

## **Inter-hemispheric asymmetry in the sea-ice response to volcanic forcing simulated by MPI-ESM (COSMOS-Mill)**

Zanchettin Davide, Oliver Bothe, Claudia Timmreck, Jürgen Bader, Alexander Beitsch, Hans-F. Graf, Dirk Notz, and Johann H. Jungclaus

### ***Response to interactive comment by Anonymous Reviewer #1***

We thank Anonymous Reviewer #1 for his/her appreciation of our work and his/her constructive criticism. We have already briefly replied in the interactive discussion to the two major concerns he/she expressed in the main comment published online in the discussion. Our initial online reply is repeated here for the sake of completeness, followed by a description of the changes we made in response to the reviewer's helpful suggestions.

Reviewer's main comment n.1: "there is no mystery about the Antarctic sea ice response to volcanic eruptions"

We disagree with the Reviewer's comment that the observed Arctic-Antarctic sea-ice dichotomy is robustly simulated by climate models, and that there is no mystery about Antarctic sea-ice response to volcanic forcing. For instance, Turner et al. (2013), who analyzed Antarctic sea-ice representation in an ensemble of CMIP5 simulations, conclude that processes responsible for the observed increase over the last 30 years are not simulated correctly. Moreover, in his recent comment published in Nature, John King concludes about Antarctic sea ice that "understanding its behavior and improving its representation in climate models must remain a high priority for climate scientists" (King, 2014).

We agree that there is not only room but need for improving our understanding of simulated sea-ice behavior, and our study primarily stems from this need. In particular, there are few assessments of simulated responses of Antarctic sea ice to volcanic forcing. Our study is to our knowledge the first description of simulated decadal climate responses to extremely large volcanic eruptions that focuses on inter-hemispheric asymmetry and sea ice. We found especially the Antarctic response interesting and its interpretation challenging, so much that disentangling the effects of dynamical and thermodynamical, as well as of local and global contributions would require dedicated additional sensitivity experiments.

As we stressed out in the manuscript, we indeed regard idealized "supervolcano" experiments as reverse analogs of warming experiments. However, as we explained in the manuscript, they do not merely describe the opposite climatic effects of positive radiative imbalances, since volcanic forcing induces complex multi-scale dynamical and thermodynamical responses.

If we are encouraged to submit a revised manuscript, we will extend the discussion about the present state of research concerning the Antarctic sea ice response to global warming.

Reviewer's main comment n.2: "Before any climate model is used, it has to be evaluated"

MPI-ESM (COSMOS-Mill) (Jungclaus et al., 2010) is based on the well known and widely used ECHAM5/MPIOM coupled general circulation model (CGCM) developed at the Max Planck Institute for Meteorology in Hamburg (MPI-M), with implementation of an interactive carbon cycle. The ECHAM5/MPIOM CGCM participated to the CMIP3 exercise, and has been extensively evaluated in that context.

MPI-ESM (COSMOS-Mill) was developed as part of the internal project “Community Simulations of the last Millennium” at MPI-M

(<http://www.mpimet.mpg.de/en/science/projects-new/projects-archive/millennium.html>)

and was further developed thereafter to allow performing idealized volcanic experiments (Timmreck et al., 2010). More details about the characteristics of MPI-ESM (COSMOS-Mill) can be found in the report by Budich et al. (2010).

Different aspects of the climate and its variability simulated by MPI-ESM (COSMOS-Mill) have been assessed in a number of studies in comparison to observations, proxy-based reconstructions as well as within a multi-model framework (e.g., Henriksson et al., 2012; Beitsch et al., 2013; Bothe et al., 2013; Fernández-Donado et al., 2013; Schubert et al., 2013; Zanchettin et al., 2013a,b). We will expand the description of the model and include a list of references regarding its evaluation if we are encouraged to submit a revised manuscript.

As a final note, the model’s name adopted in this and other studies (e.g., Bothe et al., 2013) reflects the fact that the COSMOS community used it for the Millennium Experiment. The name also prevents confusion between the ECHAM5/MPIOM ESM and its successor: the ECHAM6/MPIOM-based MPI-ESM used in CMIP5. Giorgietta et al. (2013) and Jungclaus et al. (2013) describe the characteristics of the atmospheric and ocean components of the ECHAM6/MPIOM-based MPI-ESM.

In the revised version of the manuscript we have extended the introduction including more details about the present state of research concerning the Antarctic sea ice evolution during the past decades. We now briefly introduce the characteristics of the observed trends and better discuss related modeling issues. We also expanded our discussion about the attribution of the observed changes, concerning both internal variability and the stratospheric ozone depletion. Please refer to the attached manuscript with tracked changes to see the modifications done to the text. The following references have been added: Turner et al., 2009; King, 2014; Gillett and Thompson, 2003; Sigmond and Fyfe, 2014.

Concerning the model and its evaluation, we have expanded the description of the model and included a list of references regarding the assessment of its different aspects. Please refer to the attached manuscript with tracked changes to see the modifications done to the text. The following references have been added to this purpose: Tietsche et al., 2011; Henriksson et al., 2012; Beitsch et al., 2013; Bothe et al., 2013; Fernández-Donado et al., 2013; Notz and Marotzke, 2012; Schubert et al., 2013; Zanchettin et al., 2013b; Li et al., 2013. More specifically concerning simulated sea ice, we have also added two supplementary figures (Figures S1 and S2 in the supplementary material) showing the climatological features of sea-ice concentration from the control run. The general characteristics of sea-ice simulated by MPI-ESM-COSMOS-Mill seem to agree well with

those simulated by ECHAM5/MPIOM, with both components in a higher resolution configuration than used in the present study, as described by Jungclaus et al. (2006).

***Response to interactive comment by Anonymous Reviewer #2***

We thank Anonymous Reviewer #2 for his/her appreciation of our work and his/her constructive criticism.

We agree with the Reviewer that atmospheric circulation changes are an important factor determining the climate response to the perturbation, especially as the regional scale is concerned. We have therefore deepened our investigation of the atmospheric response in our super-volcano ensemble. In doing so, we considered the following. It is unclear from the Reviewer's comment whether the necessary clarifications concern the more general aspects of the atmospheric response mechanism or, instead, more specific processes of it related to the shown sea-ice anomalies. Moreover, a detailed assessment of the atmospheric response in our "supervolcano" ensemble would require a dedicated study, beyond the aims of the present investigation. We have therefore decided to add a limited set of additional latitude-altitude plots as per Reviewer's request and have collected them in the supplementary document (Figures S3-S11). The results illustrated in the supplement are mainly summarized in the main text in the first three paragraphs of section 3.3. Please refer to the attached manuscript with tracked changes to see the modifications done to the text. The results seem to confirm that atmospheric internal processes play an important role in the first response phase, while later on the response is dominated by oceanic and ocean-atmosphere coupled processes.

Minor comments (original comments by Reviewer #2 in italics):

*Line 23, Page 126: Please briefly explain what is COSMOS-Mill.*

Following also the comment by Reviewer #1 we have expanded the explanation of the model. In particular, we now specify that "The model version's name reflects the fact that the Community Earth System Modeling (COSMOS) community used it for the Millennium Experiment." We also remark that the notation MPI-ESM COSMOS-Mill appears in the available scientific literature (Bothe et al., 2013).

*Line 25, Page 140 "a weakening of both polar and mid-latitude flow (Fig. 11d)": It seems that the polar flow is intensified.*

Climatological 10-meter winds around and over Antarctica are weak easterlies (in the model with zonal-mean values peaking at 4.5 m/s westward at ~70°S latitude) whereas the arrows depict an anomalous westerly flow (though irregular). The figure therefore correctly depicts a weakening of the polar zonal flow.

*Caption of Figure 7 "500 hPa meridional wind (b) and zonal wind (c)": 500 hPa zonal wind (b) and meridional wind (c)*

Thanks, corrected

**References**

Beitsch, A., J. H. Jungclaus, and D. Zanchettin (2013) Patterns of decadal-scale Arctic warming events in simulated climate. *Clim. Dyn.* doi:10.1007/s00382-013-2004-5

Bothe, O., J. H. Jungclauss, D. Zanchettin, and E. Zorita (2013) Climate of the last millennium: ensemble consistency of simulations and reconstructions. *Clim. Past*, 9, 1089–1110, doi:10.5194/cp-9-1-2013

Budich, R., M. Giorgetta, J. Jungclauss, R. Redler, and C. Reick (2010) The MPI-M Millennium Earth System Model: An Assembling Guide for the COSMOS Configuration [http://cosmos.enes.org/fileadmin/user\\_upload/cosmos/documents/Documentation/M-ESM-Assembling-Guide.pdf](http://cosmos.enes.org/fileadmin/user_upload/cosmos/documents/Documentation/M-ESM-Assembling-Guide.pdf)

Fernández-Donado, L., and co-authors (2013) Large-scale temperature response to external forcing in simulations and reconstructions of the last millennium, *Clim. Past*, 9, 393–421, doi:10.5194/cp-9-393-2013, 2013.

Gillett, N. P., and D. W. J. Thompson (2003) Simulation of Recent Southern Hemisphere Climate Change. *Science*, 302 (5643), 273-275, doi:10.1126/science.1087440

Giorgetta, M. A., and Coauthors (2013) Climate change from 1850 to 2100 in MPI-ESM simulations for the Coupled Model Intercomparison Project 5. *J. Adv. Model. Earth Syst.*, 5, 572–597, doi:10.1002/jame.20038

Henriksson, S.V. , P. Räisänen, J. Silén, and A. Laaksonen (2012) Quasiperiodic climate variability with a period of 50-80 years: Fourier analysis of measurements and Earth System Model simulations. *Clim. Dyn.* 39: 7-8, 1999-2011, doi:10.1007/s00382-012-1341-0

Jungclauss J.H., N. Keenlyside, M. Botzet, H. Haak, J. J. Luo, M. Latif, J. Marotzke, U. Mikolajewicz, and F. Roeckner (2006) Ocean circulation and tropical variability in the coupled model ECHAM5/MPI-OM. *J Clim* 19:3952–3972

Jungclauss, J. H., et al. (2010), Climate and carbon-cycle variability over the Last Millennium. *Clim. Past*, 6, 723-737, doi:10.5194/cp-6-723-2010

Jungclauss J. H., and Coauthors (2013) Characteristics of the ocean simulations in MPIOM, the ocean component of the Max Planck Institute Earth System Model. *J. Adv. Model. Earth Syst.*, 5, 422–446, doi:10.1002/jame.20023

Menary, M.B., W. Park, K. Lohmann, M. Vellinga, M.D. Palmer, M. Latif, and J.H. Jungclauss (2012) A multimodel comparison of centennial Atlantic meridional overturning circulation variability. *Clim. Dyn.* 38:11-12, 2377-2388

Notz, D., and J. Marotzke (2012) Observations reveal external driver for Arctic sea-ice retreat. *Geophys. Res. Lett.* 39(8), doi:10.1029/2012GL051094

Schubert, J. J., B. Stevens, T. Crueger. (2013) Madden-Julian oscillation as simulated by the MPI Earth System Model: Over the last and into the next millennium. *Journal of Advances in Modeling Earth Systems*, 5:1, 71-84, doi:10.1029/2012MS000180

Sigmond, M., and J. C. Fyfe (2014) The Antarctic Sea Ice Response to the Ozone Hole in Climate Models. *J. Climate*, 27, 1336–1342, doi: <http://dx.doi.org/10.1175/JCLI-D-13-00590.1>

Turner, J., J. C. Comiso, G. J. Marshall, T. A. Lachlan-Cope, T. Bracegirdle, T. Maksym, M. P. Meredith, Z. Wang, and A. Orr (2009) Non-annular atmospheric circulation change induced by stratospheric ozone depletion and its role in the recent increase of Antarctic sea ice extent. *Geophys. Res. Lett.*, 36, L08502, doi:10.1029/2009GL037524

Turner, John, Thomas J. Bracegirdle, Tony Phillips, Gareth J. Marshall, J. Scott Hosking, 2013: An Initial Assessment of Antarctic Sea Ice Extent in the CMIP5 Models. *J. Climate*, 26, 1473–1484. doi: <http://dx.doi.org/10.1175/JCLI-D-12-00068.1>

King, J. (2014) Climate science: A resolution of the Antarctic paradox. *Nature* 505, 491–492, doi:10.1038/505491a

Timmreck, C., H.-F. Graf, S. J. Lorenz, U. Niemeier, D. Zanchettin, D. Matei, J. H. Jungclaus and T.J. Crowley (2010), Aerosol size confines climate response to volcanic super-eruptions, *Geophys. Res. Lett.* 37, L24705, doi:10.1029/2010GL045464

Zanchettin, D., A. Rubino, D. Matei, O. Bothe, and J. H. Jungclaus (2013a) Multidecadal-to-centennial SST variability in the MPI-ESM simulation ensemble for the last millennium. *Clim. Dyn.*, 40:5, 1301-1318, doi:10.1007/s00382-012-1361-9

Zanchettin, D., O. Bothe, H. F. Graf, S. J. Lorenz, J. Luterbacher, C. Timmreck, and J. H. Jungclaus (2013b) Background conditions influence the decadal climate response to strong volcanic eruptions. *J. Geophys. Res. Atm.*, 118(10): 4090-4106, doi:10.1002/jgrd.50229

1 **Inter-hemispheric asymmetry in the sea-ice response to volcanic forcing simulated**  
2 **by MPI-ESM (COSMOS-Mill)**

3  
4 Zanchettin Davide<sup>1\*</sup>, Oliver Bothe<sup>2</sup>, Claudia Timmreck<sup>1</sup>, Jürgen Bader<sup>1</sup>, Alexander  
5 Beitsch<sup>1</sup>, Hans-F. Graf<sup>3</sup>, Dirk Notz<sup>1</sup>, and Johann H. Jungclaus<sup>1</sup>

6  
7 <sup>1</sup> Max Planck Institute for Meteorology, Bundesstr. 53, 20146 Hamburg, Germany

8 <sup>2</sup> Leibniz Institute of Atmospheric Physics at the University of Rostock, Kühlungsborn, Germany

9 <sup>3</sup> University of Cambridge, Centre for Atmospheric Science, Downing Place, Cambridge CB2  
10 3EN, United Kingdom

11  
12 \* corresponding author: [davide.zanchettin@mpimet.mpg.de](mailto:davide.zanchettin@mpimet.mpg.de)  
13  
14

15 **Abstract**

16 The decadal evolution of Arctic and Antarctic sea ice following strong volcanic eruptions  
17 is investigated in four climate simulation ensembles performed with the COSMOS-Mill  
18 version of the Max Planck Institute-Earth System Model. The ensembles differ in the  
19 magnitude of the imposed volcanic perturbations, with sizes representative of historical  
20 tropical eruptions (1991 Pinatubo and 1815 Tambora) and of tropical and extra-tropical  
21 “supervolcano” eruptions. A post-eruption Arctic sea-ice expansion is robustly detected  
22 in all ensembles, while Antarctic sea ice responds only to “supervolcano” eruptions,  
23 undergoing an initial short-lived expansion and a subsequent prolonged contraction  
24 phase. Strong volcanic forcing therefore emerges as a potential source of inter-  
25 hemispheric interannual-to-decadal climate variability, although the inter-hemispheric  
26 signature is weak in the case of historical-size eruptions. The post-eruption inter-  
27 hemispheric decadal asymmetry in sea ice is interpreted as a consequence mainly of  
28 different exposure of Arctic and Antarctic regional climates to induced meridional heat  
29 transport changes and of dominating local feedbacks that set in within the Antarctic  
30 region. “Supervolcano” experiments help clarifying differences in simulated hemispheric  
31 internal dynamics related to imposed negative net radiative imbalances, including the  
32 relative importance of the thermal and dynamical components of the sea-ice response.  
33 “Supervolcano” experiments could therefore serve the assessment of climate models’  
34 behavior under strong external forcing conditions and, consequently, favor advancements  
35 in our understanding of simulated sea-ice dynamics.

