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Modulation of quasi-biennial oscillation by volcanic sulfate aerosols

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Large and moderate volcanic eruptions significantly impact the stratosphere by releasing sulfur dioxide, thereby affecting atmospheric dynamics. Mt. Pinatubo erupted in 1991 and induced a warming of 3 K in the stratosphere that led to prolonged easterly wind regimes of the Quasi-biennial oscillation (QBO). Using a three-member ensemble mean obtained from the ECHAM6-HAMMOZ model, we show the impact of eruptive volcanoes on the tropical stratosphere and the QBO from 2003 to 2013. Our simulations with volcanoes, when compared to simulations without volcanoes, show that volcanic sulfate aerosols enhanced the stratospheric aerosol optical depth (SAOD) 2 months after the eruptions of Rabaul (0.0024 ± 0.0018), Sarychev (0.0058 ± 0.0031), and Nabro (0.0075 ± 0.0053). The increase in mean tropical SAOD over the study period 2003–2013, driven by several volcanic eruptions, produced an average radiative forcing of $-0.45 \pm 0.21 \text{ W m}^{-2}$ at the top of the atmosphere (TOA) and $-0.47 \pm 0.26 \text{ W m}^{-2}$ at the surface in the tropical region. Volcanic aerosol precursors are injected into the tropical stratosphere through volcanic plumes, leading to an increase in sulfate aerosol concentrations of $59.6 \pm 11.1 \text{ ng m}^{-3}$ and heating rates of $0.05 \pm 0.02 \times 10^{-2} \text{ K d}^{-1}$ in the tropical lower stratosphere. The model simulations show that stratospheric heating caused by the volcanoes disrupts the QBO phases, resulting in the reversal of the easterly to westerly phase and vice versa. Our study shows that moderate and large volcanoes modulate the QBO.

KEYWORDS

quasi-biennial oscillation (QBO), radiative forcing, stratospheric heating, sulfate aerosols, volcanoes

1 Introduction

Strong and moderate volcanic eruptions inject sulfur dioxide (SO_2) gas into the stratosphere, where it forms sulfate aerosol particles (e.g., Hegerl et al., 2003). Sulfate aerosols contribute to stratospheric heating by absorbing infrared radiation, thus influencing stratospheric dynamics (Hegerl et al., 2003; Sigl et al., 2015; Santer et al., 2015). Sulfate aerosols formed from volcanic SO_2 injected into the stratosphere can remain there for about 1–5 years (Carn et al., 2016). Sulfate aerosols scatter shortwave radiation, resulting in a net negative radiative forcing at the top of the atmosphere (TOA) and surface cooling (Vernier et al., 2011).

Over the past decades, multiple moderate volcanic eruptions (Volcanic Explosivity Index; $VEI \geq 3$) have augmented the stratospheric aerosol load, resulting in the enhancement of SAOD and cooling of the surface (Vernier et al., 2011). Several satellite observations show an enhancement in SAOD following volcanic eruption; for example, Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observation (CALIPSO) satellite data showed an increase in SAOD (0.006–0.012) following multiple volcanic eruptions during 2008–2013 (Andersson et al., 2015; Friberg et al., 2018). Similarly, combined satellite data from the Stratospheric Aerosol and Gas Experiment (SAGE-II), Global Ozone Monitoring by Occultation of Stars (GOMOS), and CALIPSO showed an increase in SAOD of $20.4 \times 10^{-5} \text{ yr}^{-1}$ from 2000–2009, primarily driven by a sequence of medium-size volcanic eruptions (Vernier et al., 2011). Several model simulation studies also reported a similar increase in SAOD by volcanic eruptions, for example, an enhancement in globally mean SAOD ~ 0.007 due to volcanoes occurring globally during 2005–2014 (Schmidt et al., 2018).

The increase in SAOD following volcanic eruptions leads to a reduction in radiative forcing. For example, Schmidt et al. (2018) estimated a global mean RF of -0.08 W m^{-2} for 2005–2015 using CESM1-WACCM simulations performed with and without volcanic SO_2 emissions. Similarly, Ge et al. (2016) reported a global mean RF of -0.10 W m^{-2} for 2005–2011 using GEOS-Chem simulations. Consistent with these model estimates, satellite-based observational studies also indicate reductions in global mean radiative forcing following volcanic eruptions (Andersson et al., 2015; Friberg et al., 2018).

Sulfate aerosols absorb near-infrared solar radiation and thus heat the layers in which they reside (Kremser et al., 2016; Fadnavis et al., 2021a). Model studies have reported an increase in tropical lower-stratospheric (17–35 km) heating rate following the Nabro (0.003 K d^{-1}) and El Chichón (0.0045 K d^{-1}) eruptions (Pitari et al., 2016; Fadnavis et al., 2021a). In addition, volcanic aerosols transported from the extratropics into the tropics can also contribute to the tropical stratospheric heating rates (Oman et al., 2006; Brühl et al., 2015; Robock, 2000; Timmreck, 2012).

