Compare PDFs

Find the difference between two PDF files!
Enter two PDFs and the difference will show up below.

	acp-2020-1191-manuscript-version1 (2).pdf		SWV_revision.pdf
1		1	
2	The Impact of Volcanic Eruptions of Different Magnitude on	2	The Impact of Volcanic Eruptions of Different Magnitude on
3	Stratospheric Water Vapour in the Tropics	3	Stratospheric Water Vapour in the Tropics
4	Clarissa Kroll	4	Clarissa Alicia Kroll
5	1	5	1,2
6	, Sally Dacie	6	, Sally Dacie
7	1	7	1,2
8	, Alon Azoulay	8	, Alon Azoulay
9	1,2	9	1,3
10	, Hauke Schmidt	10	, Hauke Schmidt
11	1	11	1
12	, and Claudia Timmreck	12	, and Claudia Timmreck
13	1	13	1
14	1	14	1
15	Max Planck Institute for Meteorology, Hamburg, Germany	15	Max Planck Institute for Meteorology, Hamburg, Germany
16	2	16	2
		17	International Max Planck Research School on Earth System Modelling (IMPRS-ESM), Hamburg, Germany
		18	3
17	now at: Remote Sensing Technology Institute (IMF), German Aerospace Center (DLR), Oberpfaffenhofen, Germany	19	now at: Remote Sensing Technology Institute (IMF), German Aerospace Center (DLR), Oberpfaffenhofen, Germany
18	Correspondence:Clarissa Kroll (clarissa.kroll@mpimet.mpg.de)	20	Correspondence:Clarissa Kroll (clarissa.kroll@mpimet.mpg.de)
19	Abstract.Volcanic eruptions increase the stratospheric water vapour (SWV) entry via longwave heating through the aerosol	21	Abstract.Increasing the temperature of the tropical cold point region through heating by volcanic aerosols results in increases
20	layer in the cold point region, and this additional SWV alters the atmospheric energy budget. We analyze tropical volcanic	22	in the entry value of stratospheric water vapour and subsequent changes in the atmospheric energy budget. We analyze tropical
21	eruptions of different eruption strengths with sulfur (S) injections ranging from 2.5TgS up to 40TgS using EVAens, the	23	volcanic eruptions of different strengths with sulfur (S) injections ranging from 2.5TgS up to 40TgS using EVAens, the
22	100-member ensemble of the Max Planck Institute - Earth System Model in its low resolution configuration (MPI-ESM-LR)	24	100-member ensemble of the Max Planck Institute - Earth System Model in its low resolution configuration (MPI-ESM-LR)
23	with artificial volcanic forcing generated by the Easy Volcanic Aerosol (EVA) tool. Significant increases in SWV are found for5	25	with artificial volcanic forcing generated by the Easy Volcanic Aerosol (EVA) tool. Significant increases in SWV are found5
24	the mean over all ensemble members from 2.5TgS onward ranging between [5,160] %, while for single ensemble members	26	for the mean over all ensemble members from 2.5TgS onward ranging between [5,160] %. However, for single ensemble

25	the standard deviation between the control run members (0TgS) is larger than SWV increase of single ensemble members	27	members the standard deviation between the control run members (0TgS) is larger than SWV increase of single ensemble
26	for the eruption strengths up to 20 Tg S. A historical simulation using observation based forcing files of the Mt. Pinatubo	28	members for the eruption strengths up to 20 Tg S. A historical simulation using observation based forcing files of the Mt.
27	eruption, which was estimated to have emitted(7.5±2.5) TgS, returns SWV increases slightly higher than the 10TgS	29	Pinatubo eruption, which was estimated to have emitted(7.5±2.5) TgS, returns SWV increases slightly higher than the 10
28	EVAens simulations due to differences in the aerosol profile shape. An additional amplification of the tape recorder signal is10	30	TgS EVAens simulations due to differences in the aerosol profile shape. An additional amplification of the tape recorder signal10
29	also apparent, which is not present in the 10TgS run. These differences underline that it is not only the eruption volume,	31	is also apparent, which is not present in the 10TgS run. These differences underline that it is not only the eruption volume,
30	but also the aerosol layer shape and location with respect to the cold point that have to be considered for post-eruption SWV	32	but also the aerosol layer shape and location with respect to the cold point that have to be considered for post-eruption SWV
31	increases. The additional tropical clear sky SWV forcing for the different eruption strengths amounts to [0.02, 0.65]W/m	33	increases. The additional tropical clear sky SWV forcing for the different eruption strengths amounts to [0.02, 0.65]Wm
32	2	34	−2
33	,	35	,
34	ranging between [2.5, 4] percent of the aerosol radiative forcing in the 10TgS scenario. The monthly cold point temperature	36	ranging between [2.5, 4] percent of the aerosol radiative forcing in the 10TgS scenario. The monthly cold point temperature
35	increases leading to the SWV increase are not linear with respect to AOD nor is the corresponding SWV forcing, among others,15	37	increases leading to the SWV increase are not linear with respect to AOD nor is the corresponding SWV forcing, among others,15
36	due to hysteresis effects, seasonal dependencies, aerosol profile heights, and feedbacks. However, knowledge of the cold point	38	due to hysteresis effects, seasonal dependencies, aerosol profile heights, and feedbacks. However, knowledge of the cold point
37	temperature increase allows for an estimation of SWV increases with a 12 % increase per Kelvin increase in mean cold point	39	temperature increase allows for an estimation of SWV increases of 12 % per Kelvin increase in mean cold point temperature.
38	temperature, and yearly averages show an approximately linear behaviour in the cold point warming and SWV forcing with	40	For yearly averages power functions are fitted to the cold point warming and SWV forcing with increasing AOD.
39	respect to the AOD.	41	1 Introduction
40	1 Introduction20	42	It has been established that the entry of water vapour into the stratosphere is largely controlled by the temperature of the tropical20
41	It has been established that the entry of water vapour into the stratosphere is largely controlled by the temperature of the tropical		
42	tropopause (e.g, Brewer, 1949; Mote et al., 1996; Fueglistaler et al., 2009; Dessler et al., 2014). Following up on the discussion	43	tropopause (e.g, Brewer, 1949; Mote et al., 1996; Fueglistaler et al., 2009; Dessler et al., 2014). Following up on the discussion
43	of the long term increasing trend in stratospheric water vapor (SWV) observed during the 1980s and 1990s it was proposed	44	of the long term increasing trend in stratospheric water vapor (SWV) observed during the 1980s and 1990s it was proposed
44	that volcanic eruptions could be influencing the SWV budget (e.g, Rosenlof et al., 2001; Joshi and Shine, 2003). Mainly two	45	that volcanic eruptions could be influencing the SWV budget (e.g, Rosenlof et al., 2001; Joshi and Shine, 2003). Mainly two
		46	processes are considered: the direct volcanic injection from the volcanic plume, and an indirect volcanic mechanism due to an
45	1	47	1
46		48	
47	processes are considered: the direct injection from the volcanic plume, and the indirect mechanism due to an increase of the25	49	increase of the tropopause temperature, referred to hereafter as the temperature controlled pathway. The increased SWV levels25
48	tropopause temperature. The increased SWV levels may remain in the stratosphere for more	50	may remain in the stratosphere for more than 5 years (Hall and Waugh, 1997), even though

than 5 years (Hall and Waugh,
49 1997), even though the volcanic aerosols are sedimenting out of the stratosphere within 1 - 3 years (Robock, 2000). However,
50 the magnitude of the SWV increase and the contribution from the different entry mechanisms are still unclear.
51 80 % of the eruption material is water vapour (Coffey, 1996), which could be directly injected into the stratosphere during30
52 an eruption event. But although the SWV originating form the direct injection can be detected shortly after the eruption event,
53 it is a singular event and the corresponding elevated SWV levels are spread in the stratosphere and are not distinguishable from
54 the background SWV anymore. Satellite evidence for the direct injection events exist and is discussed briefly by Schwartz et
55 al. (Schwartz et al. (2013)) and in depth by Sioris et al. (Sioris et al. (2016a), Sioris et al. (2016b)). Based on a model study
56 Joshi and Jones (2009), hypothesized that the environment surrounding the plume can also have a significant impact on the35
57 amount of SWV injected directly.
58 This study will focus on the indirect entry mechanism. In contrast to the direct entry, it can act for months or even years after
59 volcanic eruptions since it depends on the aerosol layer in the stratosphere, and not on the eruption event itself. It is caused by
60 the long wave (LW) heating by the aerosol layer, which leads to increased cold point temperatures. Consequently the saturation40
61 water vapour pressure at the cold point is increased, reducing the "freeze trap" effect originating from the increasingly low
62 temperatures and consequent loss of WV due to ice formation and fallout. The reduced freezing trap character enhances the
63 entry of water vapour into the stratosphere.
64 In an early, idealized study Joshi and Shine (2003) already underlined the importance of the aerosol profile and corresponding
65 LW-heating in the tropopause region. Despite the mechanisms being known, the analysis of the indirect entry mechanism is45
66 still complicated by scarce observational data, since the aerosols have to be in the stratosphere to open the indirect pathway and
67 even if the plume reaches the stratosphere, the amount of sulfur for aerosol formation may be too low, leading to a signal ob-
68 scured by internal variability. If a signal can be observed however, it is only one individual event occurring on the background
69 of natural variability. An assessment of how typical the respective event is, makes a larger amount of data for similar events
70 or ensemble simulations necessary. Additionally, volcanic eruptions fulfilling the criteria needed to open the indirect pathway50
71 may lead to retrieval problems or outages of observational instruments as was the case fo

the volcanic aerosols sediment out
51 of the stratosphere within 1 - 3 years (Robock, 2000). However, the magnitude of the SWV increase and the contribution from
52 the different entry mechanisms are still unclear.
53 80 % of the eruption material is water vapour (Coffey, 1996), which could be directly injected into the stratosphere during30
54 an eruption event. But although the SWV originating form the direct injection can be detected shortly after the eruption event,
55 it is a singular event. The corresponding elevated SWV levels are spread in the stratosphere and are not distinguishable from
56 the background SWV anymore. Satellite evidence for the direct injection events exist. It is discussed briefly by Schwartz et al.
57 (Schwartz et al. (2013)) and in depth by Sioris et al. (Sioris et al. (2016a), Sioris et al. (2016b)). Based on a model study Joshi
58 and Jones (2009), hypothesized that the environment surrounding the plume can also have a significant impact on the amount35
59 of SWV injected directly.
60 This study will focus on the indirect volcanic entry mechanism. In contrast to the direct entry, it can act for months or even
61 years after volcanic eruptions since it depends on the heating caused by the aerosol layer in the stratosphere and not on the
62 eruption event itself. This indirect volcanic entry is caused by the terrestrial long wave and near IR solar heating by the volcanic40
63 aerosol layer which leads to increased cold point temperatures. Consequently, the saturation water vapour pressure at the cold
64 point is increased, thereby reducing the loss of WV due to ice formation and fallout. This mechanism enhances the entry of
65 WV into the stratosphere.
66 In an early, idealized study Joshi and Shine (2003), underlined the importance of the aerosol profile and corresponding heating
67 in the tropopause region. Despite the mechanisms being known, its analysis is still complicated as internal variability and45
68 scarcity of observations has made it difficult to observe it in practice. Additionally, even if SWV increases were recorded,
69 the data usage might be discouraged, as was the case for Mt. Pinatubo by SAGE II, because discrepancies between different
70 satellites could not be satisfactory explained (Fueglistaler et al., 2013).
71 The scarcity of observational data is also reflected in the quality of the available reanalysis products for SWV, the usage of
72 which in general is discouraged in some papers (e.g, Davis et al., 2017) and which sometimes do not implicitly account for the50
73 volcanic forcing at all (Diallo et al. (2017), Tao et al. (2019)). The latter problem wa

r the most important eruption of the
72 last century, Mt. Pinatubo. Even if SWV increases were recorded, as was the case for Mt. Pinatubo by SAGE II, the data usage
73 is discouraged as discrepancies between different satellites can not be satisfactory explained (Fueglistaler et al., 2013).
74 The scarcity of observational data is also reflected in the quality of the available reanalysis products for SWV, whose usage in
75 general is discouraged in some papers (e.g, Davis et al., 2017) and which sometimes do not implicitly account for the volcanic55
76 forcing at all (Diallo et al. (2017), Tao et al. (2019)). The latter problem was also found by Löffler et al. (2016) when compar-
77 ing their simulated SWV increases after the eruption of Mt. Pinatubo with ERAinterim reanalysis. Nevertheless by performing
78 a regression analysis of water vapour entering the stratosphere as simulated by a trajectory model fed by reanalysis input,
79 Dessler et al. (2014) found a SWV peak partially overlapping with the aerosol optical depth (AOD) signal of Mt. Pinatubo.
80 2
81
82 As the SWV increase occurred before the eruption and AOD increase, the question remained if the peak in the residual might60
83 instead be caused by another source of variability. Another possible issue in the analysis was that some of the effects modeled
84 by the regressors are themselves influenced by volcanic eruptions, which may lead to the volcanic signal being attributed to a
85 different mechanism leading to the SWV increase like increases of the Brewer Dobson circulation. In the case of a volcanic
86 eruption these increases in the Brewer Dobson circulation are caused by the LW heating due to the aerosol layer and should be
87 attributed to the volcanic signal consequently. Tao et al. (2019) also undertook an indirect quantification of the SWV increase65
88 after volcanic eruption via another regression analysis. While using a Lagrangian model fed with different reanalysis sources,
89 they explicitly accounted for volcanic source terms. They found a clear volcanic signal in the expected time frame, but the
90 magnitude of the SWV increase was highly variable between the different reanalysis data sources.
91 After entering the stratosphere, the additional SWV affects both stratospheric chemistry with respect to ozone loss (Ro-70
92 brecht et al. (2019), Rosenlof (2018), Tian et al. (2009)) and SO
93 2
94 oxidation (Bekki (1995)) as well as the radiative budget of
95 the entire atmosphere (Solomon et al., 2010). Despite the forcing originating from the additional SWV often being mentioned

s also indicated by Löffler et al. (2016)
74 when discussing SWV increases simulated for the eruption of Mt. Pinatubo. Nevertheless, by performing a regression analysis
75 using a trajectory model fed by reanalysis data, Dessler et al. (2014) identified a SWV peak partially overlapping with the
76 aerosol optical depth (AOD) signal of Mt. Pinatubo. As the SWV increase occurred before the eruption and AOD increase, the
77 question remained if the peak in the residual might instead be caused by another source of variability. Another possible issue in55
78 their analysis is that some of the effects modeled by the regressors are themselves influenced by volcanic eruptions, which may
79 lead to the volcanic signal being attributed to a different source. An example is the increases of the Brewer Dobson circulation
80 (BDC), which, in the case of a volcanic eruption, are attributed to the heating due to the aerosol layer. Tao et al. (2019) also
81 undertook an indirect quantification of the SWV increase after volcanic eruptions via another regression analysis, explicitly
82 2
83
84 accounting for volcanic source terms. They found a clear volcanic signal in the expected time frame, but the magnitude of the60
85 SWV increase differed strongly between the individual reanalysis data sources.
86 After entering the stratosphere, the additional SWV affects both stratospheric chemistry with respect to ozone loss (Robrecht
87 et al. (2019), Rosenlof (2018), Tian et al. (2009)) and SO
88 2
89 oxidation (Bekki (1995), LeGrande et al. (2016)), as well as the
90 radiative budget of the entire atmosphere (Solomon et al., 2010). Despite the forcing originating from the additional SWV often65

96 as a motivation for studies, few studies exist on the forcing effect of the SWV increase after volcanic eruptions. Independent
97 of volcanic eruptions, Forster and Shine (2002) analysed the forcing impact of SWV changes in an artificial SWV profile. In a
98 study focusing on the Mt. Pinatubo eruption, Joshi and Shine (2003) calculated the additional global forcing originating from75
99 the post eruption SWV increase. As this was a side study of their paper, they did not investigate the temporal evolution nor the
100 impact of different eruption strengths. In a study on the direct injection, Joshi and Jones (2009) quantified the LW-component
101 of the SWV forcing indirectly, but their setup did not allow for a quantification of the additional contribution of the indirect
102 pathway. Most recently Krishnamohan et al. (2019) attributed changes in their TOA imbalances for different geoengineering
103 scenarios to a large influence of SWV. However, they did not separate the contributions of aerosol forcing and SWV.80
104 So far, the question remains open what the critical magnitude is for an eruption to have a significant impact on SWV content,
105 what the radiative consequences of the SWV increase are and if these effects can be predicted based on information of the
106 eruption magnitude or AOD. In this study we therefore investigate the changes in stratospheric water vapour originating from
107 the indirect pathway using a large ensemble of coupled climate model simulations with 100 ensemble members each for five85
108 eruption strengths described by changing amount of stratospheric sulfur and a control run, called the EVAens (Azoulay et al.
109 in preparation, 2020). The idealized setup using forcing files generated with the Easy Volcanic Aerosol tool (EVA) offers
110 the unique opportunity of a direct comparison between the different eruption strengths since the date and location of the
111 eruptions are identical and all ensembles have the same set of starting conditions. By comparing with the control run a direct
112 quantification of the SWV increase is possible. The large ensemble size allows us to perform an analysis of the sensitivity of90
113 the increase in stratospheric water vapour to the eruption strength along with its statistical significance. The critical eruption
114 strengths that cause stratospheric water vapour perturbations beyond the internal variability of the model are identified when
115 analyzing the individual ensemble members.
116 3

91 being mentioned as a motivation for studies, few studies exist on the forcing effect of the SWV increase after volcanic erup-
92 tions. Independent of volcanic eruptions, Forster and Shine (2002) analysed the forcing impact of SWV changes in an artificial
93 SWV profile. In a study focusing on the Mt. Pinatubo eruption, Joshi and Shine (2003) calculated the additional global forcing
94 originating from the post eruption SWV increase. As this was a side study of their paper, they did not investigate the temporal
95 evolution nor the impact of different eruption strengths. In a study on the direct injection, Joshi and Jones (2009) quantified the70
96 long wave component of the SWV forcing indirectly, but their setup did not allow for a quantification of the additional contri-
97 bution of the indirect volcanic pathway. Most recently Krishnamohan et al. (2019) attributed changes in their TOA imbalances
98 for different geoengineering scenarios to a large influence of SWV. However, they did not separate the contributions of aerosol
99 forcing and SWV.
100 75
101 So far, the questions remain open what the critical magnitude is for an eruption to have a significant impact on SWV
102 content, what the radiative consequences of the SWV increase are, and if these effects can be predicted based on eruption
103 magnitude or AOD. In this study we therefore investigate the changes in stratospheric water vapour originating from the
104 indirect volcanic pathway using a large ensemble of coupled climate model simulations with 100 ensemble members each for
105 five eruption strengths described by changing amount of stratospheric sulfur and a control run, called the EVAens (Azoulay80
106 et al., 2020). The idealized setup using forcing files generated with the Easy Volcanic Aerosol tool (EVA) offers the unique
107 opportunity of a direct comparison between the different eruption strengths, since the date and location of the eruptions are
108 identical and all ensembles have the same set of starting conditions. By comparing with the control run a direct quantification
109 of the SWV increase is possible. The large ensemble size allows us to perform an analysis of the sensitivity of the increase
110 in stratospheric water vapour to the eruption strength along with its statistical significance. The critical eruption strengths that85
111 cause stratospheric water vapour perturbations beyond the internal variability of the model are identified when analyzing the

117

118 In the following Sect. 2 the model setup of our study and the forcing calculations are described. In Sect. 3 we present results

119 starting with the top of the atmosphere imbalances and the changes in atmospheric temperature profiles. We put a particular95

120 emphasis on the changes in the annual cycle of vertically propagating water vapor signals and the intra-ensemble variability.

121 With a comparison to a historical simulation of Mt. Pinatubo we highlight the importance of the shape and position of the

122 aerosol layer for SWV entry. Finally we discuss the determination of the stratospherically adjusted forcing caused by the ad-

123 ditional water vapour. Sect. 4 is used for discussion of our results in context of earlier studies on stratospheric water vapour

124 changes. Our main findings are summarized in Sect. 5 which also gives an outlook to possible further studies.100

125 2 Methods

126 2.1 The EVAens ensemble and the GE historical simulations

127 This study is based on two sets of large ensemble simulations both covering the time frame of January 1991 to December 1993

128 but using different volcanic forcing data sets: the EVAens (Azoulay et al.in preparation, 2020) and a subset of the Max Planck105

129 Institute Grand Ensemble (MPI-GE) historical simulations (Maher et al., 2019).

130 Both ensemble simulations are performed with the Max Planck Institut - Earth System Model in its low resolution configuration

131 (MPI-ESM-LR) (version, MPI-ESM 1.1.00p2). We apply an intermediate model version between the CMIP5 version (Giorgetta

132 et al., 2013) and the CMIP6 version (Mauritsen et al. (2019)) of the MPI-ESM, with a behaviour more similar to the CMIP6

133 version. The MPI-ESM itself is a coupled model including the atmosphere component ECHAM (version echam-6.3.01p3,110

134 Stevens and Bony (2013)), the land component JSBACH (version jsbach-3.00, Reick et al. (2013), Schneck et al. (2013)), the

135 ocean component MPIOM (version mpiom-1.6.1p1, Marsland et al. (2003), Jungclaus et al. (2013)), and the biogeochemistry

136 component HAMOCC (HAMOCC5.2, Ilyina et al. (2013)).

137 In the model setup the atmosphere is run in a T63L47 configuration corresponding to a horizontal resolution of about 1.9

138 ◦

139 with

140 47 pressure levels up to 0.01hPa. The influe

112 individual ensemble members.

113 2 Methods

114 2.1 The EVAens ensemble and the GE historical simulations

115 This study is based on large ensemble simulations covering the time frame of January 1991 to December 1993 using two dif-90

116 ferent volcanic forcing data sets: the EVAens (Azoulay et al., 2020) and a subset of the Max Planck Institute Grand Ensemble

117 (MPI-GE) historical simulations (Maher et al., 2019).

118 3

119

120 Both ensemble simulations are performed with the Max Planck Institute - Earth System Model in its low resolution config-

121 uration (MPI-ESM-LR) (version, MPI-ESM 1.1.00p2). We apply an intermediate model version between the CMIP5 version

122 (Giorgetta et al., 2013) and the CMIP6 version (Mauritsen et al. (2019)) of the MPI-ESM, with a behaviour closer to the CMIP695

123 version. The MPI-ESM itself is a coupled model including the atmosphere component ECHAM (version echam-6.3.01p3,

124 Stevens and Bony (2013)), the land component JSBACH (version jsbach-3.00, Reick et al. (2013), Schneck et al. (2013)), the

125 ocean component MPIOM (version mpiom-1.6.1p1, Marsland et al. (2003), Jungclaus et al. (2013)), and the biogeochemistry

126 component HAMOCC (HAMOCC5.2, Ilyina et al. (2013)).

127 In the model setup the atmosphere is run in a T63L47 configuration corresponding to a horizontal resolution of about 1.9

128 ◦

129 with100

130 47 pressure levels up to 0.01hPa. The influen

nce of the sponge layer in the uppermost model layer reaches down to a height115

141 of 65kmwith continuously decreasing impact. As the MPI-ESM-LR does not include interactive atmospheric chemistry or

142 aerosols, the volcanic aerosols are prescribed by monthly and zonal mean values of their optical properties - the extinction,

143 single scattering albedo, and asymmetry factor - in 16 long wave and 14 short wave bands. During the simulation these monthly

144 aerosol properties are interpolated linearly in time.

145 In the MPI-GE historical simulations the optical properties of the volcanic aerosols are prescribed with an updated version of120

146 the PADS data set (Stenchikov et al. (1998), Driscoll et al. (2012), Schmidt et al. (2013)). The Mt. Pinatubo eruption in June

147 1991 lies in the investigated time frame.

148 In the EVAens the setup of the MPI-GE historical simulations was kept, changing only the representation of stratospheric

149 aerosols. The respective forcing files were generated using the Easy Volcanic Aerosol (EVA) forcing generator (Toohey et al.

150 (2016)). Six 100 member ensembles are created: A control run with zero sulfur emission only considering the EVA background125

151 aerosol and five volcanic eruption runs (2.5TgS, 5TgS, 10TgS, 20TgS, 40TgS) with EVA background aerosol and with

152 4

153

154 the volcanic eruptions occurring in June 1991 at the equator are considered, each with 100 ensemble members. Each of the

155 these 100 ensemble members was started from one of the different and independent runs of the MPI-GE historical simulations

156 in 1991 (Maher et al. (2019)). Beside the volcanic aerosols, the forcing files include an aerosol background for the industrial-

157 ized period supplied by EVA.130

158 The aerosol optical depths (AOD) for the five different eruption strengths of the EVA forcing along with the PADS Mt.