36

37 **1. Introduction**

38 Polar regional climates are in the focus of Earth system investigations owing to their  
39 strong sensitivity to external forcing and associated implications for the global climate.  
40 The so-called “polar amplification” of climate signals is mainly a consequence of positive  
41 feedbacks involving snow cover and sea ice, and it emerges more robustly in the  
42 Northern than in the Southern Hemisphere [e.g., *Parkinson*, 2004]. The different behavior  
43 of Arctic and Antarctic sea ice is largely explained by the different geographical  
44 characteristics of the two polar regions: The semi-enclosed Arctic Ocean limits sea-ice  
45 mobility and favors sea-ice thickening and persistence while making Arctic sea ice  
46 strongly susceptible to changes in the Atlantic Ocean’s northward heat transport and to  
47 anomalous atmospheric heat inflows from the surrounding landmasses. Antarctic sea-ice,  
48 by contrast, forms around the Antarctica landmass in the open Southern Ocean, its  
49 northern boundary being set by the circumpolar system of southern mid-latitude westerly  
50 winds and ocean currents. This system makes Antarctic sea ice strongly subject to  
51 equatorward drifting and melting - which explains its weak persistence - while limiting  
52 its exposure to global changes and associated anomalous atmospheric and oceanic  
53 meridional heat flows [e.g., *Zhang*, 2007, 2013]. Still, important processes driving this  
54 critical component of the Earth system remain unresolved and, hence, not robustly  
55 simulated by coupled global circulation models and Earth system models [e.g., *Maksym*  
56 | *et al.*, 2012; *Turner et al.*, [2013](#)]. Aiming at a better understanding of simulated global  
57 sea-ice behavior and of its sensitivity to external forcing, this study investigates the  
58 decadal evolution of Arctic and Antarctic sea ice in a set of idealized volcanically-forced  
59 experiments conducted with a full-complexity Earth system model. Focus is on inter-  
60 hemispheric differences in the sea-ice response.

Deleted: 2012

61 Observations covering the past three decades point to an inter-hemispheric  
62 asymmetry in recent sea-ice cover evolution: While the decline in Arctic total sea-ice  
63 cover is among the most notable features related to present climate change [e.g., *Notz and*  
64 *Marotzke*, 2012; *Stroeve et al.*, 2012; *Wang and Overland*, 2012], the Antarctic total sea-  
65 ice cover has remained steady, or even increased slightly [*Stammerjohn et al.*, 2012;  
66 | *Massonnet et al.*, 2013]. [The Antarctic sea-ice increase has been largest in autumn, with a](#)



67 dipole of a regionally significant positive trend in the Ross Sea and a negative trend in the  
68 Amundsen-Bellinghousen Sea [Turner et al., 2009]. Despite generally improved  
69 representations of observed sea-ice climatology and evolution [Stroeve et al., 2012],  
70 state-of-the-art coupled climate models fail to reproduce the observed increase in  
71 Antarctic total sea-ice cover over the last 30 years, indicating that the underlying  
72 processes are not yet simulated correctly [Turner et al., 2013]. Therefore, understanding  
73 the behavior of Antarctic sea ice and improving its representation in climate models has  
74 high priority for the aim of correctly reproducing the observed Arctic/Antarctic sea-ice  
75 dichotomy [King, 2014].

Formatted: Font: Italic

Deleted: reproduction

Deleted: of

Formatted: Font: Italic

Deleted: remains a challenge for state-of-the-art coupled climate models [Maksym et al., 2012]

Deleted: ; Turner et al., 2012].

Formatted: Font: Not Italic

Deleted: spreads

Deleted: simulated

Deleted: [Stroeve et al., 2012] as well as

Deleted: sea ice

Deleted: much more

Deleted: what

Deleted: suggest. This provides a possible explanation to the

Deleted: However

76 Internal climate variability in historical climate simulations contributes  
77 substantially to both Arctic and Antarctic sea-ice variability [Stroeve et al., 2012; Polvani  
78 and Smith, 2013]. As a consequence, simulated trends in Arctic sea ice over the last ~ 30  
79 years are generally smaller than suggested by satellite-derived sea-ice products [Stroeve  
80 et al., 2012], while simulated trends in Antarctic sea ice are characterized by large inter-  
81 model differences [Polvani and Smith, 2013]. Therefore, no conclusive assessment is  
82 available about whether the observed sea-ice asymmetry reflects a characteristic (either  
83 internally-generated or externally-forced) inter-hemispheric mode of polar climate  
84 variability or, alternatively, an extraordinary externally-forced feature.

85 Hinting towards the first hypothesis, a multicentennial control climate simulation  
86 features interdecadal periods characterized by positive trends in Antarctic sea-ice cover  
87 comparable to that observed during the last ~ 30 years [Turner et al., 2009]. The 20<sup>th</sup>

Formatted: Font: Italic

Deleted: t

88 century experienced several decades of inter-hemispheric contrast in the temperature  
89 trend [e.g., Brohan et al., 2006; Duncan et al., 2010; Chylek et al., 2010]. Inter-  
90 hemispheric out-of-phase multidecadal temperature fluctuations also emerge from  
91 reconstructed regional and continental-scale temperature variability during the last  
92 millennium and beyond [Duncan et al., 2010; Ahmed et al., 2013]. Paleoclimatic records  
93 for the last glacial maximum similarly indicate that heterogeneity and non-synchronic  
94 behavior of polar ice sheets and glacier behavior is a characteristic feature of millennial-  
95 scale climate variability [Schaefer et al., 2009; Weber et al., 2011; Shakun et al., 2012].  
96 The core processes implicated in these low-frequency inter-hemispheric climate

97 fluctuations may be similarly important for sub-centennial Arctic/Antarctic climate  
98 variability.

99 The hypothesis of an externally-forced inter-hemispheric asynchronism implies  
100 the existence of regional forcing agents and/or of response mechanisms to global forcing  
101 agents that are capable to drive a (multi)decadal inter-hemispheric climate offset.

102 Stratospheric ozone depletion in the Southern Hemisphere is among the regional factors  
103 capable of affecting Antarctic sea ice, especially so through tendential changes induced in  
104 the large-scale tropospheric circulation in the Southern Hemisphere [Gillett and  
105 Thompson, 2003; Turner et al., 2009]. Coupled climate simulations including time-  
106 varying stratospheric ozone, however, do not support a causal relationship between  
107 stratospheric ozone depletion and increased Antarctic sea ice [Sigmond and Fyfe, 2014].

108 Strong volcanic eruptions are a likely candidate for a natural forcing agent that  
109 acts globally and yet causes pronounced differences in the inter-hemispheric response:

110 For the Arctic, climate simulations indicate explosive volcanism as a major source of  
111 near-decadal [Stenchikov et al., 2009; Segschneider et al., 2012; Zanchettin et al., 2012,  
112 2013a] and multidecadal-to-centennial [Zhong et al., 2011] fluctuations in the total sea-  
113 ice area. A volcanically-forced Arctic sea-ice expansion has been suggested to be pivotal  
114 for the onset and sustenance of the Little Ice Age [Miller et al., 2012; Schleussner and  
115 Feulner, 2012], the prolonged widespread cold period spanning the 15<sup>th</sup>-18<sup>th</sup> centuries.  
116 The same period features, however, a pronounced reduction of late-summer Arctic total  
117 sea-ice extent in a recent millennial reconstruction [Kinnard et al., 2011], a  
118 counterintuitive behavior that highlights the complexity of the dynamical processes  
119 behind low-frequency variability of sea ice and our still limited knowledge about the  
120 climate state and the mechanism(s) behind specific anomalous episodes [e.g., Zanchettin  
121 et al., 2013a].

122 The scientific literature lacks studies about the susceptibility of Antarctic sea ice  
123 to volcanic forcing. There are no sufficiently-resolved reconstructions of Antarctic sea ice  
124 to assess anomalies during periods of strong volcanism before the satellite era, or they  
125 lack context as, for instance, the so-far punctual estimate of Antarctic sea-ice extent of  
126 September 1964 [Meier et al., 2013] during the aftermath of the 1963 eruption of Mount  
127 Agung. Diagnosed dynamical atmospheric responses to the strongest 20<sup>th</sup> century

Deleted: naturally

Formatted: Font: Italic

Formatted: Font: Italic

Formatted: Font: Italic

Deleted: Strong

128 eruptions are not robust in the Southern Hemisphere in observations and especially in  
129 simulations [e.g., *Robock et al.*, 2007; *Karpechko et al.*, 2010; *Charlton-Perez et al.*,  
130 2013]. Generalizing assessments based on the 20<sup>th</sup> century eruptions is prevented by the  
131 paucity and limited magnitude of the considered events, and by their concomitance with  
132 known potential disturbances to the post-eruption Antarctic climate evolution. Such  
133 disturbances include internal (e.g., a large warm event of the El Niño—Southern  
134 Oscillation or ENSO) and external ones. The latter would include, e.g., a period of weak  
135 solar activity [*Barlyaeva et al.*, 2009] and the ozone hole [*Bitz and Polvani*, 2012]. In  
136 fact, Antarctic sea ice expands considerably in the aftermath of a “supervolcano” eruption  
137 simulated by a coupled climate model [*Jones et al.*, 2005]. Confronting the  
138 “supervolcano” response with the lack of a clear response to 20<sup>th</sup> century eruptions poses  
139 the question of whether a possible southern-hemispheric dynamical response remains  
140 elusive due to a low signal-to-noise ratio for historical-size eruptions.

141 In this study, we assess the simulated inter-hemispheric sea-ice response to  
142 idealized volcanic perturbations by pursuing the following strategy: (i) investigating  
143 ensembles of Earth-system-model simulations that are sufficiently populated to yield a  
144 robust estimate of the expected forced response; (ii) comparing ensemble-average  
145 simulated responses induced by volcanic perturbations of different magnitude, ranging  
146 from that of a 1991 Pinatubo-size eruption to those of “supervolcano”-size eruptions. By  
147 including the latter we explore responses to forcing amplitudes pushing the simulated  
148 climate to its extremes. In the past, comprehensive assessments of climate responses  
149 under idealized external forcings as those used here have been proven valuable to  
150 understand simulated climate features and mechanisms and, consequently, to delimit the  
151 validity of model-based inferences about climate dynamics and variability, as well as to  
152 compare the performance of different climate models [e.g., *Stouffer et al.*, 2006].  
153 Accordingly, we focus on the inter-hemispheric asymmetry in simulated sea-ice behavior,  
154 but we also discuss possible limitations in the realism of the simulated sea-ice behavior in  
155 the light of the ocean/atmosphere/sea-ice coupled dynamics inferred from the analysis of  
156 the simulation ensemble. Our assessment therefore delineates how deficiencies in, e.g.,  
157 the representation of the Southern Ocean [e.g., *Russell et al.*, 2006; *Weijer et al.*, 2012;  
158 *Heuzé et al.*, 2013; *Salleé et al.*, 2013], of Southern Hemisphere’s atmospheric circulation

Formatted: Font: Italic

Formatted: Font: Italic

Formatted: Font: Italic

159 | [*Simpson et al.*, 2012; *Stössel et al.*, 2011] and of sea-ice processes relevant for the  
160 | Antarctic [*Landrum et al.*, 2012; *Maksym et al.*, 2012] may reverberate on simulated  
161 | transient global climate variability.

Formatted: Font: Italic

Formatted: Font: Italic

162 | We proceed as follows. First, in Section 2 we detail the experimental design of  
163 | this study, including the Earth system model, the simulations and the diagnostic tools. In  
164 | Section 3 we present the characteristics of the simulated climate responses to the imposed  
165 | forcing, focusing on post-eruption fluctuations in Arctic and Antarctic sea ice and  
166 | highlighting inter-hemispheric differences in the (forced) post-eruption signals and  
167 | associated dynamics. We discuss our results in Section 4 and provide conclusive remarks  
168 | in Section 5.

169 |

## 170 | 2. Data and methods

171 | We use the Max Planck Institute-Earth system model (MPI-ESM) in its COSMOS-Mill  
172 | version. The name of this version reflects the fact that the Community Earth System  
173 | Modeling (COSMOS) community used it for its Millennium Experiment, as described by  
174 | *Jungclaus et al.* [2010.], who provide a detailed description of the model setup. MPI-  
175 | ESM-COSMOS-Mill is based on the atmospheric general circulation model ECHAM5  
176 | [*Roeckner et al.*, 2006] coupled with the ocean model MPIOM [*Marsland et al.*, 2003;  
177 | *Jungclaus et al.*, 2006] via the OASIS3 coupler. Modules for terrestrial biosphere  
178 | (JSBACH, see: *Raddatz et al.*, 2007) and for ocean biogeochemistry (HAMOCC, see:  
179 | *Wetzel et al.*, 2005) allow for an interactive representation of the carbon cycle. The  
180 | ECHAM5/MPIOM coupled general circulation model participated to the Coupled Model  
181 | Intercomparison Project 3 (CMIP3), and has been extensively evaluated in that context.  
182 | *Jungclaus et al.* (2006) describe the general climatological oceanic features of the  
183 | ECHAM5/MPIOM included Arctic and Antarctic sea ice: simulated sea-ice  
184 | concentrations generally compare well to the observations in both hemispheres; seasonal  
185 | variation and mean distribution of Antarctic sea-ice concentrations are overall  
186 | satisfactorily simulated, though the model tends to underestimate winter sea-ice  
187 | concentration in the Weddell Sea and the Ross Sea. *Koldunov et al.* (2010) provide a  
188 | detailed evaluation of Arctic sea-ice variability simulated by ECHAM5/MPIOM  
189 | compared against late 20<sup>th</sup> century observations.

