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Key Points:

- We examine eight CMIP5 models to study ocean response to tropical volcanic eruptions
- Ocean response includes lowered SST and heat content, raised sea ice mass and AMOC
- Evidence for coupling between eruptions and El Niño is very weak

Supporting Information:

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Ocean response to volcanic eruptions in Coupled Model Intercomparison Project 5 simulations

JGR

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Abstract We examine the oceanic impact of large tropical volcanic eruptions as they appear in ensembles of historical simulations from eight Coupled Model Intercomparison Project Phase 5 models. These models show a response that includes lowering of global average sea surface temperature by 0.1–0.3 K, comparable to the observations. They show enhancement of Arctic ice cover in the years following major volcanic eruptions, with long-lived temperature anomalies extending to the middepth and deep ocean on decadal to centennial timescales. Regional ocean responses vary, although there is some consistent hemispheric asymmetry associated with the hemisphere in which the eruption occurs. Temperature decreases and salinity increases contribute to an increase in the density of surface water and an enhancement in the overturning circulation of the North Atlantic Ocean following these eruptions. The strength of this overturning increase varies considerably from model to model and is correlated with the background variability of overturning in each model. Any cause/effect relationship between eruptions and the phase of El Niño is weak.

1. Introduction

This paper examines the impact of volcanic eruptions on the ocean using recent coupled model simulations from the Coupled Model Intercomparison Project Phase 5 (CMIP5). The past decade has seen considerable progress in understanding the ocean response to volcanic aerosols in coupled models. For major volcanic eruptions, reductions in net shortwave heating lead to a drop in global annual sea surface temperature (SST) by a few tenths of a degree Kelvin for a few years and a decrease in the temperature of the upper 3000 m by -0.01 to -0.02 K spanning decades [*Robock and Mao*, 1992; *Delworth et al.*, 2005; *Stenchikov et al.*, 2009; *Gregory*, 2010; *Booth et al.*, 2012; *Driscoll et al.*, 2012; *lwi et al.*, 2012].

In one key example of the potential impact of eruptions on the background state, *Stenchikov et al.* [2009] and *Otterå et al.* [2010] present model results showing major eruptions strengthening the North Atlantic Ocean meridional overturning circulation (AMOC) by a Sverdrup ($10^6 \text{ m}^3 \text{ s}^{-1}$) or more on 5–15 year time scales. These studies suggest two mechanisms for the increase: (1) changes in wintertime surface wind stress and/or (2) increases of the density of polar surface water and thus its ability to convect. Such changes correspond to a significant 5% increase in the Atlantic's contribution to the Earth's meridional heat transport. In contrast, *Mignot et al.* [2011] and *lwi et al.* [2012] find an increase for some eruptions but not for others, while *Zhong et al.* [2011] find a decrease in AMOC due to the stabilizing impact on surface salinity of increases in sea ice export into the North Atlantic Ocean. *Zanchettin et al.* [2013] conduct a series of experiments with a single model examining the climate response to the combined Tambora and Cosigüina eruptions and suggest that some of the differences in the AMOC and gyre circulation responses to eruptions are due to the differing background states of the model climates when the eruptions occurred. One alternative explanation is that the differences are due to differences in the models themselves. To explore this model dependence, we present a composite examination of the impact of volcanic aerosols on the ocean based on analysis of five major tropical volcanic eruptions represented in 36 historical simulations using eight recent climate models.

Tropical volcanic eruptions inject sulfuric gases into the lower stratosphere where they oxidize to form sulfate aerosols. The two largest of these during our 136 year period of interest are the 1883 eruption of

Table 1. The Name, Eruption Date, Location, and the Ice-Core Volcanic Index 2 (IVI2) Stratospheric Sulfate Injections [Gao et al., 2008] of Volcanoes Considered in This Study

Name	Date	Latitude	Longitude	IVI2 (Tg)
Krakatau	26–27 Aug 1883	6.1°S	105.4°E	22
Santa María	24–25 Oct 1902	14.8°N	91.6°W	4
Agung	17 Mar, 16 May 1963	8.3°S	115.5°E	17
El Chichón	3–4 Apr 1982	17.4°N	93.2°W	14
Pinatubo	15 Jun 1991	15.1°N	120.4°E	30

Krakatau and the 1991 eruption of Pinatubo (Table 1), both of which resulted in an increase of aerosol optical depth (AOD) at 0.55 μ m of ~0.15 (~20 Tg SO₂ injection) [*Gao et al.*, 2008; *Schmidt et al.*, 2011; *Crowley and Unterman*, 2012]. The 1902 eruption of Santa María, the 1963 eruption of Agung, and the 1982 eruption of El Chichón are of similar size according to *Sato et al.* [1993] with El Chichón slightly larger. But *Ammann et al.* [2003] ranks Santa María, the strongest of the three. These stratospheric sulfate aerosols spread globally over the course of months increasing planetary albedo and reducing downwelling shortwave radiation.

