

1 **Future volcanic eruptions may delay the recovery of lower** 2 **stratospheric ozone over Antarctica and Southern Hemisphere mid-** 3 **latitudes**

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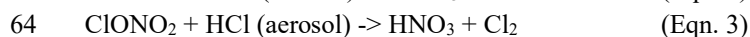
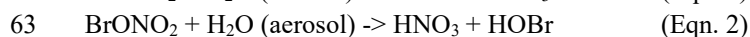
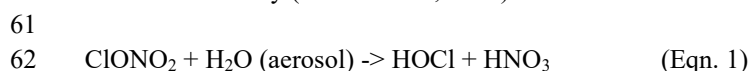
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17 **Abstract.** Sporadic explosive volcanic eruptions can inject large amounts of sulfur into the stratosphere, which forms volcanic
18 sulfate aerosols with the potential to affect stratospheric ozone chemistry. Future volcanic eruptions have been represented in
19 climate projection studies with varying degrees of realism despite their potential importance for polar ozone recovery. Climate
20 projections typically use a constant volcanic forcing based on a historical average, which very likely underestimates the
21 magnitude of future volcanic forcing and ignores the sporadic nature of volcanic eruptions. In this study, we use stochastic
22 volcanic eruption scenarios and a plume-aerosol-chemistry-climate model (UKESM-VPLUME) to assess the effect of future
23 volcanic sulfur injections on lower stratospheric ozone recovery over Antarctica and Southern Hemisphere mid-latitudes. We
24 find that sporadic eruptions can delay Antarctic total column ozone recovery by up to five years, though this delay is relatively
25 small when compared with the long-term ozone recovery timescale. Large-magnitude eruptions occurring before mid-century
26 can, however, episodically cause more substantial delays in the recovery. Based on a composite analysis we show that the
27 ozone response to volcanic sulfate aerosols over Antarctica and Southern Hemisphere mid-latitudes weakens over the 21st
28 century due to declining chlorofluorocarbon concentrations. Overall, our findings underscore the need for fully interactive
29 volcanic aerosol-chemistry coupling to assess the resilience of the Antarctic ozone layer in response to future volcanic
30 eruptions and other stratospheric perturbation events. Our results also support previous calls for sustained monitoring of
31 stratospheric composition and ozone-depleting processes to better anticipate and attribute changes in ozone recovery.

32 1 Introduction

33 The stratospheric ozone layer has been slowly recovering since about the year 2000, following the implementation of the
34 Montreal Protocol in 1987 and its subsequent amendments, which limited the production and consumption of ozone-depleting
35 substances (ODSs), including chlorofluorocarbons (CFCs) and bromine-containing halons. These ODSs are sufficiently long-
36 lived to reach the stratosphere where they undergo photolysis, releasing chlorine and bromine atoms that catalytically destroy
37 ozone. According to WMO/UNEP ozone assessments, polar ozone concentrations are projected to return to their 1980 levels
38 by 2066 (range: 2049-2077) in Antarctica and by 2045 (range: 2029-2051) in the Arctic (WMO, 2022). Although the Antarctic
39 ozone layer has shown a robust recovery trend since 2000, as observed and simulated in chemistry-climate models (WMO,
40 2022), recent years have witnessed large ozone holes related to elevated stratospheric aerosol loading from volcanic eruptions
41 and wildfires (Solomon et al., 2016; Yu et al., 2021; Solomon et al., 2023). For instance, the 2015 Calbuco eruption in Chile,
42 with a stratospheric injection of about 0.4 Tg of sulfur dioxide (SO₂), led to a record-large Antarctic ozone hole exceeding 25
43 million km² in the same year (Solomon et al., 2016; Ivy et al., 2017; Zhu et al., 2018). Model simulations by Stone et al. (2021)
44 suggest that volcanic sulfate aerosols from the 2015 Calbuco eruption reduced the total column ozone by 1 to 5% over the
45 entire Southern Hemisphere (SH) mid-latitudes region (up to 32 °S) from August to December. The unpredictability of volcanic
46 eruptions and other stratospheric perturbation events, such as wildfires, continue to complicate assessments of future ozone
47 recovery over Antarctica and SH mid-latitudes (Chipperfield and Bekki, 2024).

48
49 Stratospheric volcanic sulfate aerosols provide surfaces facilitating heterogeneous chemical reactions that catalyse the release
50 of reactive chlorine and bromine species from their respective reservoir species (Eqn. 1 to 3). At present, volcanic eruptions
51 that inject sulfur into the stratosphere result in a net decrease in Antarctic column ozone. This is because volcanic sulfate
52 aerosols enhance ozone loss via the HOx, ClOx, and BrOx catalytic cycles, which dominate over the suppression of ozone loss
53 driven by the NOx cycle (Eqn. 4). Observational and modelling studies have provided evidence of the reduction in NOx and
54 enhancement in halogen radicals at mid-latitudes after the 1991 Mt. Pinatubo eruption (Fahey et al., 1993). The ozone response
55 to volcanic sulfate aerosols is greater over Antarctica than other latitudes due to the extreme cold temperatures and the presence
56 of polar stratospheric clouds inside the Antarctic polar vortex. As anthropogenic ODSs continue to decline in this century and
57 assuming no injection of volcanic halogen or unexpected rise in CFC emissions, future volcanic eruptions with the same
58 stratospheric SO₂ injection are expected to cause less ozone loss via the ClOx and BrOx cycles. Therefore, future enhancements
59 in stratospheric aerosol loading are anticipated to lead to a net increase in Antarctic column ozone towards the middle or the
60 end of this century (Klobas et al., 2017).



66
67 Volcanic eruptions may also inject water vapour and volcanic halogen species into the stratosphere in addition to volcanic SO₂
68 and cause additional chemical ozone loss (Bobrowski et al., 2003; Pyle and Mather, 2009; Evan et al., 2023; Santee et al.,
69 2024). Recent modelling studies demonstrate that the co-injection of volcanic sulfur and halogens into the stratosphere can
70 lead to greater and prolonged ozone depletion compared to sulfur injections only (Klobas et al., 2017; Brenna et al., 2020;
71 Ming et al., 2020; Staunton-Sykes et al., 2021). Large amounts of volcanic water vapour injection, as demonstrated by the 2022
72 Hunga Tonga-Hunga Ha'apai eruption, can also perturb stratospheric ozone for 4-7 years (Zhu et al., 2022; Fleming et al.,
73 2024; Zhou et al., 2024; Zhuo et al., 2025). While volcanic halogen and water vapor emissions are important for ozone recovery
74 projections, strong stratospheric water vapor and halogen injections are rare and highly variable. Volcanic SO₂ is the most

75 commonly emitted species with comprehensive ice-core and satellite emission inventories (Carn et al., 2016; Sigl et al., 2021),
76 whereas volcanic halogen and water vapor injections lack comprehensive records, particularly prior to the satellite era. We
77 therefore focus only on SO₂ emissions in this study. Apart from volcanic halogen and water vapour emissions, very short-lived
78 (VSL) chlorine and bromine compounds, which have lifetimes of less than 6 months, are also important sources of stratospheric
79 bromine and chlorine (Sturges et al., 2000; Dorf et al., 2008; Laube et al., 2008; Sala et al., 2014; Wales et al., 2018), which
80 can lead to uncertainties in future ozone changes (Klobas et al., 2017; Villamayor et al., 2023).

81
82 Despite the importance of future volcanic sulfate aerosols enhancements for ozone recovery, few studies have investigated the
83 effects of future volcanic eruptions on stratospheric ozone over the polar and mid-latitude regions due to the unpredictability
84 of future eruptions. The first phase of the Chemistry Climate Model Initiative (CCMI-1) included volcanic aerosols in historical
85 simulations from 1850 to 2014 but assumed zero volcanic forcing in future projections from 2015 to 2100 (Dhomse et al.,
86 2018; Eyring et al., 2013). In CCMI-2, the model experiments adopted the Coupled Model Intercomparison Project (CMIP6)
87 protocol, using a prescribed constant volcanic forcing equivalent to the 1850-2014 mean volcanic forcing in future projections
88 from 2015 to 2100 (O'Neill et al., 2016). Naik et al. (2017) adopted a similar approach, prescribing a constant volcanic forcing
89 equivalent to the 1860-1999 mean to examine the volcanic effects on stratospheric ozone recovery. They concluded that
90 stratospheric volcanic sulfate aerosols enhancements in two future RCP scenarios had no impact on the polar ozone return
91 dates (Naik et al., 2017), but their model did not reproduce the chemical perturbations documented by Fahey et al. (1993)
92 following 1991 Mt. Pinatubo. Further, the use of a prescribed and constant volcanic forcing does not represent the sporadic
93 nature of volcanic eruptions. A recent study by Chim et al. (2023), using a stochastic volcanic forcing approach in a model
94 with interactive sulfur chemistry and volcanic aerosols, showed that future volcanic forcing from 2015-2100 is very likely to
95 exceed the 1850-2014 mean prescribed volcanic forcing used in CMIP6, with the median future forcing expected to be about
96 twice that of the 1850-2014 mean. This finding raises questions about the degree to which a higher and sporadic future volcanic
97 forcing would affect Antarctic ozone recovery.

98
99 In this study, we employ a plume-aerosol-chemistry-climate model with a stochastic volcanic forcing approach to simulate the
100 effects of future volcanic sulfur injections on Antarctica and SH mid-latitude ozone recovery. Our stochastic future eruption
101 scenarios resemble the statistical distribution of eruption frequency, latitude and sulfur mass in ice cores and satellite records.
102 Based on multiple ozone recovery indicators we evaluate the effects of future volcanic eruptions on stratospheric ozone. Our
103 model using stochastic scenarios indicates that future eruptions lead to delay in Antarctic ozone recovery assessed via changes
104 in total column ozone, ozone mass deficit, and ozone hole area. We also conduct a composite analysis to show that the ozone
105 response to volcanic sulfate aerosols over Antarctica and SH mid-latitudes is expected to decrease over this century. Given the
106 inherent unpredictability of volcanic eruptions, the stochastic volcanic forcing approach offers insights into the uncertainties
107 associated with future volcanic eruptions and their effects on the recovery of the Antarctic ozone layer.