Deleted: [*Jungclaus et al.*, 2010]

Formatted: Font: Italic

Formatted: Font: Italic

Formatted: Superscript

190 ECHAM5 is run in its T31L19 configuration, corresponding to a spatial resolution  
191 of 3.75° x 3.75° and 19 vertical levels with the highest one (i.e., the model top) set at 10  
192 hPa. The model's low top restrict the description of stratospheric and coupled  
193 stratosphere-troposphere dynamics [e.g., *Omrani et al.*, 2013], which may affect the  
194 dynamical atmospheric response to volcanic forcing [e.g., *Charlton-Perez et al.*, 2013].  
195 MPIOM is run in its standard configuration GR30L40, corresponding to a horizontal  
196 grid-spacing of about 3.0° and 40 vertical levels. It embeds a dynamic-thermodynamic  
197 Hibler-type sea-ice model. A detailed description of the treatment of sub grid-scale  
198 mixing and of the sea-ice dynamics and thermodynamics implemented in MPIOM is  
199 provided by *Marsland et al.* [2003].

200 A number of studies have evaluated the climate and its variability as simulated by  
201 MPI-ESM-COSMOS-Mill against observations, proxy-based reconstructions and within  
202 a multi-model framework [e.g., *Henriksson et al.*, 2012; *Beitsch et al.*, 2013; *Bothe et al.*,  
203 2013; *Fernández-Donado et al.*, 2013; *Schubert et al.*, 2013; *Zanchettin et al.*, 2012,  
204 2013a,b]. In particular, *Beitsch et al.* [2013] showed that MPI-ESM-COSMOS-Mill  
205 spontaneously generates positive decadal-scale temperature anomalies in the Arctic  
206 region that are compatible with the observed episode known as the “early twentieth  
207 century warming”. *Tietsche et al.* [2011] explored recovery mechanisms of Arctic  
208 summer sea ice through perturbation experiments conducted with ECHAM5/MPIOM in  
209 the same configuration as MPI-ESM-COSMOS-Mill. *Li et al.* [2013] used idealized  
210 global warming simulations performed with ECHAM5/MPIOM in the same  
211 configuration to explore the long-term stability of Arctic and Antarctic sea ice against  
212 slow changes in atmospheric CO<sub>2</sub> concentration. *Notz and Marotzke* [2012] showed that  
213 the internal variability of Arctic sea-ice coverage as simulated by ECHAM5/MPIOM  
214 agrees favorably with the observed internal variability. Climatological characteristics of  
215 sea-ice concentration simulated by MPI-ESM-COSMOS-Mill are provided in the  
216 supplement (Figures S1 and S2). They agree well with those described by *Jungclaus et*  
217 *al.* [2006] for ECHAM5/MPIOM.

218 Four simulation ensembles are considered describing the climatic effects of  
219 idealized volcanic perturbations of different magnitude, up to “supervolcano”-size  
220 eruptions. The four ensembles consist of (i) ten simulations forced by a 1991 Pinatubo-

**Deleted:** Modules for terrestrial biosphere (JSBACH, see: *Raddatz et al.*, 2007) and for ocean biogeochemistry (HAMOC, see: *Wetzel et al.*, 2005) allow for an interactive representation of the carbon cycle.

**Formatted:** Font: Italic

**Formatted:** Font: Italic

**Formatted:** Font: Italic

**Formatted:** Font: Italic

**Formatted:** Font: Italic

**Formatted:** Font: Italic

**Formatted:** Font: Italic

**Formatted:** Font: Italic

**Formatted:** Font: Italic

**Formatted:** Font: Italic

**Formatted:** Font: 12 pt, Italic

**Formatted:** Font: 12 pt, Subscript

**Formatted:** Font: Italic

221 like tropical eruption, (ii) ten simulations forced by a 1815 Tambora-like tropical  
222 eruption, (iii) five simulations forced by a Young Toba Tuff (Toba)-like eruption, i.e., a  
223 tropical eruption with 100-times the emission strength of the Pinatubo eruption, and (iv)  
224 ten simulations forced by a Yellowstone-like eruption (i.e., same as Toba, but located in  
225 the Northern Hemisphere’s mid-latitudes). In the following, we refer to the ensembles  
226 avoiding the volcanoes’ specific names to highlight their idealized character. We  
227 thereafter refer to the Pinatubo and Tambora simulations/eruptions as “historical”  
228 (namely HIST1 and HIST2, respectively), since these eruptions are representative of the  
229 magnitude of volcanic eruptions that occurred during the last millennium. The Toba and  
230 Yellowstone “supervolcano” ensembles are referred to as SUPER1 and SUPER2,  
231 | respectively. HIST2 corresponds to the VO2 ensemble in *Zanchettin et al.* [2013a].  
232 | SUPER1 entails the simulations used in *Timmreck et al.* [2010, 2012]. SUPER2  
233 | simulations are those described in *Segschneider et al.* [2012]. Each ensemble consists of  
234 | simulations differing only in their initial climate states, which are sampled from a multi-  
235 | millennial pre-industrial control simulation [as used in *Timmreck et al.*, 2010, and  
236 | *Zanchettin et al.*, 2013a]. In HIST1, SUPER1 and SUPER2 the eruptions start in June of  
237 | the first integration year. In HIST2 the eruption starts in April, according to historical  
238 | reconstructions of the 1815 Tambora event (*Crowley et al.* [2008], *Crowley and*  
239 | *Untermann* [2012]). HIST2 simulations include the eruptions reconstructed for the  
240 | subsequent decades, e.g., the Cosiguina eruption in the early 1830s [*Zanchettin et al.*,  
241 | 2013a].

242 | Volcanic forcing implemented in MPI-ESM is based on zonally-averaged time  
243 | series of aerosol optical depth (AOD) at 0.55  $\mu\text{m}$  and of effective particle radius ( $R_{\text{eff}}$ ).  
244 | For HIST2, we use the reconstructed 10-day average values of AOD and  $R_{\text{eff}}$  by *Crowley*  
245 | *et al.* [2008]. Data are provided for four equal-area latitudinal bands (90°S-30°S, 30°S-0°,  
246 | 0°-30°N and 30°N-90°N). Volcanic aerosols are vertically distributed between 30 and 70  
247 | hPa, with a maximum at 50 hPa [*Timmreck et al.*, 2009]. For the other ensembles, AOD  
248 | and  $R_{\text{eff}}$  are estimated following the two-step approach described by *Timmreck et al.*  
249 | [2010]. Briefly summarizing it: in the first step the formation of volcanic sulfate aerosols  
250 | is calculated from an initial volcanic sulfur injection in the stratosphere by the middle  
251 | atmosphere version of the aerosol climate model ECHAM/HAM [*Stier et al.*, 2005;

252 *Niemeier et al.*, 2009]; in the second step, the zonally-averaged monthly time series of the  
253 so calculated AOD and  $R_{\text{eff}}$  values are used as external forcing in MPI-ESM. Due to  
254 subsequent temporal interpolation of input data by ECHAM5, in MPI-ESM the eruption  
255 is tailed in the month preceding its occurrence in ECHAM/HAM.

256 In all simulations, the time-dependent AOD and  $R_{\text{eff}}$  values are used to calculate  
257 online the optical parameters of the ECHAM5 radiation scheme, including extinction,  
258 single-scattering albedo and asymmetry factor for the six solar bands (0.185-4  $\mu\text{m}$ ), and  
259 extinction for the 16 long-wave wavelength bands (3.3-100  $\mu\text{m}$ ). Aerosol sizes are  
260 assumed to be distributed with a constant standard deviation of 1.8  $\mu\text{m}$ .

261 The global-average air surface temperature drop after the 1991 Pinatubo eruption  
262 in sensitivity experiments conducted with MPI-ESM is comparable with observations  
263 [*Timmreck et al.*, 2009]. Global and hemispheric near-surface air temperature changes in  
264 a full-forcing COSMOS-Mill simulation ensemble around the Tambora eruption  
265 employing the same volcanic forcing input as the HIST2 ensemble are compatible with  
266 estimates from observations and reconstructions [*Zanchettin et al.*, 2013a].

267 Responses are diagnosed through analysis of ensemble-averages. For time series,  
268 we use deseasonalized and then low-pass filtered values. Seasonality is computed based  
269 on control-run data and then subtracted from all data. Filtering consists of 3-month  
270 centered running-mean for atmospheric variables, and 13-month centered running-mean  
271 for oceanic and sea-ice variables, unless specified otherwise. Anomalies are evaluated as  
272 deviations from the pre-eruption climatology, defined as the mean climate state during  
273 the ten years/winters/summers preceding the eruption. Post-eruption years are  
274 progressively numbered starting from the year of the eruption, which is defined as year 0.

275 A Monte-Carlo approach is used to estimate the statistical significance of the  
276 forced signals [e.g., *Graf and Zanchettin*, 2012]. Specifically, the ensemble-average  
277 signals obtained from an ensemble of  $n$  forced simulations are compared with a large set  
278 of analog ensemble-average signals (here 500) obtained by randomly sampling  $n$  years  
279 along the whole length of the control run. The empirical distribution yielded by these  
280 analog ensemble-average signals describes probabilistically the range explicable by  
281 internal variability alone, which we also interpret as the confidence level of  
282 corresponding signals in the forced ensembles having occurred by chance. We consider

283 as reference the 98% range (i.e., 1<sup>st</sup>-99<sup>th</sup> percentile band) of such distribution in order to  
 284 have a conservative estimate of internal variability signals. Since the procedure is only  
 285 based on the random selection of years, the autocorrelation is preserved in the estimation  
 286 of the significance.

287 Total sea-ice area in the Arctic and in the Antarctic is calculated as the areal sum  
 288 of sea ice covering the ocean in the Northern and Southern Hemispheres, respectively.  
 289 Analogously, total sea-ice volume is defined as the sum of local (i.e., grid-point) products  
 290 of grid-cell area and grid-cell-average sea-ice thickness. The sea-ice edge is defined as  
 291 the line denoting the sea-ice extent margin, i.e., the area enclosing sea-ice concentrations  
 292 exceeding the 0.15 threshold (in the range [0:1], where 0 indicates no sea ice in the grid-  
 293 cell and 1 indicates sea ice fully covering the grid-cell).

294 Meridional ocean heat transports  $HT$  are calculated at 60°N and 60°S as in  
 295 *Zanchettin et al.* [2012] based on the equation:  $HT = \sum_z \sum_x v T c_p \rho dx dz$ , where  $v$  is the  
 296 meridional velocity component,  $T$  is temperature,  $c_p$  is specific heat capacity at constant  
 297 pressure,  $\rho$  is density, and  $dz$  and  $dx$  represent, respectively, the integrals along depths  
 298 and longitudes. The zonal mean component of the total meridional ocean heat transports  
 299 is considered to be associated with the overturning transport; the residual component is  
 300 considered to describe the gyre contribution to the total meridional ocean heat transport.  
 301 Accordingly, deviations of  $v$  and  $T$  from the respective zonal mean values are used in the  
 302 above mentioned equation for the calculation of the gyre contribution to  $HT$ .

303 The meridional atmospheric energy transport around 60°N and 60°S are defined,  
 304 following *Keith* [1995], as the zonal integral at, respectively, 61.23°N and 61.23°S of the  
 305 convergence of the atmospheric energy transport vector  $\mathbf{F}_A$ , which can be written for  
 306 latitude  $\lambda$  as:

$$307 \quad F_\lambda = \int_\lambda -\nabla * \mathbf{F}_A dx = \int_\lambda -\nabla * \frac{1}{g} \int_0^{p_s} (c_p T + \Phi + Lq + k) \mathbf{v} dp dx \quad (1)$$

308 where  $dp$  and  $dx$  represent, respectively, the integral along pressure levels and longitudes,  
 309  $c_p T + \Phi$  represents the dry static energy, with the specific heat of the atmosphere at  
 310 constant pressure  $c_p$ , temperature  $T$ , and geopotential  $\Phi$ . The moist static energy is  
 311 depicted by  $c_p T + \Phi + Lq$ , with latent heat of evaporation/condensation  $L$ , and specific  
 312 humidity  $q$ . The horizontal wind vector is represented by  $\mathbf{v}$ . As kinetic energy  $k$  is



313 typically a small component of the energy budget, it is ignored in the following. The  
314 small latitudinal difference between atmospheric and oceanic heat transport calculations  
315 is of negligible concern, since we do not aim to close the energy budget for the two  
316 regions.

317

### 318 **3. Results**

#### 319 **3.1 Imposed forcing and global/hemispheric responses**

320 The imposed forcing is very well constrained within each of the four ensembles (Figure  
321 1). Estimates based on top-of-atmosphere radiative anomalies for individual ensembles  
322 are consistent with previously reported ones [Timmreck *et al.*, 2010; Segsneider *et al.*,  
323 | 2012; Zanchettin *et al.*, 2013a]; we therefore describe only major features and inter-  
324 ensemble differences. Especially during the first three post-eruption years, ensemble  
325 standard errors are barely distinguishable from the corresponding ensemble-average  
326 values. Peak negative anomalies in the global top-of-atmosphere net radiative flux range  
327 between  $\sim -3 \text{ Wm}^{-2}$  for HIST1 and  $\sim -27 \text{ Wm}^{-2}$  for SUPER2 (Figure 1c). SUPER2 leads  
328 to a slightly stronger forcing than SUPER1 in the net radiative flux estimate (Figure 1c).  
329 This highlights the dependence of the net forcing on the shape of post-eruption evolutions  
330 of the shortwave and longwave radiation flux anomalies, since these have otherwise  
331 similar peak values in the two ensembles (Figure 1a,b). The evolution of radiative fluxes  
332 is directly linked to the evolution of the volcanic aerosol mass, which builds up slower  
333 during the first post-eruption months in SUPER2 compared to SUPER1 (not shown).  
334 This seems to be the key to understanding the differences between the two  
335 “supervolcano” ensembles and the distinguishing traits of the former. The positive net  
336 flux anomaly around lags of 42 to 78 months is mainly a consequence of the ocean  
337 releasing less latent heat to the atmosphere [Timmreck *et al.*, 2010; Zanchettin *et al.*,  
338 | 2013a].

