transport includes contributions from meridional energy advection and energy storage in both the atmosphere and oceans.) We therefore decompose cross-equatorial OET (OET_0) into meridional transport by energy advection (OET_0^{flux}), which we obtain from monthly CESM fields, and transport implied by interhemispheric difference in OES, which we calculate as $OET_0^{\text{storage}} = OET_0 - OET_0^{\text{flux}}$.

To relate the advected component of OET to wind stress, following Afargan-Gerstman and Adam (2020), we estimate cross-equatorial Ekman mass transport using Sverdrup balance:

$$M_0^{\text{Ek}} = -\frac{1}{\beta_0} \partial_y \tau_x|_{\phi=0}, \quad (6)$$

where τ_x is the zonal surface wind stress and β_0 is the meridional gradient of the Coriolis parameter at the equator.

Vertically dependent energy transport in the atmosphere is calculated from monthly fields of the meridional fluxes of MSE and kinetic energy. Mean stratospheric energy transport $\overline{\text{SET}}$ [hereafter, $(\)$ denotes mean quantities] is calculated by integrating from the top of the atmosphere to the tropopause level (monthly mean CESM tropopause height field). Mean AET ($\overline{\text{AET}}$) is calculated by integrating over the entire atmospheric column; mean tropospheric energy transport ($\overline{\text{TET}}$) is then calculated as the difference between $\overline{\text{AET}}$ and $\overline{\text{SET}}$. Atmospheric eddy energy transport (EET) is calculated as the difference between AET and $\overline{\text{AET}}$. Equatorial values, denoted as $(\)_0$, are calculated by averaging over the latitudes equatorward of 5° for vertically integrated fluxes, and as the value at the equator for meridionally integrated fields.

d. Classification of events

We identify events using the methodology employed by Colose and LeGrande (2016): years of events are defined as years in which the global stratospheric sulfate loading was at least 8 Tg, averaged over at least one 5-month period. This classification is found to considerably improve the consistency of the transient response across all eruptions, and is not sensitive to changing the threshold by up to ± 2 Tg. Using this criterion we identify a total of 35 eruptions as events. The events are then divided into three types: events for which the sulfate loading in one hemisphere was at least 25% larger than in the opposite hemisphere were classified as either Northern (NH) or Southern (SH) Hemispheric events, according to the hemisphere with larger loading; all remaining events were classified as tropical (TROP). Overall we identify 17 NH events, 12 SH events, and 6 TROP events.

FIG. 2. (a) Monthly mean stratospheric SO_4 loading for each hemisphere, (b) the difference in stratospheric SO_4 loading between hemispheres (northern minus southern), and (c) 6-month running-mean global-mean temperature anomaly.

The monthly mean hemispheric sulfate loading for each type and the interhemispheric loading difference are shown in Fig. 2. The 6-month running mean anomaly of the global mean temperature is also provided for reference, showing a global cooling response to increased sulfate loading. The stratospheric sulfate concentration peaks around four months after the beginning of the eruption, and decays to below 5% of the peak after about three years (Stevenson et al. 2016). The hemispheric asymmetry of sulfate distribution is generally stronger after NH events. This is partly due to there being more terrestrial extratropical eruptions in the NH, but may also be related to asymmetries in the atmospheric circulation (Kang et al. 2020). The apparent oscillation in the interhemispheric difference in sulfate loading after TROP events (Fig. 2b, dotted line) is potentially caused by increased stratosphere–troposphere mass exchange during winter, which leads to seasonally varying asymmetry (Holton et al. 1995; Gao et al. 2008).

The monthly anomalies after volcanic eruptions are defined as deviations from the monthly climatology of the 30 years prior to each event. To increase the consistency across events, following Colose and LeGrande (2016), we normalize the anomaly time series such that the annual mean global sulfate loading is 20 Tg during the first year after the eruption.

There are several limitations to the stratospheric sulfate dataset that could affect our results. First, the season of the eruption, which is unknown for most events, is artificially set to April (26 out of 35 events start in April; Gao et al. 2008). Therefore, since the dynamic response to volcanic eruptions is strongly dependent on the season of the eruption (Gao et al. 2008; Stevenson et al. 2017), our results might be biased toward

FIG. 3. The 6-month averages of precipitation anomalies following (a)–(c) NH events, (d)–(f) SH events, and (g)–(i) TROP events (see text for details on the event classifications). Maps show the zonally asymmetric component of the response; side panels show the zonal mean response. For statistical significance of the response, see Fig. S4.

eruptions that occur during boreal spring. In addition, the classification of three out of six TROP events is ambiguous (this includes the Samalas eruption in 1258, which was the strongest eruption of the last millennium; this eruption might have a relatively strong effect on our results, despite the normalization of the anomalies). During these three events, the hemispheric sulfate aerosol contrast crosses the 25% threshold several times during the three years following the events. We also note that the classification is based upon the latitudinal distribution of the stratospheric sulfate aerosols, rather than the location of the volcano, which is unknown in many cases.

3. Results

a. Temperature and precipitation response

The time-dependent response to volcanic eruptions is characterized by a direct response to the sulfate aerosol forcing, most prominent during the first two years following eruptions (Colose and LeGrande 2016), with concomitant ENSO-like variations, lasting several years following eruptions (Stevenson et al. 2016). This is demonstrated in Figs. 3–5, showing the precipitation, surface temperature, and atmospheric net energy input (NEI) anomalies following the events, decomposed into zonal mean and zonally asymmetric components (i.e., deviations from the zonal mean). Three 6-month periods are shown: the second half year, indicative of the strongest direct response to the volcanic forcing; first half of the second year, when El Niño-like conditions typically appear; and the beginning of the fourth year, when La

Niña-like conditions typically appear. (We refer throughout the text to these periods as representing variations during the first year, second year, and onward).

