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## RESEARCH LETTER

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## Key Points:

- North Atlantic Tropical Cyclone frequency is lower after major volcanic eruptions
- Proxy-based timeseries are key to testing volcanic forcing effects

## Supporting Information:

- Supporting Information S1

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## Consistent decrease in North Atlantic Tropical Cyclone frequency following major volcanic eruptions in the last three centuries

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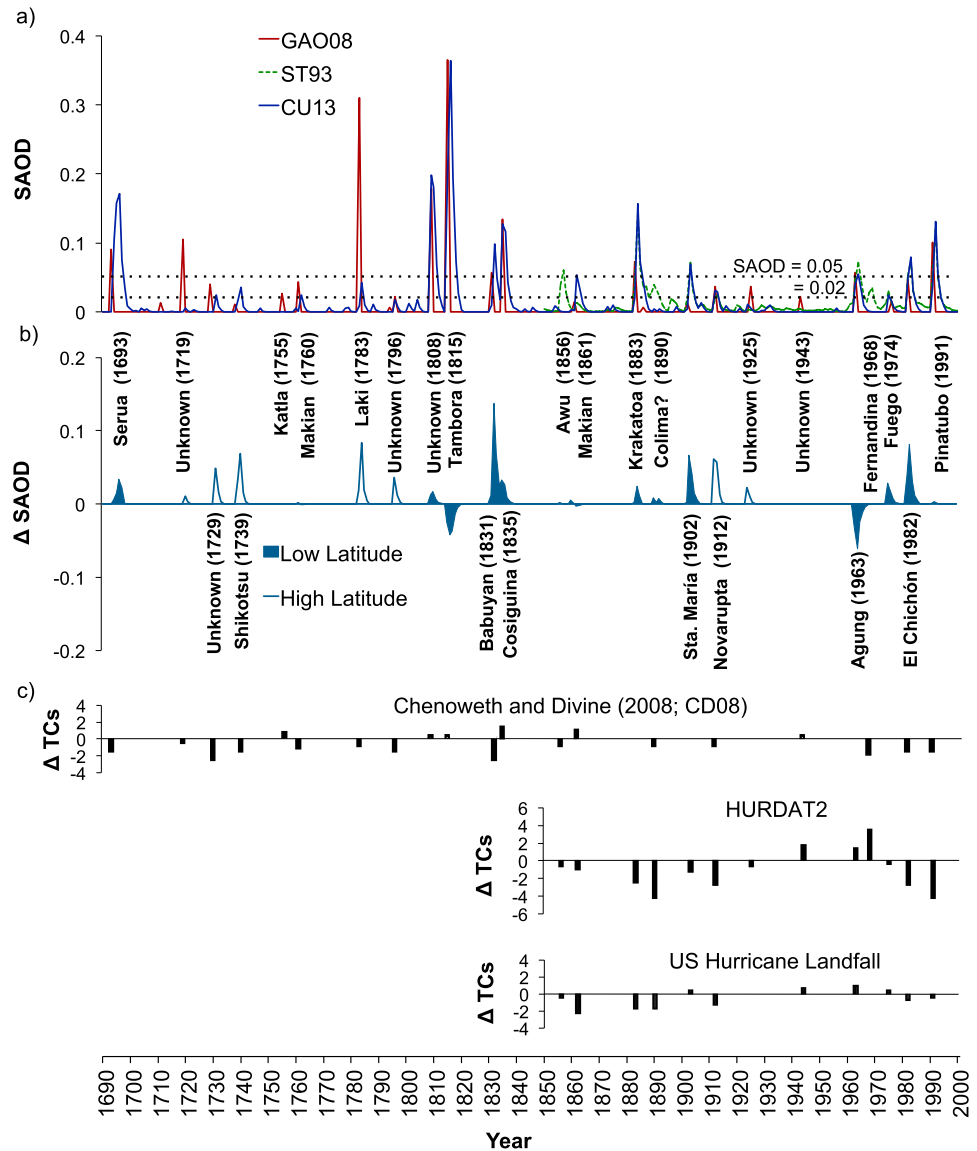
**Abstract** Injection of sulphate aerosols into the stratosphere following major volcanic eruptions alters global climate through the absorption and scattering of solar radiation. One proposed consequence is a decrease in North Atlantic Tropical Cyclone (TC) activity, as was observed following the El Chichón (1982) and Mount Pinatubo (1991) eruptions. We test this relationship using documentary and proxy reconstructions of major volcanic eruptions and TC frequency in the North Atlantic basin over the last three centuries. We find a consistent reduction in the number of TCs formed during the 3 years following major eruptions compared to the preceding 3 years, including after eruptions located at northern high latitudes. Our findings suggest that low-latitude eruptions reduce Atlantic TC frequency by decreasing local sea surface temperatures, whereas the mechanisms for the decrease in TC frequency following high-latitude eruptions are less clear and attribution is hampered by poor identification of these events.

