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2 **Radiative impact of the Hunga Tonga-Hunga Ha'apai stratospheric volcanic plume:**
3 **role of aerosols and water vapor in the southern tropical Indian Ocean**
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16 **Key Points:**

- 17 • One year after HTHH eruption, aerosols and water vapor from the volcano are still
18 present in the stratosphere
- 19 • The aerosol plume has decayed with time, while the moist layer ascended
- 20 • The descending rate of the volcanic plume is $-0.008 \text{ km day}^{-1}$; at this rate, aerosols will
21 remain in the stratosphere until mid-2025
- 22 • So far, aerosols and WV have contributed to warm the Earth-Atmosphere system in the
23 southern tropical Indian Ocean ($+0.06 \pm 0.45 \text{ W m}^{-2}$)
- 24
25

26 Abstract

27 This study attempts to quantify the radiative impact over Reunion Island (21°S, 55°E) in the
28 southern tropical Indian Ocean of the aerosols and water vapor injected in the stratosphere by the
29 eruption on 15 January 2022 in the South Pacific of the Hunga Tonga-Hunga Ha'apai underwater
30 volcano. Ground-based lidar and satellite passive instruments are used to parametrize a state-of-
31 the-art radiative transfer model for the first thirteen months after the volcano eruption. The
32 descending rate of the aerosol volcanic plume is $-0.008 \text{ km day}^{-1}$. At this rate, aerosols are expected
33 to be present in the stratosphere until the first half of 2025. The overall aerosol and water vapor
34 impact on the Earth's radiation budget for the whole period is positive (warming, $+0.06 \pm 0.45 \text{ W}$
35 m^{-2}) and dominated by the aerosol impact. However, the decreasing rate with time of the aerosol
36 warming effect is larger than that of the water vapor cooling effect, so that, in the long run, the
37 impact on the Earth's radiation budget might reduce to quasi-neutral, or even become slightly
38 negative. At the Earth's surface, aerosols are the main driver and produce a negative (cooling, $-$
39 $0.91 \pm 0.61 \text{ W m}^{-2}$) radiative impact with also a decreasing tendency with time. Heating/cooling
40 rate profiles show a clear vertical difference in the stratosphere between the aerosol warming
41 impact (17 to 25 km) and the water vapor cooling one (25 to 40 km).

42 1 Introduction

43 More than one year and a half after the eruption of Hunga Tonga-Hunga Ha'apai (HTHH)
44 underwater volcano in the South Pacific, the science community is still actively investigating the
45 climate impact of the huge amounts of ash, water, steam and gases that were injected in the
46 atmosphere. The event showed an extremely fast spatio-temporal, global dispersion of the
47 stratospheric volcanic matter that circulated the Earth in only one week (Khaykin et al., 2022) and
48 diluted pole-to-pole in three months (Taha et al., 2022), not simply passively but rather organized
49 in concentrated patches (Legras et al., 2022). Several figures are evidences of a record-breaking
50 atmospheric event. The eruption, equivalent to an energy of 110 Tg of TNT, is the most powerful
51 volcanic explosion since Krakatau (1883) and Tambora (1815) (Lac et al., 2022). The volcanic
52 plume reached an altitude of 57 km, a coincident estimation resulting from different techniques
53 (Carr et al., 2022; Proud et al., 2022), placing it in the upper stratosphere – lower mesosphere, a
54 record in the satellite era. The water injected was unprecedented: (Millán et al., 2022) estimated
55 to 146 Tg the mass of water injected in the atmosphere (.e.g. the 1991 Pinatubo eruption released
56 37 Tg of water into the atmosphere (Pitari & Mancini, 2002)). In contrast, sulfur dioxide (SO₂)
57 mass injection was not that exceptional: $\sim 0.6 - 0.7 \text{ Tg}$ (Carn et al., 2022) which is much smaller
58 than that from previous major eruptions (e.g. 20 Tg for Pinatubo (Bluth et al., 1992)). Still, the
59 stratospheric aerosol optical depth (sAOD) has been recorded as the largest since Pinatubo eruption
60 (Taha et al., 2022) and peaked at values never observed before (Baron et al., 2023).

61 The latter three variables (water, SO₂ and injection height) are some of the main factors
62 responsible for the production of volcanic sulfate and for the loss/production of ozone. The initial
63 SO₂ was fully converted into sulfates in less than two weeks under the influence of water vapor
64 (Asher et al., 2022; Legras et al., 2022), whereas volcanic sulfate and water still persist as of today.
65 The fast water vapor injection provided abundant hydroxide (OH) which reacted with SO₂ to form
66 volcanic sulfate at a faster rate than the typical ~ 30 days. Higher concentrations of volcanic sulfate
67 led to more rapid coagulation and thus larger particles. In the case of HTHH, this mechanism is
68 estimated to have halved the SO₂ lifetime and doubled the sAOD (Zhu et al., 2022). This rapid
69 growth and global persistence of volcanic sulfate aerosols have been demonstrated with

70 AERONET measurements by (Boichu et al., 2023) with the occurrence of an unusual “volcanic
71 fine mode” with a peak ranging in 0.28 – 0.50 μm . This fine mode was found to be poorly
72 absorbing, although (Kloss et al., 2022) reports from balloon-borne measurements a moderately
73 absorbing fine mode in the first 10 days after the eruption indicating small sulfate coated ash
74 particles. Volcanic sulfate is known to be a factor to impact ozone depletion by providing
75 additional surface area and suppressing the nitric oxide cycle (Tie & Brasseur, 1995). The transport
76 of volcanic sulfate from the tropics to the Antarctic by the Brewer-Dobson circulation contributed
77 to increase ozone concentrations in the middle stratosphere but to decline in the lower stratosphere
78 at mid-to-low latitudes (Lu et al., 2023), while, combined with a cold polar vortex, it contributed
79 to decrease ozone concentration in the Antarctic (Wang et al., 2022). Because ozone is not emitted
80 primarily during volcanic eruptions, its loss or production by post-eruption reactions are more
81 tedious to estimate. The effect of HTHH on stratospheric ozone is still under study.

82 Water, volcanic sulfate and the injection height are the main drivers of the impact of HTHH
83 on atmospheric global circulation (Coy et al., 2022) and climate (Zuo et al., 2022). In particular,
84 the climate forcing will depend on the radiative effect produced by the water vapor longwave
85 emission and the aerosol shortwave and longwave scattering and absorbing properties. These
86 interaction mechanisms (emission, scattering and absorption) with the shortwave and longwave
87 radiation are highly height-dependent and determine the sign of the differential of energy gained
88 (positive) or lost (negative) in all layers of the atmosphere. Several studies have demonstrated the
89 stratospheric cooling produced by the excess of water vapor injected by HTHH either locally
90 (Sellitto et al., 2022), zonally (Schoeberl et al., 2022; Vömel et al., 2022; Zhu et al., 2022) or
91 globally (Millán et al., 2022) at different time scales spanning from instantaneous estimates to 6-
92 month evolutions. As far as volcanic sulfates are concerned, these aerosols usually scatter sunlight
93 back to space, cooling the Earth’s surface, and absorb outgoing thermal radiation. Several authors
94 have made the hypothesis that HTHH eruption could impact climate not through surface cooling
95 due to sulfate aerosols, but rather through surface warming due to the radiative forcing from the
96 excess stratospheric water vapor. The impact on the Earth’s radiation budget, i.e. at the top of the
97 atmosphere, is even more uncertain since smaller impacts, hence a greater sensitivity to variations,
98 are at play. To date, assessments of the radiative effect of combined water vapor and aerosols have
99 only been performed for 3 case studies during the first 10 days after the eruption by (Sellitto et al.,
100 2022) and for the first two months after the eruption by (Zhu et al., 2022).

101 Here, the impact of water vapor and aerosols on the Earth’s radiation budget is estimated
102 over Reunion Island (21°S, 55°E) for the first thirteen months after HTHH eruption. Both water
103 vapor and aerosols are fully parametrized in a state-of-the-art radiative transfer model by means
104 of ground-based lidar and satellite measurements. The radiative effect is calculated for three
105 scenarios considering aerosols only, water vapor only and combined aerosols and water vapor.