Consistent with the expectation from energetic constraints, tropical precipitation decreases (increases) in the relatively cooling (warming) hemisphere following NH and SH events, indicating a shift of the ITCZ away from the eruption hemisphere (Figs. 3a,b and 3d,e, respectively; Colose and LeGrande 2016; Stevenson et al. 2016). During the first year following TROP events (bottom rows in Figs. 3, 4), an overall cooling and decrease in tropical precipitation is seen, with no clear zonal mean shift of the ITCZ.

The zonally asymmetric component of the direct response during the first two years following eruptions is as large as the zonal mean response (Atwood et al. 2020). Specifically, with the exception of northern Asia and Europe, which warm after all of the events³ (more so following TROP events), an overall

³ This phenomenon, referred to as NH winter warming in previous works, is associated with a positive phase of the Arctic Oscillation and the North Atlantic Oscillation due to increased pole-to-equator stratospheric temperature gradient (Robock and Mao 1992; Shindell et al. 2004; Ortega et al. 2015). [Stenchikov et al. (2002) suggest changes in the amplitude of tropospheric planetary waves as an alternative explanation.] This warming is most pronounced during the second half-year period shown in Fig. 4 due to the prescribed bias toward eruptions that occur in April, so that boreal winter dominates the response 7–12 months after the eruption.

FIG. 4. As in Fig. 3, but for surface temperatures. For statistical significance of the response, see Fig. S5.

relative cooling of land is seen during the first two years (Fig. 4, left and middle panels; Colose and LeGrande 2016; Stevenson et al. 2016). This is most pronounced in the rapidly cooling eruption hemisphere; however, relative cooling of land is also seen in the opposite hemisphere, in latitudes where there is no mean cooling (Figs. 4a,d). This suggests that ocean–land contrast, radiative, and ice feedbacks (Stevenson et al. 2016; Chen et al. 2019), all contribute to the relative cooling of land during the first two years. In particular, the stronger initial response to NH events is likely attributable to both the larger NH landmass and arctic sea ice feedbacks (we return to this point in section 3d).

Precipitation over land is similarly generally reduced following all events (Fig. 3). Following NH events, the zonally asymmetric component of the response in the African sector is opposite to the zonal mean response, leading to a weakened net response there (Fig. S4). Following TROP events, the regional ITCZ in the Asian sector shifts southward, while in the eastern Pacific it shifts northward (Fig. 3g). These contrasting regional shifts are similar but opposite to the shifts found by Mamalakis et al. (2020) in response to global warming, and are therefore likely related to the net global cooling following eruptions.

b. ENSO-like response

While the existence of an ENSO-like response to volcanic eruptions is well established, the nature of the ENSO-like response varies across studies. For example, previous works reported prominent La Niña-like cooling during the eruption year (McGregor and Timmermann 2011; Li et al. 2013), which is only weakly evident following TROP events in our analysis

(Fig. 4g). In the CESM LME, the Pacific response during the first year following NH events (Fig. 4a) shows surface warming along the equator and decreased zonal SST gradients just south of the equator (i.e., warming in the east, cooling in the west); a similar but weaker equatorial warming is seen following SH events, but with increased SST gradients just south of the equator (Stevenson et al. 2016).⁴ These anomalies differ from the El Niño- and La Niña-like anomalies seen in subsequent years (Fig. 4, middle and right panels), in particular in the equatorial regions commonly used to quantify ENSO variations. Sensitivity to the definition of the ENSO-like response may therefore contribute to the variance across studies—a point we wish to clarify in this section.

The nature of the ENSO-like response following volcanic events remains a subject of ongoing debate, as several mechanisms have been found to be at play. Predybaylo et al. (2020) suggest that event amplitude and conditions in the Pacific prior to events are critical to the evolution of the ENSO-like response. Earlier studies pointed to the “dynamic thermostat” response of the tropical Pacific to global cooling (Clement et al. 1996; Hirono 1988; Stevenson et al. 2016). Specifically, the temperature response caused by anomalous equatorial upwelling in the eastern Pacific tends to oppose the tropical mean

⁴ While based on the same data, our results differ slightly from Stevenson et al. (2016) due to differences in the classification of events, namely, a higher sulfate concentration threshold required for including eruptions in their analysis, and higher hemispheric loading asymmetry required for classification of asymmetric events (i.e., NH or SH).

FIG. 5. As in Fig. 3, but for atmospheric net energy input (NEI). For statistical significance of the response, see Fig. S6.

cooling (Clement et al. 1996; Seager et al. 2019)—i.e., promoting El Niño-like conditions following eruptions. Wind anomalies in the equatorial Pacific have been shown to influence the ENSO-like response via the Bjerknes feedback. Accordingly, anomalous westerlies in the equatorial Pacific, associated with land–ocean temperature contrasts (Ohba et al. 2013; Predybaylo et al. 2017), the weakening of the African monsoon (Khodri et al. 2017), and shifts of the subtropical jets (Pausata et al. 2020), have been suggested to promote El Niño conditions following eruptions. Of particular interest to the present analysis, shifts of the ITCZ have been suggested to favor El Niño conditions following NH events and La Niña conditions following SH events (Liu et al. 2018; Pausata et al. 2020). Specifically, due to the northward mean position of the ITCZ, the opposite shifts of the ITCZ following NH and SH eruptions lead to opposite surface zonal wind stress anomalies, and hence to contrasting ENSO-like variations following eruptions. This expectation, however, is in contrast to the La Niña-like response following either NH or SH events found by Kang et al. (2020), owing to wind-driven feedbacks.