The characteristic lifetime of stratospheric aerosols is on the order of a year or two. But the climate response to volcanic eruptions persists far longer due to their impact on the more slowly evolving components of the Earth system, notably the ocean. Historical observations confirm model results suggesting that large eruptions can reduce global mean SST by -0.3 to -0.5 K for several years relative to the centennial temperature trend [*Rampino and Self*, 1982; *Mass and Portman*, 1989; *Angell and Korshover*, 1985; *Angell*, 1988; *Minnis et al.*, 1993]. Models have shown this anomalously cool water is subducted primarily into the upper 1000 m where it persists for many decades [*Delworth et al.*, 2005; *Gleckler et al.*, 2006; *Stenchikov et al.*, 2009; *Driscoll et al.*, 2012; *Iwi et al.*, 2012]. Water found at this depth has various surface source regions: the Southern Ocean, marginal seas such as the Mediterranean, and subtropical and subpolar frontal regions [*Schmitz*, 1995]. The rates of formation vary from year to year in each location and are modulated by surface climate. At higher latitudes, temperature anomalies increasingly act like passive tracers for a circulation driven by haline processes and are thus strongly linked to the atmospheric hydrologic cycle [*Talley et al.*, 2011].

Other impacts of volcanic aerosols on natural interannual to decadal climate variability (e.g., El Niño/Southern Oscillation (ENSO)) have been discussed before [*Robock and Mao*, 1995; *Stenchikov et al.*, 2006; *Evan et al.*, 2009; *Otterå et al.*, 2010; *Zanchettin et al.*, 2012; *Booth et al.*, 2012; *Zanchettin et al.*, 2013; *Haywood et al.*, 2013]. Some recent studies of proxy data and model simulations suggest that eruptions increase the likelihood of El Niño because reduced surface radiation reduces the zonal SST gradient along the equatorial Pacific [*Adams et al.*, 2003; *Mann et al.*, 2005; *Emile-Geay et al.*, 2008; *McGregor et al.*, 2010]. In contrast, *McGregor and Timmermann* [2011] find an initially La Niña-like response. Still other studies conclude there is no causative link between volcanic eruptions and ENSO [*Self et al.*, 1997; *Robock*, 2000; *Christiansen*, 2008]. Among the five volcanic eruptions we examine in this study, four are known to have occurred with concurrent strong El Niño events: Krakatau (1883), Agung (1963–1964), El Chichón (1982), and Pinatubo (1991), while Santa María, is somewhat uncertain (see, e.g., the delta-¹⁸O records of *Cobb et al.* [2003]). However, El Niño had already initiated prior to the eruptions of El Chichón and Pinatubo [e.g., *Robock et al.*, 1995; *Self et al.*, 1997].

Volcanic aerosol-induced heating of the tropical stratosphere may also enhance the pole-to-equator temperature gradient. In Northern Hemisphere winter, this larger temperature gradient produces a stronger polar vortex, and a corresponding stationary wave pattern of tropospheric circulation, pushing the system into a positive North Atlantic Oscillation/Arctic Oscillation (NAO/AO) circulation pattern [*Robock*, 2000]. NAO/AO may also be changed by the heating reduction due to depletion of stratospheric ozone by volcanic aerosols [*Stenchikov et al.*, 2002]. To add further complexity, NAO/AO is correlated with AMOC [*Otterå et al.*, 2010; *Zanchettin et al.*, 2012], which is separately related to volcanic aerosol loading through the hydrologic cycle link mentioned above. These connections suggest further links between volcanic aerosols and the background climate which we examine. In section 2, we briefly describe the eight widely used CMIP5 coupled models and the ensembles of simulations we will examine. In section 3, we compare the behavior of the response to our target volcanoes beginning with surface meteorological changes and then progressing to the response of the ocean thermodynamic variables and then volume transport. Section 4 presents the conclusions. **Table 2.** Some Basic Information About Historical Simulations Examined in This Study: Name, Source of Aerosol Loading, Ocean ModelResolution (Longitude \times Latitude \times Vertical), the Number of Ensemble Members Using All Forcing, and the Number That Do NotInclude Volcanic Aerosols