The volcanic sulfate aerosol-induced heating in the stratosphere can disrupt the thermal structure (Santer et al., 2015) and influence the QBO (Bittner et al., 2016; Brenna et al., 2019; Toohey et al., 2014; Diallo et al., 2018). The impact of large volcanoes on the QBO has been reported in a few studies; e.g., Labitzke (1994) reported that Mt. Pinatubo in 1991 induced a stratospheric warming of 3 K, leading to a prolonged easterly phase of the QBO. The ECHAM5 model simulations showed a shift in the QBO phase from easterly to westerly after the Mt. Pinatubo eruption (Thomas et al., 2009). Brenna et al. (2021) reported that the Los Chocoyos eruption ($VEI = 8$) (14.6°N , 91.2°W) disrupted the QBO periodicity for up to 10 years. Apart from volcanic aerosols, geoengineering studies wherein the injection of 2.5–5 Tg of SO_2 into the stratosphere is considered have also shown impacts on the QBO phase induced by stratospheric heating (Aquila et al., 2014; Pitari et al., 2016). Richter et al. (2017) found that in CESM1 (WACCM) simulations with an equatorial injection of 12 Tg SO_2 per year, the QBO period increased by approximately 3.5 years. Past study reported a prolonged westerly phase of the QBO (Aquila et al., 2014). In extreme cases, some studies even show a complete shutdown of the QBO. For example,

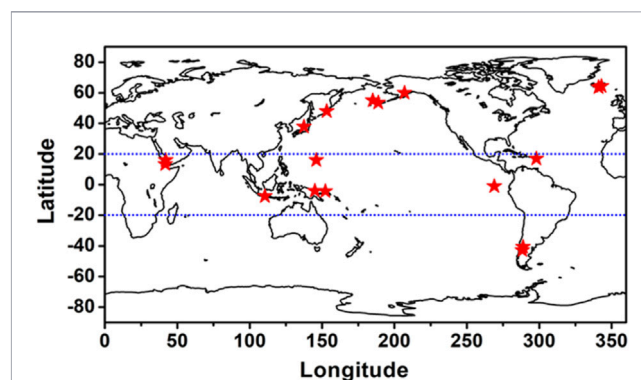


FIGURE 1
Spatial distribution of the moderate and large ($VEI \geq 3$) volcanic eruptions during 2003–2013. Red stars indicate the location of volcanoes. The blue dotted line indicates the tropical region ($20^\circ\text{S} - 20^\circ\text{N}$).

simulations using the ECHAM model indicate that a large injection of 8 Tg of SO_2 can cause such a shutdown, while the WACCM model shows the QBO shutting down with a smaller injection of just 2 Tg of SO_2 (Niemeier and Schmidt, 2017; Niemeier et al., 2020). The shutdown of the QBO depends not only on the injection rate and the model used for the simulation, but also on the injection strategy, including factors such as injection location, altitude, and injection frequency (Franke et al., 2021; Laakso et al., 2022).

Previous studies have focused on individual large tropical volcanic eruptions; here, we aim to investigate the influence of moderate volcanoes during 2003–2013 on the QBO using the state-of-the-art ECHAM6-HAMMOZ chemistry–climate model. During this period, there were a total of 17 volcanic eruptions ($VEI \geq 3$). Among these, 8 were in the tropics ($20^\circ\text{S} - 20^\circ\text{N}$), 7 in the northern extra-tropics ($20^\circ\text{N} - 90^\circ\text{N}$), and 2 in the southern extra-tropics ($20^\circ\text{S} - 90^\circ\text{S}$) (see Figure 1). The paper is structured as follows: Section 2 presents the model setup and details of the simulations performed for this investigation. Section 3 presents and discusses the outcomes derived from these simulations. Finally, in Section 4, the conclusions are made.

2 Model simulations and data analysis

2.1 Model description and experimental setup

We use the state-of-the-art aerosol–chemistry–climate model ECHAM6–HAMMOZ. It comprises the general circulation module ECHAM6, coupled to the aerosol and cloud microphysics module HAM (Stier et al., 2005; Schultz et al., 2018; Tegen et al., 2019). HAM predicts the nucleation, growth, evolution, and sinks of sulfate, black carbon (BC), particulate organic matter (POM), sea salt (SS), and mineral dust (DU) aerosols. The size distribution of the aerosol population is described by seven log-normal modes with prescribed variance as in the M7 aerosol module (Stier et al., 2005; Zhang et al., 2012). Moreover, HAM explicitly simulates the impact of aerosol species on cloud droplets and ice crystal formation. Other details of the model and emissions are reported by Fadnavis et al., 2017, 2019,

TABLE 1 Volcanic SO₂ (kt) emissions reaching the stratosphere and the uppermost tropical troposphere from the volcanic sulfur emission Inventory. The total annual amount and the explosive annual amount of global volcanic SO₂ emissions were calculated from satellite observations from 1979 to 2014 by [Carn et al. \(2016\)](#).

Year	Total (kt)	Explosive (kt)
2003	679	679
2004	2,997	410
2005	4,634	2,501
2006	1,347	661
2007	712	122
2008	2,624	2,318
2009	1,934	1,379
2010	1,470	867
2011	6,030	4,310
2012	763	563
2013	185	180

[2021b](#), [2021c](#). Anthropogenic aerosol emissions are from the Community Emissions Data System (CEDS) ([Hoesly et al., 2019](#)). The volcanic SO₂ emissions inventory used in the model simulations is based on [Carn et al. \(2017\)](#). The volcanic SO₂ emissions used in the model are tabulated in [Table 1](#).