Time-dependent anomalies of the oceanic Niño index (ONI, defined as the deviation of the 3-month running mean SST from a moving 30-yr climatology in the Niño 3.4 Pacific region) following events are shown in Fig. 6. The ONI response varies considerably across events during the first two years. However, SST variations in the equatorial Pacific following events may be driven by processes not related to ENSO, which may lead to misinterpretation of the ONI variations. Specifically, in the eastern Pacific, the ITCZ shifts northward (i.e., farther away from the equator) following SH events and southward (i.e., equatorward) following NH events. Since zonal SST gradients

in the Pacific increase as the ITCZ shifts away from the equator (Fig. 7a; Clement et al. 1996; Li and Philander 1996), SST variations in the equatorial Pacific are partly attributable to the shifts of the ITCZ—not to the ENSO-like variations.

To elucidate the role of ITCZ shifts and ENSO-like variations on SST anomalies following eruptions, in Fig. 7 we show time series of the correlations of posteruption tropical (equatorward of 30°) SST anomalies with tropical SST patterns related to: (i) northward minus southward seasonal ITCZ shifts (Figs. 7a,c; hereafter the *ITCZ shift mode*), and (ii) typical El Niño minus La Niña conditions in the CESM model⁵ (Figs. 7b,d). We note that the ITCZ shift mode is calculated based on the seasonal cycle, which has the limitation of confounding ITCZ shifts with other seasonal variations,⁶ but allows for better differentiation from the ENSO mode, which dominates interannual variations.

The use of correlations of tropical SSTs with the associated tropical SST patterns reduces the sensitivity to specific regions, as in the case of ONI, allowing differentiation of ENSO variations from the ITCZ shift mode. Indeed, the ITCZ shift

⁵ We note that known deficiencies in the representation of ENSO in the CESM model may bias our analysis of the ENSO-like response (Anchukaitis et al. 2010). Nevertheless, the key features of the ENSO-like response, especially after the first year, are consistent with other studies and with reconstructions (Stevenson et al. 2016).

⁶ Note that tropical SST anomalies are strongly correlated with NEI variations (Fig. 5), especially in the deep tropics (Adam et al. 2018a); in particular, zonally asymmetric variations are dominated by surface fluxes (Fig. S3).

FIG. 6. Monthly time series of ensemble mean anomalies of the oceanic Niño index (ONI), following NH (solid), SH (dashed), and TROP (dotted) events.

mode leads to opposite SST responses following NH versus SH events (Fig. 7c); as negligible shifts of the ITCZ are seen following TROP events, the correlations between SST anomalies and the ITCZ shift mode are generally insignificant following these events. The influence of the ITCZ shift mode decreases with time, as the ITCZ gradually returns to its mean position. Similarly, the ENSO-like response becomes considerably more consistent across events when the entire tropical region is considered, compared to ONI (Fig. 7d versus Fig. 6). Surprisingly, during the first year, La Niña conditions are most prominent following TROP events (Fig. 7d), while a weak relation with ENSO conditions is seen for both NH and SH events. It is therefore not clear that shifts of the ITCZ contribute to the instigation of ENSO events, as suggested by previous works (Liu et al. 2018; Pausata et al. 2020). A plausible interpretation of these results is that the ITCZ shift mode is superimposed on instigated ENSO variations, without directly promoting either ENSO phase.

Consistent with previous works, during the second year, El Niño-like SST warming and increased precipitation appear over the equatorial eastern Pacific for all cases (Figs. 3, 4b,e,h, 7d; Stevenson et al. 2016); about three years after the eruption, La Niña-like drying and cooling appear over the central and eastern Pacific for all three cases (Figs. 3, 4c,f,i 7d; Adams et al. 2003; Mann et al. 2005; McGregor et al. 2010; McGregor and Timmermann 2011; Li et al. 2013; Schneider et al. 2009; Stevenson et al. 2016). The uniformity of the time-dependent ENSO-like response across event types after the first year (Fig. 7d), leading to La Niña-like conditions that peak during the fourth post-eruption year for all events, is particularly perplexing given the known boreal spring predictability barrier of ENSO (Webster and Yang 1992). Specifically, the weakened SST gradients and Walker circulation in the tropical Pacific during boreal spring introduce large uncertainty in the evolution of ENSO following boreal spring. The uniformity of the response across event types therefore implies at least one of two possibilities: (i) the ENSO-like response is dictated by persistent forcing, associated with either wind anomalies or global cooling, sufficient to overcome boreal spring frailty; or (ii) the ENSO-like response is not closely related to the

canonical ENSO mode (McGregor et al. 2010). We proceed to analyze the zonal mean response, but return in section 3e to the role of the ENSO-like variations in setting the transient response of the ITCZ.

c. Zonal mean response

The time-dependent response of the zonally averaged SST anomalies is shown in Fig. 8 for the first five years after events (cf. Fig. 2 of Stevenson et al. 2016). The response to TROP and SH events exhibits equatorial surface cooling during the first year after the eruption, while for NH events the cooling is mostly confined to the NH. Consistent with the aforescribed ENSO-like response, all events exhibit equatorial surface warming (or weakening of the negative SST anomalies) during the second year, and an additional phase of equatorial cooling starting about two to three years after the eruption and ending about two years later (the equatorial SST variations are primarily driven by variations in the Niño 3 region; see Fig. S7). These variations, which continue to evolve after the stratospheric sulfate loading (and hence the radiative forcing) is already negligible (Fig. 2; Stevenson et al. 2016), illustrate the importance of dynamic and radiative feedbacks in the time-dependent response to volcanic eruptions.