339 On the global scale, the four ensembles depict significant post-eruption drops in  
340 surface (2 meters) air temperature and precipitation (Figure 2a,b). Cold temperature  
341 anomalies consistently peak in the boreal summer-autumn of year 1, i.e. slightly after the  
342 peak in the forcing (compare with Figure 1c), with larger ensemble spread for the  
343 historical eruptions (note, SUPER1 consists of only five simulations). HIST1 displays a

344 temporary initial recovery of the temperature signal to within the internal variability  
345 range in year 2, when typically a warm ENSO event sets in. The ensemble-mean  
346 simulated maximum cooling for HIST1 matches the observed maximum cooling of 0.4 K  
347 estimated for the Pinatubo by *Thompson et al.* [2009]. Annual oscillations in the post-  
348 eruption anomalous temperature evolution in the "supervolcano" ensembles indicate a  
349 seasonal differentiation of the response, with boreal winter semesters being  
350 comparatively colder than summer ones from year 2 onwards. The global temperature  
351 responses differ appreciably between SUPER1 and SUPER2. Inter-ensemble differences  
352 in global precipitation regard the timing of the post-eruption fluctuation, delayed in the  
353 case of SUPER2 compared to SUPER1, rather than the shape of the post-eruption  
354 fluctuation and its peak value. Post-eruption anomalies of hemispheric-average surface  
355 air temperature (Figure 2c,d) further highlight the differences between historical and  
356 "supervolcano" eruptions. For the two historical eruptions, inter-hemispheric differences  
357 are small and remain mostly confined within the internal variability range after the first  
358 two post-eruption years (Figure 2d). For the "supervolcano" ensembles, inter-  
359 hemispheric differences are large and remarkably independent of the location of the  
360 eruption (Figure 2d): the Northern Hemisphere undergoes a much stronger and longer  
361 lasting cooling compared to the historical ensembles (Figure 2c), with a more pronounced  
362 seasonal character than the Southern Hemisphere (compare Figure 2d). As a  
363 consequence, in both "supervolcano" ensembles the anomalous hemispheric temperature  
364 evolutions deviate considerably from the global estimate.

365 Overall, we diagnose qualitatively similar features in the different ensembles that  
366 point to an amplification of the forced global signals with increased magnitude of the  
367 eruption. "Supervolcano" simulations feature a high signal-to-noise ratio, and even the 5-  
368 member SUPER1 ensemble is suitable for robust global/hemispheric-scale inferences.  
369 Inter-hemispheric differences are apparent in the surface air temperature responses to  
370 "supervolcano" but not historical-size eruptions, suggesting that substantially different  
371 dynamical responses may characterize the different eruption sizes.

372

### 373 **3.2 Sea-ice response**

374 The post-eruption anomalies of Arctic and Antarctic sea-ice area and volume depict  
375 major inter-hemispheric differences in the sea-ice responses to both historical and  
376 “supervolcano” eruptions (Figure 3). In the Arctic, the total sea ice expands for all  
377 eruptions (Figure 3a,c). The post-eruption positive anomalies of total Arctic sea-ice area  
378 and volume are of comparable magnitude for the two historical eruptions, but their timing  
379 differs. Total Arctic sea-ice area and volume anomalies are about one order of magnitude  
380 larger in the “supervolcano” ensembles compared to historical ones. In both ensembles,  
381 total sea-ice area and volume entail a sharp increase in simulation years 1 and 2, which is  
382 followed by a decadal-scale progressive dampening of the anomaly. The larger anomalies  
383 in SUPER2 compared to SUPER1 are likely thermally driven: the volcanic cloud  
384 produced by the extra-tropical Yellowstone-like eruption is more confined to the  
385 Northern Hemisphere and produces a stronger radiative effect there, i.e. stronger cooling  
386 (Figure 2d). The system fully reverts back to within the internal variability range in about  
387 2-2.5 decades.

388         Significant post-eruption anomalies of Antarctic total sea-ice area and volume  
389 (Figure 3b,d) are only detected in the “supervolcano” ensembles. In these ensembles and  
390 especially concerning the total sea-ice volume, Antarctic sea-ice anomalies are much  
391 smaller than their Arctic counterparts (compare panels c and d of Figure 3). This is true  
392 for both the actual anomalies and their values relative to the pre-eruption climatology  
393 (which is reported in Figure 3). Initially, a short-lived Antarctic sea-ice area increase  
394 occurs approximately within the first two post-eruption years, which is not accompanied  
395 by a significant increase in sea-ice volume. This means that, in contrast to the Arctic sea-  
396 ice response, there is no post-eruption net build up of Antarctic sea-ice mass. As we will  
397 further discuss in section 3.3, the areal expansion likely results in good part from a  
398 dynamic response of the Southern Ocean’s sea ice, which is advected over a larger area.  
399 This initial expansion phase is followed by a rebound retraction phase of similar  
400 amplitude and longer duration (Figure 3b), which is characterized by a drastic reduction  
401 in the total sea-ice volume (Figure 3d). Note, reduction in sea-ice volume starts in year 1,  
402 when the positive anomaly in sea-ice area is near its peak, meaning that net losses in  
403 volume occurs already during the horizontally expanded phase of Antarctic sea ice. In  
404 other words, Antarctic sea ice covers a larger area while thinning. This second phase

405 consistently ends about eight years after the eruption. Negative anomalies of both sea-ice  
406 area and volume are larger for SUPER1, whose ensemble-spread nonetheless overlaps  
407 with that of SUPER2 during the full duration of the rebound fluctuation.

408 Generally, ensemble-spreads are larger in the Antarctic sea-ice area estimates than  
409 in their Arctic counterparts. This is true also for the spread in sea-ice volume in its  
410 relative estimates, but not in its absolute values due to smaller Antarctic climatology  
411 (Figure 3c,d). Overall, the post-eruption sea-ice evolution appears to be characterized by  
412 two distinct phases in the “supervolcano” ensembles: (i) an initial phase of tendential  
413 synchronic bi-polar expansion during integration years 1 and 2, and (ii) a subsequent,  
414 prolonged phase of inter-hemispheric asymmetry during integration years 4-6.

415 The anomalies determining the two detected phases of post-eruption sea-ice  
416 evolution feature a prominent seasonal character (Figure 4). In the historical ensembles,  
417 the significant signals detected in the deseasonalized and smoothed series of total Arctic  
418 sea-ice area (Figure 3a) originate from a significant increase during the boreal summer  
419 season (Figure 4a). In the “supervolcano” ensembles the initial post-eruption increase in  
420 total Arctic sea-ice area occurs throughout the whole year but the magnitude of  
421 departures from the climatology is more than doubled in the boreal summer compared to  
422 the boreal winter. As we will show in section 3.3, this behavior is most likely due to  
423 reduced melting, i.e., thermodynamics is very important for the initial response of Arctic  
424 sea ice to volcanic forcing. Predominance of reduced summer melting on winter growth  
425 is smeared out in the Arctic delayed response, and the annual cycle averaged over years  
426 4-6 essentially corresponds to an upward-shifted unperturbed annual cycle (Figure 4c).  
427 By contrast, the signals in total Antarctic sea-ice area are largest in the sea-ice growth  
428 season (Figure 4b), with initial post-eruption gains peaking at  $\sim 1.9\text{-}2.4$  million  $\text{km}^2$  in  
429 June-July and following losses peaking at  $\sim 1.8\text{-}2$  million  $\text{km}^2$  in August-October (Figure  
430 4d).

431 Figures 5 and 6 illustrate the regional distribution of sea-ice concentration  
432 anomalies for, respectively, the Arctic and the Antarctic region during the two detected  
433 phases of post-eruption sea-ice evolution for the SUPER1 ensemble. Mapped values refer  
434 to monthly anomalies at the end of the growing season (i.e., March for Arctic sea ice,  
435 September for Antarctic sea ice) and of the melting season (i.e., September for Arctic sea

436 ice, March for Antarctic sea ice). Immediately after the eruption, the March Arctic sea-ice  
437 concentration increases especially in the gulf of Alaska/eastern Bering Sea and in the  
438 outer Labrador Sea/western North Atlantic, where the sea-ice edge significantly advances  
439 (Figure 5a). Widespread reduced melting results in extensive increases in September  
440 Arctic sea-ice concentrations. These are particularly large in the Canadian Arctic  
441 Archipelago and in the Baffin Bay, where the sea-ice edge advances as far as reaching the  
442 Hudson and Davis Straits, along the East Greenland current, and in the Barents and Kara  
443 Seas, with the latter basin being fully sea-ice covered (Figure 5c). The same regions are  
444 important for the total Arctic sea-ice area anomaly in the second phase of the post-  
445 eruption areal evolution of sea ice. Then, the strongest contribution to the winter anomaly  
446 of Arctic total sea-ice area comes from the North Atlantic/Nordic Seas sector, where  
447 March sea-ice concentrations increase by as much as 60% (Figure 5b). September Arctic  
448 sea-ice concentration anomalies are also still significant over extensive regions, but with  
449 overall smaller amplitudes (Figure 5d).

450 In the Antarctic, total sea-ice area anomalies are of reduced amplitude and  
451 extension in austral summer during both phases (Figure 4b,d). Immediately after the  
452 eruption, there is a circumpolar tendency towards positive March anomalies of sea-ice  
453 concentration, though these are strongest and most extensive off the West Antarctic coast,  
454 where they result in a local advance of the sea-ice edge (Figure 6a). Later on, the same  
455 region faces a marked reduction of March sea-ice concentrations and a consequent retreat  
456 of the sea-ice edge (Figure 6b), which is again part of a general circumpolar tendency.  
457 The regional details of September anomalies of Antarctic sea-ice concentration during the  
458 two phases provide a more complex picture (Figure 6c,d). In both phases, negative sea-  
459 ice concentration anomalies are diagnosed off the East Antarctic coasts and in the outer  
460 Weddell Sea. In the latter region, the response has the typical traits of an open-ocean  
461 polynya, i.e. an ice-free area within the ice cover, and is surrounded by positive  
462 anomalies leading to a locally advancing sea-ice edge during the initial post-eruption  
463 phase (Figure 6c). Anomalies spatially extend more widely in the second phase, when a  
464 general retreat of the sea-ice edge is diagnosed (Figure 6d). Whereas no significant large-  
465 scale changes are detected west of the Antarctic Peninsula in the initial phase (Figure 6c),  
466 the same region faces later a reduction in sea-ice concentration which is locally as large

467 as 60% and results in a strong retreat of the sea-ice edge (Figure 6d). Whereas both  
468 phases indicate reductions in September sea-ice concentrations in the outer Ross Sea, the  
469 initial post-eruption phase entails also an extensive increase along 60°S (Figure 6c).

470 Differences between the shown SUPER1 patterns and their SUPER2 analogs (not  
471 shown) are generally minor. For historical eruptions, significant post-eruption sea-ice  
472 concentration anomalies are usually local, but generally point towards an agreement with  
473 the “supervolcano” ensembles concerning the tendencies in the key regions (not shown).

474 In summary, the sea-ice response to volcanic eruptions in MPI-ESM-COSMOS-  
475 Mill strongly depends on the amplitude of the induced global perturbation and, to a lesser  
476 extent, the location of the eruption (compare, e.g., the Arctic sea-ice response to the  
477 SUPER1/tropical and SUPER2/mid-latitude eruptions). All ensembles feature a  
478 temporary post-eruption increase in Arctic sea-ice, while no robust signature on Antarctic  
479 sea ice characterizes historical-size eruptions. The post-eruption sea-ice evolution in  
480 “supervolcano” simulations can be clearly separated into two phases: an initial one of bi-  
481 polar expansion and a delayed one marked by the contrast between persisting Arctic  
482 expansion and strong Antarctic contraction. The latter constitutes a counterintuitive  
483 simulated behavior, whose explanation seemingly lies in the anomalous seasonal  
484 behavior in a few key regions. This is explored further in the next section.

485

### 486 *3.3 Mechanism of Arctic and Antarctic sea-ice response to a “supervolcano” eruption*

487 In this section, we focus on the mechanism(s) behind the sea-ice response to the SUPER1  
488 eruption during the initial bi-polar synchronic phase and the subsequent inter-hemispheric  
489 asymmetric phase. The SUPER1 ensemble is chosen among the two “supervolcano”  
490 eruptions since previous studies on the same ensemble [Timmreck *et al.*, 2010; 2012]  
491 provide context to our inferences.

492 Immediately after the eruption, the meridional air temperature gradient  
493 temporarily increases in the lower stratosphere (supplementary Figure S3) due to the in-  
494 situ heating by the volcanic aerosols. As a consequence [see, e.g., Timmreck, 2012], the  
495 stratospheric polar vortex strengthens in both hemispheres until the volcanic cloud  
496 dissipates, i.e., for the first two post-eruption years (Figure S4). The tropospheric  
497 response is dominated by significant cooling, which persists especially in the Northern

Formatted: Font: 12 pt, Italic

498 Hemisphere winter (Figure S3), and by weakening of both the Hadley and Ferrell cells  
499 (Figures S5-S6), which is consistent with a slow-down of the global hydrological cycle  
500 (Figure S7, also compare Figure 2b). The weakening of the general circulation is further  
501 associated with weakened tropical and mid-latitude zonal flow in both hemispheres  
502 (Figure 7b). This is also concomitant with short-lived anomalous eastward polar  
503 circulations, which we interpret as part of the downward propagation of the volcanically-  
504 forced strengthened stratospheric polar vortices. Zonal-mean meridional winds at their  
505 climatological hemispheric maxima around 30°N and 50°S also depict a significant  
506 reduction of the zonal-mean northward flow in years 1-2 in the Northern Hemisphere  
507 (Figure 7c).

508 Later on, anomalies in the general atmospheric circulation become less  
509 pronounced and are only locally significant, though weaker-than-normal jet conditions  
510 remain apparent (Figure S4). So, internal atmospheric processes strongly contribute to the  
511 initial response, whereas climatic signals on the decadal scale are mostly related to  
512 oceanic and ocean-atmosphere coupled processes. This is for example the case for the  
513 significant though rather small increase of the zonal-mean southward flow in years 5-6  
514 detected in the Southern Hemisphere (Figure 7c). The structure of vertical profiles of  
515 zonal-mean anomalies of atmospheric parameters depicts an overall symmetry between  
516 the general atmospheric circulation of Northern and Southern hemispheres in both phases  
517 of post-eruption sea-ice response (Figures S3-S11). The persistent colder tropospheric  
518 anomalies over the Arctic, especially during the Northern Hemisphere winter, contrasting  
519 the comparatively weak and short-lived Antarctic anomalies represent the most apparent  
520 inter-hemispheric asymmetry in the post-eruption atmosphere (Figure S8).

521 Significant decadal anomalies characterize the post-eruption evolutions of zonal-  
522 mean surface temperature and its associated hemispheric meridional gradients (Figure  
523 7a). The post-eruption anomalies depict: (i) strong initial cooling, mostly related to quick  
524 responses over the landmasses; (ii) bipolar asymmetry in the form of delayed and  
525 prolonged (compared to tropical regions) cooling in the Arctic contrasting the reduced  
526 cooling and subsequent warming in the Antarctic; (iii) inter-hemispheric asymmetry in  
527 the equator-to-pole temperature gradient, in the form of a temporarily strengthened  
528 gradient in the Northern Hemisphere contrasting a prolonged weakened gradient in the

529 Southern Hemisphere; (iv) delayed (compared to both tropical and polar regions) cooling  
530 in the equatorial band, seemingly “phasing” ENSO to a La Niña state in year 3. The latter  
531 feature is associated to a temporary reduction of meridional gradients, which feature at  
532 this stage highly significant negative anomalies in the Southern Hemisphere.