The model simulations are performed at T63 spectral resolution, corresponding to a horizontal resolution of $1.875^\circ \times 1.875^\circ$. Vertically, 95 hybrid σ -p levels are used from the surface up to 0.001 hPa (approximately 90 km). The simulations have been carried out with a time step of 20 min. Monthly varying Atmospheric Model Intercomparison Project (AMIP) sea surface temperature (SST) and sea ice cover (SIC) ([Taylor et al., 2000](#)) were used as lower boundary conditions. The dust emission parameterisation is the same across all simulations and follows [Tegen et al. \(2002\)](#). We performed two experiments: (1) a VAL simulation, in which all aerosols, including volcanic emissions, are considered; and (2) a VAL0 simulation, in which all volcanic aerosol emissions are switched off globally. All other emissions (anthropogenic and natural) are kept the same in both experiments. Both simulations are performed for the period 2003–2013. To reduce the influence of internal variability, an ensemble-mean approach was adopted for both experiments. Three ensemble members were simulated for the period 2003–2013 after a three-year spin-up corresponding to approximately one QBO cycle (January 2000–December 2002). The ensemble members were initialized on 1, 2, and 3 January 2000 with small perturbations applied to the initial atmospheric state, and each simulation was integrated through 31 December 2013. This approach allows the ensemble members to evolve independently. In the VAL0 simulations (three ensemble members), the volcanic aerosol emissions were switched off on 1 January 2003. The VAL and VAL0 simulations ended on 31 December 2013. Each ensemble member provides daily output for the 2003–2013 simulation period. The simulations are restricted to 1 decade since simulations for longer time periods at a daily scale, extending to 80 km, are computationally expensive. To investigate

the effects of volcanic aerosol during 2003–2013, we analyse the difference between VAL and VAL0 simulations. Ensemble-mean differences between the VAL and VAL0 simulations were evaluated using a two-tailed Student's t-test at the 95% confidence level. For each month, each experiment provides approximately 30 daily values per ensemble member, resulting in about 90 samples when the three ensemble members are combined. Thus, the use of three ensemble members increases the sample size and improves the robustness of the estimated anomalies. In volcano simulations, we assumed that all volcanoes erupted at their respective times and locations. There are 17 volcanoes globally (see [Figure 1](#)) in the study period, including large and moderate volcanoes ($VEI \geq 3$) (see [Table 2](#)). VEI is used to classify the explosive magnitude of volcanic eruptions based on erupted tephra volume and eruption plume height. It is measured on a logarithmic scale from 0 to 8 ([Newhall and Self, 1982](#)).

2.2 The global space-based stratospheric aerosol climatology (GloSSAC)

We compared simulated SAOD with Global Space-based Stratospheric Aerosol Climatology (GloSSAC) SAOD. The GloSSAC provides a 43-years record of stratospheric aerosol properties, focusing on aerosol extinction coefficients at 525 nm and 1020 nm. This climatology is primarily based on data from the Stratospheric Aerosol and Gas Experiment (SAGE) instruments until mid-2005, followed by the Optical Spectrograph and InfraRed Imager System (OSIRIS) and the Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observation (CALIPSO) until 2017 ([Kovilakam et al., 2023](#)). After 2017, the SAGE III/ISS data are used in GloSSAC. GloSSAC also integrates data from other satellite platforms, as well as ground-based, airborne, and balloon-borne instruments, to achieve a global and temporally continuous dataset ([Kovilakam et al., 2023](#)). The dataset spans from 1979 to 2023, covering latitudes from 80°S to 80°N, with a horizontal resolution of 500–1,000 km and a vertical resolution from 100 m to less than 1 km, offering a comprehensive representation of stratospheric aerosol variability ([Kovilakam et al., 2023](#); [Thomason and Knepp, 2023](#); [NASA/LARC/SD/ASDC, 2022](#)). Here, we used GloSSAC Version 2.2 SAOD at 525 nm from the tropopause to 40 km for the period 2003–2013 (<https://asdc.larc.nasa.gov/project/GloSSAC>).