159 Pinatubo forcing are shown in Fig. 1 for the 550nmwaveband. All EVA aerosol distributions have very similar patterns, but

160 differ in magnitude and the duration of elevated AOD levels. Whereas the 2.5TgS run returns close to background conditions

161 within 3.5 years after the eruption, the 40TgS run only declines to the peak values of the 2.5TgS run within this time.135

162 The PADS data set has a higher background AOD level than the EVA data sets in the months before the eruption. With a sulfur

163 amount of (7.5±2.5)Tg(Timmreck et al., 2018) the Mt. Pinatubo AOD should be comparable to

ce of the sponge layer in the uppermost model layer reaches down to a height

131 of 65kmwith continuously decreasing impact. As the MPI-ESM-LR does not include interactive atmospheric chemistry or

132 aerosols, the volcanic aerosols are prescribed by the monthly zonal mean values of their optical properties - the extinction, sin-

133 gle scattering albedo, and asymmetry factor - in 16 long wave and 14 short wave bands. During the simulation these monthly

134 aerosol properties are interpolated linearly in time.105

135 In the MPI-GE historical simulations the optical properties of the volcanic aerosols are prescribed with an updated version of

136 the PADS data set (Stenchikov et al. (1998), Driscoll et al. (2012), Schmidt et al. (2013)). The Mt. Pinatubo eruption in June

137 1991 lies in the investigated time frame.

138 In the EVAens the setup of the MPI-GE historical simulations was kept, changing only the stratospheric aerosols represen-

139 tation. The respective forcing files were generated using the Easy Volcanic Aerosol (EVA) forcing generator (Toohey et al.110

140 (2016)). Six 100 member ensembles were created: A control run with zero sulfur emission only using the EVA background

141 aerosol and five volcanic eruption runs (2.5TgS, 5TgS, 10TgS, 20TgS, 40TgS) with both the EVA background aerosol

142 and a volcanic eruptions occurring in June 1991 at the equator, each with 100 ensemble members. Every ensemble member

143 was started from one of the different and independent runs of the MPI-GE historical simulations in 1991 (Maher et al. (2019)).

144 Beside the volcanic aerosols, the forcing files include an aerosol background for the industrialized period supplied by EVA.115

145 The aerosol optical depths (AOD) for the five different eruption strengths of the EVA forcing along with the PADS Mt.

146 Pinatubo forcing are shown in Fig. 1 for the 550nmwaveband. All EVA aerosol distributions have very similar patterns, but

147 differ in magnitude and duration of elevated AOD levels. Whereas the 2.5TgS run returns close to background conditions

148 within 3.5 years after the eruption, the 40TgS run only declines to the peak values of the 2.5TgS run within this time.120

149 The PADS data set has a higher background AOD level than the EVA data sets in the months before the eruption. With a

150 sulfur amount of (7.5±2.5)Tg(Timmreck et al., 2018) the Mt. Pinatubo AOD should be comparab

the 5Tgand 10TgEVA
164 data set. Generally, the AOD in the PADS data set does not spread as fast to higher latitudes after the eruption. Additionally,
165 the AOD values tend to be slightly higher than the values for the 10TgEVA data set and persist for a longer time at elevated
166 levels.140
167 5
168
169 Figure 1.Aerosol optical depth (AOD) for the five volcanically perturbed EVAens runs (2.5TgS, 5TgS, 10TgS, 20TgS and 40Tg
170 S) and the PADS Mt. Pinatubo compilation. The time evolution of the zonal average AOD is shown for all latitudes considering the 550nm
171 waveband (441nm- 625nm).)
172 6
173
174 As we merely prescribe the aerosols only the indirect pathway and not the direct injection is simulated in the EVAens.
175 In the following work all anomalies will be defined as the value difference between the volcanically perturbed (P) and unper-
176 turbed 0TgS (U) ensemble means (P-U).
177 2.2 Stratospherically adjusted clear sky forcing calculations
178 The stratospherically adjusted clear sky radiative forcing originating from the increase of stratospheric water vapour due to the145
179

le to the 5Tgand 10Tg
151 EVA data set (compare Table 1). Generally, the AOD in the PADS data set does not spread as fast to higher latitudes after the
152 eruption. Additionally, the AOD values tend to be slightly higher than the values for the 10TgEVA data set and persist longer
153 at elevated levels.125
154 4
155
156 Figure 1.Time evolution of the aerosol optical depth (AOD) for the five volcanically perturbed EVAens runs (2.5TgS, 5TgS, 10TgS,
157 20TgS and 40TgS) and the PADS Mt. Pinatubo compilation. The zonal average AOD is shown for all latitudes considering the 550nm
158 waveband (441nm- 625nm).)
159 5
160
161 Table 1.List of tropical volcanic eruptions with location, eruption date, and estimated amount of emitted sulfur. The sulfur amounts in
162 parentheses represent the best estimates.
163 volcanolocationeruption timeemitted S [Tg]reference
164 Mt Agung8
165 ◦
166 S, 115
167 ◦
168 E17 Mar 19632.5-5 (3.5)Timmreck et al. (2018) and references therein
169 El Chichón17
170 ◦
171 N, 93
172 ◦
173 W4 Apr 19822.5-5 (3.5)Timmreck et al. (2018) and references therein
174 Mt Pinatubo15
175 ◦
176 N, 120
177 ◦
178 E15 Jun 19915-10 (7)Timmreck et al. (2018) and references therein
179 Tambora8
180 ◦
181 S, 117
182 ◦
183 EApril 181515-40 (30)Marshall et al. (2018) and references therein
184 As we merely prescribe the aerosols, only the indirect volcanic pathway, which includes cold point warming and overshoot-
185 ing convection, and not the direct injection or aerosol chemistry is simulated in the EVAens.
186 In the following work all anomalies will be defined as the value difference between the volcanically perturbed (P) and unper-
187 turbed 0TgS (U) ensemble means (P-U).
188 2.2 Stratospherically adjusted clear sky forcing calculations130
189 The stratospherically adjusted clear sky radiative forcing, as defined by Hansen et al. (2005) originating from the increase of
190

179 indirect pathway (i.e. via tropopause warming by the aerosol layer) is calculated using the 1D radiative convective equilibirum

180 (RCE) model konrad (Kluft et al. (2019), Dacie et al. (2019)). Konrad is designed to represent the tropical atmosphere. It uses

181 the Rapid Radiative Transfer Model for GCMs (RRTMG) and a simple convective adjustment that fixes tropospheric tempera-

182 tures according to a moist adiabat.

183 For each eruption strength, including the 0TgS eruption, the ensemble mean of the clear sky humidity profile in the tropical150

184 region [-5,5]

185 ◦

186 latitude is determined from the EVAens output. In order to compute the adjusted radiative forcing due to the

187 increased SWV, the difference between the fluxes in an equilibrated atmosphere with and without the additional SWV must

188 be calculated. To determine the equilibrated reference without additional SWV konrad is run to equilibrium with the humidity

189 profile and chemical composition of the atmosphere fixed to the values from the 0TgS runs. The final surface temperature

190 lies between [298-301]K. Starting from this equilibrium state, only the SWV profile is replaced by the volcanically perturbed155

191 EVAens humidity profile and a new equilibrium is calculated while keeping the surface temperature fixed, but allowing for the

192 temperature above the convective top to adjust. Using both equilibrium states the adjusted SWV forcing is determined. The

193 corresponding instantaneous forcing can be calculated by calculating the flux changes without running the perturbed atmo-

194 sphere into equilibrium.

195 For the all sky case the contribution of clouds is investigated by additionally taking a 20 % fraction of high level clouds between160

190 stratospheric water vapour due to the indirect volcanic pathway (i.e. via tropopause warming by the aerosol layer), is calcu-

191 lated using the 1D radiative convective equilibirum (RCE) model konrad (Kluft et al. (2019), Dacie et al. (2019)). Konrad is

192 designed to represent the tropical atmosphere. It uses the Rapid Radiative Transfer Model for GCMs (RRTMG) and a simple

193 convective adjustment that fixes tropospheric temperatures up to the convective top according to a moist adiabat, whereas the135

194 temperatures in the higher atmospheric levels are determined by radiative-dynamical equilibrium. Being a 1D model, konrad

195 employs a highly parameterized "circulation", i.e. an upwelling term constant in time which only causes adiabatic cooling. As

196 the temperature above the convective top can adjust, this does not mean that the dynamical heating is fixed (compare Fels et al.

197 (1980)).

198 In order to calculate the stratospherically adjusted SWV forcing, the following procedure is employed: For each eruption 140

199 strength, including the 0TgS eruption, the ensemble mean of the clear sky humidity profile in the tropical region [-5,5]

200 ◦

201 lati-

202 tude is determined from the EVAens output. In order to compute the adjusted radiative forcing due to the increased SWV, the

203 difference between the fluxes in an equilibrated atmosphere with and without the additional SWV must be calculated. To deter-

204 mine the equilibrated reference without additional SWV, konrad is run to equilibrium with the humidity profile and chemical

205 composition of the atmosphere fixed to the values from the 0TgS runs. The final surface temperature lies between [298-301]145

206 K. Starting from this equilibrium state, only the SWV profile is replaced by the volcanically perturbed EVAens humidity pro-

207 file and a new equilibrium is calculated while keeping the surface temperature fixed, but allowing for the temperature above

208 the convective top to adjust. Using both equilibrium states, the adjusted SWV forcing is determined from the flux differences

209 at the top of the atmosphere. The corresponding instantaneous forcing as defined by Hansen et al. (2005) is calculated as the

210 difference of tropopause fluxes obtained without running the perturbed atmosphere into equilibrium.150

211 6

212

213 For the all sky case the contribution of clouds is investigated by additionally taking a 2

			0 % fraction of high level clouds between
196	200hPaand 300hPainto consideration, while low levels clouds are considered in the albedo settings.	214	200hPaand 300hPainto consideration, while low levels clouds are considered in the albedo settings.
197	In order to relate the SWV forcing to the aerosol forcing, the instantaneous aerosol forcing is calculated using a double	215	In order to relate the SWV forcing to the aerosol forcing, the instantaneous aerosol forcing is calculated using a double
198	radiation call in the MPI-ESM. For the double radiation call fluxes are calculated for each time step once using the atmospheric	216	radiation call in the MPI-ESM. For the double radiation call fluxes are calculated for each time step once using the atmospheric155
199	conditions with and without aerosol. The stratospheric background aerosol is corrected for by additionally calculating the forc-165	217	conditions with and without aerosol. The stratospheric background aerosol is corrected for by additionally calculating the forc-
200	ing for the corresponding 0TgS run and subtracting it.	218	ing for the corresponding 0TgS run and subtracting it.
201	7	219	7
202		220	
203	3 Results	221	3 Results
204	3.1 Effects on the time evolution of TOA radiative imbalance and surface temperature	222	3.1 Effects on the time evolution of TOA radiative imbalance and surface temperature160
205	In order to connect the amount of emitted sulfur to its impact on the energy budget of the system, we analyze the top of170	223	In order to relate the amount of emitted sulfur to its impact on the energy budget of the system, we analyze the top of the
206	the atmosphere (TOA) radiative imbalance after the volcanic eruptions as well as changes in surface temperature. The era of	224	atmosphere (TOA) radiative imbalance after the volcanic eruptions as well as changes in surface temperature. The time span of
207	negative global radiative TOA imbalance after the volcanic eruptions during which more energy leaves the earth-atmosphere	225	negative global radiative TOA imbalance after the volcanic eruptions, during which more energy leaves the earth-atmosphere
208	system than is taken up lasts between 17 and 28 months (Fig. 2). For the lower emissions (2.5TgS - 5TgS) the standard error	226	system than is taken up lasts between 17 and 28 months (Fig. 2). For the lower emissions (2.5TgS - 5TgS) the standard error
209	of the negative TOA imbalance permanently overlaps with zero imbalance. Consequently single ensemble members of the 0	227	of the negative TOA imbalance permanently overlaps with zero imbalance. Consequently single ensemble members of the 0165
210	TgS can produce similar signals as these lower emission run due to internal variability. Overall the TOA imbalance exhibits a175	228	TgS can produce similar signals to these lower emission runs due to internal variability. Overall the TOA imbalance exhibits
211	roughly linear relationship with respect to the emitted sulfur mass (see Fig. A).	229	a roughly linear relationship with respect to the emitted sulfur mass (see Fig. A).
212	As a consequence of the negative TOA imbalance the surface temperatures of the volcanically perturbed runs decrease. The	230	As a consequence of the negative TOA imbalance the surface temperatures in the volcanically perturbed runs decrease. The
213	range of ensemble mean global temperature decrease for the EVAens is -[0.09,1.30]K, when it is at its maximum (Fig. 2). As	231	range of ensemble mean maximum global temperature decrease for the EVAens is -[0.09, 1.30] K(Fig. 2). As the surface
214	the surface temperature follows the TOA imbalance, the change in surface temperature is also roughly linear with respect to	232	temperature follows the TOA imbalance, the change in surface temperature is also roughly linear with respect to the emitted S170
215	the emitted S mass (Fig. A).180	233	mass (Fig. A).
		234	Figure 2.Global top of the atmosphere (TOA) radiative imbalance and anomaly of surface temperature in the five volcanically perturbed
		235	EVAens runs (2.5TgS, 5TgS, 10TgS, 20TgS and 40TgS). For each run the standard errors of the mean are shown as shading. The
		236	vertical blue line marks the eruption time. The plots also show the values for the MPI-GE historical simulations (PADS).
216	8	237	8
217		238	
218	Figure 2.Global top of the atmosphere (TOA) radiative imbalance and anomaly of surface temperature in the five volcanically perturbed	239	In addition to the EVAens results, the TOA imbalance and global surface temperature changes for the historical Mt Pinatubo
219		240	

EVAens runs (2.5TgS, 5TgS, 10TgS, 20TgS and 40TgS). For each run the standard errors of the mean are shown. The vertical blue
220 line marks the eruption time. The plots also show the values for the MPI-GE historical simulations (PADS).
221 3.2 Effects on the cold point temperature
222 The effect of the volcanic forcing on the inner tropical temperature profile is visualized in Fig. 3 showing the EVAens ensemble
223 mean of the temperature profiles three months after the eruption. The month of September was chosen as an example since it
224 lies in the time frame within which the annual cycle of water vapour entry into the stratosphere is enhanced. Due to the location
225 of the aerosol peak above the cold point
226 1
227 the largest temperature changes occur in the lower stratosphere with increases up to185
228 24Kin the 40TgS ensemble. The cold point warming reaches maximum values of 8Kin the case of the 40TgS run. Also
229 visible in the figure is the downwards shift of the cold point with increasing sulfur burden caused by the stratospheric warming
230 leading to a downwards shift in the cold point levels. This effect is amplified by tropospheric cooling (compare surface cooling
231 in Fig. 2) due to the backscattering of solar radiation through the volcanic aerosols.
232 190
233 1
234 Point of lowest temperature between troposphere and stratosphere.

eruption are shown in black in Figure 2. The global TOA imbalance peaking at -2.4Wm
241 −2
242 compares favourably with the
243 approximately -3Wm
244 −2
245 from Earth Radiation Budget Satellite observations (Soden et al., 2002) when considering the standard
246 deviations of our ensemble. The average surface cooling between June 1991 and December 1993 is -0.26Kwith a maximum175
247 cooling of -0.36K. Compared to the cooling documented by satellite measurements from the microwave sounding unit (MSU)
248 of -0.3Kbetween June 1991 and December 1995 after ENSO removal and reached a peak cooling of -0.5K(Soden et al.,
249 2002).
250 3.2 Effects on the cold point temperature
251 In the following analysis the cold point is defined as the lowest temperature between troposphere and stratosphere lying on180
252 full pressure levels of model output remapped to a vertical spacing of 10hPain the tropical tropopause region. The associated
253 errors in average cold point temperature lie below one percent of the respective cold point temperature.
254 The effect of the volcanic forcing on the inner tropical temperature profile is visualized in Fig. 3 showing the EVAens ensemble
255 mean of the temperature profiles three months after the eruption. The month of September was chosen as an example since it
256 lies in the season of relatively large water vapour entry into the stratosphere due to maximum cold point temperature in boreal185
257 autumn and winter. Due to the location of the aerosol peak above the cold point the largest temperature changes occur in the
258 lower stratosphere, with increases up to 24Kin the 40TgS ensemble. The cold point warming reaches maximum values of
259 8Kin the case of the 40TgS run. Also visible in the figure is the downwards shift of the cold point with increasing sulfur
260 burden caused by the stratospheric warming. This effect is amplified by tropospheric cooling (compare surface cooling in Fig.
261 2) due to the back scattering of solar radiation through the volcanic aerosols.190

235 9

236

237 Figure 3.Inner tropical average of the temperature profiles for the five volcanically perturbed EVAens runs (2.5TgS, 5TgS, 10TgS, 20

238 TgS and 40TgS) in September 1991 (three months after the eruption). The temperature of the respective cold point (CP) is indicated in

239 the legend. The solid line represents the ensemble mean, the error bars symbolize the ensemble standard deviations. The 550nmextinction

240 values for the 10TgS aerosol profile are shown with a dashed black line.

241 Figure 4 shows changes in the temporal evolution of the surface temperature (a), cold point temperature (b), and 100hPa

242 specific humidity (c) with respect to the zero emission run in the inner tropics. Whereas the surface temperature decreases due

243 10

244

245 Figure 4.Temporal evolution of the inner tropical mean anomaly in surface temperature, cold point temperature, and stratospheric water

246 vapour between volcanically perturbed and unperturbed ensemble runs. The ensemble means are shown with their standard errors. The time

247 of the volcanic eruption is indicated by a vertical blue line.

248 to the scattering of incoming shortwave radiation by the volcanic aerosols, the cold point temperature increases due to warming

249 of the tropopause layer by absorption of longwave radiation by the aerosol layer.

250 When considering the ensemble mean and its standard error even the temperature changes of the low emission runs (2.5195

251 TgS and 5TgS) are significantly different from the control run for short periods of time. With maximum values of 1.93

252 Kthe ensemble mean for the cold point temperature of the 10TgS is the lowest sulfur emission to reach a mean warming

253 above the control run standard deviations

254 2

255 , cold point values below this range could be found for single control run ensemble

256 members due to internal variability. The higher emission scenarios (20TgS and 40TgS) hav

262 9

263

264 Figure 3.Inner tropical average of the temperature profiles for the five volcanically perturbed EVAens runs (2.5TgS, 5TgS, 10TgS, 20

265 TgS and 40TgS) in September 1991 (three months after the eruption). The temperature of the respective cold point (CP) is indicated in

266 the legend. The solid line represents the ensemble mean, the error bars symbolize the ensemble standard deviations. The 550nmextinction

267 values for the 10TgS aerosol profile are shown with a dashed black line.

268 Figure 4 shows the temporal evolution of the surface temperature (a), cold point temperature (b), and 70hPaspecific

269 humidity (c) as the difference to to the zero emission run in the inner tropics. Whereas the surface temperature decreases due to

270 the scattering of incoming shortwave radiation by the volcanic aerosols, the cold point temperature increases due to warming

271 of the tropopause layer by absorption of terrestrial long wave and solar near-IR radiation by the aerosol layer.195

272 10

273

274 Figure 4.Temporal evolution of the inner tropical mean anomaly in surface temperature, cold point temperature, and stratospheric water

275 vapour between volcanically perturbed and unperturbed ensemble runs. The ensemble means are shown with their standard errors. The time

276 of the volcanic eruption is indicated by a vertical blue line.

277 When considering the ensemble mean and its standard error even the temperature changes of the low emission runs (2.5

278 TgS and 5TgS) are significantly different from the control run for short periods of time. With a maximum value of 0.99

279 Kthe ensemble mean for the cold point temperature of the 5TgS is the lowest sulfur emission to reach a mean warming

280 above the control run standard deviations, which are ten times larger than the standard error for our 100 member ensemble and

281 reach maximum values of 0.73K. Cold point values below the standard deviation range could be found for single control run200

282 ensemble members due to internal variability. The higher emission scenarios (20TgS and 40TgS) have longer periods with

283 cold point temperature changes above the normally observed internal variability. In addition the 40TgS emission group shows

284 a second amplification peak of the yearly cycle of the cold point temperatures between May and September 1992.

285 The monthly changes in cold point temperature are not linear, neither with respect to emitt

e longer periods with cold point
257 temperature changes above the normally obser
ved internal variability.200
258 In addition the higher emission group shows
a second amplification peak of the yearly cy
cle of the cold point temperatures
259 between May and September 1992. In both the
20TgS and the 40TgS this second peak is red
uced with respect to the first
260 peak.
261 The monthly changes in cold point temperatur
e are not linear either with respect to emit
ted sulfur mass or to AOD in the205
262 prominent IR waveband (compare Fig. A3, A4).
Additionally the transient behaviour of the
volcanic forcing - the build up
263 2
264 The standard deviations are ten times larger
than the standard error in case of our 100 m
ember ensemble.
265 11
266
267 Figure 5.Yearly averages of adjusted tropica
l cold point temperature increases as a func
tion of tropical AOD (IR, 8475-9259 nm) for
the
268 three examined years (1991-1993). A second o
rder fit for each year with corresponding eq
uation is shown.
269 phase in 1991, the approximately constant fo
rcing in 1992, and the declining phase in 19
93, which also differs between the
270 individual eruption sizes - leads to a hyste
retic behaviour. Nevertheless, when partitio
ning the cold point warmings into these
271 individual phases, a linear relationship bet
ween IR-AOD and cold point warming arises fo
r the years 1991 and 1992 (Fig. 5).
272 In 1993 this relationship becomes quadratic
since the lower eruption strength are alrea
dy ceasing to warm the TTL region.210
273 The found cold point temperature increases a
re accompanied by an increase in the saturat
ion water vapour pressure, reducing
274 the "freeze trap" drying in the cold point r
egion. The increased saturation water vapour
pressure enables more water vapour to
275 enter the lower stratosphere as shown in Fi
g. 4 (c)
276 3
277 . Consequently the evolution of the addition
al stratospheric water vapour
278 closely follows the evolution of the cold po
int temperature changes.215
279 3.3 Effects on the tape recorder signal
280 The annual cycle of the tropical SWV is ofte
n described as the tape recorder signal (Mot
e et al., 1996): The variations of
281 the tropical cold point temperatures control

ed sulfur mass nor to AOD in205
286 the prominent IR waveband (compare Fig. A3, A
4). Additionally the transient behaviour of t
he volcanic forcing - the build
287 up phase in 1991, the approximately constant
forcing in 1992, and the declining phase in
1993, which also differ between
288 the individual eruption sizes - leads to a hy
steretic behaviour. Nevertheless, when partit
ioning the cold point warmings into
289 these individual phases, the dependence of th
e cold point warming on IR-AOD can be describ
ed with a power function of form
290 11
291
292 Figure 5.Yearly averages of adjusted tropical
cold point temperature increases as a functio
n of tropical AOD (IR, 8475-9259 nm) for the
293 three examined years (1991-1993). A power fun
ction fit of formaAOD
294 b
295 +cfor each year with corresponding equation i
s shown.
296 aAOD
297 b
298 +cfor the years 1991, 1992 and 1993 (Fig. 5).
210
299 The found cold point temperature increases ar
e accompanied by an increase in the saturatio
n water vapour pressure, reducing
300 the "freeze trap" drying in the cold point re
gion. The increased saturation water vapour p
ressure enables more water vapour
301 to enter the lower stratosphere as shown in F
ig. 4 (c). Consequently, the evolution of the
additional stratospheric water vapour
302 closely follows the evolution of the cold poi
nt temperature changes.215
303 3.3 Effects on the tape recorder signal
304 The manifestation of the annual cycle of the
tropical SWV as seen in the vertical profile
is often described as the tape recorder
305 signal (Mote et al., 1996): The variations of

ling the water vapour entry into the stratosphere via the saturation water vapour
282 pressure is imprinted on the stratospheric water vapour as music is imprinted on a tape. This leads to an annual cycle of bands
283 of high and low water vapour content propagating upwards in the stratosphere with the Brewer Dobson Circulation.220
284 As the LW-heating by the volcanic aerosols will lead to increased tropical cold point temperatures, an increase of stratospheric
285 water vapour is expected after volcanic eruptions with stratospheric aerosols. As shown in Sect. 3.2 the specific humidity shows
286 an enhancement which does not stay constant and then declines but has an annual cycle like the tape recorder. In the following
287 section we will investigate the seasonal cycle more closely.
288 Figure 6 and 7 show absolute SWV and the differences in the SWV content above 140hPawith respect to the control run225
289 for all five eruption strengths. The maximum increases are found in the eruption year itself ranging from 0.1ppmmfor the
290 2.5TgS eruption to 3.5ppmmfor the 40TgS eruption. This corresponds to 5 %, respectively 160 %, of the unperturbed
291 3
292 Here and in the following analysis we report the specific humidity as mass of water vapour per mass of moist air inppmmvalues.
293 12
294
295 SWV values (see Fig. B1). The larger the eruption, the earlier the increase becomes visible and is significant. The additional
296 SWV also follows the annual cycle of the tape recorder (Mote et al., 1996), showing maxima in the SWV enhancement around
297 September which then propagate upwards. This seasonal variation is also apparent in the behaviour of the tropopause and cold230
298 point heights: for scenarios with at least 10TgS onward the tropopause pressures and cold point pressures are higher in the
299 northern hemispheric autumn; whereas in the 20TgS and 40TgS runs the volcanic forcing leads to higher pressure levels of
300 the cold point from September 1991 until the end of 1992 accompanied by a seasonal signal in the SWV anomalies. Allowing
301 for more water vapour to transit to the lower stratosphere.
302 In the 40TgS ensemble mean the second seasonal cycle is associated with an upward propag

the tropical cold point temperatures controlling the water vapour entry into the
306 stratosphere via the saturation water vapour pressure is imprinted on the stratospheric water vapour as music is imprinted on
307 a tape. This leads to an annual cycle of bands of high and low water vapour content propagating upwards in the stratosphere220
308 with the BDC.
309 As the heating by the volcanic aerosols will lead to increased tropical cold point temperatures, an increase of stratospheric water
310 vapour is expected after volcanic eruptions with stratospheric aerosols. As shown in Sect. 3.2 the specific humidity shows an
311 enhancement which does not stay constant and then declines but has an annual cycle like the tape recorder. In the following
312 section we will investigate the seasonal cycle more closely.225
313 Figure 6 and 7 show absolute SWV and the differences in the SWV content above 140hPawith respect to the control run for
314 all five eruption strengths. The maximum increases are found in the eruption year itself ranging from 0.1ppmvfor the 2.5Tg
315 S eruption to more than 5ppmvfor the 40TgS eruption. This corresponds to 5 %, respectively 160 %, of the unperturbed
316 SWV values (see Fig. B1). The larger the eruption, the earlier the increase becomes visible and is significant. The additional
317 SWV also follows the annual cycle of the tape recorder (Mote et al., 1996), showing maxima in the SWV enhancement around230
318 September which then propagate upwards. This seasonal variation is also apparent in the behaviour of the tropopause and cold
319 12
320
321 point heights: for scenarios with at least 10TgS onward the tropopause pressures and cold point pressures are higher in the
322 northern hemispheric autumn; whereas in the 20TgS and 40TgS runs the volcanic forcing leads to higher pressure levels of
323 the cold point from September 1991 until the end of 1992 accompanied by a seasonal signal in the SWV anomalies. Allowing
324 for more water vapour to transit to the lower stratosphere.235
325 In the 40TgS ensemble mean the second seasonal cycle is associated with an upward propagat

ation of the SWV increases above235 3ppmmto even lower atmospheric pressures than in the preceding year. This behaviour can be attributed to the persistence of the high levels of aerosols in combination with the already enhanced SWV levels due to the presence of the volcanic aerosol in the previous year, the additional warming caused by the SWV and the lower lying cold point.