533 | The anomalous atmospheric energy and oceanic heat transports into the polar  
534 regions (Figure 8) provide constraints to our causal interpretation of the diagnosed  
535 regional changes. In the Northern Hemisphere, significant (i.e., outside the internal  
536 variability range) and prolonged reductions in the meridional heat transport into the  
537 Arctic region are diagnosed for both the atmosphere and the ocean: The reduction in  
538 atmospheric heat transport peaks at around lag 24-months and remains at significant  
539 levels over a 6-year period; oceanic heat transport is reduced below the lower threshold  
540 of internal variability around lag 24-months and persists in an anomalously low state for  
541 almost one decade. The estimated dry static atmospheric energy transport remains  
542 generally within the internal variability range, with a less clear ensemble-mean evolution  
543 compared to the moist static energy transport (Figure 8a). This indicates that the latent  
544 heat component – entailing reduced global ocean losses to the atmosphere (not shown)  
545 and reduced global precipitation (Figure 2b) – dominates the response over the thermal  
546 component. At this latitude, the post-eruption anomalous oceanic heat transport is  
547 dominated by the gyre component (Figure 8b), which agrees with the general behavior  
548 | typically simulated by MPI-ESM-COSMOS-Mill [e.g., Zanchettin *et al.*, 2012, 2013a].

549 | In the Southern Hemisphere, the atmospheric energy transport into the Antarctic  
550 region is significantly reduced between about 2 and 8 years after the eruption (Figure 8c),  
551 reflecting the evolution of anomalous equator-to-pole surface temperature gradient  
552 (Figure 7a). In absolute values, the associated peak post-eruption anomaly is about half of  
553 its Arctic counterpart (compare panels a and c of Figure 8). An initial, short-lived  
554 response is diagnosed in the estimated dry static atmospheric energy transport,  
555 compatible with the surface and tropospheric cooling simulated around these latitudes  
556 (compare Figure 7a), which is evidently compensated by an increase in the atmospheric  
557 latent heat component. Oceanic heat transport into the Antarctic is characterized by  
558 strong interannual variability in its post-eruption anomalous evolution as well as by  
559 strong internal variability compared to its Arctic counterpart (compare ranges in Figure

**Deleted:** Significant changes in the large-scale mid-tropospheric circulation are confined to the first 5-6 post-eruption years. Changes in the zonal-mean zonal winds describe a general weakening of the tropical and mid-latitude circulation of both hemispheres, though persisting longer in the Southern Hemisphere (Figure 7b). This is concomitant with short-lived anomalous eastward polar circulations, which we interpret as part of the downward propagation of the volcanically-forced strengthened stratospheric polar vortices (not shown). Zonal-mean meridional winds at their climatological hemispheric maxima around 30°N and 50°S depict a significant reduction of the zonal-mean northward flow in years 1-2 in the Northern Hemisphere and a significant though rather small increase of the zonal-mean southward flow in years 5-6 in the Southern Hemisphere (Figure 7c). In the first two post-eruption years, the zonal and meridional wind response can be traced back to significant weakening of both the Hadley and Ferrell cells (not shown) consistent with a slow-down of the global hydrological cycle (compare Figure 2b). During the second phase, zonal-mean vertical velocities depict a different behavior in the two hemispheres: during boreal summer the Northern Hemisphere features, as most prominent feature, a small though significant weakening of the upper and descending branches of the Hadley cell; during austral summer, the Southern Hemisphere features an expanded polar cell (not shown) consistent with the diagnosed changes in the meridional winds (Figure 7c).¶



560 8b,d). As a consequence, despite peak anomalies about twice those diagnosed in the  
561 Arctic, the post-eruption ocean heat transport into the Antarctic remains mostly within  
562 the internal variability range. The most significant feature is a temporary reduction in the  
563 total poleward transport around lag of 48 months (Figure 8d). As for the Northern  
564 Hemisphere, at these latitudes oceanic transport is dominated by the gyre component.

565 In summary, both polar regions feature a post-eruption decrease in the energy  
566 import. However, the decrease is overall more significant for the Arctic due to an overall  
567 more constrained oceanic internal variability range and to constructively superposing and  
568 comparable contributions from the atmosphere and the ocean, the former being pivotal in  
569 the initial response phase and the latter dominating the response thereafter. For the  
570 Antarctic, both oceanic and moist atmospheric energy transports remain initially  
571 unaffected, pointing towards dynamical circulation changes as cause for the initial  
572 Antarctic sea-ice response. Furthermore, the relevance of the ocean for the post-eruption  
573 Antarctic energy budget and its attribution to the imposed forcing is hampered by its  
574 strong internal variability. It is therefore important to relate anomalous ocean meridional  
575 heat transports to dynamical changes in the oceanic circulation.

576 In the Northern Hemisphere, the general response of the oceanic circulation to the  
577 SUPER1 eruption is in line with the behavior typically simulated by MPI-ESM-  
578 COSMOS-Mill after historical-size eruptions [see, e.g., *Zanchettin et al.*, 2012, 2013a].  
579 We therefore only show changes more closely related to sea ice. The post-eruption  
580 reduction in gyre-driven northward heat transport is clearly associated with a weakening  
581 of the subpolar gyre and of the North Pacific gyre in the Kuroshio-Oyashio extension  
582 region (Figure 9a). The weak Gulf Stream together with strong anomalous ocean heat  
583 losses to the atmosphere contributes to anomalous cold conditions in the upper polar  
584 ocean (not shown). The regional cold anomaly sustains the deepening of the ocean mixed  
585 layer, which occurs largely in the western portion of the subpolar gyre and in the  
586 Irminger and especially Nordic Seas (Figure 10a). Associated processes of deep water  
587 formation result in a progressive intensification of the Atlantic Meridional Overturning  
588 Circulation (AMOC) of up to 3 Sv at 30°N and 1000 m depth, which in turn allows for a  
589 delayed temporary increase in the northward ocean heat advection in the tropical and  
590 mid-latitude band (not shown). Anomalously strong oceanic convection in the Nordic

591 Seas occurs also during the second phase of the post-eruption sea-ice response (Figure  
592 10b) due to the persisting cold anomaly (compare Figure 7a). Insulation from the  
593 meanwhile advanced sea-ice edge confines the extent of the deepening region (Figure  
594 10b), exemplifying how the sea-ice evolution is fully embedded within the general  
595 coupled atmospheric/oceanic response mechanism. We further note that at this later  
596 stage, the anomalous patterns of both surface energy fluxes and near-surface atmospheric  
597 circulation do not reveal robust large-scale features (not shown), confirming the  
598 predominant role (relative to the atmosphere) of the anomalous decadal oceanic evolution  
599 for sustaining the post-eruption Arctic sea-ice anomaly.

600 In the Southern Hemisphere, the initial Antarctic sea-ice response is a dynamical  
601 consequence of a strengthened circumpolar westerly circulation along the sea-ice  
602 margins. This feature is especially evident in austral winter (Figure 11b), when it  
603 describes a poleward shift of the mid-latitude westerlies partly superposing on a positive  
604 phase of the Southern Annular Mode. An only similar feature is noticeable in summer,  
605 when the pattern describes a weakening of both polar and mid-latitude flow (Figure 11d)  
606 typical of a negative phase of the Southern Annular Mode. The marked seasonality in the  
607 near-surface wind response goes along with the marked seasonality diagnosed in the sea-  
608 ice area response (Figures 4b and 6a,c). Our interpretation is therefore consistent with the  
609 enhanced connectivity of observed sea ice variability to the overlying atmospheric  
610 circulation associated with large-scale modes like the Southern Annular Mode and ENSO  
611 [Simpkins *et al.*, 2012]. Disentangling the different contributions to the post-eruption  
612 polar circulation in the southern-hemispheric lower troposphere would require a  
613 dedicated study. We only remark the importance of the antagonism between the thermal  
614 (i.e., meridional gradient in surface cooling, Figure 7a) and dynamical (i.e., downward  
615 propagation of strengthened stratospheric polar vortex) effects. The Antarctic  
616 Circumpolar Current (ACC) weakens (Figure 9c) in response to the weakened (in  
617 summer) and southward shifted (in winter) circumpolar mid-latitude westerly flow  
618 (Figures 7b and 11b,d). During this initial response phase, the ocean undergoes also  
619 important dynamical modifications at the regional scale. In particular, the September sea-  
620 ice concentration locally decreases in the sea-ice interior region of the outer Weddell Sea  
621 (Figure 6c). This area features significantly strengthened ocean energy losses to the

622 atmosphere (Figure 11c) that are related to a significant deepening of the mixed layer  
623 (Figure 10c), hence penetration of the post-eruption cold anomaly into the deep ocean  
624 layers. Of course, the causal chain linking these features cannot be depicted without the  
625 support of dedicated sensitivity experiments. Such response is not diagnosed within the  
626 ice-covered region of the Ross Sea despite local strengthening of ocean heat losses  
627 (Figure 11c).

628         Local feedbacks involving sea-ice area, turbulent heat fluxes and modified  
629 atmospheric circulation complete the explanation for the diagnosed sea-ice behavior  
630 during the second phase of Southern Hemisphere's sea-ice response. We note particularly  
631 the anomalous equatorward near-surface atmospheric flows off the West Antarctica coast  
632 in austral summer (Figure 12b) and in the outer Ross Sea in austral winter (Figure 12d),  
633 which are associated with local negative surface energy flux anomalies over extensive  
634 regions (Figure 12a,c). The local anomalous near-surface winds set in under significantly  
635 weakened mid-latitude westerly circulation, especially during austral summer (Figure  
636 12b). The anomalous near-surface wind pattern contrasts the zonal-average tendency in  
637 the mid-troposphere (Figure 7c) and is consistent with a net reduction of atmospheric  
638 energy import into the Antarctic region (Figure 8c). The latter should therefore be  
639 regarded as mainly a consequence of local coupled ocean-atmosphere dynamics internal  
640 to the Antarctic region. The consistency with corresponding anomalous patterns in the  
641 SUPER2 ensemble (not shown) adds support to this interpretation.

642         Enhanced deep convection still takes place in the Weddell Sea region during the  
643 second response phase (Figure 10d), though its magnitude and extent are reduced  
644 compared to the initial anomaly. Again, the causal chain for this behavior cannot be fully  
645 clarified based on our experiments alone allowing for tentative hypotheses only. We  
646 accordingly interpret the strengthened oceanic convection as a likely consequence of  
647 locally strengthened surface exchange processes (Figure 12c) favored by the meanwhile  
648 decreased winter sea-ice area (Figure 6d). We thus regard the negative sea-ice anomaly as  
649 the closure element of the feedback mechanism characterizing the post-eruption ocean-  
650 atmosphere evolution in the Weddell Sea region (i.e., the regional anomaly persists until  
651 the anomalous large-scale atmospheric circulation sustains a local sea-ice reduction).

652

#### 653 4. Summarizing discussion

654 In this study we used ensemble climate simulations performed with the COSMOS-Mill  
655 version of the Max Planck Institute-Earth system model (MPI-ESM) to investigate the  
656 decadal response of Arctic and Antarctic sea ice to volcanic perturbations. We considered  
657 volcanic eruptions of different magnitude, ranging from historical-size to  
658 “supervolcano”-size, the latter with different characteristics, including tropical and extra-  
659 tropical locations. In all ensembles a sustained, largely thermally-driven expansion is  
660 robustly simulated for total area and volume of Arctic sea ice. Amplitude and duration of  
661 the anomalies essentially depend on the magnitude of the imposed forcing. In contrast,  
662 the simulated response of Antarctic sea ice is elusive for the historical-size eruptions,  
663 while “supervolcano” eruptions induce an initial short-lived, mostly dynamically-driven  
664 Antarctic sea-ice expansion which is followed by a prolonged retraction phase. For both  
665 historical and “supervolcano” eruption-types we diagnose, therefore, an inter-hemispheric  
666 asymmetry in the simulated post-eruption decadal evolution of sea ice.

667 In the case of a “supervolcano” eruption, the asymmetry primarily derives from  
668 the different sensitivity of Arctic and Antarctic regional climates to the induced global  
669 energy imbalance and from the associated large-scale atmospheric and oceanic dynamical  
670 reactions. Thermodynamics is the key for the Arctic sea-ice expansion, which is triggered  
671 by the initially reduced atmospheric heat import and is then sustained on a decadal time  
672 scale by the meanwhile reduced oceanic heat import. Noteworthy, decadal responses of  
673 North Atlantic/Arctic large-scale oceanic circulation are qualitatively similar for  
674 historical and “supervolcano” eruptions, both including a delayed strengthening of the  
675 AMOC and a north-westward compression of the subpolar gyre (compare Figure 9b with  
676 *Zanchettin et al.* [2012]). For the “supervolcano” eruptions, however, the post-eruption  
677 drop in the heat content of the global upper ocean is too large to allow for circulation-  
678 driven positive anomalies of ocean heat transport into the Arctic, as diagnosed for  
679 historical eruptions [see *Zanchettin et al.*, 2012, 2013a].

680 In contrast to Arctic sea ice, Antarctic sea ice reacts on the short-term mostly to  
681 dynamical atmospheric changes initiated by the volcanically-induced strengthening of the  
682 Southern Hemisphere’s stratospheric polar vortex. Antarctic sea ice is thereafter  
683 implicated in local surface energy exchange processes dominating the response diagnosed

684 at the hemispheric scale. The post-eruption anomalies of lateral oceanic heat flux are  
685 larger in the Antarctic than in the Arctic, but they only temporarily exceed the internal  
686 variability range (Figure 8d). We regard the temporarily, significantly decreased  
687 poleward oceanic heat transport around year 4 as a marginal contributor to the Antarctic  
688 sea-ice anomaly (negative at this stage). Post-eruption negative anomalies of atmospheric  
689 energy fluxes are likely a consequence of rather than a cause of the chain of local  
690 feedbacks within the Antarctic region. The substantially different exposure of the Arctic  
691 and Antarctic regional climates to volcanically-forced energy imbalances explains why  
692 the inter-hemispheric asymmetry becomes apparent with increasing magnitude of the  
693 eruption.

