Dear Marc,

thank you very much for guiding the editorial process of our manuscript.

Please find attached to this letter

- our replies to the referees' comments, as we have uploaded them in the discussion, (please note that referee #1 already answered our reply; this answer is not attached),
- our revised manuscript with highlighted modifications, and
- the new supplement.

In summary, the referees had three major concerns:

- 1. The methodology was not clear: To address this, we added a new section "Methodology" in which we describe in more detail our approach. A central point of this is that we analyse differences between simulation pairs and not different periods in time (e.g. before, during and after the eruptions). Moreover, we clarified throughout the revised text that we (by construction of our simulations) analyse only the effect of the "long-wave heating" induced by volcanic aerosol.
- 2. Comparison to observations: We added a new section with comparisons of our simulation results with observations and reanalysis data.
- 3. The role of the nudging approach: We address this issue by a new section (plus new supplement) based on additional sensitivity simulations, which show that our previous results are robust.

Yours,

Patrick (on behalf of all authors)

Here are our replies to the comments of referee #1:

• This paper focuses on a question of interest (at least to me) and the results look reasonable.

Reply: Yes, indeed the results are reasonable. And as we discuss, it is consistent with earlier results.

• That said, the experimental set-up does not seem robust and I cannot recommend publication until that is fixed. Let me expand on this main criticism. The authors have two runs of their model – one version has the volcanic cloud in it, while the other does not. The difference between these models is then interpreted as the impact of the eruption. Both models, however, are nudged to ECMWF-interim reanalysis, which has the impacts of the eruption in it. Because both versions of the model or being nudged towards a reanalysis that is perturbed by the volcano, it is not clear to me how to interpret the actual difference between these model runs.

Reply: It is a common misunderstanding that a nudged chemistry-climate model (CCM) is equivalent to a chemistry transport model (CTM). This is not the case, since – depending on the nudging procedure – the nudged CCM develops its own physical state on sub-synoptic scale. Therefore, our model setup is very well suited for the analysis. As we state on p. 34411 line 24 ff.:

"The Newtonian relaxation (nudging) of the prognostic variables, divergence, vorticity, temperature and the (logarithm of the) surface pressure is applied in **spectral space** with a corresponding relaxation time of 48, 6, 24 and 24 h, respectively. The sea surface temperatures (SSTs) and the sea ice concentrations (SICs) are prescribed every 12 h. A nudging of the "wave zero" (i.e., the global mean) temperature (T) is, however, excluded. This allows a temperature response of the model."

This means: The nudging is applied such that the large(r than synoptic) scale **patterns** correspond to those of ERA-Interim, but not the absolute temperature. Otherwise, the shown water vapour response on altered temperature would not be visible. Excluding the (level-dependent) global mean of the temperature (i.e., wave zero in spectral space) allows the absolute temperature to react (and propagate) as effect of the additional heating caused by the volcanic aerosol. In other words: the additional volcanic aerosols changes the absolute temperature, which then affects the hydrological cycle (cold-point temperature, convection, evaporation, etc.).

The interpretation of the results is as follows: We have two nudged simulations, both incorporate effects of the volcances as they are represented in the nudging data: effect on the SSTs (but not on the land surface), effect on the wind systems, effect on the temperature patterns (the local effect of temperature increase due to the aerosols is not visible in the nudged data!). These effects cancel out after subtracting the results of one simulations from the results of the other.

The concept is a sensitivity study, where we perturbed one simulation with the aerosol clouds of volcanic eruptions. The effect in the perturbed simulation is a strong local heating in the stratosphere and subsequent increase in stratospheric water vapour. This aspect (impact) is studied in detail by a comparison with the unperturbed simulation. The analysis of the difference between the 2 simulations gives us information on how the vertical temperature profile is modified by the heating and how the distribution of water vapour changes compared to a unperturbed simulation. Nudging keeps the model meteorology towards what is really observed, but does not prescribe the mean temperature.

The results do not give us new information about the impact of the eruptions in terms of cooling of SSTs, because this information is already in the prescribed SST.

We will add a clarifying paragraph to the description of our methodology in the revised manuscript.

• To their credit, the authors recognize this limitation, but this seems much more severe than they seem to recognize. For example, around line 25 of p. 34412, they state that the hydrological cycle is "free running". That is very misleading. Stratospheric water vapor is determined by the cold point in the TTL – and temperature is nudged towards the reanalysis. So the key parameter they are investigating, stratospheric water vapor, is indeed impacted by the nudging. In the end, I do not believe this paper should be published until the authors can better characterize the difference between the runs.

Reply: The referee is right in stating that the cold point temperature determines (to the largest extend) the amount of water vapour entering the stratosphere. However, in our case, the absolute temperature (including that at the cold point) is NOT influenced by Newtonian relaxation, because we exclude "wave-0" in spectral space (see above). Thus, the cold point temperature is affected by the volcanic aerosol and this in turn affects the water vapour. Moreover, non of the hydrological cycle variables is nudged. The setup is such that the large scale dynamics between both simulations is similar, but the hydrological cycle reacts on the altered temperature induced by the volcanic aerosol in the stratosphere. Furthermore, the free running (of freely responding) hydrological cycle is mass conserving.

• Other comments: 1) I found the paper difficult to read. The grammar was fine – I'm referring more to the overall style of writing and sentence structure. I don't have any specific suggestions other than that the authors should spend some time crafting the text.

Reply: This is unfortunately a very vague statement. Although, none of us is a native English speaker, neither our institutional internal referee, nor the editor doing the quick-access review complained about the language. Maybe it is simply a matter of taste. Nevertheless, we appreciate specific suggestions for improvement.

• 2) One recent paper on this subject not cited is Dessler et al., (2014), Variations in stratospheric water vapor over the past three decades, J. Geophys. Res., 119, 12,588 – 12,598, doi: 10.1002/2014JD021712.

Answer: Thank you very much for pointing this out. We will check the reference and add the citation, where appropriate.

3) Section 3.3 argues that tropospheric water vapor increased after the eruption. That neither makes any physical sense nor does it agree with previous research. Given that the eruption cools the troposphere, you would expect tropospheric humidity to decline, which has been seen in observations, e.g., by Soden et al. (2002), Global cooling after the eruption of Mount Pinatubo: A test of climate feedback by water vapor, Science, 296, 727-730.

Answer: We agree that we have to explain some points more clearly:

Volcanic aerosol has two effects on temperature: Warming of the stratosphere due to long wave radiation effects and cooling of the surface and thus the troposphere due to shadowing of the surface in the short-wave radiation. The latter leads to a global cooling. Both effects are in part due to the nudging and the prescribed SSTs already incorporated in both simulations. Both simulations show a cooling of the lower troposphere and subsequent lower water vapour concentrations after the eruption of Mt. Pinatubo. This is in accordance with observations, e.g. Soden et al. (2002) as stated by the referee.

However, the warming of the stratosphere partly offsets the water vapour decrease by stabilising the vertical temperature profile with subsequent lower convective activity and precipitation. Our experiment design isolates the effect of the aerosol heating in the stratosphere on the atmosphere.

We will clarify this in our discussion of the revised manuscript.

• 4) In Section 3.2, the authors turn to the "stratospheric fountain" hypothesis to explain the volcanic impacts on the monsoon regions. That is a weird argument because (to the best of my knowledge) nobody views the stratospheric fountain as a legitimate way to think about troposphere-stratosphere exchange.

Answer: Yes, this is indeed wrong as it stands. It was simply used to refer to the role of the specific region. We will remove this statement, since it is not needed for our analysis. We thank Theodore Shepherd for the very stimulating comments. Here are our replies:

• 1. There is no connection made with observations. If you are discussing real events, then it is incumbent on you to compare with observations. In your experimental set-up, you are comparing a simulation of the real world with the world that would have been (some kind of counter-factual) without aerosol loading, but assuming that everything else would have been the same. Your attribution of the effect of the aerosol loading would be more convincing if you could show that your real-world simulation agreed with observations.

Reply: The model system has been evaluated (including comparison to observations) several times (Jöckel et al., 2006, 2010, 2016). The tropical tape recorder and the different pathways of water vapour into the stratosphere in our model system have been analysed by Eichinger et al. $(2015a^1, b^2)$. A detailed analysis of stratospheric water vapour anomalies is on the way (Brinkop et al., 2015^3). Therefore, we omitted a detailed inter-comparison here. Nevertheless, we see the necessity to compare the volcanically perturbed periods in more detail here. In the revised manuscript, we provide (in a new subsection) a comparison of cold-point temperature anomaly, water vapour anomaly, and upwelling anomaly compared to ERA-Interim and water vapour anomaly compared to the recently published data of Hegglin et al. $(2014)^4$.

• I don't believe there is any observational evidence for increased stratospheric water vapour following Pinatubo (see e.g. the latest historical time series from Hegglin et al. 2014 Nature Geosci.), and – consistently with this – according to Randel et al. (2000 JGR) the tropical cold point tropopause was not significantly warmed following Pinatubo, though it was after El Chichon. This doesn't mean that the mechanism was inoperative following Pinatubo, only that it was masked by other factors. You can show that by demonstrating that your real-world simulation matches the available observations, and that your counter-factual simulation would

¹Eichinger, R., Jöckel, P., Brinkop, S., Werner, M., & Lossow, S.: Simulation of the isotopic composition of stratospheric water vapour – Part 1: Description and evaluation of the EMAC model, Atmospheric Chemistry and Physics, 15, 5537–5555, doi: 10.5194/acp-15-5537-2015, URL http://www.atmos-chem-phys.net/15/5537/2015/ (2015a).

