



RESEARCH ARTICLE

10.1002/2016JD025796

Key Points:

- In our simulations the likelihood of El Niño-like responses almost doubles after a Pinatubo-size eruption
- The central Pacific El Niños develop much stronger warming responses to volcanic forcing than the eastern Pacific El Niños
- El Niño responses are highly sensitive to the seasonal timing of a Pinatubo-size eruption

Correspondence to:

E. Predybaylo,
evgeniya.predybaylo@kaust.edu.sa

Citation:

Predybaylo, E., G. L. Stenchikov, A. T. Wittenberg, and F. Zeng (2017), Impacts of a Pinatubo-Size Volcanic Eruption on ENSO, *J. Geophys. Res. Atmos.*, 122, 925–947, doi:10.1002/2016JD025796.

Received 16 AUG 2016

Accepted 12 JAN 2017

Accepted article online 16 JAN 2017

Published online 30 JAN 2017

Impacts of a Pinatubo-size volcanic eruption on ENSO

Evgeniya Predybaylo¹, Georgiy L. Stenchikov¹, Andrew T. Wittenberg², and Fanrong Zeng²
¹Physical Science and Engineering, King Abdullah University of Science and Technology, Thuwal, Saudi Arabia, ²NOAA Geophysical Fluid Dynamics Laboratory, Princeton, New Jersey, USA

Abstract Observations and model simulations of the climate responses to strong explosive low-latitude volcanic eruptions suggest a significant increase in the likelihood of El Niño during the eruption and posteruption years, though model results have been inconclusive and have varied in magnitude and even sign. In this study, we test how this spread of responses depends on the initial phase of El Niño–Southern Oscillation (ENSO) in the eruption year and on the eruption's seasonal timing. We employ the Geophysical Fluid Dynamics Laboratory CM2.1 global coupled general circulation model to investigate the impact of the Pinatubo 1991 eruption, assuming that in 1991 ENSO would otherwise be in central or eastern Pacific El Niño, La Niña, or neutral phases. We obtain statistically significant El Niño responses in a year after the eruption for all cases except La Niña, which shows no response in the eastern equatorial Pacific. The eruption has a weaker impact on eastern Pacific El Niños than on central Pacific El Niños. We find that the ocean dynamical thermostat and (to a lesser extent) wind changes due to land-ocean temperature gradients are the main feedbacks affecting El Niño development after the eruption. The El Niño responses to eruptions occurring in summer are more pronounced than for winter and spring eruptions. That the climate response depends on eruption season and initial ENSO phase may help to reconcile apparent inconsistencies among previous studies.

1. Introduction

Volcanic radiative impacts are important climate drivers on multiple time scales [Robock, 2000; Stenchikov, 2009; Timmreck, 2012; Meehl et al., 2015; Timmreck et al., 2016]. Large explosive volcanic eruptions inject sulfur-rich gases into the stratosphere [Newhall and Self, 1982; Schnetzler et al., 1997], where they get converted into sulfate aerosols [Turco et al., 1983; Lamb, 1970; LeGrande et al., 2016]. These aerosols scatter light in ultraviolet and visible spectra, absorb and scatter in the near infrared, and absorb, scatter, and emit thermal longwave radiation [Lacis et al., 1992; Hansen et al., 1997; Stenchikov et al., 1998], affecting the energy balance of the planet. As a result, the global mean surface temperature cools, and the lower stratosphere heats up [Minnis et al., 1993; Stenchikov et al., 1998; Rind and Lacis, 1993; Turco et al., 1983; De Silva and Zielinski, 1998; Briffa et al., 1998; Santer et al., 2014]. The associated redistribution of radiative heating directly impacts atmospheric circulation [Rind et al., 1992; Stenchikov et al., 2006] and cools the ocean [Church et al., 2005; Gleckler et al., 2006; Stenchikov et al., 2009; Otterå et al., 2010], producing global and regional changes in the Earth's climate system [Otterå et al., 2010; Fischer et al., 2007; Haywood et al., 2013] and affecting the major modes of climate variability. Impacts of volcanic eruptions on the North Atlantic/Arctic oscillation have been a subject of active research over the past 15–20 years [Robock and Mao, 1995; Stenchikov et al., 2002, 2006].

The El Niño–Southern Oscillation (ENSO) is one of the most important climate variability modes, which controls the climate not only in the equatorial Pacific [Soden, 2000] but also impacts the entire globe [Brönnimann et al., 2007; Ineson and Scaife, 2009; Graf and Zanchettin, 2012]. It perturbs the hydrological cycle [Soden, 2000], multidecadal biological and biogeochemical cycles of the ocean [Chavez et al., 2003; Yoder and Kennelly, 2003], and affects hurricane [Goldenberg et al., 2001; Gray, 1984; Vecchi et al., 2014] and tornado [Lee et al., 2016] activity and precipitation patterns [Ropelewski and Halpert, 1987, 1996; Ratnam et al., 2014; Jia et al., 2015]. ENSO causes an anomalous change in the air-sea interaction in the equatorial Pacific every 2–7 years [Trenberth, 1997; D'Arrigo et al., 2005]. The positive (El Niño) and negative (La Niña) phases are asymmetric in magnitude, and have different spatial-temporal appearance and intensity [Choi et al., 2013, 2015]. Generally, El Niños have a strong stochastic component associated with westerly wind bursts (WWBs), while La Niñas are more likely to occur immediately after strong El Niños. For this reason, it is usually easier to predict La Niñas than El Niños.

© 2017. The Authors.

This is an open access article under the terms of the Creative Commons Attribution-NonCommercial-NoDerivs License, which permits use and distribution in any medium, provided the original work is properly cited, the use is non-commercial and no modifications or adaptations are made.

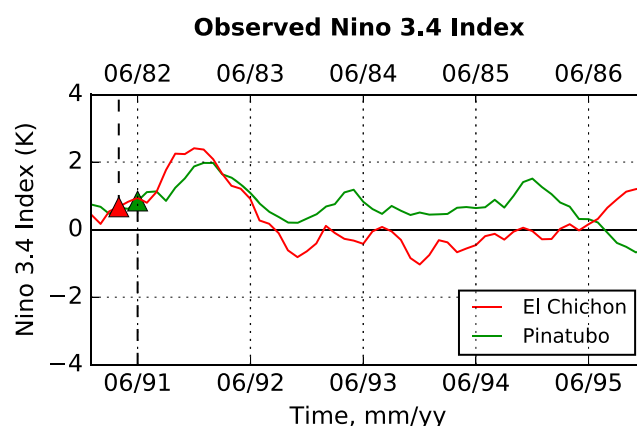


Figure 1. Observed Niño 3.4 index (K) evolution after the El Chichón (red, top x axis) and Pinatubo (green, bottom x axis) eruptions, calculated using the Extended Reconstructed SST (ERSST) V4 data set. Eruption dates are marked with triangles.

Although predictions and projections of ENSO remain a challenge due to the strong nonlinearity of ENSO processes and their high sensitivity to external forcing [Wittenberg, 2002; Collins *et al.*, 2010; DiNezio *et al.*, 2012; Watanabe *et al.*, 2012; Choi *et al.*, 2013; Lee *et al.*, 2014; Wittenberg *et al.*, 2014], accurate future projections of ENSO behavior are crucial to assess future climate risks [Vecchi and Wittenberg, 2010; Capotondi *et al.*, 2015; Wittenberg, 2015; Guilyardi *et al.*, 2016], and for societal decision making [Cash *et al.*, 2006].

Most of the largest eruptions of the twentieth century occurred in El Niño years, e.g., El Chichón in April 1982, and Pinatubo in June 1991 (Figure 1). It was confirmed recently that the Tambora

eruption, which produced about 3 times more sulfur dioxide than Pinatubo and caused the “Year without a summer” in 1816, was also accompanied by an El Niño [Raible *et al.*, 2016]. The nature of these relationships is not well understood, but they have dramatic consequences for the entire planet—thus, it is important to better investigate the mechanism of volcanic impacts on ENSO [Stenchikov, 2009; Li *et al.*, 2011; Timmerreck, 2012; Wittenberg *et al.*, 2014], which could explain some of its temporal modulation in historical and paleorecords, and their relation to internal ENSO dynamics [Emile-Geay *et al.*, 2008; Vecchi and Wittenberg, 2010; Emile-Geay *et al.*, 2013; McGregor *et al.*, 2013; Ogata *et al.*, 2013; Atwood *et al.*, 2016].

Given the brevity of in situ and satellite observational records, the actual volcanic forcing impact on ENSO cannot easily be empirically determined. Studies based on paleodata [Adams *et al.*, 2003; McGregor *et al.*, 2010; Wahl *et al.*, 2014; Li *et al.*, 2013] detected a remarkable shift in tropical Pacific climate in postvolcanic years toward an El Niño-like state or a multiyear El Niño. The direct effect of volcanic forcing cools the surface; e.g., Li *et al.* [2013] emphasized the importance of tropical Pacific sea surface temperature (SST) cooling shortly after the eruption. This cooling takes place prior to the development of an extra El Niño-like warming the year after an eruption. The physical interpretation of the initial cooling is still under question. McGregor and Timmermann [2011] captured this phenomenon using the Community Climate System Model (CCSM3); however, in their study the amplitude of simulated cooling was overestimated, and the subsequent warming was quantitatively inconsistent with temperatures inferred from proxy records.

In one of the first modeling studies on volcanic impacts on ENSO, Hirano [1988] suggested that absorption of solar and terrestrial radiation by volcanic aerosols led to atmospheric heating, which produced a wind anomaly that triggered an El Niño event. This interaction was further studied by Robock *et al.* [1995] with the help of an atmospheric general circulation model (GCM) CCM1. Robock *et al.* [1995] calculated the effect of the El Chichón eruption and concluded that at the time of the eruption, the El Niño in the spring of 1982 was already underway, so it was not caused by the eruption; however, volcanic forcing might have affected the El Niño amplitude.

Mann *et al.* [2005] and Emile-Geay *et al.* [2008] studied volcanic impacts on ENSO, using the simplified coupled atmosphere-ocean model of Zebiak and Cane [1987]. Emile-Geay *et al.* [2008] performed large ensemble experiments testing the tropical Pacific response to strong volcanic forcing. They found that only very powerful eruptions of more than an order of magnitude stronger than Pinatubo could lead to a correlation between volcanic forcing and El Niño and therefore affect El Niño likelihood and/or magnitude. The simplicity of the Cane-Zebiak model precluded a reliable quantitative determination of the level of volcanic forcing needed for an ENSO response, leaving uncertain whether Pinatubo was above or below this threshold. Both papers, however, suggested that strong volcanic forcing affects ENSO and tropical Pacific climate via the ocean dynamical thermostat (ODT) mechanism [Seager *et al.*, 1988; Clement *et al.*, 1996], in which surface perturbations are displaced by upwelling of deeper, unperturbed waters.

Ohba *et al.* [2013] confirmed the findings of Adams *et al.* [2003] and McGregor *et al.* [2010] using an interim version of the Model for Interdisciplinary Research on Climate (MIROC) [Watanabe *et al.*, 2010]. They investigated the sensitivity of ENSO to volcanic forcings of realistic strength ($1.5 \times$ Pinatubo and 0.5, 1.5, and 2 scaling of that value) as well as the background ENSO phase: neutral, positive, and negative. They suggested that the ODT is not the sole mechanism affecting the SST response; there is also a strong contribution of the atmospheric response to the changes in the land-ocean temperature gradient in the Western Pacific (WP).

Maher *et al.* [2015] analyzed the tropical Pacific climate state in the composite Coupled Model Intercomparison Project phase 3 (CMIP3) and CMIP5 historical simulations after the five strongest eruptions. They also found a tendency toward an El Niño-like (La Niña-like) SST response in the first (third) year after an eruption. However, only a third of the examined models were able to simulate a realistic ENSO [Kim *et al.*, 2014; Kim and Jin, 2011].

The most recent studies on the topic [Pausata *et al.*, 2015; Lim *et al.*, 2015; Stevenson *et al.*, 2016] discussed a shift of the Intertropical Convergence Zone (ITCZ) as an alternative mechanism of volcano-El Niño interaction, induced by a very strong (more than 15 W m^{-2}) radiative forcing that is asymmetric with respect to the equator. Stevenson *et al.* [2017] used the Community Earth System Model (CESM) to study responses to volcanic eruptions occurred in January, April, July, and October, calculating aerosol distributions interactively within the model. They found a cooling response during the first 6 months after an eruption, which they interpreted as resulting from reduced downward shortwave flux (especially in the relatively cloud-clear east Pacific) which strengthened the zonal SST gradient along the equator, and shifted the ITCZ northward. The initial La Niña-like cooling in the CESM was then followed by an El Niño-like warming of the east Pacific, which Stevenson *et al.* [2017] suggested being caused by an induced off-equatorial anticyclonic wind stress curl that forced an equatorward Sverdrup transport of heat in the upper ocean.

Thus, the sensitivity studies conducted so far [Mann *et al.*, 2005; Emile-Geay *et al.*, 2008; McGregor and Timmermann, 2011; Ohba *et al.*, 2013; Lim *et al.*, 2015; Stevenson *et al.*, 2017] reveal that simulation results are model dependent and do not fully illuminate the mechanisms of volcanic impacts on ENSO.

Many studies [Ashok *et al.*, 2007; Kug *et al.*, 2010; Lee *et al.*, 2014; Chen *et al.*, 2015; Capotondi *et al.*, 2015; Chen *et al.*, 2016] have highlighted the diversity of ENSO events, mechanisms, and impacts. The Pinatubo eruption coincided with a moderate Central Pacific (CP) El Niño that lasted for about 2 years [Kessler and McPhaden, 1995], and the eruption of El Chichón was accompanied by a very strong eastern Pacific (EP) El Niño that behaved differently from that in the Pinatubo case. Here we hypothesize that the initial ENSO state, including the different El Niño types, may play a key role in the diverse tropical Pacific responses to volcanic eruptions. We focus on the following questions:

1. What causes the diversity of ENSO responses to Pinatubo-size volcanic forcing in observations and model simulations?
2. What atmospheric and oceanic feedbacks tend to amplify or damp the ENSO response?
3. How do ENSO responses and feedbacks depend on the preconditioning of the tropical Pacific climate system?
4. How sensitive is ENSO to small perturbations, and how different might ENSO responses be to volcanic eruptions occurring at different times of year?