694 In both the Arctic and the Antarctic, regions of strongest simulated sea-ice  
695 response correspond to key regions for sea-ice and ice-cap variability found in  
696 reconstructions and observations. For instance, extensive increases in September Arctic  
697 sea-ice concentrations are simulated in the Canadian Arctic Archipelago and in the Baffin  
698 Bay (Figure 5c). This is in agreement with records of ice-cap growth from Arctic Canada  
699 covering the last millennium indicating a strong link to large volcanic eruptions  
700 [Anderson *et al.*, 2008; Miller *et al.*, 2012]. In our simulations, internal variability of  
701 Antarctic sea ice is stronger for total area and weaker for total volume compared to Arctic  
702 sea ice (Figure 3, note that total Antarctic sea-ice volume is almost half its Arctic  
703 counterpart). As shown by the forced responses, hemispheric metrics for the Antarctic  
704 often mask strong spatially-heterogeneous variability (Figure 6c), as also indicated by  
705 observations [e.g., Simpkins *et al.*, 2012, 2013]. The interplay between large-scale  
706 dynamics and local processes highlights several relevant mechanisms and features, which  
707 need to be reliably represented in models to build confidence in the simulated  
708 representation of post-eruption sea-ice evolutions, particularly for the Antarctic. These  
709 include, among others, the global hydrological cycle, the downward propagation of polar  
710 vortex signals, ENSO, the global oceanic conveyor of heat, and the atmospheric forcing  
711 of the Antarctic circumpolar current (ACC).

712 The downward propagation of volcanically-forced stratospheric signals,  
713 especially the post-eruption strengthening of the stratospheric polar vortex, is important  
714 for the initial dynamical atmospheric response to explosive volcanic eruptions [e.g.,

715 *Stenchikov et al.*, 2006; *Fischer et al.*, 2007; *Zanchettin et al.*, 2012]. The stratospheric  
716 polar vortex significantly strengthens after the eruption in both hemispheres in HIST2,  
717 SUPER1 ([Figure S4](#)) and SUPER2, with duration tracing that of the imposed radiative  
718 anomaly, but it does not in HIST1 (not shown). Signals in the polar mid-troposphere are  
719 robust only for “supervolcano” eruptions (compare Figure 7b). Larger ensembles could  
720 fully clarify whether the lack of robust dynamical atmospheric responses for historical  
721 eruptions reflects a low signal-to-noise ratio rather than a truly lacking dynamical  
722 response. The latter hypothesis, however, is supported by the deficient representation of  
723 stratospheric dynamics and stratospheric-tropospheric coupling in latest-generation “low-  
724 top” coupled general circulation models (CGCMs) [*Charlton-Perez et al.*, 2013; *Omrani*  
725 *et al.*, 2013], a characteristic which is shared by the version of MPI-ESM used here.

726 Similar concerns arise about the simulated southern tropospheric mid-latitude jet,  
727 e.g., its too equatorward climatological position [e.g., *Swart and Fyfe*, 2012].  
728 Furthermore, simulated Antarctic sea-ice variability and sensitivity to external  
729 disturbances may as well suffer from an imperfect description of tropospheric internal  
730 dynamics, e.g., those related to variability of the Southern Annular Mode [*Simpson et al.*,  
731 2012] and of the associated surface wind variability [*Zhang*, 2013]. In particular, the  
732 strength and latitudinal position of the circumpolar winds affect the Antarctic sea ice via  
733 the Ekman transport [*Maksym et al.*, 2012; *Landrum et al.*, 2012; *Weijer et al.*, 2012]. In  
734 our simulations the total Antarctic sea-ice volume does not support the early post-  
735 eruption horizontal expansion phase (Figure 3b,d) leading to sea-ice thinning. This  
736 contrasts other model-based indications that wind intensification tends to increase  
737 Antarctic sea-ice volume through increased ridged ice production [*Zhang*, 2013]. The  
738 post-eruption initial resilience of total Antarctic sea-ice volume may therefore reflect a  
739 truly distinct dynamical behavior related to extremely strong volcanic forcing, but it may  
740 also reflect poor representation of near-surface atmospheric circulation. This is important  
741 for the case discussed here, given also the marked seasonal character of Antarctic sea-ice  
742 response to the volcanic perturbation during the delayed contraction phase (Figure 4d).

743 The timing of the strongest post-eruption surface cooling at equatorial latitudes,  
744 delayed with respect to that at mid-latitudes strongly contributes to the post-eruption  
745 strengthening of meridional gradients (Figure 7). This equatorial cooling has a strong

746 imprint in the Pacific in the form of an apparent phasing of ENSO on a delayed La Niña  
747 state. This robust response of ENSO to the volcanic perturbation occurs after the  
748 maximum global surface cooling (compare with Figure 2a), when the Toba/SUPER1  
749 ensemble indicates a non-significant, though tendentially warm, ENSO response  
750 [Timmreck *et al.*, 2010]. ENSO in this version of MPI-ESM was consistently found to be  
751 only weakly sensitive to volcanic forcing for a selection of historical-size eruptions in  
752 transient climate simulations covering the last millennium, with an only tendential  
753 response towards a cold (La Niña) anomaly [Zanchettin *et al.*, 2012]. Nonetheless, two  
754 considerations limit our confidence on the simulated forced behavior of ENSO. First,  
755 ENSO's representation still is a challenge for coupled climate models [e.g., Guilyardi *et*  
756 *al.*, 2009] and MPI-ESM-COSMOS-Mill produces too strong and too regular ENSO  
757 fluctuations [compare Jungclaus *et al.*, 2006]. Second, while observations are still  
758 insufficient to draw confident conclusions about the role of ENSO for post-eruption  
759 dynamics of tropical and extra-tropical climates, a recent reconstruction points to a robust  
760 ENSO response to the largest historical tropical eruptions consisting of immediate  
761 cooling followed by anomalous warming one year after [Li *et al.*, 2013]. The  
762 disagreement between indications from this paleoclimate record and from MPI-ESM-  
763 COSMOS-Mill simulations asks for additional dynamical investigations that are beyond  
764 the scope of this study.

765         The strength of the ACC is overestimated in MPI-ESM-COSMOS-Mill  
766 [Marsland *et al.*, 2003; compare also: Jungclaus *et al.*, 2013]. Implications for the  
767 diagnosed post-eruption Antarctic sea-ice evolutions and for the global redistribution of  
768 ocean heat anomalies are difficult to disentangle without dedicated sensitivity  
769 experiments. We note, however, that an overly strong and displaced ACC corresponds to  
770 a biased structure and strength of the subpolar gyres and of associated oceanic convective  
771 activity. In fact, the ocean model MPIOM largely overestimates the mixed layer depth in  
772 the Ross Sea and Weddell Sea gyres [Griffies *et al.*, 2009]. The latter is associated with  
773 the occurrence of a small permanent polynya [Marsland *et al.*, 2003]. Biases of this kind  
774 could be relevant for the delayed reduction diagnosed in Antarctic sea ice (Figure 6d) as  
775 well as for the propagation of heat anomalies into the deep ocean (Figure 11c,d). As  
776 shown by different models, ocean/sea-ice mechanisms during the sea-ice growth phase

777 are strongly interrelated with oceanic stratification and ocean vertical heat transport. In a  
778 weakly stratified Southern Ocean, ice melting from ocean heat flux decreases faster than  
779 the ice growth, leading to an increase in the net ice production and hence an increase in  
780 ice mass [Zhang, 2007]. In the Community Climate System Model version 3, the  
781 freshwater flux between Antarctic sea ice and the Southern Ocean is closely intertwined  
782 with ocean convection and deep-ocean heat uptake [Kirkman and Bitz, 2011]. On the one  
783 hand, a (climatologically) excessively mixed Southern Ocean, as in MPIOM [Griffies *et al.*,  
784 2009], implies reduced efficiency of external forcings to produce anomalous heat  
785 fluxes. On the other hand, locally excessive convective strength as in the Weddell Sea  
786 would imply enhanced oceanic heat losses to the atmosphere. Consequently, simulated  
787 estimates of post-eruption global air-surface cooling [e.g., Timmreck *et al.*, 2010] may be  
788 biased towards being too conservative.

789         A linkage exists between internally-generated multidecadal- and centennial-scale  
790 variability of the Weddell Sea sea-ice cover and of the AMOC in the Kiel Climate Model  
791 [Park and Latif, 2008], where oceanic deep convection within the Weddell Sea gyre  
792 plays a central role in the inter-hemispheric connection [Martin *et al.*, 2013]. Ascribing a  
793 similar bipolar ocean seesaw to the decadal-scale volcanically-forced evolutions  
794 presented here, one would expect that the enhanced deep convection in the Weddell Sea  
795 (Figure 11c,d) hampers the southward deep water flow in the North Atlantic, i.e.,  
796 promote an AMOC slow-down. Overly-strong ocean convection in the Weddell Sea  
797 could accordingly help explain the overall weaker post-eruption AMOC strengthening in  
798 MPI-ESM-COSMOS-Mill compared to other CGCMs, as discussed in Zanchettin *et al.*  
799 [2012]. We note nonetheless that this has likely only faint implications for the post-  
800 eruption decadal Arctic sea-ice evolution, since the northward oceanic heat transport at  
801 subpolar and polar latitudes is largely determined by the gyre circulation (Figure 8b, see  
802 | also Zanchettin *et al.*, 2012, 2013a).

803         Perturbation experiments like the described “supervolcano” experiments highlight  
804 known limits and less understood features of CGCMs and Earth system models, and  
805 might therefore help to delimit the reliability of their forced dynamical climate responses  
806 in more general contexts. We foresee several advantages in a more extensive employment  
807 of “supervolcano” simulations as analog of, e.g., sudden warming experiments. The



808 restoration from the induced cold anomaly highlights variability modes and  
809 teleconnections that would arise under background warming conditions due  
810 predominantly to internal climate variability, whereas externally-forced warming  
811 experiments produce forced anomalous patterns of climate variability. We remark that  
812 our 5-member Toba simulation ensemble was sufficient to yield largely significant, hence  
813 coherent, dynamical responses. Thus, such a small-size ensemble allows for confident  
814 inferences about simulated forced global [Timmreck *et al.*, 2010] as well as regional  
815 [Timmreck *et al.*, 2012] changes. As shown here, “supervolcano” experiments allow  
816 gaining insights on the relative importance of the thermodynamical and dynamical  
817 components of sea-ice responses to imposed negative net radiative imbalances. Such  
818 separation highlights differences in the internal hemispheric dynamics related to external  
819 radiative perturbations.

820

## 821 **5. Conclusions**

822 Ensemble Earth-system-model simulations depict inter-hemispheric differences in the  
823 decadal sea-ice response to strong volcanic eruptions regarding both the sensitivity to the  
824 forcing and the sign of the induced anomalies. Arctic sea ice is very sensitive to volcanic  
825 forcing owing especially to its strong exposure to externally-forced changes in meridional  
826 heat transport. By contrast, Antarctic sea ice appears to be less susceptible to volcanic  
827 forcing and responds only to extremely large (so-called “supervolcano”) eruptions. In  
828 further contrast to Arctic sea ice, the post-eruption evolution of Antarctic sea ice is  
829 mostly determined by feedbacks that set in within the Antarctic region. Whereas Arctic  
830 sea ice robustly expands for a prolonged period after major volcanic eruptions, the post-  
831 eruption Antarctic sea ice evolution includes an initial short-lived expansion and a  
832 subsequent prolonged contraction phase. This sea-ice asymmetry reflects the potential of  
833 volcanic forcing to significantly affect inter-hemispheric interannual-to-decadal climate  
834 variability in a broader context. Nonetheless, key processes implied in the generation of  
835 the asymmetry include less understood, hence poorly simulated features. This poses non-  
836 negligible caveats when extrapolating simulation-based inferences to the real climate  
837 system. In this sense, idealized “supervolcano” perturbation experiments could serve the  
838 assessment of climate models’ performances under strong forcing conditions.

839

840

## 841 Acknowledgements

842

843 The authors thank Achim Stössel for the helpful discussions and comments on an earlier

844 version of the manuscript, and Kay Hübner for the motivating initial analyses. [Comments](#)

845 [from two anonymous reviewers helped improving the clarity of the manuscript.](#) This

846 work benefitted from the MPI-M integrated projects “Millennium” and “Super

847 Volcano”. This work was supported by the Federal Ministry for Education and Research

848 in Germany (BMBF) through the research program "MiKlip"

849 (FKZ:01LP1158A(DZ):/01LP1130A(CT)).

850

851

852

## 853 References

854 Ahmed, M., et al. (2013) Continental-scale temperature variability during the past two millennia. *Nat.*

855 *Geosc.*, 6, 339-346, doi:10.1038/NNGEO1797

856 Anderson, R. K., Miller, G. H., Briner, J. P., Lifton, N. A., and S. B. DeVogel (2008), A millennial

857 perspective on Arctic warming from 14C in quartz and plants emerging from beneath ice caps.

858 *Geophys. Res. Lett.* 35:L01502, doi:10.1029/2007GL032057

859 Barlyaeva, T. V., I. A. Mironova, and D. I. Ponyavin (2009), Nature of decadal variations in the climatic

860 data of the second half of the 20<sup>th</sup> century. *Dokl. Earth. Sci.* 425A(3):419–423

861 [Beitsch, A., J. H. Jungclaus, and D. Zanchettin \(2013\) Patterns of decadal-scale Arctic warming events in](#)

862 [simulated climate. \*Clim. Dyn.\* doi:10.1007/s00382-013-2004-5](#)

863 Bitz, C.M., and L.M. Polvani (2012), Antarctic climate response to stratospheric ozone depletion in a fine

864 resolution ocean climate model, *Geophys. Res. Lett.*, 39, L20705, doi:10.1029/2012GL053393

865 [Bothe, O., J. H. Jungclaus, D. Zanchettin, and E. Zorita \(2013\) Climate of the last millennium: ensemble](#)

866 [consistency of simulations and reconstructions. \*Clim. Past\*, 9, 1089–1110, doi:10.5194/cp-9-1-](#)

867 [2013](#)

868 Brohan P, Kennedy JJ, Harris I, Tett SFB, Jones PD (2006) Uncertainty estimates in regional and global

869 observed temperature changes: a new dataset from 1850. *J. Geophys. Res.* 111:D12106,

870 doi:12110.11029/12005JD006548

871 Charlton-Perez, A. J., et al. (2013), On the lack of stratospheric dynamical variability in low-top versions of

872 the CMIP5 models. *J. Geophys. Res.*, 118, 2494–2505, doi:10.1002/jgrd.50125

873 Chylek, P., Folland, C.K., Lesins, G., and M. K. Dubey (2010) Twentieth century bipolar seesaw of the

874 Arctic and Antarctic surface air temperatures. *Geophys. Res. Lett.* 37, L08703,

875 doi:10.1029/2010GL042793

876 Crowley, T. J., and M. B. Unterman (2012) Technical details concerning development of a 1200-yr proxy

877 index for global volcanism, *Earth Syst. Sci. Data Discuss.*, 5, 1-28, doi:10.5194/essdd-5-1-2012

878 Crowley, T. J., et al. (2008), Volcanism and the little ice age, *PAGES News*, 16:22–23

879 Duncan, R. P., P. Fenwick, J. G. Palmer, M. S. McGlone, and C. S. M. Turney (2010) Non-uniform

880 interhemispheric temperature trends over the past 550 years. *Clim. Dyn.* 35:1429–1438,