The vertical profile of air temperature anomalies following events is decomposed into hemispherically symmetric and antisymmetric components in Fig. 9 (top and bottom panels, respectively). Similar symmetric responses are prominent following all event types. Specifically, global cooling following events is seen in the troposphere, while pronounced warming is seen in the stratosphere, owing to the direct diabatic radiative heating and lower thermal inertia in the stratosphere (Stenchikov et al. 1998). The differential heating of the stratosphere induces a lowering of the tropopause by 10–20 hPa, seen after all events. A pronounced antisymmetric component is seen following asymmetric events (i.e., NH and SH). Following these events, differential heating of the lower stratosphere relative to the troposphere is seen in the eruption hemisphere, leading to contrasting interhemispheric heating of the stratosphere and troposphere (Figs. 9e,f). As shown next, these contrasting effects have a critical role in the energetic response to volcanic forcing.

d. Energetic constraints on ITCZ shifts

Since ITCZ shifts and asymmetric heating are seen primarily following NH and SH events, our analysis hereon does not include TROP events. ITCZ shifts following asymmetric (i.e., NH and SH) events are significantly smaller than seasonal variations (Fig. 1b), but are nevertheless larger than interannual natural variability (ITCZ shifts of $\sim 1^\circ$ following eruptions versus interannual variability of $\sim 0.1^\circ$). This is equally true for the associated temperature anomalies (Figs. 8 and 9a–c). Therefore, to reduce noise related to natural and inter-revent variability, time-dependent asymmetric variations are shown with a centered 6-month running mean.

The anomalies in equatorial atmospheric net energy input (NEI_0) following eruptions are shown in Fig. 10. During the later half of the first year, NEI_0 decreases following SH, but increases following NH events. From Eq. (2), NEI_0 anomalies

FIG. 7. Typical SST pattern and monthly time series of the correlation of tropical ($|\phi| < 30^\circ$) SST response following events with each SST pattern for (a),(c) northward seasonal shifts of the ITCZ (Aug–Sep minus Mar–Apr means). (b),(d) Typical El Niño minus La Niña conditions (ONI values above 0.5 and below -0.5). Maps in the top panels show the zonally asymmetric component of each SST pattern; side panels show the zonal means of each pattern. Green rectangles show the Niño 3.4 region, used in the calculation of ONI. All calculations are done on months at least 10 years after eruptions.

therefore amplify the ITCZ shifts during the first year, by causing the ITCZ to shift farther poleward following SH events and farther equatorward following NH events. After the first year, NEI_0 increases during the El Niño phase, and decreases during the La Niña phase, in accordance with the variations in equatorial surface temperatures (Fig. 8). These variations are consistent with the observed equatorward shift of the ITCZ during El Niño episodes, and poleward shift during La Niña episodes (Adam et al. 2016b; Boos and Korty 2016).

To study the common features of the asymmetric response, we next examine normalized composites of the anomalies following NH and SH events, where positive (negative)

anomalies in the position of the ITCZ indicate shifts toward (away from) the eruption hemisphere (i.e., anomalies following SH events are added a minus sign). The asymmetric time-dependent responses of the positions of the energy flux equator (EFE) and precipitation centroid (ϕ_I) are shown in Fig. 11a for the first three years following asymmetric events. As expected, the precipitation centroid shifts away from the hemisphere of the eruption immediately after the eruption, and gradually returns to its climatological position after about 3–5 years. As in the seasonal cycle (Fig. S1), the peak shift in ϕ_I , about 7–8 months after the eruption, lags the peak in the radiative forcing (Fig. 2) by about 2 months.

FIG. 8. Monthly time series of normalized anomalies of the zonal mean SST following (a) TROP events, (b) NH events, and (c) SH events. The time series is smoothed using a centered 6-month running mean.

FIG. 9. Average temperature anomalies during the first 6 months after the eruption for (a),(d) TROP events; (b),(e) NH events; and (c),(f) SH events, decomposed into (top) hemispherically symmetric and (bottom) hemispherically antisymmetric components. Solid and dashed lines show the annual mean tropopause height prior to and during the first 6 months following events, respectively.

In contrast to the expectation from energetic constraints, the EFE shifts in the opposite direction to the ITCZ (i.e., toward the cooling hemisphere) in the months immediately following the eruption. The EFE then shifts toward the warming hemisphere, crossing its climatological position about ten months after the eruption. Only after 14 months is the EFE seen to covary with the gradual return of the ITCZ to its climatological position. The strong correlation between the positions of the EFE and ITCZ following eruptions, found in previous studies (Colose and LeGrande 2016), therefore holds only after the first year following eruptions, and does not reveal contrasting initial transient responses of the EFE and ITCZ.

The contrasting response of the EFE six months after eruptions is primarily due to the anomaly in cross-equatorial AET (AET_0), which, in contrast with the theoretical expectation, transports energy away from the cooling hemisphere (Fig. 11b). As a result, total interhemispheric energy transport, caused by the anomalies in the top of the atmosphere energy budget (TOA_0) is primarily achieved by cross-equatorial ocean energy transport (OET_0), peaking about five months after the eruption, when AET_0 is weakest (Fig. 11b). After the first year, AET_0 dominates the total interhemispheric transport.