Shortly after the eruption a decrease in water vapour content above 50hPais visible. At these altitudes SWV increases with height due to its production by methane oxidation

4

. Heating in the aerosol layer below leads to lofting of air parcels, bringing240

lower humidity air upwards to higher altitudes, where it causes a net reduction in humidity. However, this effect never exceeds 0.5 ppmm and is offset as soon as the lifted air becomes more moist due to enhanced SWV entry through the tropopause region.

4

The MPI-ESM uses a parameterized methane oxidation scheme (Schmidt et al., 2013).

13

Figure 6.Tropical average in [-23,23]° latitude of WV above 140hPafor sulfur injections of 2.5TgS, 5TgS, 10TgS, 20TgS and 40TgS as well as the PADS dataset. The WMO-tropopause pressure is indicated by a black line, the cold point pressure is shown as black dashed line. Absolute values are shown. In regions not covered by black crosses statistical significant difference between water vapour values of the perturbed and unperturbed runs (t-test at p=0.05) were found.

14

Figure 7.Tropical average in [-23,23]° latitude of WV anomalies above 140 hPa for the sulfur injections of 2.5TgS, 5TgS, 10TgS, 20TgS and 40TgS as well as the PADS dataset. The lowermost panel shows the MPI-GE historical simulations for Mt. Pinatubo using the PADS forcing data set discussed in Sect. 3.5. The WMO-tropopause pressure is indicated by a black line, the cold point pressure is shown as black dashed line. Differences with respect to the 0TgS control run are shown. In regions not covered by black crosses statistical significant difference between water vapour values of the perturbed and unperturbed runs (t-test at p=0.05) were found.

ion of the SWV increases above 4ppmvto even lower atmospheric pressures than in the preceding year. This behaviour can be attributed to the persistence of the high levels of aerosols in combination with the already enhanced SWV levels due to the presence of the aerosol in the previous year, the additional warming caused by the SWV and the lower lying cold point.

Shortly after the eruption a decrease in water vapour content above 50hPais visible. At these altitudes SWV increases with240 height due to its production by methane oxidation, which is parameterized in the MPI-ESM (Schmidt et al., 2013). Heating in the aerosol layer below leads to lofting of air parcels, bringing lower humidity air upwards to higher altitudes, where it causes a net reduction in humidity. However, this effect never exceeds 0.8 ppmv and is offset as soon as the lifted air becomes more moist due to enhanced SWV entry through the tropopause region.

13

Figure 6.Tropical average in [-23,23]° latitude of WV above 140hPafor sulfur injections of 2.5TgS, 5TgS, 10TgS, 20TgS and 40TgS as well as the PADS dataset. The WMO-tropopause pressure is indicated by a black line and the cold point pressure is shown as black dashed line. Absolute values are shown. In regions not covered by black crosses statistical significant difference between water vapour values of the perturbed and unperturbed runs (t-test at p=0.05) were found.

14

Figure 7.Tropical average in [-23,23]° latitude of WV anomalies above 140 hPa for the sulfur injections of 2.5TgS, 5TgS, 10TgS, 20TgS and 40TgS as well as the PADS dataset. The lowermost panel shows the MPI-GE historical simulations for Mt. Pinatubo using the PADS forcing data set discussed in Sect. 3.5. The WMO-tropopause pressure is indicated by a black line and the cold point pressure is shown as black dashed line. Differences with respect to the 0TgS control run are shown. In regions not covered by black crosses statistical significant difference between water vapour values of the perturbed and unperturbed runs (t-test at p=0.05) were found.

	15		15
328		348	
329	3.4 Intra-ensemble variability	349	3.4 Intra-ensemble variability245
330	The investigation of single ensemble members is of interest because - unlike for the ensemble mean - their physical character-245	350	The investigation of single ensemble members is of interest because - unlike for the ensemble mean - their physical character-
331	istics would actually be "observable". Figure 8 shows seasonal averages of the specific humidity at the cold point as a function	351	istics would actually be "observable". Figure 8 shows seasonal averages of the specific humidity at the cold point as a function
332	of cold point temperature for the inner tropical average in 1991 and 1992. The season SON was chosen as it is the period of	352	of cold point temperature for the inner tropical average in 1991 and 1992. The season SON was chosen, as follows the period
333	highest SWV values in the lower stratosphere. Each point in Fig. 8 symbolizes one single ensemble member. Orange crosses	353	where the entry value of stratospheric water vapor is highest, leading to the highest SWV in the tropical lower stratosphere.
334	show the theoretically possible maximum specific humidity curve calculated as the saturation humidity for water vapour over	354	Each point in Fig. 8 symbolizes one single ensemble member. Orange crosses show the theoretically possible maximum spe-250
335	ice at the average cold point temperature, using the corresponding exact solution for the Clausius Clapeyron equation by Mur-250	355	cific humidity curve calculated as the saturation humidity for water vapour over ice at the average cold point temperature,
336	phy and Koop (2005)	356	using the corresponding exact solution for the Clausius Clapeyron equation by Murphy and Koop (2005)
337	5	357	1
338	. An approximation for the Clausius Clapeyron Equation at this temperature range with an 12%increase	358	. An approximation
339	of specific humidity perKis shown with a dashed grey line. This approximation is based on the assumption that the atmo-	359	for the Clausius Clapeyron Equation at this temperature range with an 12%increase of specific humidity perKis shown with
340	sphere around the cold point is saturated in the 0TgS run. The single ensemble members follow the slopes and values of	360	a dashed grey line. This approximation is based on the assumption that the atmosphere around the cold point is saturated in the
341	the calculated water vapour amount at saturation in the inner tropics ([-5,5]	361	0TgS run. The single ensemble members follow the slopes and values of the calculated water vapour amount at saturation in255
		362	the inner tropics ([-5,5]
342	◦	363	◦
343	latitude). Next to water vapour, sublimated lofted	364	latitude). In addition to water vapour, sublimated lofted ice can also contribute to the SWV leading
344	ice can also contribute to the SWV leading to specific humidity values exceeding the ones expected when calculating the the255	365	to higher values than expected based on the saturation water vapour alone. These overshooting events are considered in our
345	SWV based on the saturation water vapour alone as soon as temperatures rise again and the saturation water vapour increases.	366	convective parameterisation (Möbis and Stevens, 2012). However, at the location of the cold point, this lofted ice would still
346	However at the location of the cold point this lofted ice would still be in the ice state and not accounted for in the specific	367	be in the ice state and not accounted for in the specific humidity term.
347	humidity term.	368	In the inner tropics the simulated specific humidity values agree nicely with those from the Clausius Clapeyron equation and260
348	In the inner tropics the found specific humidity values agree nicely with the values from the Clausius Clapeyron equation and	369	its approximated form of a 12 % increase of SWV per Kelvin. Considering the simplification of taking the average cold point
349	its approximated form of a 12 % per Kelvin. However this is only true for the inner tropics. For the entire tropics values up260	370	temperatures between [-5,5]
350	to around 1ppmmlower could be found (compare Fig. C1). As the cold point temperatures increase with increasing latitude,	371	◦
351	higher specific humidity values would be expected at higher latitudes. However, the wate	372	latitude instead of the minimum cold point temperatures this agreement is surprisingly goo

r vapour enters the tropics mainly in
352 the inner tropical region and then spreads throughout the globe, leading to values lower than expected according to the Clausius
353 Clapeyron equation.

354 265
355 In the first SON after the eruption in 1991 the SWV values and cold point temperatures are larger than in 1992. Additionally
356 the separation by sulfur content is more pronounced in 1991. Table 1 lists the number of individual ensemble members lying
357 within two standard deviations of the temperature or humidity value spread of the control run. In 1991 several individual group
358 members up to the 5TgS run overlap with the control run spread as far as cold point temperature changes and humidity
359 values are concerned. The 10TgS run marks the emission strength at which single ensemble members start to be significantly270
360 different from the control run as only one ensemble member can not be distinguished from the control run using only the
361 temperature values. The 20TgS run is the first emission strength with no control run overlap and for which all individual
362 ensemble members show significant increase of SWV content and cold point temperatures in 1991 and even in 1992 where the
363 lower emission runs start to show an increasing number of ensemble members overlapping with the control run values again.
364 This analysis shows the difficulty to register the SWV increases in observational data which collect data for a single realisa-275
365 5
366 Formula (7) giving the saturation water vapour pressure above icep
367 ice
368 asp
369 ice
370 =exp(9.550426−5723.265/T+ 3.53068ln(T)−0.00728332T)is
371 used. This formula is valid for temperatures above 110K. With the knowledge ofp
372 ice
373 and the respective total pressure a calculation of the specific humidity
374 is possible.

d.
373 Oman et al. (2008), for example, only found good agreement when considering the minimum cold point temperatures in the
374 tropics. As they were analyzing a band between [-10,10]
375 ◦
376 latitude, a factor contributing to the difference to our result could
377 be that the main contribution to dehydration of air parcels during their horizontal motion in the vertical ascent takes place in265
378 the inner tropics (Schoeberl and Dessler, 2011), the region to which our study is restricted. Consistent with this analysis is the
379 stronger discrepancy of approx 1.6 ppmv between the SWV values predicted using the Clausius Clapeyron equation and the
380 SWV output by the model when averaging over the entire tropics (compare Fig. C1).
381 In the first SON after the eruption in 1991 the SWV values and cold point temperatures are larger than in 1992. Additionally270
382 the separation by sulfur content is more pronounced in 1991. Table 2 lists the number of individual ensemble members lying
383 within two standard deviations of the temperature or humidity value spread of the control run. In 1991 several individual group
384 members up to the 5TgS run overlap with the control run spread as far as cold point temperature changes and humidity
385 values are concerned. The 10TgS run marks the emission strength at which single ensemble members start to be significantly
386 different from the control run as only one ensemble member can not be distinguished from the control run using only the275
387 temperature values. The 20TgS run is the first emission strength with no control run overlap and for which all individual
388 1
389 Formula (7) giving the saturation water vapour pressure above icep
390 ice
391 asp
392 ice
393 =exp(9.550426−5723.265/T+ 3.53068ln(T)−0.00728332T)is
394 used. This formula is valid for temperatures above 110K. With the knowledge ofp
395 ice
396 and the respective total pressure a calculation of the specific humidity
397 is possible.

375	16	398	16
376		399	
377	tions of a volcanic eruption. Although signals similar to those of individual eruptions can be produced by internal variability	400	ensemble members show significant increase of SWV content and cold point temperatures in 1991 and even in 1992 where the
		401	lower emission runs start to show an increasing number of ensemble members overlapping with the control run values again.
		402	This analysis shows the difficulty to register the SWV increases in observational data which collect data for a single realisa-
		403	tions of a volcanic eruption. Although signals similar to those of individual eruptions can be produced by internal variability280
378	of an unperturbed scenario, our larger ensemble size allows to extract a robust signal which may be obscured in one single	404	of an unperturbed scenario, our larger ensemble size allows to extract a robust signal which may be obscured in one single
379	observational record.	405	observational record.
380	Figure 8.Seasonal averages of specific humidity at the cold point as a function of cold point temperature for SON 1991 (a) and 1992 (b).	406	Figure 8.Seasonal averages of specific humidity at the cold point as a function of cold point temperature for SON 1991 (a) and 1992 (b).
381	Values for each individual ensemble member are shown as dots for the inner tropics. An approximation (see text) for the Clausius Clapeyron	407	Values for each individual ensemble member are shown as dots for the inner tropics. An approximation (see text) for the Clausius Clapeyron
382	equation at this temperature range with an 12%increase of specific humidity perKis shown with a dashed grey line. The exact solution for	408	equation at this temperature range with an 12%increase of specific humidity perKis shown with a dashed grey line. The exact solution for
383	the Clausius Clapeyron Equation over ice by Murphy and Koop (2005) is calculated for the average ensemble cold point temperatures and	409	the Clausius Clapeyron Equation over ice by Murphy and Koop (2005) is calculated for the average ensemble cold point temperatures and
384	pressure and shown in orange.	410	pressure and shown in orange.
385	Table 1.Number of ensemble members lying within the two standard deviations of the corresponding control run values with respect to	411	Table 2.Number of ensemble members lying within the two standard deviations of the corresponding control run values with respect to
386	temperature (T) or specific humidity (Q) or their union(QUT)in 1991 and 1992.	412	temperature (T) or specific humidity (Q) or their union(QUT)in 1991 and 1992.
387	Case2.5 Tg S5 Tg S10 Tg S20 Tg S40 Tg S	413	Case2.5 Tg S5 Tg S10 Tg S20 Tg S40 Tg S
388	Q	414	Q
389	1991	415	1991
390	9376000	416	9376000
391	T	417	T
392	1991	418	1991
393	9279100	419	9279100
394	(QUT)	420	(QUT)
395	1991	421	1991
396	9382100	422	9382100
397	Q	423	Q
398	1992	424	1992
399	97895200	425	97895200
400	T	426	T
401	1992	427	1992
402	98906500	428	98906500
403	(QUT)	429	(QUT)
404	1992	430	1992
405	99906500	431	99906500
406	3.5 Comparison to the MPI-GE historical simulations (Mt. Pinatubo)280		
407	The importance of the aerosol profile shape and particularly the extinction at the cold point for the SWV entry becomes		

408 apparent when comparing the results of the EVAens members to the Mt. Pinatubo period in the MPI-GE historical simulations.

409 17

410

411 In terms of the amount of emitted sulfur the EVAens simulations for 5TgS and 10TgS can be seen as bounds of recent

412 estimates of the sulfur emission of the Mt. Pinatubo eruption of (7.5±2.5)TgS (Timmreck et al., 2018). As the PADS data

413 set describing the Mt. Pinatubo eruption is based on observational evidence rather than simulated as is the case for the EVA285

414 data sets only a range and not a set amount of emitted sulfur is given.

415 In Fig. 2, which shows the global mean TOA radiative imbalance for the MPI-GE historical simulations along with the EVAens

416 simulations, the values for the Mt. Pinatubo eruption in the MPI-GE historical simulations lie between the 5TgS and the 10

417 TgS imbalances in 1991. In 1992 the mean MPI-GE TOA imbalance is slightly more negative than the ensemble mean of even

418 the 10TgS run, but overall the deviations between the 10TgS run and Mt. Pinatubo run in the MPI-GE historical simulations290

419 are small compared to the standard error of the ensemble means.

420 However the tropical 1992 SWV anomalies in the historical simulations show a stronger increase than in the10Tg S EVAens

421 simulations (absolute values are shown in Fig. 7, for percental changes see Fig. B1). In particular the seasonal cycle of the tape

422 recorder is more strongly amplified in the Mt. Pinatubo run in 1992, where the SWV increase of 0.7ppmmexceeds the

423 0.5ppmmvalues of 1991. The heights the corresponding SWV increases reach also differ: the 0.5ppmmsignal propagates295

424 upwards to 20hPain the historical run, whereas in the 10TgS run the 0.5ppmmsignal only reaches pressure levels of up to

425 60hPa.

426 The higher SWV values are caused by a stronger heating of the atmospheric region around the cold point controlling the

427 indirect SWV entry. Fig. 9 shows cold point temperatures for the example of SON 1992. Mt. Pinatubo in the MPI-GE historical

428 simulations reaches values lying between those of the 10TgS and 20TgS run.300

432 17

433

434 3.5 Comparison to the MPI-GE historical simulations (Mt. Pinatubo)

435 The importance of the aerosol profile shape and particularly the extinction at the cold point for the SWV entry becomes285

436 apparent when comparing the results of the EVAens members to the Mt. Pinatubo period in the MPI-GE historical simulations.

437 In terms of the amount of emitted sulfur the EVAens simulations for 5TgS and 10TgS can be seen as bounds of recent

438 estimates of the sulfur emission of the Mt. Pinatubo eruption of (7.5±2.5)TgS (Timmreck et al., 2018). As the PADS data

439 set describing the Mt. Pinatubo eruption is based on observational evidence rather than simulation results, as is the case for the

440 EVA data sets, only a range and not a set amount of emitted sulfur is given.290

441 In Fig. 2, which shows the global mean TOA radiative imbalance for the MPI-GE historical simulations along with the EVAens

442 simulations, the values for the Mt. Pinatubo eruption in the MPI-GE historical simulations lie between the 5TgS and the 10

443 TgS imbalances in 1991. In 1992 the mean MPI-GE TOA imbalance is slightly more negative than the ensemble mean of even

444 the 10TgS run, but overall the deviations between the 10TgS run and Mt. Pinatubo run in the MPI-GE historical simulations

445 are small compared to the standard error of the ensemble means.295

446 However, the tropical 1992 SWV anomalies in the historical simulations show a stronger increase than in the10Tg S

447 EVAens simulations (absolute values are shown in Fig. 7, for percental changes see Fig. B1). In particular, the SWV increases

448 show a seasonality enhancing the tape recorder amplitude in the historical simulations but not to that extent in the EVAens

449 simulations. Notable is also a stronger SWV increase of 1.2ppmvin the Mt. Pinatubo simulations of 1992 exceeding the 1.0

450 ppmvof 1991. The heights the corresponding SWV increases reach also differ: the 0.7ppmvsignal propagates upwards to 20300

451 hPain the historical run, whereas in the 10TgS run the 0.7ppmvsignal only reaches pressure levels of up to 50hPa.

452 The higher SWV values are caused by a stronger heating of the atmospheric region around the cold point, controlling the

453 indirect volcanic SWV entry. Fig. 9 shows cold point temperatures for the example of SON 1992. Mt. Pinatubo in the MPI-GE historical

454 simulations reaches values lying between those of the 10TgS and 20TgS run. T

429 18

430

431 Figure 9.Average cold point temperature in the inner tropical region [-5,5]° latitude as a function of emitted sulfur for all EVAens members in

432 SON 1992. Each point symbolizes one ensemble member, the horizontal lines denote the ensemble mean. The average cold point temperature

433 range for the historical Mt Pinatubo eruptions (PADS) is shown as a grey line, the shaded region shows the extent between corresponding

434 maximum and minimum cold point temperature.

435 This can be understood when comparing the aerosol extinction profiles of EVA and PADS as proxy for the heating generated

436 by the aerosol layer (Fig. 10).

437 Although the PADS data set Mt. Pinatubo forcing has a peak value lying between the 5TgS and 10TgS EVA extinctions,

438 the extinction at the tropopause and cold point pressure is slightly higher than the 10TgS extinction in September 1991 (Fig.

439 10a and 10c). In September 1992 the PADS data set reaches values of the 20TgS EVA in the cold point region and a peak305

440 extinction comparable to the 10TgS profile (Fig. 10b and 10d)

441 6

442 . As the warming of the cold point determines the SWV entry

443 anomaly, the extinction values at this point will have a significant impact.

444 In 1991 the 10TgS EVA extinction values at the cold point are only slightly weaker than the extinction values of the PADS

445 forcing set. The SWV values in the MPI-GE historical simulations are nearer to but not equal to those in the 10TgS as the

446 peak extinction values in the 10TgS are more than two times larger than the PADS extinction values and partially compensate310

447 the lower values at the cold point. In 1992 the situation changes and the MPI-GE shows higher SWV values than the 10TgS

448 ensemble. The amplification of the SWV entry in the second SON after the eruption can be attributed to three phenomena. First,

449 the LW-heating was only building up in the northern hemisphere summer of 1991 and had not yet reached its maximum value

450 as can be seen in the plot for the cold point temperatures (Figure 4). Second, the peak extinction values of the PADS data set

451 start to exceed the 10TgS values in 1992, which may lead to higher SWV-entry values (Fig

his can be understood when

455 comparing the aerosol extinction profiles of EVA and PADS as proxy for the heating generated by the aerosol layer (Fig. 10).305

456 18

457

458 Figure 9.Average cold point temperature in the inner tropical region [-5,5]° latitude as a function of emitted sulfur for all EVAens members in

459 SON 1992. Each point symbolizes one ensemble member, the horizontal lines denote the ensemble mean. The average cold point temperature

460 range for the historical Mt Pinatubo eruptions (PADS) is shown as a grey line, the shaded region shows the extent between corresponding

461 maximum and minimum cold point temperature.

462 Although the PADS data set Mt. Pinatubo forcing has a peak value lying between the 5TgS and 10TgS EVA extinctions,

463 the extinction at the tropopause and cold point pressure is slightly higher than the 10TgS extinction in September 1991 (Fig.

464 10a and 10c). In September 1992 the PADS data set reaches values of the 20TgS EVA in the cold point region and a peak

465 extinction comparable to the 10TgS profile (Fig. 10b and 10d). A similar behaviour is apparent in the solar extinction bands

466 (s. Appendix E1). As the warming of the cold point determines the SWV entry anomaly, the extinction values at this point will310

467 have a significant impact.

468 In 1991 the 10TgS EVA extinction values at the cold point are only slightly weaker than the extinction values of the PADS

469 forcing set. The SWV values in the MPI-GE historical simulations are nearer to, but not equal, to those in the 10TgS as the

470 peak extinction values in the 10TgS are more than two times larger than the PADS extinction values and partially compensate

471 the lower values at the cold point. In 1992 the situation changes and the MPI-GE shows higher SWV values than the 10Tg315

472 S ensemble. The amplification of the SWV entry in the second SON after the eruption can be attributed to three phenomena.