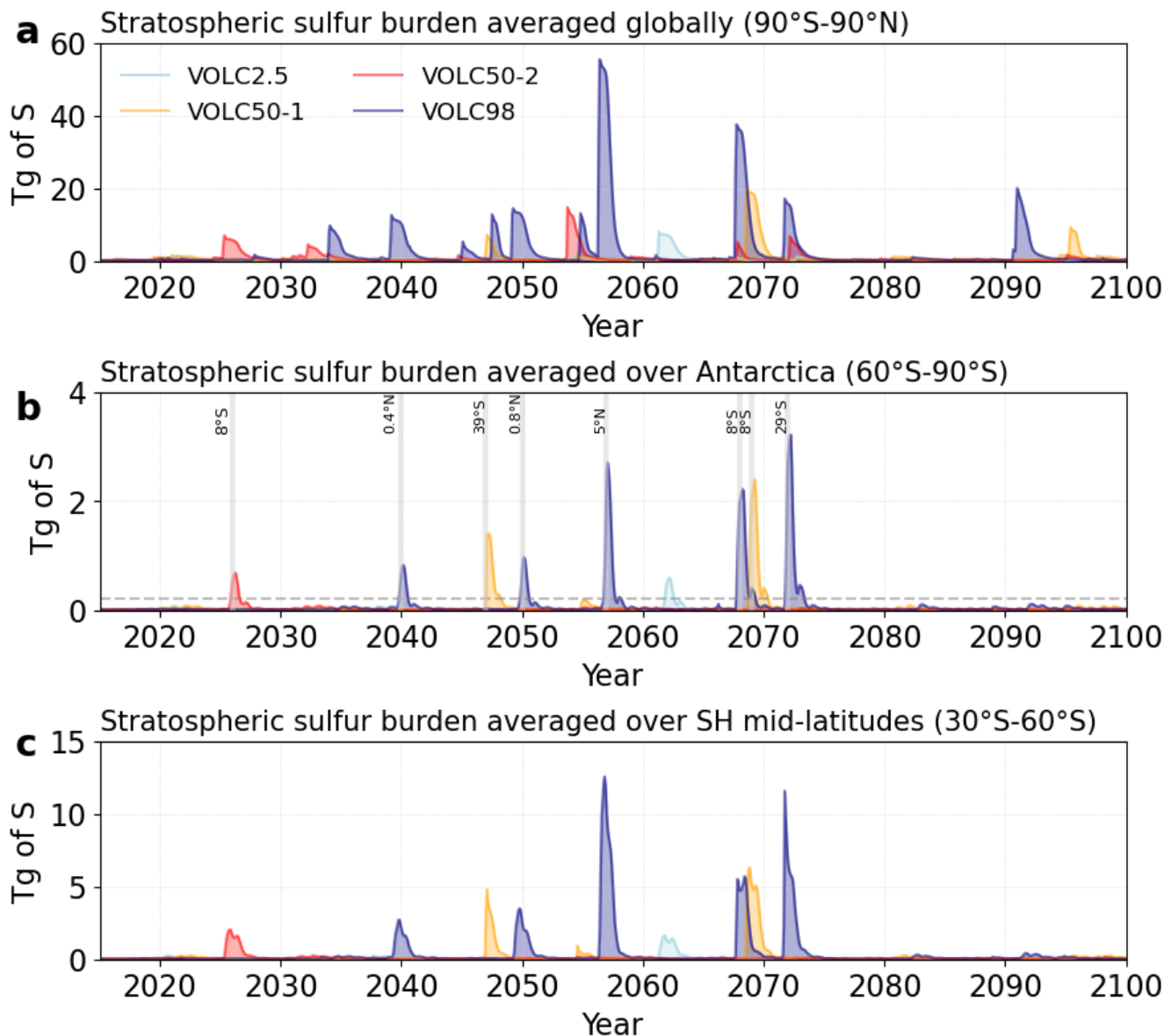
108 **2 Methods**

109 **2.1 Model Setup**

110 We use the UKESM simulations from Chim et al. (2023) and briefly summarise their design here. We first generate 1000
111 stochastic future eruption scenarios from 2015 to 2100 by resampling eruptions recorded in bipolar ice cores and satellite
112 observations over the past 11,500 years (Sigl et al., 2022; Carn, 2022). Our stochastic scenarios statistically resemble the
113 eruption magnitude (in terms of volcanic SO₂ mass), eruption frequency, and eruption latitude of past volcanic eruptions.
114 Based on the methodology developed in previous studies (Aubry et al., 2016; Bethke et al., 2017), we resample large-
115 magnitude eruptions (defined as > 3 Tg of SO₂ mass injection) using volcanic emission datasets from both ice cores and

116 satellite observations, and small-magnitude eruptions (defined as < 3 Tg of SO_2 mass injection) using satellite volcanic
117 emission records only. We do not resample the emissions of volcanic halogen species and water vapour due to the lack of a
118 comprehensive record of such species in the current ice-cores and satellite datasets. Consequently, we consider only the
119 stratospheric volcanic SO_2 mass in the resampling method.

120 We consider four scenarios as input for our future climate simulations: the low-end scenario (VOLC2.5, with 0.64 Tg of SO_2
121 yr^{-1}), the two median scenarios (VOLC50-1 and VOLC50-2, with 1.44 Tg of SO_2 yr^{-1}) and the high-end scenario (VOLC98,
122 with 5.23 Tg of SO_2 yr^{-1}), each correspond to a scenario sampled near to the 2.5th, 50th, and 97.5th percentiles, respectively,
123 of the ranked total sulfur mass of the 1000 stochastic scenarios. We perform simulations of the two median scenarios with
124 small-magnitude eruptions only (VOLC50-1s and VOLC50-2s, with 0.64 and 0.57 Tg of SO_2 yr^{-1}) to evaluate the effects of
125 small-magnitude eruptions. The results of the VOLC runs are compared against control simulations with constant volcanic
126 forcing as in CMIP6 ScenarioMIP (CONST, with 0.77 to 0.86 Tg of SO_2 yr^{-1} according to current volcanic emission records)
127 and without explosive volcanic emissions (NOVOLC). All runs used the SSP3-7.0 scenario for anthropogenic forcings with
128 simulation years between 2015 and 2100, with three ensemble members each. Fig. 1 shows the stratospheric sulfur burden
129 globally, over the Antarctic, and SH mid-latitudes for the four selected stochastic scenarios (VOLC2.5, VOLC50-1, VOLC50-
130 2, VOLC98).



131

132 **Figure 1.** Stratospheric sulfur burden (in Tg of S) averaged (a) globally, (b) over Antarctica, and (c) over SH mid-latitudes from
 133 2015 to 2100 for the four stochastic scenarios. The dotted line in panel (b) shows the stratospheric sulfur burden threshold of 0.2 Tg
 134 of S, which is used to select eruptions for composite analysis. The injection latitudes of the selected eruptions are shown in panel (b).
 135

136 We employ the UKESM-VPLUME framework, which integrates the fully-coupled UK Earth System Model version 1.1 that
 137 includes ocean-atmosphere interactions (UKESM1.1; Mulcahy et al., 2023) with a one-dimensional eruptive plume model
 138 (Plumeria; Mastin, 2007, 2014). This framework allows for interactive stratospheric sulfur chemistry and volcanic sulfate
 139 aerosols with plume injection heights consistent with background climate conditions, enabling the simulation of stratospheric
 140 aerosol life cycle consistent with simulated future climate conditions. We evaluate the performance of UKESM1.1 in assessing

141 polar ozone concentrations by comparing the UKESM historical simulations with satellite observations (see Sect. 3.1). To
142 evaluate the volcanic-induced ozone response in UKESM1.1, we perform a five-member ensemble of UKESM1.1 historical
143 simulations from 1991 to 1993 to assess the model-simulated ozone response after the 1991 Mt. Pinatubo eruption. We inject
144 10 Tg of SO₂ across 13 latitude bands between 0° and 15°N to allow the aerosols to be distributed in both hemispheres. This
145 injection approach is consistent with previous modelling studies using UM-UKCA and CESM (Dhomse et al., 2014; Mills et
146 al., 2016). We also perform a five-member ensemble without volcanic emissions as the control simulations.

147

148 We perform future climate simulations under a high-end future shared socio-economic pathway (SSP3-7.0) from 2015 to 2100,
149 with three ensemble members for each stochastic scenario. The emission projections in our simulations follow the ScenarioMIP
150 experiment under SSP3-7.0, which do not include the emission of VSL chlorine and bromine compounds.

151 **2.2 Ozone recovery indicators**

152 We assess the effects of future eruptions on Antarctic ozone recovery by evaluating three ozone recovery indicators over the
153 Antarctica (60°S to 90°S) and SH mid-latitudes (30°S to 60°S), including (1) total column ozone, (2) ozone mass deficit, and
154 (3) ozone hole area. Total column ozone and ozone mass deficit have been used as ozone recovery indicators in previous
155 studies to assess the recovery trend of ozone in the historical and future periods (Dhomse et al., 2018; de Laat et al., 2017;
156 Stone et al., 2021). We assess the impact of volcanic forcing on Antarctic ozone return dates by comparing results of the VOLC
157 runs versus the NOVOLC control run.