2. Methodology

To answer the above questions, we employ a global coupled ocean-atmosphere GCM, CM2.1, developed by the Geophysical Fluid Dynamics Laboratory (GFDL) [Delworth *et al.*, 2006], which was used for the Coupled Model Intercomparison Project phase 3 (CMIP3) and the Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC AR4), as well as in our previous studies on the volcanic impact on atmospheric and oceanic circulations involving the Arctic Oscillation and Atlantic Meridional Overturning Circulation [Stenchikov *et al.*, 2006; Stenchikov, 2009; Stenchikov *et al.*, 2009].

2.1. Model Description

Here we briefly summarize the formulation of the CM2.1 global coupled GCM. Its atmospheric component [Anderson *et al.*, 2004] has a horizontal grid spacing of 2° latitude by 2.5° longitude, with 24 vertical levels and a finite volume dynamical core [Lin, 2004]. The land surface component [Milly and Shmakin, 2002] has the same horizontal resolution as the atmospheric component. The ocean component [Griffies *et al.*, 2005; Gnanadesikan *et al.*, 2006] is implemented on a tripolar horizontal grid, with zonal spacing of 1° and meridional

Table 1. Comparison of ENSO Phase Probabilities in the Observational Data and CM2.1 Model Output for Neutral ENSO, Central Pacific El Niño, Eastern Pacific El Niño, and La Niña

	Neutral ENSO	CP El Niño	EP El Niño	La Niña
Observations	0.43	0.23	0.11	0.23
CM2.1	0.49	0.16	0.08	0.27

spacing telescoping from 1° at high latitudes to 1/3° near the equator. The ocean model has 50 vertical levels, with 10 m spacing over the top 220 m. The ocean and atmosphere are coupled every 2 h. The atmospheric composition, incoming solar radiation, and land cover are kept at the 1990 level.

CM2.1's tropical Pacific and ENSO simulation characteristics have been extensively discussed [e.g., Wittenberg *et al.*, 2006; Wittenberg, 2009; Kug *et al.*, 2010; Wittenberg *et al.*, 2014; Karamperidou *et al.*, 2014; Atwood *et al.*, 2016; Chen *et al.*, 2016]. While the simulated SST, winds, surface fluxes, and oceanic subsurface temperature do have biases in some regions, they generally agree well with observations. Wittenberg *et al.* [2006] and Wittenberg [2009] showed that CM2.1 captures the main aspects of tropical Pacific climate and ENSO. In addition, Kim and Jin [2011] showed that CM2.1 is one of the few models able to produce a stable, realistic ENSO under various external forcing perturbations. Kug *et al.* [2010] and Capotondi *et al.* [2015] discussed CM2.1's ability to successfully reproduce realistic CP and EP El Niño patterns and frequencies (Table 1).

2.2. Experimental Setup

Generally, ENSO comprises El Niño, La Niña, and neutral phases. However, El Niños can be of multiple types that can be split roughly into CP and EP groups [Ashok *et al.*, 2007; Kug *et al.*, 2010; Lee *et al.*, 2014; Chen *et al.*, 2015; Capotondi *et al.*, 2015; Chen *et al.*, 2016]. CP and EP El Niño types are characterized by a distinct genesis. Observations show that weak and moderate El Niños mostly tend to be of the CP type, while the strong El Niños usually follow a canonical EP pattern [Rasmusson and Carpenter, 1982; Zheng *et al.*, 2014; Fang *et al.*, 2015]. Kug *et al.* [2010] showed that the formation of the moderate El Niño is the product of zonal advection, while the strong El Niño involves a greater role for vertical advection. Chen *et al.* [2015] argued that chaotically generated WWBs play a key role in El Niño formation. WWBs occur sporadically during November (in the year before the event)–May (in the year of the event) and strongly impact ENSO variability [Vecchi *et al.*, 2006]. WWBs have been shown to play a role in triggering and amplifying El Niño [Gebbie *et al.*, 2007; Zavala-Garay *et al.*, 2008; Wittenberg *et al.*, 2014; Atwood *et al.*, 2016], as well as affecting its type [Chen *et al.*, 2015].

The development of a moderate El Niño is initiated by WWBs that cause eastward advection of warm water toward the CP. Consequently, the SST gradient between the WP and CP decreases and results in a reduction of the easterly trade winds over the WP. This causes CP SST warming via the Bjerknes feedback mechanism [Bjerknes, 1969] and further decreases the SST gradient. The resultant CP El Niño onset slightly reduces the EP upwelling but does not shut it down completely, allowing the EP to remain cold.

Strong WWBs frequently serve as forerunners of extremely strong El Niño events, by driving the multiple oceanic Kelvin wave pulses that accumulate warm water in the EP. Thus, an expanded warm pool develops early in September–October of an El Niño year, and SST anomalies peak in the boreal winter near the eastern boundary, almost completely shutting down the equatorial upwelling. The related equatorial Pacific westerly wind anomaly is greater for EP than CP El Niños, due to a stronger response of the zonal surface pressure gradient. CM2.1 generally captures all these features, though Wittenberg *et al.* [2006] noted that in CM2.1 westerly wind anomalies formed due to the Bjerknes effect are located farther west than in the real world.

In our simulations, we examine the impact of Pinatubo forcing on different El Niños, which tend to peak around December of 1991, coinciding with the peak in the Pinatubo volcanic forcing that develops half a year after the eruption. Below we refer to 1991 as the initial year of the simulation or the year of eruption, and 1992 and 1993 as the first and second years after the eruption, respectively.

Having noted the above differences in amplitude, spatial pattern, and genesis of the CP and EP El Niños, it is important to study the different ways in which EP and CP El Niños respond to an eruption. We consider a complete set of initial conditions (ICs) to cover the possible initial ENSO phases that could occur in the year of an eruption. We do not suggest a causal relationship between volcanic eruptions and ENSO at the time of eruption; instead, we assume that the eruption could happen in neutral ENSO, El Niño (CP or EP), or La Niña conditions, which occur with different probabilities (see Table 1). We then examine the evolution of the volcanically perturbed ENSO probability distribution, relative to that of unperturbed or infinitesimally perturbed simulations.

Table 2. Summary of the Experiments Used in Sections 3.1–3.3 and 3.4

Name	ENSO Type	ENSO Strength	Ensemble Members	Pinatubo Forcing Coefficient	Forcing Start Month
<i>Sections 3.1–3.3</i>					
NTE CTR	Neutral	Neutral	10	0	-
NTE PRT			10	1	June
CPE CTR	Central Pacific	Moderate	10	0	-
CPE PRT	El Niño		10	1	June
EPE CTR	Eastern Pacific	Strong	10	0	-
EPE PRT	El Niño		10	1	June
LAN CTR	La Niña	Moderate	10	0	-
LAN PRT			10	1	June
<i>Section 3.4</i>					
CTR ₀			1	0	-
CTR _α ^{Feb}	Eastern		50	[0.001, ..., 0.05]	February
PRT _α ^{Feb}	Pacific	Strong	50	[1.001, ..., 1.05]	February
CTR _α ^{Jun}	El Niño		50	[0.001, ..., 0.05]	June
PRT _α ^{Jun}			50	[1.001, ..., 1.05]	June

In order to simulate the volcanic perturbation, we prescribe the Pinatubo aerosols' optical properties according to *Stenchikov et al.* [2006, 2009] using optical depth from *Sato et al.* [1993]. This volcanic data set provides zonally averaged monthly mean spectral dependent aerosol extinction, single scattering albedo, and an asymmetry parameter which are required to conduct radiative transfer simulations within CM2.1. The experimental design allows the corresponding ENSO conditions to freely develop before the eruption, exactly as in the present-day control run, and then the Pinatubo forcing is applied in June 1991. The control and perturbed runs first diverge in mid-May 1991, because the monthly mean aerosol characteristics are interpolated between the months.

We conducted four 10-member ensemble integrations of control (CTR, without volcanic aerosols) and perturbed (PRT, with volcanic aerosols) experiments with ENSO-neutral, La Niña, CP El Niño, and EP El Niño occurring in the initial year of each control experiment (see Table 2). The ensemble of 10 different ICs from each ENSO category is intended to capture the intrinsic variability of the climate system. To increase the signal-to-noise ratio, we present our results using ensemble means and use the ensemble spread to calculate statistical significance. We implement three methods of calculating anomalies. Based on the climatology computed from the present-day control run conducted by *Delworth et al.* [2006], using monthly averages over years 101–200 of the 300 year simulation, we calculate the control and perturbed anomalies by subtracting the climatology from the control and perturbed runs, respectively. CTR-CLM represents a pure ENSO signal, and the difference PRT-CTR (which we call the “response”) shows the ENSO response to volcanic forcing, as the “control ENSO” effect is removed. Statistical significance of the ensemble mean differences is computed using a two-tailed Student's *t* test at the 0.05 significance level.

We start each ensemble simulation from 1 January 1991 with the coupled atmosphere-ocean IC extracted from the 300 year present-day control run. The ENSOs present in this run were categorized by their strength and associated spatial pattern according to the threshold values of the boreal winter (October/November/December, OND) NINO3.4 index. The chosen ENSOs are then combined into 10-member control ensembles of ENSO-neutral (with OND Niño3.4 < 0.15 K), weak/CP El Niño (with OND Niño3.4 index range 1.4–1.8 K), strong/EP El Niño (with OND Niño3.4 index range 3.4–4.7 K), and La Niña (with OND NINO3.4 < –1 K) named NTE, CPE, EPE, and LAN ensembles, respectively. Figure 2 shows control 10-member ensemble mean surface temperature anomalies, wind stress anomalies, and total precipitation in December of 1991 for all experiments.

2.3. Observational Data

For calculations of the observed Niño 3.4 index (Figure 1), we use monthly mean SSTs from the NOAA Extended Reconstructed SST (ERSST) V4 data set [*Huang et al.*, 2015]. The data have a spatial resolution of 2° latitude

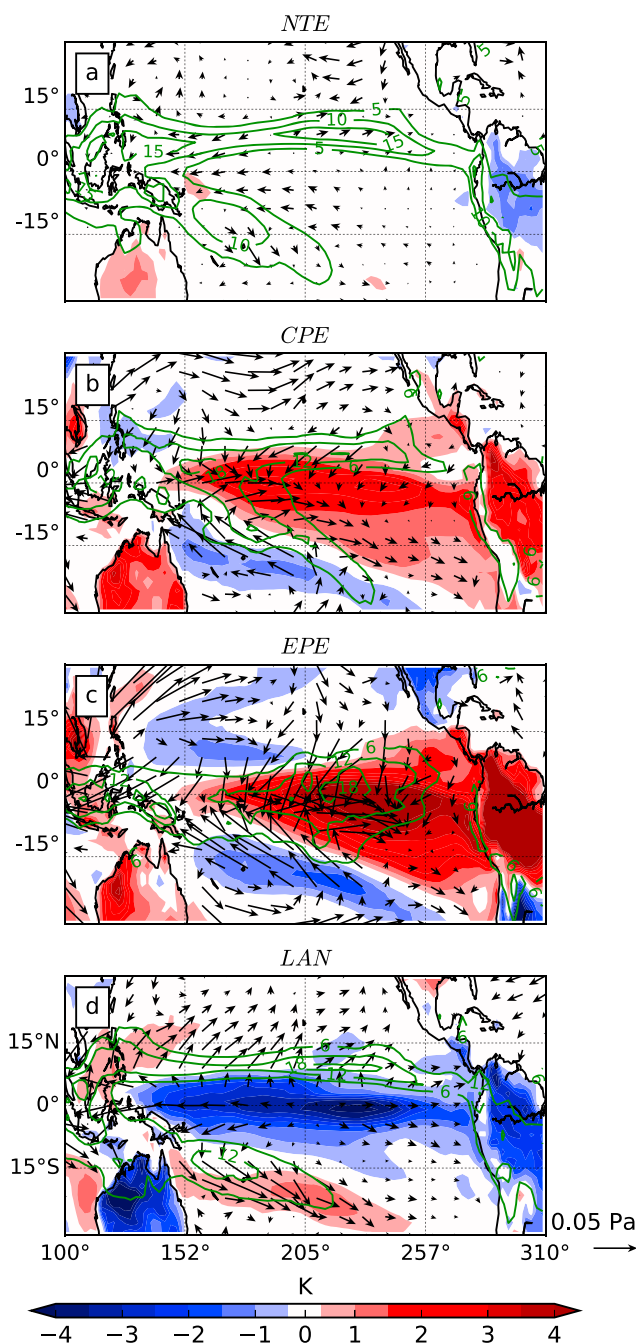


Figure 2. Simulated surface temperature (K, shading) and surface wind stress vector (Pa, arrows) anomalies with respect to the climatology (CTR-CLM) and total precipitation (mm day⁻¹, contours) of the 10-member control ensemble mean ENSO phases in December 1991: (a) neutral ENSO (NTE), (b) central Pacific El Niño (CPE), (c) eastern Pacific El Niño (EPE), and (d) La Niña (LAN).

by 2° longitude and cover January 1854 up to present. The climatology is computed as monthly averages from the data for years 1880–2000. Figure 1 shows that the Niño 3.4 SST anomaly trajectories following the El Chichón (1982) and Pinatubo (1991) eruptions evolve quite differently. The El Niño of 1982 is stronger than that of 1991 but terminates sooner. The 1982 El Niño is followed by a prolonged La Niña, while the weaker and longer El Niño of 1991 is followed by prolonged El Niño warm anomalies.

The observed ENSO frequencies (Table 1) are calculated using the ERSST V4 data set for the period 1980–2010. CP El Niño events are identified following *Pascolini-Campbell et al.* [2015] method based on combination of

nine different SST-based indices (see Table 1 in *Pascolini-Campbell et al.* [2015]) complemented by the visual examination of spatial patterns.

3. Results

Imposing the prescribed volcanic aerosols in the model results in a reduction of the shortwave (SW) radiation reaching the surface. The longwave (LW) effect of volcanic aerosols at the surface is relatively small and insignificant [*Stenchikov et al.*, 1998]. Figure 3 represents the time evolution of the 10-member ensemble mean (PRT-CTR) all-sky net (down-up) surface SW radiative flux response, averaged globally and over the tropical belt (20°S–20°N) for NTE, CPE, EPE, and LAN ensembles. The net SW radiative surface response in the tropics peaks at -5.5 W m^{-2} in the fall of 1991, as the SO_2 mass conversion e-folding time is about 35 days, and the aerosol layer fully develops by then [*Stenchikov et al.*, 1998]. The global net SW radiative surface response of -3 W m^{-2} peaks 6 months after the eruption, because of the interaction of conversion and transport processes. In the CPE and EPE experiments, we see weaker forcing in the winter of 1991, due to El Niño-induced increases in cloud cover. The volcanic radiative effect is present for more than 2 years and is quite similar in the different ensembles—apart from incoherent chaotic fluctuations among the different cases, which are not completely removed by the 10-member ensemble mean.