106 **2 Materials and Methods**

107 **2.1 The Maïdo instrumentation**

108 Two lidar systems located at the Observatoire de Physique de l’Atmosphère à La Réunion (OPAR)
109 Maïdo station (21.079°S, 55.383°E, 2160m asl; Baray et al., 2013) are used in this study: the
110 Li1200 lidar operating at 355 nm with 87 nights of observations between 19 January 2022 and 15
111 February 2023 and the LiO3T lidar operating at 532 nm with 55 coincident nights (19 January

112 2022 to 13 September 2022). Details on these systems can be found in (Baron et al., 2023) and
113 references therein.

114 The extinction coefficients presented in this work are obtained applying the elastic, 2-component
115 inversion algorithm (Klett, 1985) using a constant lidar ratio (LR). Several LR values were tested
116 between 40 and 70 sr at both wavelengths. The value of 60 sr was fixed for this study: it is the
117 value that gives the best agreement between OMPS and lidar sAOD after complete, homogenous
118 global dilution of the plume in the last 10 months of our dataset (see Section 3). The transmittance
119 method initially used in (Baron et al., 2023) for the thick plume observed during the first days after
120 the eruption over Reunion Island was not retained for at least two reasons: with decreasing aerosol
121 loads, the transmittance method would have led to large uncertainties in the LR retrieval; the
122 unreliability of the method for ground-based systems in low aerosol loads and at such altitude
123 levels (17 – 32 km). However, both Klett and transmittance methods were compared for significant
124 aerosol loads (i.e. in the first three months after the eruption) and a good agreement was obtained
125 for the value of 60 sr. Note that this value is also in the range of what is expected for sulfate
126 aerosols in the available literature (Lopes et al., 2019). During the first 55 measurements for which
127 coincident 355- and 532-nm sAOD are available, namely between 19 January until 13 September,
128 the Ångström exponent calculated between the lidar-derived sAOD at 355 and 532 nm, $AE_{355/532}$,
129 was computed. For the rest of the measurements at 355 nm, i.e. after 13 September 2022, $AE_{355/532}$
130 was set to the value of 1.5 calculated for the stabilized period 13 June 2022 – 13 September 2022
131 (see Section 3). The uncertainty associated to the extinction profiles at 355 nm, and by extension
132 to the sAOD at 355 nm, has been calculated considering an uncertainty on the lidar ratio of ± 10
133 sr. This uncertainty of ± 10 sr on the lidar ratio is the largest uncertainty calculated at 355 nm by
134 (Baron et al., 2023) for the HTHH plume over Reunion Island in January 2022. The uncertainty
135 associated to the sAOD at 745 nm has been calculated considering both the uncertainty associated
136 to sAOD at 355 nm and the uncertainty associated to $AE_{355/532}$, fixed to a constant value of ± 1.0
137 (Baron et al., 2023).

138 It is our belief that the results in Reunion Island can be easily generalized throughout the southern
139 tropical Indian Ocean region. (Mallet et al., 2018) reported for the first time the pristine
140 characteristics of the southern Indian Ocean region located between 10 and 40°S and between 50
141 and 110°E. Except its very northern boundary, this domain is not impacted by the longitudinal
142 transport of the Asian monsoon over the northern Indian Ocean. Tropospheric sea salt aerosols are
143 the dominant and the AOD-modulating aerosol type (Mallet et al., 2018). The same statement is
144 true over Reunion Island (DufLOT et al., 2022). The synoptic circulation in the southern tropical
145 Indian Ocean is strongly connected with the Mascarene anticyclone, which, because of its location
146 in the middle of this basin, limits the transport of terrestrial aerosols to this region. Several
147 indicators of the homogenous dilution of the volcanic plume in the stratosphere are exposed in
148 Section 3 and reinforce the assumption made from now that the results in Reunion Island can be
149 generalized to the whole southern tropical Indian Ocean region.

150 2.2 Satellite and reanalysis data

151 The Ozone Mapper and Profiler Suite (OMPS) Limb profiler has been on-board the Suomi
152 National Polar Partnership (NPP) since October 2011. Using limb scattering solar radiation, OMPS
153 provides good quality of aerosols extinction retrievals at several wavelengths: 510, 600, 675, 745,
154 869 and 997 nm (Taha et al., 2021). As recommended by the latter, we use data product version
155 2.0 of aerosol extinction profile at 745 nm to follow the aerosol volcanic plume over Reunion
156 Island, from January 2022 to mid-April 2023. These data are provided from 10 to 40 km height on

157 a vertical grid of 1 km. Stratospheric aerosol optical depth calculations are made by integrating the
158 extinction profiles from 17 km to 40 km, where 17 km corresponds to tropopause height over
159 Reunion Island (Bègue et al., 2010). Based on previous studies in the Southern Hemisphere (Bègue
160 et al., 2017; Tidiga et al., 2022), background periods extend from 2012 to February 2014 and from
161 January 2017 to April 2018, to exclude volcanic eruptions (Kelud, Calbuco, Ambae and Ulawun)
162 and Australian 2019/2020 biomass burning episode (the Black Summer).

163 The MERRA-2 Stratospheric Composition Reanalysis of Aura MLS (M2-SCREAM) products are
164 used for characterizing the water vapor (WV) and ozone vertical distribution, in particular the 3D,
165 3-hourly GMAO_M2SCREAM_INST3_CHEM product available until 1 April 2023 (doi:
166 10.5067/7PR3XRD6Q3NQ). This product, produced at NASA's Global Modeling and
167 Assimilation Office (GMAO), is generated by assimilating MLS (Microwave Limb Sounder) and
168 OMI (Ozone Monitoring Instrument) retrievals into the GEOS (Goddard Earth Observing System)
169 Constituent Data Assimilation System (CoDAS) driven by meteorological fields from MERRA-2.
170 Stratospheric water vapor and ozone, among other compounds, are assimilated in M2-SCREAM.
171 Assimilated fields are provided globally at 0.5° (latitude) by 0.625° (longitude) resolution from
172 approximately 10 km up to the lower thermosphere. Concretely, the variables of specific humidity
173 (QV, kg kg^{-1}), ozone (O3, ppmv), mid-layer pressure (PL, Pa) and mid-layer height (H, m) were
174 used. The specific humidity was converted to the actual water vapor pressure and then to water
175 vapor mixing ratio. All variables were averaged over four pixels surrounding the Maïdo
176 coordinates. Assimilation uncertainties for each of the assimilated constituents are calculated from
177 the CoDAS statistical output (Wargan et al., 2023). For the period January 2022 to September
178 2022 and in the height interval of interest of this study (17 – 32 km) the uncertainty on the water
179 vapor and ozone are less than 0.2 and 0.13 ppmv (Wargan et al., 2023), respectively.

180 MLS version 5.0, level 3 data are also used to extract the monthly mean water vapor over our site
181 and in the stratosphere during 2021 to serve as a climatological reference. See
182 https://mls.jpl.nasa.gov/data/v5-0_data_quality_document.pdf for more details about this MLS
183 product. The monthly mean of the water vapor in the altitude range of interest in 2021 is 4.5 ppmv.
184 This value sets the climatological reference necessary to parametrize the unperturbed conditions
185 of the water vapor.