158 Total column ozone quantifies the ozone mass within the entire atmospheric column, measured in Dobson units (DU). The
159 total column ozone recovery back to historical baseline (i.e., 1980 conditions) is a metric used by the UNEP/WMO Ozone
160 Assessment Report (WMO, 2022). We use the 1978-1982 mean October total column ozone over Antarctica from UKESM1.1
161 historical simulations (hereafter referred to as “1980 baseline”) to evaluate the return year of total column ozone concentrations
162 relative to pre-industrial levels. We define the return year of the total ozone column as the last year of having an October mean
163 total column ozone averaged over Antarctica that is lower than the 1980 baseline total column ozone value. We then assess
164 the delay in total column ozone recovery by comparing the return dates of VOLC runs with the control simulation NOVOLC.

165 Ozone mass deficit represents the deviation of the observed or model-simulated ozone mass from a reference value of the total
166 ozone column over Antarctica. The value of 220 DU serves as a reference point frequently used to identify the depletion of
167 Antarctic stratospheric ozone, as values less than 220 DU were not observed before 1979. In this study, we define two metrics
168 for ozone mass deficit, one with a reference value of 220 DU and the second with a reference value of 175 DU. We define the
169 return year of ozone mass deficit as the last year of having an October mean ozone mass deficit averaged over Antarctica that
170 is lower than the 220 DU and 175 DU thresholds.

171 To quantify the uncertainty in return dates arising from interannual variability, we apply a Monte Carlo analysis to the ensemble
172 mean of the October total column ozone and ozone mass deficit time series. Each series is first decomposed into a smoothed
173 long-term trend and residuals. We then generate 1,000 realizations by resampling the residual variability and superimposing it
174 on the trend. This approach yields a probabilistic distribution of return dates for crossing a specified ozone threshold, from
175 which we report the median return year as well as the 5th to 95th percentile range.

176 Ozone hole area represents the spatial extent over the Antarctic where the total ozone column falls below a reference value.
177 To define the presence or absence of an ozone hole in any given year, we use thresholds of total column ozone less than 220
178 DU, 175 DU and 150 DU, respectively. We assess recovery trends of the October Antarctic ozone hole area by performing
179 linear regression to evaluate the return dates, and compare against the control simulation NOVOLC.

180 **2.3 Composite analysis of ozone chemical loss**

181 We evaluate the chemical loss of Antarctic ozone using the stratospheric diagnostics developed by Lee et al. (2002). This
182 method estimates the rate of odd oxygen ($Ox = O_3 + O(^3P) + O(^1D)$) destruction for each catalytic cycle by determining the
183 reaction rates of the rate-limiting steps. Under the assumption that $[Ox] \approx [O_3]$, the rate of odd oxygen loss is approximately
184 equal to the rate of ozone loss for each catalytic cycle. We calculate the cumulative stratospheric ozone loss over Antarctica
185 for the halogen cycles (i.e., the sum of ClOx and BrOx cycles), NOx cycle, and HOx cycle by integrating the monthly-mean
186 ozone loss rate over time (from October to March) and altitude (up to 25 km). We calculate the cumulative stratospheric ozone
187 loss over SH mid-latitudes by integrating the monthly-mean ozone loss rate up to 25 km for 12 months since the eruption
188 month.

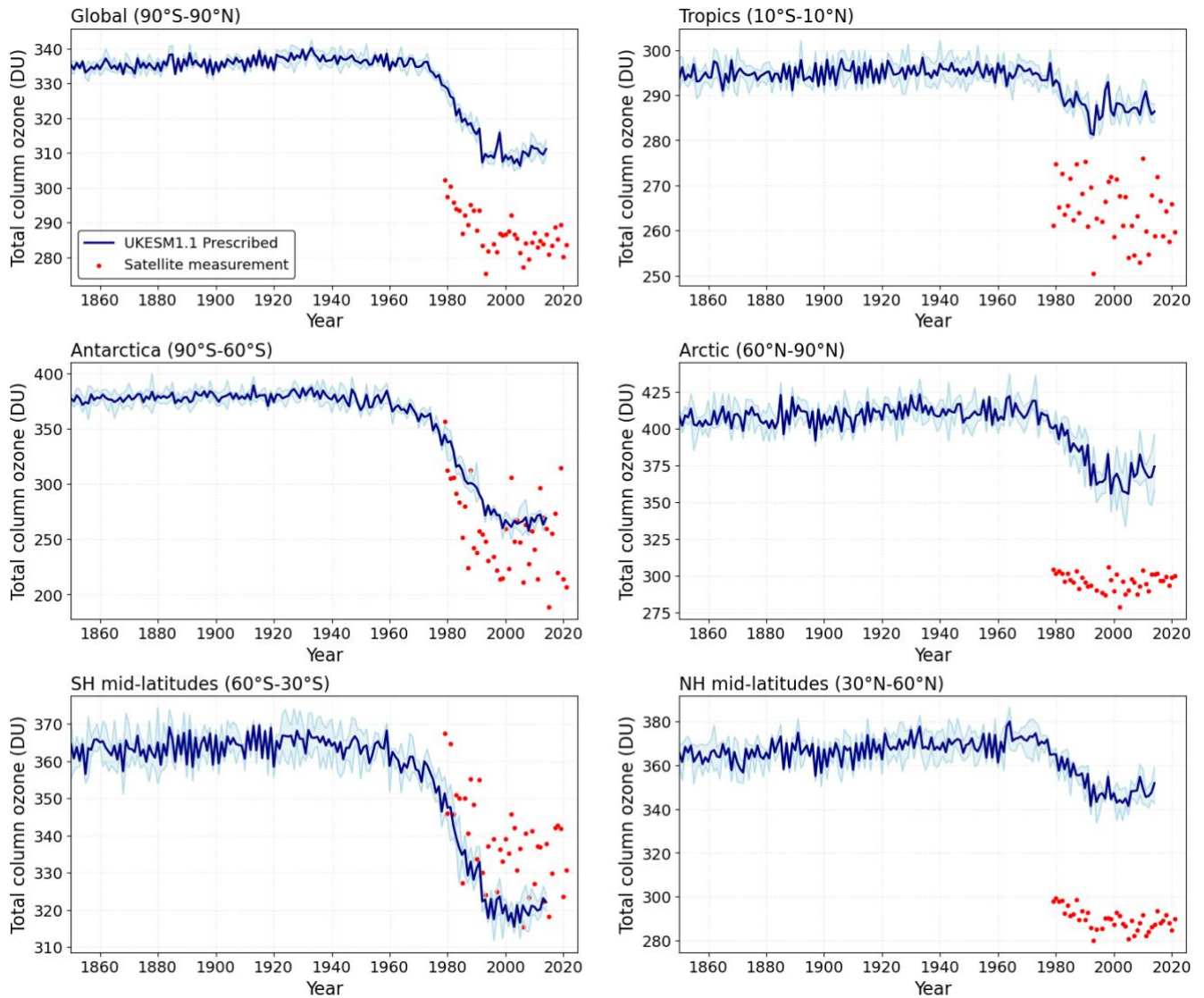
189 To evaluate the volcanic-induced ozone loss via catalytic cycles, we perform a composite analysis of the cumulative
190 stratospheric ozone loss for selected eruptions. We identify 9 large-magnitude eruptions with stratospheric sulfur burden peaks
191 greater than 0.2 Tg of S over the Antarctic (Fig. 2b) for performing composite analysis to evaluate the chemical changes via
192 each catalytic cycle.

193 To assess the volcanic effects on Antarctic stratospheric ozone, we evaluate the relative differences in the chemical loss
194 between the VOLC and NOVOLC runs. We calculate the cumulative stratospheric ozone loss anomaly relative to the control
195 run (NOVOLC) for each catalytic cycle over Antarctica and SH mid-latitudes, using a 2-year window prior to the eruption as
196 reference to compare with the 5-year post-eruption period. The peak ozone response (i.e., at year 0 since the eruption) for each
197 selected eruption is normalised with the respective total sulfur mass and plotted against the eruption year for comparison.

198 **3 Results**

199 **3.1 Model Evaluation**

200 We assess the performance of UKESM in simulating ozone concentrations by comparing the total column ozone simulated by
201 UKESM1.1 during the historical period (1850-2014) with satellite observations (Fig. 3). UKESM1.1 is the latest version of
202 UKESM, incorporating several improvements compared to the previous model version, which reduced the cold bias in the
203 historical global mean surface temperature by 50% (Mulcahy et al., 2023).