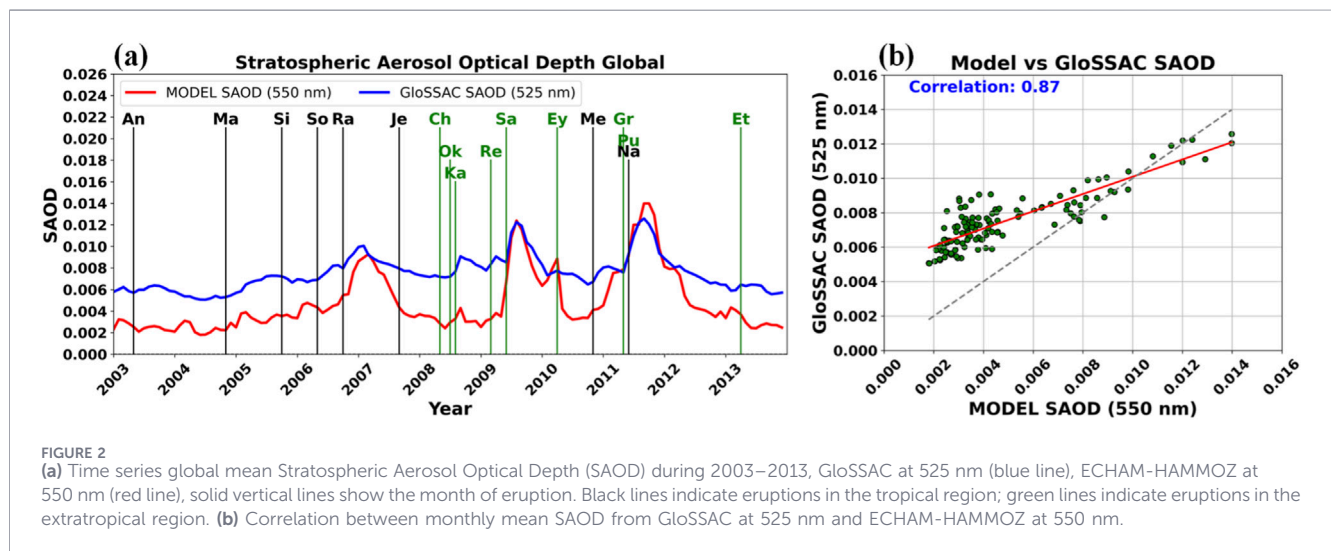
3 Results

3.1 Model evaluation

[Figure 2a](#) compares the simulated global mean SAOD at 550 nm with GloSSAC observations at 525 nm during 2003–2013. The model captures the major temporal variations in SAOD, including enhanced SAOD following the Rabaul (2006), Sarychev (2009), and Nabro (2011) eruptions and low aerosol loading during the volcanically quiescent period of 2003–2005. The agreement is particularly good for the Sarychev eruption, while the model slightly underestimates the Rabaul (by ~ 0.0002) and overestimates the Nabro peak (by ~ 0.0013). The model simulation shows a strong correlation with GloSSAC SAOD ($r = 0.87$; [Figure 2b](#)), indicating that the model reasonably captures the timing and relative variability of volcanic aerosol perturbations.

TABLE 2 List of volcanoes from 2003 to 2013 and their Volcanic Explosivity Index (VEI).

Sr No.	Volcano	Month year	Location	VEI
1	Anatahan (An)	May 2003	16 °N, 146 °E	3
2	Manam (Ma)	November 2004	4.08 °S, 145.04 °E	4
3	Sierra negra (Si)	Oct 2005	1 °S 91 °W	3
4	Soufrière hills (So)	May 2006	17 °N 62 °W	3
5	Rabaul (Ra)	October 2006	4.27 °S, 152.20 °E	4
6	Jebel at tair (Je)	Sep 2007	16 °N 42 °E	3
7	Chaiten (Ch)	May 2008	42.83 °S, 72.65 °W	4
8	Mt okmok (Ok)	July 2008	53.48 °N, 168.17 °W	4
9	Kasatochi (Ka)	August 2008	55.00 °N, 175.00 °W	4
10	Redoubt (Re)	Mar 2009	60 °N 153 °W	3
11	Sarychev (Sa)	June 2009	48.00 °N, 153.20 °E	4
12	Eyjafjallajökull (Ey)	April 2010	63.63 °N, 19.60 °W	4
13	Merapi (Me)	November 2010	7.54 °S, 110.44 °E	4
14	Grimsvotn (Gr)	May 2011	64.42 °N, 17.33 °W	4
15	Nabro (Na)	June 2011	13.37 °N, 41.70 °E	4
16	Puyehue-cordon caulle (Pu)	June 2011	40.59 °S, 72.12 °W	5
17	Etna (Et)	April 2013	37.89 °N, 137.48 °E	3



Despite this good temporal agreement, there are differences in simulated SAOD and GloSSAC during 2003–2013, particularly during the volcanically quiescent period of 2003–2005 (~ 0.003) and after 2012 (~ 0.0025). Part of this difference arises from uncertainties in the GloSSAC record associated with satellite merging procedures, instrument transitions, and wavelength differences between GloSSAC and the model (Kovilakam et al., 2020). Also, differences are partially due to uncertainties in the model, e.g., volcanic emission inventories, transport, and aerosol optical properties (Fadnavis et al., 2013; Tilmes et al., 2023). The

underestimation of model SAOD will relatively affect the simulated radiative forcing, heating rates, and QBO.

3.2 Impact of volcanoes on the SAOD, radiative forcing, and heating rates

Figure 3 illustrates simulated anomalies of SAOD (VAL-VAL0) for (1) the global mean and (2) the zonal mean averaged over the tropical region (20° S – 20° N). During the volcanically quiescent period from 2003 to 2005, SAOD anomalies increased

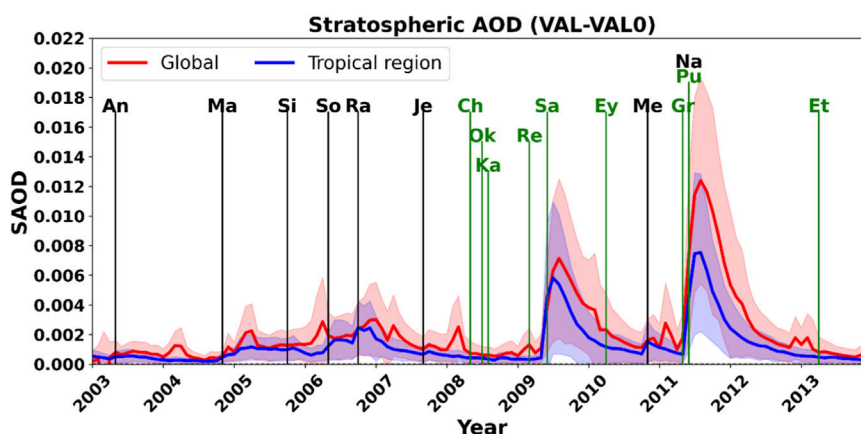


FIGURE 3
Time series of stratospheric aerosol optical depth (SAOD) (above tropopause to 1 hPa) during 2003–2013 using ECHAM6–HAMMOZ model simulations (VAL–VAL0). Global mean (red line) and zonal mean averaged over the tropical region (20° S – 20° N). Solid vertical lines show the month of eruption. Black lines indicate the eruption in the tropical region, and green lines indicate the eruption in the extratropical region.