186 2.3 The GAME radiative transfer model: code and parametrization

187 Radiative fluxes propagating through the atmosphere were calculated with the radiative transfer
188 (RT) model GAME (Dubuisson et al., 1996; Dubuisson, 2004; Dubuisson et al., 2006). For this
189 study, GAME was set up to calculate spectrally integrated upward and downward radiative fluxes
190 in 40 plane and homogeneous layers from 0 to 100 km with a 1 km resolution from 0 to 30 km and
191 a coarser resolution above. The shortwave (SW) spectral range was set from 0.2 to $4.0 \mu\text{m}$ (wave
192 number resolution of 400 cm^{-1} from 0.2 to $0.7 \mu\text{m}$ and 100 cm^{-1} from 0.7 to $4.0 \mu\text{m}$). In the
193 longwave (LW) spectral range, spectral limits were defined between 4.0 and $50.0 \mu\text{m}$ (115 points
194 at a wave number resolution of 20 cm^{-1}). GAME calculates solar flux values at the boundary of
195 plane and homogenous atmospheric layers by using the discrete ordinates method (Stamnes et al.,
196 1988). Gas (H_2O , CO_2 , O_3 , N_2O , CO , CH_4 , and N_2 are considered) absorption is calculated from
197 the correlated k distribution (Lacis & Oinas, 1991). More details about the computation of the gas
198 transmission functions can be found in (Dubuisson, 2004) and (Sicard et al., 2014). In the
199 longwave spectral range, GAME presents the advantage of the complete representation of the long-
200 wave aerosol scattering, in addition to their absorption (Sicard et al., 2014).

201 For the sake of clarity and comparability with other works, we recall the definition of the direct
 202 radiative effect (*DRE*) of a perturbed vs. unperturbed atmospheric compound on the Earth-
 203 Atmosphere radiative budget. At a given height level, L :

$$204 \quad DRE(L) = [F_p^\downarrow(L) - F_p^\uparrow(L)] - [F_u^\downarrow(L) - F_u^\uparrow(L)] \quad (1)$$

205 where F are the radiative flux values for the perturbed (p subindex) and unperturbed (u subindex),
 206 while the \downarrow and \uparrow arrows indicate, respectively, the downward and upward flux direction. By that
 207 definition, negative (positive) *DRE* values represent a cooling (warming) effect. The *DRE* was
 208 calculated at two climate-relevant altitude levels: at the top of atmosphere (TOA) and at the bottom
 209 of atmosphere (BOA). The contribution in the atmospheric column is quantified by the
 210 atmospheric direct radiative effect, *DRE(ATM)*, which is defined as follows:

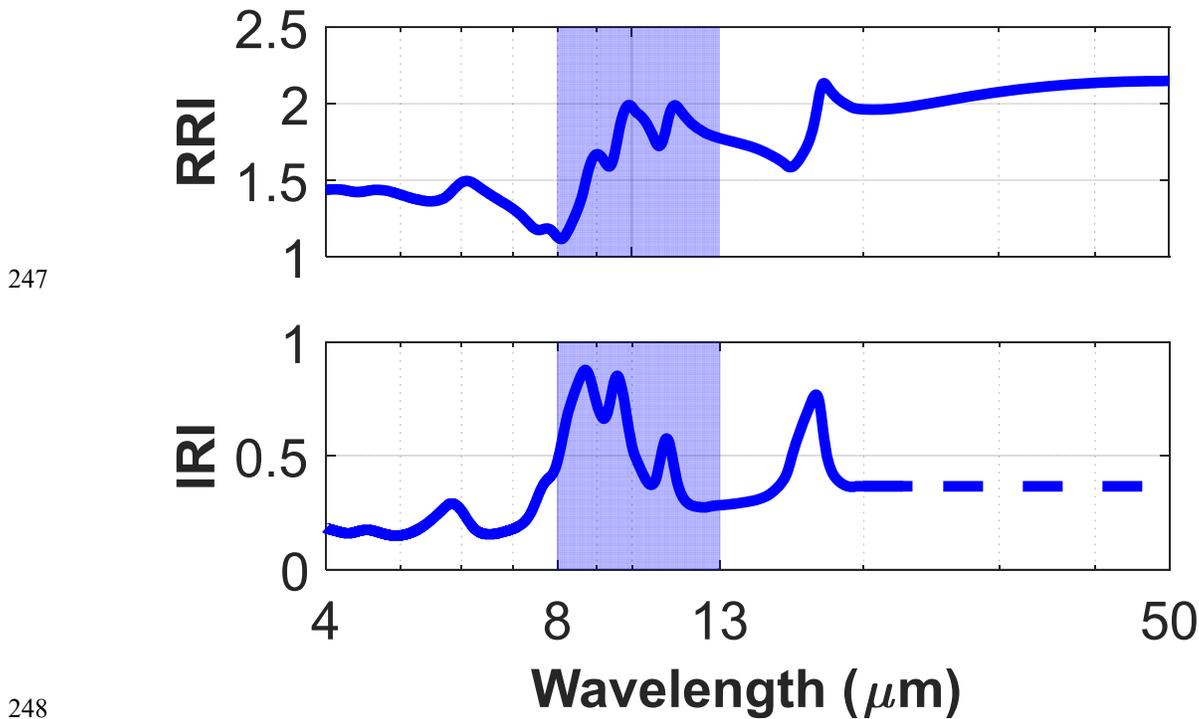
$$211 \quad DRE(ATM) = DRE(TOA) - DRE(BOA) \quad (2)$$

212 As far as GAME parametrization is concerned, temperature and pressure profiles used in both SW
 213 LW simulations are taken from radiosoundings launched from Saint-Denis, the state capital of
 214 Reunion Island, 20 km North of Maïdo, every night at 00:00 Local Time. Aerosols are fully
 215 parameterized in GAME by the user in terms of spectrally and vertically resolved aerosol optical
 216 depth (AOD), single scattering albedo (SSA), and asymmetry factor (asyF).

217 In the SW spectral range, the spectral sAOD was calculated with the lidar extinction coefficients
 218 at 355 nm and $AE_{355/532}$. Spectral SSA and asyF were interpolated from the four-wavelength
 219 AERONET L1.5 monthly means and assumed constant above 1020 nm. The spectral surface
 220 albedo was interpolated from the four-wavelength AERONET L2.0 annual (2022) mean and
 221 assumed constant above 1020 nm.

222 In the LW spectral range, a Mie code was used to calculate the LW spectral AOD, SSA and asyF.
 223 The bimodal, lognormal size distribution was considered. Geometric median radii and standard
 224 deviations were calculated from AERONET L1.5 fine and coarse mode volume median radii and
 225 standard deviations applying Eq. (A2) and (A3) of (Sicard et al., 2014). For the refractive index
 226 we used the GEISA (Gestion et Étude des Informations Spectroscopiques Atmosphériques:
 227 Management and Study of Spectroscopic Information) spectroscopic database (Jacquinet-Husson
 228 et al., 2008). In particular the refractive index of the binary system H_2SO_4/H_2O with a H_2SO_4
 229 mixing ratios (in mass, i.e., the ratio of the H_2SO_4 mass to the total mass of the droplets) of 75 %
 230 and at temperature of 215 K (this temperature corresponds in average to the atmospheric
 231 temperature at the height of the volcanic plume) was used. For comparison, (Bernath et al., 2023),
 232 who analyzed measurements of atmospheric infrared transmittance of the HTHH sulfate aerosol
 233 plume by the Atmospheric Chemistry Experiment satellite 20 days after the eruption, found a
 234 H_2SO_4 mixing ratio of 62.5 %. The real part is defined over the range 0.61 – 5000.00 μm (wave
 235 number resolution of 2 cm^{-1}), while the imaginary part is defined over the range 2.36 – 23.15 μm
 236 (wave number resolution of 0.96 cm^{-1}). The reader is referred to (Biermann et al., 2000) for more
 237 details on this dataset. Figure 1 shows the real part (RRI) and imaginary part (IRI) of the refractive
 238 index used. Large spectral variations in the infrared atmospheric window (8–13 μm), which have
 239 an important impact on the infrared radiative budget of the atmosphere, are visible. The most
 240 astonishing feature of the figure is probably the high absolute values of the IRI which emphasizes
 241 the high absorbing properties of sulfate aerosols in the longwave spectral range. For comparison
 242 IRI (this study) is 2 to 3 times larger in the atmospheric window than IRI for mineral dust (Sicard
 243 et al., 2014). Finally, for the LW broadband surface albedo, we used IASI (Infrared Atmospheric

244 Sounding Interferometer) December nighttime monthly mean climatology of the surface
 245 emissivity (i.e. $1 - \text{surface albedo}$) at 890 cm^{-1} , i.e. $11.24 \mu\text{m}$, over the Indian Ocean (Zhou et al.,
 246 2013), and set the surface albedo value to 0.01.