### 1. Introduction

Tropical Cyclone (TC) activity over the North Atlantic basin was reduced after the volcanic eruptions of Mount Pinatubo (Philippines, 1991) and El Chichón (Mexico, 1982) [Evan, 2012]. These two eruptions have become the model for predicting how radiative forcing by stratospheric aerosols can alter TC formation [e.g., Emanuel et al., 2013; Caron et al., 2015]. Strong eruptions emit SO<sub>2</sub> that oxidizes to form long-lived stratospheric sulphate aerosols, which decrease sea surface temperatures by backscattering solar radiation and increase temperatures at the tropopause and in the lower stratosphere through absorption of longwave and short-wave radiation [Robock, 2000]. These climate system responses should affect TC number and intensity in the North Atlantic [Evan, 2012; Korty et al., 2012]; understanding how sulphate aerosols affect TC activity is important not only for understanding climate impacts of volcanic activity but also for anticipating the effects of geoengineering strategies that involve stratospheric radiation management [e.g., Robock et al., 2008; MacCracken, 2009; Robock et al., 2013].

Direct attribution of rare forcing events is hampered by high interannual to decadal background TC variability in the North Atlantic basin of both natural [e.g., Goldenberg et al., 2001; Camargo et al., 2010] and anthropogenic [e.g., Holland and Webster, 2007; Dunstone et al., 2013] origin. Eruption impact on TC activity may also depend on season, climate boundary conditions, and volcano location. For example, the consequences for hemispheric and zonal temperature gradients are very different for major low-latitude tropical eruptions, where aerosols have long stratospheric residence times and disperse globally [Robock, 2000], compared with high-latitude eruptions where aerosols are dispersed regionally and rarely reach the tropics [e.g., Oman et al., 2005; Schneider et al., 2009]. Moreover, both the Pinatubo and El Chichón eruptions occurred during the positive phase of the El Niño–Southern Oscillation (ENSO), when TC activity is reduced because of enhanced vertical wind shear [Gray, 1984; Goldenberg and Shapiro, 1996] and suppressed axisymmetric organization of deep convection [Goldenberg et al., 2001]. Assessing the impact of the June 1991 Pinatubo eruption is further complicated by a second stratospheric eruption—Cerro Hudson, Chile (46°S)—that occurred 2 months later.

Importantly, short instrumental observation records limit the analysis of the possible response of Atlantic TCs to major volcanic eruptions: aircraft reconnaissance started in 1944 and satellite data dates from the 1960s [Jarvinen et al., 1984; Neumann et al., 1985]. Longer records of TC frequency can be obtained from historical documents such as ship logs, meteorological journals, and newspapers [e.g., Chenoweth and Divine, 2008; Chenoweth, 2014]. Past volcanic aerosol records can also be reconstructed from sulphate in ice cores [e.g., Sato et al., 1993; Crowley et al., 2008; Gao et al., 2008]. Although the level of accuracy is lower than direct



**Figure 1.** Volcanic aerosol forcing records and post-eruption TC anomalies. (a) Global annual mean stratospheric aerosol optical depth (SAOD) since 1690 [Crowley and Unterman, 2013; Gao et al., 2008] and since 1850 [Sato et al., 1993]. (b) SAOD distribution between Northern (NH) and Southern (SH) Hemispheres for low-latitude and high-latitude volcanic eruptions (using CU13 data set) where a positive value means higher SAOD in the NH relative to the SH. (c) Mean difference in TC frequency for each major volcanic eruption  $\left( (\overline{\Delta TC_f})_y = \frac{(TC_y + TC_{y+1} + TC_{y+2})}{3} - \frac{(TC_{y-1} + TC_{y-2} + TC_{y-3})}{3} \right)$ , where  $y$  is the first post-eruption year, calculated using the reconstructed TC time series of CD08 [Chenoweth and Divine, 2008], HURDAT2, and the U.S. Hurricane landfall database [Landsea et al., 2004; Landsea and Franklin, 2013].

observations, employing these proxy sources of information increases the record length available for study. The same data permit hypothesis testing, such as whether TC genesis consistently shifts from the Mean Development Region (MDR; 8°–20°N and 20°–65°W) towards the north and west, as identified by Evan [2012] following the Pinatubo and El Chichón eruptions. In this study we use TC counts following eruptions over the last three centuries to address the following questions: (1) Do major volcanic eruptions reduce TC frequency in the North Atlantic Basin? (2) Is TC frequency sensitive to the location (latitude) of the volcanic eruption? (3) Does the location of North Atlantic TC formation shift following major volcanic eruptions? Finally, we explore the various volcanic aerosol-induced mechanisms that may explain the TC responses observed following eruptions over the last three centuries.