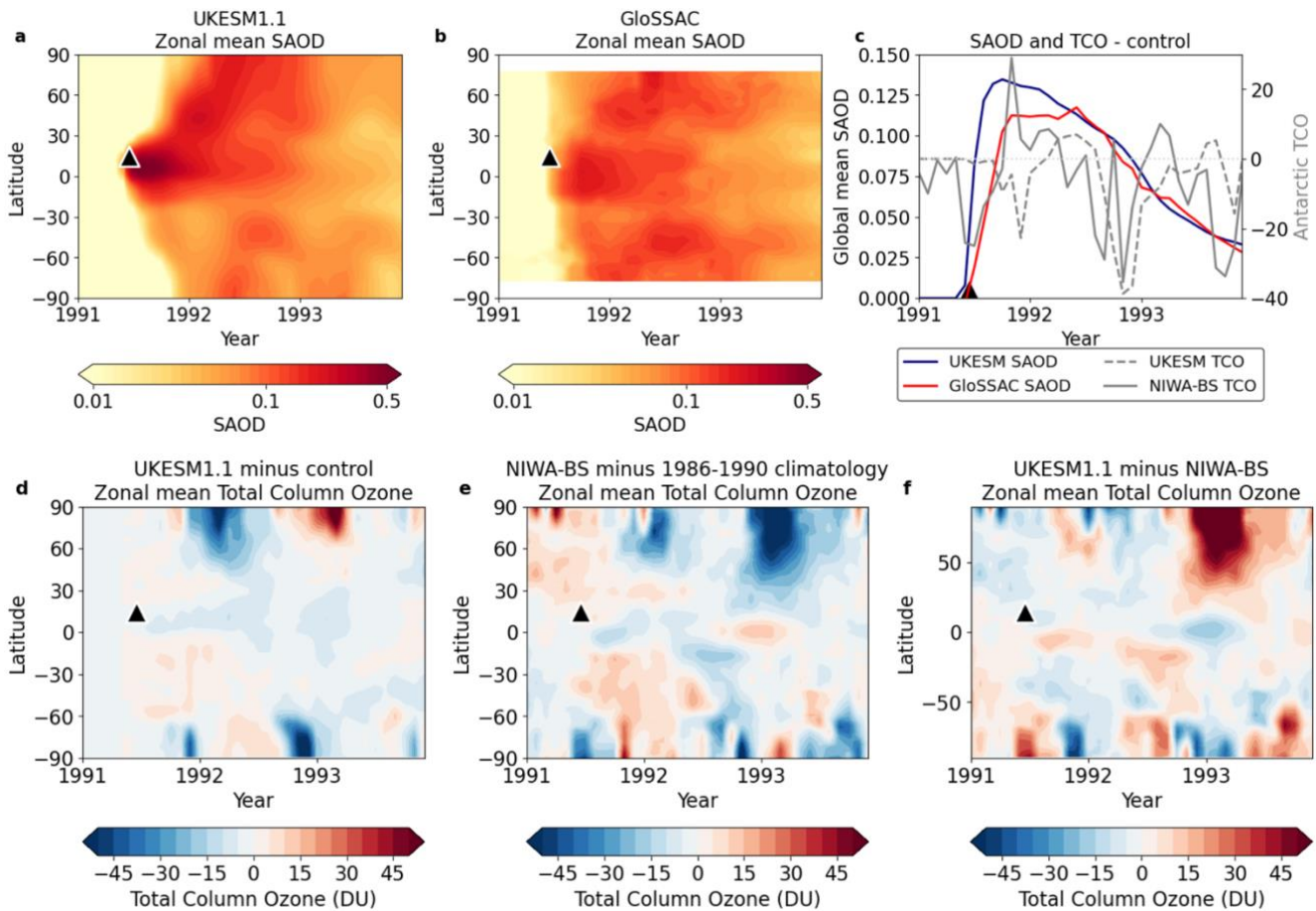
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205 **Figure 2.** Annual mean total column ozone (in DU) averaged (a) globally (90°S to 90°N), over the (b) tropics (20°S to 20°N), (c)
 206 Antarctica (60°S to 90°S), (d) Arctic (60°N to 90°N), (e) SH mid-latitudes (30°S to 60°S), and (f) NH mid-latitudes (30°N to 60°N). The
 207 blue lines represent the UKESM historical simulations from 1850 to 2014, and the red markers represent satellite measurements
 208 from 1978 to 2023. The blue shading shows the maximum and minimum values of the ensemble members.

209 Fig. 2 shows the total column ozone over different latitude bands simulated by UKESM1.1 with prescribed volcanic forcing in the historical period from 1850 to 2014, along with satellite measurements from 1978 to 2023 obtained from the NASA
 210 Ozone Watch (NASA Ozone Watch, 2023). UKESM1.1 tends to overestimate the global, tropical and NH mid- and high-
 211 latitudes total column ozone as compared to satellite measurements. The total column ozone over SH mid-latitudes and
 212 Antarctica shows good correspondence with satellite measurements. Nonetheless, UKESM has a high bias (+20%) in
 213 stratospheric ozone compared to other CMIP6 models (Keeble et al., 2021), which stems from the bias over the tropics and
 214

215 Northern Hemisphere (NH). Fig. S1 shows the comparison of the zonal mean ozone mass mixing ratio averaged from 2000 to
216 2014 for UKESM1.1, ML-TOMCAT merged 4-D ozone dataset (Dhomse et al., 2021; Dhomse and Chipperfield, 2023), and
217 the CMIP6 multi-model mean (Keeble et al., 2021). We find that the zonal mean ozone distribution in UKESM1.1 in general
218 agrees well with the ML-TOMCAT dataset. UKESM1.1 simulates higher ozone mass mixing ratio over the tropics and
219 extratropical regions, lower ozone over Antarctic stratosphere from 1-10hPa, and higher ozone in the Antarctic lower
220 stratosphere between 10-30 hPa compared to both ML-TOMCAT and the CMIP6 multi-model mean (Fig. S1d and e). This
221 suggests that UKESM1.1 may overestimate lower stratospheric (10-30 hPa) ozone concentration over Antarctica relative to
222 other CMIP6 models and ML-TOMCAT. UKESM1.1 also simulates higher total column ozone between 30°S and 60°N, and
223 a deeper, more persistent Antarctic ozone hole compared to the CMIP6 multi-model mean (Fig. S2).

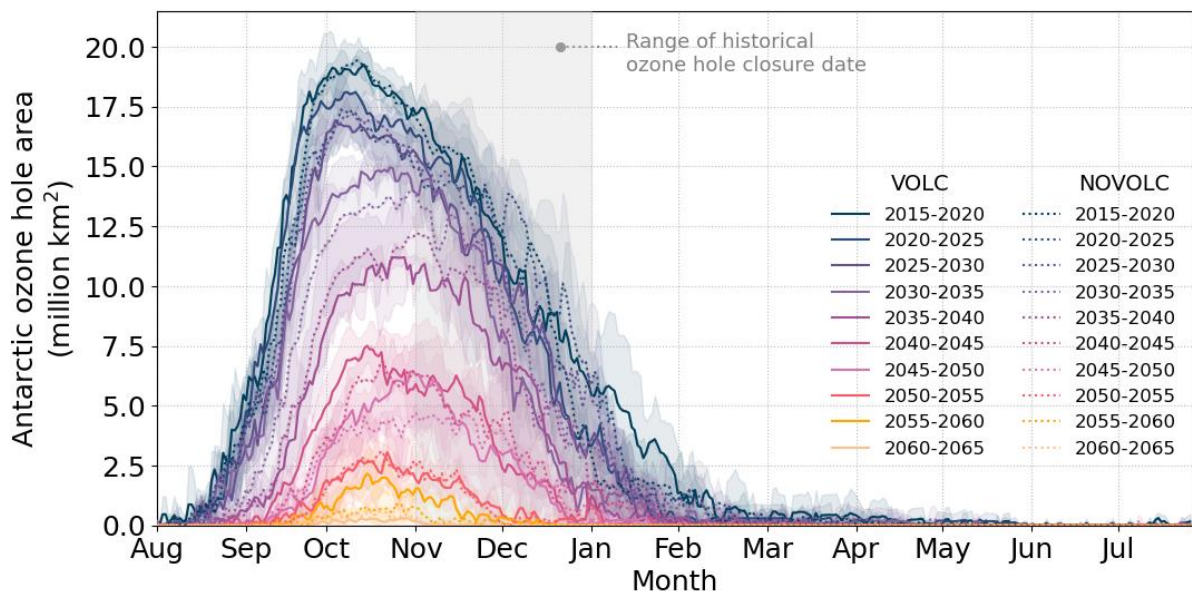
224 To assess the ozone response after volcanic eruptions in UKESM1.1, we performed a five-member ensemble of UKESM1.1
225 historical simulations for a case study of the 1991 Mt. Pinatubo eruption. We compared the UKESM1.1 model-simulated
226 stratospheric aerosol optical depth (SAOD) and total column ozone with the Global Space-based Stratospheric Aerosol
227 Climatology (GloSSAC) v2.2 dataset (Kovilakam et al., 2020) and the National Institute of Water and Atmospheric Research
228 - Bodeker Scientific (NIWA-BS) filled total column ozone dataset v3.5.2 (Bodeker et al., 2021) to evaluate the model
229 performance (Fig. 4). Our results show that UKESM1.1 simulates a higher SAOD over the tropics and NH (Fig. 4a and 4b),
230 and the peak global mean SAOD value is 22% higher than that of GloSSAC (Fig. 4c). The ensemble-mean of the UKESM1.1-
231 simulated total column ozone is able to capture total column ozone loss over Antarctic summer in 1991 and 1992, with a
232 magnitude comparable to NIWA-BS total column ozone loss (Fig 4d to 4f). Although our simulations capture the magnitude
233 of the Antarctic total column ozone response, the timing of ozone loss in 1991 does not match NIWA-BS. This discrepancy
234 likely reflects differences between our free-running ensemble climatology and the 1991 atmospheric conditions, as our
235 Pinatubo simulations are not nudged to observations. However, UKESM1.1 simulates a prolonged Antarctic total column
236 ozone loss until spring when compared to NIWA-BS dataset (Fig. 4f).



237

238 **Figure 3. Stratospheric aerosol optical depth (SAOD) and total column ozone responses following the 1991 Mt. Pinatubo eruption.**
 239 **(a,b) Zonal monthly mean SAOD from the UKESM1.1 simulation and GloSSAC. (c) Time series of global monthly mean SAOD**
 240 **anomalies (UKESM1.1 relative to control and GloSSAC relative to 1986-1990 climatological mean) and Antarctic monthly mean**
 241 **total column ozone anomalies (UKESM1.1 relative to control and NIWA-BS relative to 1986-1990 climatological mean).** **(d,e) Zonal**
 242 **monthly mean total column ozone anomalies from UKESM1.1 (relative to control) and NIWA-BS (relative to 1986-1990**
 243 **climatological mean). (f) Difference in total column ozone anomalies between UKESM1.1 and NIWA-BS (panel d minus e).**

244 Fig. 4 presents our model-simulated daily size of the Antarctic ozone hole averaged over a 5-year period from 2015 to 2065
 245 for one of our stochastic scenarios (VOLC50-1). Satellite observations from 1979 to 2022 suggest that the closure date of the
 246 Antarctic ozone hole typically occurs between mid-November and late December (Copernicus Atmosphere Monitoring Service
 247 (CAMS), 2023). However, the timing of the Antarctic ozone hole in our model results deviate significantly from historical
 248 observations. Our model simulations show a considerably prolonged duration of the Antarctic ozone hole extending well into
 249 February during the earlier decades of this century (Fig. 4). This discrepancy is attributed to the stratospheric cold bias of
 250 UKESM over Antarctica, which is associated with a strong polar night jet and the strong downdraught of mesosphere air during
 251 polar winter (Sellar et al., 2019), resulting in a stronger and more persistent polar vortex over Antarctica.