Name	Aerosol Source	Ocean Resolution	Ensembles
CCSM4	Ammann et al. [2003]	$1^{^\circ} \times (0.27^{^\circ} \sim 0.54^{^\circ}) \times 60$ lev	6/4
GFDL-CM3	Sato et al. [1993]	$1^{\circ} \times (1/3^{\circ} \sim 1^{\circ}) \times 50$ lev	5/3
GFDL-ESM2G	Sato et al. [1993]	$1^{\circ} \times (1/3^{\circ} \sim 1^{\circ}) \times 63$ layers ^a	3/0
GFDL-ESM2M	Sato et al. [1993]	$1^{\circ} \times (1/3^{\circ} \sim 1^{\circ}) \times 50$ lev	2/0
HadCM3	Sato et al. [1993]	$1.25^{\circ} \times 1.25^{\circ} \times 20$ lev	10/0
HadGEM2-ES	Sato et al. [1993]	$1^{\circ} \times (1/3^{\circ} \sim 1^{\circ}) \times 40$ lev	4/0
MPI-ESM-LR	Sato et al. [1993]	$1.5^{\circ} \times 1.5^{\circ} \times 40$ lev	3/0
MRI-CGCM3	<i>Interactive</i> ^b	$1.0^{\circ} \times 0.5^{\circ} \times 51$ lev	3/0

^aThe 63 layers are remapped onto 50 fixed depth levels for analysis.

^bDriscoll et al. [2012].

2. Data and Methods

Here we briefly describe the ensembles of simulations from eight models listed in Table 2. All models are driven by both natural and anthropogenic forcings. In addition, for two models we have ensemble members that lack volcanic forcing. Each ensemble member begins with different initial conditions, and in our study, we are interested in the period 1871–2005.

1. The National Center for Atmospheric Research Community Climate System Model, version 4 (CCSM4) atmospheric component model has a resolution of $0.9^{\circ} \times 1.25^{\circ}$ lat-lon and 26 vertical levels (L26), with a top at 2.9 mb [*Gent et al.*, 2011]. The ocean model uses Parallel Ocean Program (version 2) with $0.9^{\circ} \times 0.27 - 0.54^{\circ}$ L60



Figure 1. Global average all-sky and clear-sky downwelling surface shortwave flux anomalies (red and green) from the ensemble means of each of the eight models (units: W m⁻²). Black curves show aerosol optical depth of (top, scale on right) *Ammann et al.* [2003] and (second plot, scale on right) *Sato et al.* [1993]. Dashed lines show the starting time of the five eruptions (in chronological order: Krakatau, St. María, Agung, El Chichón, and Pinatubo).



Figure 2. Ensemble means of the rotated extended empirical orthogonal function (REEOF) first (red, decay ENSO phase) and second (green, buildup ENSO phase) principal components after removal of the climatological seasonal cycle. Dashed lines show the starting time of the five eruptions (in chronological order: Krakatau, St. María, Agung, El Chichón, and Pinatubo). Bottom plot shows the average of the time series for all the model ensembles. No trend has been removed.

resolution, a displaced North Pole, and a coupled sea ice model (true of all models considered here). CCSM4 uses volcanic aerosols provided by *Ammann et al.* [2003], which are roughly 20–30% larger in AOD than the estimates of *Sato et al.* [1993] and *Solomon et al.* [2007]. Here we examine an ensemble of six historical simulations with full natural and anthropogenic forcing and four that exclude volcanic aerosols.

- 2. We consider three models from the Geophysical Fluid Dynamics Laboratory (GFDL). The first is the Climate Model version 3 (CM3) [Donner et al., 2011, Griffies et al., 2011]. The atmospheric component model has a resolution of $2.0^{\circ} \times 2.5^{\circ}$ L48, with a model top at 0.01 mb. This model is characterized by a particularly strong indirect aerosol effect. The ocean component used in CM3 is Modular Ocean Model version 4.0 (MOM4.0), whose simulation is very similar to that of GFDL CM2.1 (MOM4.0, *Delworth et al.*, 2006; *Gnanadesikan*, 2006). The horizontal resolution is approximately $1^{\circ} \times (1/3^{\circ} \sim 1^{\circ}) \times 50$ /ev and a tripolar grid is applied [*Griffies et al.*, 2011]. We examine five ensemble members of the historical simulations and an additional three ensemble members that exclude volcanic aerosols.
- 3. The second is the GFDL Earth System Model GFDL-ESM2M [*Dunne et al.*, 2012] whose atmospheric horizontal resolution is the same as CM3, but the model top is in the lower stratosphere and it uses 24 vertical levels. The ocean component is MOM4.1 [*Griffies*, 2009], which differs from MOM4.0. The differences between the two models are described in detail in *Dunne et al.*, 2012. We examine two ensemble members of GFDL-ESM2M that are driven by historical forcing.
- 4. The third in this series, GFDL-ESM2G [*Dunne et al.*, 2012], differs from GFDL-ESM2M mainly in that is uses the Generalized Ocean Layered Dynamics (GOLD) isopycnal coordinate ocean model. The horizontal resolution of the model is approximately the same at the other two GFDL models. While the model's