881 doi:10.1007/s00382-010-0794-2

882 [Fernández-Donado, L., and co-authors \(2013\) Large-scale temperature response to external forcing in](#)

883 [simulations and reconstructions of the last millennium. \*Clim. Past\*, 9, 393–421, doi:10.5194/cp-9-](#)

884 [393-2013](#)

885 Fischer, E., J. Luterbacher, E. Zorita, S. F. B. Tett, C. Casty, and H. Wanner (2007), European climate

886 response to tropical volcanic eruptions over the last half millennium, *Geophys. Res. Lett.*, 34,

887 L05707, doi:10.1029/2006GL027992

888 | [Gillett, N. P., and D. W. J. Thompson \(2003\) Simulation of Recent Southern Hemisphere Climate Change. Science, 302 \(5643\), 273-275. doi:10.1126/science.1087440](#)

889 |

890 | Graf, H.F., and D. Zanchettin (2012), Central Pacific El Niño, the “subtropical bridge” and Eurasian

891 | Climate. *J. Geophys. Res.*, 117, D01102, doi:10.1029/2011JD016493

892 | Griffies, S. M., et al. (2009), Coordinated Ocean-ice Reference Experiments (COREs). *Oce. Modelling*, 26:

893 | 1–46. doi:http://dx.doi.org/10.1016/j.ocemod.2008.08.007

894 | Guilyardi, E., A. Wittenberg, A. Fedorov, M. Collins, C.Z. Wang, A. Capotondi, G.J. van Oldenborgh, and

895 | T. Stockdale (2009) Understanding El Niño in ocean-atmosphere General Circulation Models:

896 | Progress and challenges. *Bull. Amer. Met. Soc.*, 90, 325–340

897 | [Henriksson, S.V. , P. Räisänen, J. Silén, and A. Laaksonen \(2012\) Quasiperiodic climate variability with a](#)

898 | [period of 50-80 years: Fourier analysis of measurements and Earth System Model simulations.](#)

899 | [Clim. Dyn. 39: 7-8, 1999-2011. doi:10.1007/s00382-012-1341-0](#)

900 | Heuzé, C., K. J. Heywood, D. P. Stevens, and J. K. Ridley (2013) Southern Ocean bottom water

901 | characteristics in CMIP5 models. *Geophys. Res. Lett.*, 40, 1409–1414, doi:10.1002/grl.50287

902 | Holden, P. B. N. R. Edwards, E.W. Wolff, N. J. Lang, J. S. Singarayer, P. J. Valdes, and T. F. Stocker

903 | (2010) Interhemispheric coupling, the West Antarctic Ice Sheet and warm Antarctic interglacials.

904 | *Clim. Past*, 6, 431-443, doi:10.5194/cp-6-431-2010

905 | Jones, G. S., J. M. Gregory, P. A. Stott, S. F. B. Tett, and R. B. Thorpe (2005) An AOGCM simulation of

906 | the climate response to a volcanic super-eruption. *Clim. Dyn.* 25, 725-738

907 | Jungclaus J.H., N. Keenlyside, M. Botzet, H. Haak, J. J. Luo, M. Latif, J. Marotzke, U. Mikolajewicz, and

908 | F. Roeckner (2006) Ocean circulation and tropical variability in the coupled model

909 | ECHAM5/MPI-OM. *J. Clim* 19:3952–3972

910 | Jungclaus, J. H., et al. (2010), Climate and carbon-cycle variability over the Last Millennium. *Clim. Past*, 6,

911 | 723-737, doi:10.5194/cp-6-723-2010

912 | Jungclaus J. H., et al. (2013) Characteristics of the ocean simulations in MPIOM, the ocean component of

913 | the Max Planck Institute Earth System Model. *J. Adv. Model. Earth Syst.*,

914 | doi:10.1002/jame.20023

915 | Karpechko, A. Y., N. P. Gillett, M. Dall'Amico, and L. J. Gray (2010), Southern Hemisphere atmospheric

916 | circulation response to the El Chichón and Pinatubo eruptions in coupled climate models, *Q. J. R.*

917 | *Meteorol. Soc.*, 136, 1813-1822, doi:10.1002/qj.683

918 | Keith, D.W. (1995) Meridional energy transport: uncertainty in zonal means. *Tellus*, 47A, 30-44

919 | [King, J. \(2014\) Climate science: A resolution of the Antarctic paradox. Nature 505, 491–492.](#)

920 | [doi:10.1038/505491a](#)

921 | Kinnard, G., Zdanowicz, C.M., Fisher, D.A., Isakkson, E., de Vernal, A., Thompson, L.G. (2011)

922 | Reconstructed changes in Arctic sea ice over the past 1,450 years. *Nature* 479: 509-513,

923 | doi:10.1038/nature10581

924 | Kirkman, C. and C.M. Bitz (2011), The Effect of the Sea Ice Freshwater Flux on Southern Ocean

925 | Temperatures in CCSM3: Deep Ocean Warming and Delayed Surface Warming, *J. Climate*, 24,

926 | pp. 2224-2237 doi: 10.1175/2010JCLI3625

927 | [Koldunov, N. V., D. Stammer, and J. Marotzke \(2010\) Present-Day Arctic Sea Ice Variability in the](#)

928 | [Coupled ECHAM5/MPI-OM Model. J. Climate, 23, 2520–2543.](#)

929 | [doi:http://dx.doi.org/10.1175/2009JCLI3065.1](#)

930 | Landrum, L., M. M. Holland, D. P. Schneider, and E. Hunke (2012) Antarctic Sea Ice Climatology,

931 | Variability, and Late Twentieth-Century Change in CCSM4. *J. Clim.*25, 4817-4838,

932 | doi:10.1175/JCLI-D-11-00289.1

933 | [Li, C., D. Notz, S. Tietsche, and J. Marotzke \(2013\) The Transient versus the Equilibrium Response of Sea](#)

934 | [Ice to Global Warming. J. Climate, 26, 5624–5636. doi: http://dx.doi.org/10.1175/JCLI-D-12-](#)

935 | [00492.1](#)

936 | Li, J., S.-P. Xie, E. R. Cook, M. S. Morales, D. A. Christie, N. C. Johnson, F. Chen, R. D’Arrigo, A. M.

937 | Fowler, X. Gou, and K. Fang. (2013) El Niño modulations over the past seven centuries. *Nature*

938 | *Cl. Ch.*, 3:9, 822-826, doi:10.1038/NCLIMATE1936

939 | Maksym, T., S.E. Stammerjohn, S. Ackley, and R. Massom (2012) Antarctic sea ice—A polar opposite?

940 | *Oceanography* 25(3):140–151, http://dx.doi.org/10.5670/oceanog.2012.88.

941 | Marshall, G.J., (2003) Trends in the Southern Annular Mode from observations and reanalyses. *J. Clim.*,

942 | 16:4,134–4,143, http://dx.doi.org/10.1175/1520-0442(2003)016<4134:TITSAM>2.0.CO;2

943 Marsland, S.J., H. Haak, J.H. Jungclaus, M. Latif, and F. Röske (2003), The Max Planck Institute global  
944 ocean/sea ice model with orthogonal curvilinear coordinates, *Ocean Modell.*, 5, 91-127

945 Martin, T., W. Park, and M. Latif (2013) Multi-centennial variability controlled by Southern Ocean  
946 convection in the Kiel Climate Model. *Clim. Dyn.*, 40:2005–2022, doi:10.1007/s00382-012-1586-  
947 7

948 Massonnet, F., P. Mathiot, T. Fichefet, H. Goosse, C. K. Beatty, M. Vancoppenolle, and T. Lavergne  
949 (2013), Model reconstruction of the Antarctic sea ice thickness and volume changes over 1980–  
950 2008 using data assimilation. *Oce. Modelling*, 64: 67-75,  
951 <http://dx.doi.org/10.1016/j.ocemod.2013.01.003>

952 Meier, W. N., D. Gallaher, and G. G. Campbell (2013), New estimates of Arctic and Antarctic sea ice  
953 extent during September 1964 from recovered Nimbus I satellite imagery. *Cryosph.*, 7:699-705,  
954 doi:10.5194/tc-7-699-2013

955 [Menary, M.B., W. Park, K. Lohmann, M. Vellinga, M.D. Palmer, M. Latif, and J.H. Jungclaus \(2012\) A](#)  
956 [multimodel comparison of centennial Atlantic meridional overturning circulation variability. \*Clim.\*](#)  
957 [\*Dyn.\* 38:11-12, 2377-2388](#)

958 Miller, G. H., et al. (2012), Abrupt onset of the Little Ice Age triggered by volcanism and sustained by sea-  
959 ice/ocean feedbacks, *Geophys. Res. Lett.*, 39, L02708, doi:10.1029/2011GL050168

960 Niemeier, U., et al. (2009), Initial fate of fine ash and sulfur from large volcanic eruptions, *Atmos. Chem.*  
961 *Phys.*, 9, 9043–9057

962 Notz, D., and J. Marotzke (2012) Observations reveal external driver for Arctic sea-ice retreat. *Geophys.*  
963 *Res. Lett.* 39(8), doi:10.1029/2012GL051094

964 Omrani, N.-E., Keenlyside, N., Bader, J., & Manzini, E. (2013). Stratosphere key for wintertime  
965 atmospheric response to warm Atlantic decadal conditions. *Clim. Dyn.*, doi:10.1007/s00382-013-  
966 1860-3

967 Park, W., and M. Latif (2008), Multidecadal and multicentennial variability of the meridional overturning  
968 circulation. *Geophys. Res. Lett.* 35:L22703, doi:10.1029/2008GL035779

969 Parkinson, C. L. (2004) Southern Ocean sea ice and its wider linkages: insights revealed from models and  
970 observations. *Antarctic Science* 16 (4): 387–400, doi:10.1017/S0954102004002214

971 Polvani, L. M., and K. L. Smith (2013) Can natural variability explain observed Antarctic sea ice trends?  
972 New modeling evidence from CMIP5. *Geophys. Res. Lett.* 40, 3195–3199, doi:10.1002/grl.50578

973 Raddatz, T.J., et al (2007), Will the tropical land biosphere dominate the climate-carbon cycle feedback  
974 during the twenty-first century? *Clim Dyn* 29:565–574

975 Robock, A., T. Adams, M. Moore, L. Oman, and G. Stenchikov (2007) Southern Hemisphere atmospheric  
976 circulation effects of the 1991 Mount Pinatubo eruption, *Geophys. Res. Lett.* 34,  
977 doi:10.1029/2007GL031403

978 Roeckner, E., et al. (2006), Sensitivity of simulated climate to horizontal and vertical resolution in the  
979 ECHAM5 atmosphere model, *J. Clim.*, 19, 3771–3791, doi:http://dx.doi.org/10.1175/JCLI3824.1

980 Russell, J. L., R. J. Stouffer, and K. W. Dixon (2006) Intercomparison of the Southern Ocean Circulations  
981 in IPCC Coupled Model Control Simulations, *J. Clim.* 19:4560-4575

982 Sallée, J.-B., E. Shuckburgh, N. Bruneau, A. J. S. Meijers, T. J. Bracegirdle, and Z. Wang (2013),  
983 Assessment of Southern Ocean mixed layer depths in CMIP5 models: Historical bias and forcing  
984 response, *J. Geophys. Res. Oceans*, 118, 1845–1862, doi:10.1002/jgrc.20157

985 Schaefer, J.M., Denton, G.D., Kaplan, M.R., et al. (2009) High-frequency Holocene glacier fluctuations in  
986 New Zealand differ from the northern signature. *Science* 324: 622–625

987 Schleussner, C. F., and G. Feulner (2012) A volcanically triggered regime shift in the subpolar North  
988 Atlantic ocean as a possible origin of the Little Ice Age. *Clim. Past Discuss.*, 8, 6199–6219,  
989 doi:10.5194/cpd-8-6199-2012

990 [Schubert, J. J., B. Stevens, T. Crueger. \(2013\) Madden-Julian oscillation as simulated by the MPI Earth](#)  
991 [System Model: Over the last and into the next millennium. \*J. Adv. Model. Earth Sys.\*, 5:1, 71-84,](#)  
992 [doi:10.1029/2012MS000180](#)

993 Segsneider J., A. Beitsch, C. Timmreck, V. Brovkin, T. Ilyina, J. Jungclaus, S. J. Lorenz, K. D. Six, and  
994 D. Zanchettin (2013) Impact of an extremely large magnitude volcanic eruption on the global  
995 climate and carbon cycle estimated from ensemble Earth System Model simulations. *Biogeosc.* 10,  
996 669–687, doi:10.5194/bg-10-669-2013

997 Shakun, J.D., et al. (2012) Global warming preceded by increasing carbon dioxide concentrations during  
998 the last deglaciation, *Nature* 484, 49–54, doi:10.1038/nature10915