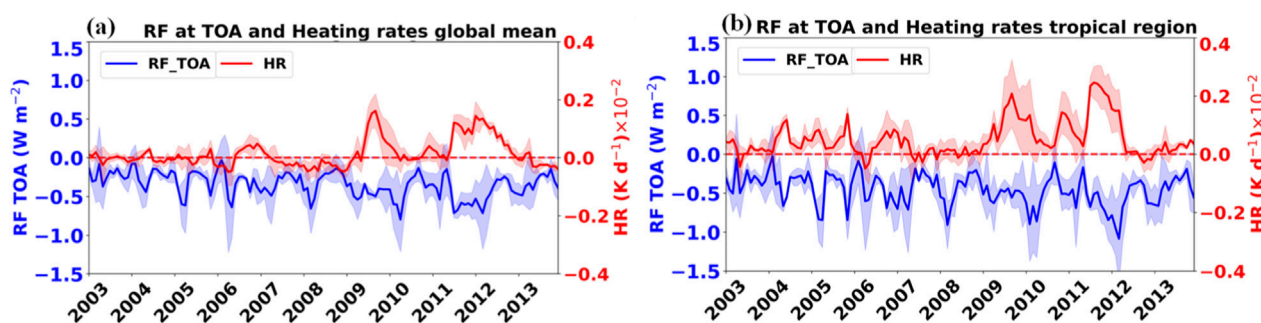


FIGURE 4
(a) Global mean radiative forcing at TOA (W m^{-2}) (blue line), stratospheric heating rates ($\text{K d}^{-1} \times 10^{-2}$) (red line) (above tropopause to 1 hPa) from VAL–VAL0 during 2003–2013, (b) same as (a) but zonally averaged over the tropical region (20° S – 20° N).

by 0.0006 ± 0.0004 in the tropical region and by 0.00096 ± 0.00081 globally. During 2005–2013, when a series of volcanic eruptions occurred (Table 2), SAOD averaged over this period increased to 0.0014 ± 0.0011 in the tropics and 0.0023 ± 0.0019 globally.

Notably, Figure 3 highlights prominent SAOD peaks in 2006 (Rabaul), 2009 (Sarychev), and 2011 (Nabro) in the tropical region (Supplementary Table S1). Andersson et al. (2015) reported that the Kasatochi eruption in 2009, the Sarychev eruption in 2010, and the Nabro eruption in 2011 collectively enhanced global SAOD by 30% between 2008 and 2012 relative to the reference period 1999–2002. Our ECHAM–HAMMOZ model simulations show an increase in SAOD by 38.2% due to Kasatochi, Sarychev, and Nabro eruptions during 2008–2012, which is in good agreement with the findings of Andersson et al. (2015), indicating the reliable performance of the model. Additionally, GOMOS satellite data in the tropical region show SAOD ranging from 0.002 to 0.0055 during 2002–2008 (Vernier et al., 2011). Our model shows SAOD values ranging from 0.001 to 0.005 during the same period, which also aligns closely with the observations, indicating good agreement.

Further, we show changes in RF and heating rates caused by volcanic eruptions. Figures 4a,b show a decrease in the global mean RF at the TOA by $0.34 \pm 0.17 \text{ W m}^{-2}$ and by $0.45 \pm 0.21 \text{ W m}^{-2}$ over the tropics, averaged over the years 2003–2013. There is a reduction in global mean surface radiative forcing by $0.37 \pm 0.23 \text{ W m}^{-2}$ and within the tropics by $0.47 \pm 0.26 \text{ W m}^{-2}$ (Supplementary Figure S1). The impacts are clearly visible in the tropical mean (Figure 4b) while it is subdued in the global mean (Figure 4a). The radiative forcing within the tropical region is more negative ($-0.43 \pm 0.20 \text{ W m}^{-2}$) than the global mean ($-0.32 \pm 0.16 \text{ W m}^{-2}$) since the majority of volcanoes are tropical (Table 2) and solar intensity is greater in the tropics. Individual eruption contributions including Rabaul (2006), Sarychev (2009), and Nabro (2011) are summarised in Supplementary Table S1. Past studies also illustrated a similar impact of volcanic eruptions on radiative forcing. For example, the El Chichón eruption in April 1982 caused a reduction in radiative forcing by $2\text{--}4 \text{ W m}^{-2}$ over a year within the region $30^\circ \text{ S} - 30^\circ \text{ N}$ (Robock and Mao, 1995), while the Pinatubo eruption in June 1991 reduced a TOA radiative forcing by 4.5 W m^{-2} within the region $40^\circ \text{ S} - 40^\circ \text{ N}$ (Minnis et al., 1993; Yang et al., 2019).

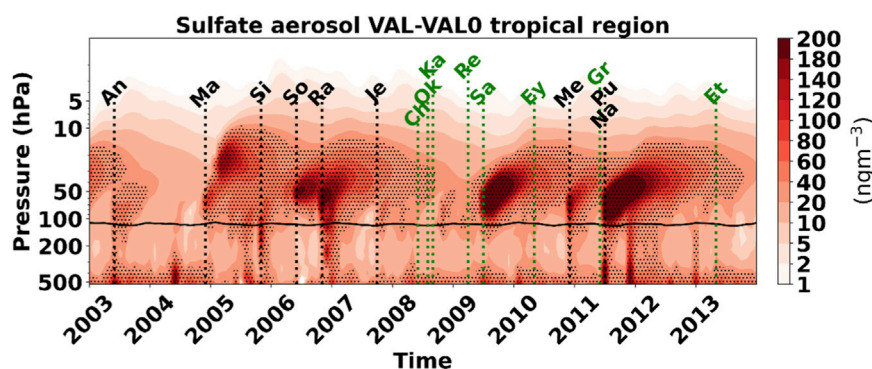


FIGURE 5
Time series of volcanic sulfate aerosol (ng m^{-3}). The black dots indicate significance at a 95% confidence level, obtained using a Student's t-test for the mean difference. Dotted vertical lines show the month of eruption. A thick black line in the figure indicates the tropopause.