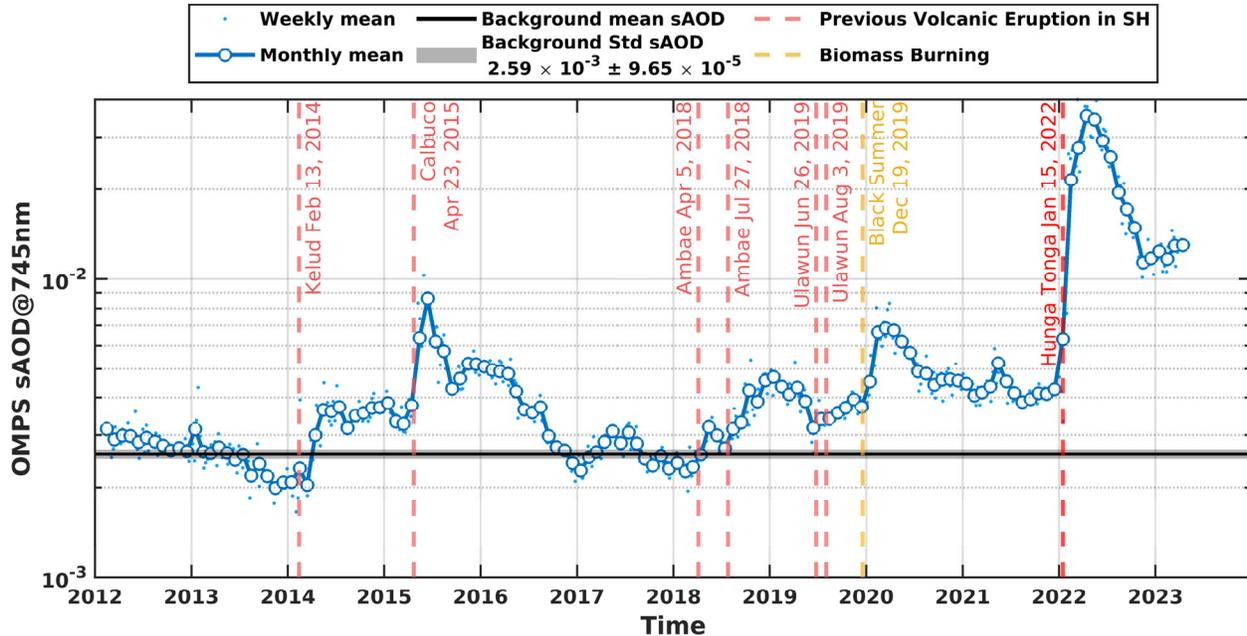
249 **Figure 1.** Spectral complex refractive index considered for the calculation of the aerosol radiative
 250 properties in the longwave spectral range. See text for details. The infrared atmospheric window
 251 (8–13 μm) is indicated by the blue shaded area. The dash line in IRI is the extrapolation of the
 252 dataset used up to 50 μm .

253 So as to avoid the dependency on solar zenith angle, only daily radiative effects are presented in
 254 this work. To do so, each nighttime measurement and parametrization is assumed to be constant
 255 for the 24 hours of the day considered and both SW and LW radiative effects are calculated at an
 256 hourly time resolution between 00:00 and 23:00 UT. In these calculations, the solar zenith angle
 257 is the only parameter that varies. The daily radiative effect is the average of the 24 hourly *DRE*.

258 **3 Vertical/temporal evolution of the HTHH volcanic plume over Reunion Island**

259 The historical context of the aerosol load over Reunion Island is shown by the temporal evolution
 260 of the stratospheric AOD at 745 nm measured by OMPS in the last decade (Figure 2). The
 261 background sAOD is measured over the unperturbed years 2012 and 2013. It is $(2.59 \pm 0.10) \times 10^{-3}$.
 262 At each exceptional event the sAOD takes off from this background sAOD and since the eruption
 263 of Ambae in July 2018 the sAOD over Reunion Island has never turned back to its background
 264 value. The sAOD peak produced by HTHH (0.035) is the highest in the last decade and it is a
 265 factor 4 times higher than the second highest event (0.009, Calbuco eruption in April 2015). Zonal
 266 averages between 30°S and 15°N for HTHH and 20°S and 90°S for Calbuco showed that HTHH
 267 sAOD was more than double that for the 2015 Calbuco eruption (Taha et al., 2022). The reason
 268 why the local and zonal sAOD differences between HTHH and Calbuco differ lies in the zonal
 269 mean stratospheric conditions. In the case of HTHH, a marked easterly band (Khaykin et al., 2022;

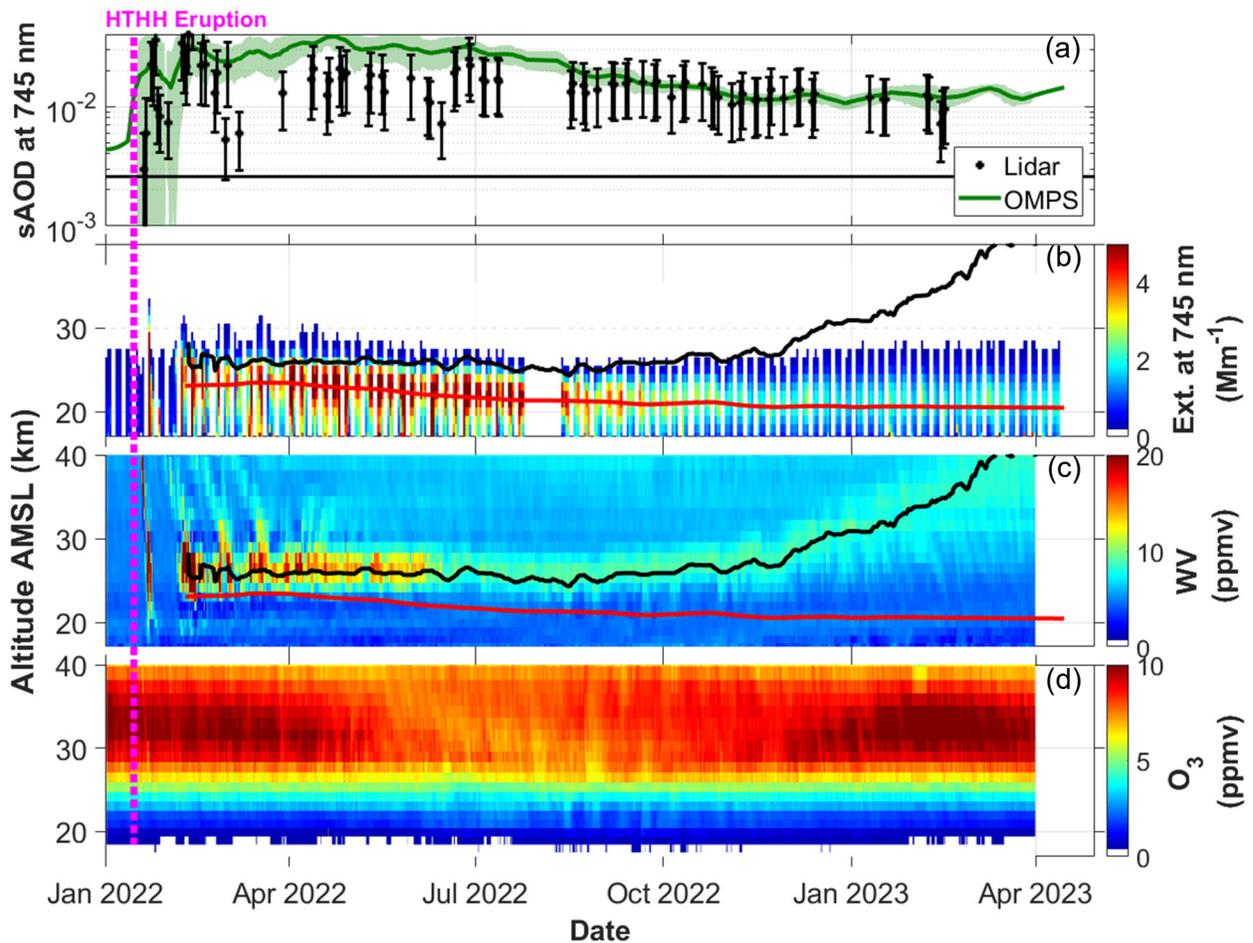
270 Legras et al., 2022) favored a direct transport from HTHH towards Reunion Island (both being
 271 approximately at the same latitude). Further back historically, the 40+ year satellite record of
 272 monthly sAOD for the 60°S – 60°N latitude band in (Khaykin et al., 2022) shows that only the
 273 eruptions of Pinatubo (1991) and El Chichón (1982) exceeded the HTHH one in terms of absolute
 274 stratospheric AOD (by a factor of 6 and 3, respectively).