**Table 1.** List of Major Volcanic Eruptions in Last 300 Years Used in this Study<sup>a</sup>

Volcano	Latitude	Longitude	VEI	Eruption Month	Year	SAOD Range	ST93	GAO08	CU13	First Post-Eruption Year
<b>Serua<sup>b</sup></b>	<b>6.3°S</b>	<b>130°E</b>	<b>4</b>	<b>June</b>	<b>1693</b>	<b>0.09–0.17</b>		✓	✓	<b>1693</b>
Unknown	NH				1719	0.11		✓	-	1719
Unknown	NH				1729	0.04		✓	-	1730
Shikotsu	42.7°N	141.4°E	5	August	1739	0.04		-	✓	1740
Katla	63.6°N	19.1°W	5	October	1755	0.03		✓	-	1756
Makian	0.3°N	127.4°E	4	September	1760	0.04		✓	-	1761
Laki	64.4°N	17.3°W	4	May	1783	0.04–0.31		✓	✓	1783
Unknown	NH				1796	0.02		✓	-	1796
<b>Unknown<sup>b,c</sup></b>	<b>TROP</b>			<b>December</b>	<b>1808</b>	<b>0.18–0.2</b>		✓	✓	<b>1809</b>
<b>Tambora</b>	<b>8.3°S</b>	<b>118°E</b>	<b>7</b>	<b>April</b>	<b>1815</b>	<b>0.36–0.37</b>		✓	✓	<b>1815</b>
<b>Babuyan Claro<sup>b</sup></b>	<b>19.5°N</b>	<b>121.9°E</b>	<b>4</b>	<b>September</b>	<b>1831</b>	<b>0.06–0.1</b>		✓	✓	<b>1832</b>
<b>Cosiguina<sup>b</sup></b>	<b>12.9°N</b>	<b>87.6°W</b>	<b>5</b>	<b>January</b>	<b>1835</b>	<b>0.13–0.13</b>		✓	✓	<b>1835</b>
<b>Awu</b>	<b>3.7°N</b>	<b>125.5°E</b>	<b>3</b>	<b>March</b>	<b>1856</b>	<b>0.06</b>	✓	-	-	<b>1856</b>
<b>Makian</b>	<b>0.3°N</b>	<b>127.4°E</b>	<b>4</b>	<b>December</b>	<b>1861</b>	<b>0.05</b>	-	-	✓	<b>1862</b>
<b>Krakatau<sup>b</sup></b>	<b>6.1°S</b>	<b>105.4°E</b>	<b>6</b>	<b>August</b>	<b>1883</b>	<b>0.07–0.16</b>	✓	✓	✓	<b>1883</b>
Colima?	19.5°N	103.6°W	4	January	1890	0.04	✓	-	-	1890
<b>Santa María<sup>b</sup></b>	<b>14.8°N</b>	<b>91.6°W</b>	<b>6</b>	<b>October</b>	<b>1902</b>	<b>0.07–0.07</b>	✓	-	✓	<b>1903</b>
Novarupta	58.3°N	155.2°W	6	June	1912	0.02–0.04	✓	✓	✓	1912
Unknown	NH				1925	0.04	-	✓	-	1925
Unknown	NH				1943	0.02	-	✓	-	1944
<b>Agung</b>	<b>8.3°S</b>	<b>115.5°E</b>	<b>5</b>	<b>March</b>	<b>1963</b>	<b>0.06–0.07</b>	✓	✓	✓	<b>1963</b>
Fernandina	0.4°S	91.6°W	4	June	1968	0.03	✓	-	-	1968
Fuego	14.5°N	90.9°W	4	October	1974	0.03	✓	-	-	1975
<b>El Chichón<sup>b</sup></b>	<b>17.4°N</b>	<b>93.2°W</b>	<b>5</b>	<b>March</b>	<b>1982</b>	<b>0.05–0.08</b>	✓	✓	✓	<b>1982</b>
<b>Pinatubo</b>	<b>15.1°N</b>	<b>120.4°E</b>	<b>6</b>	<b>June</b>	<b>1991</b>	<b>0.1–0.13</b>	✓	✓	✓	<b>1991</b>