Figure 4 depicts the global and tropical mean surface temperature responses (PRT-CTR) to volcanic forcing. Half a year after the eruption the surface temperature decreases by about 0.4 K globally, which is in a good agreement with observations and other modeling studies [*Dutton and Christy*, 1992; *Hansen et al.*, 1992; *McCormick et al.*, 1995; *Soden et al.*, 2002]. The surface temperature relaxes slowly because of the high thermal capacity of the ocean. According to *Stenchikov et al.* [2009] it takes about a decade for the atmosphere and upper ocean system to equilibrate (not shown). In the tropics, the surface temperature fluctuations are higher and depend on the initial ocean state, demonstrating warmer anomalies in comparison with the global ones in the NTE and CPE cases during 1992 and beginning of 1993.

Figure 5 shows the 10-member ensemble mean perturbations of the Niño 3.4 SST. Red curves show the isolated effect of volcanic forcing on ENSO. The Pinatubo volcanic forcing causes a statistically significant increase of the Niño 3.4 index in summer 1992 in the NTE and CPE ensembles. The NTE ensemble transforms into a moderate-to-weak El Niño-like warming in the second winter after the eruption that lasts more than a year (Figure 5a); and in the CPE 10-member ensemble mean, El Niño extends for an extra year (Figure 5b). Distinctively, in the EPE ensemble the El Niño weakens in winter of 1991 and is followed by less of a prolongation of El Niño than in the NTE and CPE cases (Figure 5c). The LAN ensemble does not show any notable changes (Figure 5d), and for this reason we do not show the LAN responses further.

3.1. Volcanic Impacts at Different ENSO Phases

Figure 6 displays Hovmöller diagrams of SST, zonal wind, net surface heat flux, all-sky net (down-up) shortwave flux at the surface, latent heat flux, and total cloud amount responses (PRT-CTR) for the NTE, CPE, and EPE ensembles. Shortly after the eruption, the equatorial Pacific region is subject to a strong SW radiation reduction (Figures 6a, 6i, and 6q). The uniformly distributed clear-sky radiative forcing is modulated by clouds. All-sky radiative forcing drives climate changes and could itself cause about 0.3 K global SST cooling [*Stenchikov et al.*, 2009].

The land initially cools faster than the ocean, and the associated land-ocean temperature gradient (LOTG) both in the WP and EP generates zonal wind anomalies, which, in turn, affect ocean temperature in the initial stage of the process [*Ohba et al.*, 2013]. The ocean dynamical thermostat (ODT) mechanism [*Seager et al.*, 1988; *Clement et al.*, 1996] comes into play later, as the ocean responds more slowly to the radiative forcing. The ODT mechanism triggers nonuniform SST changes in the equatorial Pacific due to a differential ocean response to the spatially uniform radiative forcing. The efficiency and duration of these mechanisms depend on the background ocean state; therefore, ENSO responses appear to be different for different ICs (Figure 5).

3.1.1. NTE Ensemble

Figure 6b shows that in the NTE case during the first 3 months after the eruption, the LOTG (PRT-CTR) response in the WP is stronger than in the EP, because the EP ocean surface and nearby land areas in the background state are cold (Figure 2a). The developed WP LOTG decreases the trade winds in the WP (Figure 6c) west of 140°E. This westerly wind anomaly allows more warm water to advect eastward and favors WWBs. Both effects (enhanced westerlies and WWBs) are usually important forerunners of an El Niño event. However, because of

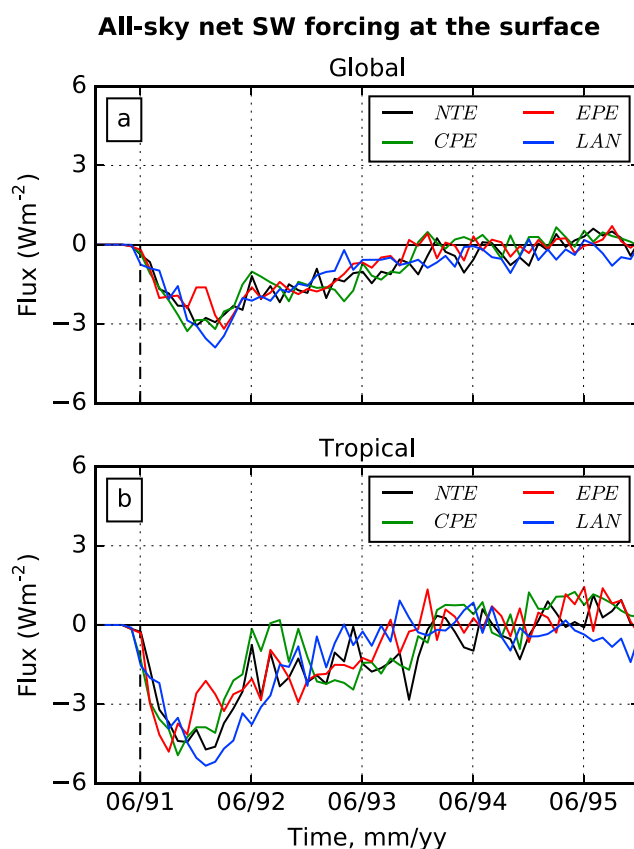


Figure 3. Simulated (a) global and (b) tropical (20°S – 20°N) all-sky net (down-up) shortwave (SW) radiative forcing at the surface (W m^{-2}) calculated as a 10-member ensemble mean difference (PRT-CTR) for neutral ENSO (NTE, black), central Pacific El Niño (CPE, green), eastern Pacific El Niño (EPE, red), and La Niña (LAN, blue) experiments.

the land's low thermal capacity, this LOTG mechanism is short lived and lasts for 2–3 months until the ocean temperature adjusts to the radiative forcing.

The SST changes are further supported by the ODT mechanism. Specifically, in Figure 6b we see a significant WP SST reduction starting in October of 1991, while the EP SST remains unperturbed. Such a spatially varying response of the ocean surface to a uniform atmospheric forcing is due to the upwelling which is strong in the CP and EP and regulates the SST there. Cooling of the WP SST increases sea level pressure (SLP, not shown) in the WP, enhances CP westerly wind anomalies, and results in a further reduction of the easterly trade winds (Figure 6c). This activates the Bjerknes feedback [Bjerknes, 1969] and leads to a positive SST anomaly in the EP (Figure 6b). It is important to note that the relatively weak El Niño-like SST response in the NTE case does not completely shut down the upwelling (Figure 2a); and therefore, the ODT mechanism is functioning throughout the period of volcanic forcing. Although the wind feedback is controlled by the ODT mechanism, it is also intensified by the eastward shift of the deep convection (Figure 6g), which tends to follow the warmest SSTs as explained in detail in Choi *et al.* [2015].

To evaluate the effect of the surface energy balance change, we calculate the net energy flux $\text{NF} = \text{SW} + \text{LW} + \text{SH} + \text{LH}$, where SW is the net (down-up) shortwave flux, LW is the net (down-up) longwave flux, SH is the sensible heat flux, and LH is the latent heat flux at the surface. The net surface heat flux response is negative in the areas with a positive SST response, thus tending to damp the latter. Figures 6d–6f show that the all-sky surface radiative and LH fluxes in the first and second years after the eruption mostly damp the SST response. The main contributors to the net energy flux anomaly are the SW and LH fluxes (see Figures 6e and 6f). The presence of clouds in the tropical Pacific changes the distribution of the incoming solar radiation: the all-sky SW response is negative over the warmer ocean areas, as the warmer ocean favors more shading by convective clouds. Similarly, the SW response is positive over the colder ocean areas, as the colder ocean favors less shading by convective clouds. This indicates that the SST response is partly damped by changes in

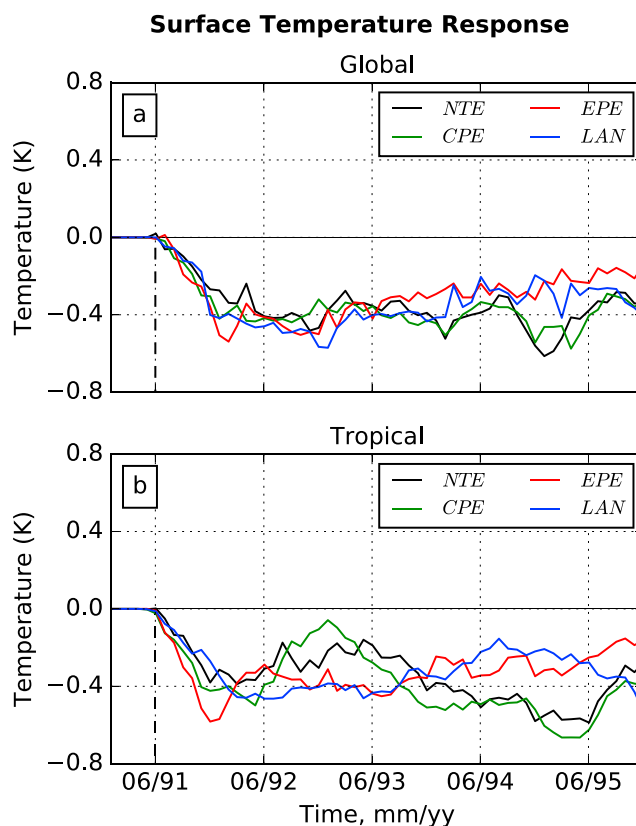


Figure 4. Same as in Figure 3 but for the surface temperature response (K).

overshooting deep convection. LH flux also responds to the SST changes and mostly works to damp the SST response. Compared to SW and LH fluxes, the effect of the LW and SH (not shown) on the net surface heat flux changes is small.

3.1.2. CPE Ensemble

In the initial stage associated with the fast land cooling during 2 or 3 months after the eruption, Figure 6j shows, in contrast to the NTE ensemble, a weak negative surface temperature response in the EP for the CPE ensemble. This is because the LOTG mechanism develops not only in the WP but also in the EP, since the control EP SST and nearby land are warmer than normal (see Figure 2b) despite the maximum SST is in the CP, thus enhancing westerly (easterly) wind anomaly in the WP (EP) (Figure 6k). This causes an eastward transport of warm water in the WP and intensification of the upwelling and cooling SST in the EP. As in the NTE ensemble, this effect is short lived. In the WP the net-flux cooling of ocean in 1991 (Figure 6l) turns on the ODT mechanism [Seager *et al.*, 1988; Clement *et al.*, 1996] starting from October 1991. Due to the CP position of the warm pool in the CPE case, the upwelling is partially suppressed, but still functions. Therefore, the ODT mechanism is slightly weaker in the CPE case than in the NTE case, but still is relatively long lasting. The positive SST response survives until the end of 1992, slightly shorter than in NTE case. As in the NTE ensemble, an eastward shift of deep convection (Figure 6o) enhances the response.

The westerly wind response, driven by the weakened WP SST gradient, pushes the equatorial Pacific toward El Niño-like conditions (as in the NTE case) and warms the EP SST (Figures 6j and 6k). The volcanically induced warming in the EP finally shuts down the upwelling and ODT in February of 1993. As in the NTE case, the cloudiness and net surface energy fluxes tend to damp the SST response (PRT-CTR) after May 1992.

According to the observational study by Li *et al.* [2010], at the time of the Mount Pinatubo eruption the moderate CP El Niño of 1991 also transitioned to EP in 1992, similar to what we find in our CPE simulations.

3.1.3. EPE Ensemble

The SST and atmospheric responses of the EPE case significantly deviate from those of the CPE and NTE cases. A much more pronounced EP cooling that starts soon after the eruption and lasts for more than half a year is a distinct feature of the EPE response (Figure 6r). A newly developed zonal SST gradient enhances the trade

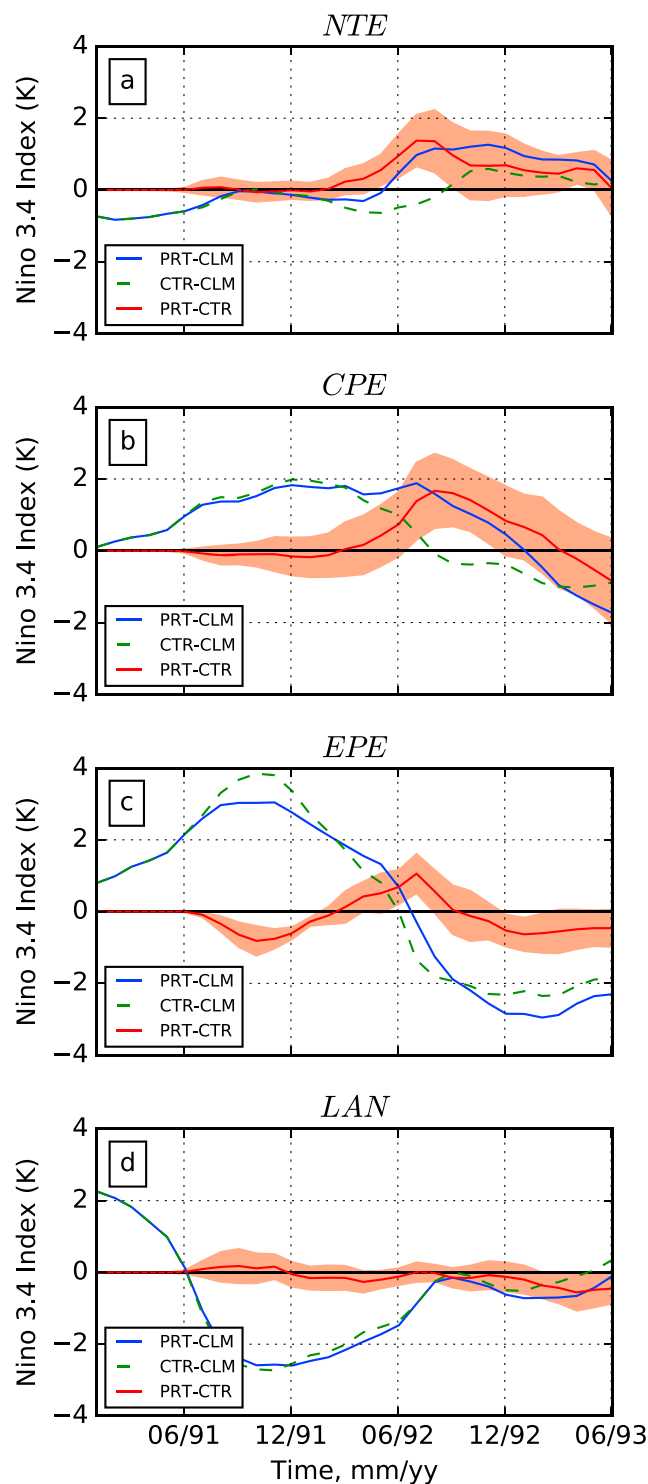


Figure 5. Simulated control (green dash) and perturbed (blue) Niño 3.4 index (K) calculated using 10-member ensemble means and difference between them (PRT-CTR) (red) in the (a) neutral ENSO (NTE), (b) central Pacific El Niño (CPE), (c) eastern Pacific El Niño (EPE), and (d) La Niña (LAN) experiments. Red shading depicts 95% confidence intervals for the ensemble mean difference.