To better understand the dominance of OET_0 during the first year, in Figs. 12a,b we show OET_0 decomposed into fluxes advected across the equator (OET_0^{flux} , orange) and cross-equatorial energy transport implied by interhemispheric differences in ocean energy storage (OET_0^{storage} , gray), separately for SH and NH events (left and right panels). The seamless and similar but opposite evolution of OET_0 following SH and NH events masks strong variations in both OET_0^{flux} and OET_0^{storage} .

Specifically, OET_0 anomalies evolve through synchronous but contrasting oscillations of OET_0^{flux} and OET_0^{storage} on annual and semiannual time scales. Common to SH and NH events, peak OET_0 is equi-partitioned between OET_0^{flux} and OET_0^{storage} , with peak OET_0^{flux} lagging peak OET_0 by about 4 months.

To further understand the ocean response, Figs. 12c,d shows the anomalies in cross-equatorial oceanic Ekman mass transport (M_0^{Ek} , blue), which are dominated by Pacific wind stress anomalies ($M_0^{\text{Ek,Pacific}}$, green). Surprisingly, Ekman mass transport is southward during the first year following both SH and NH events; during the second year, the Ekman mass transport changes sign after SH events but vanishes after NH events. This is inconsistent with the expectation for contrasting M_0^{Ek} responses to the poleward versus equatorward shifts of the ITCZ following SH and NH events, respectively (Kang et al. 2020). Other sources of meridional gradients in zonal surface wind stress [Eq. (6)], such as global cooling and land–ocean contrasts (Pausata et al. 2020), may therefore account for the similar initial responses of M_0^{Ek} following both events. Further, no correlation is seen between the time-dependent variations of M_0^{Ek} and OET_0^{flux} . This questions the role of wind-driven Ekman ocean energy transport in the emergent transient ocean response. Specifically, in the equilibrated state, ITCZ shifts are known to induce tropical wind-driven meridional ocean currents which transport energy toward the differentially cooling hemisphere (Held 2001; Green and Marshall 2017; Schneider 2017). However, transient wind-driven ocean energy transport (i.e., far from equilibrium) is not well understood. The lack of a consistent relation between the Ekman mass transport and OET_0^{flux}

FIG. 10. Monthly time series of the normalized anomalies in equatorial atmospheric net energy input (NEI_0) following NH (orange) and SH (blue) events. The shading indicates ± 1 standard deviation of the mean across all events. The blue bar at the right end shows the standard deviation of interannual variations, indicating natural variability. The time series are smoothed using a 6-month running mean.

following SH and NH events indicate that: (i) wind-driven Ekman energy transport contributes very little to the transient OET_0 response, and therefore (ii) ocean energy storage and eddy transport (associated with either turbulent wind stress or ocean eddies) have a critical role in the transient response (Czaja and Marshall 2006).

Ocean dynamics play a critical role in the climate response to anomalies in sea ice extent, and especially in mediating these anomalies to tropical regions, leading to changes in the distribution of tropical SSTs and precipitation (Deser et al. 2015; England et al. 2020). We therefore also examine the response of hemispheric sea ice area to SH and NH events (Figs. 12e,f). Following SH events, increased sea ice area is seen in the SH, with negligible changes in NH ice cover. In contrast, following NH events, sea ice area increases in the NH but decreases in the SH. Changes in sea ice area therefore reinforce interhemispheric heating through albedo effects, contributing to the stronger response following NH events. Moreover, the sea ice anomalies persist for over a decade (Schneider et al. 2009), and therefore have a critical role in sustaining the anomalous interhemispheric heating for time scales much longer than the volcanic forcing.

We now return to the atmospheric response, aiming to understand the relatively weak AET_0 response during the first year. Figure 11c shows that during the first few months, energy transport by the stratosphere and troposphere are approximately equal and opposite, leading to negligible net mean AET (AET_0). The cancellation of stratospheric and tropospheric transport is partially due to the lowering of the tropopause in the equatorial region (Fig. 9). However, this lowering of the tropopause neither qualitatively affects the contrasting responses of the stratosphere and troposphere, nor the net cancellation of AET_0 over the entire column. Contrasting stratospheric and tropospheric energy transport persist until the stratospheric diabatic heating by sulfate aerosols subsides, about 18 months after the eruption.

The normalized contrasting dynamic and energetic responses of the stratosphere and troposphere to the asymmetric radiative

FIG. 11. Monthly time series following asymmetric events of the normalized anomalies of (a) the precipitation centroid (ϕ_I ; red), energy flux equator (EFE; blue), and the predicted position of the ITCZ based on top of atmosphere radiative balance (ϕ_{TOA} ; yellow; see text for information on its calculation); (b) total cross-equatorial meridional energy transport (TOA_0 ; yellow), given by the sum of cross-equatorial energy transport by the atmosphere (AET_0 ; green) and ocean (OET_0 ; purple); and (c) cross-equatorial eddy energy transport (EET_0 ; lilac) and mean cross-equatorial atmospheric energy transport (AET_0 ; green), given by the sum of mean cross-equatorial transport in the troposphere (TET_0 ; bright blue) and stratosphere (SET_0 ; maroon). The shading indicates ± 1 standard deviation of the mean across all asymmetric events. To facilitate the comparison with the time-dependent shifts of the precipitation centroid (Fig. 11a), the vertical axis in (b) and (c) is inverted. The time series are smoothed using a centered 6-month running mean. See text for details on the normalization.

forcing are shown in Fig. 13 for the first two years following eruptions. The anomalous stratospheric circulation transports energy away from the eruption hemisphere (i.e., southward), and decays in proportion to the stratospheric aerosol forcing (Fig. 2). In contrast, the anomalous tropospheric circulation, driven by the interhemispheric surface temperature difference, transports energy toward the cooling eruption hemisphere (Figs. 13g–i; the upper branch of the Hadley circulation dominates the energy transport). The anomalous tropospheric transport persists for several years after the aerosol radiative forcing has completely decayed (not shown), sustained by the persistent interhemispheric temperature difference (Fig. 8).