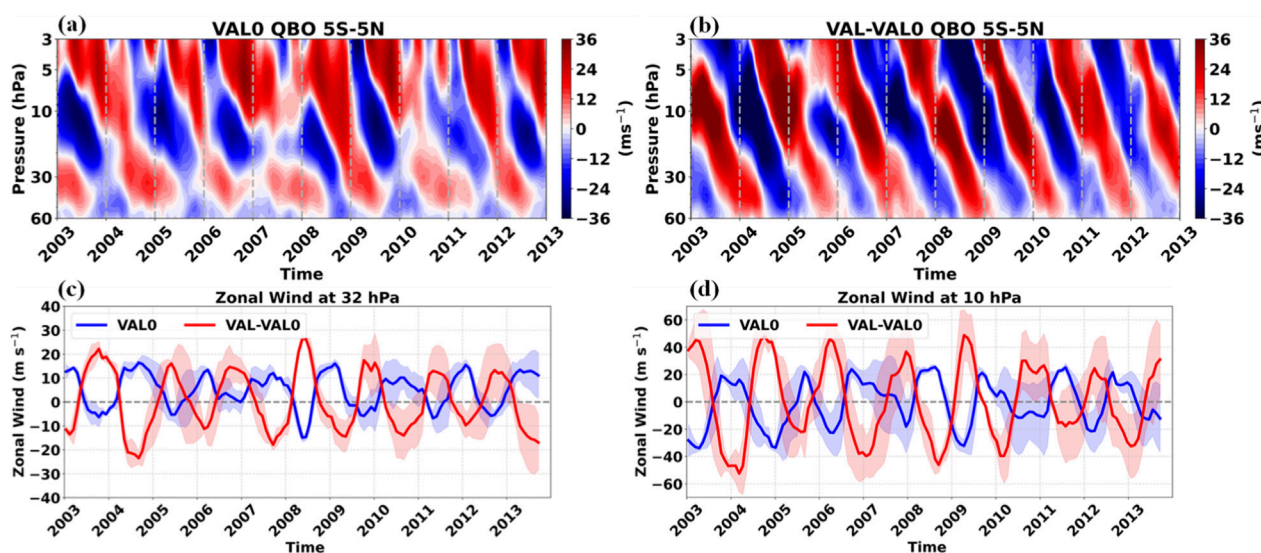


FIGURE 6
Time series of zonal wind (m s^{-1}). (a) zonal wind from VAL0 simulation and averaged over the equatorial region ($5^{\circ}\text{S} - 5^{\circ}\text{N}$), (b) same as (a) but averaged from VAL - VAL0 simulation, in Figs. a–b, red shading indicates the westerly phase, and blue shading indicates the easterly phase (c) zonal wind at 32 hPa, (d) same as (c) but for 10 hPa.

Figures 4a,b also show significant enhancements in stratospheric heating rates (locally) due to volcanic sulfate aerosols. Figure 4a shows an increase in the global mean heating rates by $0.01 \pm 0.009 \times 10^{-2} \text{ K d}^{-1}$, and within the tropics by $0.051 \pm 0.04 \times 10^{-2} \text{ K d}^{-1}$. Similar to SAOD, the heating rate peaks are more prominent in the tropical region than in the global mean, with the largest contributions from Rabaul (2006), Sarychev (2009), and Nabro (2011) (Figure 4b; Supplementary Table S1). However, the heating rate decreased after 2012 due to the reduced amount of stratospheric sulfate aerosols (Figure 3b). Due to the higher numbers of tropical volcanoes and SAOD during the study period, the heating rates within the tropical region ($0.051 \pm 0.04 \times 10^{-2}$) exceed the global average ($0.01 \pm 0.009 \times 10^{-2} \text{ K d}^{-1}$).

Further, we analyse the distribution of volcanic sulfate aerosols in the tropical stratosphere. Figure 5 illustrates the large enhancement of sulfate aerosols in the stratosphere following

volcanic eruptions. This figure shows less enhancement in sulfate aerosols ($23.32 \pm 2 \text{ ng m}^{-3}$) in the volcanically quiescent period from 2003–2004 in the VAL-VAL0 simulation. However, there is a significant enhancement by $59.60 \pm 11 \text{ ng m}^{-3}$ seen during 2005–2013 due to a series of moderate and large volcanoes. This figure suggests that volcanic sulfate aerosols persist for 2–3 years in the stratosphere (100–five hPa) after the eruption. The increase in stratospheric sulfate aerosol concentrations during the initial 2 months following the eruption in the tropics is presented in Supplementary Table S1. There is an enhancement in stratospheric sulfate aerosols in the tropics due to extratropical volcanoes. This may be due to transport from the extra-tropics to the tropics. It should be noted that aerosols are transported from the extratropics to the tropics and vice versa (Günther et al., 2018; Timmreck, 2012; Haywood et al., 2010; Robock, 2000). However, such analysis is out of the scope of the manuscript.