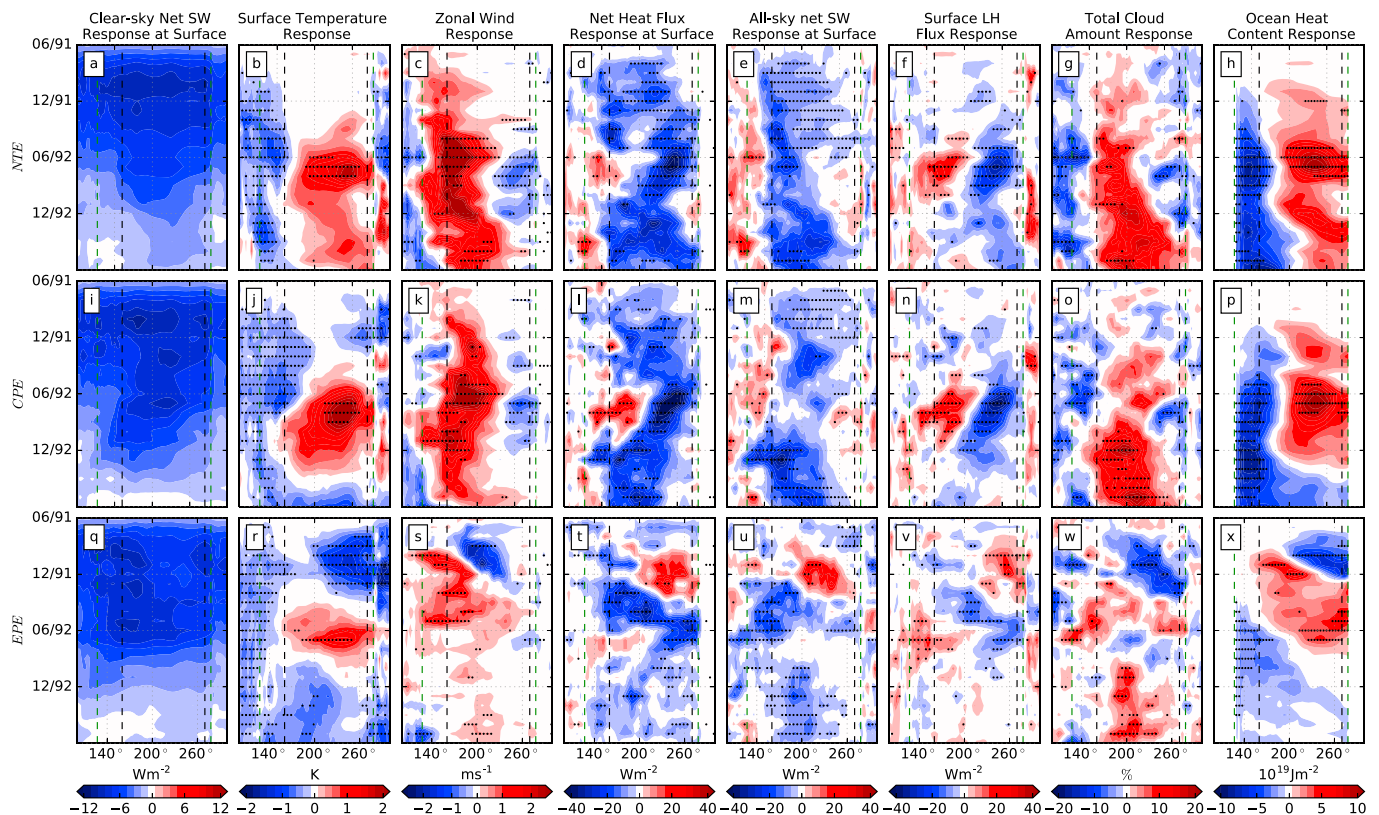


Figure 6. Hovmöller diagrams of simulated net (down-up) clear-sky shortwave (SW) surface flux (W m^{-2}), surface temperature (K), zonal wind (m s^{-1}), net surface heat flux, and all-sky net (down-up) shortwave (SW) fluxes at the surface, latent heat (LH) flux (W m^{-2}), total cloud amount (%), and top 300 m ocean heat content (10^{19} J m^{-2}) 10-member ensemble mean response (PRT-CTR) to a Pinatubo-size volcanic perturbation. Values are averaged over 5°S – 5°N for (a–h) neutral ENSO (NTE), (i–p) central Pacific El Niño (CPE), and (q–x) eastern Pacific El Niño (EPE) experiments for June 1991 to June 1993. Black dots define the statistical significance at the 0.05 level. Dashed black and green lines indicate the bounds of the Niño3 + Niño4 region and coastal boundaries of the equatorial Pacific Ocean, respectively. Red and blue colors of wind correspond to westerly and easterly anomalies, respectively.

winds in the CP (Figure 6s), boosting the upwelling and facilitating a prompt westward expansion of the negative SST response toward the CP in October 1991–February 1992. At the same time, similar to the NTE and CPE cases, the westerly wind response develops in the WP initially due to the LOTG mechanism and later due to the ODT mechanism. This gradually leads to a relaxation of the negative temperature response in the CP and EP, causing a positive SST response in May–September 1992. Because of the weak ODT and deficit of available heat content in the EPE case, due to the shutdown of background upwelling and the poleward discharge of heat content associated with the strong El Niño, the westerly wind anomaly is short lived and vanishes by the end of summer 1992 (Figure 6s). Thus, the El Niño-like signal is shorter than that in the NTE and CPE cases.

The cloud cover in the EPE case is the broadest among all the cases, but it responds relatively moderately, except for a strong decline in EP and CP in 1991 (Figure 6w). Due to the EP SST cooling in the second half of 1991, both the deep convection zone and cloudiness anomaly decline in the EP (Figure 6w). In the first half of 1992 the colder EP SST leads to the westward shift of convection and precipitation, reducing the amount of clouds in the EP and increasing downward SW radiation (Figure 6u). Generally, in all cases, the SW flux mainly responds to cloud changes, which increase where SST warms and decrease where it cools, effectively damping the SST response (Figures 6e, 6m, and 6u).

Finally, we consider the ocean heat content response, which is less affected by the direct surface cooling caused by volcanic radiative forcing. Figures 6h, 6p, and 6x show the response (PRT-CTR) of the top 300 m ocean heat content of the 2°S – 2°N averaged equatorial Pacific region. It demonstrates roughly the same effects, discussed above in sections 3.1.1 and 3.1.2 using SST changes as diagnostics, but more clearly defines the termination of different development phases. Positive responses of the ocean heat content extend to the

end of 1992 for the NTE case and the end of 1993 for the CPE case. In both cases, the ODT effects are evident until the end of the warming phase. For the EPE case, in contrast, we see a significant negative ocean heat content response in the EP that lasts until the end of 1991.

Because the strong background El Niño warms SSTs in the EP already at the time of eruption and later on (see Figure 2c), the sea and land surface temperatures near the South American continent are much higher in the EPE than in CPE and NTE cases. When the land surface temperature in the EP decreases after the eruption, the EP LOTG is stronger than in NTE or CPE cases due to the anomalously warm coastal waters. However, because of the land's low thermal capacity, the LOTG effect is short lived as in NTE and CPE cases and cannot maintain the negative temperature response beyond half a year after the eruption (Figure 6r). This is further clarified in sections 3.3 and 3.4.

3.2. Ocean Heat Budget

To better understand the interplay of different processes and compare their contributions in changing the mixed layer temperature and heat content, we analyze the ocean mixed layer heat budget using the equation for mixed layer temperature from *Stevenson and Niiler* [1983] and *Huang et al.* [2010] to calculate monthly mean ocean temperature tendencies:

$$\frac{dT_a}{dt} = -w_H \frac{T_a - T_H}{H} - u_a \frac{\partial T_a}{\partial x} - v_a \frac{\partial T_a}{\partial y} + \frac{Q_{NF}}{\rho C_p H} + Q_{Res} \quad (1)$$

where T_a , u_a , and v_a are the temperature, zonal, and meridional currents; subscript a denotes the quantities vertically averaged between surface and the bottom of the mixed layer at $H=50$ m; T_H and w_H are the temperature and ocean vertical velocity at $H=50$ m, Q_{NF} is net surface heat flux, $\rho=1029$ kg m⁻³ is the seawater density, and $C_p=3990$ J kg⁻¹ K⁻¹ is the specific heat capacity of seawater at constant pressure. The first term in the right side of equation (1) denotes the thermocline and ODT effect, the second and third terms are zonal and meridional advection, the fourth term is heating/cooling caused by surface heat fluxes, and Q_{Res} is a residual term which includes vertical diffusion and subgrid scale mixing.

Each element of the heat budget is integrated over the narrowed Niño3 + Niño4 region (2°S–2°N, 160°E–90°W) for the perturbed and control ensembles, and the difference response is presented in Figure 7. The results are shown for five half-year periods: (a) July–December 1991; (b) January–June 1992; (c) July–December 1993; (d) January–June 1993, and (e) July–December 1993.

During period *a* (Figure 7a), the mixed layer responds to the radiative cooling and the change of the surface wind stresses induced by the increased LOTG and slowly developing ODT and Bjerknes feedbacks. The upper ocean warms only for NTE, while the strongest cooling happens in EPE due to zonal and meridional advective cooling in the EP.

Figure 7b shows the important stage of strengthening of the El Niño-like response due to the volcanic impact. Period *b* in the control CPE and EPE ensembles corresponds to the transition from El Niño to La Niña. However, due to the strong ocean dynamical response to volcanic forcing, an intense warming is observed in all cases even though the surface fluxes tend to damp the temperature anomalies. In the NTE and CPE cases, all three advective terms contribute roughly equally to this warming. In contrast, the EPE warming is mostly caused by the thermocline effect. Zonal and meridional advection also contribute to this warming, but to a lesser extent.

During period *c*, in all cases the El Niño-like responses enter a decay phase, which is the fastest in the EPE ensemble due to a fast transition from a strong El Niño into a strong La Niña, mostly caused by upwelling of anomalously cold water due to arrival of an upwelling oceanic Kelvin wave.

During period *d*, the CPE cooling is caused by a strong zonal advection feedback. The NTE case reaches the final relaxation state at a later time than CPE (period *e*) by means of the almost equal contributions of the thermocline and zonal advection feedbacks.

3.3. Initial ENSO Responses in Period *a*

The initial stage of the ENSO response (period *a* in Figure 7) is characterized by competition between the LOTG and ODT mechanisms and reveals drastic differences between the EPE and NTE/CPE SST responses. To better demonstrate the onsets of the LOTG and ODT mechanisms and differences in the SST evolution in NTE, CPE, and EPE cases, in Figure 8 we show maps of the monthly 10-member ensemble mean responses (PRT-CTR) of surface temperature and wind stress for all experiments for July–December of 1991. Figures 8a–8f correspond to the NTE experiment, 8g–8l to the CPE, and 8m–8r to the EPE. In all cases, the land mass cooling is present

10-member ens mean ocean heat budget (PRT-CTR)
integrated over top 50 m, NINO3+4
July-December 1991

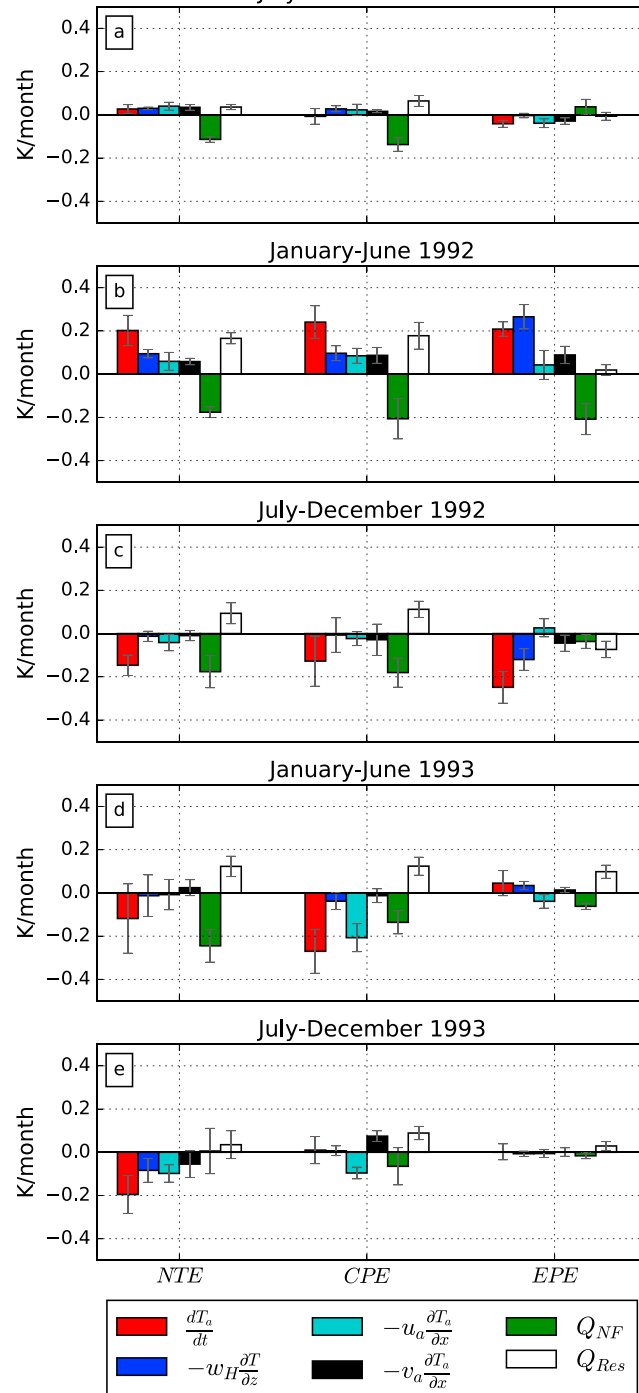


Figure 7. Simulated eruption response of the oceanic top 50 m total temperature tendency (K/month) (red), thermocline feedback and ocean dynamical thermostat (blue), zonal advection (cyan), meridional advection (black), net surface heat flux (green), and residual (white) for the time intervals: (a) July–December 1991, (b) January–June 1992, (c) July–December 1992, (d) January–June 1993, and (e) July–December 1993. Within each time interval, the 10-member ensemble mean NTE, CPE, and EPE responses to the Pinatubo eruption (PRT-CTR) are shown. The values are integrated over the narrowed Niño3 + Niño4 region (2°S–2°N, 160°E–90°W). Error bars show the standard error of the ensemble mean difference, which increases with time as both ensembles chaotically disperse.