²Eichinger, R., Jöckel, P., & Lossow, S.: Simulation of the isotopic composition of stratospheric water vapour – Part 2: Investigation of HDO / H2O variations, Atmospheric Chemistry and Physics, 15, 7003–7015, doi: 10.5194/acp-15-7003-2015, URL http://www.atmos-chemphys.net/15/7003/2015/ (2015b).

³Brinkop, S., Dameris, M., Jöckel, P., Garny, H., Lossow, S., & Stiller, G.: The millennium water vapour drop in chemistry-climate model simulations, Atmospheric Chemistry and Physics Discussions, 15, 24 909–24 953, doi: 10.5194/acpd-15-24909-2015, URL http://www.atmos-chem-phys-discuss.net/15/24909/2015/ (2015)

⁴Hegglin, M. I., D. A. Plummer, T. G. Shepherd, J. F. Scinocca, J. Anderson, L. Froidevaux, B. Funke, D. Hurst, A. Rozanov, J. Urban, T. von Clarmann, K. A. Walker, R. Wang, S. Tegtmeier, and K. Weigel, Vertical structure of stratospheric water vapour trends derived from merged satellite data, Nature Geoscience, 7, 768–776, (2014), doi:10.1038/ngeo2236

have had lower CPTs and lower water vapour than observed. I appreciate that the water vapour observations are uncertain, and unavailable in the tropical lower stratosphere right after the aerosol injection, but there are observations at other latitudes and altitudes. And the temperature is also an important validation field since it is not constrained by your nudging and is key to the water vapour response.

Reply: We completely agree. In the revised manuscript we show and discuss the cold-point temperature anomalies of both simulations in comparison to ERA-Interim, as well as the water vapour anomalies at 80 hPa. The results confirm exactly your suggestion: The NOVOL simulation shows a lower cold-point temperature anomaly and consistently lower water vapour. The VOL simulation, in contrast, overestimates both, the cold-point temperature anomaly and correspondingly the water vapour anomaly.

From this, we conclude that the masking factor you mention (see also Fueglistaler et al., 2012^5) is indeed captured correctly by your nudged model: We see an attenuated upwelling in the period right after the Mt. Pinatubo eruption, which cools the tropical tropopause, and which is (over)compensated by an (in our case) too strong volcanic heating.

• 2. I do have some concerns about the nudging. I appreciate that there is no perfect approach here, but by nudging the divergence field, it must be that the vertical motion is strongly constrained. Yet after an aerosol injection, basic dynamics tells us that the additional heating will lead both to warming of the atmosphere and to vertical motion, during the transient phase of the response. (This is the classic Eliassen 1950 response.) By suppressing any changes in the vertical motion, the heating must go entirely into warming and the warming will thus be too strong. This would not be a big issue if the transient phase was short, but in the tropics it could be the better part of a year because of the flywheel effect (Scott & Haynes 1998 QJRMS). By comparing your model temperature with the observed temperature, you could determine how much of an error you are thereby incurring.

Reply: The nudging by Newtonian relaxation as we applied it, does neither "strongly constrain", nor "suppress any changes" in the vertical motion. The model retains all its degrees of freedom, although possibly damped. Indeed, the vertical velocity between NOVOL and VOL does change, as we expect it. This is shown by an additional figure (in the Supplement of the revised manuscript) of the differences in upwelling (w^{*} calculated according to the TEM (Transformed Eulerian Mean) method as described by Holton, 2004^6).

To show that the nudging indeed mostly increases the signal-to-noise ratio,

⁵Fueglistaler, S. (2012), Stepwise changes in stratospheric water vapor?, J. Geophys. Res., 117, D13302, doi:10.1029/2012JD017582.

⁶Holton, J. R.: An Introduction to Dynamic Meteorology, International Geophysics Series, 4th edn., Academic Press, San Diego, New York, USA, 2004.

if two different simulations are compared point-by-point, we performed an additional sensitivity study of two simulations, in which we only nudge the (logarithm of the) surface pressure. The resulting signals of the differences are very similar, though superimposed by "meteorological noise". An additional section on sensitivity studies in the revised manuscript discusses this.

• A second issue with the nudging is that since you leave only the globalmean temperature free to respond, there will presumably be some artefact in the extratropics because the radiative imbalance from the aerosol loading is only in the tropics. In other words, the entire global mean has to adjust to the level of the tropical adjustment. Is it obvious that this would not affect your results concerning the influence of the monsoon, for example?

Reply: We think that this is a wrong view on what the nudging is doing. As we explain above, the model keeps all its degrees of freedom, implying the possibility that local temperature perturbations can propagate and be converted to vertical motion. The relaxation towards the local temperature anomaly does not imply that the model response can only be by global adjustment. Only the signal propagation might be damped due to the continuous relaxation towards a defined (local) state.

As above, we refer to the additional sensitivity study presented in the revised manuscript. We find a similar monsoon signal also in a quasi-free running simulation. We thank referee #2 for the comments. Here are our replies:

• The paper investigates the impact of past volcanic eruptions on stratospheric water vapour. The subject of the paper is suitable for ACP and understanding long-term variability in climatic gases, in particular water vapour, is of importance for our understanding of the Earth system. However, I do have some concerns regarding the paper, which I hope can be rectified.

Reply: Thank you for the summary. We hope that our replies will dispel your concerns.

• 1) Experimental setup: I understand that all nudging methods have to compromise. However, it is not clear to me why the authors have chosen to exclude the optical but not the chemical impact of volcanic aerosols on the system. Presumably the heating change will be the leading order effect, but changes in composition might counteract or support the heating rate changes (e.g. ozone anomalies will impact heating rates as well). Please explain in more detail what you can or cannot conclude from the experimental setup.

Reply: The reason was to exclude the secondary effects of the volcanic aerosol (via heterogeneous chemistry and corresponding radiative forcing, e.g. via ozone), but rather to isolate the primary effect of the direct aerosol induced heating. We add this information to the revised manuscript (Section 2.2).

Moreover, the experimental setup is explained in more detail in the revised Section 2.3 (Methodology) and the discussion on conclusions and limitations is extended.

• 2) Observational backing: The paper lives in the "model world" only. I really miss some direct links to observations (apart from sometimes confusing citations of observational evidence in the text, see example below). It would be good to know if the model with the aerosol effect fully included captures the observed variability well. At the moment the paper is a nice technical sensitivity study without a solid link to the real world. This in itself is not a problem, but the authors sketch out a mechanism, which might not have a link to the "real world".

Reply: Here you agree with a similar comment by Ted Shepherd. And as we point out in our reply to him, we also agree. The revised manuscript contains a comparison with observations.

• In summary: I feel the paper requires a significant re-work with a better link (illustrated with one or two figures) to observational evidence and a contextual link to the data used to "drive" the model: How freely can the temperature structure evolve compared to the reanalysis? What is the water vapour signal in the reanalysis? Reply: As stated above, we show an inter-comparison to observational data in the revised manuscript. A comparison to ERA-Interim data is also included.

• Unfortunately I found the paper hard to read. Here are some examples of what could be improved: P34408, 118: Everything is know? Please rephrase ...

Reply: We reformulated the sentence.

- P34408, l20: Please avoid negations ("not only") ... Reply: Removed.
- P34412, l25: What does "in some parts" mean?

Reply: This and the subsequent section have been reformulated in accordance with other comments referring to the nudging procedure.

- P34417, ll19: What has this to do with the stratospheric fountain? Reply: Nothing. We removed this sentence as it is misleading.
- P34420, l23: Not sure what the message of this paragraph is. Different peaks in different plots? Please simplify (short sentences, direct comparison, is it your model or the reanalysis you are using that is causing the differences?)

Reply: This paragraph is part of a comparison with other model simulations and observations and subsequent discussion. The message in line 23 is that we found two maxima in our water vapour signal, similar as Considine et al. $(2001)^1$. Nevertheless, we agree that this text passage should be simplified and we did so in the revised manuscript.

• P34422, 16: Why not use Figure 10 to summarise in a concise way what you think is going on?

Reply: Indeed, the purpose of Figure 10 was exactly this. In the revised manuscript we refer to it, where appropriate and expand the summary accordingly.

¹Considine, D. B., Rosenfield, J. E., and Fleming, E. L.: An interactive model study of the influence of the Mount Pinatubo aerosol on stratospheric methane and water trends, J. Geophys. Res.-Atmos., 106, 27711-27727, doi:10.1029/2001JD000331, 2001. 34410, 34419, 34420, 34421

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Impact of major volcanic eruptions on stratospheric water vapour

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Abstract

Volcanic eruptions can have significant impact on the earth's weather and climate system. Besides the subsequent tropospheric changes also the stratosphere is influenced by large eruptions. Here changes in stratospheric water vapour after the two major volcanic eruptions of El Chichón in Mexico in 1982 and Mount Pinatubo on the Philippines in 1991 are investigated with chemistry-climate model simulations. This study is based on two simulations with specified dynamics of the EMAC model, performed within the Earth System Chemistry integrated Modelling (ESCiMo) project, of which only one includes the long-wave volcanic forcing through prescribed aerosol optical properties. The results show a significant increase in stratospheric water vapour after induced by the eruptions, resulting from increased heating rates and the subsequent changes in stratospheric and tropopause temperatures in the tropics. The tropical vertical advection and the South Asian summer monsoon are identified as important sources for the additional water vapour in the stratosphere. Additionally, volcanic influences on the tropospheric water vapour and ENSO are evident-, if the long-wave forcing is strong enough. Our results are corroborated by additional sensitivity simulations of the Mount Pinatubo period with reduced nudging and reduced volcanic aerosol extinction.