473 First, the heating was only building up in the northern hemisphere summer of 1991 and had not yet reached its maximum value,

474 as can be seen in the plot for the cold point temperatures (Figure 4). Second, the peak extinction values of the PADS data set

475 start to exceed the 10TgS values in 1992, which may lead to higher SWV-entry values (Figur

ure 11a). Third, the LW-extinction315
452 at the cold point reaches a second, larger maximum around the northern summer of 1992 with values comparable to the 20
453 6
454 A similar behaviour is apparent in the solar extinction bands (s. Appendix E1).
455 19
456
457 TgS. This second maximum is not represented in the EVA-forcing (Figure 11b) and goes along with a decrease in the peak
458 forcing difference in the EVAens which lead to higher SWV entry values in the EVAens in 1991
459 7
460 .
461 Figure 10.Average tropical aerosol extinction in the 8475-9259nminfrared waveband from EVA for 5TgS, 10TgS, and 20TgS
462 emissions as well as from PADS in September 1991 and 1992. The horizontal lines indicate the pressure levels of the cold point in the region
463 between [-5,5]
464 ◦
465 latitude for each eruption strength. (a) and (b) show the entire profile. (c) and (d) are zooms on the cold point region.
466 7
467 The larger difference between peak extinction and extinction at the cold point as well as the faster decline of extinction values compensated for the very
468 similar extinction values at the cold point in 1991.
469 20
470
471 Figure 11.Temporal evolution of the aerosol extinction at the profile peak (a) and the cold point region (b) for the 8475-9259 nm IR-
472 waveband.
473 3.6 Adjusted forcing caused by the SWV increases
474 In the following the adjusted SWV forcing is investigated using the one-dimensional model konrad, as described in Sect. 2.2.320
475 Fig. 12a shows flux anomalies between an unperturbed run and a run perturbed with the increased SWV levels of the 40Tg
476 S run but without the volcanic aerosols. Although SWV levels are increased within the complete stratospheric column, the
477 main flux changes occur in the tropopause region where the enhancement is largest, the atmosphere denser and the effect of
478 stratospheric water vapour on the radiative forcing strongest (Solomon et al., 2010). In this region, the incoming solar radia-
479 tion (SWd) is reduced by absorption, while at lower altitudes, there is no difference in the shortwave flux. This is caused by325
480

e 11a). Third, the LW-extinction
476 at the cold point reaches a second, larger maximum around the northern summer of 1992 with values comparable to the 20Tg320
477 S. This second maximum is not represented in the EVA-forcing (Figure 11b) and goes along with a decrease in the peak forcing
478 19
479
480 difference in the EVAens, which lead to higher SWV entry values in the EVAens in 1991. The larger difference between peak
481 extinction and extinction at the cold point compensated for the very similar extinction values at the cold point in 1991.
482 Figure 10.Average tropical aerosol extinction in the 8475-9259nminfrared waveband from EVA for 5TgS, 10TgS, and 20TgS
483 emissions as well as from PADS in September 1991 and 1992. The horizontal lines indicate the pressure levels of the cold point in the region
484 between [-5,5]
485 ◦
486 latitude for each eruption strength. (a) and (b) show the entire profile. (c) and (d) are zooms on the cold point region.
487 20
488
489 Figure 11.Temporal evolution of the aerosol extinction at the profile peak (a) and the cold point region (b) for the 8475-9259 nm IR-
490 waveband.
491 3.6 Adjusted forcing caused by the SWV increases
492 In the following the adjusted SWV forcing is investigated using the one-dimensional model konrad, as described in Sect. 2.2.325
493 Fig. 12a shows flux anomalies between an unperturbed run and a run perturbed with the increased SWV levels of the 40Tg
494 S run but without the volcanic aerosols. Although SWV levels are increased within the complete stratospheric column, the
495 main flux changes occur in the tropopause region where the enhancement is largest, the atmosphere denser, and the effect of
496 stratospheric water vapour on the radiative forcing strongest (Solomon et al., 2010). In this region, the incoming solar radiation
497 (SWd) is reduced by absorption, while at lower altitudes there is no difference in the shortwave flux. This is caused by the SWV330
498

480 the SWV absorbing part of the solar spectrum which amongst others tropospheric water vapour completely absorbs at lower

481 altitudes in the unperturbed run. The reflected solar radiation (SWu) at the surface is only changed negligibly sinceΔSWd at

482 the surface is also negligible. Consequently the SW contribution (SWd-SWu) to the adjusted forcing at the TOA is very small

483 and negative - accounting for a slightly increased scattering due to the SWV.

484 The emitted long wave radiation (LWu) near the surface is unchanged. This is due to our setup: As the surface temperature330

485 is fixed in the perturbed run, the surface LWu is fixed as well. The SWV leads to an increase in the tropopause temperatures,

486 following this increase the emitted long wave radiation at the main SWV levels is also increased as can be seen in the increased

487 downward long wave radiation (LWd). The outgoing long wave radiation (LWu) above the tropopause region is substantially

488 reduced as the SWV acts as a greenhouse gas and traps part of the outgoing radiation. This leads to the characteristic net

489 positive forcing of the greenhouse gas at the TOA.335

490 A comparison to the instantaneous forcing (Fig. 12b) shows that the temperature adaptations in the stratosphere significantly

491 amplify the SWV forcing. Whereas the atmospheric levels with increased SWV are warmed, stratospheric cooling is found in

492 higher atmospheric levels, reducing the LWu. This effect has to build up though and consequently is more pronounced in the

493 21

494

495 adjusted SWV forcing.

496 340

497 Figure 12.SW and LW contributions to the total adjusted SWV radiative forcing in the tropical stratosphere [-5,5]

498 ◦

499 latitude for November

500 1992 of the 40TgS eruption. The upward fluxes are in solid lines and positive upward, the downward fluxes are in dashed lines and positive

501 downward. The difference between the perturbed and unperturbed equilibrium state are shown. (a) adjusted forcing (b) instantaneous forcing.

502 The temporal evolution of the adjusted inner tropical SWV forcing is shown for all five eruptions strengths in Fig. 13. The

503 adjusted forcing caused by the SWV, corresponding to the 2.5TgS to 10TgS run, is below t

498 absorbing part of the solar spectrum which, amongst others, tropospheric water vapour completely absorbs at lower altitudes in

499 the unperturbed run. The reflected solar radiation (SWu) at the surface is only changed negligibly sinceΔSWd at the surface

500 is also negligible.

501 Consequently, the SW contribution (SWd-SWu) to the adjusted forcing at the TOA is very small and negative - accounting

502 for a slightly increased scattering due to the SWV.335

503 The emitted long wave radiation (LWu) near the surface is unchanged. This is due to our setup: As the surface temperature

504 is fixed in the perturbed run, the surface LWu is fixed as well. The SWV leads to an increase in the tropopause temperatures,

505 following this increase the emitted long wave radiation at the main SWV levels is also increased, as can be seen in the increased

506 downward long wave radiation (LWd). The outgoing long wave radiation (LWu) above the tropopause region is substantially

507 reduced as the SWV acts as a greenhouse gas and traps part of the outgoing radiation. This leads to the characteristic net340

508 positive forcing of the greenhouse gas at the TOA.

509 A comparison to the instantaneous forcing (Fig. 12b) shows that the temperature adaptations in the stratosphere significantly

510 amplify the SWV forcing. Whereas the atmospheric levels with increased SWV are warmed, stratospheric cooling is found in

511 higher atmospheric levels, reducing the LWu. This effect has to build up though and consequently is more pronounced in the

512 21

513

514 adjusted SWV forcing. If addition to the strastosphere also the troposphere were allowed to adjust, part of this effect may be345

515 counterbalanced (Huang et al., 2020).

516 Figure 12.SW and LW contributions to the total adjusted SWV radiative forcing in the tropical stratosphere [-5,5]

517 ◦

518 latitude for November

519 1992 of the 40TgS eruption. The upward fluxes are in solid lines and positive upward, the downward fluxes are in dashed lines and positive

520 downward. The difference between the perturbed and unperturbed equilibrium state are shown. (a) adjusted forcing (b) instantaneous forcing.

521 The temporal evolution of the adjusted inner tropical SWV forcing is shown for all five eruptions strengths in Fig. 13. The

522 adjusted forcing caused by the SWV, corresponding to the 2.5TgS to 10TgS run, is below the

he maximum fluctuations caused
504 by internal variability denoted by the grey
line in Fig. 13. The forcing found for thes
e runs could be found for one ensemble
505 member in the 0Tgcase as well, although the
time span in which it would occur would mos
t probably be shorter. The adjusted
506 SWV forcing for the 40TgS run reaches values
of up to 0.65Wm
507 −2
508 .345
509 The signal evolution, especially in the 20Tg
S and 40TgS cases, matches the evolution alr
eady observed for the SWV in-
510 creases in the tropical region, with two sea
sonal peaks. However, since also SWV at pres
sures lower than 100hPacontribute
511 to the total forcing - although to a smaller
extent - and the transport out of the tropic
al region with the BDC is slow the first
512 peak forcing times are longer than the peak
specific humidity values at 100hPain Fig.
4.
513 When considering clouds the stratosphericall
y adjusted forcing caused by the additional
SWV is slightly increased by0.1350
514 Wm
515 −2
516 at most (compare Appendix Fig. E2). As the c
louds reflect part of the downcoming SW radi
ation some of the SW
517 bands, which could be completely absorbed wh
ile traveling through the complete stratosph
ere and troposphere, are not ab-
518 sorbed entirely. The additional SWV in the s
tratosphere increases the absorption in thes
e bands and reduces the outgoing SW
519 radiation. This leads to an increase of the
forcing.
520 355
521 22
522
523 Figure 13.Time evolution of the TOA adjusted
clear sky forcing in the tropical region [-
5,5]
524 ◦
525 latitude for the ensemble mean SWV increases
526 caused by all eruption strengths. The blue l
ine marks the eruptions time, the dashed gre
y line shows the threshold up to which devia
tions in
527 radiative fluxes can be caused by internal v
ariability. Shaded areas indicate the flux a
nomaly range originating from the standard d
eviations
528 of the SWV profiles are plotted to visualize
the signal range.
529 The monthly SWV forcing does not exhibit a l
inear behaviour with respect to the mass of
emitted sulfur or main IR AOD
530 waveband (Fig. A5, A6) as was to be expected
since the monthly cold point temperatures al
so did not change in a linear man-
531 ner with respect to the emitted sulfur mass
and the temperature increases showed a hyst

maximum fluctuations caused
523 by internal variability denoted by the grey l
ine in Fig. 13. The forcing found for these r
uns could be found for one ensemble350
524 member in the 0Tgcase as well, although the t
ime span in which it would occur would most p
robably be shorter. The adjusted
525 SWV forcing for the 40TgS run reaches values
of up to 0.65Wm
526 −2
527 .
528 The signal evolution, especially in the 20TgS
and 40TgS cases, matches the evolution alread
y observed for the SWV in-
529 creases in the tropical region, with two seas
onal peaks. However, since also SWV at pressu
res lower than 100hPacontribute
530 to the total forcing - although to a smaller
extent - and the transport out of the tropic
al region with the BDC is slow, the first355
531 peak forcing times are longer than the peak s
pecific humidity values at 100hPain Fig. 4.
532 When considering clouds the stratospherically
adjusted forcing caused by the additional SWV
is slightly increased by0.1
533 Wm
534 −2
535 at most (compare Appendix Fig. E2). As the cl
ouds reflect part of the down coming SW radia
tion, some of the SW
536 bands, which could be completely absorbed whi
le traveling through the complete stratospher
e and troposphere, are not ab-
537 sorbed entirely. The additional SWV in the st
ratosphere increases the absorption in these
bands and reduces the outgoing SW360
538 radiation. This leads to an increase of the f
orcing.
539 22
540
541 Figure 13.Time evolution of the TOA adjusted
clear sky forcing in the tropical region [-
5,5]
542 ◦
543 latitude for the ensemble mean SWV increases
544 caused by all eruption strengths. The blue li
ne marks the eruptions time, the dashed grey
line shows the threshold up to which deviati
ons in
545 radiative fluxes can be caused by internal va
riability. Shaded areas indicate the flux ano
maly range originating from the standard devi
ations
546 of the SWV profiles are plotted to visualize
the signal range.
547 The monthly SWV forcing does not exhibit a li
near behaviour with respect to the mass of em
itted sulfur or main IR AOD
548 waveband (Fig. A5, A6) as was to be expected
since the monthly cold point temperatures al
so did not change in a linear man-
549 ner with respect to the emitted sulfur mass a
nd the temperature increases showed a hystere

eretic behaviour. However again, when

532 averaging out the seasonal dependencies and partitioning the time after the eruption in the signal build up phase (1991, after

533 the eruption), the phase of approximately constant forcing (1992), and the phase of declining signal (1993) the relationship360

534 is quasi linear

535 8

536 . Deviations from the linear trend are introduced by the higher eruption strength not reaching the maximum

537 forcing values as fast as the lower eruption strengths in 1991. Additionally, the lower eruption strengths already are relaxing

538 to the background state in 1993. Compared to the cold point - AOD relation the signal build up and relaxation back to the

539 ground state is damped as the transport of the SWV into the stratosphere and out of the tropical region takes place over longer

540 timescales than the cold point warming or cooling respectively.365

541 8

542 Although the relationship is of second order the second order term does not lead to a large contribution within the examined region.

543 23

544

545 Figure 14.Yearly averages of adjusted forcing due to the additional SWV as a function of AOD (IR, 8475-9259 nm) for the three examined

546 years (1991-1993). A second order fit for each year with corresponding equation is shown.

547 As the SWV forcing counteracts the volcanic forcing its relation to the aerosol forcing is of interest. Fig. 15a shows the

548 tropical aerosol forcing as calculated using the double radiation call in the MPI-ESM. For the evaluation of the forcing in the

549 double radiation call different conventions exists as far as the evaluation at the TOA or the tropopause are concerned (e.g.

550 compare Forster et al. (2016)). As the double radiation call is used to determine an instantaneous forcing the readout at the370

551 tropopause level would be following the standard convention as defined by Hansen et al. (2005) or in the IPCC. However we

552 present both values for the entire and inner tropics to allow for an easy comparison to other studies. The relative magnitude of

553 the SWV adjusted forcing with respect to the aerosol forcing increases approximately linearly until September 1992 and then

554 reaches a constant values of around 2.5 % for the readout at the tropopause and 4 % percent for the readout at the TOA (Fig.

555 15b).375

556 24

557

558 Figure 15.(a) Time evolution of the tropical aerosol forcing for the 10TgS run as calcula

tic behaviour. However again, when365

550 averaging out the seasonal dependencies and partitioning the time after the eruption in the signal build up phase (1991, after

551 the eruption), the phase of approximately constant forcing (1992) and the phase of declining signal (1993), the relationship

552 can be fitted to a power function of formaAOD

553 b

554 +cin the region of interest. Deviations from the trend are introduced by the

555 proportionally less warming in the cold point region in the higher eruption strengths as the main AOD increases occur in the

556 region of peak forcing. Compared to the cold point - AOD relation the signal build up and relaxation back to the ground state370

557 is damped as the transport of the SWV into the stratosphere and out of the tropical region takes place over longer timescales

558 than the cold point warming or cooling respectively.

559 23

560

561 Figure 14.Yearly averages of adjusted forcing due to the additional SWV as a function of AOD (IR, 8475-9259 nm) for the three examined

562 years (1991-1993). A power function fit for each year with corresponding equation is shown.

563 As the SWV forcing counteracts the volcanic forcing, its relation to the aerosol forcing is of interest. Fig. 15a shows the

564 tropical aerosol forcing as calculated using the double radiation call in the MPI-ESM. For the evaluation of the forcing in the375

565 double radiation call different conventions exists as far as the evaluation at the TOA or the tropopause are concerned (e.g.

566 compare Forster et al. (2016)). As the double radiation call is used to determine an instantaneous forcing, the readout at the

567 tropopause level would be following the standard convention as defined by Hansen et al. (2005) or in the IPCC. However, we

568 present both values for the entire and inner tropics to allow for an easy comparison to other studies. The relative magnitude of

569 the SWV adjusted forcing with respect to the aerosol forcing increases approximately linearly until September 1992 and then380

570 reaches a constant values of around 2.5 % for the readout at the tropopause and 4 % percent for the readout at the TOA (Fig.

571 15b).

572 24

573

574 Figure 15.(a) Time evolution of the tropical aerosol forcing for the 10TgS run as calcula

ted using the double radiation call in the MPI-
559 ESM. The forcing is evaluated for the inner and entire tropics at the tropopause and the TOA. (b) Time evolution of the percentage of aerosol
560 forcing counterbalanced by the SWV forcing in the inner and entire tropics .
561 4 Discussion
562 4.1 Magnitude of SWV increases due to indirect mechanism
563 In general, the annual cycle of the tropical tropopause temperatures account for a variation of SWV content of±1.4ppmv≈
564 0.87ppmmaround the mean background at 110hPa in SAGE II data of the early 1990s (Mote et al., 1996). In the MPI-GE the
565 variations of the SWV tape recorder signal at the same height is lower: it reaches±0.69ppmmat most. However the SAGE380
566 II data which Mote et al. (1996) used fell in the era of the Mt. Pinatubo eruption and may be biased due to an amplification of
567 the seasonal signal by the presence of the aerosol layer, an effect leading to maximum deviations of up to±1.0ppmmin the
568 10TgS scenarios of the EVAens simulations.
569 In the case of volcanic perturbations, the alterations in SWV content caused by the indirect entry mechanism via tropopause
570 warming after volcanic eruptions can surpass the SWV variations due to the annual cycle. In our simulations, deviations of385
571 more than±0.69ppmmare produced in the simulations for emissions equal to or larger than 10TgS eruption, which is an
572 upper bound of the emission estimate for the Mt. Pinatubo eruption (Timmreck et al., 2018). The SWV anomaly caused by
573 the 2.5TgS eruption is comparable to changes caused by the Quasi Biennial Oscillation (QBO) which are [0.1-0.2]ppmm
574 according to the regression analysis by Dessler et al. (2013), whereas the 10TgS has an impact stronger than the changes in
575 the Brewer Dobson Circulations (maximum 0.4ppmm).390
576 In climate change studies the stratospheric water vapour is also affected by doubledCO
577 2
578 concentrations. The SST increase
579 leads to a consequent atmospheric humidity increase amounting up to 6-10 % in the stratosphere, with values exceeding 10
580 % in the lower stratosphere (Wang et al. (2020)). These changes of SWV due to increasedC

ted using the double radiation call in the MPI-
575 ESM. The forcing is evaluated for the inner and entire tropics at the tropopause and the TOA. (b) Time evolution of the percentage of aerosol
576 forcing counterbalanced by the SWV forcing in the inner and entire tropics .
577 4 Discussion
578 4.1 Magnitude of SWV increases due to indirect volcanic mechanism
579 In general, the annual cycle of the tropical tropopause temperatures accounts for a variation of the SWV content of±1.4ppmv385
580 around the mean background at 110hPain SAGE II data of the early 1990s (Mote et al., 1996) and±1.0-1.3ppmvbased on
581 the SPARC Data Initiative multi-insturment mean (SDI MIM) at 100hPain 2005-2010 (Davis et al., 2017). In the MPI-GE
582 the variation in the SWV tape recorder signal at the same height reaches up to±1.1ppmv. This is in accordance with the
583 SDI MIM and only slightly lower than the SAGE II data. The slightly higher values reported for early 1990 fall into the era
584 of the Mt. Pinatubo eruption, however, and may be increased compared to the multiyear mean due to an amplification of the390
585 seasonal signal by the presence of the aerosol layer, an effect leading to maximum deviations of up to±1.6ppmvin the 10
586 TgS scenario of the EVAens simulations.
587 In the case of volcanic perturbations, the alterations in SWV content caused by the indirect temperature controlled entry
588 mechanism via tropopause warming after volcanic eruptions can surpass the SWV variations due to the annual cycle. In our
589 simulations deviations of more than±1.1ppmvare produced in the simulations for emissions equal to or larger than 10TgS395
590 eruption, which is an upper bound of the emission estimate for the Mt. Pinatubo eruption (Timmreck et al., 2018). The SWV
591 anomaly caused by the 2.5TgS eruption is comparable to changes caused by the Quasi Biennial Oscillation (QBO), which are
592 [0.16-0.32]ppmvaccording to the regression analysis by Dessler et al. (2013), whereas the 10TgS has an impact stronger
593 than the changes in the BDC (maximum 0.65ppmv).
594 In climate change studies, the stratospheric water vapour is also affected by doubledCO
595 2
596 concentrations. The SST increase400
597 leads to a consequent atmospheric humidity increase, amounting up to 6-10 % in the stratosphere, with values exceeding 10
598 % in the lower stratosphere (Wang et al. (2020)). These changes of SWV due to increasedCO

0

581 2

582 levels are comparable to

583 increases caused by the smaller eruptions of 2.5TgS and 5TgS. However, the SWV changes due to volcanic eruptions are

584 25

585

586 only temporary.395

587 4.2 Comparison to studies based on reanalysis data

588 While our model study has the advantage that the large ensemble size allows to determine the statistical significance of SWV

589 increases and the spread of responses to a volcanic eruption due to internal variability of the earth system, a model study

590 is always limited by the ability of the model to represent the earth system realistically. Individual models may differ in the400

591 parametrization of convection, the entry of water vapour into the stratosphere and tropopause height. The aerosol profiles used

592 in this study are artificial in case of the EVAens and will include uncertainties of the retrieval in case of the PADS data set. As

593 our results are therefore at first only representative for the model used in our study, a comparison to results based on or derived

594 from observational evidence of SWV changes after volcanic eruptions is desirable.

595 405

596 During the Mt. Pinatubo period observations of SWV are scarce, especially as solar occultation measurements done by

597 HALOE and SAGE II in this time period suffered from aerosol interference. Therefore, the data usage is discouraged (e.g.

598 Fueglistaler et al. (2013)). Although Fueglistaler et al. (2013) mention "anomalously large anommalies" of SWV values shortly

599 after the Mt. Pinatubo eruption in SAGE II data, they also warn that the measurements may be biased by aerosol artifacts since

600 the SWV signal is not present in the HALOE data. We will therefore compare our results not directly to satellite observations410

601 but to the outcome of two regression analysis studies of SWV entry fed by reanalysis input by Dessler et al. (2014) and Tao

602 et al. (2019).

599 2

600 levels are comparable to

601 25

602

603 increases caused by the smaller eruptions of 2.5TgS and 5TgS. However, the SWV changes due to volcanic eruptions are

604 only temporary.

605 405

606 4.2 Comparison to studies based on reanalysis data

607 While our model study has the advantage that the large ensemble size allows determining the statistical significance of SWV

608 increases, and the spread of responses to a volcanic eruption due to internal variability of the earth system, a model study is

609 always limited by the ability of the model to represent the earth system realistically. Individual models may differ in the pa-

610 rameterization of convection, the entry of water vapour into the stratosphere and tropopause height. The aerosol profiles used410

611 in this study are artificial in case of the EVAens and will include uncertainties of the retrieval in case of the PADS data set. As

612 our results are therefore at first only representative for the model used in our study, a comparison to results based on or derived

613 from observational evidence of SWV changes after volcanic eruptions is desirable.

614 During the Mt. Pinatubo period, observations of SWV are scarce, especially as solar occultation measurements by HALOE415

615 and SAGE II suffered from aerosol interference in this time period. Therefore, the data usage is discouraged (e.g. Fueglistaler

616 et al. (2013)). Although Fueglistaler et al. (2013) mention "anomalously large anomalies" of SWV values shortly after the Mt.

617 Pinatubo eruption in SAGE II data, they also warn that the measurements may be biased by aerosol artifacts since the SWV

618 signal is not present in the HALOE data. However, in the Boulder balloon data a 1-2ppmvincrease of SWV at 24-26kmand

619 18-20kmlevels is registered for 1992 (Oltmans et al., 2000). Most likely due to additional contribution from e.g. methane420

620 oxidation or ENSO signals, the maximum increase is slightly larger than the 1.0-1.2ppmvshown for the tropical region in

621 Figure 7 or the value of approximately 0.5ppmvat 70hPaat corresponding latitude (compare Appendix F1). Angell (1997)

622 report the stratospheric warming due the eruption of Mt. Pinatubo based on radiosonde data. The warmest seasonal anomaly

623 in the 100-50hPalayer at the equator is (2.0±0.8)K, where we find 1.8Kfor the 5TgS and 5.2Kfor the 10TgS scenario

603 Dessler et al. (2014) studied the different contributors to SWV entry at 82hPaslightly above the tropical tropopause in their

604 model by using a regression analysis on data output from a trajectory model fed by MERRA reanalysis input (Rienecker et al.,415

605 2011). They found a maximum increase of 0.34 ppmv≈0.21ppmmin the residual which partially overlapped with the AOD

606 increase caused by the Mt. Pinatubo eruption. This value is lower than our finding of up to (0.5-0.7)ppmmincrease in the first

607 and second post eruption year. Our higher value may be explained by the different quantification approaches: whereas Dessler

608 et al. (2014) indirectly quantified the SWV increase due to Mt. Pinatubo in the residuals after subtracting contributing terms

609 like the Brewer Dobson Circulation (BDC), we directly quantified the additional SWV entry by taking the difference between420

610 the MPI-GE historical ensemble with volcanic aerosol and the EVAens control run. This allowed us to avoid the subtraction

611 of SWV entry caused by the volcanic eruption but attributed to other sources. For example, in the Dessler et al. (2014) study,

612 the BDC term is described by the heating in the 82hParegion, but part of this may be caused by the presence of the volcanic

613 aerosol layer. Our method also eliminates other sources of SWV, that may have caused the SWV increase prior to the Mt.

614 Pinatubo eruption in the residual of Dessler et al. (2014). Additionally the tropopause in our historical simulation is located at425

615 pressures larger than 100hPa. Dessler et al. (2014) report that 82hPais slightly above the tropopause. Their possibly higher

616 lying tropopause may also explain part of the difference.

617 26

618

619 Tao et al. (2019) use MERRA-2 (Gelaro et al., 2017), JRA-55 (Kobayashi et al., 2015) and ERAinterim (Dee et al., 2011)

620 reanalysis data for the Mt. Pinatubo eruption in their trajectory model. The SWV increase attributed to the volcanic eruption430

621 ranges between 0.4ppmv≈0.2ppmm(ERAinterim) and 0.8ppmv≈0.5ppmm(MERRA-2 and JRA-55). Only MERRA-2

622 explicitly accounts for the volcanic aerosol

624 in OND 1991. An estimate of the corresponding SWV increase at background levels of 4.5-5ppmvis 1.13-1.3ppmv, in425

625 accordance with our findings (Figure 7).