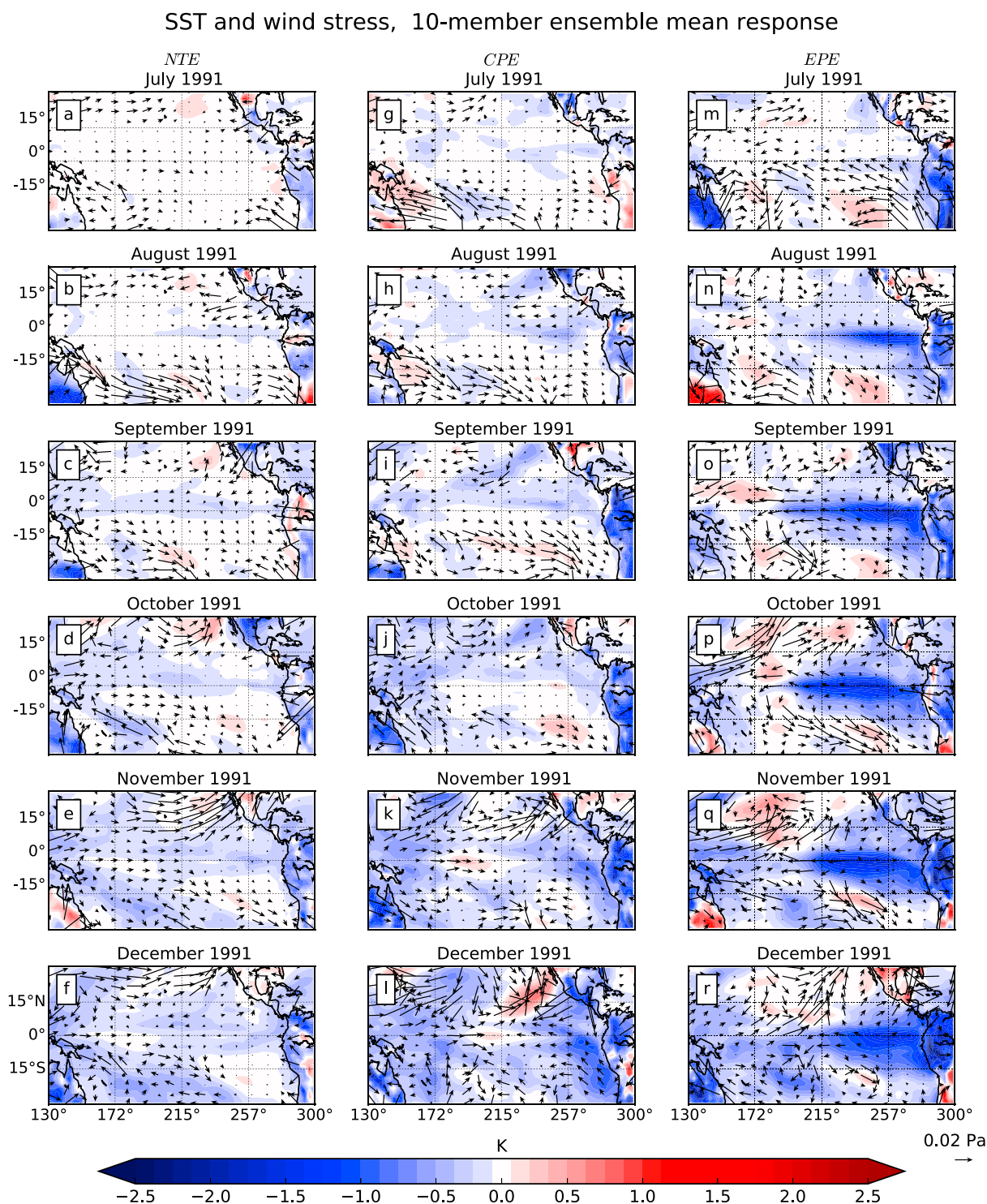


Figure 8. Simulated surface temperature (K, shading) and surface wind stress vector (Pa, arrows) 10-member ensemble mean response (PRT-CTR) to the Pinatubo radiative forcing in the (a–f) neutral ENSO (NTE), (g–l) central Pacific El Niño (CPE), and (m–r) eastern Pacific El Niño (EPE) experiments for July–December 1991.

in Australia and the Americas, although it is partially opposed by advection of warm air from the ocean as happens in the NTE in July and November, in CPE in July, and in EPE in August, November, and slightly in October 1991.

In the NTE and CPE ensembles, both WP and EP LOTG effects on equatorial (5°S – 5°N) wind stress are seen in August–September. The WP SST cooling onset associated with the ODT mechanism takes place in October and November 1991 in the CPE and NTE experiments, respectively, and develops a westerly wind stress response in the WP and CP (Figures 8d, 8e, 8j, and 8k).

In the EPE experiment, we see the strong off-land wind stress response in July 1991 (Figure 8m). The significant SST cooling of the EP starts from August 1991 (Figure 8n). This SST cooling is so strong that it initiates a significant positive feedback of the wind stress in the CP in September–December and continues to expand westward (Figures 8o–8r). At the same time, the ODT mechanism turns on in October 1991 (Figure 8p). As we mentioned earlier, the strong EP cooling in the EPE case cannot be fully explained by only LOTG and ODT mechanisms. Our working hypothesis is that this ocean cooling is partially caused by the predominant weakening of the strong El Niños in the perturbed EPE ensemble. This effect was discussed by Wittenberg *et al.* [2014] and based on the asymmetry of a strong El Niño response probability distribution function. It essentially means that a very strong El Niño disturbed by any perturbation more probably leads to an outcome of a weaker El Niño. To better quantify this effect, a specific set of sensitivity experiments are conducted and analyzed in the section 3.4.

3.4. Sensitivity of El Niño Response to Small Perturbations

Wittenberg *et al.* [2014] demonstrated that El Niño is very sensitive to the small perturbations that limit its predictability. In their simulations, a slight perturbation of ICs at the beginning of a calendar year drastically affected the El Niño later in the year and onward. Stricken by the IC change, the sporadically developing WWBs stochastically impacted the amplitude of El Niño and were even able to reverse the ENSO phase with respect to the unperturbed case. At first glance, this contradicts our observation of a statistically significant ENSO response to volcanic forcing in the EPE and CPE ensembles and has to be explained. In addition, this stochastic behavior could be useful to understand the initial stage of the EPE ensemble development, as the stochastic component of the response might be responsible for the initial drastic cooling in the EPE case, as we hypothesized in section 3.3.

Therefore, in this section we specifically test the sensitivity of the El Niño response to small perturbations in the control and perturbed runs. We use a single IC, which results in a strong EP El Niño in the initial year of the simulation (Figure 9, black dashed line). We name this control simulation CTR_0 . The perturbations are introduced by a small radiative forcing that is generated by imposing the Pinatubo aerosol extinctions [Sato *et al.*, 1993; Stenchikov *et al.*, 2009] multiplied by a factor α , where α ranges from 0.001 to 0.05 with a step of 0.001. The corresponding SW radiative forcings scale proportionally. We apply these small perturbations in the control runs with zero background volcanic forcing and in the perturbed runs on the top of the full-scale Pinatubo forcing. In both perturbed and control runs, the small forcing is imposed at the beginning of 1991 (in our case in February), as in Wittenberg *et al.* [2014], and in June, at the time of the actual observed Pinatubo eruption. We assume that forcing imposed early in the year has more time to disrupt the control El Niño than that imposed in June.

In order to evaluate the spread of responses to these small perturbations, we calculate two sets of 50-member “perturbed control” and “perturbed Pinatubo” ensembles, named $\text{CTR}_\alpha^{\text{Feb}}$, $\text{CTR}_\alpha^{\text{Jun}}$, $\text{PRT}_\alpha^{\text{Feb}}$, and $\text{PRT}_\alpha^{\text{Jun}}$, respectively (see Table 2). Thus, in “June” cases we preserve the control El Niño until the time of a perturbation, so the control El Niño develops unaffected until a perturbation is imposed. In “February” cases, the El Niño is actually first affected in January, due to time interpolation of the prescribed volcanic optical depth.

Figure 9 compares the changes of Niño 3.4 index in all four ensembles forced by α perturbations imposed in February (a and c) and in June (b and d) at the time of Pinatubo eruption. Perturbations imposed in February lead to a drastic change of the Niño 3.4 index, almost completely suppressing the strong EP El Niño in most members of the $\text{CTR}_\alpha^{\text{Feb}}$ and $\text{PRT}_\alpha^{\text{Feb}}$ ensembles, while the June perturbations in $\text{CTR}_\alpha^{\text{Jun}}$ runs disturb the El Niño only slightly. The ensemble average in $\text{PRT}_\alpha^{\text{Jun}}$ runs repeats the main features of the Niño 3.4 response revealed in Figure 5c, confirming the statistical stability of our 10-member ensemble ENSO responses.

Thus, the small α perturbations imposed in June do not have enough time to grow and affect the ENSO phase in the current and next year (Figure 9b). Although the variability inside the $\text{CTR}_\alpha^{\text{Jun}}$ ensemble grows in the

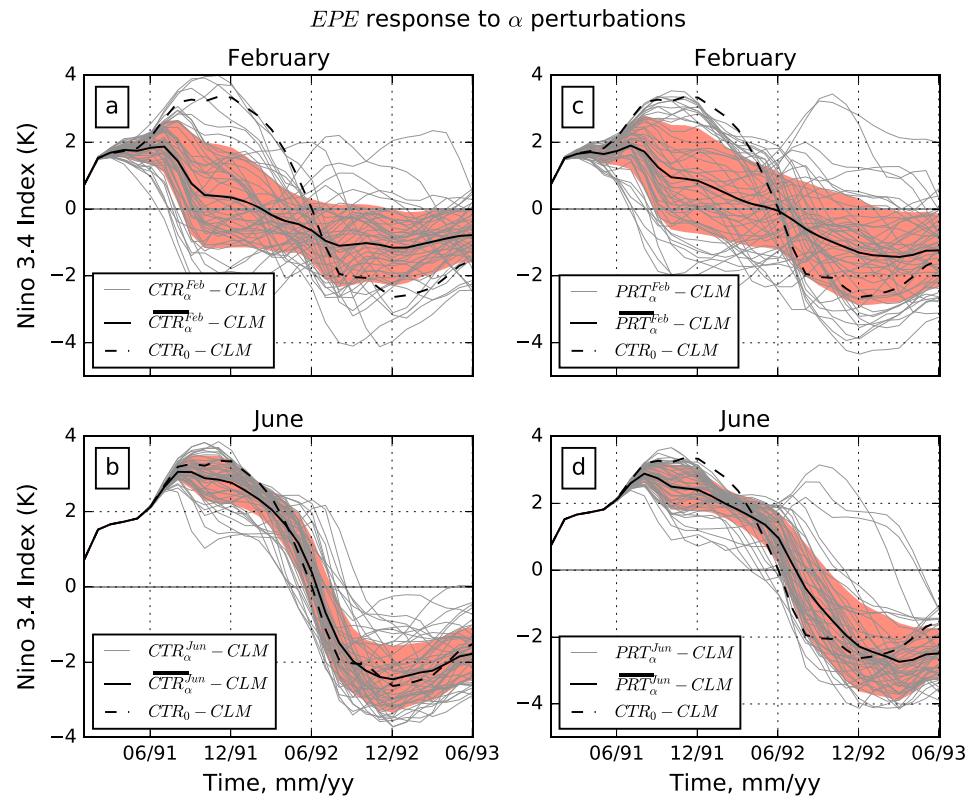


Figure 9. Comparison of simulated Niño 3.4 indexes (K) in a single reference eastern Pacific El Niño realization (CTR_0) and 50-member control and perturbed ensembles with α perturbations in (a and c) February (CTR_α^{Feb} ; PRT_α^{Feb}) and (b and d) June (CTR_α^{Jun} ; PRT_α^{Jun}). Niño 3.4 indexes in single realizations (grey spaghetti curves) and ensemble means (black solid curve) are calculated with respect to the climatology. Bars denote the ensemble mean quantities. Red shading depicts $\pm\sigma$ ensemble variability.

second year after perturbation, the ensemble mean still captures the El Niño trajectory fairly well. However, the February perturbations change ENSO drastically. Experiments conducted imposing disturbances in different months (not shown) suggest that the signal-to-noise ratio in case of winter and spring eruptions is smaller than in the case of summer and fall eruptions, i.e., the ENSO response to eruptions like El Chichón or Tambora in April (apart from the response dependence on the magnitude of an eruption) could be less pronounced than that of Pinatubo which happened later in the year. This finding is consistent with the concept of the “ENSO spring predictability barrier,” which suggests that the persistence of El Niño is lower during the late boreal winter and spring, while it increases starting in June. McPhaden [2003] examined this concept in the observations, and Levine and McPhaden [2015] confirmed it using a simplified conceptual model, while our study finds consistent behavior in a coupled ocean-atmosphere GCM.

Another interesting feature in Figure 9 is a decrease of the Niño 3.4 index in both CTR_α^{Jun} and PRT_α^{Jun} ensembles with respect to the CTR_0 at the peak of El Niño in November–December 1991. This effect is directly related to the drastic SST cooling in the EPE case shown in Figure 8 and discussed in section 3.3. It is also consistent with the results from our 10-member ensemble (Figure 5c). To clarify this issue, in Figure 10 we present the 50-member ensemble mean monthly surface temperature responses calculated from PRT_α^{Jun} and CTR_α^{Jun} experiments with respect to CTR_0 and their difference ($PRT_\alpha^{Jun} - CTR_\alpha^{Jun}$). The idea is to split up the “perturbed Pinatubo” (Figures 10a–10f) response into the stochastic (Figures 10g–10l) and the forced deterministic volcanic (Figures 10m–10r) components.