- 999 Shin, S.-I., Z. Liu, B. L. Otto-Bliesner, J. E. Kutzbach, and S. J. Vavrus (2003) Southern Ocean sea-ice  
1000 control of the glacial North Atlantic thermohaline circulation. *Geophys. Res. Lett.* 30(2), 1096,  
1001 doi:10.1029/2002GL015513
- 1002 [Sigmond, M., and J. C. Fyfe \(2014\) The Antarctic Sea Ice Response to the Ozone Hole in Climate Models.](#)  
1003 [J. Climate, 27, 1336–1342, doi: http://dx.doi.org/10.1175/JCLI-D-13-00590.1](#)
- 1004 Simpkins, G. R., L. M. Ciasto, D. W. J. Thompson, and M. H. England (2012) Seasonal Relationships  
1005 between Large-Scale Climate Variability and Antarctic Sea Ice Concentration. *J. Climate*, 25,  
1006 5451–5469, doi: <http://dx.doi.org/10.1175/JCLI-D-11-00367.1>
- 1007 Simpkins, G. R., L. M. Ciasto, and M. H. England (2013) Observed variations in multidecadal Antarctic sea  
1008 ice trends during 1979–2012. *Geophys. Res. Lett.* 40, 1–6, doi:10.1002/grl.50715
- 1009 Simpson, I., P. Hitchcock, T. Shepherd, and J. Scinocca (2012) Southern Annular Mode Dynamics in  
1010 Observations and Models. Part 1: the Influence of Climatological Zonal Wind Biases in a  
1011 Comprehensive GCM. *J. Climate*, doi:10.1175/JCLI-D-12-00348.1, in press
- 1012 Stammerjohn, S., R. Massom, D. Rind, and D. Martinson (2012), Regions of rapid sea ice change: An  
1013 inter-hemispheric seasonal comparison. *Geophys. Res. Lett.*, 39, L06501,  
1014 doi:10.1029/2012GL050874
- 1015 Stenchikov, G., K. Hamilton, R. J. Stouffer, A. Robock, V. Ramaswamy, B. Santer, and H.-F. Graf (2006), Arctic  
1016 Oscillation response to volcanic eruptions in the IPCC AR4 climate models, *J. Geophys. Res.*, 111,  
1017 D07107, doi:10.1029/2005JD006286
- 1018 Stenchikov, G., T. L. Delworth, V. Ramaswamy, R. J. Stouffer, A. Wittenberg, and F. Zeng (2009), Volcanic  
1019 signals in oceans, *J. Geophys. Res.*, 114, D16104, doi:10.1029/2008JD011673
- 1020 Stier, P., et al. (2005), The aerosol climate model ECHAM5-AM, *Atmos. Chem. Phys.*, 5, 1125–1156
- 1021 Stössel, A., Zhang, Z., and Vihma, T. (2011) The effect of alternative real-time wind forcing on Southern  
1022 Ocean sea-ice simulations. *J. Geophys. Res.* 116, C11021, doi:10.1029/2011JC007328
- 1023 Stouffer, R. J., and Coauthors (2006), GFDL's CM2 Global Coupled Climate Models. Part IV: Idealized  
1024 Climate Response. *J. Clim.*, 19, 723–740, doi:<http://dx.doi.org/10.1175/JCLI3632.1>
- 1025 Stroeve, J. C., V. Kattsov, A. P. Barrett, M. C. Serreze, T. Pavlova, M. M. Holland, and W. N. Meier  
1026 (2012), Trends in Arctic sea ice extent from CMIP5, CMIP3 and observations, *Geophys. Res.*  
1027 *Lett.*, 39(16), L16502, doi:10.1029/2012GL052676
- 1028 Swart, N. C., and J. C. Fyfe (2012) Observed and simulated changes in southern hemisphere surface  
1029 westerlies. *Geophys. Res. Lett.*, 39, L16 711, doi:10.1029/2012GL052810
- 1030 Thompson, D. W. J., J. M. Wallace, P. D. Jones, and J. J. Kennedy (2009) Identifying Signatures of Natural  
1031 Climate Variability in Time Series of Global-Mean Surface Temperature: Methodology and  
1032 Insights. *J. Climate*, 22, 6120–6141.
- 1033 [Tietsche, S., D. Notz, J. H. Jungclaus and J. Marotzke \(2011\) Recovery mechanisms of Arctic summer sea](#)  
1034 [ice. Geophys. Res. Lett., 38, L02707, doi:10.1029/2010GL045698](#)
- 1035
- 1036 Timmreck, C., S. J. Lorenz, T. J. Crowley, S. Kinne, T. J. Raddatz, M. A. Thomas, and J. H. Jungclaus  
1037 (2009), Limited temperature response to the very large AD 1258 volcanic eruption, *Geophys. Res.*  
1038 *Lett.*, 36, L21708, doi:10.1029/2009GL040083.
- 1039 Timmreck, C., H.-F. Graf, S. J. Lorenz, U. Niemeier, D. Zanchettin, D. Matei, J. H. Jungclaus and T. J.  
1040 Crowley (2010), Aerosol size confines climate response to volcanic super-eruptions, *Geophys.*  
1041 *Res. Lett.* 37, L24705, doi:10.1029/2010GL045464.
- 1042 Timmreck, C. (2012), Modeling the climatic effects of large volcanic eruptions. *WIREs Clim. Change*, 3,  
1043 545–564, doi:10.1002/wcc.192
- 1044 Timmreck C, H.-F. Graf, D. Zanchettin, S. Hagemann, T. Kleinen and K. Krüger (2012) Climate response  
1045 to the Toba eruption: regional changes. *Quat Int.* 258,30-44.
- 1046 [Turner, J., J. C. Comiso, G. J. Marshall, T. A. Lachlan-Cope, T. Bracegirdle, T. Maksym, M. P. Meredith,](#)  
1047 [Z. Wang, and A. Orr \(2009\) Non-annular atmospheric circulation change induced by stratospheric](#)  
1048 [ozone depletion and its role in the recent increase of Antarctic sea ice extent. Geophys. Res. Lett.,](#)  
1049 [36, L08502, doi:10.1029/2009GL037524](#)
- 1050 Turner, J., T. Bracegirdle, T. Phillips, G. J. Marshall, and J. S. Hosking (2013), An Initial Assessment of  
1051 Antarctic Sea Ice Extent in the CMIP5 Models. *J. Climate*, 26, 1473–1484,  
1052 doi:<http://dx.doi.org/10.1175/JCLI-D-12-00068.1>

Deleted: 2012

Deleted: , in press

- 1053 Wang, M. and J. E. E. Overland (2012) A sea ice free summer Arctic within 30 years—an update from  
1054 CMIP5 models, *Geophys. Res. Lett.*, doi:10.1029/2012GL052868, in press
- 1055 Weber, M. E., P. U. Clark, W. Ricken, J. X. Mitrovica, S. W. Hostetler, and G. Kuhn (2011)  
1056 Interhemispheric Ice-Sheet Synchronicity During the Last Glacial Maximum. *Science*, 334(6060),  
1057 1265-1269, doi:10.1126/science.1209299
- 1058 Weijer, W., et al. (2012) The Southern Ocean and its climate in CCSM4. *J. Climate*, 25, 2652–2675,  
1059 doi:10.1175/JCLI-D-11-00302.1
- 1060 Wetzel, P., Winguth, A., and E. Maier-Reimer (2005), Sea-to-air CO<sub>2</sub> fluxes from 1948 to 2003. *Glob*  
1061 *Biogeochem Cycles* 19 GB2005. doi:10.1029/2004GB002339
- 1062 Zanchettin, D., C. Timmreck, H.-F. Graf, A. Rubino, S. Lorenz, K. Lohmann, K. Krueger, and J. H.  
1063 Jungclaus (2012), Bi-decadal variability excited in the coupled ocean–atmosphere system by  
1064 strong tropical volcanic eruptions. *Clim. Dyn.*, 39:1-2, 419-444, DOI:10.1007/s00382-011-1167-1
- 1065 Zanchettin, D., O. Bothe, H. F. Graf, S. J. Lorenz, J. Luterbacher, C. Timmreck, and J. H. Jungclaus  
1066 (2013a) Background conditions influence the decadal climate response to strong volcanic  
1067 eruptions. *J. Geophys. Res. Atm.*, 118(10): 4090-4106, doi:10.1002/jgrd.50229
- 1068 [Zanchettin, D., A. Rubino, D. Matei, O. Bothe, and J. H. Jungclaus \(2013b\) Multidecadal-to-centennial](#)  
1069 [SST variability in the MPI-ESM simulation ensemble for the last millennium. \*Clim. Dyn.\*, 40:5,](#)  
1070 [1301-1318, doi:10.1007/s00382-012-1361-9](#)
- 1071 Zhang, J. (2007) Increasing Antarctic Sea Ice under Warming Atmospheric and Oceanic Conditions, *J.*  
1072 *Climate*, 20, 2515–2529, doi: http://dx.doi.org/10.1175/JCLI4136.1
- 1073 Zhang, J. (2013) Modeling the impact of wind intensification on Antarctic sea ice volume. *J. Clim.*, doi:  
1074 http://dx.doi.org/10.1175/JCLI-D-12-00139.1
- 1075 Zhong, Y., et al (2011) Centennial-scale climate change from decadal-paced explosive volcanism: a  
1076 coupled sea ice-ocean mechanism. *Clim Dyn* 37:11-12, 2373-2387, doi:10.1007/s00382-010-  
1077 0967-z

1078

1079

## 1080 **Figure captions**

1081

1082 **Figure 1** – Simulated imposed forcing in the two historical and the two “supervolcano”  
1083 ensembles as diagnosed through anomalies in the global-average top-of-atmosphere solar  
1084 (shortwave, a), thermal (longwave, b) and net (c) radiation. Lines (shading): mean  
1085 (standard error of the mean). Black dotted lines indicate the internal variability range  
1086 ( $n=10$ , see methods). The magenta vertical hatched line indicates the approximate start of  
1087 the eruptions. Lag(0) corresponds to January of the eruption year. Positive anomalies  
1088 correspond to increased downward flux. No smoothing was applied to the series.

1089

1090 **Figure 2** – Simulated post-eruption anomalies of global-average surface (2 meters) air  
1091 temperature (SAT) (panel a) and total precipitation (b), and Northern-hemispheric  
1092 average SAT (c) and difference between Northern and Southern-hemispheric average  
1093 SAT (d) for the two historical and the two “supervolcano” ensembles. Lines (shading):  
1094 mean (standard error of the mean). Black dashed lines indicate the internal variability  
1095 range ( $n=10$ , see methods). The magenta vertical hatched line indicates the approximate  
1096 start of the eruptions. Lag(0) corresponds to January of the eruption year. Note that the y-  
1097 axis in panels c and d has the same scale, highlighting the relative magnitude of inter-  
1098 hemispheric differences in the temperature response.

1099

1100 **Figure 3** – Simulated post-eruption anomalies of Arctic (top) and Antarctic (bottom) total  
1101 sea-ice area (top panels) and volume (bottom) for the two historical and the two  
1102 “supervolcano” ensembles. Lines (shading): mean anomaly (standard error of the mean).  
1103 Anomalies are smoothed with a 13-months centered moving average. Black dashed lines



1104 indicate the internal variability range ( $n=10$ , see methods). The magenta vertical hatched  
1105 line indicates the approximate start of the eruptions. Lag(0) corresponds to January of the  
1106 eruption year. The number on top-right of each panel is the approximate pre-eruption  
1107 climatology. The y-axis has the same scale in panels a and b, and in panels c and d,  
1108 highlighting the different magnitude of the hemispheric responses.

1109

1110 **Figure 4** – Ensemble-mean simulated seasonal evolutions of hemispheric sea-ice area for  
1111 integration years 1-2 (panels a, b) and 4-6 (panels c,d) for the two historical and the two  
1112 “supervolcano” ensembles. Gray shading (hatched white line) represents the 98% range  
1113 (mean) for signal occurrence in the control run. Signal in the control run corresponds to  
1114 the annual evolution averaged over three randomly chosen consecutive years, for a 10-  
1115 member ensemble.

1116

1117 **Figure 5** – Ensemble-mean simulated March (top panels) and September (bottom panels)  
1118 Arctic sea-ice concentration anomalies for integration years 1-2 (panels a, b) and 4-6  
1119 (panels c,d) of the SUPER1 ensemble. Only grid points where the anomaly is significant  
1120 at 95% confidence ( $n=5$ , see methods) are shown. The green and orange lines indicate,  
1121 respectively, the pre-eruption average and post-eruption average sea-ice edge.

1122

1123 **Figure 6** – Ensemble-mean simulated March (top panels) and September (bottom panels)  
1124 Antarctic sea-ice concentration anomalies for integration years 1-2 (panels a, b) and 4-6  
1125 (panels c,d) of the SUPER1 ensemble. Only grid points where the anomaly is significant  
1126 at 95% confidence ( $n=5$ , see methods) are shown. The green and orange lines indicate,  
1127 respectively, the pre-eruption average and post-eruption average sea-ice edge.

1128 **Figure 7** – Ensemble-mean post-eruption evolution of zonal-mean surface temperature  
1129 (panel a), zonal-mean 500 hPa zonal wind (panel b) and meridional wind (panel c)  
1130 anomalies for the SUPER1 ensemble. Positive zonal and meridional winds are,  
1131 respectively, eastward and northward. Filled contours in panels a,b: Hovmoeller  
1132 diagrams, only changes statistically significant at 99% confidence are shown. Line plots  
1133 in panel a: anomalies of equator-to-pole gradient (EPG) for the Northern (top) and  
1134 Southern (bottom) hemispheres. EPG is defined, for each hemisphere, as the difference  
1135 between values at the grid latitude closest to the equator and the first grid latitude  
1136 poleward of 70°. Dotted lines in panels a,c are 98% confidence ranges. A 13-month  
1137 running-average smoothing has been applied to all data.

1138

1139

1140 **Figure 8** - Ensemble-mean simulated anomalies of zonally-integrated atmospheric energy  
1141 transport at ~60°N (panel a) and ~60°S (c), and oceanic heat transport by advection at  
1142 60°N (b) and at 60°S (d) for the SUPER1 ensemble. Lines (shading): mean anomaly  
1143 (standard error of the mean). Dashed lines indicate the internal variability range ( $n=5$ , see  
1144 methods). Internal variability ranges for the gyre component of ocean heat transport are  
1145 not shown, since barely distinguishable from that of the total transport. The magenta  
1146 vertical hatched line indicates the approximate start of the eruptions. Lag(0) corresponds  
1147 to January of the eruption year. Anomalies are smoothed with a 13-months centered  
1148 moving average. Positive values correspond to northward transport anomalies (y-axis is

Deleted: meridional

Deleted: zonal

1149 inverted in panels c and d) to ease comparison of poleward transports in the two  
1150 hemispheres.

1151

1152 **Figure 9** – Ensemble-mean simulated annual-average oceanic barotropic streamfunction  
1153 for integration years 1-2 (panels a, b) and 4-6 (panels c,d) of the SUPER1 ensemble. Only  
1154 changes statistically significant at 95% confidence are shown. The green and orange lines  
1155 indicate, respectively, the pre-eruption and post-eruption average winter (top: DJF,  
1156 bottom: JJA) sea-ice edge.

1157

1158 **Figure 10** – Ensemble-mean simulated annual-average mixed layer thickness in two  
1159 oceanic deep convection regions for integration years 1-2 (panels a, b) and 4-6 (panels  
1160 c,d) of the SUPER1 ensemble. Only changes statistically significant at 95% confidence  
1161 are shown. The green and orange lines indicate, respectively, the pre-eruption and post-  
1162 eruption average winter (top: DJF, bottom: JJA) sea-ice edge.

1163

1164 **Figure 11** – Ensemble-mean simulated Southern Hemisphere summer (DJF, top) and  
1165 winter (JJA, bottom) total net surface energy flux (latent and sensible heat, short- and  
1166 long-wave radiation, panels a,c) and 10-meter wind anomalies (panels b,d) for integration  
1167 years 1-2 of the SUPER1 ensemble. Panels a,c: black dots indicate grid points where  
1168 changes are non significant at the 95% confidence level; panels b,d: only changes  
1169 statistically significant at 95% confidence for at least one of the wind components are  
1170 shown

1171

1172 **Figure 12** – Ensemble-mean simulated Southern Hemisphere summer (DJF, top) and  
1173 winter (JJA, bottom) total net surface energy flux (latent and sensible heat, short- and  
1174 long-wave radiation, panels a,c) and 10-meter wind anomalies (panels b,d) for integration  
1175 years 4-6 of the SUPER1 ensemble. Panels a,c: black dots indicate grid points where  
1176 changes are non significant at the 95% confidence level; panels b,d: only changes  
1177 statistically significant at 95% confidence for at least one of the wind components are  
1178 shown.

1179