<sup>a</sup>The “check” and (“hyphen”) marks the eruptions included (not included) in the corresponding volcanic forcing index and reported as having at least (less than) 20% of the global annual mean SAOD of the Pinatubo (1991) eruption. VEI corresponds to the volcanic explosivity index, which measures the explosiveness of a volcanic eruption on a scale from 0 to 8 [Newhall and Self, 1982]. Low-latitude volcanic eruptions in bold had SAOD > 0.05.

<sup>b</sup>Indicates eruptions (SAOD > 0.05) that injected more aerosols into the NH with respect to the SH according to the CU13 volcanic forcing index (Figure 1).

<sup>c</sup>Date is from Guevara-Murua et al. [2014].

## 2. Data

### 2.1. Volcanic Forcing Indices

We use three global volcanic forcing indices to reconstruct past volcanism (Figure 1). Sato et al. [1993] (hereafter ST93) estimates Stratospheric Aerosol Optical Depth (SAOD) at 550 nm from ejecta volume (from Mitchell [1970] for 1850 to 1882), ground-based measurements of atmospheric extinctions (1883 to 1978), and satellite measurements (from 1979). Crowley and Unterman [2013] (CU13) also estimate SAOD at 550 nm but as derived from sulphate concentrations in 22 ice core records from Greenland and Antarctica over the last 1200 years. Gao et al. [2008] (GAO08) estimate the mass of stratospheric aerosols produced by eruptions in the last 1500 years based on 54 polar ice core records; we convert mass into SAOD by dividing the loadings by 150 Tg [Stothers, 1984].

For each volcanic forcing index, we classify eruptions based on magnitude, annual global average SAOD, and latitude (Table 1). The largest category (*All*) includes every eruption that caused 20% of the global annual mean SAOD of the Pinatubo eruption (SAOD ≈ 0.02). This threshold allows us to evaluate eruptions with the potential to impact climate (regional or global), while limiting inclusion of minor high-latitude eruptions recorded because of their proximity to the ice cores. In total, there were 25 major volcanic eruptions (16 low latitude and 9 high latitude) between 1690 and 2000 that fulfilled our criteria (Table 1). We subdivide the *All* category into two further categories based on volcano location: *Low Latitude* (20°N–20°S) and *High Latitude* (poleward of 40°N). Note that no confirmed annual average SAOD > 0.02 signals exist for subtropical or Southern Hemisphere (SH) high-latitude eruptions during this time. Finally, we consider a fourth category of the most extreme low-latitude volcanic eruptions (annual average SAOD > 0.05, ~40% of the global mean SAOD of the Pinatubo eruption; *Low Latitude* (SAOD > 0.05)).

### 2.2. North Atlantic Tropical Cyclones

We compare historical TC frequency data from the last 300 years using two data sets, each covering a different area of the North Atlantic Basin. The revised Atlantic Hurricane Database (HURDAT2) [Landsea and Franklin, 2013]

includes the annual number of TCs formed in the North Atlantic Basin since 1851. It uses data from land stations and ships at sea in the early period, supplemented with aircraft reconnaissance after 1944 and satellite data after the 1960s [Jarvinen et al., 1984]. We also use the HURDAT2 subset of U.S. Hurricane Landfalls reported from 1851 [Landsea et al., 2004; Landsea and Franklin, 2013]. Chenoweth and Divine [2008] (hereafter CD08) is the most comprehensive and longest document-based time series of TC activity within the North Atlantic Basin; it reconstructs the annual number of TCs since 1690 that passed near the Lesser Antilles (along the 61.5°W meridian from the coast of South America to 25°N). From both HURDAT2 and CD08 databases, we consider only tropical storms and hurricanes (sustained surface winds >17 m/s and >33 m/s, respectively).