252

253 **Figure 4.** Daily size of the ozone hole area (5-year averaged) over Antarctica (60°S to 90°S, in million km²) for VOLC50-1 (solid lines)
 254 and NOVOLC (dotted lines). The colour shading shows the range of the maximum and minimum UKESM ensemble members. The
 255 grey shading shows the range of historical ozone hole (defined as area < 220 DU) closure date from satellite observations over 1979-
 256 2022 (CAMS, 2023).

257 Despite these limitations, we show that UKESM reproduces reasonably well the historical evolution of the October total ozone
 258 column (see Fig. 1c), when the ozone hole is at its deepest. This provides confidence in the model's ability to assess Antarctic
 259 ozone recovery by assessing the October-mean values of the ozone recovery indicators. The magnitude of Antarctic total
 260 column ozone response in our Pinatubo simulations also show a good agreement with observational datasets, suggesting that
 261 our estimates of relative impacts across different volcanic scenarios are robust. Since the stratospheric cold bias and strong
 262 winter polar vortex in UKESM (Sellar et al., 2019; Hall et al., 2021) likely hinder the transport of volcanic aerosols into the
 263 Antarctic stratosphere, we assess the ozone responses over both Antarctica and SH mid-latitudes.

264 3.2 Volcanic effects on Antarctic ozone recovery indicators

265 We use three ozone recovery indicators to assess the volcanic effects on Antarctic ozone return years, as summarised in Table
 266 1. We find that the return year of ozone mass deficit to 220 DU threshold is delayed by 0 to 2 years for the low-end and median
 267 future scenarios (VOLC2.5 and VOLC50, Table 1), and 5 years for a high-end future eruption scenario (VOLC98, Table 1).
 268 Although some stochastic scenarios indicate a later median return year, the overlap in uncertainty ranges suggests that the
 269 delay is not distinguishable from internal variability (Table 1). On the other hand, the return year of ozone mass deficit to the
 270 175 DU threshold ranges between 0 to 5 years across the stochastic scenarios.

271

272 **Table 1.** Return years of the ozone recovery indicators for Antarctic October-mean values, including total column ozone (median
 273 return year relative to the model 1980 baseline), the Antarctic October-mean ozone mass deficit (median return year relative to the
 274 220DU and 175DU thresholds), and the Antarctic October-mean ozone hole area (220 DU threshold). Numbers in brackets indicate
 275 the 5th to 95th percentile uncertainty range from the Monte Carlo uncertainty analysis. For the ozone hole area recovery trend, the
 276 uncertainty range corresponds to the spread across the three ensemble members.

Indicator	NOVOLC	VOLC2.5	VOLC50-1	VOLC50-1s	VOLC50-2	VOLC50-2s	VOLC98
Total ozone column	2066 (2062-2072)	2065 (2063-2074)	2064 (2058-2072)	2068 (2059-2075)	2069 (2062-2075)	2069 (2062-2077)	2071 (2066-2074)
Ozone mass deficit (220 DU)	2062 (2054-2065)	2064 (2059-2066)	2060 (2050-2069)	2060 (2052-2062)	2062 (2051-2064)	2062 (2053-2064)	2068 (2047-2073)
Ozone mass deficit (175 DU)	2038 (2022-2041)	2041 (2024-2044)	2038 (2022-2042)	2039 (2023-2043)	2044 (2036-2046)	2039 (2030-2041)	2041 (2030-2044)
Ozone hole area (220 DU)	2058 (2058-2059)	2058 (2056-2060)	2059 (2058-2060)	2059 (2059-2060)	2059 (2058-2060)	2059 (2058-2059)	2061 (2060-2062)

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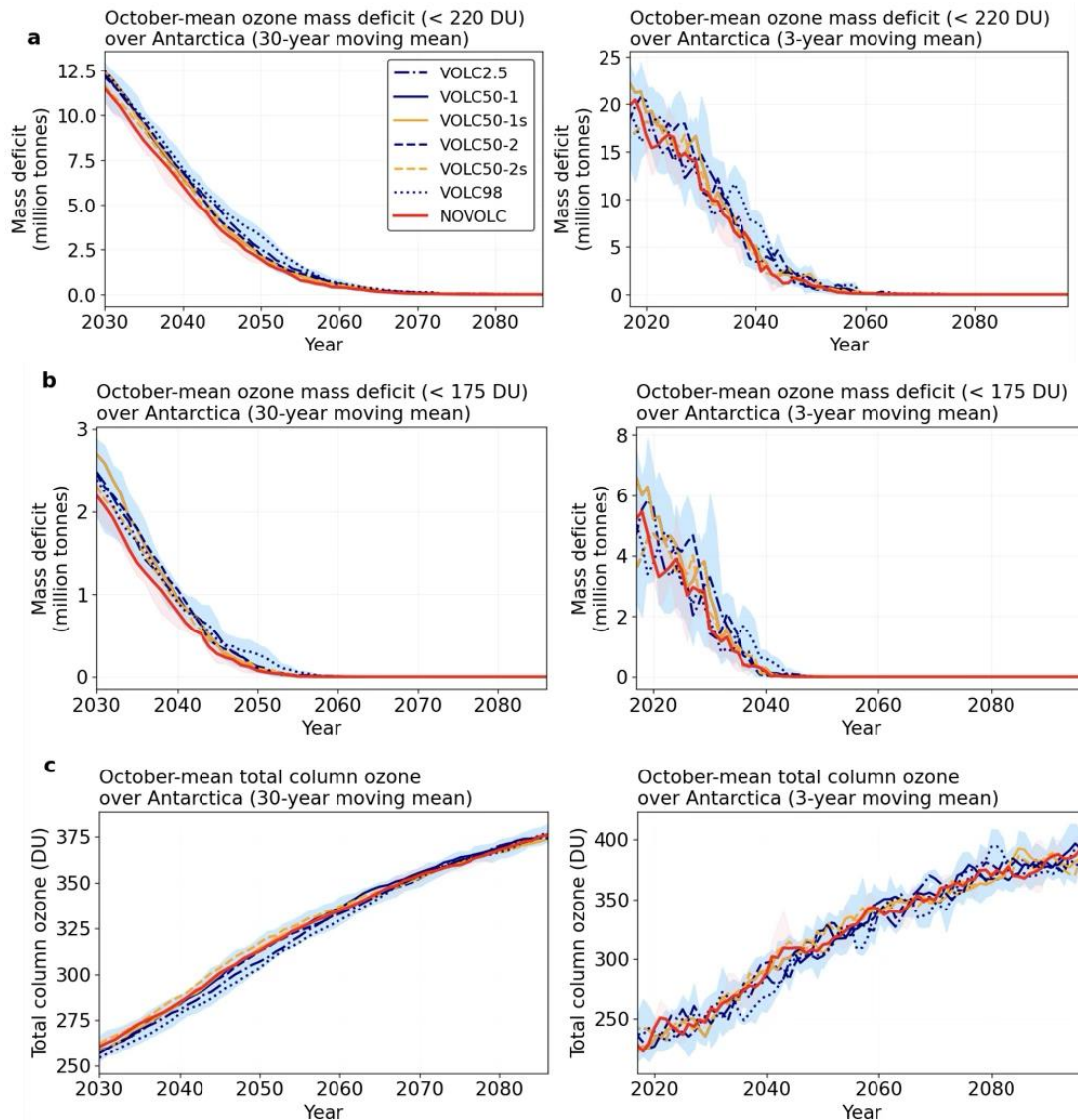
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Figure 5a and 5b shows the October-mean ozone mass deficit over Antarctica with 220 DU and 175 DU thresholds respectively, shown as 30-year and 3-year moving means. The 3-year moving mean reveals substantial interannual variability in ozone mass deficit across scenarios, driven by individual eruptions and internal variability. When smoothed with a 30-year moving mean to show longer-term trends, all stochastic scenarios consistently exhibit higher ozone mass deficit compared to NOVOLC prior to their respective return years (Fig. 5a and 5b). The relative contribution of volcanic effects to ozone mass deficit in year 2030 ranges between 5.7% to 8.7% (0.7 million tonnes to 1.0 million tonnes) using a 220 DU threshold, and 8.9% to 23.0% (0.2 million tonnes to 0.5 million tonnes) using a 175 DU threshold across the stochastic scenarios (Fig. 5a and 5b). The median return years for ozone mass deficit with a 220 DU threshold are the same for the VOLC50 scenarios and their respective runs with small-magnitude eruptions only (VOLC50-1s and VOLC50-2s). For the 175 DU threshold, VOLC50-2s shows an earlier return by 5 years for both the median and 5th to 95th percentile ranges as compared to VOLC50-2, while VOLC-501s shows a 1-year delay in return year as compared to VOLC50-1.