FIG. 12. Monthly time series of the normalized anomalies following SH and NH events in (a),(b) OET_0 (purple) decomposed into advected flux (OET_0^{flux} ; orange) and cross-equatorial OET implied by interhemispheric differences in energy storage ($OET_0^{storage}$; gray); (c),(d) cross-equatorial Ekman mass transport [Eq. (6)], for zonal mean zonal wind stress over oceans (blue; M_0^{Ek}) and over the Pacific ($M_0^{Ek,Pacific}$; $160^\circ\text{--}270^\circ\text{E}$; $1\text{ Sv} = 10^9\text{ kg s}^{-1}$); and (e),(f) hemispheric sea ice area in the SH (red) and NH (bright blue). Shading indicates ± 1 standard deviation of the mean across event types (not shown for $OET_0^{storage}$, which is calculated as the residual of OET_0 and OET_0^{flux}). The time series are smoothed using a centered 6-month running mean. Northward mass and energy fluxes are positive. See text for details on the normalization.

In addition to the contrasting tropospheric and stratospheric responses, the partitioning of energy transport between eddy and mean AET (EET_0 and AET_0) shows similar contrasting transports during the first year (Fig. 11c). Specifically, the initial negative anomaly in AET_0 is primarily due to EET_0 , which opposes AET_0 during the first year. EET_0 then becomes negligible during the second year, after which $AET_0 \approx \overline{AET_0} \approx \overline{EET_0}$, in accordance with the underlying assumptions of the energetic constraints.

Thus, the contrasting responses by the troposphere and stratosphere and by the mean and eddy components of AET render the EFE (derived from net AET) an invalid predictor of ITCZ shifts during the first year following eruptions. Surprisingly, despite the complex and contrasting responses by the different energy transport elements, we find that the total interhemispheric energy transport, derived from the top of atmosphere radiative balance (TOA_0 , Fig. 11b) is a good predictor of ITCZ shifts following eruptions. A linear regression between the monthly position of the precipitation centroid and TOA_0 for all years excluding the 10 years following events gives the empirical estimated position of the precipitation centroid:

$$\phi_{TOA} = -2.6^\circ \text{PW}^{-1} \times TOA_0. \quad (7)$$

We find ϕ_{TOA} to be as good a predictor of the position of the precipitation centroid following eruptions ($R = 0.92$; Fig. 11a) as tropospheric cross-equatorial energy transport ($\overline{EET_0}$, $R = -0.89$), which, according to theory, should be the best predictor of ITCZ shifts based solely on cross-equatorial energy transport. Moreover, changes in ϕ_{TOA} precede shifts of the precipitation centroid by about a month, and may therefore be particularly useful in predicting ITCZ shifts following eruptions. However, the robustness of TOA_0 as a predictor of ITCZ shifts following volcanic eruptions should be tested in other datasets.

e. ENSO modulation of the energetic constraints

ENSO variations are known to be related to cross-equatorial atmospheric energy fluxes (Adam et al. 2016b; Lintner and Boos 2019) and ocean energy transport (Cheng et al. 2019). The consistent ENSO-like response following events (Fig. 7) may therefore affect the energetic response. To assess the influence of the ENSO-like variations following events on the energetic constraints, Fig. 14 shows lead-lag regressions of ensemble-mean ONI and components of cross-equatorial energy transport in the atmosphere and oceans, during months at least ten years after eruptions (a similar plot for the anomalous response following events is not shown, due to the sensitivity of ONI to

FIG. 13. The 4-month averages of (a)–(f) anomalous radiative heating (colors) and meridional streamfunction (contours) and (g)–(l) anomaly in meridional MSE transport. Maroon lines show the tropopause height. The streamfunction intervals are $2 \times 10^9 \text{ m}^2 \text{ s}^{-1}$ for positive values (solid lines, clockwise rotation) and $3 \times 10^9 \text{ m}^2 \text{ s}^{-1}$ for negative values (dashed lines, counterclockwise rotation).

processes not related to ENSO, as discussed in section 3b). ENSO variations generally coincide with or lead the cross-equatorial fluxes, with some exceptions discussed below.

Southward AET_0 and northward OET_0 are seen during Niño episodes, with a total net northward transport. The associated OET_0 anomaly is confined to the months immediately prior to and after ENSO, whereas AET_0 lagging Niño episodes by more than 3 months becomes northward and persists for over a year after events (Fig. 14a). Based on typical ONI values during the Niño and Niña phases following eruptions (Fig. 6), cross-equatorial energy fluxes associated with ENSO are of the same order as the anomalous cross-equatorial fluxes following eruptions. The net northward (southward) fluxes associated with Niño (Niña) conditions in the CESM LME therefore introduce asymmetry to TROP events, weaken (amplify) NH (SH) events during the second year, and amplify (weaken) NH (SH) events after the second year. This also holds for the ITCZ shifts following events, as cross-equatorial tropospheric energy transport $\overline{\text{TET}}_0$ is also positive during and after Niño events (Fig. 14c).