1 Introduction

As the most important greenhouse gas in the troposphere water vapour plays a key role in the climate feedback loop. This feedback is known to amplify the already known greenhouse effect through an increase in amplifies the greenhouse effect of CO₂ by about 60 % (Forster et al., 2007). Relevant for the climate is not only the Besides the water vapour distribution in the troposphere, but also the amount of water vapour in the stratosphere is relevant for climate (Solomon et al., 2010). The abundance of stratospheric water vapour (SWV) is mainly controlled by the temperatures at the tropical tropopause (Mote et al., 1996). This leads lt is further subject to a high inter-annual and multi-decadal variability, mostly dom-

inated by the Quasi-Biennial Oscillation (QBO) and El Niño-Southern Oscillation (ENSO), which both affect tropical tropopause temperatures (<u>Dessler et al., 2014b</u>). Besides the two well known phenomena also changes in the chemical balance (like higher methane oxidation rates through an increase of stratospheric chlorine, hydroxyl and ozone), changes in circulation patterns, as well as volcanic influences have to be taken into account (Forster et al., 2007; Stenke and Grewe, 2005).

The effect of volcanoes on the climate system was investigated with models under different aspects. Stenchikov et al. (1998) addressed the radiative impact of Mount Pinatubo aerosols. They used observational data of aerosol extinctions and effective radii to calculate the radiative forcing with the ECHAM4 general circulation model and found that the stratospheric heating was mainly due to near-infrared solar forcing. Arfeuille et al. (2013) investigated the uncertainties of the Mount Pinatubo aerosol extinction and the effect on temperature. Thomas et al. (2009a, b) were the first to use a comprehensive and complex simulation of the ECHAM5 general circulation model to evaluate the effects of the Mount Pinatubo eruption under different boundary conditions taking into account observed SSTs sea surface temperatures (SSTs) and volcanically induced ozone anomalies, as well as QBO and ENSO. They were able to realistically reproduce the observed lower stratospheric temperature response with the combined effects of the prescribed SSTs, ozone anomalies and the state of the QBO. Graf et al. (1993) used the ECHAM2 general circulation model to investigate the stratospheric aerosol effects of El Chichón and Mount Pinatubo on the Northern Hemisphere's (NH) climate. They identified short-term dynamical responses and a warming of the lower troposphere in the first winter season after the eruptions, which they found to be in good agreement with observations. With the MAECHAM4/CHEM model Timmreck and Graf (2006) performed a model study to simulate the radiative effects and the dispersal of the aerosols after the eruption of a super volcano in the mid-latitudes of the NH. They discovered that the initial dispersal of the aerosol cloud mainly depends on the season of the eruption. In summer, the cloud is transported west- and northward, in winter more south- and eastward. Contrary to the heating found in other studies after the Pinatubo eruption, Timmreck and Graf (2006) determined

a strong cooling of $\sim -1.6~{\rm K~d^{-1}}$ in the upper stratosphere shortly after the eruption of the super volcano. The dispersal of the Mount Pinatubo aerosol cloud was investigated with the ECHAM4 model by Timmreck et al. (1999a) using a Newtonian relaxation technique. Their simulated aerosol distribution was mostly in good agreement with observational data. Timmreck et al. (1999b) also used the ECHAM4 model to perform an interactive simulation with prognostic aerosols for the Mount Pinatubo eruption. They then compared the results for the years 1991 and 1992 with a non-interactive study, satellite observations and in situ measurements. They were able to reproduce dynamical effects in accordance with observations and found maximum heating rates of $\sim 0.3~{\rm K~d^{-1}}$ in October 1991.

Regarding the changes in stratospheric water vapour in the aftermath of major volcanic eruptions, Oltmans et al. (2000) Angell (1997) found in the limited observational data available that there is a SWV increase after volcanic eruptions, which disappeared after approximately 2 years. Joshi and Jones (2009) proposed that within the following 2 years after the eruption, a heating of the tropopause layer imposed by volcanic aerosol clouds allowed more water vapour to pass into the stratosphere. They supposed that the total SWV perturbation (globally averaged) after the Mount Pinatubo eruption would account for at least 15 % more SWV. The increase in tropopause temperatures supported the investigations of Considine et al. (2001). They also made volcanic eruptions partly responsible for the SWV-trend in the 1990s. In their model simulation they determined an increase in tropopause temperatures of about 0.5 K, an increase in stratospheric temperatures of around 2–3 K at a height between 20 hPa and 50 hPa, and a maximum increase in water vapour of only $\sim 6\%$ in the upper stratosphere after Mount Pinatubo.

The present study investigates the SWV perturbations of the long-wave radiative effect of the major volcanic eruptions of El Chichón (Mexico, 1982) and Mount Pinatubo (Philippines, 1991) using the state of the art general circulation chemistry-climate model EMAC. The main objective is to analyse the perturbation of SWV and the transport paths of water vapour into the stratosphere. In Sect. 2 Section 2 provides a brief description of the EMAC modeland, the setup of the used simulationsare given, and the methodology behind our analyses. In Sect. 3 our results of 3 we show a comparison of our model results with observations an reanalysis data. The resulting volcanic SWV perturbations are presented.Section 4 analysed in Sect. 4, the corresponding sensitivity simulations in Sect. 5. Section 6 discusses the results and Sect. 5. T summaries the findings and provides an outlook on further studies.

2 Model simulations

2.1 Model description

The ECHAM/MESSy Atmospheric Chemistry (EMAC) model is a numerical chemistry and climate simulation system that includes sub-models describing tropospheric and middle atmosphere processes and their interaction with oceans, land and human influences (Jöckel et al., 2010). It uses the second version of the Modular Earth Submodel System (MESSy2) to link multi-institutional computer codes. The core atmospheric model is the 5th generation European Centre Hamburg general circulation model (Roeckner et al., 2006). For the present study we applied EMAC (ECHAM5 version 5.3.02, MESSy version 2.51) in the T42L90MA-resolution, i.e., with a spherical truncation of T42 (corresponding to a quadratic Gaussian grid of approx. 2.8° by 2.8° in latitude and longitude) with 90 vertical hybrid pressure levels up to 0.01 hPa.

2.2 ESCiMo consortial simulations

Multiple simulations with different boundary conditions were performed within the "Earth System Chemistry integrated Modelling (ESCiMo)" initiative (?) (Jöckel et al., 2016). These model simulations were defined to improve the understanding of processes in the atmosphere and also to help answer questions related to climate change, ozone depletion and air quality. Besides the scientific relevance, the obtained results are supposed to have also political and social impact. This is especially important for the contribution to the WMO/UNEP ozone and IPCC climate assessments.

In our study we focus on 2 simulations with specified dynamics. They were "nudged" with a Newtonian relaxation technique towards 6 hourly ECMWF reanalysis data (ERA-Interim, Dee et al., 2011), which are available from the year 1979 to 2012.

The Newtonian relaxation (nudging) of the prognostic variables, divergence, vorticity, temperature and the (logarithm of the) surface pressure is applied in spectral space with a corresponding relaxation time of 48, 6, 24 and 24 h, respectively. The sea surface temperatures (SSTs) SSTs and the sea ice concentrations (SICs) are prescribed every 12 h. A nudging of the "wave zero" (i.e., the global mean) temperature (\overline{T}) is, however, excluded. This allows a temperature response of the model. This means: the nudging is applied such that the synoptic scale **patterns** correspond to those of ERA-Interim, but not the absolute temperature. Otherwise, no water vapour response on altered temperature would be visible.

The nudging is applied in the troposphere from above the boundary layer up to 10 hPa. The nudging coefficients increase from zero in a vertical transition region with a maximum between 84 and 10 hPa and decrease again. The boundary layer (the lowest three model levels) and the middle atmosphere (less than 10 hPa) are left unaffected.

The simulations RC1SD-base-01 (in the following referred as VOL) and RC1SD-base-10 (from now on referred as NOVOL) range from 1979 to 2012 and 2013, respectively. They are nearly identical simulations, which differ mainly with respect to volcanic perturbations in the system: In VOL the volcanic perturbation is considered, whereas NOVOL has no volcanic perturbation, with the exception that the aerosol surfaces (which are relevant for the heterogeneous chemistry) are also prescribed in the NOVOL simulation. The reason was to exclude the secondary effects of the volcanic aerosol (via heterogeneous chemistry and corresponding radiative forcing, e.g. via ozone) from the analysis, but rather to isolate the primary effect of the direct aerosol induced heating. In VOL the dynamically relevant volcanic sulfate aerosol effect is achieved by prescribing zonally and monthly averaged values of the aerosol radiative properties: extinction coefficients for the 16 spectral bands (short wave: 4 bands, long wave: 12 bands) of EMAC, single scattering albedo and asymmetry factor (B. Luo, personal communication, 2013; ftp://iacftp.ethz.ch/pub_read/luo/ccmi/). These

are used by the radiation scheme to calculate the corresponding heating rates (Fig. 1). Atmospheric chemistry in troposphere and stratosphere is calculated interactively. Volcanic temperature changes might therefore influence methane oxidation, a water vapour source, in the stratosphere.