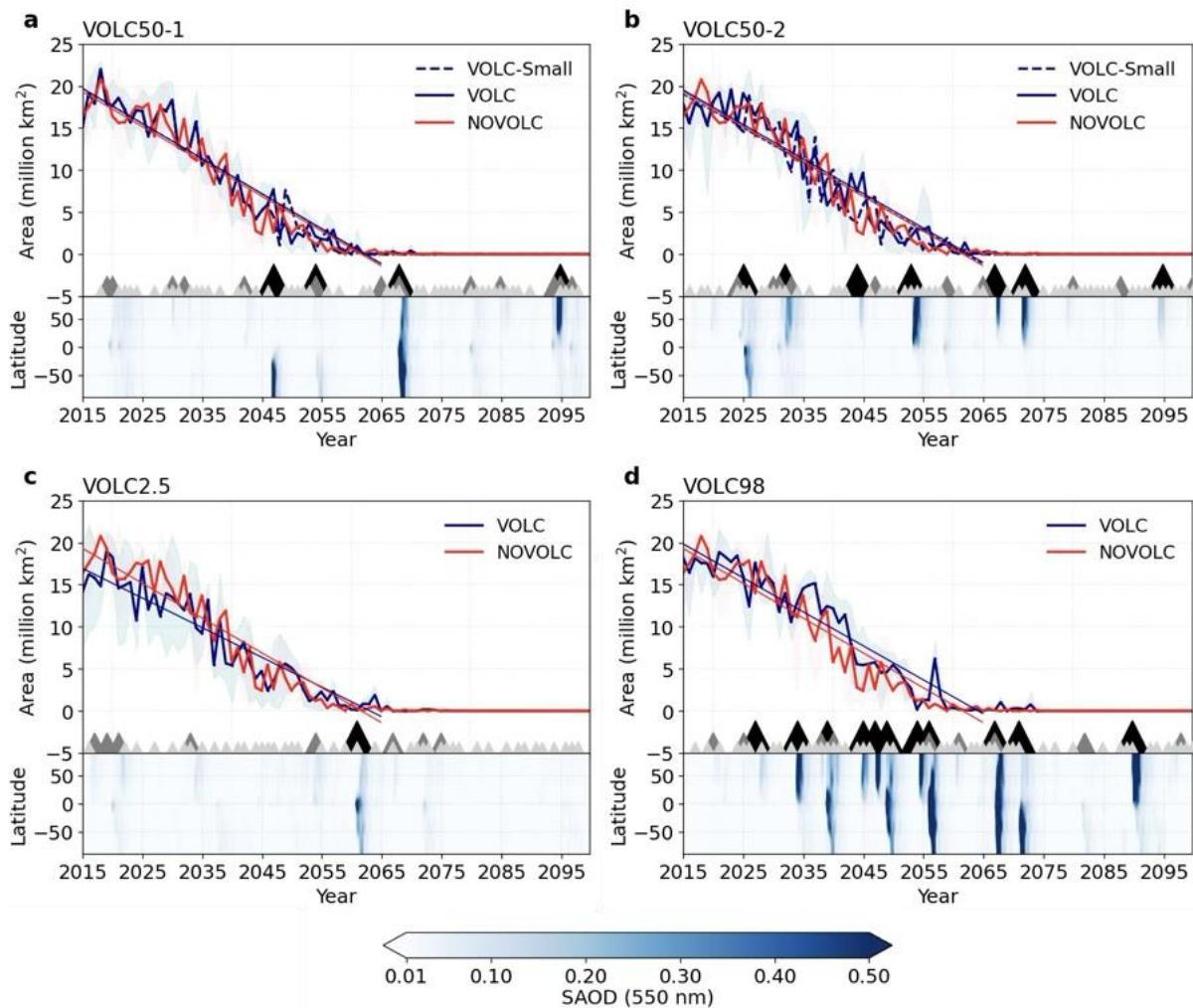
289
 290 **Figure 5. October-mean ozone recovery indicators over Antarctica. (a) Ozone mass deficit using the 220 DU threshold (in million**
 291 **tonnes), (b) ozone mass deficit using the 175 DU threshold (in million tonnes), and (c) total column ozone (in DU). The left column**
 292 **shows the values applied with a 30-year moving mean, and the right column shows the values applied with a 3-year moving mean.**
 293

294 Figure 5c shows the time series of the October-mean Antarctic total column ozone. Compared to the NOVOLC run, our
 295 stochastic scenarios show a 0.5% to 2.8% (1.5 DU to 8 DU) lower Antarctic total column ozone averaged over 2030 to 2050,
 296 which lasts until around 2060s for some stochastic scenarios. The greater volcanic-induced impact on ozone loss in the near-
 297 term is likely due to the higher chlorine concentrations in the atmosphere as compared to the later decades. In our model and
 298 without considering future volcanic eruptions, Antarctic total column ozone returns to its 1980 baseline in 2066 (range: 2062-
 299 2072, NOVOLC). Among all the simulated stochastic scenarios, only the median scenario VOLC50-2 and the high-end
 300 scenario VOLC98 show delays in the Antarctic total column ozone return date, by 3 years and 5 years respectively. One of the
 301 median scenarios with small-magnitude eruptions only (VOLC50-1s) has a 4-year delay in the median return date of total

302 column ozone compared to VOLC50-1, while the other median stochastic scenario (VOLC50-2) has the same return year as
303 the scenario with small-magnitude eruptions only (VOLC50-2s).

304
305 We also examine the responses of the annual mean total column ozone globally and over SH mid-latitudes (Figure S3a and
306 S3b). Our results show that volcanic sulfate aerosols induce a reduction of 0.4% to 0.6% (1.2 DU to 1.9 DU) in total column
307 ozone over SH mid-latitudes, and 0.3% to 0.5% (0.9 DU to 1.6 DU) globally, relative to the NOVOLC simulation for the
308 period 2030–2086. The magnitude of response in global and SH mid-latitude total column ozone is consistently negative
309 throughout this century and comparable to or slightly greater than that over Antarctica (+0.1% to -0.5%, Fig. S3d), where the
310 response shows greater variability.

311
312 Figure 6 shows the October-mean Antarctic ozone hole area for the VOLC and NOVOLC runs, with the zonal-mean
313 stratospheric aerosol optical depth of the respective VOLC runs shown in each panel. To assess the return year of the Antarctic
314 ozone hole area, we assess the recovery trend of the Antarctic ozone hole area. We find that the recovery of the Antarctic
315 ozone hole area back to 220 DU is delayed by 1 to 3 years, except for the VOLC2.5 scenario (Table 1). Small-magnitude
316 eruptions (VOLC50-1s and VOLC50-2s) have no impact on the median return dates of the Antarctic ozone hole area. The
317 longer delay in the VOLC98 scenario can be attributed to the occurrence of a large-magnitude tropical eruption in 2056
318 emitting 114 Tg of SO₂ (Fig. 1), which results in a high SAOD over Antarctica and an Antarctic October ozone hole area of 5
319 million km² relative to NOVOLC (Fig. 6). Although we see a delay in the recovery of the Antarctic ozone hole area in most
320 of the VOLC scenarios for a 220 DU threshold, our simulation results also show that the Antarctic ozone hole area is highly
321 variable between ensemble members. We are not able to perform robust statistical tests due to the limited number of ensemble
322 members. We also examine the Antarctic ozone hole area recovery under a lower total column ozone threshold of 175 DU and
323 150 DU. The Antarctic ozone hole area recovery is delayed by 1 year in most VOLC scenarios except for VOLC2.5 for a 175
324 DU threshold, and there is no delay in return date in all scenarios for a 150 DU threshold (Supplementary Fig. S4 and S5).



325

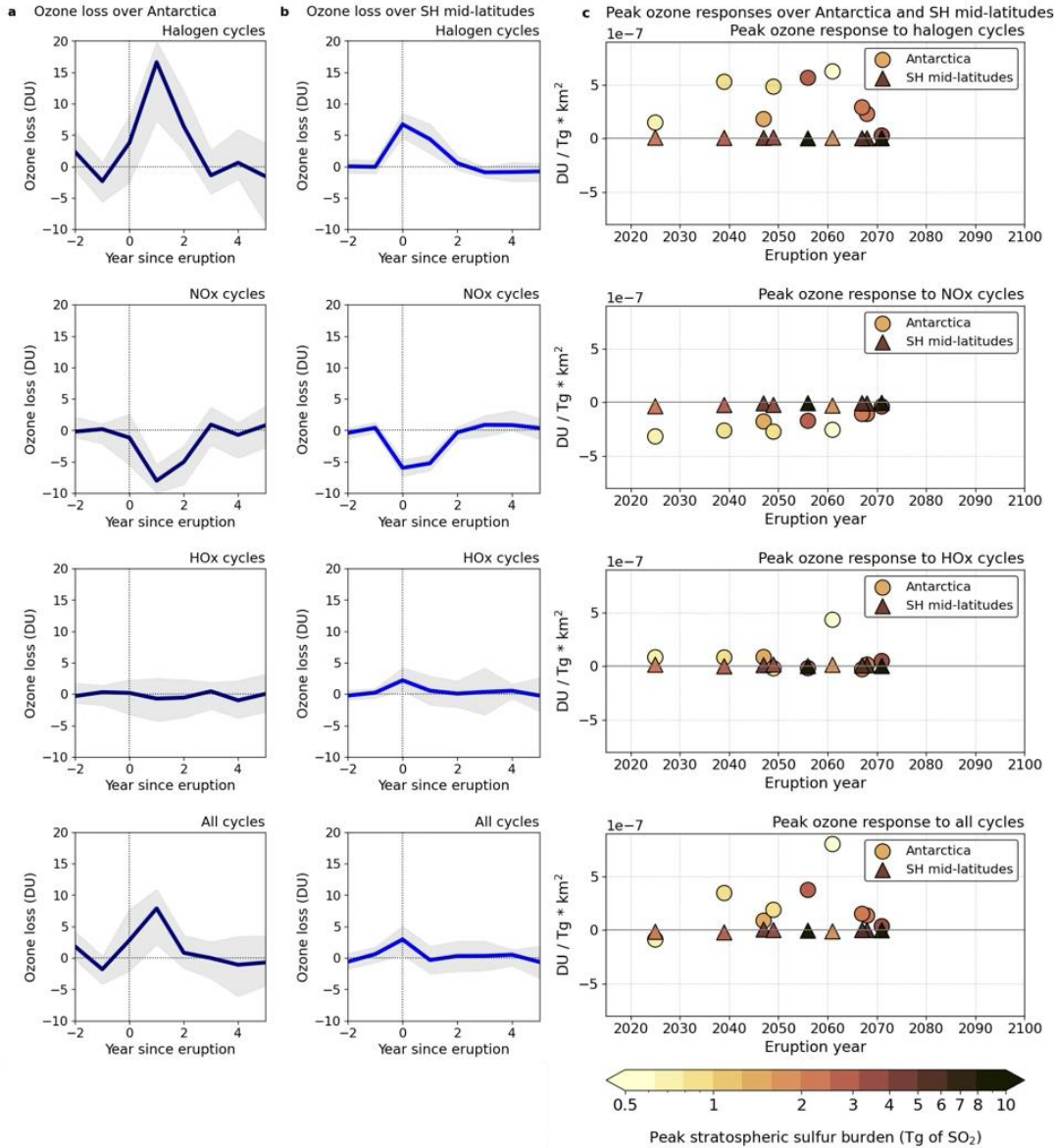
326 **Figure 6. October-mean ozone hole area (in million km²) averaged over Antarctica for (a) VOLC50-1, (b) VOLC50-2, (c) VOLC2.5,**
 327 **and (d) VOLC98. The blue lines show the ensemble mean values and the linear regression for each VOLC scenario, the red lines**
 328 **show the ensemble mean values for NOVOLC, and the shading shows the values of the maximum and minimum values across the**
 329 **ensemble members. The triangles refer to the occurrence of eruptions in each stochastic scenario: the black triangles refer to**
 330 **eruptions with > 3 Tg of SO₂ injection, the grey triangles refer to eruptions with 1 to 3 Tg of SO₂ injection, and the light grey**
 331 **triangles refer to eruptions with 0.1 to 1 Tg of SO₂ injection. The lower panel shows the zonal mean stratospheric aerosol optical depth at 550**
 332 **nm for the four VOLC runs.**
 333