### 3. Methods

To establish the first year in which TC frequency was affected by each eruption (*first post-eruption year*; Table 1), we consider the response time of mechanisms involved in TC formation. The El Chichón and Mount Pinatubo eruptions caused upper troposphere warming (70 hPa) and sea surface temperature (SST) cooling over the MDR coincident with the stratospheric aerosol injection [Evan, 2012]. Stratospheric forcing from low-latitude eruptions should affect TC formation rapidly and continue until the stratospheric aerosol concentrations decay to pre-eruption values (assumed to follow an *e*-folding residence time of 1 year) [Robock, 2000]. The most active months of TC activity over the North Atlantic Basin are August–October [Neumann et al., 1985]; we assume that at least two of these months need to be affected by aerosol forcing to consider the year as *post-eruption* (i.e., the first post-eruption year for low-latitude eruptions will only be the same calendar year as the eruption if it occurs before September). One third of the selected SAOD peaks are derived from Northern Hemisphere (NH) high-latitude eruptions, and over half of these are of “unknown” origin (Table 1). We propose that only eruptions from the first half of the year (January–June) will affect the TC season of the same calendar year. If the eruption is “unknown,” the first post-eruption year is assigned based on the first appearance date of the aerosols in the volcanic forcing index.

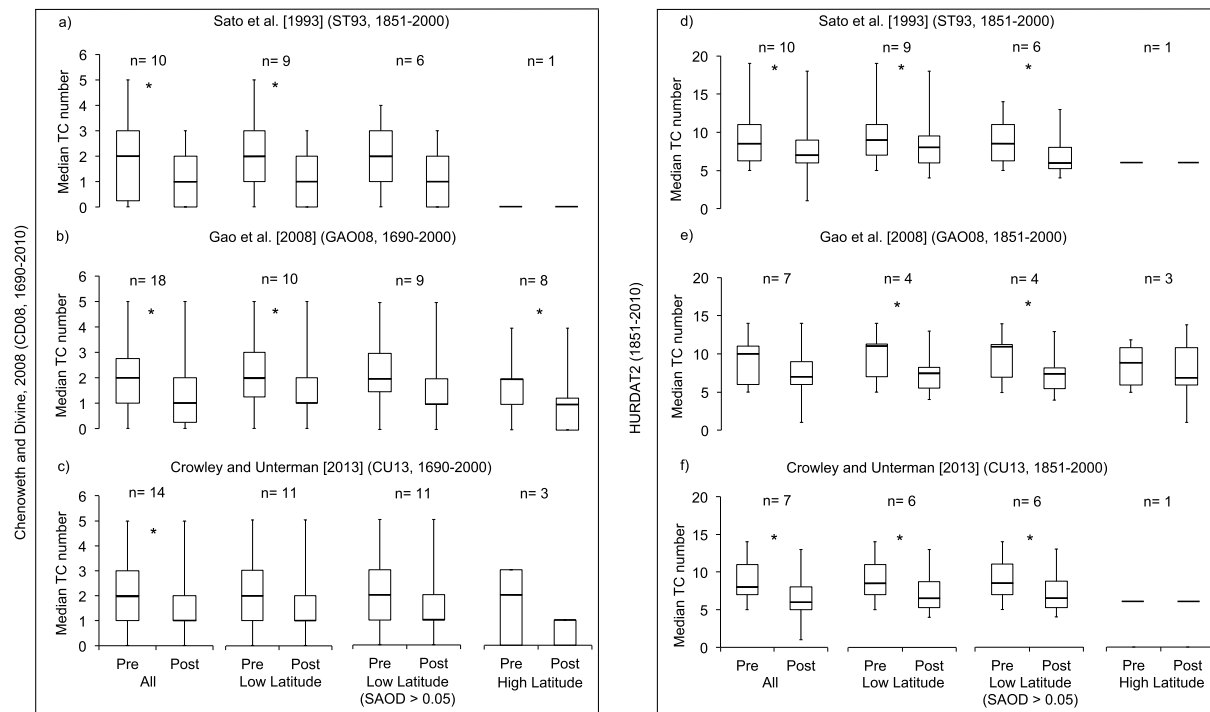
To evaluate which eruption categories decrease TC frequency, we compare the number of TCs formed during the 3 years preceding the eruptions (pre-eruption years) to those formed during the 3 years following the eruptions (post-eruption years). The statistical significance of any decrease in TC counts in each category is assessed by a Wilcoxon rank sum test and by a one-tailed *T* test (if normally distributed). We also examine changes in TC genesis location over the MDR between pre-eruption and post-eruption years using the HURDAT2 database. The genesis location is defined as the point where each storm reached tropical or extratropical storm status (maximum sustained surface winds >17 m/s). We determine the annual percentage of TCs formed in the MDR latitudes with respect to the whole North Atlantic Basin and test for a decrease in the post-eruption years as described above.