Figure 3. Ensemble mean global net surface heat flux anomalies from the 6 year average prior to each eruption. The seasonal cycle and the ENSO signal have been removed separately from each, and the time series are then smoothed with a running four-season average. Bars show the standard error from the ensemble mean. Bottom right-hand plot shows the average of all the ensembles for each eruption. The unsmoothed figure version can be found in supporting information Figure S6.

native vertical grid is 63 isopycnal layers, we analyze output that has been remapped to a 50-level depth-based coordinate. GFDL-ESM2G differs from the MOM depth-based vertical coordinate models in that the numerics are computed on the isopycnal surfaces. A comparison of the simulations obtained from both the MOM and GOLD ocean models is found in *Dunne et al.* [2012]. We examine three ensemble members of the historical simulations from GFDL-ESM2G.

- 5. We consider two models produced by the Hadley Centre. The first, the Hadley Centre Coupled Model, version 3 (HadCM3), has an atmosphere with 2.5°×3.75°×19*lev* resolution with top pressure 5 mb [*Jones et al.*, 2003]. The HadOM3 ocean model has 1.25°×1.25°×20*lev* resolution. We examine 10 ensemble members driven by historical forcing.
- 6. The second Hadley Centre model we consider is the Hadley Centre Global Environmental Model version 2-Earth System (HadGEM2-ES) with a somewhat higher resolution atmosphere (1.25°×1.875°×38*lev*, with top pressure 3 mb) and ocean 1°×(1/3° ~ 1°)×40*lev* [*Collins et al.*, 2011]. We examine four ensemble members driven by historical forcing.
- 7. The Max Planck Institute-Earth System Model-Low Resolution (MPI-ESM-LR) atmosphere model, known as ECHAM6, has 1.875°×1.875°×47*lev* resolution with top pressure 10 mb. The MPIOM ocean model uses a bipolar orthogonal spherical coordinate system with approximately 1.5°×1.5°×40*lev* resolution [*Mauritsen et al.*, 2012]. We examine three ensemble members driven by historical forcing.
- 8. The final model we consider is the Meteorological Research Institute Coupled General Circulation Model version 3 (MRI-CGCM3). The MRI-AGCM3 atmosphere has 1.125°×1.125°×48*lev* resolution, and is notable for having a high top at 0.01 mb. The ocean-ice model is the MRI Community Ocean Model Version 3 with a tripolar grid and approximately 1.0°×0.5°×51*lev* resolution [*Yukimoto et al.*, 2012]. For MRI-CGCM3, as for MPI-ESM-LR, we have a more limited ensemble member set of three historical ensemble members.



Figure 4. Ensemble mean global average SST anomaly from the 6 year average prior to each eruption (similar to Figure 3). The seasonal cycle and the ENSO signal have been removed separately from each, and the time series have then been smoothed with a running fourseason average. Bottom right-hand plot shows observed SST processed in the same way (*y* axis same as other plots) and the average SST of all ensembles with *y* axis on the right. Bars show the standard error from the ensemble mean. The unsmoothed figure version can be found in supporting information Figure S7.

Most of the ocean component grids are irregular. For this comparison, all ocean variables have been remapped onto a regular 1° \times 1° horizontal grid using the same conservative mapping routine. All data are monthly averaged, and a seasonal cycle has been removed from each ensemble member as well as from the observation sets used for comparison. In some situations when examining the impact of volcanic eruptions on subsurface temperature, a centennial trend has also been removed in order to reduce the impact of centennial warming associated with greenhouse gasses on the multidecadal response to volcanoes.