334 3.3 Chemical changes in the Antarctic and SH mid-latitudes lower stratosphere

335

336 To assess the volcanic effects on the chemical loss of ozone, we select 9 large-magnitude eruptions with stratospheric sulfur
 337 burden > 0.2 Tg of S over Antarctica for composite analysis (Figure 1b). Fig. 7a shows the composite analysis results of the
 338 cumulative stratospheric ozone loss anomaly relative to NOVOLC over Antarctic and SH mid-latitudes for each catalytic
 339 cycle. A positive value indicates a net ozone loss via the catalytic cycle, while a negative value indicates a net ozone gain. Our

340 analysis shows a net ozone loss via halogen cycles at year zero after volcanic eruptions, and a net ozone increase via NO_x
 341 cycle over Antarctica and SH mid-latitudes (Fig. 7a and 7b). The volcanic-induced ozone response via halogen and NO_x cycles
 342 is stronger over Antarctica than the SH mid-latitudes, as shown in Figure 10a. Stratospheric volcanic sulfate aerosols catalyse
 343 the release of active chlorine from reservoir species via heterogeneous chemistry, forming ClO_x (Fig. S6). This enhances
 344 stratospheric ozone loss via halogen cycles over Antarctica and SH mid-latitudes (Figure S8 and S9). The sulfate aerosols also
 345 catalyse the hydrolysis of N₂O₅ (Fig. S7), suppressing the NO_x-catalysed ozone loss for 1 to 2 years after the eruption (Fig.
 346 S8 and S9).
 347



348 **Figure 7. Composite analysis of the cumulative stratospheric ozone loss (up to 25 km) relative to NOVOLC for the 9 selected**
 349 **eruptions (a) over Antarctica, integrated by time from August to the following July, and (b) SH mid-latitudes, cumulative loss for**
 350 **12 months after the onset of each eruption. Year 0 denotes the year of eruption. The shaded region shows the range of the first and**
 351

352 **third quartiles. (c) Peak ozone response of the selected eruptions to halogen cycles, NO_x cycle, HO_x cycle, and all cycles (halogen +**
353 **NO_x + HO_x) over Antarctica (circles) and SH mid-latitudes (triangles). The values are normalised with the respective peak**
354 **stratospheric S burden for each eruption and total area over Antarctica and SH mid-latitudes, respectively.**
355

356 Fig. 7c shows the peak ozone response (i.e., the response at year 0 or 1 of the composite analysis) via the catalytic cycles,
357 normalised by the respective peak stratospheric sulfur burden and total area over Antarctica and SH mid-latitudes. The
358 normalised peak ozone response exhibits a decreasing trend with eruption year for both regions and is stronger over Antarctica
359 than the SH mid-latitudes (Fig. 7a and 7b). However, the overall magnitude of ozone response, summing all catalytic cycles,
360 is comparable between the Antarctica and SH mid-latitudes (Fig. 7c). Although we see clear decreasing trends in ozone
361 response over both regions, we are unable to identify the shift in the ozone chemistry response due to a lack of large-magnitude
362 eruptions with > 0.2 Tg of S over Antarctica after 2070 in these simulations.

364 **4 Discussion**

365
366 Using a plume-aerosol-chemistry-climate model UKESM-VPLUME with an improved stochastic volcanic forcing, we assess
367 the effect of future volcanic eruptions with sulfur injections on Antarctic and SH mid-latitude ozone recovery by evaluating
368 the October-mean values of three ozone recovery indicators. We find that the extent of the delay in Antarctic ozone return
369 dates depends on the eruption timing, latitude and aerosol distribution in the stochastic scenarios, with two scenarios showing
370 delays in ozone mass deficit of 2 years (VOLC2.5) and up to 5 years (VOLC98) with a 220 DU ozone threshold (Table 1). If
371 we use a lower ozone threshold of 175 DU, the relative effect of volcanic eruptions on ozone mass deficit can be up to 23% in
372 2030 for a high-end future scenario (Fig. 5b). For the 175 DU threshold, VOLC50-2 shows the longest delay (6 years), while
373 VOLC2.5 and VOLC98 both show 3-year delays, and VOLC50-1 has the same return year as NOVOLC. The apparent
374 inconsistencies in the delays of recovery across scenarios arise from the different recovery timelines, which is due to
375 differences in the timing of volcanic eruptions across scenarios, as well as varying sensitivities of the two ozone mass deficit
376 thresholds. The recovery for the 175 DU threshold is much faster (median year of 2038 in NOVOLC) than the 220 DU
377 threshold (median year of 2062 in NOVOLC), making each metric sensitive to eruptions occurring at different times. For
378 instance, VOLC2.5 and VOLC98 both have large-magnitude eruptions with SH aerosol distribution before 2062, thus showing
379 greater delay in return years for the 220 DU threshold. VOLC50-2 shows the longest delay for the 175 DU threshold because
380 it has higher SH SAOD than VOLC2.5 and VOLC98 before 2038 (Fig. 6). In contrast, the first large-magnitude eruption in
381 VOLC50-1 occurs in 2047, after the 175 DU threshold recovery year, and thus showing no delay for this metric.

382 We also show that future scenarios with small-magnitude eruptions only (VOLC50-1s and VOLC50-2s) have no effect on the
383 return years of Antarctic ozone mass deficit with a 220 DU threshold, but have a mixed effect on the recovery of deeper ozone
384 holes at the 175 DU threshold. One of the scenarios with small-magnitude eruptions only (VOLC50-2s), which has the lowest
385 annual SO₂ flux (0.57 Tg SO₂ yr⁻¹), shows earlier recovery by 5 years for the 175 DU threshold metric, while the other small-
386 magnitude scenario (VOLC50-1s) shows a 1-year delay compared to VOLC50-1 (Table 1). This suggests that ozone recovery
387 timing is primarily governed by large-magnitude eruptions, with small-magnitude eruptions playing a secondary
388 role. However, the large uncertainty ranges and overlapping values in the return years between VOLC and NOVOLC scenarios
389 reflect substantial internal variability between ensemble members; this highlights that recovery projections should be
390 interpreted as ranges rather than median values alone. Our simulations are also limited by the number of ensemble members.
391 Future modelling experiments with larger ensemble sizes would further quantify the uncertainty in volcanic effects on ozone
392 recovery and distinguish between volcanic forcing effects and internal variability.

393 In terms of the total column ozone, we find a delay of up to 5 years in the high-end volcanic scenario (VOLC98), which
394 includes volcanic eruptions with large stratospheric sulfur burdens and SAOD over Antarctica between the years 2055 and
395 2065 (Fig. 1 and 5). Our results show that an adopted large-magnitude tropical eruption in the year 2056 in the VOLC98
396 scenario, with an Antarctic sulfur burden of 2.7 Tg of S, leads to an Antarctic ozone hole exceeding 5 million km² in October
397 (Fig. 6d). This occurs despite the Antarctic ozone column being otherwise close to returning to 1980 values in the 2050s. The
398 simulated ozone hole is comparable in size to that caused by the 2015 Calbuco eruption, which emitted around 0.4 Tg of SO₂
399 and resulted in an October ozone hole of about 4.5 million km² larger (Solomon et al., 2016). The comparable ozone hole size
400 is attributable to the substantially higher stratospheric chlorine loading in 2015 compared to that projected for the mid-21st
401 century. These findings demonstrate that future eruptions with significant volcanic sulfur injections have the potential to cause
402 large Antarctic ozone holes in the middle of this century according to our simulations.