The heating rates lead to increased local temperatures (Fig. 1), which cause perturbations in the atmospheric chemistry and dynamics (i.e., upward motion in the stratosphere). The dynamical effect in turn has an impact on the vertical distribution of chemical constituents and water vapour.

2.3 Methodology

The nudging data in some parts already include the dynamical effects of the volcances and therefore both simulations are influenced. The results in this study will mainly represent the difference are presented as differences between the VOL and NOVOL simulation, if not stated otherwise. This configuration allows a The nudging allows this one-by-one comparison but still (differences), because it increases the signal-to-noise ratio, due to the similar synoptic conditions in the reference (NOVOL) and perturbed (VOL) simulations. On the other hand, it limits the analysis of the dynamical effects. Note, however, the hydrological cycle itself is "free running" and not nudged in the simulationsNevertheless, since the absolute temperature (including that at the cold point) is not influenced by Newtonian relaxation, because we exclude "wave-0" in spectral space (see above), the cold point temperature is affected by the volcanic aerosol, and this in turn affects the water vapour. As none of the hydrological variables is nudged, the hydrological cycle (convection, evaporation, etc.) reacts on the altered temperature (e.g., cold point) induced by the volcanic aerosol in the stratosphere.

Because we nudge both (VOL and NOVOL) simulations with ERA-Interim data and prescribe SSTs, the SST part of the volcanic signal is included in both simulations, and cancels out in the differences calculations. Therefore, our approach isolates the radiative effect of volcanic aerosol via temperature changes on water vapour.

Last but not least, our approach (of looking at the differences between two nudged simulations) is considerably different to a comparison of volcanically perturbed periods with periods prior or past the event. This can, at first sight, lead to counter-intuitive results if the differences between two simulations are wrongly interpreted as differences between different periods.

We performed additional sensitivity studies for the Mount Pinatubo eruption, to underpin our findings from the two original simulations with full chemistry (further denoted as FC). To reduce the computational costs, we first repeated the FC simulation pair, however, without interactive chemistry. To still include in good approximation the chemical effects, we prescribed the monthly average chemical constituents, which are either relevant for the radiative forcing (O₃, N₂O, CO₂, CF₂Cl₂, CFCl₃), or the methane oxidation (OH,O¹D,Cl) from the VOL and NOVOL simulations, respectively. In total, we performed 3 such pairs of simulations, one of each pair with and one without representation of the Mount Pinatubo eruption (prescribed aerosol optical properties), i.e., corresponding to the original VOL and NOVOL simulations, respectively:

- In the first pair, further denoted RE (for "remake" of FC) the full chemistry is replaced by the prescribed chemistry without further modifications. This pair serves as a new control simulation for the subsequent pairs.
- For a second pair, further denoted as LA (for "low aerosol") we repeated the RE simulation with Mount Pinatubo eruption, however, with the aerosol extinction coefficients scaled by a factor of 0.5.
- In a third pair, further denoted QF (for "guasi free-running"), only the (logarithm of the) surface pressure was nudged, in contrast to FC, in which also divergence, vorticity and temperature (without wave-0) have been nudged.

As for the FC pair (VOL and NOVOL), we analyse the differences ("with volcano" minus "without volcano") of the additional pairs (RE, LA, QF).

The heating rates lead to increased local temperatures (

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3 Comparison with observations and reanalysis data

The EMAC model system has been evaluated (including comparison to observations) several times (Jöckel et al., 2006, 2010; Jöckel et al., 2016). The tropical tape recorder and the different pathways of water vapour into the stratosphere in our model system have been analysed by Eichinger et al. (2015a). A detailed analysis of stratospheric water vapour anomalies is provided by Brinkop et al. (2015).

In this section we provide a comparison of our simulated cold point temperature anomaly and the corresponding water vapour anomaly with reanalysis data from ERA-Interim (Dee et al., 2011), and the merged water vapour data set of Hegglin et al. (2014).

To assess the change in water vapour after major volcanic eruptions, it is necessary to simulate a realistic anomaly of the cold point temperature including the heating due to the volcanic aerosol in the stratosphere. The cold point anomaly is influenced by the QBO phase, the strength of the Brewer-Dobson circulation (strength of upwelling), ENSO, the temperature of the troposphere (Dessler et al., 2014a; Fueglistaler and Haynes, 2005), and through the volcanic eruption (Randel et al., 1995). Our nudged simulations include these processes, and the nudging ensures the right "timing". The period shortly after the El Chichón eruption is characterised by a westerly phase of the QBO and a subsequent warming, and the period after the Mount Pinatubo is influenced by an east phase of the QBO accompanied with cooling. Both eruptions are followed by an El Niño at the subsequent turn of the year, which contributes to changes of the cold point temperature. Fueglistaler (2012) points out the complicated situation influencing the stratospheric temperature evolution after the eruption of Mount Pinatubo. He describes two competing effects, a strongly intensified residual circulation (cooling of the tropopause in the tropics) and the radiative heating through the volcanic aerosol in the stratosphere: "Hence, in the situation where the aerosol heating outruns the dynamical effect, one would indeed expect a jump in water entering the stratosphere in mid-1991." Observations, however, do not show a clear picture after the eruptions (Fueglistaler, 2012).

ERA-Interim cold point temperature anomalies show an inter-annual variability, reaching a minimum of about -1 K and maxima of about +0.75 K (Fig. 1), which cause perturbations in the atmospheric chemistry and dynamics (i.e., upward motion in the stratosphere) 2). The dynamical effect in turn has an impact on the vertical distribution of chemical constituents and water vapour The period after the two large eruptions both show a similar increase in temperature of 0.75 K, although the eruption of Mount Pinatubo was stronger. Our VOL and NOVOL simulations show a similar cold point temperature anomaly, except for the periods after the eruptions. These discrepancies are expected for the NOVOL simulation, because the stratospheric heating effect of the eruption is not included in the simulation. In consistence with Fueglistaler (2012), we find a strong negative cold point anomaly (Fig. 2) with a corresponding negative water vapour (Fig. 3) and upwelling anomaly (Fig. 4, negative upwelling anomaly means stronger upwelling). The VOL simulation overestimates the cold point temperature and, compared to ERA-Interim, underestimates the water vapour anomaly in the period after the Mount Pinatubo eruption. The sensitivity simulation including the volcanic aerosol with (by a factor of 0.5) reduced extinction (see Sect. 2.3), hereafter denoted as VOL-LA, yields similar results for the cold point anomaly in the period after Mount Pinatubo (1992) compared to ERA-Interim Fig. 2. However, none of the simulations captures the ERA-Interim cold point temperature anomaly at the period close to the eruption of Mount Pinatubo.

The water vapour anomaly at 80 hPa in ERA-interim does not always follow the cold point temperature anomaly (Fig. 3). We find two large maxima one half and about two years after the eruptions of El Chichón and Mount Pinatubo, respectively in both ERA-Interim data and our VOL simulation. ERA-Interim exhibits the largest maxima. These maxima are in agreement with results of Dessler et al. (2014a, see their Figure 4), who show by their regression analysis that both volcanic eruptions (residual of the regression analysis) explains about 0.3 ppmv of the water vapour anomaly.

The merged data set (MSD in Fig. 3) of Hegglin et al. (2014) is unfortunately limited to the period around the Mount Pinatubo eruption. The MSD shows larger amplitudes before the eruption and is completely different to ERA-Interim. In particular 2 years after the eruption,

the MSD data show a large minimum, whereas ERA-Interim shows a large maximum. Unfortunately, directly after the eruption no MSD data is available. The large minimum of water vapour anomaly occurring in ERA-Interim before the eruption of Mount Pinatubo is captured by none of our simulations. It coincides with a maximum of the ERA-Interim cold point temperature anomaly. This is in contradiction to the expected correlation between stratospheric water vapour and cold point temperature.

The development of upwelling (see anomaly in Fig. 4) is similar in VOL, NOVOL and VOL-LA, except after the eruption of Mount Pinatubo. We conclude that the heating through the volcanic aerosol decreases the upwelling in the stratosphere (cf. positive anomaly after 1992). Whereas from 1984 to 1996 ERA-Interim upwelling shows the same anomaly as our simulations, it shows large discrepancies in the period after El Chichón, of which the origin is unknown. Part of the negative upwelling anomaly 19982/83 and 1991/92 is caused by the subsequent El Niño events. This might delay the increase of cold point temperature and water vapour by about one year, although the maximum of heating due to the volcanic aerosol in the stratosphere occurs in the year of the eruption (Fig. 1).

4 Perturbation of stratospheric water vapour

In this section we present the findings of our study of the perturbations of stratospheric water vapour <u>due to the long-wave heating of the volcanic aerosol</u> for the time periods following the eruptions of El Chichón in 1982 and Mount Pinatubo in 1991. We first concentrate on tropical stratospheric aspects and then also take possible influences from the South Asian monsoon region into account as well as impacts on tropospheric water vapour and ENSO.

4.1 Tropics

The maximum heating rates for the tropical mean (5° S– 5° N) for both volcanoes are found at a height of about 20 hPa with an amplitude of around 0.45 and 0.6 K d⁻¹, respectively (Fig. 1). For Mount Pinatubo there is a second local maximum of heating rates located

around 40 hPa occurring approximately 3 months after the eruption. These maxima coincide well with the maxima of the aerosol extinction.