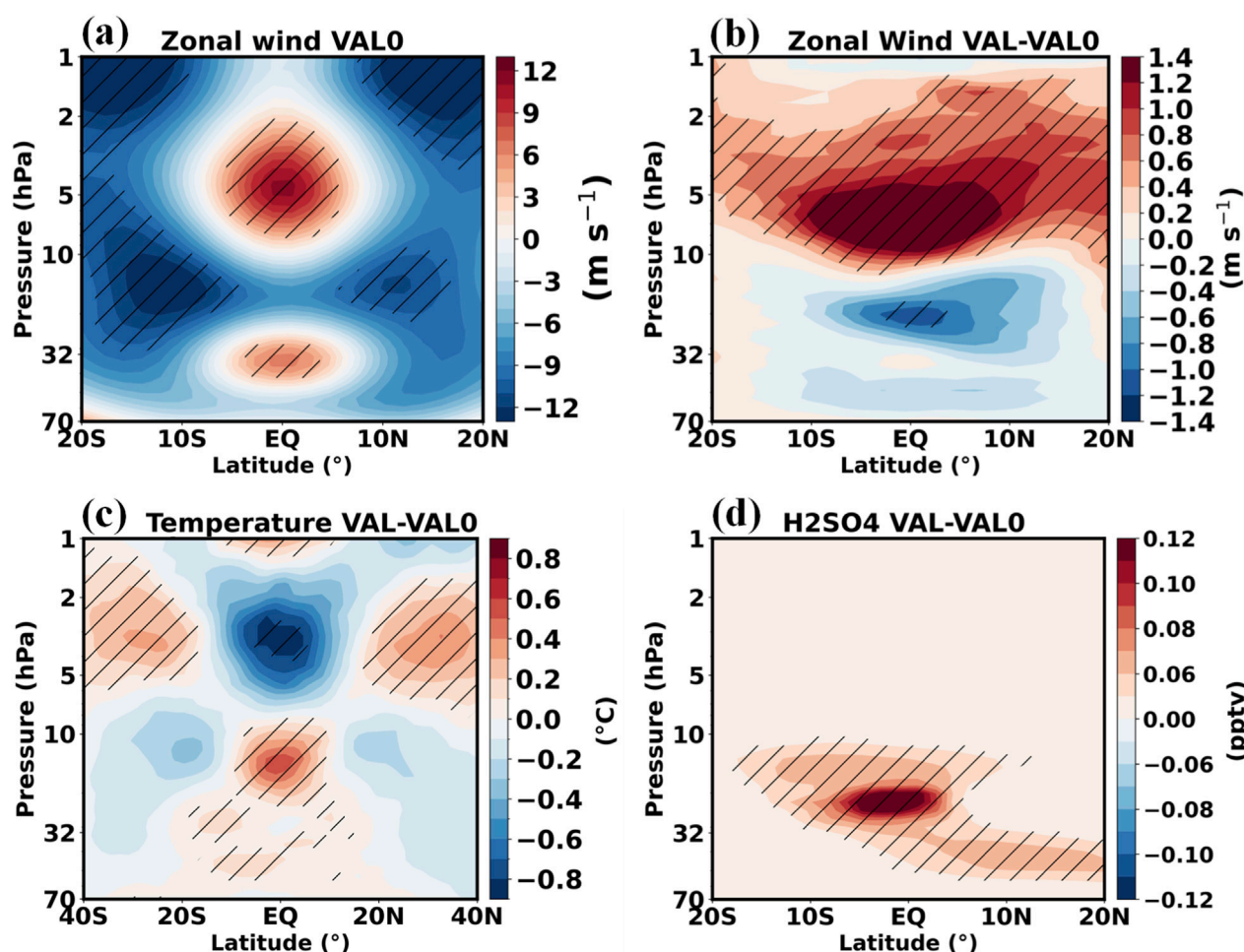


FIGURE 7

(a) Latitude pressure cross-section of the amplitude of QBO in zonal wind (m s^{-1}) averaged for the period 2003–2013 from VAL0 simulation, (b) same as (a) but for VAL-VAL0 simulation, (c) same as (b) but for temperature ($^{\circ}\text{C}$), (d) same as (b) but for H₂SO₄ (ppt). The black hatches indicate significance at a 95% confidence level, obtained using a Student's t-test for the mean difference.

3.3 Impact of volcanic sulfate aerosol on the QBO

In this section, we show the impact of volcanic sulfate aerosol on the QBO. Figure 6a shows the altitude variation of the QBO in zonal mean zonal winds for the $5^{\circ}\text{S} - 5^{\circ}\text{N}$ from the VAL0 simulation. It shows that the downward propagation of the easterly is faster than (~ 22 – 26 months cycle) the westerly (~ 24 – 36 months cycle). Volcanoes cause a reversal of the easterly phase to the westerly phase and vice versa, as seen from the anomalies (VAL-VAL0) plotted in Figures 6b,c,d. show the time series of zonal mean zonal wind at 32 hPa and 10 hPa, respectively. In both pressure levels, the VAL0 simulation shows alternating easterly and westerly winds, completing one oscillation ~ 24 – 30 months. In contrast, the anomalies (VAL-VAL0) are largely out of phase with VAL0 at both pressure levels. This suggests that the volcanic perturbation reverses the phase of the QBO at both pressure levels (see Figures 6c,d).

Previous studies have also reported that tropical volcanic eruptions can delay the progression of the QBO phases. For example, using the UM-UKCA aerosol-climate model, Brown et al. (2023) reported that tropical eruptions (e.g., Mt. Tambora

in 1815 and Mt. Pinatubo in 1991) delayed the progression of the QBO phases. Our study extends this by considering both large and moderate volcanic eruptions globally, rather than being limited to the tropical region, allowing us to investigate the broader impact of volcanic forcing on the QBO.