### 4. Results

#### 4.1. Testing for a Volcanic Impact on TC Number

Figure 2 shows the median and population distribution of TC counts for the 3 years before and after volcanic eruptions since 1851 (Figure 2a) and since 1690 (Figures 2b and 2c). From the CD08 database, these show a consistent reduction of 1 TC per year in the 3 post-eruption years, independent of the volcanic eruption category. The reduction in TC counts for *All* volcanic eruptions is statistically significant ( $p$  value < 0.05), independent of the volcanic forcing index used, while the decrease for *Low Latitude* eruptions is statistically significant in the GAO08 and ST93 volcanic forcing indices (statistics in Table S1 in the supporting information). The GAO08 volcanic forcing index also shows a statistically significant reduction in TCs ( $p$  value < 0.05) following the category of NH *High Latitude* eruptions; however, the relationship is not significant for CU13.

Results are consistent for both TC databases: HURDAT2 (Figures 2d–2f) shows a significant decrease of up to 2 TCs per year for *All* eruptions (CU13 and ST93 volcanic forcing indices) and up to 3.5 TCs per year for the eruption subsets *Low Latitude* and *Low Latitude* ( $SAOD > 0.05$ ) (Figures 2d–2f; supporting information Table S1). The reduction of 2 TCs per year after NH *High Latitude* eruptions (GAO08 since 1851,  $n = 3$ ), however, is not statistically significant (Figure 2e). By comparing TC counts of three post-eruption years with respect to the three pre-eruption years, our results are insensitive to potential bias associated with storm underrecording in the past. We further demonstrate this by repeating the exercise using (1) the Vecchi and Knutson [2008] (1878–2006) adjustment for the HURDAT2 database, which estimates TC undercounts by overlapping ship



**Figure 2.** TC counts (from (a–c) the *Chenoweth and Divine* [2008] database and (d–f) from the HURDAT2 database [*Landsea and Franklin*, 2013]) in the years before and after stratospheric volcanic aerosol forcing by major volcanic eruptions, according to volcanic forcing indices of *Sato et al.* [1993] (Figures 2a and 2d); *Gao et al.* [2008] (Figures 2b and 2e); and *Crowley and Unterman* [2013] (Figures 2c and 2f). Four categories of eruptions are considered, *All*, *Low Latitude*, *Low Latitude (SAOD > 0.05)* and *NH High Latitude*, and “*n*” indicates the number of eruptions in each category. The box plot outlines the median, 25 and 75% quartiles and range of TC counts for each population; only the median is displayed if *n* = 1. An “asterisk” denotes a statistically significant decrease in TCs between pre-eruption and post-eruption years at *p* value < 0.05 (Wilcoxon rank sum test; full statistics are given in the supporting information Table S1).

positions for the presatellite period with the TC tracks of the satellite era (supporting information Table S1), and (2) the *Chenoweth* [2014] (1851–1898) TC database of the full North Atlantic, which includes an average of 4 TCs per year more than the HURDAT2 database for the same period. Finally, we see a statistically significant decrease in Hurricane landfall frequency (1 Hurricane per year) using the U.S. Hurricane landfall database for *All*, *Low Latitude*, and *Low Latitude (SAOD > 0.05)* categories of volcanic eruptions in CU13, although not for GAO08 and ST93 (supporting information Figure S2 and Table S1).

#### 4.2. Changes in TC Genesis Location

There is a significant (14–19%) decrease in the proportion of TCs formed across MDR latitudes relative to total TC counts in the North Atlantic Basin after (1) *Low Latitude* and *Low Latitude (SAOD > 0.05)* eruptions in all volcanic forcing indices, and (2) for *All* eruptions in ST93 and CU13 (Wilcoxon rank sum test; supporting information Figure S1 and Table S1). The decrease in the proportion of TCs formed in the MDR after *NH High Latitude* volcanic eruptions (GAO08) is not statistically significant.

### 5. Discussion and Conclusion

The reduction in TC activity following the eruptions of Mount Pinatubo (1991) and El Chichón (1982) has been attributed to aerosol radiative forcing [*Evan*, 2012]. The present study supports this hypothesis and shows a consistent decrease in the number of TCs for a period of 3 years following major volcanic eruptions in the last three centuries (Figures 1 and 2). The decrease is also associated with a small reduction in the proportion of TCs formed within MDR latitudes relative to the total TC counts in the North Atlantic Basin (HURDAT2; Figure S1 and Table S1). Although this result is dominated by the strong shift in TC genesis location following both the Pinatubo (1991) and El Chichón (1982) eruptions and confounded by the incomplete HURDAT2 record for the MDR prior to the 1940s [e.g., *Landsea*, 2007], the post-eruption reduction of TCs recorded in the Lesser Antilles document-based TC time series [*Chenoweth and Divine*, 2008] (Figures 2a–2c) provides

additional support for a northward and westward shift in TC genesis away from these islands following major volcanic eruptions.