The response of SST to an eruption is approximately a factor of 10 less than the SST variations associated with ENSO in the tropical Pacific and so it would require averaging 100 ensemble members to suppress an uncorrelated ENSO signal simply by ensemble averaging in this region. Lacking so many ensemble members, we carry out this separation by adopting the method of *Guan and Nigam* [2008] to identify the buildup and decay phases of ENSO and then ensemble average those time series, looking for a temporal relationship between the evolution of ENSO in the ensemble average and the timing of the eruptions. The method of *Guan and Nigam* [2008] involves computing the rotated extended empirical orthogonal functions (REEOFs) of SST in the Pan-Pacific domain (20°S–60°N, 120°E–60°W) for each ensemble member (seasonal cycle removed). We then associate the first two REEOFs with the canonical ENSO decay and buildup phases so that by averaging the REEOFs associated with individual ensemble members we can isolate the ENSO signal in the ensemble averages. This method is also applied to filter the impact of the ENSO buildup and decay phases from surface ocean variables such as SST and sea surface salinity (SSS).

While the emphasis in this study is on examining the model's behavior, we do include brief comparison to historical temperature observations. Simulated SST and subsurface heat content for the top 1 km of the

ocean $\left(\rho C_p \int_{-1000m}^{0} T dz\right)$ is represented here by the temperature anomaly from its climatological annual average, since ρ and C_p are essentially constant. The heat contents are compared to the Simple Ocean Data



Figure 5. Ensemble mean net surface heat flux difference between Northern and Southern Hemispheres (Northern minus Southern Hemispheres) from the 6 year average difference prior to each eruption (similar to Figure 3). Units are W m^{-2} . The seasonal cycle and the ENSO signal have been removed separately from each, and the time series have then been smoothed with a running four-season average. Bottom right-hand plot shows the average of all model ensembles for each eruption. Bars show the standard error from the ensemble mean. The unsmoothed figure version can be found in supporting information Figure S9.

Assimilation version 2.2.4 reanalysis estimates of *Carton and Giese* [2008] and *Giese and Ray* [2011]. SST is provided by HadISST1 (Hadley Centre Sea Ice and Sea Surface Temperature) [*Rayner et al.*, 2003]. The algorithms used to estimate SST from satellite infrared radiances are themselves quite sensitive to aerosol loading, with corrections in excess of 0.4 K. This connection between the phenomena and observing system means that satellite infrared SST observations of the ocean's response to El Chichón and Pinatubo must be viewed cautiously [*Zhang et al.*, 2004].

3. Results

We begin by examining global averaged atmospheric variables, identifying the response to volcanic aerosols by comparing average conditions during the 6 years prior with average conditions in the years following each eruption. Of the five eruptions, Krakatau causes the strongest reduction of ocean surface net surface shortwave flux (Figure 1). For this eruption on average, the all-sky shortwave flux is reduced by 4.5 W m⁻², a reduction followed in size by Pinatubo with a reduction of 3.4 W m⁻². In contrast the weakest, Agung and El Chichón, reduced all-sky shortwave radiation on average by 2.0 W m⁻². Among the eight models, CCSM4 has the greatest reduction, especially for Santa María, whose annual average all-sky shortwave radiative forcing in CCSM4 was reduced by an average of 5.5 W m⁻². The large reduction in this case is likely due to the use in CCSM4 of *Ammann et al.* [2003] aerosol loading estimates. The reductions of allsky radiation (which include cloudy regions) are about 70% of the clear-sky reductions in all models, which is comparable to the results in *Stenchikov et al.* [1998]. This fairly constant ratio is surprising since global mean cloud fraction varies widely among models, ranging from 50% in CCSM4 and HadCM3 to over 70% in GFDL-CM3. Surprisingly also, the global mean cloud fraction does not show a pronounced volcanic signal, except in CCSM4 (see supporting information Figure S4).



Figure 6. Ensemble mean Northern Hemisphere sea ice mass anomaly $(0-90^{\circ} N)$ from the 6 year average prior to the eruption (similar to Figure 3) (units: 1×10^{15} kg). The seasonal cycle and the ENSO signal have been removed separately from each, and the time series have then been smoothed with a running four-season average. Bottom right-hand plot shows the average of all model ensembles for each eruption. Bars show the standard error from the ensemble mean. The unsmoothed figure is in supporting information Figure S10.

All ensemble members show significant variability associated with ENSO. The buildup and decay phases of ENSO in the ensemble averages are shown in Figure 2. GFDL-ESM2M (which only has two ensemble members) and MRI-CGCM3 show buildup and decay phases of El Niño associated with the Pinatubo eruption although in both cases the buildup began prior to the eruption. Several other models, CCSM4, HadCM3, and MPI-ESM-LR, show an El Niño even further in progress when the eruption occurred. A different set of models, CCSM4, HadGEM2-ES, and MRI-CGCM3, show development of a small to medium sized El Niño in the year following Krakatau, while GFDL-ESM2G shifts into the La Niña phase. The evidence of a connection to the other eruptions is even weaker. If we accept the average of the eight model ensembles as our most



Figure 7. Ensemble mean sea surface salinity (units: psu) in the northern Atlantic sector for the 2 years following the Krakatau eruption minus the 6 year average prior to eruption. Two models are shown: (left) CCSM4 which has a weak AMOC response to volcanic eruptions and (right) GFDL-CM3 which has a strong AMOC response to volcanic eruptions.