Black dots in Figures 10a–10f and 10g–10l show the areas where the CTR_0 surface temperature is below the 10th percentile or above the 90th percentile of the PRT_α^{Jun} (10a–10f) and CTR_α^{Jun} (10g–10l) ensembles, respectively. We choose to use percentiles here because the probability distributions of both PRT_α^{Jun} and CTR_α^{Jun} ensembles are skewed as discussed above. In Figures 10m–10r black dots show statistically

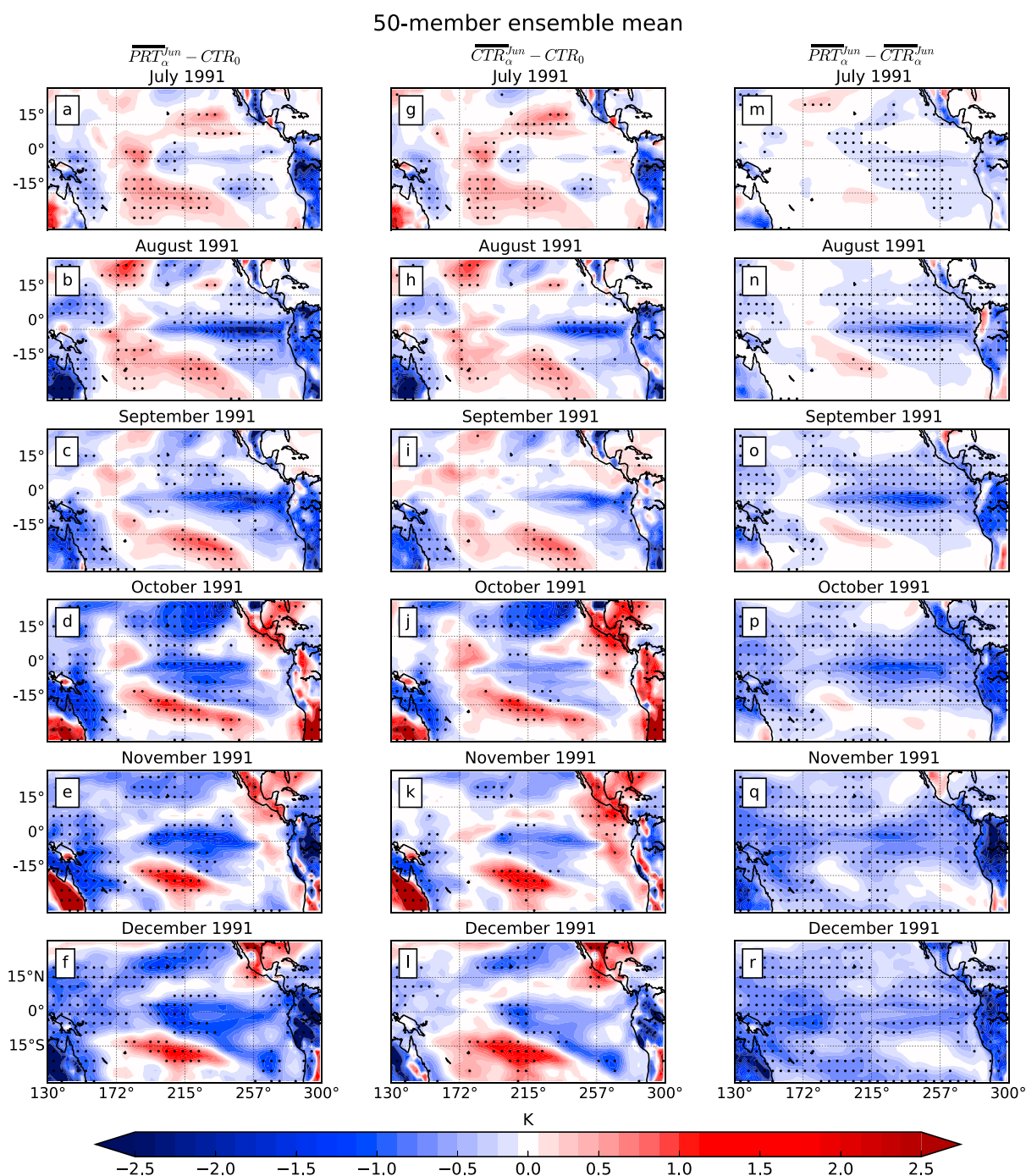


Figure 10. Simulated monthly averaged surface temperature (K, shading) 50-member (a–f) perturbed (PRT_{α}^{Jun}) and (g–l) control (CTR_{α}^{Jun}) ensemble means with α perturbations calculated as anomalies with respect to a single reference eastern Pacific El Niño realization (CTR_0) and (m–r) difference between PRT_{α}^{Jun} and CTR_{α}^{Jun} for July–December 1991. Black dots indicate areas where the reference surface temperature is below the 10th percentile or above the 90th percentile for the 50-member CTR_{α}^{Jun} (Figures 10a–10f) and PRT_{α}^{Jun} (Figures 10g–10l) ensemble distributions, and statistical significance of the ensemble mean surface temperature difference calculated using two-tailed Student's t test at the 0.05 significance level (Figures 10m–10r).

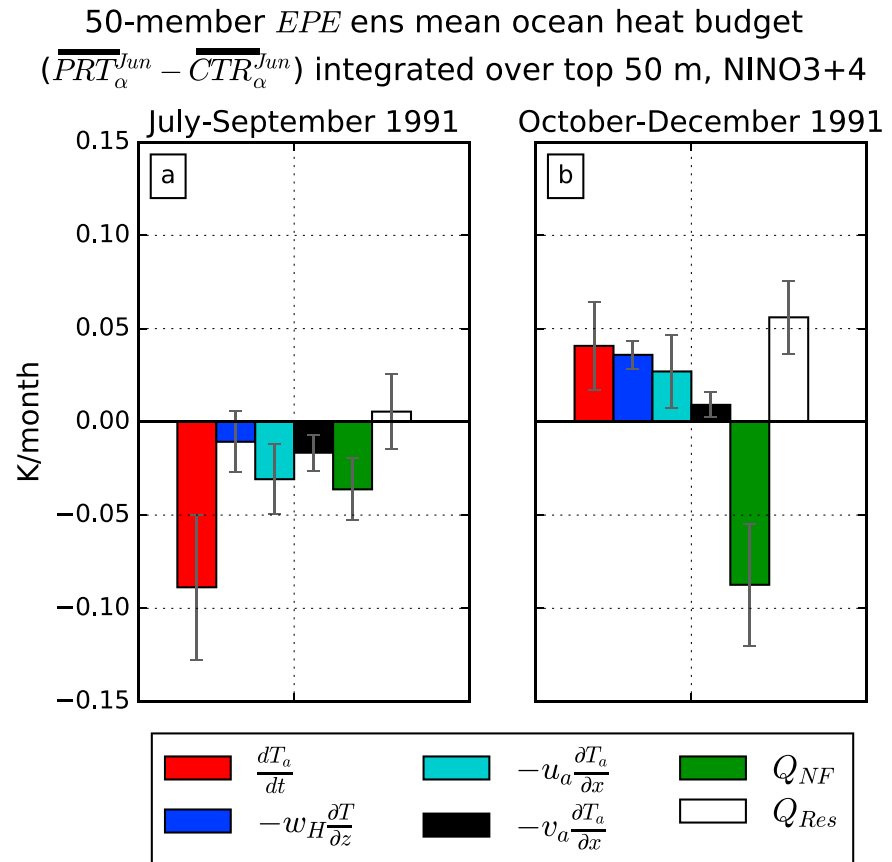


Figure 11. Simulated oceanic top 50 m total temperature tendency (K/month) (red), thermocline feedback (blue), zonal advection (cyan), meridional advection (black), net surface heat flux (green), and residual (white) for the time intervals: (a) July–September 1991 and (b) October–December 1991. Within each time interval, the 50-member ensemble mean eastern Pacific El Niño response to the Pinatubo eruption ($\overline{PRT}_\alpha^{Jun} - \overline{CTR}_\alpha^{Jun}$) is shown. The values are integrated over the narrowed Niño3 + Niño4 region (2°S–2°N, 160°E–90°W). Error bars show the standard error of the mean difference.

significant $\overline{PRT}_\alpha^{Jun} - \overline{CTR}_\alpha^{Jun}$ surface temperature responses at the 0.05 significance level, calculated using a two-tailed Student's *t* test, as their probability distribution is close to normal.

The magnitude of the stochastic response in the EP is evident in Figures 10g–10l, which shows the purely chaotic dispersion of the unforced ensemble. The anomaly is predominantly negative in the equatorial region, indicating a shift to weaker El Niños.

The volcanically induced SST cooling with the stochastic component removed (Figures 10m–10r) is statistically significant, and reaches more than 1 K in the equatorial region. The “deterministic” SST cooling (Figure 10m–r) is slightly greater than the stochastic cooling caused by the infinitesimal forcing (Figure 10m–r).

In order to better separate the radiative and dynamical feedbacks in July–December 1991 for this single EPE case, we performed the ocean heat budget analysis using differences of $\overline{PRT}_\alpha^{Jun}$ and $\overline{CTR}_\alpha^{Jun}$ ensembles (i.e., removing the stochastic component of the response) as described in section 3.2, but for two 3 month intervals. Figure 11 shows that during the first 3 months (July–September 1991) of the radiative cooling, the zonal and meridional advection along with the thermocline feedbacks strengthen the cooling in the EP. Starting from October 1991 the ODT overwhelms this, and, despite the intensified radiative cooling, the mixed layer warms due to thermocline and zonal advection feedbacks.

4. Discussion and Conclusions

This research article focuses on mechanisms of the response of ENSO to Pinatubo-size volcanic forcing and aims to reconcile apparent inconsistencies in previous studies [Mann *et al.*, 2005; Emile-Geay *et al.*, 2008; McGregor and Timmermann, 2011; Ohba *et al.*, 2013]. Using the coupled ocean-atmosphere model CM2.1,

we simulate the climate impact of a Pinatubo-size eruption in ENSO-neutral, CP and EP El Niño, and La Niña years. We show that the initial ENSO phase, El Niño amplitude and type, and the seasonal timing of the eruption affect the climate and ENSO responses to volcanic forcing. We find that the eruption causes different climate responses in CP and EP El Niño years, and we study the sensitivity of the volcano-induced ENSO response to small perturbations, illuminating the contributions of stochastic and deterministic responses to volcanic radiative forcing.

4.1. ENSO Response

A Pinatubo-size eruption induces an El Niño-like response in the first year after the eruption in ENSO-neutral years and prolongs moderate CP El Niños. The EP El Niño weakens due to a combined effect of the enhanced LOTG mechanism and the ENSO stochastic response, and its termination is delayed. The ENSO in La Niña years is largely insensitive to volcanic forcing in CM2.1. The absence of any La Niña response is caused by a weak LOTG mechanism due to the anomalously cold equatorial SST and suppressed ODT mechanism because of the weak zonal upwelling gradient in the entire equatorial Pacific. This effect might be exaggerated by CM2.1, as its La Niñas tend to expand upwelling farther west than observed.

4.2. Deterministic Mechanisms

The LOTG mechanism described by *Ohba et al.* [2013] as the main driver of the ENSO response to volcanic forcing is also at work in CM2.1. However, we find it to be relatively short lived and work only for the first 2–3 months after the eruption, being the forerunner of stronger dynamical responses. Depending on the ENSO phase, the LOTG enhances wind anomalies in the WP and/or EP. During the response of neutral ENSO or CP El Niño, the WP westerly wind anomaly dominates and causes an El Niño-like warming in the perturbed experiments. In the response of the EP El Niño the EP easterly wind anomaly prevails, although it is not fully related to the LOTG mechanism. Taking into account the findings of *Wittenberg et al.* [2014] and using the “perturbed forcing” technique, we find that the total cooling response during the first 6 months after an eruption is a combination of stochastic and deterministic components.

The ODT mechanism of *Clement et al.* [1996] takes over after the LOTG in October–November 1991, maintaining a deterministic El Niño-like response for a year in the CP El Niño case, and almost 2 years in the ENSO-neutral case. In the EPE case, weakened upwelling due to the strong El Niño tempers the ODT, resulting in only short-term SST warming.

4.3. Stochastic Mechanisms

Wittenberg et al. [2014] indicated the high sensitivity of El Niños to small perturbations that limits their predictability. In particular, our sensitivity analysis suggests that the eruption response of the strong EP El Niño contains both stochastic and deterministic components. The role of the stochastic component is large if small perturbations are imposed at the beginning of the calendar year but decreases if disturbances are imposed later in the year. Because the Pinatubo eruption occurred in June, the deterministic part of the EP El Niño response prevails over the stochastic one. This suggests that the timing of a volcanic eruption is critical for the El Niño response. The EP El Niño responses to volcanic eruptions occurring in the winter-spring are more contaminated by chaotic dispersion than those occurring in summer-fall. Therefore, Pinatubo-size eruptions in winter and spring are less likely to produce a clean impact on strong EP El Niños than the actual Pinatubo, which occurred in June. In this context, the conclusion in *Robock et al.* [1995] that the huge “El Niño of the century” in 1982 after the El Chichón eruption in April could not be significantly influenced by volcanic forcing, as it was already in the development phase at the time of eruption, might be questioned.

4.4. Consistency With Other Studies and Observations

Our results are consistent with the findings by *Ohba et al.* [2013] in terms of development of an El Niño-like response in the first year after a Pinatubo-size eruption. However, they did not consider different flavors of El Niño. In addition, the La Niña in the CM2.1 does not show a tendency toward warmer ocean conditions after the eruption, unlike in the MIROC5i model used by *Ohba et al.* [2013].

The responses of different El Niño types—CP and EP in our study—are consistent, even quantitatively, with the ERSST observational data displayed in Figure 1 and discussed in *Li et al.* [2010] and *Wahl et al.* [2014]. The strong EP El Niño at the time of the El Chichón eruption was comparatively short and followed by a La Niña, while the moderate CP El Niño during the Pinatubo eruption lasted longer and transformed from CP to EP in the second year after the eruption, as seen in our simulations. In fact, a La Niña-like response often follows the

El Niño-like response [Maher *et al.*, 2015], which we have not fully addressed in this study, which has focused on the immediate posteruption period. The induced La Niña-like response will be a subject of a different study.

As Pinatubo-size volcanic impacts in ENSO-neutral and CP El Niño years generate El Niño like responses in at least the first year after an eruption, the frequency of El Niño conditions in this year might reach 0.65 (see Table 1: NTE+CPE). This is consistent with paleoanalyses that suggest a doubling of the probability of warm EP SSTs in posteruption years, compared to the model climatology (0.27, see Table 1: CPE+EPE).

4.5. Recommendations for Further Analysis

We have demonstrated a sensitivity of the volcanic response to the magnitude and shape of an ENSO event that otherwise would have developed in the unperturbed case. This might partially explain the spread of previous model results, as simulations in different studies have been conducted for different initial ENSO phases. Moreover, many up-to-date models produce only one type of El Niño or produce unrealistically weak or overly frequent ENSO events [Ham and Kug, 2012; Yu and Kim, 2010]. This intrinsic model behavior inevitably affects the simulated ENSO sensitivity and may not be able to completely cover the full set of possible cases in nature.