As in the response to asymmetric events, contrasting sensitivity of $\text{OET}_0^{\text{flux}}$ and $\text{OET}_0^{\text{storage}}$ is seen to ENSO, with ENSO always leading the oceanic fluxes (Fig. 14b). In contrast, the implied cross-equatorial AET due to atmospheric energy storage $\text{AET}_0^{\text{storage}}$ is positive prior to and during Niño episodes and negative following Niño episodes. Similarly, mean atmospheric ($\overline{\text{AET}}_0$) and stratospheric ($\overline{\text{SET}}_0$) fluxes, as well as atmospheric eddy transport (EET_0), lead ENSO by up to a year (Fig. 14c); the $\overline{\text{AET}}_0$ and EET_0 show contrasting sensitivities to ENSO, which peak about 6 and 4 months following ENSO events, respectively. This suggests that the contrasting AET_0 and EET_0 responses following events (Fig. 11c) are in part due to the Niño-like phase during the second year. More generally, by affecting the atmospheric and oceanic energy fluxes (as well

as surface conditions), the ENSO-like variations following events modulate the transient response of the ITCZ to volcanic events, and contribute to the contrasting responses by the various energy transport elements.

4. Discussion

Volcanic eruptions can lead to strong climatic impacts that vary over time. These include global cooling and shifts of the tropical rain belt, which entail major socioeconomic implications. Understanding the transient and long-term response to volcanic eruptions is therefore of prime importance (National Academies of Sciences and Engineering 2021). The present analyses focus on the mechanisms and energetic constraints that determine the time-dependent shifts of the ITCZ in response to volcanic eruptions.

The response of the ITCZ to volcanic eruptions is characterized by an initial shift away from the eruption hemisphere, typically peaking about eight months after the eruption—lagging peak radiative forcing by about a season (Fig. 2; Angell 1997); the ITCZ then gradually returns to its pre-perturbation position over a period of 4 years or longer, which exceeds the period of direct stratospheric aerosol forcing (~ 3 years). The shifts of the ITCZ affect sea surface temperatures (SSTs), especially in the Pacific, leading to enhanced (weakened) equatorial zonal SST gradients following SH (NH) events. These variations are superimposed on concomitant ENSO-like variations, typically with El Niño-like conditions appearing during the second year after eruptions, and La Niña-like conditions after the third year (Fig. 7; Stevenson et al. 2016).

A smooth response of the ITCZ is seen following SH and NH events, despite contrasting variations in cross-equatorial energy transport by the stratosphere and troposphere, as well as by the eddy and mean components of atmospheric energy

FIG. 14. Lead-lag plot of the regression coefficients of the oceanic Niño index (ONI) and cross-equatorial energy transport quantities: (a) total cross-equatorial meridional energy transport (TOA_0 ; yellow), given by the sum of cross-equatorial energy transport by the atmosphere (AET_0 ; green) and ocean (OET_0 ; purple); (b) advected OET (OET_0^{flux} ; orange), and cross-equatorial energy transport implied by interhemispheric differences in ocean energy storage (OET_0^{storage} ; gray) and atmospheric energy storage (AET_0^{storage} ; blue); (c) cross-equatorial atmospheric eddy energy transport (EET_0 ; lilac), and mean cross-equatorial atmospheric energy transport (AET_0 ; green), given by mean cross-equatorial transport in the troposphere (TET_0 ; bright blue) and stratosphere (SET_0 ; maroon).

transport (AET). Specifically, in the eruption hemisphere, stratospheric sulfate aerosols warm the stratosphere by increased absorbance of longwave radiation but cool the surface and troposphere by increased shortwave reflection, leading to contrasting energy transport anomalies in the stratosphere and troposphere. Consequently, the Hadley circulation cannot be assumed to dominate AET, and the link between the positions of the ITCZ and EFE (derived from net AET) weakens. Indeed, unlike Colose and LeGrande (2016) who found strong correlation between ITCZ and EFE variations in the second year following eruptions, we find contrasting shifts of the ITCZ and EFE during the first year (Fig. 11a). Moreover, due to the balancing between stratospheric and tropospheric energy transport, ocean energy transport (OET), driven by radiative forcing and by wind stress anomalies, dominates the total interhemispheric

energy transport during the first year (Fig. 11b). Ocean turbulent transport and energy storage have a critical role in the transient OET response, which is not well understood on seasonal time scales. As the asymmetric forcing by stratospheric aerosols subsides, tropospheric energy transport becomes the dominant transporter of anomalous interhemispheric energy fluxes, causing the EFE and ITCZ responses to converge.

ENSO variations are found to be associated with cross-equatorial atmospheric and oceanic energy fluxes in the CESM LME (Fig. 14). The ENSO-like variations following eruptions therefore modulate the transient shifts of the ITCZ by affecting both surface conditions and the energetic constraints on the ITCZ response. Specifically, Niño (Niña) conditions are found to be associated with net northward (southward) energy transport, implying that the ENSO response weakens (amplifies) NH (SH) events during the first year and amplifies (weakens) NH (SH) events after the third year. The contrasting mean and eddy atmospheric responses during the first year (Fig. 11) may also be partly attributable to the ENSO-like response (14). Nevertheless, the energetic relation to ENSO may be sensitive to the representation of ENSO in CESM (Tan et al. 2020), and should be tested in other datasets.