Figure 8. Ensemble mean 2 year running averaged global ocean heat content anomalies (0-1000 m) from both models and Simple Ocean Data Assimilation (SODA) data (units: K). The seasonal cycle of SODA data from 1970 to 2005 is removed from all model outputs and the average of 1970 is subtracted from all data. No linear trend has been removed. Vertical dashed lines indicate the starting time of the five eruptions.



Figure 9. Global mean temperature anomaly scaled by the RMS annual temperature variability at each level. A quadratic trend computed separately for each model, at each level, is removed prior to computing anomalies in order to highlight the response to volcanic eruptions. Vertical axis is depth, (units: m).



Figure 10. Ensemble mean Atlantic meridional overturning stream function anomaly from the 6 year average prior to each eruption (units: Sv). The time series have been smoothed with a running eight-season average after removal of the climatological seasonal cycle. Bottom right-hand plot shows the average of all the ensembles for each eruption. The El Chichón time series are truncated so they do not continue past the 1991 Pinatubo eruption. Bars show the standard error from the ensemble mean. The unsmoothed version can be found in supporting information Figure S16.

stable estimate of how the climate responds to volcanic aerosols (treating each model as equally valid), then the lack of a coherent ENSO signal suggests there is no compelling evidence of a link between the timing of an eruption and a shift in phase of ENSO. We have made a similar effort to examine the relationship between the volcanic eruptions and the phase of NAO in boreal winter sea level pressure (not shown) and found similarly tenuous connections.

We next examine the changes in surface variables in response to the eruptions by comparing the global average values for the years following the eruption with the 6 year average prior to the eruptions. Surface net outgoing longwave flux and latent heat loss both show a negative anomaly with a lag of a year or so (resulting from their connection to SST). As a result of this lag and their lower amplitudes, the surface heat flux anomaly following the eruptions is dominated by the impact of the immediate reduction in net shortwave radiation, with the strongest and statistically significant response associated with Krakatau followed by Pinatubo (Figure 3, bottom right). Of the different models, we consider CCSM4 has the largest net flux reduction, particularly for Krakatau and Santa María. MRI-CGCM3 seems to be almost missing El Chichón, and its response to Agung is delayed by a year because the volcanic aerosol forcing data differ from either of the two analyses used to force the other models. The recovery time for all models is 3 years or so, reflecting the atmospheric lifetime of stratospheric sulfate aerosols.

We next turn our attention to the oceanic response. The lowest SSTs occur 1–2 years following each eruption (Figure 4). Among the eruptions, we consider the maximum seasonal cooling of -0.2 to -0.4 K is associated with Krakatau, values reasonably consistent with the (poorly constrained) observed estimates shown in the lower right-hand plot. Combining this cooling estimate with the net surface flux anomalies shown in Figure 3 allows us to estimate an effective mixing depth over which the cooling from Krakatau is distributed of about 85 m. For some models the Santa María and El Chichón eruptions elicit fairly weak responses in SST, the latter likely because its aerosol loading is significantly lower than the others.



Figure 11. AMOC volume transport response to eruptions versus intrinsic AMOC variability. AMOC transport response is the increase during years 7–8 following eruptions versus the natural AMOC RMS annually averaged variability computed during the 41 year gap in huge tropical eruptions 1920–1960. Comparisons are shown for the strong Krakatau eruptions, the weak El Chichón eruption, and the average response to the five eruptions.

The relaxation time of SST back to its pre-eruption value is ~5–7 years, a little longer than the recovery time of surface net fluxes (Figure 3). Among the eruptions we consider, Krakatau is again unusual in that SSTs remain depressed (although the response to Krakatau is affected by additional aerosol loading from the 1886 Tarawera eruption that followed Krakatau, Figure 1, top). Although correcting for ENSO generally makes the Pinatubo signal stronger (compared to supporting information Figure S8), Pinatubo still has a weaker SST response than Krakatau in several of the models. Models with high equilibrium climate sensitivity, such as HadGEM2-ES, GFDL-CM3, and MPI-ESM-LR [*Andrews et al.*, 2012], are sensitive to volcanic forcing as well. For example, with ~2.5 W m⁻² net flux decrease after Krakatau, SST is reduced by ~0.27 K in HadGEM2-ES.