Furthermore, we show the latitude pressure cross-section of the zonal mean zonal wind for the VAL0 simulation (Figure 7a). The zonal mean zonal wind shows a double-peak structure near the equator ($10^{\circ}\text{S} - 10^{\circ}\text{N}$), with maximum centred at 32 hPa and five hPa and minimum at 20 hPa (Figure 7a). A similar double-peak structure in the zonal wind at the equator is reported by Fadnavis and Beig (2009). It is produced by the secondary meridional circulation induced by the QBO (Plumb and Bell, 1982; Punge et al., 2009).

When the volcanic forcing is turned on, it leads to a noticeable weakening of the zonal wind speed and increases temperature in the tropical stratosphere (32–10 hPa), while it increases zonal wind speed and decreases temperature in the upper stratosphere (five to one hPa) near the equator (Figures 7b,c). The mechanism for heating of the lower stratosphere by volcanic aerosol, which leads to QBO phase disruption, can be understood through changes in thermal-wind and residual circulation. Volcanic sulfate aerosols absorb near-infrared

and longwave radiation, producing anomalous heating of the tropical lower stratosphere (Figure 7c; Supplementary Figure S2). This generates a meridional temperature gradient between the heated equatorial region and cooler subtropics (Figure 7c). Through thermal wind balance, this temperature gradient induces an equatorial positive wind shear, resulting in easterly winds (70–20 hPa) at the equator (Figure 7b). Past studies report that sulfate aerosols injected via geoengineering lead to strong heating anomalies at the equator thereby perturbing the QBO phase (Niemeier et al., 2020).

Interestingly, our results show that volcanoes lead to increased H_2SO_4 concentrations in the tropical lower stratosphere, where there is a decrease in wind speed and increase in temperature (32–10 hPa) (Figure 7d). The elevated H_2SO_4 supports chemical pathways that contribute to ozone loss in the tropical lower stratosphere (Hazra et al., 2019). However, detailed investigation of ozone chemistry is beyond the scope of this manuscript.

4 Conclusion

ECHAM6-HAMMOZ chemistry-climate model simulations for volcanic eruptions during the period 2003–2013 were analysed using the three-member ensemble mean and compared with control simulations to study the impact of volcanic sulfate aerosols on radiative forcing and the QBO. Our findings are:

1. Our simulations show that volcanoes during the study period elevated tropical SAOD by 0.0012 ± 0.00095 , leading to a negative radiative forcing in the tropical region ($20^\circ\text{S} - 20^\circ\text{N}$) (TOA: $-0.45 \pm 0.21 \text{ W m}^{-2}$; surface: $-0.47 \pm 0.26 \text{ W m}^{-2}$). Comparison of our findings with other models shows slight differences in simulated SOAD and radiative forcing values. For example, Brühl et al. (2015) using the EMAC model reported an increase in SAOD by 0.001–0.01 in the tropics for tropical volcanic eruptions from 2002 to 2012, producing a decrease in radiative forcing by -0.1 to -0.26 W m^{-2} . Schmidt et al. (2018) using the CESM1-WACCM model estimated an enhancement in SAOD by 0.0076 during 2005–2014 that caused a reduction in global mean radiative forcing by -0.10 W m^{-2} . The differences between model results arise from variations in volcanic SO_2 inventory, injection methodology, and the aerosol parameterization, used in the model (Mills et al., 2016).
2. The volcanic eruptions increased the sulfate aerosol amount in the tropical lower stratosphere by $56.60 \pm 11 \text{ ng m}^{-3}$ during 2003–2013. These aerosols enter the tropical stratosphere and elevate the lower stratospheric heating by $0.05 \pm 0.02 \times 10^{-2} \text{ K d}^{-1}$. This heating contributes to the reversal of the easterly to westerly phases of the QBO and vice versa.
3. Volcanic sulfate aerosols lead to a strengthening of QBO wind speeds at $\sim 5 \text{ hPa}$ and a weakening at $\sim 20 \text{ hPa}$ near the equator, accompanied by increased temperatures and enhanced H_2SO_4 concentrations in the same region. The enhanced H_2SO_4 concentrations in the lower stratosphere may lead to ozone loss.

Our study points out that moderate and large volcanoes modulate the QBO. However, the results may vary from model to model. The uncertainties in a particular model arise from the use of emission inventories, transport processes, and various parametrizations

(Fadnavis et al., 2013; 2017; 2019). These uncertainties can affect the simulated SAOD, radiative forcing, heating rates, and QBO response. The QBO modulates tropical convection and cyclones, causing an impact on extreme weather (Fadnavis et al., 2011; Kroll et al., 2024). Hence, representation of volcanic aerosol forcing and its interaction with the QBO may improve climate projections.

Data availability statement

The datasets presented in this study can be found in online repositories. The names of the repository/repositories and accession number(s) can be found in the article/Supplementary Material.

Author contributions

PC: Validation, Writing – original draft, Formal Analysis, Conceptualization, Writing – review and editing, Methodology, Data curation, Software, Investigation, Visualization. SF: Methodology, Supervision, Conceptualization, Investigation, Writing – review and editing. AL: Writing – review and editing, Methodology, Supervision. J-PV: Supervision, Writing – review and editing, Methodology. ST: Writing – review and editing, Supervision, Methodology. CS: Writing – review and editing, Visualization, Methodology, Software. VS: Data curation, Validation, Visualization, Writing – review and editing. RM: Methodology, Writing – review and editing, Supervision.

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Conflict of interest

Author RM was employed by Forschungszentrum Jülich GmbH. The remaining author(s) declared that this work was conducted in the absence of any commercial or financial relationships that could be construed as a potential conflict of interest.

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Supplementary material

The Supplementary Material for this article can be found online at: <https://www.frontiersin.org/articles/10.3389/fenvs.2026.1831183/full#supplementary-material>

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