The significant decrease in TC frequency after *Low Latitude* eruptions coincides with a reduction in SSTs over the MDR according to the Hadley Centre Sea Ice and Sea Surface Temperature (HadISST1) database [Rayner *et al.*, 2003] (supporting information Figure S3). SST is a well-recognized factor controlling TC formation [e.g., Gray, 1968, 1979, 1981] and is a mechanism by which volcanic eruptions could affect TC frequency [Evan, 2012]. ENSO events also affect TC activity [Gray, 1984; Goldenberg and Shapiro, 1996], and since large-volcanic eruptions (SAOD of Pinatubo or greater) may increase the likelihood of El Niño [Emile-Geay *et al.*, 2008], any indirect El Niño-related dynamics would also contribute to the post-eruption decrease in TC activity. Since 1871, the extended instrumental Multivariate ENSO Index (MEI.tex; 1871–2005) [Wolter and Timlin, 2011] shows an ~18% increase in the number of moderate to strong El Niño years within the post-eruption relative to pre-eruption years used in our analysis. While these results indicate a role for ENSO in our results, this bias toward El Niño events is not found for post-eruption years earlier in the record (1690–1871) based on the ENSO proxy reconstruction of Gergis and Fowler [2009], although the same decrease in TC counts (1 TC per year according to the CD08 database) is observed for both the early (1690–1871) and late (1871–2005) post-eruption years. In support of Emile-Geay *et al.* [2008]'s conclusion that El Niño events are not triggered by aerosol forcing, and that even large low-latitude volcanic eruptions (e.g., Pinatubo; Figure 2) only increase the probability of a positive ENSO phase by 50%, we note that the eruptions of Serua (1693), Babuyan Claro (1831), and Krakatau (1883) were not directly followed by moderate to strong El Niños. The fact that a strong decrease in TC counts is observed after each of these eruptions (Figure 1)—Serua (4 TCs in total for 3 post-eruption years; CD08), Babuyan Claro (7 TCs in total; CD08), and Krakatau (1883; 8 TCs in total; HURDAT2)—reinforces the direct link between TC activity and radiative forcing by stratospheric aerosols.

Insights are also gained from considering TC frequency data after specific eruptions, including those that do not fit the general trends (Figure 1). There are a few strong (SAOD > 0.05) low-latitude volcanic eruptions, such as the “Unknown” eruption in December 1808 [Guevara-Murua *et al.*, 2014], Tambora (Indonesia, 8.3°S, 1815), Cosiguina (Nicaragua, 12.9°N, 1835), and Agung (Indonesia, 8.3°S, 1963), that are not followed by a decrease in TC counts (Figure 1). The apparent increase in TC frequency after the Cosiguina eruption could reflect the small number of TCs for the three Cosiguina pre-eruption years, which overlapped with the post-eruption years of the Babuyan Claro (Philippines, 19.5°N, 1831) eruption. The increase in TCs after the “Unknown” volcanic explosivity index 6 (VEI6) eruption of 1808 [Guevara-Murua *et al.*, 2014] may reflect the opposing influence of the strong 1809–1811 La Niña event [Gergis and Fowler, 2009], which may have countered volcanic forcing from an eruption that injected approximately twice the stratospheric aerosols of the Mount Pinatubo 1991 eruption [Crowley and Unterman, 2013].

The increased TC counts after both the Tambora (1 TC in total for the three post-eruption years; Figure 1) and Agung eruptions (0–4 TCs in total for the three post-eruption years depending on TC database; Figure 1) are more puzzling, since both produced abundant stratospheric sulphate and the Agung eruption was followed by two El Niño years (first and third post-eruption years). The case of Tambora (Indonesia, 1815), one of the largest volcanic eruptions of the last millennium, is particularly interesting since it had a strong influence on global climate and occurred within the coldest decade in the last 500 years [e.g., Jones *et al.*, 1995; Briffa *et al.*, 1998; D'Arrigo *et al.*, 2009]. Although small, the increase in TC frequency after the Tambora 1815 eruption also conflicts with the lower TC activity postevent predicted in a modeling study [Korty *et al.*, 2012]. Could this divergence reflect the pattern of aerosol distribution between the hemispheres? Both Tambora and Agung volcanoes lie at 8°S. Importantly, these two eruptions are the only major low-latitude eruptions since 1690 that injected more aerosols into the SH than the NH (Figure 1) [Crowley and Unterman, 2013; Arfeuille *et al.*, 2014]; and in the case of the 1963 Agung eruption, the stratospheric aerosols were mainly confined to the SH [Crowley and Unterman, 2013; Arfeuille *et al.*, 2014]. To understand the consequences for TC frequency, we need information from modeling experiments and more historic SH eruptions with an asymmetric distribution of aerosols between hemispheres.