The overall larger values for the Mount Pinatubo eruption are explainable through the higher mass of ejected SO₂ (with a factor of 2–3) and therefore larger aerosol extinction, which lead to increased heating rates and stronger temperature changes (Figs. 1 and 5). The volcanically induced temperature increase has its maximum values in the middle stratosphere and reaches within the same month ~ 1 K for El Chichón and ~ 3 K for Mount Pinatubo, respectively, at 20 and 30 hPa. For the 50 hPa level the amplitude of the temperature change reaches about 1.5 K for El Chichón and 4 K for Mount Pinatubo, respectively, approximately 6 months after the eruptions. For both volcanoes the induced temperature increases decline back to unperturbed values within about 2 years. Both volcanoes' signals are clearly evident in a difference plot. The signal of El Chichón is about half the standard deviation of the annual temperature variation, while that of Mount Pinatubo is larger than this amplitude (Fig. 5).

As the volcanic aerosols were mostly injected near the equator and their effects of increasing heating rates and temperatures are also concentrated in the tropical region, we find the main SWV perturbations in the tropical stratosphere as seen in Fig. 6 as a difference plot (VOL-NOVOL) for both volcanoes. The increased amount of water vapour in the stratosphere ranges from below the cold point at 90 hPa up to 10 hPa and is located in the upwelling region of the Brewer–Dobson circulation (BDC) in the tropics.

Both volcanic periods show an increase in SWV shortly after the eruption compared to the simulation without volcanoes (NOVOL). The absolute maximum for El Chichón of around 0.3 ppmv^1 is located around 90 hPa and is reached approximately one year after the eruption in the NH summer season, whereas the increases in water vapour for Mount Pinatubo result in a double peak maximum. The first peak is located at a height of around 80 hPa some 9 months after the eruption and the second is propagating from near 100 hPa with a total increase of 1 ppmv starting one year after the eruption. The signals are then propagating similar to the tropical tape recorder to higher altitudes of the stratosphere.

¹1 ppmv = 1 μ mol mol⁻¹ in SI units.

The relative increases compared to the background values of NOVOL are up to 20 % for El Chichón at a height between 90 and 80 hPa occurring in the first winter after the eruption (Fig. 6). For Mount Pinatubo there is some kind of a triple peak structure of relative maxima around the same height, with the first maximum showing an increase of 40 % also in the first winter after the eruption. The second is following in the same year's summer season with a magnitude of 50 % increase in SWV and is tailed by a third local maximum in the following winter season (1992/93) with a relative increase of about 45 %.

The small local SWV maximum found between 20 and 30 hPa shortly after the eruption of El Chichón coincides with the local maximum of temperature increase of around 1 K (Fig. 1). Also the maximum of relative increases in water vapour around one year after the El Chichón eruption is related to the temperature increase of 1.5 K, which also occurs one year after the eruption at the same pressure level. Correspondingly, for Mount Pinatubo the absolute maximum of temperature increase can be found at a height of around 40 hPa in November 1991 to February 1992, where also a small absolute (around 0.3 ppmvppmv) and a relative increase (around 10 %) in SWV occurs in the winter months after the eruption.

The negative values of water vapour changes found at a pressure level of around 50 hPa and above for both volcanoes are associated with the uplifting of air characterised by lower SWV mixing ratios through the additional volcanic heating in this area. The SWV minima are propagating similar to the tropical tape recorder to higher altitudes.

The time lag between the local maximum of induced heating and the propagation of the SWV signal into the same height is about 27–28 months for El Chichón and \sim 26 months for Mount Pinatubo.

As the tropopause is located within the area affected by the induced heating of the stratosphere, changes in cold point characteristics can be found in the simulation (Fig. 7). The cold point temperature increases with a maximum of around 1.4 and 2.4 K, respectively, approximately one year after the eruptions. The pressure at the cold point changes by up to 1 and 3 hPa, respectively, within 9–12 months. The specific humidity is also increasing with a maximum of about 0.25 and 0.55×10^{-6} kg kg⁻¹, respectively, approximately 18 months after both volcanic eruptions. With rising temperatures at the cold point more water vapour is transported into the stratosphere (Randel et al., 2004). The increased humidity at the cold point supports this conclusion. The displayed changes in pressure at the cold point can be explained through changes in the local temperature gradient, which differs between both simulations. Because of the temperature changes the local cold point is found around 1.2 and 3 hPa, respectively, below its unperturbed altitude.

Figure 8 shows the water vapour perturbation at the height of the local relative maxima at about 80 hPa. The time series shows a triple peak (compare relative changes of Fig. 6) of differences in absolute values for Mount Pinatubo. The first is occurring in late 1991, the second in the beginning of 1992 and the third in late 1992 with a maximum of about $\sim 0.9 \text{ ppmv}$. For El Chichón the increase in water vapour at this height results in a longer lasting peak, starting shortly after the eruption and rising to a maximum of about 0.4 ppmv in late 1983. There is a small second maximum in mid 1984 with an amplitude of 0.2 ppmv.

Figure 9 represents a latitude-time cross-section for the changes in water vapour for both volcanic periods near the 90 hPa level. As it can be seen, the largest absolute increases of water vapour are in the extra-tropical region of the Northern Hemisphere. The maxima reach 0.6 ppmv for El Chichón and 1.2 ppmv for Mount Pinatubo, respectively, in the second year after the eruption. Both extra-tropical maxima are located mainly between 10 and 40° N. However, the maximum relative increase in SWV is in the tropics about 20 % for El Chichón and 40–50 % (a double peak) for Pinatubo compared to the background value of NOVOL. Interesting is also the periodicity of the maxima, especially for the El Chichón eruption, as it occurs in the NH summer months for the three summer seasons after the eruption. The Pinatubo period is similar to El Chichón, but lacks a third maximum in 1993.

These results underline the special importance of the extra-tropical region for the transport of water vapour into the stratosphere. This aspect will be addressed in the following section.

4.2 Monsoon and extra-tropical influences

Dessler et al. (1995) found in observational data that moist air is able to enter the stratosphere in subtropical regions by traveling travelling along isentropic surfaces. Figure 10

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shows latitude-height cross-sections for the NH summer months (June, July, August) one year after the eruption of El Chichón and Mount Pinatubo, respectively. The increase in SWV appears globally. Both volcanoes exhibit a strong signal around 20 to 40° N in the NH, propagating through the tropopause into the stratosphere, and reaching a height of 90– 80 hPa. The time period shown represents the SWV increases with mixing ratios of around 0.6 ppmv ppmv for El Chichón and 1.2 ppmv for Mount Pinatubo, respectively. As the temperature of the tropopause in the subtropics is far higher (around ~ 200–225 K) than in the tropics (normally below 190 K), the ascending air parcels are characterised by a higher saturation vapour pressure and therefore by a higher mixing ratio of water vapour leading to a moistening of the extra-tropical stratosphere.

Upward transport of water vapour occurs predominantly during the South Asian summer monsoon (SASM), which is determined as a significant source of moisture for the upperlevel of the monsoon anticyclone and the lower extra-tropical stratosphere (Dethof et al., 1999; Eichinger et al., 2015b). The SASM is located mainly over northern India, the Tibetan plateau, central Asia and China and is associated with strong seasonal circulation anomalies and the isolation of air masses, which starts in June and ends in September.

In the month of August, directly after both eruptions took place, a significant increase in SWV with a magnitude of one standard deviation (of the NOVOL simulation) showed up in that region until August of the second year (Fig. 11). The third year's August does not show any signs of a further increase. According to one of the "stratospheric fountains" (Newell and Gould-Stewart, 1981) the Asian summer monsoon can be accounted for increasing SWV amounts after volcanic eruptions through transport of water vapour from the troposphere into the stratosphere.

The large-scale circulation patterns of the SASM can reach deep into the subtropics and are primarily driven by thermal processes and linked to convective latent heating. These processes lead to an appearance of anticyclones in the upper troposphere and lower stratosphere, able to reach up to around 70 hPa (Dunkerton, 1995), and to an exchange of air masses through deep convection. The moist air passes the dynamical tropopause traveling travelling along isentropes, which cross the tropopause in this region. Through the extent of

the anticyclone into the lower stratosphere the air is not freeze-dried by passing regions with low temperatures. This moistening can only be found in the NH and reaches its maximum in the boreal summer months, however, the total strength varies from year to year.

As we will further explain in Sect. 5 the moistening of the stratosphere due to the SASM depends on the amount of heating of the lower stratosphere due to the volcanic aerosol.

4.3 Influence on tropospheric water vapour and ENSO

In the first winter season after both eruptions increased amounts of water vapour (VOL compared to NOVOL) propagate from the surface up to the tropopause. Because of the large background values compared to the stratosphere, the relative increase is small. This is in agreement with the findings of Soden et al. (2002), because we analyse here the difference between VOL and NOVOL (i.e., the effect of the long-wave heating of the volcanic aerosol), instead of the transient periods before and after the eruption. The decreased tropospheric water vapour after the eruptions, as found by Soden et al. (2002), is included in both simulations and therefore cancels out in the difference. Our result therefore suggests that the drying effect after the eruptions is weakened by the aerosol induced heating.

The water vapour "column" reaches a pressure level of $\sim 150\,h\text{Pa}$. Large amounts of water vapour are able to penetrate the tropopause at the time with the largest increases in cold point temperature around one year after the eruptions (Fig. 7).

The water vapour anomalies in the troposphere coincide with ENSO-events, at least the signals beginning in the first December after the eruptions. El Niños are generally strongest in the season from December to April and have a large impact on the weather system. They result in increased tropical convection and general changes in the circulation around the tropical tropopause (upwelling) due to positive temperature effects. The occurring temperature perturbations can expand well above the tropopause into the stratosphere (see Scherllin-Pirscher et al., 2012 and references therein).