The hemisphere in which eruptions occur alters the strength of the ocean response because of the weakness of cross-equatorial stratospheric aerosol exchange and the greater land mass in the Northern Hemisphere. For example, the sulfate aerosols from Agung were four times larger in the Southern Hemisphere than in the Northern Hemisphere, while El Chichón put twice as much aerosol loading into the Northern Hemisphere as into the Southern Hemisphere [*Sato et al.*, 1993]. This positive interhemispheric difference (Northern minus Southern Hemispheres) in net surface heat flux due to Agung is evident to some extent for all the eight models (Figure 5). The asymmetry is most evident in the two GFDL-ESM models.

We note two additional impacts of aerosols on the surface ocean. The first is an increase in both Arctic sea ice extent and total mass ranging from a few percent up to 10% within 2–3 years following an eruption (the sea ice mass change is shown in Figure 6). This sea ice mass increase persists for years longer than the SST decrease. Among the models we consider, MRI-CGCM3 shows the largest changes in sea ice areal extent and GFDL-CM3 shows the greatest change in mass, but these models also have the greatest variability from year to year. Interestingly, the sea ice mass recovery is more rapid after El Chichón and Pinatubo than the previous eruptions.

The second impact is an increase in SSS following the eruptions, which we believe is associated with multiyear aerosol-induced changes in the atmospheric hydrologic cycle (time series of global SSS are included in supporting information Figure S11). This effect is most evident in the North Atlantic subpolar gyre, two examples of which are shown for Krakatau (Figure 7).



Figure 12. Ensemble mean zonal integrated Atlantic meridional overturning transport stream function (color, units: Sv) and zonal mean density (contours, Cl (contour interval): 25 g m⁻³) response to the Krakatau eruption estimated from the difference between the 2 year average (years 7–8) following the eruption minus the average during the 6 years prior to the eruption.

We next consider the response of the subsurface ocean, beginning by examining the global average ocean heat content in the upper 1000 m (Figure 8). By focusing on the global average, we eliminate the impact of horizontal heat transport convergence. Restricting our attention to the upper 1000 m reduces the impact of model drift which in these models is <0.025 K/100 yr in this layer (e.g., supporting information Figure S21). Averaged over the upper 1000m temperature is depressed for multiple decades by as much as 0.03 K, comparable to observations. For HadCM3, the accumulated effect of Krakatau and Santa María may have obscured the anthropogenic warming trend for a century. On such long-timescales, the anomalies due to volcanic aerosols cannot easily be separated from the centennial 0–1000 m warming trend, which in this model is 0.3–0.4 K per century (equivalent to $\sim 4 \times 10^{23}$ J per century).

For two of our models (CCSM4 and GFDL-CM3), we also have historical run ensemble members lacking volcanic aerosols (shown in supporting information Figures S14 and S15). Comparison of global SST for the ensemble averages with and without volcanic aerosols confirms that the impact of the eruptions on SST can persist for 7 years or more and their concentration at the beginning and end of the 20th century reduces the curvature of a quadratic trend of SST (in other words, the *acceleration* of warming) computed over the 20th century.

We next consider the vertical structure of the global average temperature anomalies over the upper 2 km in which it is apparent that the signals associated with the volcanic eruptions are still evident at this depth (now, with trends removed, Figure 9). It is also interesting to note that the Krakatau/Santa María and Agung cool anomalies are particularly prominent and last longer in the deep ocean (below 1000 m) in GFDL-CM3 and HadGEM2-ES. The presence of the strong volcanically induced cooling is a reason why the centennial warming trends in these two models are the weakest among the models we consider (Figure 8). For these models, the eruptions are masking some of the effects of increasing greenhouse gasses.

The reduction in SST and increase in SSS in response to eruptions have the potential to reduce the stability of the oceanic water column and enhance overturning. These changes in the Atlantic overturning circulation



Figure 13. Same as Figure 12 except showing anomalous temperature (colors, units: K) and salinity (contour, CI: 0.05 psu).

contribute to meridional heat transport as suggested by *Stenchikov et al.* [2009] and *Otterå et al.* [2010]. To revisit this issue, we define an AMOC index time series as the time series of the maximum value of the zonally integrated volume transport stream function from 30°N to 60°N (*Otterå et al.* [2010] uses a similar definition, except with latitudinal limits of 20–50°N). The time series show AMOC is generally enhanced by volcanic eruptions, but the magnitudes vary, with GFDL-CM3 and GFDL-ESM2G showing transport enhancements of up to 2 Sv (Figure 10). In contrast, AMOC increases in CCSM4 and HadCM3 are only one quarter to one third as large. The average response is only a small fraction of a Sverdrup.