275
 276 **Figure 2.** sAOD (17 – 40 km) at 745 nm from OMPS over Reunion Island. The most important
 277 volcanic eruptions (name and date) in the southern hemisphere are indicated by red vertical lines.
 278 The Australian 2019/2020 biomass burning episode is indicated in orange.

279 The vertical and temporal (January 2022 – April 2023) evolution of the HTHH volcanic plume
 280 over Reunion Island is analyzed by means of sAOD at 745 nm and profiles of extinction coefficient
 281 at 745 nm, water vapor and ozone (Figure 3). While the monthly OMPS sAOD peak is reached in
 282 April and May 2022, the instantaneous lidar sAOD peaks just a few days after the eruption,
 283 reaching 0.54 on 21 January. This time difference is an indication of the dilution time of the
 284 volcanic matter injected by HTHH in the stratosphere at the global scale. Other studies confirm
 285 that the volcanic plume dispersed nearly pole-to-pole in three months (Khaykin et al., 2022; Taha
 286 et al., 2022). Another indicator of this dilution is the standard deviation (calculated as a 15-day
 287 rolling standard deviation) associated to OMPS monthly sAOD: once passed the first month, it
 288 steadily decreases all along year 2022. The agreement between monthly and instantaneous sAOD
 289 which becomes excellent as of September 2022 is also an indicator of the homogenous dilution of
 290 the volcanic plume in the stratosphere. It also reinforces our belief that these results in Reunion
 291 Island could probably be generalized throughout the southern tropical Indian Ocean region. A
 292 decrease of the monthly sAOD is observed after April/May and until November. Then sAOD
 293 stabilizes until today (sAOD = 0.012, almost 5 times the background sAOD). The Ångström
 294 exponent calculated between the lidar-derived sAOD at 355 and 532 nm, $AE_{355/532}$, has near-zero
 295 values during the first two weeks after the eruption (Baron et al., 2023), indicating a probable
 296 mixing of fine-mode sulfate and coarse-mode non-sulfate (ash/ice particles) aerosols as suggested
 297 by (Sellitto et al., 2022). After one month (not shown) $AE_{355/532}$ reaches values above 1.3 that
 298 stabilize rapidly at 1.5.

299 The time-height plot of the extinction coefficient (Figure 3b) shows clearly the height and vertical
 300 extension of the volcanic plume which is still present on 15 April 2023 and located at 18.5 – 23.5
 301 km height (sAOD = 0.012). The plume peak height has a decreasing tendency since April 2022 at
 302 an average steady rate of -244 m per month or ~ -0.008 km day⁻¹. Assuming this rate constant in
 303 time and a tropopause height in Reunion Island of 17 km (Bègue et al., 2010), the remaining life
 304 time of the volcanic plume in the stratosphere is estimated to be between 2 and 2.5 years after 15
 305 April 2023. Except during the first week of detection above Reunion Island, the HTHH volcanic
 306 plume is not detected above 30 km. The water vapor plume (Figure 3c) reveals also clearly the
 307 unusually high water vapor concentration caused by the volcanic plume. A local peak of 65 ppmv
 308 is reached on 13 February 2022. It is almost 15 times higher than the climatological reference
 309 value of 4.5 ppmv (see Section 2.2). On a monthly basis, the water vapor stratospheric peak in
 310 February 2022 is approximately 5 times higher than the climatological reference (4.5 ppmv). This
 311 ratio decreases to almost 2 in February 2023. The water vapor plume is thinner than the aerosol
 312 one and located at a higher altitude, 3 to 4 km higher. Such a difference is observed at the zonal



313

314 **Figure 3.** (a) Monthly (OMPS) and instantaneous (lidar, nighttime) sAOD (17 – 40 km) at 745 nm
 315 over Reunion Island; Time-height plots of (b) extinction coefficient at 745 nm from OMPS, (c)
 316 water vapor and (d) ozone mixing ratio from M2-SCREAM. The red and black lines in (b) and (c)
 317 report the peak height of the aerosol and water vapor plumes, respectively.

318 scale (at 15° S), although not so accentuated, during the first six months of year 2022 (Schoeberl
 319 et al., 2022). The height of the peak of the aerosol and water vapor layers (respectively, red and
 320 black lines in Figure 3b and Figure 3c) have opposite tendencies as of April 2022: the aerosol
 321 plume is slowly descending whereas the moist layer is ascending slowly until October 2022 and
 322 at a higher rate afterwards. (Schoeberl et al., 2022) explain that “the water vapor is transported
 323 upward with the diabatic circulation that gives rise to the tropical trace gas tape recorders
 324 (Schoeberl et al., 2018) whereas the aerosols are gravitationally settling”. (Legras et al., 2022) who
 325 analyze the same period (first six months of year 2022) precise that the ascent of the moist layer
 326 is due to the Brewer–Dobson circulation. The ozone cycle (Figure 3d) with highs in the austral
 327 summer (January-April) and lows in the austral winter (July-October) reflects the higher
 328 production of ozone in summer due to the peak of solar radiation compared to winter
 329 (Abdoulwahab, 2016). Apart from this natural cycle of stratospheric ozone at subtropical latitudes,
 330 no other spatio-temporal variation, potentially caused by HTHH eruption, is visible at naked eye
 331 in Figure 3d. Some authors mention that, following HTHH eruption, ozone concentrations increase
 332 in the middle stratosphere and decrease in the lower stratosphere were caused by enhanced sulfate
 333 aerosol (Lu et al., 2023), others claimed that the midlatitude and tropics ozone reduction observed
 334 by MLS was mainly linked to circulation effects (Wang et al., 2022). At this early stage of our
 335 understanding of the effects of HTHH on the stratospheric ozone, the present study does not
 336 consider any potential increase/decrease of stratospheric ozone due to HTHH eruption.

337 **4 Impact of the HTHH volcanic plume on the Earth’s radiation budget**

338 In order to analyze the radiative impact of the aerosols and the water vapor, separately and
 339 altogether, 4 runs of GAME are performed and summarized in Table 1. The perturbed condition
 340 is the full parametrization with observed sAOD and water vapor mixing ratio. For the unperturbed
 341 conditions, the impact of aerosols is assessed by assuming an aerosol-free stratosphere; the impact
 342 of water vapor is assessed by assuming that the water vapor mixing ratio in the HTHH moist layer
 343 is equal to the climatological value of 4.5 ppmv obtained from the MLS 2021 monthly means; the
 344 impact of aerosols and water vapor is assessed by assuming both an aerosol-free stratosphere and
 345 a water vapor mixing ratio of 4.5 ppmv in the HTHH moist layer.