626 As more regression analyses exist based on reanalysis data, we now compare our results not directly to observations but to the

627 outcome of two regression analysis studies of SWV entry fed by the reanalysis input by Dessler et al. (2014) and Tao et al.

628 (2019).

629 430

630 Dessler et al. (2014) studied the different contributors to SWV entry at 82hPaslightly above the tropical tropopause in their

631 model by using a regression analysis on data output from a trajectory model fed by MERRA reanalysis input (Rienecker et al.,

632 2011). They found a maximum increase of 0.34ppmvin the residual, which partially overlapped the AOD increase caused by

633 the Mt. Pinatubo eruption. This value is lower than our finding of up to (0.8-1.1)ppmvincrease in the first and second post

634 eruption year. Our higher value may be explained by the different quantification approaches: whereas Dessler et al. (2014) in-435

635 directly quantified the SWV increase due to Mt. Pinatubo in the residuals after subtracting contributing terms like the BDC, we

636 26

637

638 directly quantified the additional SWV entry by taking the difference between the MPI-GE historical ensemble with volcanic

639 aerosol and the EVAens control run. This allowed us to avoid the subtraction of SWV entry caused by the volcanic eruption

640 but attributed to other sources. For example, in the Dessler et al. (2014) study, the BDC term is described by the heating in

641 the 82hParegion, but part of this may be caus

(Fujiwara et al. (2017)); the corresponding 0.5ppmmSWV increase is in good
623 agreement with the SWV increases found in the first post eruption year in our historical simulations, although our SWV increase
624 in the second post eruption year of 0.7ppmmis larger. As analyzed in Sect. 3.5 this higher increase is mainly caused by the
625 height of the aerosol layer with respect to the cold point. Since the aerosol profiles are imported on fixed pressure levels, with-435
626 out prescribing the distance between the tropopause and the aerosol layer, and the tropopause heights differ amongst various
627 models, the comparison between different studies and models would be facilitated if not only the total volcanic forcing but also
628 the heating in the region were reported when quantifying and comparing volcanically induced SWV increases. A tropopause
629 located at larger pressure in the second post eruption summer in the reanalysis data could explain the differences between the
630 maximum values of 0.5ppmmby Tao et al. (2019) using the MERRA-2 reanalysis and our results amounting up to 0.7ppmm440
631 in the second post eruption summer as it would lead to a lower heating rate in the cold point region and consequently a reduced
632 water vapour entry compared to the control years. Additionally our model does not include interactive chemistry -H
633 2
634 Osinks
635 could reduce the amount of water vapour and especially the build up in the second poster eruption summer. However Löffler
636 et al. (2016) also found a stronger SWV increase in the second post eruption summer when investigating the perturbations of

ed by the presence of the volcanic aerosol layer. Our method also eliminates440
642 other sources of SWV that may have caused the SWV increase prior to the Mt. Pinatubo eruption in the residual of Dessler
643 et al. (2014). Additionally, the tropopause in our historical simulation is located at pressures larger than 100hPa. Dessler et al.
644 (2014) report that 82hPais slightly above the tropopause. Their possibly higher lying tropopause may also explain part of the
645 difference.
646 Tao et al. (2019) use MERRA-2 (Gelaro et al., 2017), JRA-55 (Kobayashi et al., 2015) and ERAinterim (Dee et al., 2011)445
647 reanalysis data for the Mt. Pinatubo eruption in their trajectory model. The SWV increase attributed to the volcanic eruption
648 ranges between 0.4ppmv(ERAinterim) and 0.8ppmv(MERRA-2 and JRA-55). Only MERRA-2 explicitly accounts for the
649 volcanic aerosol (Fujiwara et al. (2017)): their corresponding 0.8ppmvSWV increase is in good agreement with the SWV
650 increases found in the first post eruption year in our historical simulations, although our SWV increase in the second post
651 eruption year of 1.1ppmvis larger. As analyzed in Sect. 3.5, this higher increase is mainly caused by the height of the aerosol450
652 layer with respect to the cold point. Since the aerosol profiles are imported on fixed pressure levels, without prescribing the
653 distance between the tropopause and the aerosol layer, and the tropopause heights differ amongst various models, the compar-
654 ison between different studies and models would be facilitated if not only the total volcanic forcing but also the heating in the
655 region were reported when quantifying and comparing volcanically induced SWV increases. A tropopause located at larger
656 pressure in the second post eruption summer in the reanalysis data could explain the differences between the maximum values455
657 of 0.8ppmvby Tao et al. (2019) using the MERRA-2 reanalysis and our results amounting up to 1.1ppmvin the second post
658 eruption summer, as it would lead to a lower heating rate in the cold point region and consequently a reduced water vapour
659 entry compared to the control years. Additionally, our model does not include interactive chemistry -H
660 2
661 Osinks could reduce
662 the amount of water vapour and especially the build up in the second post eruption summer. However, Löffler et al. (2016)
663 also found a stronger SWV increase in the second post eruption summer when investigating the perturbations of stratospheric460

637 stratospheric water vapour using nudged chemistry-climate model simulations with prescribed aerosol for the Mt. Pinatubo445

638 eruption. The SWV increases for the Mt. Pinatubo eruption reach values of up to 35 % compared to the unperturbed run in

639 the inner tropical average. Thus our finding of an increase of 25 % above the unperturbed levels for the historical simulations

640 (compare Fig. B1) lies between the estimates from reanalysis data by Tao et al. (2019) and the chemistry climate studies by

641 Löffler et al. (2016)

642 In the EVAens, where the temporal evolution of the extinction profile does not lead to as drastical changes in the vertical shape450

643 as in the PADS forcing data set, the temporal evolution of the increase in SWV is similar to the evolution found by Tao et al.

644 (2019), with only one prominent maximum. Maximum values of SWV increase are in the range of 0.2ppmmand 0.7ppmm

645 for the 5TgS and 10TgS EVAens runs - covering the estimated sulfur emission range of Mt. Pinatubo and in agreement with

646 Tao et al. (2019).

647 455

648 4.3 SWV contributions due to the indirect and direct injection

649 The main difference between the direct and indirect mechanism is that the indirect pathway is active for the entire life time

650 of the volcanic aerosol in the lower stratosphere, whereas the direct injection is a singular event. There is no study known to

651 us comparing the SWV entry due to both events within one framework. Although up to 80 % of the eruption volume can be

652 water vapour (Coffey (1996)), rapid condensation can remove 80-90 % of this humidity on the way to the stratosphere (Glaze460

653 et al. (1997)). There are only a few cases of direct injections reported, which cover relatively small eruption events. The water

654 vapour within the eruption column which reached the stratosphere did not lead to elevated SWV above background for more

655 27

656

657 than a week locally in the few cases reported, i.e the eruption of Kasatochi (2008) with maximum SWV content of 9 ppmv≈

658 5.6 ppmm (Schwartz et al., 2013) lasting for around 1 day, the eruption of Calbuco (2015) with 10 ppmv≈6.2ppmmvalues

659 for approximately a week (Sioris et al. (2016a)) and the eruption of Mt. Saint Helens with (64±4)ppmv≈(40±2)ppmm465

660 (Murcray et al. (1981)) detectable above background for around a week, but only on local and not global scale. In the case of

661 the Mt. Pinatubo eruption the model estimate

664 water vapour using nudged chemistry-climate model simulations with prescribed aerosol for the Mt. Pinatubo eruption. The

665 SWV increases for the Mt. Pinatubo eruption reach values of up to 35 % compared to the unperturbed run in the inner tropical

666 average. Thus, our finding of an increase of 25 % above the unperturbed levels for the historical simulations (compare Fig. B1)

667 lies between the estimates from reanalysis data by Tao et al. (2019) and the chemistry climate studies by Löffler et al. (2016).

668 In the EVAens, where the temporal evolution of the extinction profile does not lead to as drastic changes in the vertical shape465

669 as in the PADS forcing data set, the temporal evolution of the increase in SWV is similar to the evolution found by Tao et al.

670 (2019), with only one prominent maximum. Maximum values of SWV increase are in the range of 0.3ppmvand 1.1ppmv

671 for the 5TgS and 10TgS EVAens runs - covering the estimated sulfur emission range of Mt. Pinatubo and in agreement with

672 Tao et al. (2019).

673 470

674 27

675

676 4.3 SWV contributions due to the indirect volcanic and direct volcanic injection

677 Unlike the direct volcanic mechanism the indirect volcanic pathway is active for the entire life time of the volcanic aerosol in

678 the lower stratosphere, whereas the direct volcanic injection is a singular event. There is no study known to us comparing the

679 SWV entry due to both events within one framework. Although up to 80 % of the eruption volume can be water vapour (Coffey

680 (1996)), rapid condensation can remove 80-90

s for the direct injection tended to have a larger range and maximum values than
662 the estimates based on observational data (Joshi and Jones, 2009). In contrast the indirect pathway allows for slower, but more
663 continuous, raised stratospheric water vapour levels ultimately spread throughout the globe but with lower peak values. These
664 enhanced SWV levels are detectable in our ensemble mean even years after the actual volcanic eruption if the emitted sulfur470
665 amount is large enough (larger than 10TgS). Depending on the explosivity of the volcano the relative contributions of the
666 direct and indirect injection mechanism to SWV increases should change: for small eruptions the direct injection will lead to
667 a relatively high increase in SWV, which is short lived and spatially confined. For the larger eruptions this short lived SWV
668 enhancement is followed by a relatively stable increase of SWV which spreads throughout the entire globe, dominating the
669 SWV increase due to the volcanic eruption.475

670 4.4 SWV Forcing

671 In our simulations the adjusted forcing caused by the increase of SWV in the tropical region amounts to maximally 2.5 to
672 4 percent of the tropical aerosol forcing in the same time frame for the 10TgS eruption. However, although a decline of
673 the tropical forcing within the three year time-frame is found, the forcing around the complete globe will last longer than the
674 impact of the volcanic aerosols. The decline in tropical stratospheric water vapour is caused by its transport to the poles by480
675 the Brewer Dobson Circulation, leading to a regional shift of the location of the SWV forcing (see Fig. F1). Forster and Shine
676 (2002) found the polar SWV forcing to be 2.5 times stronger than the tropical forcing: Since in the polar region the tropopause
677 height is lower and the water vapour content is generally lower, SWV increases of the same magnitude have a much larger
678 impact there than in the tropical region. Based on the factor of 2.5, the 40TgS run 1993 values would cause an adjusted
679 forcing of up to 0.5Wm

% of this humidity on the way to the stratosphere (Glaze et al. (1997)). There are475
681 only a few cases of direct volcanic injections reported, which cover relatively small eruption events. The water vapour within
682 the eruption column reaching the stratosphere did not lead to elevated SWV above background for more than a week locally
683 in the few cases reported, i.e the eruption of Kasatochi (2008) with maximum SWV content of 9ppmv(Schwartz et al., 2013)
684 lasting for around 1 day, the eruption of Calbuco (2015) with 10 ppmv values for approximately a week (Sioris et al. (2016a)),
685 and the eruption of Mt. Saint Helens with (64 ±4)ppmv(Murcray et al. (1981)) detectable above background for around a480
686 week, but only on local and not global scale. In the case of the Mt. Pinatubo eruption the model estimates for the direct volcanic
687 injection tended to have a larger range and maximum values than the estimates based on observational data (Joshi and Jones,
688 2009). In contrast, the indirect volcanic pathway allows for slower, but more continuous, raised stratospheric water vapour
689 levels ultimately spreading throughout the globe but with lower peak values. These enhanced SWV levels are detectable in our
690 ensemble mean even years after the actual volcanic eruption if the emitted sulfur amount is larger than 10TgS. Depending485
691 on the explosivity of the volcano, the relative contributions of the direct and indirect volcanic injection mechanisms to SWV
692 increases should change: for small eruptions the direct volcanic injection will lead to a relatively high increase in SWV, which
693 is short lived and spatially confined. For the larger eruptions this short lived SWV enhancement is followed by a relatively
694 stable increase of SWV which spreads throughout the entire globe, dominating the SWV increase due to the volcanic eruption.
695 4.4 SWV Forcing490

696 In our simulations the adjusted forcing caused by the increase of SWV in the tropical region amounts to maximally 2.5 to 4
697 percent of the tropical aerosol forcing over the time frame for the 10TgS eruption. However, although a decline of the tropical
698 forcing within the three year time-frame is found, the forcing around the complete globe will last longer than the impact of
699 the volcanic aerosols. The decline in tropical stratospheric water vapour is caused by its transport to the poles by the BDC,
700 leading to a regional shift of the location of the SWV forcing (see Fig. F1). Forster and Shine (2002) found the polar SWV495
701 forcing to be 2.5 times stronger than the tropical forcing: Since in the polar region the tropopause height is lower and the water

680 −2
681 in the polar region, whereas the aerosol forcing is back to the background levels of the 2.5TgS485
682 run by 1993. This shift in relative magnitude is one of the factors contributing to the positive TOA imbalance at the end of
683 1993 (for 20 Tg S and 40 Tg S in Fig. 2).
684 For the eruption of Mt. Pinatubo with (7.5 ±2.5)TgS emitted, earlier estimates of SWV forcing exist. Joshi and Shine
685 (2003) calculated the global SWV forcing to be 0.1Wm
686 −2
687 . Our 10TgS adjusted forcing results of up to 0.11Wm
688 −2
689 are490
690 nearer to the estimate by Joshi and Shine (2003) than our 5TgS results of up to 0.03Wm
691 −2
692 . In 1992, the adjusted forcing
693 in the inner tropics lies between [0.02 - 0.03]Wm
694 −2
695 for the 5TgS and [0.06 - 0.11]Wm
696 −2
697 for the 10TgS scenarios. The
698 better agreement of our 10TgS adjusted SWV forcing with Joshi and Shine (2003) value for the Mt. Pinatubo SWV forcing
699 can be attributed to two points: First, the form and location of the forcing profile with respect to the cold point is crucial when
700 comparing model results of increased SWV levels and the consequent changes in SWV forcing (s. Sect. 3.5). This may con-495
701 tribute to differences between values from Joshi and Shine (2003) and our analysis. Second, the polar forcing will be stronger
702 28
703
704 than the tropical forcing which was calculated with konrad. Using the ratio of polar forcing to tropical forcing by Forster and
705 Shine (2002) the polar estimate would be [0.05 - 0.08]Wm
706 −2
707 for 5TgS and [0.15 - 0.28]Wm
708 −2
709 for 10TgS.
710 In their study on direct SWV entry for the K

702 vapour content is generally lower, SWV increases of the same magnitude have a much larger impact there than in the tropical
703 region. Based on the factor of 2.5, the 40TgS run 1993 values would cause an adjusted forcing of up to 0.5Wm
704 −2
705 in the
706 polar region, whereas the aerosol forcing is back to the levels of the 2.5TgS run by 1993. This shift in relative magnitude is
707 one of the factors contributing to the positive TOA imbalance at the end of 1993 (for 20 Tg S and 40 Tg S in Fig. 2).500
708 For the eruption of Mt. Pinatubo, with (7.5 ±2.5)TgS emitted, earlier estimates of SWV forcing exist. Joshi and Shine
709 (2003) calculated the global SWV forcing to be 0.1Wm
710 −2
711 . Our 10TgS adjusted forcing results of up to 0.11Wm
712 −2
713 are
714 28
715
716 nearer to the estimate by Joshi and Shine (2003) than our 5TgS results of up to 0.03Wm
717 −2
718 . In 1992, the adjusted forcing
719 in the inner tropics lies between [0.02 - 0.03]Wm
720 −2
721 for the 5TgS and [0.06 - 0.11]Wm
722 −2
723 for the 10TgS scenarios.505
724 The better agreement of our 10TgS adjusted SWV forcing with Joshi and Shine (2003) value for the Mt. Pinatubo SWV
725 forcing can be attributed to two points: First, the form and location of the forcing profile with respect to the cold point is crucial
726 when comparing model results of increased SWV levels and the consequent changes in SWV forcing (s. Sect. 3.5). This may
727 contribute to differences between our values and the ones in the Joshi and Shine (2003) analysis. Second, the polar forcing will
728 be stronger than the tropical forcing which was calculated with konrad. Using the ratio of polar forcing to tropical forcing by510
729 Forster and Shine (2002), the polar estimate would be [0.05 - 0.08]Wm
730 −2
731 for 5TgS and [0.15 - 0.28]Wm
732 −2
733 for 10TgS.
734 In their study on direct volcanic SWV entry f

rakatau eruption Joshi and Jones (2009) found a LW-forcing of +(0.33±0.09)500
Wm
−2
for a direct entry above 100hPaof 1.5ppmv≈0.933ppmmusing the downward TOA heat flux, climate feedback
parameter and near surface temperature changes. They chose a relatively high estimate of SWV increase, which would approx-
imately correspond to the SWV increases in our 20TgS eruption run. As the TOA-SW contribution to our forcing is negligible
(s. Fig. 12) a comparison to our total forcing is possible. The corresponding value of [0.21 - 0.33]Wm
−2
is in the lower part
of the range found by Joshi and Jones (2009), which is likely caused mainly by the restriction of our study to the tropical region.505
Krishnamohan et al. (2019) also mentioned the contribution of aerosol induced SWV changes to the flux changes at the
TOA in a geoengineering study, however they did not explicitly calculate it. The amount of short wave forcing they attribute
to additional SWV is much higher than our value and also positive. Presumably these differences are caused by the forcing
including tropospheric adjustments and using a 2 xCO
2
reference frame and thus can not be compared to our study.510
In order to put the adjusted radiative SWV forcing due to the indirect pathway into a broader context we compare it with
SWV forcing due to anthropogenicCO
2
and methane releases: the rate of radiative forcing increase due toCO
2
in the 2000s
9
reached values of almost 0.03Wm
−2
year
−1
and a total of (1.82±0.19)Wm
−2
for the time frame from 1750 to 2011, whereas
the additional radiative forcing due to methane in the same time frame is (0.48±0.05)Wm
−2
(Myhre et al., 2013)
10
. The515
forcing caused by the SWV entering the stratosphere via the indirect entry mechanism exceeds the yearly increase of forcing

or the Krakatau eruption Joshi and Jones (2009) found a LW-forcing of +(0.33±0.09)Wm
−2
for a direct volcanic entry above 100hPaof 1.5ppmvusing the downward TOA heat flux, climate feedback
parameter, and near surface temperature changes. They chose a relatively high estimate of SWV increase, which would approx-515
imately correspond to the SWV increases in our 20TgS eruption run. As the TOA-SW contribution to our forcing is negligible
(s. Fig. 12) a comparison to our total forcing is possible. The corresponding value of [0.21 - 0.33]Wm
−2
is in the lower part
of the range found by Joshi and Jones (2009), which is likely caused mainly by the restriction of our study to the tropical region.

Krishnamohan et al. (2019) also mentioned the contribution of aerosol induced SWV changes to the flux changes at the520
TOA in a geoengineering study, however they did not explicitly calculate it. The amount of short wave forcing they attribute
to additional SWV is much higher than our value and also positive. Presumably these differences are caused by the forcing
including tropospheric adjustments and using a 2 xCO
2
reference frame and thus can not be compared to our study.
In order to put the adjusted radiative SWV forcing due to the indirect volcanic pathway into a broader context we compare525
it with SWV forcing due to anthropogenicCO
2
and methane releases: the rate of radiative forcing increase due toCO
2
in
the 2000s
2
reached values of almost 0.03Wm
−2
year
−1
and a total of (1.82±0.19)Wm
−2
for the 1750 to 2011 time frame,

whereas the additional radiative forcing due to methane in the same time frame is (0.48±0.05)Wm
−2
(Myhre et al., 2013)
3
.
The forcing caused by the SWV entering the stratosphere via the indirect volcanic entry mechanism exceeds the yearly increase

746 due toCO
747 2
748 in the 2000s starting with the 5TgS run. The peak adjusted tropical SWV forcing for the 40TgS scenario
749 amounts to one third of the totalCO
750 2
751 forcing (due to the accumulated emissions from 1750 to 2011) and is larger than the
752 forcing due to methane emissions for the same period.
753 4.5 Predictability of responses520
754 The increase in cold point temperature and SWV are delayed events. Time is required for it to build up the signals as the
755 aerosols warm the cold point region allowing more water vapour transit into the stratosphere. An approximately stable phase
756 with fluctuations due to the seasonal cycle is attained a couple of months after eruption. The cold point warming has stabilized.
757 The SWV forcing is mainly determined by the additional SWV entering the stratosphere each month as the tropopause region
758 is the region in which WV has the strongest radiative effect. The corresponding build up of tropical SWV forcing due to an525
759 accumulation of SWV in the stratosphere is therefore counteracted by the transport to higher altitudes and from the tropics to
760 the polar region by the Brewer Dobson Circulation.
761 Due to the transient character of the processes - the build up of the forcing and the consequent decline as well as the time
762 9
763 AtmosphericCO
764 2
765 concentration increased from (278±2)ppmmto (390.5±0.2)ppmm.
766 10
767 Methane concentration increased from (722±25)ppbto (1803±2)ppb.
768 29
769
770 shift between effect and response - our analysis showed that a linear relationship between AOD and cold point warming, or
771 SWV, can not be determined for the entire er

771 of forcing due toCO
772 2
773 in the 2000s starting with the 5TgS run. The peak adjusted tropical SWV forcing for the 40TgS530
774 scenario amounts to one third of the totalCO
775 2
776 forcing (due to the accumulated emissions from 1750 to 2011) and is larger
777 than the forcing due to methane emissions for the same period.
778 4.5 Predictability of responses
779 The increase in cold point temperature and SWV are delayed events. Time is required for it to build up the signals as the
780 aerosols warm the cold point region allowing more water vapour transit into the stratosphere. An approximately stable phase535
781 2
782 AtmosphericCO
783 2
784 concentration increased from (278±2)ppmmto (390.5±0.2)ppmm.
785 3
786 Methane concentration increased from (722±25)ppbto (1803±2)ppb.
787 29
788
789 with fluctuations due to the seasonal cycle is attained a couple of months after eruption. The cold point warming has stabilized.
790 The SWV forcing is mainly determined by the additional SWV entering the stratosphere each month as the tropopause region
791 is the region in which WV has the strongest radiative effect. The corresponding build up of tropical SWV forcing due to an
792 accumulation of SWV in the stratosphere is therefore counteracted by the transport to higher altitudes and to the polar region
793 by the BDC.540
794 Due to the transient character of the processes - the build up of the forcing and the consequent decline as well as the time
795 shift between effect and response - our analysis showed that a linear relationship between AOD and cold point warming, or
796 SWV, can not be determined for the entire era

	a after the volcanic eruption. Time phases have to be considered as well to account530		after the volcanic eruption. Time phases have to be considered as well to account
772	for the resulting hysteresis. Other parameters additionally constrain the cold point warming and SWV response. The season	797	for the resulting hysteresis. Other parameters additionally constrain the cold point warming and SWV response. The season
773	during which the eruption occurs will influence the impact of the aerosols: The CP warming will be most effective during the	798	during which the eruption occurs will influence the impact of the aerosols: The CP warming will be most effective during the545
774	peak times of the tape recorder signal around September. Each volcanic eruption is different and parameters like the eruption	799	peak times of the tape recorder signal around September. Each volcanic eruption is different and parameters like the eruption
775	location, amount of emitted sulfur, and specifically the location of the aerosol layer will have a massive influence both on cold	800	location, amount of emitted sulfur, and specifically the location of the aerosol layer will have a massive influence both on cold
776	point warming and SWV entry/forcing. Feedbacks like the SWV forcing in the TTL will enhance the CP warming and lead to535	801	point warming and SWV entry/forcing. Feedbacks like the SWV forcing in the TTL will enhance the CP warming and lead to
777	higher SWV forcing at same AODs eventually before the aerosols fall out.	802	higher SWV forcing at same AODs eventually before the aerosols fall out.
778	Nevertheless an approximately linear relationship	803	Nevertheless, approximately a power function realtionship of formaAOD
779	11	804	b
780	can be found in our simulations for CP warming and SWV forcing with	805	+ccan be found in our simulations for CP warming550
781	respect to IR-AOD after the volcanic eruptions when taking yearly averages. These results however must be interpreted with	806	and SWV forcing with respect to IR-AOD after the volcanic eruptions when taking yearly averages. These results however must
782	care and the derived formulas are only applicable for tropical eruptions occurring in June and reaching to similar heights in the	807	be interpreted with care and the derived formulas are only applicable for tropical eruptions occurring in June and reaching to
783	stratosphere.540	808	similar heights in the stratosphere.
784	Generally, the knowledge both of the respective CP warming in the inner tropics for a specific volcanic eruption and the	809	Generally, the knowledge - both of the respective CP warming in the inner tropics for a specific volcanic eruption and the
785	respective background SWV in the cold point region allows for a relatively accurate prediction of the increase in SWV levels	810	respective background SWV in the cold point region - allows for a relatively accurate prediction of the increase in SWV levels555
786	when using a 12 % SWV increase per Kelvin warming in the mean CP values.	811	when using a 12 % SWV increase per Kelvin warming in the mean CP values.
787	5 Conclusion and Outlook	812	5 Conclusion and Outlook
788	Our study of EVAens and the historical simulations led us to draw the following conclusions:545	813	Our study of EVAens and the historical simulations led us to draw the following conclusions:
789	1. The analysis of the ensemble runs showed the difficulty to extract the volcanic signal in the SWV if only individual	814	1. The analysis of the ensemble runs showed the difficulty to extract the volcanic signal in the SWV if only individual
790	observations are available as internal variability of SWV in the control run could produce SWV values as large as the	815	observations are available as internal variability of SWV in the control run could produce SWV values as large as the560
791	variation found for some of the 10TgS ensemble members in our simulations. Ensemble mean SWV increases are	816	variation found for some of the 10TgS ensemble members in our simulations. Ensemble mean SWV increases are
792	already significant	817	already significant (T-test with p=0.05) in the 2.5TgS eruption.
793	12	818	2. The increase in stratospheric water vapour does not remain constant throughout year, but can vary with the seasons. An
794	in the 2.5TgS eruption.		
795	2. The increase in stratospheric water vapour does not remain constant throughout year but can vary with the seasons. An550		
796	amplification of the seasonal cycle could be observed, especially in the 20TgS and 40TgS scenarios.	819	amplification of the seasonal cycle could be observed, especially in the 20TgS and 40TgS scenarios.