We suggest that when analyzing the ENSO sensitivity to volcanic eruptions in models or observations, special attention needs to be paid to the initial phase of ENSO and the seasonal timing of the eruption. Previous studies did not take into account the composites of different initial ENSO phases, eruption timings. Meehl *et al.* [2015] also noted that especially multimodel composites could obscure the ENSO response signals. Moreover, as in a specific model one perturbs an intrinsic ENSO, the results inevitably become model dependent, as different models generate different (and often inadequate) ENSO cycles [Gabriel and Robock, 2015], so meaningful composites are difficult to obtain.

The numerical experiments we conducted are consistent with those proposed in the “volc-pinatubo-full” task from the VolMIP activity planned for CMIP6 [Zanchettin *et al.*, 2016] and will be contributing to that inter-comparison study. Our recommendations for a more accurate and selective compositing of the results will help to better identify the ENSO responses to volcanic forcing. The “volc-pinatubo-slab” experimental design [Zanchettin *et al.*, 2016], however, is insufficient to cover all aspects of Volcano-ENSO interaction, as a full ocean model is needed.

Acknowledgments

We thank Hans-F. Graf, Alan Robock, and Jim Carton for valuable discussions. The research reported in this publication was supported by King Abdullah University of Science and Technology. In particular, for computer time, the resources of the KAUST Supercomputing Laboratory were used, in addition to those of NOAA GFDL. The simulation results and supporting data sets are available from the corresponding author upon request. NOAA ERSST V4 data are provided by the NOAA/OAR/ESRL PSD, Boulder, Colorado, USA, from their Web site at <http://www.esrl.noaa.gov/psd/>.

References

- Adams, J. B., M. E. Mann, and C. M. Ammann (2003), Proxy evidence for an El Niño-like response to volcanic forcing, *Nature*, 426(6964), 274–278.
- Anderson, J. L., et al. (2004), The new GFDL global atmosphere and land model AM2-LM2: Evaluation with prescribed SST simulations, *J. Clim.*, 17(24), 4641.
- Ashok, K., S. K. Behera, S. A. Rao, H. Weng, and T. Yamagata (2007), El Niño Modoki and its possible teleconnection, *J. Geophys. Res.*, 112, C11007, doi:10.1029/2006JC003798.
- Atwood, A. R., D. S. Battisti, A. T. Wittenberg, W. G. H. Roberts, and D. J. Vimont (2016), Characterizing unforced multi-decadal variability of ENSO: A case study with the GFDL CM2.1 coupled GCM, *Clim. Dyn.*, 1–18, doi:10.1007/s00382-016-3477-9.
- Bjerknes, J. (1969), Atmospheric teleconnections from the equatorial Pacific, *Mon. Weather Rev.*, 97(3), 163–172.
- Briffa, K. R., P. D. Jones, F. H. Schweingruber, and T. J. Osborn (1998), Influence of volcanic eruptions on Northern Hemisphere summer temperature over the past 600 years, *Nature*, 393(6684), 450–455.
- Brönnimann, S., E. Xoplaki, C. Casty, A. Pauling, and J. Luterbacher (2007), ENSO influence on Europe during the last centuries, *Clim. Dyn.*, 28(2–3), 181–197.
- Capotondi, A., et al. (2015), Understanding ENSO diversity, *Bull. Am. Meteorol. Soc.*, 96, 921–938, doi:10.1175/BAMS-D-13-00117.1.
- Cash, D. W., J. C. Borck, and A. G. Patt (2006), Countering the loading-dock approach to linking science and decision making comparative analysis of El Niño/Southern Oscillation (ENSO) forecasting systems, *Sci. Technol. Hum. Values*, 31(4), 465–494.
- Chavez, F. P., J. Ryan, S. E. Lluch-Cota, and M. Niquen (2003), From anchovies to sardines and back: Multidecadal change in the Pacific Ocean, *Science*, 299(5604), 217–221.
- Chen, C., M. A. Cane, A. T. Wittenberg, and D. Chen (2016), ENSO in the CMIP5 simulations: Lifecycles, diversity, and responses to climate change, *J. Clim.*, 30, 775–801 doi:10.1175/JCLI-D-15-0901.1.
- Chen, D., T. Lian, C. Fu, M. A. Cane, Y. Tang, R. Murtugudde, X. Song, Q. Wu, and L. Zhou (2015), Strong influence of westerly wind bursts on El Niño diversity, *Nat. Geosci.*, 8(5), 339–345.
- Choi, K.-Y., G. A. Vecchi, and A. T. Wittenberg (2013), ENSO transition, duration, and amplitude asymmetries: Role of the nonlinear wind stress coupling in a conceptual model, *J. Clim.*, 26(23), 9462–9476.
- Choi, K.-Y., G. A. Vecchi, and A. T. Wittenberg (2015), Nonlinear zonal wind response to ENSO in the CMIP5 models: Roles of the zonal and meridional shift of the ITCZ/SPCZ and the simulated climatological precipitation, *J. Clim.*, 28(21), 8556–8573.
- Church, J. A., N. J. White, and J. M. Arblaster (2005), Significant decadal-scale impact of volcanic eruptions on sea level and ocean heat content, *Nature*, 438(7064), 74–77.
- Clement, A. C., R. Seager, M. A. Cane, and S. E. Zebiak (1996), An ocean dynamical thermostat, *J. Clim.*, 9(9), 2190–2196.
- Collins, M., et al. (2010), The impact of global warming on the tropical Pacific Ocean and El Niño, *Nat. Geosci.*, 3(6), 391–397.
- D’Arrigo, R., E. R. Cook, R. J. Wilson, R. Allan, and M. E. Mann (2005), On the variability of ENSO over the past six centuries, *Geophys. Res. Lett.*, 32, L03711, doi:10.1029/2004GL022055.
- De Silva, S. L., and G. A. Zielinski (1998), Global influence of the AD 1600 eruption of Huaynaputina, Peru, *Nature*, 393(6684), 455–458.

- Delworth, T. L., et al. (2006), GFDL's CM2 global coupled climate models. Part I: Formulation and simulation characteristics, *J. Clim.*, 19(5), 643–674.
- DiNezio, P. N., B. P. Kirtman, A. C. Clement, S.-K. Lee, G. A. Vecchi, and A. Wittenberg (2012), Mean climate controls on the simulated response of ENSO to increasing greenhouse gases, *J. Clim.*, 25, 7399–7420, doi:10.1175/JCLI-D-11-00494.1.
- Dutton, E. G., and J. R. Christy (1992), Solar radiative forcing at selected locations and evidence for global lower tropospheric cooling following the eruptions of El Chichón and Pinatubo, *Geophys. Res. Lett.*, 19(23), 2313–2316.
- Emile-Geay, J., R. Seager, M. A. Cane, E. R. Cook, and G. H. Haug (2008), Volcanoes and ENSO over the past millennium, *J. Clim.*, 21(13), 3134–3148.
- Emile-Geay, J., K. Cobb, M. Mann, and A. T. Wittenberg (2013), Estimating central equatorial Pacific SST variability over the past millennium. Part II: Reconstructions and implications, *J. Clim.*, 26, 2329–2352, doi:10.1175/JCLI-D-11-00511.1.
- Fang, X.-H., F. Zheng, and J. Zhu (2015), The cloud-radiative effect when simulating strength asymmetry in two types of El Niño events using CMIP5 models, *J. Geophys. Res. Oceans*, 120, 4357–4369, doi:10.1002/2014JC010683.
- Fischer, E., J. Luterbacher, E. Zorita, S. Tett, C. Casty, and H. Wanner (2007), European climate response to tropical volcanic eruptions over the last half millennium, *Geophys. Res. Lett.*, 34, L05707, doi:10.1029/2006GL027992.
- Gabriel, C., and A. Robock (2015), Stratospheric geoengineering impacts on El Niño/Southern Oscillation, *Atmos. Chem. Phys.*, 15(20), 11,949–11,966.
- Gebbie, G., I. Eisenman, A. Wittenberg, and E. Tziperman (2007), Modulation of westerly wind bursts by sea surface temperature: A semistochastic feedback for ENSO, *J. Atmos. Sci.*, 64, 3281–3295, doi:10.1175/JAS4029.1.
- Gleckler, P., T. Wigley, B. Santer, J. Gregory, K. AchutaRao, and K. Taylor (2006), Volcanoes and climate: Krakatoa's signature persists in the ocean, *Nature*, 439(7077), 675–675.
- Gnanadesikan, A., et al. (2006), GFDL's CM2 global coupled climate models. Part II: The baseline ocean simulation, *J. Clim.*, 19(5), 675–697.
- Goldenberg, S. B., C. W. Landsea, A. M. Mestas-Núñez, and W. M. Gray (2001), The recent increase in Atlantic hurricane activity: Causes and implications, *Science*, 293(5529), 474–479.
- Graf, H.-F., and D. Zanchettin (2012), Central Pacific El Niño, the “subtropical bridge,” and Eurasian climate, *J. Geophys. Res.*, 117, D01102, doi:10.1029/2011JD016493.
- Gray, W. M. (1984), Atlantic seasonal hurricane frequency. Part I: El Niño and 30 mb quasi-biennial oscillation influences, *Mon. Weather Rev.*, 112(9), 1649–1668.
- Griffies, S., et al. (2005), Formulation of an ocean model for global climate simulations, *Ocean Sci.*, 1(1), 45–79.
- Guilyardi, E., A. Wittenberg, M. Balmaseda, W. Cai, M. Collins, M. J. McPhaden, M. Watanabe, and S.-W. Yeh (2016), Fourth CLIVAR workshop on the evaluation of ENSO processes in climate models: ENSO in a changing climate, *Bull. Am. Meteorol. Soc.*, 97(5), 817–820, doi:10.1175/BAMS-D-15-00287.1.
- Ham, Y.-G., and J.-S. Kug (2012), How well do current climate models simulate two types of El Niño?, *Clim. Dyn.*, 39(1–2), 383–398.
- Hansen, J., A. Lacis, R. Ruedy, and M. Sato (1992), Potential climate impact of Mount Pinatubo eruption, *Geophys. Res. Lett.*, 19(2), 215–218, doi:10.1029/91GL02788.
- Hansen, J., M. Sato, and R. Ruedy (1997), Radiative forcing and climate response, *J. Geophys. Res.*, 102, 6831–6864.
- Haywood, J. M., A. Jones, N. Bellouin, and D. Stephenson (2013), Asymmetric forcing from stratospheric aerosols impacts Sahelian rainfall, *Nat. Clim. Change*, 3(7), 660–665.
- Hirano, M. (1988), On the trigger of El Niño Southern Oscillation by the forcing of early El Chichon volcanic aerosols, *J. Geophys. Res.*, 93(D5), 5365–5384.
- Huang, B., Y. Xue, D. Zhang, A. Kumar, and M. J. McPhaden (2010), The NCEP GODAS ocean analysis of the tropical Pacific mixed layer heat budget on seasonal to interannual time scales, *J. Clim.*, 23(18), 4901–4925.
- Huang, B., V. F. Banzon, E. Freeman, J. Lawrimore, W. Liu, T. C. Peterson, T. M. Smith, P. W. Thorne, S. D. Woodruff, and H.-M. Zhang (2015), Extended reconstructed sea surface temperature version 4 (ERSST.v4). Part I: Upgrades and intercomparisons, *J. Clim.*, 28(3), 911–930.
- Ineson, S., and A. Scaife (2009), The role of the stratosphere in the European climate response to El Niño, *Nat. Geosci.*, 2(1), 32–36.
- Jia, L., et al. (2015), Improved seasonal prediction of temperature and precipitation over land in a high-resolution GFDL climate model, *J. Clim.*, 28(5), 2044–2062.
- Karamperidou, C., M. A. Cane, U. Lall, and A. T. Wittenberg (2014), Intrinsic modulation of ENSO predictability viewed through a local Lyapunov lens, *Clim. Dyn.*, 42, 253–270, doi:10.1007/s00382-013-1759-z.
- Kessler, W. S., and M. J. McPhaden (1995), The 1991–1993 El Niño in the central Pacific, *Deep Sea Res., Part II*, 42(2), 295–333.
- Kim, S. T., and F.-F. Jin (2011), An ENSO stability analysis. Part II: Results from the twentieth and twenty-first century simulations of the CMIP3 models, *Clim. Dyn.*, 36(7–8), 1609–1627.
- Kim, S. T., W. Cai, F.-F. Jin, A. Santoso, L. Wu, E. Guilyardi, and S.-I. An (2014), Response of El Niño sea surface temperature variability to greenhouse warming, *Nat. Clim. Change*, 4(9), 786–790.
- Kug, J.-S., J. Choi, S.-I. An, F.-F. Jin, and A. T. Wittenberg (2010), Warm pool and cold tongue El Niño events as simulated by the GFDL 2.1 coupled GCM, *J. Clim.*, 23(5), 1226–1239.
- Lacis, A., J. Hansen, and M. Sato (1992), Climate forcing by stratospheric aerosols, *Geophys. Res. Lett.*, 19(15), 1607–1610.
- Lamb, H. H. (1970), Volcanic dust in the atmosphere; with a chronology and assessment of its meteorological significance, *Philos. Trans. R. Soc. A*, 266(1178), 425–533.
- Lee, S.-K., P. N. DiNezio, E.-S. Chung, S.-W. Yeh, A. T. Wittenberg, and C. Wang (2014), Spring persistence, transition, and resurgence of El Niño, *Geophys. Res. Lett.*, 41, 8578–8585, doi:10.1002/2014GL02484.
- Lee, S.-K., A. T. Wittenberg, D. B. Enfield, S. J. Weaver, C. Wang, and R. Atlas (2016), US regional tornado outbreaks and their links to spring ENSO phases and North Atlantic SST variability, *Environ. Res. Lett.*, 11(4), 044008.
- LeGrande, A. N., K. Tsigaridis, and S. E. Bauer (2016), Role of atmospheric chemistry in the climate impacts of stratospheric volcanic injections, *Nat. Geosci.*, 9, 652–655.
- Levine, A. F. Z., and M. J. McPhaden (2015), The annual cycle in ENSO growth rate as a cause of the spring predictability barrier, *Geophys. Res. Lett.*, 42, 5034–5041, doi:10.1002/2015GL064309.
- Li, G., B. Ren, C. Yang, and J. Zheng (2010), Indices of El Niño and El Niño Modoki: An improved El Niño Modoki index, *Adv. Atmos. Sci.*, 27, 1210–1220.
- Li, J., S.-P. Xie, E. R. Cook, G. Huang, R. D'Arrigo, F. Liu, J. Ma, and X.-T. Zheng (2011), Interdecadal modulation of El Niño amplitude during the past millennium, *Nat. Clim. Change*, 1(2), 114–118.
- Li, J., et al. (2013), El Niño modulations over the past seven centuries, *Nat. Clim. Change*, 3(9), 822–826.
- Lim, H.-G., S.-W. Yeh, J.-S. Kug, Y.-G. Park, J.-H. Park, R. Park, and C.-K. Song (2015), Threshold of the volcanic forcing that leads the El Niño-like warming in the last millennium: Results from the ERIK simulation, *Clim. Dyn.*, 46, 3725–3736.