These differences in AMOC sensitivity to volcanic aerosols seem to reflect the overall variability of AMOC in various models. Those models with low AMOC variability in general show a weak response to volcances as well, while those with high variability show strong AMOC transport responses to eruptions (the relationship is most evident for the stronger eruptions, Figure 11). The dependence of the sensitivity of AMOC on the model variability likely explains some of the conflicting conclusions of previous studies noted in section 1. Comparisons of the GFDL-CM3 and CCSM4 ensemble mean results with the simulations excluding volcanic forcing (presented in supporting information) are consistent with this model dependence.

The variations in the AMOC time series among different models are reflected in variations in the latitudinal and vertical structure of their stratification and overturning circulation (Figures 12 and 13 show the changes in circulation and stratification in response to Krakatau). The stream function change is closely correlated with surface water density change, for which both salinity and temperature variations are important. Models with reduced surface temperature and increased salinity after volcanic eruptions have the most significant AMOC increase. For example, in GFDL-CM3 sea water density increases extending from surface to several hundred meters deep and from the equator to 60° N. In other models, temperature and salinity changes are generally in-phase, which causes a smaller density and stream function increase. When temperature and salinity

changes are generally in phase, the relative magnitudes become important. The two GFDL ESM models have larger surface temperature and salinity changes than other models. For other models such as CCSM4, HadCM3, and HadGEM2-ES, the small size of the in-phase changes of temperature and salinity after eruptions may help to explain their low AMOC response to volcanic eruptions.

4. Conclusions

Questions regarding the ocean response to changes in atmospheric aerosol loading in general have arisen in recent years [e.g., *Evan et al.*, 2009; *Otterå et al.*, 2010; *Booth et al.*, 2012; *Zhang et al.*, 2013]. Here we exploit the availability of new simulations produced as part of CMIP5 to revisit the ocean response to the five largest tropical volcanic eruptions of the last 135 years (Krakatau, Santa María, Agung, El Chichón, and Pinatubo) in a set of 36 historical climate simulations produced using eight widely used climate models. Our goals are to clarify the common features of the oceanic response and those properties that vary among models and simulations.

The first part of this study examines the response of global average and hemispheric properties to volcanic aerosols. All models show an annual average reduction in net surface solar radiation of $1-5 \text{ W m}^{-2}$, a drop in net surface heat flux of $1-3 \text{ W m}^{-2}$, and a resulting decline in SST of 0.1-0.3 K. Sea ice extent and mass also increase by about 5%. For smaller eruptions, SST may recover in a few years, but our results confirm the suggestion of previous studies [e.g., *Delworth et al.*, 2005; *Gleckler et al.*, 2006; *Stenchikov et al.*, 2009] that the impacts on ocean heat content of major eruptions may persist for decades. The increase in sea ice area and mass also persists well beyond the lifetime of stratospheric aerosols due to the reinforcing impact of solar albedo feedback and reductions in thermodynamic surface heat loss. The cool SST signal also penetrates into the subsurface ocean, lowering 0–1000 m temperature by an average of roughly 0.03 K, and persisting for many decades, masking some of the anthropogenic warming signal. Indeed, comparisons of simulations with and without volcanic aerosols (available for CCSM4 and GFDL-CM3) show that the concentration of eruptions in the early years of the 20th century and again near the end of the century may mask some of the acceleration of ocean heating that might otherwise have been observed.

A number of previous studies have explored the connection between volcanic eruptions and interannual to decadal climate variability. However, all have suffered from insufficient samples to make this discrimination (we estimate that approximately 100 ensemble members would be required to simply average out an incoherent ENSO signal). Using a rotated extended empirical orthogonal function analysis to further discriminate the natural and forced response, we find there is no compelling evidence of a link between the timing of an eruption and a shift in phase of ENSO. We have made a similar effort to examine the relationship between the volcanic eruptions and the phase of NAO in boreal winter sea level pressure and found similarly tenuous connections.

Finally, we examine the impact of eruptions on the overturning circulation, most particularly the overturning in the North Atlantic and confirm previous results suggesting that volcanic eruptions may enhance the overturning circulation (and consequently increase northward heat transport) due to increases in ocean surface density in the northern Atlantic sector. A comparison shows that the models vary by at least a factor of four in their sensitivity, the most sensitive models being those which have the most Atlantic meridional overturning variability in general.

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