797 3. The average WV at the cold point enterin
g the stratosphere after volcanic eruptions
can be approximated with only minor
798 errors using the mean saturation water vapou
r pressure over ice at the respective averag
e tropical cold point temperature
799 for all investigated aerosol profiles of the
EVAens. When given a base value of an unpert
urbed atmospheric state, a 12 %
800 SWV increase per K increase in cold point te
mperature is a good first estimate for erupt
ion strengths up to 40TgS.555
801 4. The comparison of the idealized EVAens s
imulations and MPI-GE historical simulations
of Mt. Pinatubo show that
802 neither the simulated TOA radiative imbalanc
e nor the estimated amount of emitted strato
spheric sulfur suffice to con-
803 11
804 The second order term does not contribute si
gnificantly since all input values are small
er than 1 in the AOD range of the volcanic e
ruptions.
805 12
806 T-test comparison to reference run with p=0.
05.
807 30
808
809 strain the SWV increases. However the aeroso
l layer shape and height with respect to the
tropopause play a crucial and
810 dominating role when estimating the SWV incr
eases.
811 5. The adjusted tropical forcing caused by
the additional SWV last longer than the for
cing caused by the aerosol layer itself560
812 and its contribution to the total volcanic f
orcing grows with time as the aerosols fall
out. For the 20Tgand the 40Tg
813 scenarios the TOA radiative imbalance shows
a positive value by the end of the simulati
on. Part of this positive TOA
814 radiative imbalance can be attributed to the
forcing by the additional SWV.
815 6. When considering also the time dependenc
e an approximately linear relationship betwe
en yearly averaged tropical cold
816 point warming/adjusted SWV forcing and IR-AO
D can be deduced. This relationship however
only holds for comparable565
817 eruptions occurring at the same time within
the year in the tropics. Additionally the f
inal eruption height and aerosol
818 profile shape has to match the one used with
in our framework.
819 Based on our study different follow up quest
ions for future investigations arise. Our st
udy only focuses on the indirect
820 injection and although other studies focus o
n the direct injection no study know to us c

820 3. The average WV at the cold point entering
the stratosphere after volcanic eruptions can
be approximated with only minor565
821 errors using the mean saturation water vapour
pressure over ice at the respective average t
ropical cold point temperature
822 30
823
824 for all investigated aerosol profiles of the
EVAens. When given a base value of an unpert
urbed atmospheric state, a 12 %
825 SWV increase per K increase in cold point tem
perature is a good first estimate for eruptio
n strengths up to 40TgS.
826 4. The comparison of the idealized EVAens si
mulations and MPI-GE historical simulations o
f Mt. Pinatubo show that
827 neither the simulated TOA radiative imbalance
nor the estimated amount of emitted stratosph
eric sulfur suffice to con-570
828 strain the SWV increases. However, the aeroso
l layer shape and height with respect to the
tropopause play a crucial and
829 dominating role when estimating the SWV incre
ases.
830 5. The adjusted tropical forcing caused by t
he additional SWV last longer than the forcin
g caused by the aerosol layer itself
831 and its contribution to the total volcanic fo
rcing grows with time as the aerosols fall ou
t. For the 20Tgand the 40Tg
832 scenarios the TOA radiative imbalance shows a
positive value by the end of the simulation.
Part of this positive TOA575
833 radiative imbalance can be attributed to the
forcing by the additional SWV.
834 6. When considering also the time dependence
a power function relationship of formaAOD
835 b
836 +cbetween yearly averaged
837 tropical cold point warming/adjusted SWV forc
ing and IR-AOD can be deduced. This relations
hip however only holds
838 for comparable eruptions occurring at the sam
e time of the year in the tropics. Additional
ly, the final eruption height and
839 aerosol profile shape has to match the one us
ed within our framework.580

ombines both effects allowing for
821 a direct comparison within one single framework. Using integrated plume models for a combined study would allow the570
822 quantification of the entire SWV changes and for an estimation of their relative importance. As we use no interactive chemistry
823 our estimate of the H
824 2
825 O might be overestimating the SWV from the indirect injection remaining in the stratosphere. A study
826 using interactive chemistry would enable the assessment of the impact of H
827 2
828 O sinks on the volcanically induced SWV increase
829 as well as ozone chemistry. Of particular interest are the impact of the additional H
830 2
831 O on the oxidation process of SO
832 2
833 , as well
834 as on sulfate particle formation and growth, and a follow up study on this topic could enable a lower boundary estimate to be575
835 made for the SWV increases and the changed aerosol lifetime (compare Case et al. (2015), Kilian et al. (2020)). Additionally
836 the mechanism of the indirect injection has implications beyond that of a volcanic eruption: As geoengineering scenarios also
837 apply sulfur derivatives in the stratosphere, an investigation of the long term SWV signal within these scenarios may be of
838 interest as well.
839 Code and data availability.Details to the EVAens primary data is published with the first publication on the EVAens by Azoulay et al.580
840 (2020). The MPI-ESM historical simulations are described by Maher et al. (2019). The 1D RCE model konrad is available online under
841 https://github.com/atmtools/konrad.
842 31

840 Based on our study follow up questions for future investigations arise. Our study only focuses on the indirect temperature
841 controlled injection and, although other studies focus on the direct injection, no study known to us combines both effects
842 allowing for a direct comparison within one single framework. Using integrated plume models for a combined study would
843 allow the quantification of the entire SWV changes and for an estimation of their relative importance. As we use no interac-
844 tive chemistry, our estimate of the H
845 2
846 O might be overestimating the SWV from the indirect temperature controlled injection585
847 remaining in the stratosphere. A study using interactive chemistry would allow the assessment of the impact of H
848 2
849 O sinks
850 on the volcanically induced SWV increase and ozone chemistry. Of particular interest are the impact of the additional H
851 2
852 O
853 on the oxidation process of SO
854 2
855 , as well as on sulfate particle formation and growth. A follow up study on this topic could
856 give a lower boundary estimate for the SWV increases and the changed aerosol lifetime (compare Case et al. (2015), Kilian
857 et al. (2020)). Additionally, the mechanism of the indirect temperature controlled injection has implications beyond that of a590
858 volcanic eruption: As geoengineering scenarios also apply sulfur derivatives in the stratosphere, an investigation of the long
859 term SWV signal within these scenarios may be of interest as well (e.g. Boucher et al. (2017)).
860 Code and data availability.Primary data and scripts used in the analysis that may be useful in reproducing the author’s work are archived by
861 the Max Planck Institute for Meteorology and can be obtained via https://pure.mpg.de/pubman/faces/ViewItemOverviewPage.jsp?itemId=item_3270686.
862 A more comprehensive description of the EVAens is provided by Azoulay et al., submitted to JGR (2020). Further information was archived595
863 by the Max Planck Institute for Meteorology under http://hdl.handle.net/21.11116/0000-0007-8B38-E. The 1D RCE model konrad is avail-
864 able online under https://github.com/atmtools/konrad.
865 31

843

844 Appendix A: Scaling of different physical parameters altered by the presence of volcanic aerosols

845 Following the discussion in Sect. 3.1, 3.2 and 3.6 the graphs for TOA imbalance (Fig. A1), surface temperature (Fig. A2), cold

846 point temperature changes (Fig. A3) and SWV forcing (Figure A5) are shown which each ensemble mean divided by the mass585

847 of emitted sulfur. In case of the cold point temperature changes and the SWV forcing the dependence of monthly mean values

848 of cold point temperature change and SWV forcing are shown as a function of AOD in the IR waveband of [8475,9259] nm in

849 Fig. A4 and A6.

850 Figure A1.Scaled top of the atmosphere (TOA) imbalance for the five volcanically perturbed EVAens runs (2.5TgS, 5TgS, 10TgS, 20

851 TgS and 40TgS). The ensemble means are shown. The vertical blue line marks the eruption time. All TOA imbalances are divided by the

852 mass of emitted sulfur. Incoming fluxes are defined positive, outgoing fluxes are defined negative.

853 Figure A2.Scaled surface temperature for the five volcanically perturbed EVAens runs (2.5TgS, 5TgS, 10TgS, 20TgS and 40TgS).

854 The ensemble means are shown. The vertical blue line marks the eruption time. All surface temperatures are divided by the mass of emitted

855 sulfur.

856 32

857

858 Figure A3.Scaled temporal evolution of the mean CP temperature anomaly. The time of the volcanic eruption is indicated by a vertical blue

859 line. All changes in cold point temperature are divided by the mass of emitted sulfur.

860 Figure A4.Scatter plot of AOD in the IR waveband (8475-9250 nm) and CP temperature anomaly for all time steps in the first 2.5 years

861 after the eruption.

862 33

863

864 Figure A5.Scaled time evolution of the adjusted clear sky SW-forcing in the tropical region [-5,5]

865 ◦

866 latitude for all eruption strengths. The

867 blue line marks the eruptions time. All clear sky forcings are divided by the mass of emitted sulfur. Incoming fluxes are defined positive,

868 outgoing fluxes are defined negative

869 Figure A6.Scatter plot of AOD in the IR waveband (8475-9250 nm) and the clear sky SWV forcing for all time steps in the first 2.5 years

870 after the eruption.

871 Appendix B: Percental changes in the tape recorder signal

872 Complementary to the plot in Sect. 3.3 we sh

866

867 Appendix A: Scaling of different physical parameters altered by the presence of volcanic aerosols

868 Following the discussion in Sect. 3.1, 3.2 and 3.6 the graphs for TOA imbalance (Fig. A1), surface temperature (Fig. A2), cold

869 point temperature changes (Fig. A3) and SWV forcing (Figure A5) are shown which each ensemble mean divided by the mass600

870 of emitted sulfur. In case of the cold point temperature changes and the SWV forcing the dependence of monthly mean values

871 of cold point temperature change and SWV forcing are shown as a function of AOD in the IR waveband of [8475,9259] nm in

872 Fig. A4 and A6.

873 Figure A1.Scaled top of the atmosphere (TOA) imbalance for the five volcanically perturbed EVAens runs (2.5TgS, 5TgS, 10TgS, 20

874 TgS and 40TgS). The ensemble means are shown. The vertical blue line marks the eruption time. All TOA imbalances are divided by the

875 mass of emitted sulfur. Incoming fluxes are defined positive, outgoing fluxes are defined negative.

876 Figure A2.Scaled surface temperature for the five volcanically perturbed EVAens runs (2.5TgS, 5TgS, 10TgS, 20TgS and 40TgS).

877 The ensemble means are shown. The vertical blue line marks the eruption time. All surface temperatures are divided by the mass of emitted

878 sulfur.

879 32

880

881 Figure A3.Scaled temporal evolution of the mean CP temperature anomaly. The time of the volcanic eruption is indicated by a vertical blue

882 line. All changes in cold point temperature are divided by the mass of emitted sulfur.

883 Figure A4.Scatter plot of AOD in the IR waveband (8475-9250 nm) and CP temperature anomaly for all time steps in the first 2.5 years

884 after the eruption.

885 33

886

887 Figure A5.Scaled time evolution of the adjusted clear sky SW-forcing in the tropical region [-5,5]

888 ◦

889 latitude for all eruption strengths. The

890 blue line marks the eruptions time. All clear sky forcings are divided by the mass of emitted sulfur. Incoming fluxes are defined positive,

891 outgoing fluxes are defined negative

892 Figure A6.Scatter plot of AOD in the IR waveband (8475-9250 nm) and the clear sky SWV forcing for all time steps in the first 2.5 years

893 after the eruption.

894 Appendix B: Percental changes in the tape recorder signal

895 Complementary to the plot in Sect. 3.3 we sho

ow the differences in water vapour between perturbed and unperturbed state with590
873 respect to the unperturbed state in percent.
874 34
875
876 Figure B1.Percental difference in water vapour above 140hPain the tropical average over [-23,23]° latitude for the pure sulfur injections
877 of the EVAens (2.5TgS, 5TgS, 10TgS, 20TgS and 40TgS). The lowermost panel shows the MPI-GE historical simulations for Mt.
878 Pinatubo using the PADS forcing data set. The height of the WMO-tropopause is indicated by a black line, the cold point pressure is shown
879 as black dashed line. In regions not covered by black crosses statistical significant difference between stratospheric water vapour values of
880 the perturbed and unperturbed runs (t-test at p=0.05) were found.
881 35
882
883 Appendix C: Intra-ensemble variability in the entire tropics
884 Here we show the complementary plots to those in Sect. 3.4 for the intra-ensemble variability in the entire tropics [-23,23]
885 ◦
886 latitude.
887 Figure C1.Seasonal averages of specific humidity at cold point as a function of cold point temperature for SON 1991 (a) and 1992 (b)
888 accounting for the entire tropics. Values for each individual ensemble member are shown as dots for the entire tropics. An approximation
889 (see text) for the Clausius Clapeyron equation at this temperature range with an 12%increase of specific humidity perKis shown with a
890 dashed grey line. The exact solution for the Clausius Clapeyron Equation over ice by Murphy and Koop (2005) is calculated for the average
891 ensemble cold point temperatures and pressure and shown in orange.
892 Appendix D: aerosol extinction profiles - 550nmsolar waveband595
893 Complementary to the discussion on the infrared extinction profiles in Sect. 3.5 the 550nmsolar waveband is shown in the
894 following plots.
895 36
896
897 Figure D1.Tropical average of the aerosol extinction profile in the 550nmsolar waveband for the EVA forcing corresponding to 5TgS,
898 10TgS and 20TgS as well as the PADS forcing for the Mt. Pinatubo eruption.
899 Appendix E: SWV forcing - SW component
900 Fig. E1 shows the very small contribution of the SW component to the total adjusted SWV forcing presented in Sect. 3.6.
901 37

w the differences in water vapour between perturbed and unperturbed state with605
896 respect to the unperturbed state in percent.
897 34
898
899 Figure B1.Percental difference in water vapour above 140hPain the tropical average over [-23,23]° latitude for the pure sulfur injections
900 of the EVAens (2.5TgS, 5TgS, 10TgS, 20TgS and 40TgS). The lowermost panel shows the MPI-GE historical simulations for Mt.
901 Pinatubo using the PADS forcing data set. The height of the WMO-tropopause is indicated by a black line, the cold point pressure is shown
902 as black dashed line. In regions not covered by black crosses statistical significant difference between stratospheric water vapour values of
903 the perturbed and unperturbed runs (t-test at p=0.05) were found.
904 35
905
906 Appendix C: Intra-ensemble variability in the entire tropics
907 Here we show the complementary plots to those in Sect. 3.4 for the intra-ensemble variability in the entire tropics [-23,23]
908 ◦
909 latitude.
910 Figure C1.Seasonal averages of specific humidity at cold point as a function of cold point temperature for SON 1991 (a) and 1992 (b)
911 accounting for the entire tropics. Values for each individual ensemble member are shown as dots for the entire tropics. An approximation
912 (see text) for the Clausius Clapeyron equation at this temperature range with an 12%increase of specific humidity perKis shown with a
913 dashed grey line. The exact solution for the Clausius Clapeyron Equation over ice by Murphy and Koop (2005) is calculated for the average
914 ensemble cold point temperatures and pressure and shown in orange.
915 Appendix D: aerosol extinction profiles - 550nmsolar waveband610
916 Complementary to the discussion on the infrared extinction profiles in Sect. 3.5 the 550nmsolar waveband is shown in the
917 following plots.
918 36
919
920 Figure D1.Tropical average of the aerosol extinction profile in the 550nmsolar waveband for the EVA forcing corresponding to 5TgS,
921 10TgS and 20TgS as well as the PADS forcing for the Mt. Pinatubo eruption.
922 Appendix E: SWV forcing - SW component
923 Fig. E1 shows the very small contribution of the SW component to the total adjusted SWV forcing presented in Sect. 3.6.
924 37

902		925	
903	Figure E1.Time evolution of the SW-component to the adjusted clear sky forcing in the tropical region [-5,5]	926	Figure E1.Time evolution of the SW-component to the adjusted clear sky forcing in the tropical region [-5,5]
904	◦	927	◦
905	latitude for the ensemble	928	latitude for the ensemble
906	mean SWV increases caused by all eruption strengths. The blue line marks the eruptions time. The flux range originating from the standard	929	mean SWV increases caused by all eruption strengths. The blue line marks the eruptions time. The flux range originating from the standard
907	deviations of the SWV profiles are plotted to visualize the signal range.	930	deviations of the SWV profiles are plotted to visualize the signal range.
908	The total forcing and its SW component for the cloudy sky case as discussed in Sect. 3.6 is shown in Fig. E2.600	931	The total forcing and its SW component for the cloudy sky case as discussed in Sect. 3.6 is shown in Fig. E2.615
909	Figure E2.Time evolution of the (a) total (b) SW-component to the adjusted all sky forcing in the tropical region [-5,5]	932	Figure E2.Time evolution of the (a) total (b) SW-component to the adjusted all sky forcing in the tropical region [-5,5]
910	◦	933	◦
911	latitude for the	934	latitude for the
912	ensemble mean SWV increases caused by all eruption strengths. The blue line marks the eruptions time. The flux range originating from the	935	ensemble mean SWV increases caused by all eruption strengths. The blue line marks the eruptions time. The flux range originating from the
913	standard deviations of the SWV profiles are plotted to visualize the signal range.	936	standard deviations of the SWV profiles are plotted to visualize the signal range.
914	Appendix F: Global spread of the stratospheric water vapour	937	Appendix F: Global spread of the stratospheric water vapour
915	Figure F1 shows the spread of the additional SWV around the globe as mentioned in the Discussion Sect. 4.4.	938	Figure F1 shows the spread of the additional SWV around the globe as mentioned in the Discussion Sect. 4.4.
916	38	939	38
917		940	
918	Figure F1.Difference in stratospheric water vapour content at 70 hPa as a function of time and latitude. Black crosses mark the regions of	941	Figure F1.Difference in stratospheric water vapour content at 70hPaas a function of time and latitude. Black crosses mark the regions of
919	statistical significance of the data (Mann Whitney U Test at p=0.05).	942	statistical significance of the data (Mann Whitney U Test at p=0.05).
920	39	943	39
921		944	
922	Author contributions.CK, CT and HS designed the study. CK conducted the analysis/investigation and wrote the paper. SD contributed to	945	Author contributions.CK, CT and HS designed the study. CK conducted the analysis/investigation and wrote the paper. SD contributed to
923	the development of the methodology to calculate the SWV forcing with konrad and the interpretation of the corresponding result. AA set up	946	the development of the methodology to calculate the SWV forcing with konrad and the interpretation of the corresponding result. AA set up
924	the EVAens simulations. CT, SD and HS contributed to the writing of the paper.605	947	the EVAens simulations. CT, SD and HS contributed to the writing of the paper.620
925	Competing interests.The authors declare that they have no conflict of interest.	948	Competing interests.The authors declare that they have no conflict of interest.
926	Acknowledgements.This research has been supported by the Deutsche Forschungsgemeinschaft (DFG) Research Unit VolImpact (FOR2820)	949	Acknowledgements.This research has been supported by the Deutsche Forschungsgemeinschaft (DFG) Research Unit VolImpact (FOR2820)
927	within the projects VolDyn (TO 967/2-1) and VolClim (TI 344/2-1). SD and CK were/are members of the International Max Planck Research	950	within the projects VolDyn (TO 967/2-1) and VolClim (TI 344/2-1). SD and CK were/are members of the International Max Planck Research
928	School (IMPRS). The data was processed using CDO (https://code.mpimet.mpg.de/projects/cdo/embedded/cdo.pdf) and using the computing	951	School (IMPRS). The data was processed using CDO (https://code.mpimet.mpg.de/projects/cdo/embedded/cdo.pdf) and using the computing
929	facilities at the Deutsche Klimarechenzentrum (DKRZ). We thank Lukas Kluft for advising	952	facilities at the Deutsche Klimarechenzentrum (DKRZ). We thank Lukas Kluft for advising us on the usage of the 1D RCE model konrad.625

us on the usage of the 1D RCE model konrad. 610
930 40
931
932 References

933 Azoulay, A., Schmidt, H., and Timmreck, C.: The Arctic polar vortex response to volcanic forcing of different strengths,in preparation,
934 2020.
935 Bekki, S.: Oxidation of volcanic SO_2: A sink for stratospheric OH and H_2O, Geophysical Research Letters, 22, 913–916,
936 https://doi.org/10.1029/95GL00534, 1995.615
937 Brewer, A. W.: Evidence for a world circulation provided by the measurements of helium and water vapour distribution in the stratosphere,
938 Quarterly Journal of the Royal Meteorological Society, 75, 351–363, https://doi.org/10.1002/qj.49707532603, 1949.
939 Case, P. A., Tsigaridis, K., and LeGrande, A. N.: The Effect of Stratospheric Water Vapor in Large Volcanic Eruptions on Climate and
940 Atmospheric Composition, in: AGU Fall Meeting, San Fracisco, California, 14-18 December 2015, A51N-0270, AGU, https://agu.confex.
941 com/agu/fm15/meetingapp.cgi/Paper/79947, 2015.620
942 Coffey, M. T.: Observations of the impact of volcanic activity on stratospheric chemistry, Journal of Geophysical Research: Atmospheres,
943 101, 6767–6780, 1996.
944 Dacie, S., Kluft, L., Schmidt, H., Stevens, B., Buehler, S. A., Nowack, P. J., Dietmüller, S., Abraham, N. L., and Birner, T.: A 1D RCE Study of
945 Factors Affecting the Tropical Tropopause Layer and Surface Climate, Journal of Climate, 32, 6769–6782, https://doi.org/10.1175/JCLI-
946 D-18-0778.1, 2019.625
947 Davis, S. M., Hegglin, M. I., Fujiwara, M., Dragani, R., Harada, Y., Kobayashi, C., Lon

953 40
954
955 References

956 Angell, J. K.: Stratospheric warming due to Agung, El Chichón, and Pinatubo taking into account the quasi-biennial oscillation, Journal of
957 Geophysical Research: Atmospheres, 102, 9479–9485, https://doi.org/https://doi.org/10.1029/96JD03588, https://agupubs.onlinelibrary.
958 wiley.com/doi/abs/10.1029/96JD03588, 1997.
959 Azoulay, A., Schmidt, H., and Timmreck, C.: The Arctic polar vortex response to volcanic forcing of different strengths, J. Geophys. Res.,630
960 submitted, 2020.
961 Bekki, S.: Oxidation of volcanic SO_2: A sink for stratospheric OH and H_2O, Geophysical Research Letters, 22, 913–916,
962 https://doi.org/10.1029/95GL00534, 1995.
963 Boucher, O., Kleinschmitt, C., and Myhre, G.: Quasi-Additivity of the Radiative Effects of Marine Cloud Brightening and Stratospheric
964 Sulfate Aerosol Injection, Geophysical Research Letters, 44, 11,158–11,165, https://doi.org/https://doi.org/10.1002/2017GL074647, https:635
965 //agupubs.onlinelibrary.wiley.com/doi/abs/10.1002/2017GL074647, 2017.
966 Brewer, A. W.: Evidence for a world circulation provided by the measurements of helium and water vapour distribution in the stratosphere,
967 Quarterly Journal of the Royal Meteorological Society, 75, 351–363, https://doi.org/10.1002/qj.49707532603, 1949.
968 Case, P. A., Tsigaridis, K., and LeGrande, A. N.: The Effect of Stratospheric Water Vapor in Large Volcanic Eruptions on Climate and
969 Atmospheric Composition, in: AGU Fall Meeting, San Fracisco, California, 14-18 December 2015, A51N-0270, AGU, https://agu.confex.640
970 com/agu/fm15/meetingapp.cgi/Paper/79947, 2015.
971 Coffey, M. T.: Observations of the impact of volcanic activity on stratospheric chemistry, Journal of Geophysical Research: Atmospheres,
972 101, 6767–6780, 1996.
973 Dacie, S., Kluft, L., Schmidt, H., Stevens, B., Buehler, S. A., Nowack, P. J., Dietmüller, S., Abraham, N. L., and Birner, T.: A 1D RCE Study of
974 Factors Affecting the Tropical Tropopause Layer and Surface Climate, Journal of Climate, 32, 6769–6782, https://doi.org/10.1175/JCLI-645
975 D-18-0778.1, 2019.
976 Davis, S. M., Hegglin, M. I., Fujiwara, M., Dragani, R., Harada, Y., Kobayashi, C., Long,

g, C., Manney, G. L., Nash, E. R., Potter, G. L.,

Tegtmeier, S., Wang, T., Wargan, K., and Wright, J. S.: Assessment of upper tropospheric and stratospheric water vapor and ozone in

reanalyses as part of S-RIP, Atmospheric Chemistry and Physics, 17, 12 743–12 778, https://doi.org/10.5194/acp-17-12743-2017, 2017.