403 Our results differ from with the findings of Naik et al. (2017), another previous modelling study that assessed the effects of
404 future volcanic eruptions on stratospheric ozone recovery in climate projections. Naik et al. (2017) use a prescribed constant
405 future volcanic forcing equivalent to the 1860-1999 mean volcanic forcing to evaluate stratospheric ozone responses in climate
406 projections using the GFDL-CM3 model under the RCP4.5 and RCP8.5 scenarios. Their model results suggest that an
407 enhanced volcanic sulfate aerosol burden does not affect Antarctic ozone column return dates but leads to an earlier recovery
408 of global stratospheric ozone. The disagreement between Naik et al. (2017) and our study potentially arises from differences
409 in forcing magnitude, model responses and chemistry representations between GFDL-CM3 and UKESM. For instance, the
410 ClO/Cly ratio at 50°N in Naik et al. (2017) is a factor of 4 lower than observations on 22 March 1992, following the 1991 Mt.
411 Pinatubo eruption (Fahey et al., 1993), suggesting that the GFDL model may not adequately capture the response of halogen
412 chemistry. In addition, the UKESM model version used in our study has limited heterogeneous reactions, whereas a recent
413 study by Ming et al. (2020) used a new heterogeneous scheme in UKESM with eight additional reactions, producing a better
414 match with the observed total column ozone over Antarctica. Beyond differences in model chemistry responses, the contrasting
415 results are likely stemming from the experimental design (prescribed versus interactive volcanic forcing) and different future
416 emission scenarios used (RCP versus SSP). These differences hinder direct comparison between the two studies. Future model
417 comparison studies using consistent future emission scenarios and volcanic forcing magnitudes are necessary to assess the
418 projected ozone responses to volcanic eruptions across different models.

419
420 Our composite analysis of cumulative stratospheric ozone response reveals a net ozone loss for the 9 selected large-magnitude
421 eruptions (Fig. 7c). The stratospheric ozone response to volcanic eruptions is sensitive to the amount of aerosol mass
422 transported into the polar vortex (Fig. 1). Our results show a linear decreasing trend of peak ozone response after large-
423 magnitude eruptions. However, due to the lack of large-magnitude eruptions after 2070 with high sulfur burden over Antarctica
424 in our stochastic scenarios, we do not see a clear shift from net stratospheric ozone loss to production over Antarctica. We
425 expect that eruptions occurring after year 2070 will eventually lead to a net ozone gain instead of ozone loss. Such shifting in
426 the chemistry will affect the future ozone recovery trend. Our findings indicate that even a small-magnitude eruption in 2092,
427 with a SO₂ injection mass of less than 3 Tg of SO₂ and an Antarctic sulfur burden as low as 0.08 Tg of S, can lead to a small
428 net ozone loss over Antarctica (Fig. S8). The stratospheric ozone response to volcanic eruptions is also influenced by
429 stratospheric dynamics. However, due to the coupled nature of our model simulations, we are unable to isolate and quantify
430 the dynamical component of the ozone response. Furthermore, the stochastic nature of our scenarios and the limited model run
431 length constrain our ability to identify the timing of the shift in stratospheric ozone response to volcanic eruptions, as the shift
432 in the ozone chemistry may occur after 2100. Nonetheless, our study demonstrates that the composite analysis of the
433 stratospheric ozone loss via catalytic cycles is a valuable tool for evaluating the chemical response of stratospheric ozone to
434 volcanic eruptions.

435 Although our study primarily focuses on the ozone response over Antarctica and SH mid-latitudes, volcanic eruptions can also
436 induce significant changes in ozone concentrations over the Arctic and NH mid-latitudes (Fig. S3c, S10 and S11). Composite
437 analysis of 15 selected eruptions (with a stratospheric sulfur burden exceeding 0.5 Tg of S over the NH mid-latitudes) reveals
438 substantial net ozone loss of 9 DU over the Arctic (interquartile range: 18 DU to -2 DU) and 2 DU loss over the NH mid-
439 latitudes (interquartile range: 6 DU to -3 DU) (Figures S12). However, the ozone response over these regions exhibits greater
440 uncertainties and variability compared to the SH. The limited ensemble size in this study constrains our ability to draw robust
441 conclusions regarding the NH ozone response. Furthermore, the occurrence of multiple volcanic eruptions between 2040 and
442 2060 in the VOLC98 scenario (Fig. 1) may introduce biases in the composite analysis, as the overlapping ozone responses
443 from these closely timed eruptions could potentially interfere with one another (Fig. S12). Consequently, while our findings
444 highlight the potential for volcanic eruptions to impact stratospheric ozone recovery over polar regions, additional research
445 with larger ensemble simulations is necessary to more comprehensively characterise the ozone response over the Arctic and
446 NH mid-latitudes.

447 We acknowledge that UKESM1.1 exhibits a stratospheric cold bias and excessively strong polar vortex that leads to prolonged
448 ozone depletion over Antarctica (Fig. 4). The comparison of the UKESM1.1 simulation with ML-TOMCAT shows that the
449 climatological mean of UKESM1.1 overestimates lower stratospheric ozone loss over Antarctica and SH mid-latitudes. Due
450 to these model biases, our results likely overestimate the cumulative ozone loss over Antarctica and SH mid-latitudes.
451 However, the relative volcanic effects on ozone in our simulated scenarios remain robust. In addition, our stochastic scenarios
452 include stratospheric volcanic SO₂ emissions only, but not volcanic halogen species, water vapour and VSL chlorine and
453 bromine compounds, which affect stratospheric ozone recovery. Klobas et al. (2017) emphasised the sensitivity of future ozone
454 changes to a Pinatubo-like eruption with VSL bromine injection between 0 to 8 pptv, which can lead to changes in total column
455 ozone over Antarctica between -3% and 3% under one future representative concentration pathway (i.e., RCP6.0). Although
456 VSL chlorine has a small contribution to total stratospheric chlorine (about 3%) between 2010 to 2019 (Bednarz et al., 2022),
457 the increasing emission of VSL chlorine highlights the potential importance of its impact on future ozone changes. By not
458 accounting for VSL species in our simulations, we likely underestimate the full magnitude of volcanic impacts on ozone
459 recovery. The lack of comprehensive historical records on volcanic halogen emissions, the variability in halogen injections
460 across different eruptions, and the uncertainties in future VSL halogenated compounds emissions make it challenging to project
461 future stratospheric halogen loadings using our stochastic approach. Therefore, our model-simulated effects on ozone represent
462 a lower bound of the potential effects of future volcanic eruptions on ozone depletion. We anticipate that including volcanic
463 halogen, water vapour, and VSL chlorine and bromine emissions in our stochastic scenarios would likely result in further
464 delays to Antarctic and SH mid-latitude ozone recovery. It is essential to improve current volcanic halogen emission datasets
465 and proxy records, and account for the emission of halogenated VSL compounds to better assess the impact of future volcanic
466 eruptions on stratospheric ozone recovery.

467
468 Our model experiments use a high-end future anthropogenic emission scenario (SSP3-7.0) which has high future methane
469 emissions (Meinshausen et al., 2020). Previous studies showed that stratospheric ozone responses to volcanic eruptions are
470 sensitive to the background anthropogenic greenhouse gas emissions (Naik et al., 2017; Klobas et al., 2017). For instance,
471 Klobas et al. (2017) demonstrate that ozone depletion after volcanic eruption is stronger in low-end RCP scenarios, which is
472 attributed to a warmer stratosphere and lower methane emissions. As methane reacts with chlorine in the atmosphere, higher
473 methane emissions in the future will lead to lower stratospheric reactive chlorine species and thus suppress the ClO_x-catalysed
474 ozone loss. For the same model simulation under a low-end SSP scenario, we expect a stronger volcanic-induced ozone
475 response and potentially a later year of shift in ozone response (net loss versus net gain, see Section 3.3). Future studies can
476 design specific model experiments with the co-injection of halogen species and water vapour to quantify the timing of the shift
477 in Antarctic ozone chemistry under different future emission scenarios.

478

479 **4 Conclusion**

480

481 Using the plume-aerosol-chemistry-climate model UKESM-VPLUME with a stochastic representation of volcanic sulfur
482 emissions, we quantified the effects of six stochastic scenarios on projected ozone recovery over Antarctica and Southern
483 Hemisphere mid-latitudes. We showed that a high-end future eruption scenario delays the recovery of the Antarctic total
484 column ozone and ozone mass deficit for up to 5 and 6 years respectively. These delays represented relatively small
485 perturbations to the overall Antarctic and SH mid-latitudes ozone recovery. Our results also showed that the peak ozone loss
486 due to large-magnitude volcanic eruptions in our stochastic scenarios weakens from 2015 to 2070 (Fig. 7c) as a result of
487 declining CFC concentrations. Although our stochastic future eruption scenarios caused only a few years of delay in the ozone
488 recovery metrics, large-magnitude eruptions can still potentially lead to substantial perturbations in stratospheric ozone prior
489 to its recovery. For example, the mid-century eruption in the VOLC98 scenario led to an Antarctic ozone hole up to 5 million
490 km² (Fig. 5c and 6d). Our results also showed that small-magnitude eruptions have little effect on the recovery of Antarctic
491 and SH mid-latitudes stratospheric ozone. These results highlighted the importance of accounting for volcanic variability in
492 assessing stratospheric ozone recovery over Antarctica and SH mid-latitudes. Volcanic eruptions along with other events that
493 perturb the stratospheric aerosol layer, such as wildfires and deliberate stratospheric aerosol injections, remain as future
494 challenges in assessing ozone recovery (WMO, 2022; Solomon et al., 2023; Chipperfield et al., 2024). Enhancing our
495 understanding of volcano-induced ozone responses, including the effects of volcanic halogen emissions, is a crucial aspect for
496 accurately assessing the healing of the Antarctic ozone layer. Continued improvements in models, volcanic emission datasets,
497 and observational monitoring of the stratosphere will be critical to assess the stability and resilience of ozone layer recovery
498 in the coming decades.

499 **Code availability**

500 The code to reproduce the figures in the main text and supplementary information are available on Github:

501 https://github.com/maychim/volc_ozone

502

503 **Data availability**

504 The data used in this study are available here: 10.5281/zenodo.15838168

505

506 **Author Contribution**

507 MMC conceptualised the paper with the advice from TJA and AS. MMC performed the UKESM simulations with the support
508 from NLA. MMC performed the data analysis, visualisation, and wrote the first draft of the paper. HG, BJ, and SS provided
509 critical feedback and helped shape the research and analysis. All authors discussed the results and commented on the
510 manuscript.

511

512 **Competing interest**

513 The authors declare no competing interests.

514

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523

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