- Lin, S.-J. (2004), A 'vertically Lagrangian' finite-volume dynamical core for global models, *Mon. Weather Rev.*, *132*(10), 2293–2307.
- Maher, N., S. McGregor, M. H. England, and A. S. Gupta (2015), Effects of volcanism on tropical variability, *Geophys. Res. Lett.*, *42*, 6024–6033, doi:10.1002/2015GL064751.
- Mann, M. E., M. A. Cane, S. E. Zebiak, and A. Clement (2005), Volcanic and solar forcing of the tropical Pacific over the past 1000 years, *J. Clim.*, *18*(3), 447–456.
- McCormick, M. P., L. W. Thomason, and C. R. Trepte (1995), Atmospheric effects of the Mt Pinatubo eruption, *Nature*, *373*(6513), 399–404.
- McGregor, S., and A. Timmermann (2011), The effect of explosive tropical volcanism on ENSO, *J. Clim.*, *24*(8), 2178–2191.
- McGregor, S., A. Timmermann, and O. Timm (2010), A unified proxy for ENSO and PDO variability since 1650, *Clim. Past*, *6*(1), 1–17.
- McGregor, S., A. Timmermann, M. H. England, O. Elison Timm, and A. T. Wittenberg (2013), Inferred changes in El Niño–Southern Oscillation variance over the past six centuries, *Clim. Past*, *9*, 2269–2284, doi:10.5194/cp-9-2269-2013.
- McPhaden, M. J. (2003), Tropical Pacific Ocean heat content variations and ENSO persistence barriers, *Geophys. Res. Lett.*, *30*(9), 1480, doi:10.1029/2003GL016872.
- Meehl, G. A., H. Teng, N. Maher, and M. H. England (2015), Effects of the Mount Pinatubo eruption on decadal climate prediction skill of Pacific sea surface temperatures, *Geophys. Res. Lett.*, *42*, 10,840–10,846, doi:10.1002/2015GL066608.
- Milly, P., and A. Shmakin (2002), Global modeling of land water and energy balances. Part I: The land dynamics (LaD) model, *J. Hydrometeorol.*, *3*(3), 283–299.
- Minnis, P., E. Harrison, L. Stowe, G. Gibson, F. Denn, D. Doelling, and W. Smith (1993), Radiative climate forcing by the Mount Pinatubo eruption, *Science*, *259*(5100), 1411–1415.
- Newhall, C. G., and S. Self (1982), The volcanic explosivity index (VEI) an estimate of explosive magnitude for historical volcanism, *J. Geophys. Res.*, *87*(C2), 1231–1238.
- Ogata, T., S.-P. Xie, A. Wittenberg, and D.-Z. Sun (2013), Interdecadal amplitude modulation of El Niño/Southern Oscillation and its impacts on tropical Pacific decadal variability, *J. Clim.*, *26*, 7280–7297, doi:10.1175/JCLI-D-12-00415.1.
- Ohba, M., H. Shioyama, T. Yokohata, and M. Watanabe (2013), Impact of strong tropical volcanic eruptions on ENSO simulated in a coupled GCM, *J. Clim.*, *26*(14), 5169–5182.
- Otterå, O. H., M. Bentsen, H. Drange, and L. Suo (2010), External forcing as a metronome for Atlantic multidecadal variability, *Nat. Geosci.*, *3*(10), 688–694.
- Pascolini-Campbell, M., D. Zanchettin, O. Bothe, C. Timmreck, D. Matei, J. H. Jungclaus, and H.-F. Graf (2015), Toward a record of Central Pacific El Niño events since 1880, *Theor. Appl. Climatol.*, *119*(1–2), 379–389.
- Pausata, F. S., L. Chafik, R. Caballero, and D. S. Battisti (2015), Impacts of high-latitude volcanic eruptions on ENSO and AMOC, *Proc. Natl. Acad. Sci. U.S.A.*, *112*(45), 13,784–13,788.
- Raible, C. C., et al. (2016), Tambora 1815 as a test case for high impact volcanic eruptions: Earth system effects, *Wiley Interdiscip. Rev. Clim. Change*, *7*, 569–589.
- Rasmusson, E. M., and T. H. Carpenter (1982), Variations in tropical sea surface temperature and surface wind fields associated with the southern oscillation/El niño, *Mon. Weather Rev.*, *110*(5), 354–384.
- Ratnam, J., S. Behera, Y. Masumoto, and T. Yamagata (2014), Remote effects of El Niño and Modoki events on the austral summer precipitation of southern Africa, *J. Clim.*, *27*(10), 3802–3815.
- Rind, D., and A. Lacis (1993), The role of the stratosphere in climate change, *Surv. Geophys.*, *14*(2), 133–165.
- Rind, D., N. Balachandran, and R. Suozzo (1992), Climate change and the middle atmosphere. Part II: The impact of volcanic aerosols, *J. Clim.*, *5*(3), 189–208.
- Robock, A. (2000), Volcanic eruptions and climate, *Rev. Geophys.*, *38*(2), 191–219.
- Robock, A., and J. Mao (1995), The volcanic signal in surface temperature observations, *J. Clim.*, *8*(5), 1086–1103.
- Robock, A., K. E. Taylor, G. L. Stenchikov, and Y. Liu (1995), GCM evaluation of a mechanism for El Niño triggering by the El Chichón ash cloud, *Geophys. Res. Lett.*, *22*(17), 2369–2372.
- Ropelewski, C. F., and M. S. Halpert (1987), Global and regional scale precipitation patterns associated with the El Niño/Southern Oscillation, *Mon. Weather Rev.*, *115*(8), 1606–1626.
- Ropelewski, C. F., and M. S. Halpert (1996), Quantifying Southern Oscillation-precipitation relationships, *J. Clim.*, *9*(5), 1043–1059.
- Santer, B. D., et al. (2014), Volcanic contribution to decadal changes in tropospheric temperature, *Nat. Geosci.*, *7*(3), 185–189.
- Sato, M., J. E. Hansen, M. P. McCormick, and J. B. Pollack (1993), Stratospheric aerosol optical depths, 1850–1990, *J. Geophys. Res.*, *98*(D12), 22,987–22,994.
- Schnetzler, C., G. Bluth, A. Krueger, and L. Walter (1997), A proposed volcanic sulfur dioxide index (VSI), *J. Geophys. Res.*, *102*(B9), 20,087–20,091.
- Seager, R., S. E. Zebiak, and M. A. Cane (1988), A model of the tropical Pacific sea surface temperature climatology, *J. Geophys. Res.*, *93*(C2), 1265–1280.
- Soden, B. J. (2000), The sensitivity of the tropical hydrological cycle to ENSO, *J. Clim.*, *13*(3), 538–549.
- Soden, B. J., R. T. Wetherald, G. L. Stenchikov, and A. Robock (2002), Global cooling after the eruption of Mount Pinatubo: A test of climate feedback by water vapor, *Science*, *296*(5568), 727–730.
- Stenchikov, G. (2009), Chapter 4—The role of volcanic activity in climate and global change, in *Climate Change: Observed Impacts on Planet Earth*, edited by T. M. Letcher, pp. 77–102, Elsevier, Amsterdam.
- Stenchikov, G., A. Robock, V. Ramaswamy, M. D. Schwarzkopf, K. Hamilton, and S. Ramachandran (2002), Arctic Oscillation response to the 1991 Mount Pinatubo eruption: Effects of volcanic aerosols and ozone depletion, *J. Geophys. Res.*, *107*(D24), 4803, doi:10.1029/2002JD002090.
- Stenchikov, G., K. Hamilton, R. J. Stouffer, A. Robock, V. Ramaswamy, B. Santer, and H.-F. Graf (2006), Arctic Oscillation response to volcanic eruptions in the IPCC AR4 climate models, *J. Geophys. Res.*, *111*, D07107, doi:10.1029/2005JD006286.
- Stenchikov, G., T. L. Delworth, V. Ramaswamy, R. J. Stouffer, A. Wittenberg, and F. Zeng (2009), Volcanic signals in oceans, *J. Geophys. Res.*, *114*, D16104, doi:10.1029/2008JD011673.
- Stenchikov, G. L., I. Kirchner, A. Robock, H.-F. Graf, J. C. Antuna, R. Grainger, A. Lambert, and L. Thomason (1998), Radiative forcing from the 1991 Mount Pinatubo volcanic eruption, *J. Geophys. Res.*, *103*(D12), 13,837–13,857.
- Stevenson, J. W., and P. P. Niiler (1983), Upper ocean heat budget during the Hawaii-to-Tahiti shuttle experiment, *J. Phys. Oceanogr.*, *13*(10), 1894–1907.
- Stevenson, S., B. Otto-Bliesner, J. Fasullo, and E. Brady (2016), “El Niño like” hydroclimate responses to last millennium volcanic eruptions, *J. Clim.*, *29*(8), 2907–2921.
- Stevenson, S., J. Fasullo, B. Otto-Bliesner, R. Tomas, and C. Gao (2017), The role of eruption season in reconciling model and proxy responses to tropical volcanism, *Proc. Natl. Acad. Sci. U.S.A.*, in press.

- Timmreck, C. (2012), Modeling the climatic effects of large explosive volcanic eruptions, *Wiley Interdiscip. Rev. Clim. Change*, 3(6), 545–564.
- Timmreck, C., H. Pohlmann, S. Illing, and C. Kadow (2016), The impact of stratospheric volcanic aerosol on decadal-scale climate predictions, *Geophys. Res. Lett.*, 43, 834–842, doi:10.1002/2015GL067431.
- Trenberth, K. E. (1997), The definition of El Niño, *Bull. Am. Meteorol. Soc.*, 78(12), 2771–2777.
- Turco, R., O. Toon, R. Whitten, P. Hamill, and R. Keesee (1983), The 1980 eruptions of Mount St. Helens: Physical and chemical processes in the stratospheric clouds, *J. Geophys. Res.*, 88(C9), 5299–5319.
- Vecchi, G., A. Wittenberg, and A. Rosati (2006), Reassessing the role of stochastic forcing in the 1997–1998 El Niño, *Geophys. Res. Lett.*, 33, L01706, doi:10.1029/2005GL024738.
- Vecchi, G. A., and A. T. Wittenberg (2010), El Niño and our future climate: Where do we stand?, *Wiley Interdiscip. Rev. Clim. Change*, 1(2), 260–270.
- Vecchi, G. A., et al. (2014), On the seasonal forecasting of regional tropical cyclone activity, *J. Clim.*, 27(21), 7994–8016.
- Wahl, E. R., H. F. Diaz, J. E. Smerdon, and C. M. Ammann (2014), Late winter temperature response to large tropical volcanic eruptions in temperate western North America: Relationship to ENSO phases, *Global Planet. Change*, 122, 238–250.
- Watanabe, M., et al. (2010), Improved climate simulation by MIROC5: Mean states, variability, and climate sensitivity, *J. Clim.*, 23(23), 6312–6335.
- Watanabe, M., J.-S. Kug, F.-F. Jin, M. Collins, M. Ohba, and A. T. Wittenberg (2012), Uncertainty in the ENSO amplitude change from the past to the future, *Geophys. Res. Lett.*, 39, L20703, doi:10.1029/2012GL053305.
- Wittenberg, A. T. (2002), ENSO response to altered climates, PhD thesis, 475 pp., Princeton Univ., Princeton, N. J. [Available at <http://www.gfdl.noaa.gov/~atw/research/thesis/>]
- Wittenberg, A. T. (2009), Are historical records sufficient to constrain ENSO simulations?, *Geophys. Res. Lett.*, 36, L12702, doi:10.1029/2009GL038710.
- Wittenberg, A. T. (2015), Low-frequency variations of ENSO, *US CLIVAR Variations*, 13(1), 26–31.
- Wittenberg, A. T., A. Rosati, N.-C. Lau, and J. J. Plushay (2006), GFDL's CM2 global coupled climate models. Part III: Tropical Pacific climate and ENSO, *J. Clim.*, 19(5), 698–722, doi:10.1175/JCLI3631.1.
- Wittenberg, A. T., A. Rosati, T. L. Delworth, G. A. Vecchi, and F. Zeng (2014), ENSO modulation: Is it decadal predictability?, *J. Clim.*, 27(7), 2667–2681, doi:10.1175/JCLI-D-13-00577.1.
- Yoder, J. A., and M. A. Kennelly (2003), Seasonal and ENSO variability in global ocean phytoplankton chlorophyll derived from 4 years of seaWiFS measurements, *Global Biogeochem. Cycles*, 17(4), 1112, doi:10.1029/2002GB001942.
- Yu, J.-Y., and S. T. Kim (2010), Identification of central-Pacific and eastern-Pacific types of ENSO in CMIP3 models, *Geophys. Res. Lett.*, 37, L15705, doi:10.1029/2010GL044082.
- Zanchettin, D., et al. (2016), The model intercomparison project on the climatic response to volcanic forcing (VolMIP): Experimental design and forcing input data for CMIP6, *Geosci. Model Dev.*, 9(8), 2701–2719, doi:10.5194/gmd-9-2701-2016.
- Zavala-Garay, J., C. Zhang, A. M. Moore, A. T. Wittenberg, M. J. Harrison, A. Rosati, J. Vialard, and R. Kleeman (2008), Sensitivity of hybrid ENSO models to unresolved atmospheric variability, *J. Clim.*, 21, 3704–3721, doi:10.1175/2007JCLI1188.1.
- Zebiak, S. E., and M. A. Cane (1987), A model El Niño–Southern Oscillation, *Mon. Weather Rev.*, 115(10), 2262–2278.
- Zheng, F., X.-H. Fang, J.-Y. Yu, and J. Zhu (2014), Asymmetry of the Bjerknes positive feedback between the two types of El Niño, *Geophys. Res. Lett.*, 41, 7651–7657, doi:10.1002/2014GL062125.