Impact of...	Perturbed	Unperturbed
Aerosols	Measured sAOD Measured WV	sAOD = 0 Measured WV
Water vapor		Measured sAOD WV = 4.5 ppmv above 20 km (climatology from MLS monthly means in 2021)
Aerosol and water vapor		sAOD = 0 WV = 4.5 ppmv above 20 km (climatology from MLS monthly means in 2021)

346 **Table 1.** Aerosol and water vapor parametrization for the perturbed/unperturbed simulations of
 347 GAME.

348 Figure 4, Figure 5 and Figure 6 show the radiative impact of aerosols only, of the water vapor
349 only, and of both aerosols and water vapor, respectively, in terms of time plots of $DRE(BOA)$ and
350 $DRE(TOA)$ as well as time-height plots of heating/cooling (H/C) rate anomaly in the SW, LW and
351 SW+LW spectral ranges.

352 In the first four months after the eruption, the aerosol LW dominates over the SW component at
353 TOA (Figure 4a), producing a net warming (positive) radiative effect. After May 2022, both
354 components roughly compensate: $|DRE(TOA)| < 0.6 \text{ W m}^{-2}$. Interestingly, during the first week
355 $DRE(TOA)$ is negative. It presents a peak at -13.8 W m^{-2} which falls in the range of instantaneous
356 values found by (Sellitto et al., 2022) for the fresh plume (-19.4 and -12.1 W m^{-2}). This change of
357 sign is attributed to a combination of facts: the high sAOD measured over Reunion Island during
358 the first week after the eruption (Baron et al., 2023) and smaller SW asymmetry factors for the
359 month of January 2022 (implying an increase of the solar radiation reflected back to space). We
360 also observed a smaller fine mode radius in January 2022 compared to the rest of the period, also
361 evidenced by (Boichu et al., 2023) over Reunion Island, but this is not expected to affect
362 significantly the LW radiative properties (Sicard et al., 2014). (Zhu et al., 2022) who use a global
363 climate model to simulate the radiative effect in the first two months of 2022 find that when only
364 volcanic sulfur dioxide is present (without water vapor) the zonal $DRE(TOA)$ at the latitude of
365 Reunion Island is negative. This is opposite of our results, but somehow expected since their model
366 underestimates the production of sulfate particles (i.e., it underestimates sAOD) when water
367 injection is not considered. At BOA the aerosol LW component is nearly zero, so that the aerosol
368 net DRE is that of the SW component: a cooling is observed. A peak is observed during the first
369 week after the eruption at -26.8 W m^{-2} and again it falls in the range of instantaneous values found
370 by (Sellitto et al., 2022) for the fresh plume (-27.9 and -17.5 W m^{-2}). As a consequence of
371 $DRE(TOA)$ and $DRE(BOA)$, and as illustrated by the time-height plots of H/C rates (Figure 4d),
372 the atmosphere warms, mostly because of the aerosol LW warming. The SW+LW aerosol impact
373 in the stratosphere is mostly positive with daily heating rates of $\sim +1.00 \text{ }^\circ\text{K day}^{-1}$ in the first 4-6
374 months, and $\sim +0.06 - +0.15 \text{ }^\circ\text{K day}^{-1}$ afterwards. The H/C rate profiles follow the shape of the
375 aerosol extinction profiles with a decreasing tendency starting in April 2022 (Figure 3b). In
376 conclusion, the effect caused by aerosols only is what is usually expected for stratospheric sulfate
377 aerosols (Bernath et al., 2023): HTHH stratospheric particles measured over Reunion Island scatter
378 sunlight back to space, cooling the Earth's surface and absorb outgoing longwave radiation,
379 heating the stratosphere.

380 The water vapor radiative effect (Figure 5) is dominated by the cooling effect of water vapor
381 longwave emission in the moist layer. This layer produces a slightly negative effect at TOA
382 ($|DRE(TOA)| < 0.3 \text{ W m}^{-2}$) and a neutral effect ($\sim 0 \text{ W m}^{-2}$) at BOA. (Sellitto et al., 2022)
383 estimated also a negative $DRE(TOA)$ caused by water vapor for the fresh plume (instantaneous
384 values of -0.7 and -0.4 W m^{-2}), but a positive $DRE(TOA)$ caused by water vapor for the aged
385 plume (8 February 2022) of $+0.8 \text{ W m}^{-2}$ and attributed to the descent in altitude of the moist layer.
386 Our analysis does not support this change of sign. (Zhu et al., 2022) find that when only water
387 vapor is injected in their model (without sulfur dioxide; see their supplementary material) the zonal
388 $DRE(TOA)$ at the latitude of Reunion Island is positive and much smaller than that caused by
389 aerosols. The discrepancy with our findings stems from an excess of water vapor in (Zhu et al.,
390 2022) simulations since the reaction of sulfur dioxide (not present) and hydroxide is not happening
391 in their simulation; different heights of the moist layer; and/or zonal vs. local computations. The
392 SW+LW water vapor radiative impact in the stratosphere is mostly negative with daily cooling

393 rates of $\sim -0.20 - -0.40$ °K day⁻¹ in the first 4-6 months, and $\sim -0.06 - -0.15$ °K day⁻¹ afterwards.
 394 The same ascending behavior of the water vapor concentration (Figure 3c) is observed on the
 395 profiles of the LW water vapor cooling rate (Figure 5c). (Schoeberl et al., 2022), who estimated
 396 the LW zonal impact of water vapor at 15° South for the first 6 months of year 2022, show a
 397 cooling effect in the stratosphere with a peak at ~ -0.5 °K day⁻¹ at the end of February and
 398 decreasing afterwards. (Sellitto et al., 2022) calculated instantaneous cooling rates peaking
 399 between -4.0 and -10 °K day⁻¹.