The increased temperatures also lead to more water vapour in the troposphere, which subsequently propagates into the stratosphere. Fueglistaler and Haynes (2005), as well as Scaife et al. (2003) therefore related El Niño situations with a moistening of the stratosphere.

events) after 1992 seem to coincide well with the increases in water vapour (Fig. 12). After the volcanic eruptions The volcanic aerosol induced long-wave heating in VOL reduces the total available potential energy (CAPE) featured a drop in comparison to NOVOL (not shown). This results in a reduced convective activity due to an increase in atmospheric stability caused by the volcanic heating and the steeper temperature gradient in the upper troposphere. Less convection leads to a subsequent decrease in precipitation and a weakening of the hydrological cycle. The water vapour in the atmosphere is not converted into precipitation and hence is transported to higher altitudes, where increased tropopause temperatures allow more water vapour to enter the stratosphere. This supposes that major volcanic eruptions influence El Niños and significantly amplify the moistening of the tropical stratosphere.

So far the results can be summarised that volcanic eruptions increase aerosol induced heating increases the concentrations of tropospheric and stratospheric water vapour. Increased temperatures and therefore higher saturation vapour pressures at the tropopause allow additional water vapour to transit into the stratosphere. There, the SWV gets dispersed along characteristic circulation patterns (i.e., BDC, tropical tape recorder and tropical pipe; Plumb, 1996). The values of water vapour in the stratosphere reach their peak after about 18 months and it takes the signal 3 to 4 years to decay.

5 **Sensitivity studies**

As outlined in Sect. 2.3, we additionally performed 3 pairs of sensitivity studies to assess the uncertainties of our findings. The Figures (named Sn (with n = 1, 2, 3, ...), shown in the Supplement, repeat the results of the VOL-NOVOL pair and show the corresponding results of the sensitivity pairs in addition.

Our first sensitivity simulation pair "RE" repeats the VOL and NOVOL simulations, but (solely to reduce computational costs) with prescribed (monthly average) instead of fully interactive chemistry. As such, "RE" serves as a new reference simulation for "QF" and "LA". The zonally averaged differences in heating rates of RE (Fig. S1, upper panels) are a bit smaller compared to those of the VOL-NOVOL pair. The prescribed chemistry therefore seems to reduce the volcanic aerosol temperature effect in the stratosphere (Figs. S2; S3 upper panel; and S4). Correspondingly, the water vapour perturbance in RE is also smaller than in the VOL-NOVOL pair (Fig. S5, upper panels; S6a,b; S7, upper panels).

We know from Sect. 3 that in our VOL simulation the heating effect of the volcanic aerosol is probably overestimated, because the resulting cold point temperature anomaly is too large compared to reanalysis data. This is also shown by other models. Arfeuille et al. (2013) discussed the potential causes of this overestimation. They found that uncertainties arise from the aerosol extinctions and the method of how they are derived from observations, and from radiative and dynamical model artefacts. To examine the uncertainty related to the aerosol load and characteristics and the simulated heating rates, we performed a simulation, in which we scaled the prescribed aerosol extinction by one half (LA). The resulting vertical heating rate distribution is accordingly reduced by about one half (Fig. S1, lower right panel). Also the temperature differences in the tropics (shown at different pressure levels in Fig. S2 and at the cold point in Fig. S3, upper panel) of the LA pair is half as large as those of the RE pair. However, the water vapour difference of LA (Fig. S3 middle and lower panel; Fig. 4 for 80 hPa; Fig. S5 for zonal average at 90 hPa over time; Fig. S7 for 5°S-5°N average vertical cross section over time) is only roughly 1/3 of that of RE.

To examine the effect of the applied nudging on the results, a pair of simulations, in which only the (logarithm of the) surface pressure is nudged, was performed (QF). The calculated differences in heating rates (Fig. S1, lower left panel) show similar patterns as the RE pair, however, more noisy. The tropical tropopause is heated more as in RE (Fig. S2). However, the resulting cold point temperature change is comparable with that of RE (Fig. S3, upper panel). The warming of the upper stratosphere in QF is the largest of all

simulation pairs. The corresponding water vapour changes in 3 distinct periods of time, but similar in magnitude to RE (Fig. S5), and with similar patterns in the tropical tape-recorder (Fig. S7).

Overall, the influence of the Mount Pinatubo induced aerosol heating in the stratosphere is similar in all simulation pairs (RE, QF, LA) compared to FC, although the magnitude differs. We find a moistening of the stratosphere and troposphere in the months June to August in the year after the eruption (1992), which has decayed already in 1993 (Figs. S6a–d). The effect is smallest for LA, because the heating of the tropopause is the smallest. The moistening effect of the troposphere (due to heating of the upper troposphere with subsequent stabilisation of the temperature stratification), however, is only visible in FC and RE. In QF it is overlayed by meteorological noise, and in LA the heating is too small.

The zonal mean vertical upwelling differences (VOL-NOVOL) of all simulation pairs show the same structure (Fig. S8): stronger upwelling in the middle/upper stratosphere due to the heating of the volcanic aerosol, and reduced upwelling until 1993 at the tropopause, which, however, lasts longer in FC. The volcanic aerosol induced heating in the stratosphere reduces the upwelling at the tropopause, and thus warms it (at least partly) dynamically. This also holds for the "guasi free-running" simulation pair QF.

We conclude that our main findings presented in Sect. 4 about the influence of the volcanically induced heating in the stratosphere on tropospheric and stratospheric water vapour changes are independent of the applied nudging, but depend on the heating strength.

6 Discussion

Volcanic aerosol has two effects on atmospheric temperature: it warms the stratosphere due to long wave radiation effects and cools the surface and thus the troposphere due to a shadowing of the surface in the short-wave radiation. Indeed, all our simulations show a cooling of the lower troposphere up to the cold point (Fig. 2) and subsequent lower water vapour mixing ratios (not shown) after both eruptions.

This is due to the nudging and prescribed SSTs and in accordance with observations (e.g. Parker et al., 1996; Soden et al., 2002). The stratospheric warming through the volcanic aerosol is, however, only accounted for in the VOL simulation. Hence, the cooling signal at the tropopause cancels out and the lower stratospheric warming signal remains, if we subtract VOL-NOVOL.

The volcanic forcing Similar as other simulations of volcanic eruptions. our VOL simulation tends to overestimate the eruption of Mount Pinatubo (our Fig. 2; SPARC, 2010; Gettelman et al., 2010), whereas El Chichón is captured well in terms of cold point temperature and water vapour at 80 hPa. Unfortunately, the lack of observations of stratospheric water vapour directly after the eruptions renders a more in-depth comparison difficult (Fueglistaler, 2012). Water vapour anomalies from ERA-Interim have to be analysed with caution, although they have been improved in comparison to ERA-40 reanalysis data Dee et al. (2011). Yet, the results differ significantly from a recently published merged data set based on satellite observations (MSD; our Fig. 3; Hegglin et al., 2014). For instance, at the time of the eruption of Mount Pinatubo ERA-Interim shows a minimum, and MSD a maximum in the water vapour anomaly. Moreover, for the ERA-Interim reanalysis an ozone climatology has been applied for the calculation of heating rates (Wright and Fueglistaler, 2013), whereas ozone is calculated interactively in VOL and NOVOL.

The NOVOL simulation shows a strong minimum in cold point temperature anomaly and consequently in water vapour anomaly after the eruption of Mount Pinatubo. This means that the synoptic situation offsets a part of the volcanic heating in the lower stratosphere as was already suggest by Fueglistaler (2012), because observations do not show a significant increase in cold point temperature or water vapour after the Mount Pinatubo eruption. Therefore, it is guestionable that simulations not capturing the cooling of the tropopause due to the specific synoptic conditions can simulate the effect of Mount Pinatubo correctly. Actually, such simulations must necessarily overestimate the stratospheric heating effect of the eruption and hence the water vapour anomaly. Arfeuille et al. (2013) pointed out the role of the appropriate size distribution retrieval for the prescribed aerosol. Although

we followed their suggestions and used the most recent data, we still find our Mount Pinatubo warming signal overestimated (Fig. 2). Whether this is related to deficiencies in the forcing data, or can be attributed to radiative artefacts remains unresolved. A sensitivity simulation with only half the aerosol extinction values of VOL leads to lower heating rates in the stratosphere and a cold point temperature, which is more in accordance with ERA-Interim.

The volcanic long-wave radiative forcing heats the lower stratosphere mainly in the tropical region, leading to an increase in temperatures for about 2 years after the eruptions with a maximum of 1.5 K for El Chichón and 4 K for Mount Pinatubo, respectively. Both can be identified as significant changes, though the overall temperature increase may be overestimated by our model by about 1 The large temperature change (VOL-NOVOL) for the Mount Pinatubo eruption seems to be in contradiction to observations, which state a heating of roughly 2 K similar to the El Chichón eruption (Angell, 1997; Randel et al., 2000). However, the large warming of the cold point temperature of VOL-NOVOL (Fig. 7, upper panel) shows the long-wave warming due to the volcanic aerosol, only. Concurrently, NOVOL shows a decrease of the cold point temperature, which is probably caused by an increase in upwelling in the period around the eruption of Mount Pinatubo.