The relay of stratosphere–troposphere–ocean energy transport, modulated by ENSO-like variations, therefore governs the time-dependent response of the ITCZ, which can be divided into three stages, as illustrated in Fig. 15:

- 1) First year: direct response to stratospheric aerosol forcing. The ITCZ shifts away from the eruption hemisphere. Anomalous total interhemispheric energy transport is dominated by ocean energy transport (OET), due to the cancellation by the contrasting stratospheric (SET) and tropospheric energy transport (TET). Due to the northward mean position of the ITCZ, zonal SST gradients in the Pacific increase following SH events (Figs. 4d, 7a), amplifying the poleward shift of the ITCZ due to decreased NEI_0 [Fig. 10, Eq. (2)]. Similarly, equatorward shifts of the ITCZ following NH events are amplified by reduced zonal SST gradients and increased NEI_0 .
- 2) Second year: as stratospheric energy transport subsides with decreasing aerosol forcing, interhemispheric energy transport is increasingly dominated by TET. Therefore, the correlations between the positions of the ITCZ and EFE, as well as with cross-equatorial AET, strengthen. El Niño-like conditions appear. The El Niño-like conditions support northward cross-equatorial energy transport thereby amplifying (weakening) SH (NH) events (Fig. 14). Increased NEI_0 , associated with El Niño conditions, supports the equatorward shift of the ITCZ following NH events and damps the northward shift of the ITCZ following SH events.
- 3) Third to fifth years (or longer): the hemispherically asymmetric response subsides; La Niña-like conditions typically appear, modulating energy transport and ITCZ shifts, in opposite manner to the second year. Interhemispheric differential heating is sustained by sea ice feedbacks.

Much of the above characterizations hold for the response following both NH and SH events. We nevertheless note several important differences between the NH and SH events in the data analyzed here. First, surface cooling is generally

FIG. 15. Depiction of the relation of ITCZ shifts to stratospheric, tropospheric and oceanic energy transport during three phases following asymmetric (SH) eruptions. Orange (blue) colors indicate warming (cooling). Year 1: the ITCZ shifts away from the eruption hemisphere. Ocean energy transport dominates interhemispheric energy transport due to the contrasting stratospheric and tropospheric energy transport anomalies. Year 2: asymmetric aerosol forcing subsides in the stratosphere. The troposphere becomes the main transporter of interhemispheric energy anomalies, and the ITCZ and EFE covary. El Niño-like warming in the Pacific typically appears. Years 3–5: as interhemispheric temperature difference gradually decays, the ITCZ converges to its pre-eruption position. La Niña-like conditions typically appear in the Pacific.

stronger following NH events (Fig. 4a versus Fig. 4d, and Fig. 8b versus Fig. 8c), due to the larger land area in the NH, and due to sea ice feedbacks (Fig. 12e,f). Similarly, the changes in precipitation are slightly more pronounced following NH events (Fig. 3a versus Fig. 3d), especially over monsoonal regions, associated with reduced cross-equatorial moisture divergence, which weakens the summer monsoon in the eruption hemisphere (Liu et al. 2016). Furthermore, while the surface cooling following NH events is mostly confined to the eruption hemisphere, the cooling following SH events extends to the opposite hemisphere (Fig. 8c). This is consistent with the ITCZ blocking effect (Kang et al. 2020), where the mean position of the ITCZ north of the equator favors penetration of SH anomalies into the NH. Nevertheless, differences in the degree of asymmetry between NH and SH events are partly attributable to the difference in the asymmetry of the forcing (Fig. 2b), which may be related to the larger number of volcanoes in NH high latitudes, but may also be related to differences in sulfate distribution associated with the atmospheric circulation, in particular due to the dominance of events following boreal spring (Stevenson et al. 2017).

The strong correlation of the transient ITCZ response to tropospheric energy transport indicates that the theoretical underpinning of the energetic constraints holds true following volcanic eruptions. However, several limitations in the application of these constraints remain. First, the assumption that tropical energy transport is dominated by the Hadley circulation (i.e., that stratospheric and eddy transports are negligible in the tropics) is not valid following eruptions (Roberts et al. 2017; Xiang et al. 2018; Afargan-Gerstman and Adam 2020). Second, the processes controlling the lagged response of the ITCZ to anomalies in the atmospheric energy budget are poorly understood. Third, the dependence of the ITCZ on coupled ocean-atmosphere processes is not well understood, especially on seasonal time scales. In particular, we find that

wind-driven Ekman energy transport, generally assumed to constrain shifts of the ITCZ (Schneider et al. 2014; Schneider 2017; Green and Marshall 2017; Kang 2020), has a negligible role in the anomalous transient ocean energy transport.

Surprisingly, despite the complex relay of energy between the different atmospheric and oceanic components, we find that the total interhemispheric energy transport (TOA_0 , derived from top of atmosphere radiative balance) is a potentially useful predictor of ITCZ shifts following eruptions (Fig. 11a). Given the complex and nonmonotonic responses of the atmospheric and oceanic components (Figs. 11 and 12), assessing the generality of the empirical TOA constraint requires further testing. Nevertheless, as TOA energy balance variations precede ITCZ shifts by several weeks, constraints based on readily available TOA data may offer a pathway for practical real-time predictions of ITCZ shifts following eruptions.

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Data availability statement. The CESM LME data are publicly available via the Earth System Grid (www.earthsystemgrid.org). Stratospheric sulfate loading data can be obtained from <http://climate.envsci.rutgers.edu/IV12/> (original dataset 2).

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