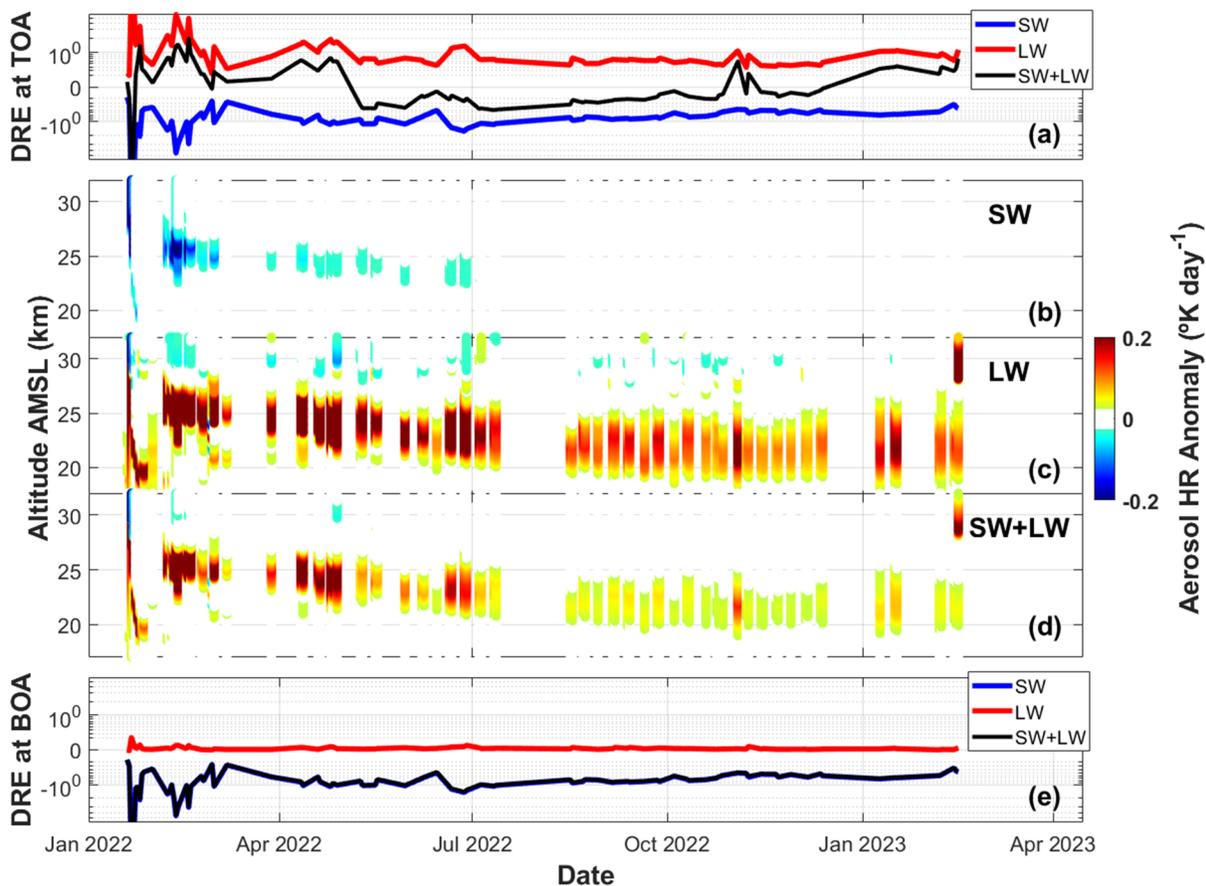
400 When aerosols and water vapor HTHH perturbations are considered together (Figure 6), the
 401 radiative effect both at TOA and BOA is similar to that of the aerosols only (Figure 4): positive at
 402 TOA until April 2022 and switching to negative-to-neutral afterwards, and negative at BOA. These
 403 effects (warming/cooling/neutral) are in agreement with the estimations made by (Sellitto et al.,
 404 2022) for what they call the “aged plume”. In a third simulation performed by (Zhu et al., 2022)
 405 where both sulfur dioxide and water vapor are injected, these authors find a negative radiative
 406 effect at both TOA and BOA which, averaged globally and limited to the first 2 months of 2022,
 407 are equal to -0.21 and -0.21 W m⁻², respectively. The discrepancy with our results at TOA seems
 408 to speak against the different methodologies used, and in particular to the amount of volcanic
 409 aerosols used in both works: a function of the water vapor injected (Zhu et al., 2022) and measured
 410 by lidar (this study). Our results imply that the missing energy at the surface is trapped in the
 411 stratospheric volcanic layer which warms: the total atmospheric radiative budget is positive and
 412 generally exceeds $+1.0$ W m⁻² in the first 4-6 months after the eruption and is on the order of $+0.4$
 413 $- +1.0$ W m⁻² afterwards, the causing agent being the positive aerosol LW *DRE(ATM)* (nearly 10
 414 times larger than the WV component of opposite sign) reflecting the strong aerosol LW warming
 415 w.r.t. a moderate water vapor LW cooling.

416 An interesting result at this point is how the aerosols and water vapor H/C rates distribute vertically
 417 in the atmosphere. It is clear from Figure 6d that the longwave cooling caused by water vapor and
 418 warming caused by the aerosols coexist at different altitude levels. To further discuss this issue,
 419 we analyze three different periods of time, excluding from now on the first two weeks after the
 420 eruption to allow for some dilution to happen:

- 421 • The entire period from Feb. 2022 to Feb. 2023 (M2 – M13, M1 being Jan. 2022).
- 422 • Feb. 2022 – Apr. 2022 (M2 – M4), short-term period: aerosol and WV *DRE(TOA)* is
 423 positive.
- 424 • May 2022 – Feb. 2023 (M5 – M13), mid-term period: aerosol and WV *DRE(TOA)* is
 425 negative-to-neutral.

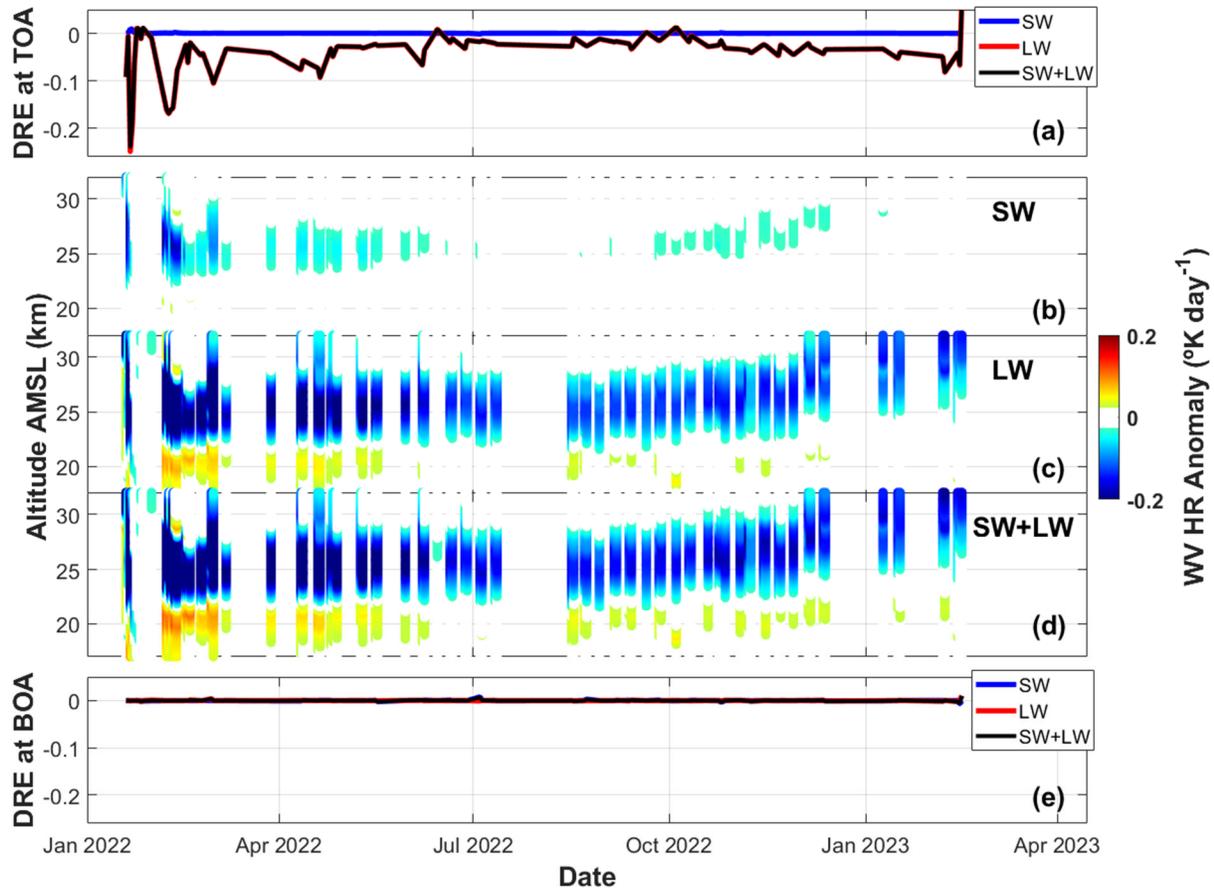
426 The first period is representative of the radiative impact of HTHH since the eruption to date, while
 427 the second and third periods are representative of the short- and mid-term tendencies, respectively.
 428 Figure 7 shows the averaged H/C rate anomaly profiles over the three periods for the last
 429 simulation performed (aerosols and water vapor). The radiative effects of the three simulations
 430 (aerosols only, WV only, aerosols and WV) associated to the three periods at the three atmospheric
 431 level (BOA, ATM, TOA) are summarized in Table 2. Independently of the period considered, the
 432 WV cooling peak, related to unusually high WV concentration at that altitude in the first 4-6
 433 months after the eruption (see Figure 3c), occurs at 27 km. The aerosol positive peak occurs at 25
 434 km during M2 – M4 and drops to 22 km during M5 – M13. The change of sign of the profiles
 435 occurs at 26 km for the entire period, while it is a little higher ($\sim 26-27$ km) during M2 – M4 and
 436 lower ($\sim 24-25$ km) during M5 – M13, in agreement with the descent of the volcanic aerosol layer.