Considine et al. (2001) used an interactive 2-D model simulation to evaluate the effects of the volcanic aerosols of Mount Pinatubo. They used observations of extinction rates, size distribution and aerosol surface area densities to simulate the aerosol effects. They also compared their resulting temperature changes after the eruption with the findings of Angell (1997), who used radiosonde data and removed QBO effects, and values from NCEP analysis on three different pressure levels (20, 30 and 50 hPa, see Fig. 8 of Considine et al., 2001). The results from Angell (1997) peaked with a temperature increase of approximately 3–4 K at a height between 30 and 50 hPa in late 1991, which agrees well with the results in the present study (Fig. 5 and also Fig. 1 in Sect. 2.2) with a peak of 4 K occurring around 40 hPa between 1991 and 1992. Also the duration of the temperature signal of approximately 2 years is in agreement with the results presented here. This is supported by the simulation of Joshi and Shine (2003), who obtained similar results like in the NCEP analy-

sis with a more complex GCM than the 2-D model of Considine et al. (2001). On 20 hPa it agrees better with the model of Considine et al. (2001), on 50 hPa however, better with the results from Angell (1997). The results exceed the regional standard deviation.

The stratospheric temperature increase leads to elevated temperatures at the cold point by about 1.4 and 2.4 K. The resulting higher saturation vapour pressure of the air allowed more water vapour to enter the stratosphere through the tropopause, leading to SWV increases in the tropics of 20 % for El Chichón and 50 % for Mount Pinatubo, respectively, in the lower stratosphere. Two (Mount Pinatubo) to 3 (El Chichón) summer seasons after the eruptions, the South Asian summer monsoon could be determined as a significant source of additional stratospheric moisture. In the NH summer months increased amounts of water vapour entered the stratosphere over the SASM anticyclone, peaking in the second year after the eruptions with an increase of 0.5 ppmv-ppmv for El Chichón and 1 ppmv for Mount Pinatubo, respectively.

Considine et al. (2001) also studied the changes in stratospheric water vapour that occurred after the volcanic aerosol forcing of Mount Pinatubo. They mostly concentrated on changes in trends, but also showed the response of SWV perturbations in their model (their Fig. 16b) in a very similar way like in Fig. 6 in the previous section of our study.

Some of the facts of their figure results of Considine et al. (2001) are worth being pointed out: their model also simulated some kind of a double peak signal. The first peak occurring occurs in early 1992 with its maximum (approximately a 30–35% increase) at a height of ~ 80 hPaand a second peak in the change of year 1992–1993. The second peak appears at the turn of the year 1992/93, with a signal that is smaller in magnitude (around 25%). A comparable peak structure for the specific humidity changes was shown by Joshi and Shine (2003) for the NCEP analysis data(their Fig. 3, bottom), but their model wasn't able to reproduce this – (their Fig. 3, bottom). They referred to Angell (1997) for an explanation of the occurring double peak as an influence of the QBO. Both signals in Considine et al. (2001) are propagating into higher regions of the stratosphere, which. This is in good agreement with our model results. Additionally, they also simulated a SWV minimum shortly after the eruption, which propagates from around 20 hPa higher into the stratosphere. The,

to some extent, lower relative values can probably be explained through smaller temperature changes associated with the volcanic forcing. They also mentioned that HALOE H_2O data is lacking a clear signal of water vapour increase for the Mount Pinatubo period and therefore assumed that it may be possible that the simulated temperature changes at the tropopause, which control the entry value of water vapour in that region, are too high. They further concluded that in reality a Mount Pinatubo signal in tropopause temperatures was masked by the inter-annual variability of the tropopause of 1-2 K.

Joshi and Shine (2003) also found the maximum increases in their model over the equatorial regions, but their results did not indicate any sign of increased transport from the extratropical troposphere into the stratosphere (their Fig. 4). In our simulations the occurrence of a tropospheric drying² effect after volcanic eruptions is masked by the slight differences between the two used simulations is represented in both simulations (due to nudging and prescribed SSTs/SICs) and therefore cancels out of the differences calculations (Fig. 12).

A comparison of our results with observations is difficult, because water vapour measurements prior to 1994 may be noisy and biased due to the volcanic aerosol layer (Fueglistaler, 2012). Interestingly, our results of the VOL simulation show a period with less water vapour compared to the NOVOL simulation without volcanic eruption (Fig. 6) directly after the eruption of Mount Pinatubo similar as present in the HALOE data shown by Fueglistaler (2012), his Fig. 5a. This period of reduced water vapour amount is also visible shortly after the eruption of El Chichón. We suggest that this effect is associated with the uplifting of air through the additional volcanic heating in this area.

Moreover, a significant increase of water vapour is propagating from the troposphere into the stratosphere during the El Niños in 1982–1983 and 1991–19921982/83 and 1991/92, in addition to the anyway elevated values during El Niños. The volcanic heating increased the upper tropospheric stability and therefore reduced the convective activity, which led to less precipitation and more available water vapour in the atmosphere. Thus, our results suggest that the drying effect after the eruptions is weakened by the aerosol induced heating. Once

²The drying results from decreasing tropospheric temperatures due to less absorption of solar radiation, which is "blocked" by the volcanic aerosols (e.g., Soden et al., 2002).

the water vapour signal reaches the stratosphere, it gets dissipated by the typical stratospheric circulation patterns of the BDC, the tropical pipe and the tropical tape recorder. An overview of the mainly affected tropospheric processes is given in Fig. 13.

The reaction of the model to the volcanic perturbation also resulted in dynamical changes on a sub-synoptic scale, which influenced the vertical and horizontal winds. Most changes were located within regions of altered water vapour abundance. Strengthening of vertical motion was found mainly in the tropical region in the stratosphere, resulting directly from local heating rates. The differences in vertical velocity accounted for about 0.1– $0.2 \, 10^{-3} Pa s^{-1}$ after the eruption of Mount Pinatubo. Though, these results have to be considered with care, as the simulations were influenced through the applied Newtonian relaxation technique, which affected the dynamical variables vorticity and divergence.

7 Summary and outlook

The two nudged simulations RC1SD-base-01 (with volcanic perturbation, VOL) and RC1SD-base-10 (no volcanic perturbation, NOVOL) were used to carry out a sensitivity analysis of the effects of two major volcanic eruptions (El Chichón and Mount Pinatubo) on stratospheric water vapour (SWV). In particular, we analysed the long-wave heating effect of the volcanic aerosol on the temperature and water vapour distribution. To simulate the effects of the volcanic eruptions, the VOL simulation used prescribed monthly and zonally averaged optical properties (optical thickness, asymmetry factor and single scattering albedo) of the volcanic aerosol. These were derived from observational data and used in the model to calculate the heating rates, which result in thermal, dynamical and chemical changes. The nudging setup was chosen to allow a temperature response to the volcanic aerosol driven heating. Here the effects on the hydrological cycle, in particular the SWV distribution and time development, was investigated.

Our results (as sketched in Fig. 13) are in good agreement with findings from other literature previous studies, especially in structure and duration of the volcanic signals. We found increased that volcanic aerosol heats the stratosphere including the cold point,

which leads to increased and upward propagating amounts of water vapour in the stratosphere shortly after the eruptions, which propagate into higher regions of the stratosphere. The South-Asian summer monsoon (SASM, Fig. 13) was identified as a source of additional SWV for at least two years after the eruptions. A simulation with lowered volcanic aerosol extinction confirms this result, but shows a smaller amplitude. Additionally, the burden of tropospheric water vapour was increased during the El Niños of 1982/83 and 1991/92. We found that the burden of tropospheric water vapour was increased during the El Niños 1982/83 and 1991/92 only for the VOL simulation, when the volcanic aerosol caused the largest heating rates in our simulations. Volcanic aerosol induced long-wave heating has therefore the potential to further attenuate the hydrological cycle appearing primarily do to the (short-wave shielding induced) cooling of SSTs and the troposphere: if the stabilisation of the vertical temperature profile is strong enough to suppress convection and consequently convective precipitation. Because of a weakened hydrological cycle (i.e., less precipitation) and increased temperatures at the tropopause the water vapour was able to propagate into the stratosphere. In conclusion, strong volcanic eruptions block the sun by the injection of aerosol particles into the stratosphere, where they remain for years leading to a cooling of the surface. Locally these aerosols heat the middle stratosphere down to the tropopause and subsequently increase the amount of water vapour transported into the stratosphere. Additionally, periods of strong convective activity (e.g., El Niños) in the tropics are can be influenced by the stabilisation of the vertical temperature gradient, if the heating reaches sufficiently down. Thus, the modification of the atmospheric water vapour due to the long-wave heating of the atmosphere comprises not only the stratosphere but the whole vertical column. Our focus in this study was the estimation of SWV increases and the related transport paths. Volcanic eruptions, however, also influence chemical processes in the stratosphere. For instance the ozone generation is influenced by temperature changes. Moreover, SWV formed by methane oxidation as a source of SWV needs further investigation.

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Figure 1. Zonally averaged heating rates [K d⁻¹] as differences (VOL-NOVOL) in the tropics (5° S- 5° N) for **(a)** the 1982 El Chichón and **(b)** the 1991 Mount Pinatubo eruption. Contours indicate absolute temperature changes (interval 0.5 K) due to the heating rates.