The opposite situation, an asymmetric SAOD with a NH bias, was a contributing factor proposed by Evan [2012] to explain the decrease in TC formation and the north and westward shift in TC genesis location following the El Chichón eruption (Mexico, 17°N, 1982). Aerosols from El Chichón were retained almost exclusively within the NH stratosphere (Figure 1) causing a stronger aerosol direct radiative forcing in the NH with respect to the SH tropics [Evan, 2012]. This asymmetry forced a southward meridional SST gradient

( $-0.7^{\circ}\text{C}$  from  $30^{\circ}\text{S}$  to  $30^{\circ}\text{N}$ ) [Evan, 2012], a feature linked to an increase in vertical wind shear over the tropical North Atlantic, reduced TC activity, and a movement of the TC genesis location away from the MDR [Kossin and Vimont, 2007; Vimont and Kossin, 2007]. Of the low-latitude eruptions since 1690 that injected more aerosols into the NH, only Babuyan Claro (Philippines,  $19.5^{\circ}\text{N}$ , 1831) and Santa María (Guatemala,  $14.8^{\circ}\text{N}$ , 1902) caused NH-biased stratospheric aerosol distributions equivalent to or more extreme than the El Chichón eruption (Figure 1; CU13). The reduction of 2 TCs per year observed after Babuyan Claro supports Evan's [2012] hypothesis, although the effects of the Santa María eruption are less clear: TC counts decrease for the second and third post-eruption years in HURDAT2 but not in CD08 database. As with the asymmetric SAOD events where aerosol dispersal was disproportionately in the SH (Tambora and Agung), further evidence for the relation between radiative forcing and SST gradient requires data from a greater number of historic NH SAOD biased events like Babuyan Claro, Santa María, and El Chichón.

We also see a consistent decrease in the number of TCs (CD08) after the category *High Latitude* eruptions (eight NH volcanic events in GAO08 volcanic forcing index; Figure 2 and Table S1). As aerosols produced by NH high-latitude eruptions do not usually reach the tropics [e.g., Oman et al., 2005; Oman et al., 2006; Kravitz and Robock, 2011], the mechanism for TC reduction is not expected to involve changes in local upper tropospheric temperature and SSTs over the MDR. Indeed, no post-eruption SST signal is observed for high-latitude eruptions in the MDR (supporting information Figure S3 and Table S1), and there is no shift in TC genesis location (Table S1). Instead, the decrease in TC counts after high-latitude eruptions may be associated with the latitudinal asymmetry of aerosol distribution. Haywood et al. [2013] modeled stratospheric aerosol injection from NH volcanic eruptions and identified a southward Atlantic SST gradient (stronger SST decrease to the north) and regional hydroclimatic changes (weakening of the West African monsoon and reduced Western Sahel precipitation). The paucity of volcanic events and the relatively high number of high-latitude eruptions of unknown source add considerable uncertainty to interpretations of ways in which these eruptions affect North Atlantic climate. For example, our assessment that SST and genesis location in the MDR are not affected by NH high-latitude eruptions is based on only three eruptions (GAO08; Figure 1) that overlap with the SST (HadISST1 database) [Rayner et al., 2003] and HURDAT2 records, two of which are twentieth century "unknown" eruptions recorded only by GAO08 and omitted from other volcanic forcing indices [e.g., Arfeuille et al., 2014]. Improving the identification of NH high-latitude volcanic eruptions is therefore a critical next step.

To summarize, our study demonstrates the importance and utility of proxy-based time series for testing volcanic forcing effects on TC frequency over the North Atlantic basin. We demonstrate that TC counts are sensitive to the stratospheric aerosol radiative forcing caused by volcanic aerosols. Moreover, we highlight the importance of combining data from historical documents and other proxy sources with modeling studies to improve our understanding of how volcanic eruptions at all latitudes affect TCs and climate.

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