Dee, D. P., Uppala, S. M., Simmons, A. J., Berrisford, P., Poli, P., Kobayashi, S., Andrae, U., Balmaseda, M. A., Balsamo, G., Bauer,

P., Bechtold, P., Beljaars, A. C. M., van de Berg, L., Bidlot, J., Bormann, N., Delsol, C., Dragani, R., Fuentes, M., Geer, A. J., Haim-630

berger, L., Healy, S. B., Hersbach, H., Hólm, E. V., Isaksen, L., Kållberg, P., Köhler, M., Matricardi, M., McNally, A. P., Monge-Sanz,

B. M., Morcrette, J.-J., Park, B.-K., Peubey, C., de Rosnay, P., Tavolato, C., Thépaut, J.-N., and Vitart, F.: The ERA-Interim reanalysis:

configuration and performance of the data assimilation system, Quarterly Journal of the Royal Meteorological Society, 137, 553–597,

https://doi.org/10.1002/qj.828, 2011.

Dessler, A. E., Schoeberl, M. R., Wang, T., Davis, S. M., and Rosenlof, K. H.: Stratospheric water vapor feedback, Proceedings of the635

National Academy of Sciences of the United States of America, 110, 18 087–18 091, https://doi.org/10.1073/pnas.1310344110, 2013.

Dessler, A. E., Schoeberl, M. R., Wang, T., Davis, S. M., Rosenlof, K. H., and Vernier, J.-P.: Variations of stratospheric water vapor over the

past three decades, Journal of Geophysical Research: Atmospheres, 119, 12,588–12,598, https://doi.org/10.1002/2014JD021712, 2014.

Diallo, M., Ploeger, F., Konopka, P., Birner, T., Müller, R., Riese, M., Garny, H., Legras, B., Ray, E., Berthet, G., and Jegou, F.: Significant

Contributions of Volcanic Aerosols to Decadal Changes in the Stratospheric Circulation, Geophysical Research Letters, 44, 10,780–640

10,791, https://doi.org/10.1002/2017GL074662, 2017.

Driscoll, S., Bozzo, A., Gray, L. J., Robock, A., and Stenchikov, G.: Coupled Model Intercomparison Project 5 (CMIP5) simulations of cli-

mate following volcanic eruptions, Journal of Geophysical Research: Atmospheres, 117, n/a–n/a, https://doi.org/10.1029/2012JD017607,

2012.

C., Manney, G. L., Nash, E. R., Potter, G. L.,

Tegtmeier, S., Wang, T., Wargan, K., and Wright, J. S.: Assessment of upper tropospheric and stratospheric water vapor and ozone in

reanalyses as part of S-RIP, Atmospheric Chemistry and Physics, 17, 12 743–12 778, https://doi.org/10.5194/acp-17-12743-2017, 2017.

Dee, D. P., Uppala, S. M., Simmons, A. J., Berrisford, P., Poli, P., Kobayashi, S., Andrae, U., Balmaseda, M. A., Balsamo, G., Bauer,650

P., Bechtold, P., Beljaars, A. C. M., van de Berg, L., Bidlot, J., Bormann, N., Delsol, C., Dragani, R., Fuentes, M., Geer, A. J., Haim-

berger, L., Healy, S. B., Hersbach, H., Hólm, E. V., Isaksen, L., Kållberg, P., Köhler, M., Matricardi, M., McNally, A. P., Monge-Sanz,

B. M., Morcrette, J.-J., Park, B.-K., Peubey, C., de Rosnay, P., Tavolato, C., Thépaut, J.-N., and Vitart, F.: The ERA-Interim reanalysis:

configuration and performance of the data assimilation system, Quarterly Journal of the Royal Meteorological Society, 137, 553–597,

https://doi.org/10.1002/qj.828, 2011.655

Dessler, A. E., Schoeberl, M. R., Wang, T., Davis, S. M., and Rosenlof, K. H.: Stratospheric water vapor feedback, Proceedings of the

National Academy of Sciences of the United States of America, 110, 18 087–18 091, https://doi.org/10.1073/pnas.1310344110, 2013.

Dessler, A. E., Schoeberl, M. R., Wang, T., Davis, S. M., Rosenlof, K. H., and Vernier, J.-P.: Variations of stratospheric water vapor over the

past three decades, Journal of Geophysical Research: Atmospheres, 119, 12,588–12,598, https://doi.org/10.1002/2014JD021712, 2014.

Diallo, M., Ploeger, F., Konopka, P., Birner, T., Müller, R., Riese, M., Garny, H., Legras, B., Ray, E., Berthet, G., and Jegou, F.: Significant660

Contributions of Volcanic Aerosols to Decadal Changes in the Stratospheric Circulation, Geophysical Research Letters, 44, 10,780–

10,791, https://doi.org/10.1002/2017GL074662, 2017.

41

Driscoll, S., Bozzo, A., Gray, L. J., Robock, A., and Stenchikov, G.: Coupled Model Intercomparison Project 5 (CMIP5) simulations of cli-

mate following volcanic eruptions, Journal of Geophysical Research: Atmospheres, 117, n/a–n/a, https://doi.org/10.1029/2012JD017607,

2012.665

Fels, S. B., Mahlman, J., Schwarzkopf, M., and Sinclair, R.: Stratosph

966 Forster, P. M., Richardson, T., Maycock, A. C., Christopher J. Smith, C. J., Samset, B. H., Myhre, G., Timothy Andrews, Timothy Pincus,
967 R., and Schulz, M.: Recommendations for diagnosing effective radiative forcing from climate models for CMIP6, Journal of Geophysical
968 Resarch:
969 A
970 tmosphere,121,12460--12475, https://doi.org/10.1002/2016JD025320,2016.
971 41
972
973 Forster, P. M. d. F. and Shine, K. P.: Assessing the climate impact of trends in stratospheric water vapor, Geophysical Research Letters, 29,645
974 10–1–10–4, https://doi.org/10.1029/2001GL013909, 2002.
975 Fueglistaler, S., Dessler, A. E., Dunkerton, T. J., Folkins, I., Fu, Q., and Mote, P. W.: Tropical tropopause layer, Reviews of Geophysics, 47,
976 1606, https://doi.org/10.1029/2008RG000267, 2009.
977 Fueglistaler, S., Liu, Y. S., Flannaghan, T. J., Haynes, P. H., Dee, D. P., Read, W. J., Remsberg, E. E., Thomason, L. W., Hurst, D. F., Lanzante,
978 J. R., and Bernath, P. F.: The relation between atmospheric humidity and temperature trends for stratospheric water, Journal of Geophysical650
979 Research: Atmospheres, 118, 1052–1074, https://doi.org/10.1002/jgrd.50157, 2013.
980 Fujiwara, M., Wright, J. S., Manney, G. L., Gray, L. J., Anstey, J., Birner, T., Davis, S., Gerber, E. P., Harvey, V. L., Hegglin, M. I., Homeyer,
981 C. R., Knox, J. A., Krüger, K., Lambert, A., Long, C. S., Martineau, P., Molod, A., Monge-Sanz, B. M., Santee, M. L., Tegtmeier, S.,
982 Chabrillat, S., Tan, D. G. H., Jackson, D. R., Polavarapu, S., Compo, G. P., Dragani, R., Ebisuzaki, W., Harada, Y., Kobayashi, C., McCarty,
983 W., Onogi, K., Pawson, S., Simmons, A., Wargan, K., Whitaker, J. S., and Zou, C.-Z.: Introduction to the SPARC Reanalysis Intercomparison655
984 Project (S-RIP) and overview of the reanalysis systems, Atmospheric Chemistry and Physics, 17, 1417–1452, https://doi.org/10.5194/acp-

eric Sensitivity to Perturbations in Ozone and Carbon
998 Dioxide: Radiative and Dynamical Response., Journal of Atmospheric Sciences, 37, 2265–2297, https://doi.org/10.1175/1520-
999 0469(1980)037<2265:SSTPIO>2.0.CO;2, 1980.
1000 Forster, P. M., Richardson, T., Maycock, A. C., Christopher J. Smith, C. J., Samset, B. H., Myhre, G., Timothy Andrews, Timothy Pincus,
1001 R., and Schulz, M.: Recommendations for diagnosing effective radiative forcing from climate models for CMIP6, Journal of Geophysical
1002 Resarch:
1003 A
1004 tmosphere,121,12460--12475, https://doi.org/10.1002/2016JD025320,2016.
1005 Forster, P. M. d. F. and Shine, K. P.: Assessing the climate impact of trends in stratospheric water vapor, Geophysical Research Letters, 29,
1006 10–1–10–4, https://doi.org/10.1029/2001GL013909, 2002.670
1007 Fueglistaler, S., Dessler, A. E., Dunkerton, T. J., Folkins, I., Fu, Q., and Mote, P. W.: Tropical tropopause layer, Reviews of Geophysics, 47,
1008 1606, https://doi.org/10.1029/2008RG000267, 2009.
1009 Fueglistaler, S., Liu, Y. S., Flannaghan, T. J., Haynes, P. H., Dee, D. P., Read, W. J., Remsberg, E. E., Thomason, L. W., Hurst, D. F., Lanzante,
1010 J. R., and Bernath, P. F.: The relation between atmospheric humidity and temperature trends for stratospheric water, Journal of Geophysical
1011 Research: Atmospheres, 118, 1052–1074, https://doi.org/10.1002/jgrd.50157, 2013.675
1012 Fujiwara, M., Wright, J. S., Manney, G. L., Gray, L. J., Anstey, J., Birner, T., Davis, S., Gerber, E. P., Harvey, V. L., Hegglin, M. I., Homeyer,
1013 C. R., Knox, J. A., Krüger, K., Lambert, A., Long, C. S., Martineau, P., Molod, A., Monge-Sanz, B. M., Santee, M. L., Tegtmeier, S.,
1014 Chabrillat, S., Tan, D. G. H., Jackson, D. R., Polavarapu, S., Compo, G. P., Dragani, R., Ebisuzaki, W., Harada, Y., Kobayashi, C., McCarty,
1015 W., Onogi, K., Pawson, S., Simmons, A., Wargan, K., Whitaker, J. S., and Zou, C.-Z.: Introduction to the SPARC Reanalysis Intercomparison
1016 Project (S-RIP) and overview of the reanalysis systems, Atmospheric Chemistry and Physics, 17, 1417–1452, https://doi.org/10.5194/acp-680

985	17-1417-2017, 2017.	1017	17-1417-2017, 2017.
986	Gelaro, R., McCarty, W., Suárez, M. J., Todling, R., Molod, A., Takacs, L., Randles, C., Darmenov, A., Bosilovich, M. G., Reichle, R., Wargan,	1018	Gelaro, R., McCarty, W., Suárez, M. J., Todling, R., Molod, A., Takacs, L., Randles, C., Darmenov, A., Bosilovich, M. G., Reichle, R., Wargan,
987	K., Coy, L., Cullather, R., Draper, C., Akella, S., Buchard, V., Conaty, A., da Silva, A., Gu, W., Kim, G.-K., Koster, R., Lucchesi, R.,	1019	K., Coy, L., Cullather, R., Draper, C., Akella, S., Buchard, V., Conaty, A., da Silva, A., Gu, W., Kim, G.-K., Koster, R., Lucchesi, R.,
988	Merkova, D., Nielsen, J. E., Partyka, G., Pawson, S., Putman, W., Rienecker, M., Schubert, S. D., Sienkiewicz, M., and Zhao, B.: The 660	1020	Merkova, D., Nielsen, J. E., Partyka, G., Pawson, S., Putman, W., Rienecker, M., Schubert, S. D., Sienkiewicz, M., and Zhao, B.: The
989	Modern-Era Retrospective Analysis for Research and Applications, Version 2 (MERRA-2), Journal of Climate, Volume 30, 5419–5454,	1021	Modern-Era Retrospective Analysis for Research and Applications, Version 2 (MERRA-2), Journal of Climate, Volume 30, 5419–5454,685
990	https://doi.org/10.1175/JCLI-D-16-0758.1, 2017.	1022	https://doi.org/10.1175/JCLI-D-16-0758.1, 2017.
991	Giorgetta, M. A., Jungclaus, J., Reick, C. H., Legutke, S., Bader, J., Böttinger, M., Brovkin, V., Crueger, T., Esch, M., Fieg, K., Glushak, K.,	1023	Giorgetta, M. A., Jungclaus, J., Reick, C. H., Legutke, S., Bader, J., Böttinger, M., Brovkin, V., Crueger, T., Esch, M., Fieg, K., Glushak, K.,
992	Gayler, V., Haak, H., Hollweg, H.-D., Ilyina, T., Kinne, S., Kornblueh, L., Matei, D., Mauritsen, T., Mikolajewicz, U., Mueller, W., Notz, D.,	1024	Gayler, V., Haak, H., Hollweg, H.-D., Ilyina, T., Kinne, S., Kornblueh, L., Matei, D., Mauritsen, T., Mikolajewicz, U., Mueller, W., Notz, D.,
993	Pithan, F., Raddatz, T., Rast, S., Redler, R., Roeckner, E., Schmidt, H., Schnur, R., Segschneider, J., Six, K. D., Stockhause, M., Timmreck,665	1025	Pithan, F., Raddatz, T., Rast, S., Redler, R., Roeckner, E., Schmidt, H., Schnur, R., Segschneider, J., Six, K. D., Stockhause, M., Timmreck,
994	C., Wegner, J., Widmann, H., Wieners, K.-H., Claussen, M., Marotzke, J., and Stevens, B.: Climate and carbon cycle changes from 1850 to	1026	C., Wegner, J., Widmann, H., Wieners, K.-H., Claussen, M., Marotzke, J., and Stevens, B.: Climate and carbon cycle changes from 1850 to 690
995	2100 in MPI-ESM simulations for the Coupled Model Intercomparison Project phase 5, Journal of Advances in Modeling Earth Systems, 5,	1027	2100 in MPI-ESM simulations for the Coupled Model Intercomparison Project phase 5, Journal of Advances in Modeling Earth Systems, 5,
996	572–597, https://doi.org/10.1002/jame.20038, 2013.	1028	572–597, https://doi.org/10.1002/jame.20038, 2013.
997	Glaze, L. S., Baloga, S. M., and Wilson, L.: Transport of atmospheric water vapor by volcanic eruption columns, Journal of Geophysical	1029	Glaze, L. S., Baloga, S. M., and Wilson, L.: Transport of atmospheric water vapor by volcanic eruption columns, Journal of Geophysical
998	Research: Atmospheres, 102, 6099–6108, https://doi.org/10.1029/96JD03125, 1997.670	1030	Research: Atmospheres, 102, 6099–6108, https://doi.org/10.1029/96JD03125, 1997.
999	Hall, T. M. and Waugh, D.: Tracer transport in the tropical stratosphere due to vertical diffusion and horizontal mixing, Geophysical Research	1031	Hall, T. M. and Waugh, D.: Tracer transport in the tropical stratosphere due to vertical diffusion and horizontal mixing, Geophysical Research695
1000	Letters, 24, 1383–1386, https://doi.org/10.1029/97GL01289, 1997.	1032	Letters, 24, 1383–1386, https://doi.org/10.1029/97GL01289, 1997.
		1033	42
		1034	
1001	Hansen, J., Sato, M., Ruedy, R., Nazarenko, L., Lacis, A., Schmidt, G. A., Russell, G., Aleinov, I., Bauer, M., Bauer, S., Bell, N., Cairns, B.,	1035	Hansen, J., Sato, M., Ruedy, R., Nazarenko, L., Lacis, A., Schmidt, G. A., Russell, G., Aleinov, I., Bauer, M., Bauer, S., Bell, N., Cairns, B.,
1002	Canuto, V., Chandler, M., Cheng, Y., Del Genio, A., Faluvegi, G., Fleming, E., Friend, A., Hall, T., Jackman, C., Kelley, M., Kiang, N.,	1036	Canuto, V., Chandler, M., Cheng, Y., Del Genio, A., Faluvegi, G., Fleming, E., Friend, A., Hall, T., Jackman, C., Kelley, M., Kiang, N.,
1003	Koch, D., Lean, J., Lerner, J., Lo, K., Menon, S., Miller, R., Minnis, P., Novakov, T.,	1037	Koch, D., Lean, J., Lerner, J., Lo, K., Menon, S., Miller, R., Minnis, P., Novakov, T., O

	Oinas, V., Perlwitz, J., Perlwitz, J., Rind, D., Romanou,675		inas, V., Perlwitz, J., Perlwitz, J., Rind, D., Romanou,
1004	A., Shindell, D., Stone, P., Sun, S., Tausnev, N., Thresher, D., Wielicki, B., Wong, T., Yao, M., and Zhang, S.: Efficacy of climate forcings,	1038	A., Shindell, D., Stone, P., Sun, S., Tausnev, N., Thresher, D., Wielicki, B., Wong, T., Yao, M., and Zhang, S.: Efficacy of climate forcings,700
1005	Journal of Geophysical Research: Atmospheres, 110, https://doi.org/10.1029/2005JD005776, https://agupubs.onlinelibrary.wiley.com/doi/	1039	Journal of Geophysical Research: Atmospheres, 110, https://doi.org/10.1029/2005JD005776, https://agupubs.onlinelibrary.wiley.com/doi/
1006	abs/10.1029/2005JD005776, 2005.	1040	abs/10.1029/2005JD005776, 2005.
		1041	Huang, Y., Wang, Y., and Huang, H.: Stratospheric Water Vapor Feedback Disclosed by a Locking Experiment, Geophysical Research Let-
		1042	ters, 47, e2020GL087 987, https://doi.org/https://doi.org/10.1029/2020GL087987, https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/
		1043	2020GL087987, e2020GL087987 2020GL087987, 2020.705
1007	Ilyina, T., Six, K. D., Segschneider, J., Maier-Reimer, E., Li, H., and Núñez-Riboni, I.: Global ocean biogeochemistry model HAMOCC:	1044	Ilyina, T., Six, K. D., Segschneider, J., Maier-Reimer, E., Li, H., and Núñez-Riboni, I.: Global ocean biogeochemistry model HAMOCC:
1008	Model architecture and performance as component of the MPI-Earth system model in different CMIP5 experimental realizations, Journal of680	1045	Model architecture and performance as component of the MPI-Earth system model in different CMIP5 experimental realizations, Journal of
1009	Advances in Modeling Earth Systems, 5, 287–315, https://doi.org/10.1029/2012MS000178, 2013.	1046	Advances in Modeling Earth Systems, 5, 287–315, https://doi.org/10.1029/2012MS000178, 2013.
1010	42		
1011			
1012	Joshi, M. M. and Jones, G. S.: The climatic effects of the direct injection of water vapour into the stratosphere by large volcanic eruptions,	1047	Joshi, M. M. and Jones, G. S.: The climatic effects of the direct injection of water vapour into the stratosphere by large volcanic eruptions,
1013	Atmospheric Chemistry and Physics, 9, 6109–6118, https://doi.org/10.5194/acp-9-6109-2009, 2009.	1048	Atmospheric Chemistry and Physics, 9, 6109–6118, https://doi.org/10.5194/acp-9-6109-2009, 2009.710
1014	Joshi, M. M. and Shine, K. P.: A GCM Study of Volcanic Eruptions as a Cause of Increased Stratospheric Water Vapor, Journal of Climate, 16,	1049	Joshi, M. M. and Shine, K. P.: A GCM Study of Volcanic Eruptions as a Cause of Increased Stratospheric Water Vapor, Journal of Climate, 16,
1015	3525–3534, https://doi.org/10.1175/1520-0442(2003)016<3525:AGSOVE>2.0.CO;2, 2003.685	1050	3525–3534, https://doi.org/10.1175/1520-0442(2003)016<3525:AGSOVE>2.0.CO;2, 2003.
1016	Jungclaus, J. H., Fischer, N., Haak, H., Lohmann, K., Marotzke, J., Matei, D., Mikolajewicz, U., Notz, D., and von Storch, J. S.: Characteristics	1051	Jungclaus, J. H., Fischer, N., Haak, H., Lohmann, K., Marotzke, J., Matei, D., Mikolajewicz, U., Notz, D., and von Storch, J. S.: Characteristics
1017	of the ocean simulations in the Max Planck Institute Ocean Model (MPIOM) the ocean component of the MPI-Earth system model, Journal	1052	of the ocean simulations in the Max Planck Institute Ocean Model (MPIOM) the ocean component of the MPI-Earth system model, Journal
1018	of Advances in Modeling Earth Systems, 5, 422–446, https://doi.org/10.1002/jame.20023, 2013.	1053	of Advances in Modeling Earth Systems, 5, 422–446, https://doi.org/10.1002/jame.20023, 2013.715
1019	Kilian, M., Brinkop, S., and Jöckel, P.: Impact of the eruption of Mt Pinatubo on the chemical composition of the stratosphere, Atmospheric	1054	Kilian, M., Brinkop, S., and Jöckel, P.: Impact of the eruption of Mt Pinatubo on the chemical composition of the stratosphere, Atmospheric
1020	Chemistry and Physics, 20, 11 697–11 715, https://doi.org/10.5194/acp-20-11697-2020, https://acp.copernicus.org/articles/20/11697/2020/,690	1055	Chemistry and Physics, 20, 11 697–11 715, https://doi.org/10.5194/acp-20-11697-2020, https://acp.copernicus.org/articles/20/11697/2020/,
1021	2020.	1056	2020.

1022 Kluft, L., Dacie, S., Buehler, S. A., Schmidt, H., and Stevens, B.: Re-Examining the First Climate Models: Climate Sensitivity of a Modern

1023 Radiative-Convective Equilibrium Model, Journal of Climate, 32, 8111–8125, https://doi.org/10.1175/JCLI-D-18-0774.1, 2019.

1024 Kobayashi, S., Ota, Y., Harada, Y., Ebita, A., Miyaoka, M., Onoda, H., Onogi, K., Kamahori, H., Kobayashi, C., Endo, H., Miyaoka, K., and

1025 Takahshini, K.: The JRA-55 Reanalysis: General Specifications and Basic Characteristics, Journal of the Meteorological Society of Japan.695

1026 Ser. II, 93, 5–48, https://doi.org/10.2151/jmsj.2015-001, 2015.

1027 Krishnamohan, K.-P. S.-P., Bala, G., Cao, L., Duan, L., and Caldeira, K.: Climate system response to stratospheric sulfate aerosols: sensitivity

1028 to altitude of aerosol layer, Earth System Dynamics, 10, 885–900, https://doi.org/10.5194/esd-10-885-2019, 2019.

1029 Löffler, M., Brinkop, S., and Jöckel, P.: Impact of major volcanic eruptions on stratospheric water vapour, Atmospheric Chemistry and Physics,

1030 16, 6547–6562, https://doi.org/10.5194/acp-16-6547-2016, 2016.700

1031 Maher, N., Milinski, S., Suarez-Gutierrez, L., Botzet, M., Dobrynin, M., Kornblueh, L., Kröger, J., Takano, Y., Ghosh, R., Hedemann, C., Li, C.,

1032 Li, H., Manzini, E., Notz, D., Putrasahan, D., Boysen, L., Claussen, M., Ilyina, T., Olonscheck, D., Raddatz, T., Stevens, B., and Marotzke,

1033 J.: The Max Planck Institute Grand Ensemble: Enabling the Exploration of Climate System Variability, Journal of Advances in Modeling

1034 Earth Systems, 11, 2050–2069, https://doi.org/10.1029/2019MS001639, 2019.

1035 Marsland, S. J., Haak, H., Jungclaus, J. H., Latif, M., and Röske, F.: The Max-Planck-Institute global ocean/sea ice model with orthogonal705

1036 curvilinear coordinates, Ocean Modelling, 5, 91–127, https://doi.org/10.1016/S1463-5003(02)00015-X, 2003.

1057 Kluft, L., Dacie, S., Buehler, S. A., Schmidt, H., and Stevens, B.: Re-Examining the First Climate Models: Climate Sensitivity of a Modern

1058 Radiative-Convective Equilibrium Model, Journal of Climate, 32, 8111–8125, https://doi.org/10.1175/JCLI-D-18-0774.1, 2019.720

1059 Kobayashi, S., Ota, Y., Harada, Y., Ebita, A., Miyaoka, M., Onoda, H., Onogi, K., Kamahori, H., Kobayashi, C., Endo, H., Miyaoka, K., and

1060 Takahshini, K.: The JRA-55 Reanalysis: General Specifications and Basic Characteristics, Journal of the Meteorological Society of Japan.