Figure 2. Anomaly of the monthly and zonally averaged tropical (20°S–20°N) cold point temperature [K] calculated from ERA-Interim (black) and as simulated in the NOVOL (blue) and VOL (green) simulations. The purple line (VOL-LA) shows the result of a sensitivity simulation (Mount Pinatubo period, only), for which the stratospheric aerosol extinction was scaled by a factor of 0.5. The anomalies have been calculated based on the period 1980–1996 for ERA-Interim, NOVOL and VOL, and 1990–1996 for VOL-LA, respectively, and further smoothed with a box-smoother of 3 months length. The vertically oriented dashed lines indicate the eruptions of El Chichón (March 1982) and Mount Pinatubo (June 1991), respectively.



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Figure 3. Anomaly of the monthly and zonally averaged tropical (20°S–20°N) water vapour [ppmv] at 80 hPa calculated from ERA-Interim (black) and as simulated in the NOVOL (blue) and VOL (green) simulations. The purple line (VOL-LA) shows the result of a sensitivity simulation (Mount Pinatubo period, only), for which the stratospheric aerosol extinction was scaled by a factor of 0.5. The anomalies have been calculated based on the period 1980–1996 for ERA-Interim, NOVOL and VOL, and 1990–1996 for VOL-LA, respectively, and further smoothed with a box-smoother of 3 months length. The red-line (MSD) shows the anomaly based on the merged satellite data as derived by Hegglin et al. (2014, their Fig. 2, upper panel). The vertically oriented dashed lines indicate the eruptions of El Chichón (March 1982) and Mount Pinatubo (June 1991), respectively.



Figure 4. Anomaly of the monthly averaged tropical (20°S–20°N) upwelling [10⁻⁶ Pa/s] at 100 hPa calculated from ERA-Interim (black) and as simulated in the NOVOL (blue) and VOL (green) simulations. The purple line (VOL-LA) shows the result of a sensitivity simulation (Mount Pinatubo period, only), for which the stratospheric aerosol extinction was scaled by a factor of 0.5. The anomalies have been calculated based on the period 1980–1996 for ERA-Interim, NOVOL and VOL, and 1990–1996 for VOL-LA, respectively, and further smoothed with a box-smoother of 3 months length. The upwelling (w*) has been calculated according to the TEM (Transformed Eulerian Mean) method as described by Holton (2004). The vertically oriented dashed lines indicate the eruptions of El Chichón (March 1982) and Mount Pinatubo (June 1991), respectively.

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Figure 5. Temperature [K] differences (VOL-NOVOL) for the tropics (5° S– 5° N), zonally averaged after the March 1982 El Chichón and the June 1991 Mount Pinatubo eruption for **(a)** 20 hPa, **(b)** 30 hPa and **(c)** 50 hPa. Red dashed lines indicate the standard deviation [K] for the unperturbed NOVOL simulation in the same region, calculated over the whole time series of 1979–2013.



Figure 6. SWV [ppmv, colours] as absolute differences (VOL-NOVOL) in the tropics (5° S– 5° N), zonally averaged for **(a)** the 1982 El Chichón and **(b)** the 1991 Mount Pinatubo eruption. Contours indicate relative changes of water vapour (interval 5%) compared to the background value of NO-VOL. The small signal seen in 1986 is associated with the November 1985 eruption of Nevado del Ruiz (Colombia).



Figure 7. Zonally averaged differences (VOL-NOVOL) in temperature, pressure and humidity at the cold point in the tropics (5° S– 5° N), zonally averaged for the El Chichón period (left) and the Mount Pinatubo period (right).



Figure 8. Differences (VOL-NOVOL) in water vapour [ppmv] for the tropics (5° S– 5° N), zonally averaged after the March 1982 EI Chichón and the June 1991 Mount Pinatubo eruption for the 80 hPa level. The red dashed line indicates the standard deviation [ppmv] for the unperturbed NOVOL simulation in the same region, calculated over the whole time series of 1979–2013.

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Figure 9. SWV [ppmv, colours] as absolute differences (VOL-NOVOL) zonally averaged near the 90 hPa level for **(a)** the El Chichón period (1982–1986) and **(b)** the Mount Pinatubo period (1991–1995). Contours indicate relative changes in water vapour (interval 5%) compared to the background value of NOVOL. The small signal seen in 1986 is associated with the November 1985 eruption of Nevado del Ruiz (Colombia).



Figure 10. SWV [ppmv, colours] as absolute differences (VOL-NOVOL), zonally averaged as a near global (60° S– 60° N) vertical cross-section at a height between 120 and 20 hPa for the months of June, July and August in the year following the eruptions (1983 and 1992, respectively). White contours indicate the relative increase in SWV (intervals 5, 10, 20 and 50%) compared to the background value of NOVOL. Black dashed contours mark increases in units of standard deviation for the particular month in NOVOL (calculated over the time period 1979–2013). The left column shows the El Chichón period, the right column the Mount Pinatubo period.



Figure 11. SWV [ppmv, colours] as absolute differences (VOL-NOVOL) as a near global (60° S- 60° N) horizontal cross-section at 75 hPa for the month of August in the year of the eruptions and in the following two years. White contours indicate the relative increase in SWV (intervals 5, 10, 20 and 50%) compared to the background value of NOVOL. Black dashed contours mark significant increases in units of standard deviation for the month of August in NOVOL (calculated over the time period 1979–2013). The left column shows the El Chichón period, the right column the Mount Pinatubo period.



Figure 12. Same as Fig. 6 but for 1000–10 hPa with nearly logarithmic contour intervals from 1 to 20%.



<u>Equato</u>r

Figure 13. Schematic of tropospheric processes influenced by a strong volcanic eruption.

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Supplemental information on "Impact of major volcanic eruptions on stratospheric water vapour"

by

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This supplement provides the results of additional sensitivity simulations, of which the model setups are summarised in Table S1. Details about the model setups are described in Section 2.2 of the main manuscript, and the results are discussed in Section 3 of the main manuscript.

Table S1: Description of simulation pairs, one including the stratospheric aerosol extinction and one excluding it.

simulation pair	description of setup
FC	original (nudged) simulations with full chemistry
RE	remake simulations (nudged) with prescribed monthly average chemistry
\mathbf{QF}	quasi free running simulations with prescribed monthly average chemistry
	and nudging of (logarithm of) surface pressure, only
LA	low a erosol (nudged) simulations with prescribed monthly average chemistry
	and by a factor of 0.5 reduced aerosol optical depth



Figure S1: Zonally averaged heating rates $[K d^{-1}]$ as differences (VOL-NOVOL) in the tropics (5°S–5°N) for the Mount Pinatubo eruption. Upper left panel: original simulation (nudged) with full chemistry (FC), upper right panel: "remake" simulation with prescribed chemistry (RE), lower left panel: "quasi free" running simulation (QF), and lower right panel: "low aerosol" simulation (LA). Contours indicate absolute temperature changes (interval 0.5 K) due to the heating rates.



Figure S2: Temperature [K] differences (VOL-NOVOL) for the tropics (5° S- 5° N), zonally averaged after the June 1991 Mount Pinatubo eruption for 20 hPa (upper panel), 30 hPa (middle panel) and 50 hPa (lower panel). The different simulation pairs are coloured as labelled in the upper panel.



Figure S3: Zonally averaged differences (VOL-NOVOL) in temperature (upper panel), pressure (middle panel) and specific humidity (lower panel) at the cold point in the tropics (5°S-5°N), zonally averaged for the Mount Pinatubo period. The different simulation pairs are coloured as labelled in the upper panel.



Figure S4: Differences (VOL-NOVOL) in water vapour [ppmv] for the tropics (5° S- 5° N), zonally averaged after the June 1991 Mount Pinatubo eruption for the 80 hPa level. The different simulation pairs are coloured as indicated.



Figure S5: SWV [ppmv, colours] as absolute differences (VOL-NOVOL) zonally averaged near the 90 hPa level for the Mount Pinatubo period (1991–1995) for the different simulation pairs as indicated. Contours indicate relative changes in water vapour (interval 5%) compared to the background value of NOVOL. The different simulation pairs are labelled according to Table S1.



Figure S6a: Left column: SWV [ppmv, colours] as absolute differences (VOL-NOVOL), zonally averaged as a near global (60° S- 60° N) vertical cross-section at a height between 120 and 20 hPa for the months of June, July and August in the year following the eruption of Mount Pinatubo. White contours indicate the relative increase in SWV (intervals 5, 10, 20 and 50 %) compared to the background value of NOVOL. Right column: SWV [ppmv, colours] as absolute differences (VOL-NOVOL) as a near global (60° S- 60° N) horizontal cross-section at 75 hPa for the month of August in the year of the eruption and in the following two years. White contours indicate the relative increase in SWV (intervals 5, 10, 20 and 50 %) compared to the background value of NOVOL. Shown are the results of the simulation pair FC.



Figure S6b: Shown are the results of the simulation pair **RE**.



Figure S6c: Shown are the results of the simulation pair QF.



Figure S6d: Shown are the results of the simulation pair LA.



Figure S7: SWV [ppmv, colours] as absolute differences (VOL-NOVOL) in the tropics ($5^{o}S-5^{o}N$), zonally averaged for the 1991 Mount Pinatubo eruption. Contours indicate relative changes of water vapour (interval 5 %) compared to the background value of NOVOL. The different simulation pairs are labelled according to Table S1.


Figure S8: Upwelling $[10^{-6} \text{ Pa s}^{-1}]$ as absolute differences (VOL-NOVOL) in the tropics $(20^{\circ}\text{S}-20^{\circ}\text{N})$ for the 1991 Mount Pinatubo eruption. The different simulation pairs are labelled according to Table S1.