1061 Ser. II, 93, 5–48, https://doi.org/10.2151/jmsj.2015-001, 2015.

1062 Krishnamohan, K.-P. S.-P., Bala, G., Cao, L., Duan, L., and Caldeira, K.: Climate system response to stratospheric sulfate aerosols: sensitivity

1063 to altitude of aerosol layer, Earth System Dynamics, 10, 885–900, https://doi.org/10.5194/esd-10-885-2019, 2019.725

1064 LeGrande, A. N., Tsigaridis, K., and Bauer, S. E.: Role of atmospheric chemistry in the climate impacts of stratospheric volcanic injections,

1065 Nature Geoscience, 9, 652–655, https://doi.org/10.1038/ngeo2771, 2016.

1066 Löffler, M., Brinkop, S., and Jöckel, P.: Impact of major volcanic eruptions on stratospheric water vapour, Atmospheric Chemistry and Physics,

1067 16, 6547–6562, https://doi.org/10.5194/acp-16-6547-2016, 2016.

1068 Maher, N., Milinski, S., Suarez-Gutierrez, L., Botzet, M., Dobrynin, M., Kornblueh, L., Kröger, J., Takano, Y., Ghosh, R., Hedemann, C., Li, C.,730

1069 Li, H., Manzini, E., Notz, D., Putrasahan, D., Boysen, L., Claussen, M., Ilyina, T., Olonscheck, D., Raddatz, T., Stevens, B., and Marotzke,

1070 J.: The Max Planck Institute Grand Ensemble: Enabling the Exploration of Climate System Variability, Journal of Advances in Modeling

1071 Earth Systems, 11, 2050–2069, https://doi.org/10.1029/2019MS001639, 2019.

1072 43

1073

1074 Marshall, L., Schmidt, A., Toohey, M., Carslaw, K. S., Mann, G. W., Sigl, M., Khodri, M., Timmreck, C., Zanchettin, D., Ball, W. T., Bekki, S.,

1075 Brooke, J. S. A., Dhomse, S., Johnson, C., Lamarque, J.-F., LeGrande, A. N., Mills, M. J., Niemeier, U., Pope, J. O., Poulain, V., Robock,735

1076 A., Rozanov, E., Stenke, A., Sukhodolov, T.,

1037 Mauritsen, T., Bader, J., Becker, T., Behrens, J., Bittner, M., Brokopf, R., Brovkin, V., Claussen, M., Crueger, T., Esch, M., Fast, I., Fiedler, S.,

1038 Fläschner, D., Gayler, V., Giorgetta, M., Goll, D. S., Haak, H., Hagemann, S., Hedemann, C., Hohenegger, C., Ilyina, T., Jahns, T., Jiménéz-

1039 de-la Cuesta, D., Jungclaus, J., Kleinen, T., Kloster, S., Kracher, D., Kinne, S., Kleberg, D., Lasslop, G., Kornblueh, L., Marotzke, J., Matei,

1040 D., Meraner, K., Mikolajewicz, U., Modali, K., Möbis, B., Müller, W. A., Nabel, J. E. M. S., Nam, C. C. W., Notz, D., Nyawira, S.-S.,710

1041 Paulsen, H., Peters, K., Pincus, R., Pohlmann, H., Pongratz, J., Popp, M., Raddatz, T. J., Rast, S., Redler, R., Reick, C. H., Rohrschnei-

1042 der, T., Schemann, V., Schmidt, H., Schnur, R., Schulzweida, U., Six, K. D., Stein, L., Stemmler, I., Stevens, B., Storch, J.-S., Tian,

1043 F., Voigt, A., Vrese, P., Wieners, K.-H., Wilkenskjeld, S., Winkler, A., and Roeckner, E.: Developments in the MPI‐M Earth System

1044 Model version 1.2 (MPI‐ESM1.2) and Its Response to Increasing CO 2, Journal of Advances in Modeling Earth Systems, 11, 998–1038,

1045 https://doi.org/10.1029/2018MS001400, 2019.715

1046 Mote, P. W., Rosenlof, K. H., McIntyre, M. E., Carr, E. S., Gille, J. C., Holton, J. R., Kinnersley, J. S., Pumphrey, H. C., Russell, J. M.,

1047 and Waters, J. W.: An atmospheric tape recorder: The imprint of tropical tropopause temperatures on stratospheric water vapor, Journal of

1048 Geophysical Research: Atmospheres, 101, 3989–4006, https://doi.org/10.1029/95JD03422, 1996.

1049 43

1050

1051 Murcray, D. G., Murcray, F. J., Barker, D. B., and Mastenbrook, H. J.: Changes in stratospheric water vapor associated with the mount st. Helens

Tilmes, S., Tsigaridis, K., and Tummon, F.: Multi-model comparison of the volcanic sulfate

1077 deposition from the 1815 eruption of Mt. Tambora, Atmospheric Chemistry and Physics, 18, 2307–2328, https://doi.org/10.5194/acp-18-

1078 2307-2018, https://acp.copernicus.org/articles/18/2307/2018/, 2018.

1079 Marsland, S. J., Haak, H., Jungclaus, J. H., Latif, M., and Röske, F.: The Max-Planck-Institute global ocean/sea ice model with orthogonal

1080 curvilinear coordinates, Ocean Modelling, 5, 91–127, https://doi.org/10.1016/S1463-5003(02)00015-X, 2003.740

1081 Mauritsen, T., Bader, J., Becker, T., Behrens, J., Bittner, M., Brokopf, R., Brovkin, V., Claussen, M., Crueger, T., Esch, M., Fast, I., Fiedler, S.,

1082 Fläschner, D., Gayler, V., Giorgetta, M., Goll, D. S., Haak, H., Hagemann, S., Hedemann, C., Hohenegger, C., Ilyina, T., Jahns, T., Jiménéz-

1083 de-la Cuesta, D., Jungclaus, J., Kleinen, T., Kloster, S., Kracher, D., Kinne, S., Kleberg, D., Lasslop, G., Kornblueh, L., Marotzke, J., Matei,

1084 D., Meraner, K., Mikolajewicz, U., Modali, K., Möbis, B., Müller, W. A., Nabel, J. E. M. S., Nam, C. C. W., Notz, D., Nyawira, S.-S.,

1085 Paulsen, H., Peters, K., Pincus, R., Pohlmann, H., Pongratz, J., Popp, M., Raddatz, T. J., Rast, S., Redler, R., Reick, C. H., Rohrschnei-745

1086 der, T., Schemann, V., Schmidt, H., Schnur, R., Schulzweida, U., Six, K. D., Stein, L., Stemmler, I., Stevens, B., Storch, J.-S., Tian,

1087 F., Voigt, A., Vrese, P., Wieners, K.-H., Wilkenskjeld, S., Winkler, A., and Roeckner, E.: Developments in the MPI‐M Earth System

1088 Model version 1.2 (MPI‐ESM1.2) and Its Response to Increasing CO 2, Journal of Advances in Modeling Earth Systems, 11, 998–1038,

1089 https://doi.org/10.1029/2018MS001400, 2019.

1090 Mote, P. W., Rosenlof, K. H., McIntyre, M. E., Carr, E. S., Gille, J. C., Holton, J. R., Kinnersley, J. S., Pumphrey, H. C., Russell, J. M.,750

1091 and Waters, J. W.: An atmospheric tape recorder: The imprint of tropical tropopause temperatures on stratospheric water vapor, Journal of

1092 Geophysical Research: Atmospheres, 101, 3989–4006, https://doi.org/10.1029/95JD03422, 1996.

1093 Murcray, D. G., Murcray, F. J., Barker, D. B., and Mastenbrook, H. J.: Changes in stratospheric water vapor associated with the mount st. Helens

1052 eruption, Science (New York, N.Y.), 211, 823–824, https://doi.org/10.1126/science.211.4484.823, 1981.720
1053 Murphy, D. M. and Koop, T.: Review of the vapour pressures of ice and supercooled water for atmospheric applications, Quarterly Journal of
1054 the Royal Meteorological Society, 131, 1539–1565, https://doi.org/10.1256/qj.04.94, 2005.
1055 Myhre, Shindell, Bréon, Collins, Fuglestvedt, Huang, Koch, Lamarque, Lee, Mendoza, Nakajima, Robock, Stephens, and Takemura and Zhang:
1056 Anthropogenic and Natural Radiative Forcing: Climate Change 2013: The Physical Science Basis., Contribution of Working Group I to the
1057 Fiith Assessment Report of the Intergovernmental Panel of Climate Change, https://www.ipcc.ch/site/assets/uploads/2018/02/WG1AR5_725
1058 Chapter08_FINAL.pdf, 2013.
1059 Reick, C. H., Raddatz, T., Brovkin, V., and Gayler, V.: Representation of natural and anthropogenic land cover change in MPI-ESM, Journal of
1060 Advances in Modeling Earth Systems, 5, 459–482, https://doi.org/10.1002/jame.20022, 2013.
1061 Rienecker, M. M., Suarez, M. J., Gelaro, R., Todling, R., Bacmeister, J., Liu, E., Bosilovich, M. G., Schubert, S. D., Takacs, L., Kim, G.-K.,
1062 Bloom, S., Chen, J., Collins, D., Conaty, A., da Silva, A., Gu, W., Joiner, J., Koster, R. D., Lucchesi, R., Molod, A., Owens, T., Pawson,730
1063 S., Pegion, P., Redder, C. R., Reichle, R.,

1094 eruption, Science (New York, N.Y.), 211, 823–824, https://doi.org/10.1126/science.211.4484.823, 1981.
1095 Murphy, D. M. and Koop, T.: Review of the vapour pressures of ice and supercooled water for atmospheric applications, Quarterly Journal of755
1096 the Royal Meteorological Society, 131, 1539–1565, https://doi.org/10.1256/qj.04.94, 2005.
1097 Myhre, Shindell, Bréon, Collins, Fuglestvedt, Huang, Koch, Lamarque, Lee, Mendoza, Nakajima, Robock, Stephens, and Takemura and Zhang:
1098 Anthropogenic and Natural Radiative Forcing: Climate Change 2013: The Physical Science Basis., Contribution of Working Group I to the
1099 Fiith Assessment Report of the Intergovernmental Panel of Climate Change, https://www.ipcc.ch/site/assets/uploads/2018/02/WG1AR5_
1100 Chapter08_FINAL.pdf, 2013.760
1101 Möbis, B. and Stevens, B.: Factors controlling the position of the Intertropical Convergence Zone on an aquaplanet, Journal of Advances
1102 in Modeling Earth Systems, 4, https://doi.org/https://doi.org/10.1029/2012MS000199, https://agupubs.onlinelibrary.wiley.com/doi/abs/10.
1103 1029/2012MS000199, 2012.
1104 Oltmans, S., Voemel, H., Kley, D., Hofmann, K. P., and Rosenlof, K.: Increase in stratospheric water vapor from balloon-borne, frostpoint
1105 hygrometer measurements at Washington, DC and Boulder, Colorado, Geophysical Research Letters, 27, 2000.765
1106 Oman, L., Waugh, D. W., Pawson, S., Stolarski, R. S., and Nielsen, J. E.: Understanding the Changes of Stratospheric Water Vapor in Coupled
1107 Chemistry Climate Model Simulations, Journal of the Atmospheric Sciences, 65, 3278 – 3291, https://doi.org/10.1175/2008JAS2696.1,
1108 https://journals.ametsoc.org/view/journals/atsc/65/10/2008jas2696.1.xml, 2008.
1109 Reick, C. H., Raddatz, T., Brovkin, V., and Gayler, V.: Representation of natural and anthropogenic land cover change in MPI-ESM, Journal of
1110 Advances in Modeling Earth Systems, 5, 459–482, https://doi.org/10.1002/jame.20022, 2013.770
1111 44
1112
1113 Rienecker, M. M., Suarez, M. J., Gelaro, R., Todling, R., Bacmeister, J., Liu, E., Bosilovich, M. G., Schubert, S. D., Takacs, L., Kim, G.-K.,
1114 Bloom, S., Chen, J., Collins, D., Conaty, A., da Silva, A., Gu, W., Joiner, J., Koster, R. D., Lucchesi, R., Molod, A., Owens, T., Pawson,
1115 S., Pegion, P., Redder, C. R., Reichle, R., R

	Robertson, F. R., Ruddick, A. G., Sienkiewicz, M., and Woollen, J.: MERRA: NASA’s Modern-		obertson, F. R., Ruddick, A. G., Sienkiewicz, M., and Woollen, J.: MERRA: NASA’s Modern-
1064	Era Retrospective Analysis for Research and Applications, Journal of Climate, 24, 3624–3648, https://doi.org/10.1175/JCLI-D-11-00015.1,	1116	Era Retrospective Analysis for Research and Applications, Journal of Climate, 24, 3624–3648, https://doi.org/10.1175/JCLI-D-11-00015.1,
1065	2011.	1117	2011.775
1066	Robock, A.: Volcanic eruptions and climate, Reviews of Geophysics, 38, 191–219, https://doi.org/10.1029/1998RG000054, 2000.	1118	Robock, A.: Volcanic eruptions and climate, Reviews of Geophysics, 38, 191–219, https://doi.org/10.1029/1998RG000054, 2000.
1067	Robrecht, S., Vogel, B., Grooß, J.-U., Rosenlof, K., Thornberry, T., Rollins, A., Krämer, M., Christensen, L., and Müller, R.: Mechanism of735	1119	Robrecht, S., Vogel, B., Grooß, J.-U., Rosenlof, K., Thornberry, T., Rollins, A., Krämer, M., Christensen, L., and Müller, R.: Mechanism of
1068	ozone loss under enhanced water vapour conditions in the mid-latitude lower stratosphere in summer, Atmospheric Chemistry and Physics,	1120	ozone loss under enhanced water vapour conditions in the mid-latitude lower stratosphere in summer, Atmospheric Chemistry and Physics,
1069	19, 5805–5833, https://doi.org/10.5194/acp-19-5805-2019, 2019.	1121	19, 5805–5833, https://doi.org/10.5194/acp-19-5805-2019, 2019.
1070	Rosenlof, K. H.: Changes in water vapor and aerosols and their relation to stratospheric ozone, Comptes Rendus Geoscience, 350, 376–383,	1122	Rosenlof, K. H.: Changes in water vapor and aerosols and their relation to stratospheric ozone, Comptes Rendus Geoscience, 350, 376–383,780
1071	https://doi.org/10.1016/j.crte.2018.06.014, 2018.	1123	https://doi.org/10.1016/j.crte.2018.06.014, 2018.
1072	Rosenlof, K. H., Oltmans, S. J., Kley, D., Russell, J. M., Chiou, E.-W., Chu, W. P., Johnson, D. G., Kelly, K. K., Michelsen, H. A., Nedoluha,740	1124	Rosenlof, K. H., Oltmans, S. J., Kley, D., Russell, J. M., Chiou, E.-W., Chu, W. P., Johnson, D. G., Kelly, K. K., Michelsen, H. A., Nedoluha,
1073	G. E., Remsberg, E. E., Toon, G. C., and McCormick, M. P.: Stratospheric water vapor increases over the past half-century, Geophysical	1125	G. E., Remsberg, E. E., Toon, G. C., and McCormick, M. P.: Stratospheric water vapor increases over the past half-century, Geophysical
1074	Research Letters, 28, 1195–1198, https://doi.org/10.1029/2000GL012502, 2001.	1126	Research Letters, 28, 1195–1198, https://doi.org/10.1029/2000GL012502, 2001.
1075	Schmidt, H., Rast, S., Bunzel, F., Esch, M., Giorgetta, M., Kinne, S., Krismer, T., Stenchikov, G., Timmreck, C., Tomassini, L., and Walz, M.:	1127	Schmidt, H., Rast, S., Bunzel, F., Esch, M., Giorgetta, M., Kinne, S., Krismer, T., Stenchikov, G., Timmreck, C., Tomassini, L., and Walz, M.:785
1076	Response of the middle atmosphere to anthropogenic and natural forcings in the CMIP5 simulations with the Max Planck Institute Earth	1128	Response of the middle atmosphere to anthropogenic and natural forcings in the CMIP5 simulations with the Max Planck Institute Earth
1077	system model, Journal of Advances in Modeling Earth Systems, 5, 98–116, https://doi.org/10.1002/jame.20014, 2013.745	1129	system model, Journal of Advances in Modeling Earth Systems, 5, 98–116, https://doi.org/10.1002/jame.20014, 2013.
1078	Schneck, R., Reick, C. H., and Raddatz, T.: Land contribution to natural CO 2 variability on time scales of centuries, Journal of Advances in	1130	Schneck, R., Reick, C. H., and Raddatz, T.: Land contribution to natural CO 2 variability on time scales of centuries, Journal of Advances in
1079	Modeling Earth Systems, 5, 354–365, https://doi.org/10.1002/jame.20029, 2013.	1131	Modeling Earth Systems, 5, 354–365, https://doi.org/10.1002/jame.20029, 2013.
		1132	Schoeberl, M. R. and Dessler, A. E.: Dehydration of the stratosphere, Atmospheric Chemistry and Physics, 11, 8433–8446,790
		1133	https://doi.org/10.5194/acp-11-8433-2011, https://acp.copernicus.org/articles/11/8433/2011/, 2011.
1080	Schwartz, M. J., Read, W. G., Santee, M. L., Livesey, N. J., Froidevaux, L., Lambert, A., and Manney, G. L.: Convectively injected water vapor	1134	Schwartz, M. J., Read, W. G., Santee, M. L., Livesey, N. J., Froidevaux, L., Lambert, A., and Manney, G. L.: Convectively injected water vapor
1081	in the North American summer lowermost strat	1135	in the North American summer lowermost strato

osphere, Geophysical Research Letters, 40, 2316–2321, https://doi.org/10.1002/grl.50421,
1082 2013.750
1083 Sioris, C. E., Malo, A., McLinden, C. A., and D’Amours, R.: Direct injection of water vapor into the stratosphere by volcanic eruptions,
1084 Geophysical Research Letters, 43, 7694–7700, https://doi.org/10.1002/2016GL069918, 2016a.
1085 Sioris, C. E., Zou, J., McElroy, C. T., Boone, C. D., Sheese, P. E., and Bernath, P. F.: Water vapour variability in the high-latitude upper
1086 troposphere – Part 2: Impact of volcanic eruptions, Atmospheric Chemistry and Physics, 16, 2207–2219, https://doi.org/10.5194/acp-16-
1087 2207-2016, 2016b.755
1088 44
1089
1090 Solomon, S., Rosenlof, K. H., Portmann, R. W., Daniel, J. S., Davis, S. M., Sanford, T. J., and Plattner, G.-K.: Contributions
1091 of stratospheric water vapor to decadal changes in the rate of global warming, Science (New York, N.Y.), 327, 1219–1223,
1092 https://doi.org/10.1126/science.1182488, 2010.
1093 Stenchikov, G. L., Kirchner, I., Robock, A., Graf, H.-F., Antuña, J. C., Grainger, R. G., Lambert, A., and Thomason, L.: Radia-
1094 tive forcing from the 1991 Mount Pinatubo volcanic eruption, Journal of Geophysical Research: Atmospheres, 103, 13 837–13 857,760
1095 https://doi.org/10.1029/98JD00693, 1998.
1096 Stevens, B. and Bony, S.: Water in the atmosphere, Physics Today, 66, 29–34, https://doi.org/10.1063/PT.3.2009, 2013.
1097 Tao, M., Konopka, P., Ploeger, F., Yan, X., Wright, J. S., Diallo, M., Fueglistaler, S., and Riese, M.: Multitimescale variations in mod-
1098 eled stratospheric water vapor derived from three modern reanalysis products, Atmospheric Chemistry and Physics, 19, 6509–6534,
1099 https://doi.org/10.5194/acp-19-6509-2019, 2019.765
1100 Tian, W., Chipperfield, M. P., and Lü, D.: Impact of increasing stratospheric water vapor on ozone depletion and temperature change, Advances
1101 in Atmospheric Sciences, 26, 423–437, https://doi.org/10.1007/s00376-009-0423-3, 2009.

sphere, Geophysical Research Letters, 40, 2316–2321, https://doi.org/10.1002/grl.50421,
1136 2013.
1137 Sioris, C. E., Malo, A., McLinden, C. A., and D’Amours, R.: Direct injection of water vapor into the stratosphere by volcanic eruptions,795
1138 Geophysical Research Letters, 43, 7694–7700, https://doi.org/10.1002/2016GL069918, 2016a.
1139 Sioris, C. E., Zou, J., McElroy, C. T., Boone, C. D., Sheese, P. E., and Bernath, P. F.: Water vapour variability in the high-latitude upper
1140 troposphere – Part 2: Impact of volcanic eruptions, Atmospheric Chemistry and Physics, 16, 2207–2219, https://doi.org/10.5194/acp-16-
1141 2207-2016, 2016b.
1142 Soden, B. J., Wetherald, R. T., Stenchikov, G. L., and Robock, A.: Global cooling after the eruption of Mount Pinatubo: a test of climate800
1143 feedback by water vapor, Science (New York, N.Y.), 296, 727–730, https://doi.org/10.1126/science.296.5568.727, 2002.
1144 Solomon, S., Rosenlof, K. H., Portmann, R. W., Daniel, J. S., Davis, S. M., Sanford, T. J., and Plattner, G.-K.: Contributions
1145 of stratospheric water vapor to decadal changes in the rate of global warming, Science (New York, N.Y.), 327, 1219–1223,
1146 https://doi.org/10.1126/science.1182488, 2010.
1147 Stenchikov, G. L., Kirchner, I., Robock, A., Graf, H.-F., Antuña, J. C., Grainger, R. G., Lambert, A., and Thomason, L.: Radia-805
1148 tive forcing from the 1991 Mount Pinatubo volcanic eruption, Journal of Geophysical Research: Atmospheres, 103, 13 837–13 857,
1149 https://doi.org/10.1029/98JD00693, 1998.
1150 Stevens, B. and Bony, S.: Water in the atmosphere, Physics Today, 66, 29–34, https://doi.org/10.1063/PT.3.2009, 2013.
1151 45
1152
1153 Tao, M., Konopka, P., Ploeger, F., Yan, X., Wright, J. S., Diallo, M., Fueglistaler, S., and Riese, M.: Multitimescale variations in mod-
1154 eled stratospheric water vapor derived from three modern reanalysis products, Atmospheric Chemistry and Physics, 19, 6509–6534,810
1155 https://doi.org/10.5194/acp-19-6509-2019, 2019.
1156 Tian, W., Chipperfield, M. P., and Lü, D.: Impact of increasing stratospheric water vapor on ozone depletion and temperature change, Advances
1157 in Atmospheric Sciences, 26, 423–437, https://doi.org/10.1007/s00376-009-0423-3, 2009.

1102	Timmreck, C., Mann, G. W., Aquila, V., Hommel, R., Lee, L. A., Schmidt, A., Brühl, C., Carn, S., Chin, M., Dhomse, S. S., Diehl, T., English,	1158	Timmreck, C., Mann, G. W., Aquila, V., Hommel, R., Lee, L. A., Schmidt, A., Brühl, C., Carn, S., Chin, M., Dhomse, S. S., Diehl, T., English,
1103	J. M., Mills, M. J., Neely, R., Sheng, J., Toohey, M., and Weisenstein, D.: The Interactive Stratospheric Aerosol Model Intercomparison	1159	J. M., Mills, M. J., Neely, R., Sheng, J., Toohey, M., and Weisenstein, D.: The Interactive Stratospheric Aerosol Model Intercomparison 815
1104	Project (ISA-MIP): motivation and experimental design, Geoscientific Model Development, 11, 2581–2608, https://doi.org/10.5194/gmd-770	1160	Project (ISA-MIP): motivation and experimental design, Geoscientific Model Development, 11, 2581–2608, https://doi.org/10.5194/gmd-
1105	11-2581-2018, 2018.	1161	11-2581-2018, 2018.
1106	Toohey, M., Stevens, B., Schmidt, H., and Timmreck, C.: Easy Volcanic Aerosol (EVA v1.0): an idealized forcing generator for climate simu-	1162	Toohey, M., Stevens, B., Schmidt, H., and Timmreck, C.: Easy Volcanic Aerosol (EVA v1.0): an idealized forcing generator for climate simu-
1107	lations, Geoscientific Model Development, 9, 4049–4070, https://doi.org/10.5194/gmd-9-4049-2016, 2016.	1163	lations, Geoscientific Model Development, 9, 4049–4070, https://doi.org/10.5194/gmd-9-4049-2016, 2016.
1108	Wang, T., Zhang, Q., Kuilman, M., and Hannachi, A.: Response of stratospheric water vapour to CO_2 doubling in WACCM, Climate Dynamics,	1164	Wang, T., Zhang, Q., Kuilman, M., and Hannachi, A.: Response of stratospheric water vapour to CO_2 doubling in WACCM, Climate Dynamics,820
1109	54, 4877–4889, https://doi.org/10.1007/s00382-020-05260-z, 2020.775	1165	54, 4877–4889, https://doi.org/10.1007/s00382-020-05260-z, 2020.
1110	45	1166	46