437 At the southern hemisphere scale (60°S – 10°S), (Wang et al., 2022) estimated a stratospheric
 438 temperature anomaly from MLS measurements at 25 hPa (~ 25 km at the latitude of Reunion
 439 Island) of ~ -1°K on 1 November, 2022. This negative temperature anomaly seems in agreement
 440 with the sign (negative) of the heating/cooling rate profile observed at 25 km for the mid-term
 441 period M5 – M13. Interestingly, note the effect of the rapid vertical displacements of
 442 positive/negative rate anomalies (Figure 6d) in the short-term period which, once averaged,
 443 compensate at 23 and 24 km and thus result in a neutral (zero) H/C rate at these altitudes (Figure
 444 7). In terms of shape and magnitude, the short-term period M2 – M4 exhibits marked and steep
 445 variations with positive and negative peaks of similar intensity (0.27 °K day⁻¹ in absolute value).
 446 During the mid-term period (M5 – M13) the homogeneous dilution of the plume makes the H/C
 447 rate vertical distribution much smoother with also comparable positive and negative peaks at ~0.13
 448 °K day⁻¹ in absolute value. However, because the WV cooling occurs at higher altitudes than the
 449 aerosol warming, what drives the stratospheric radiative effect is that of the aerosols (warming):
 450 $DRE(ATM) = +1.79 \pm 1.43 \text{ W m}^{-2}$ for the short-term period (M2 – M4) and $+0.63 \pm 0.22 \text{ W m}^{-2}$
 451 for the mid-term period (M5 – M13). It is $+0.97 \pm 0.95 \text{ W m}^{-2}$ over the first thirteen months after
 452 the eruption. We can state without any doubt that this warming effect would have been reduced to
 453 a likely neutral effect ($DRE(ATM) \sim 0 \text{ W m}^{-2}$) if aerosols and water vapor had been present at the
 454 same altitude levels.



455

456 **Figure 4.** Aerosol direct radiative impact. (a) Radiative effect (W m^{-2}) at TOA; Time-height
 457 evolution of the H/C rate anomaly in the (b) SW; (c) LW; and (d) SW+LW spectral ranges; (e)
 458 Radiative effect (W m^{-2}) at BOA.

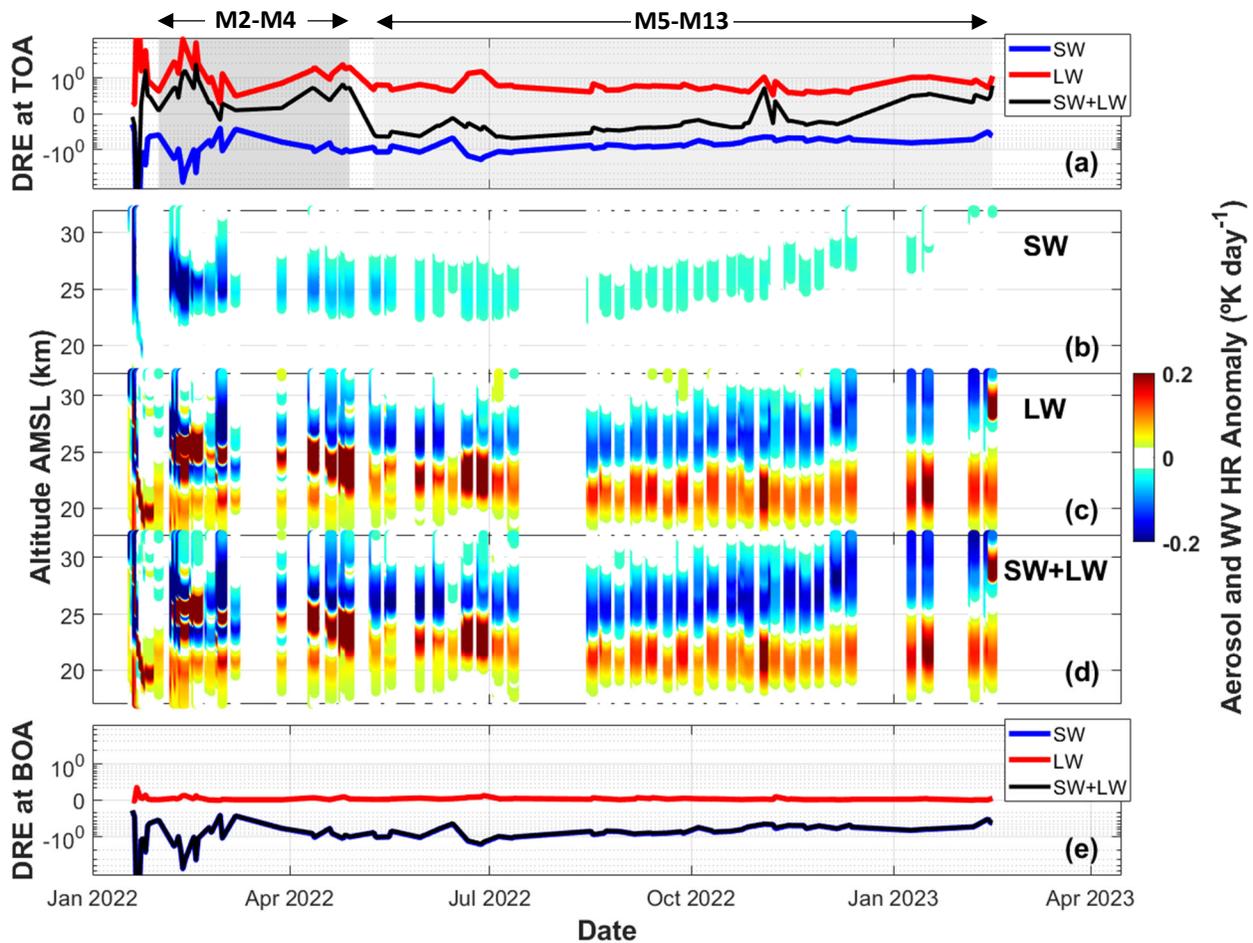


459

460 **Figure 5.** Idem as Figure 4 for the water vapor. Note that the left axis scale in (a) and (e) is different
 461 from that of Figure 4.

462 Finally, the overall HTHH aerosol and water vapor impact on the Earth's radiation budget (Table
 463 2) is positive (warming) for the first thirteen months after the eruption: $DRE(TOA) = +0.06 \pm 0.45$
 464 $W m^{-2}$. The aerosol warming dominates, so far, for this period. However, the breakdown in short-
 465 and mid-term tendencies shows that the predominance of the aerosols at TOA decreases with time:
 466 $DRE(TOA)$ is strongly positive ($+0.53 \pm 0.48 W m^{-2}$) during M2 – M4 and switches sign ($-0.13 \pm$
 467 $0.26 W m^{-2}$) during M5 – M13. Given the large standard deviations associated, these numbers
 468 should be taken with caution. Several facts and hypothesis support this evolution. By looking back
 469 to Figure 3a, one sees that, between April and November 2022, sAOD is quantitatively decreasing
 470 with time faster than the slow decaying stratospheric water vapor (Khaykin et al., 2022). In
 471 addition, the volcanic aerosol plume decreases in height while the moist layer ascends. Both effects
 472 reinforce the radiative efficiency at TOA of water vapor against volcanic aerosols, and are part of
 473 the reason why $DRE(TOA)$ changes sign from positive to negative. In the longer term,
 474 $DRE(TOA)$, averaged over the whole post-eruption period, might reduce to quasi-neutral, or even
 475 become slightly negative. At the global scale, (Zhu et al., 2022), mentioned earlier, find a
 476 $DRE(TOA)$ of $-0.21 W m^{-2}$ for the first two months of 2022; their estimation is of opposite sign
 477 compared to this study (see two paragraphs above for explanation). At the surface, the results are
 478 more conclusive: a cooling produced by HTHH volcanic aerosols is observed ($DRE(BOA) = -0.91$

479



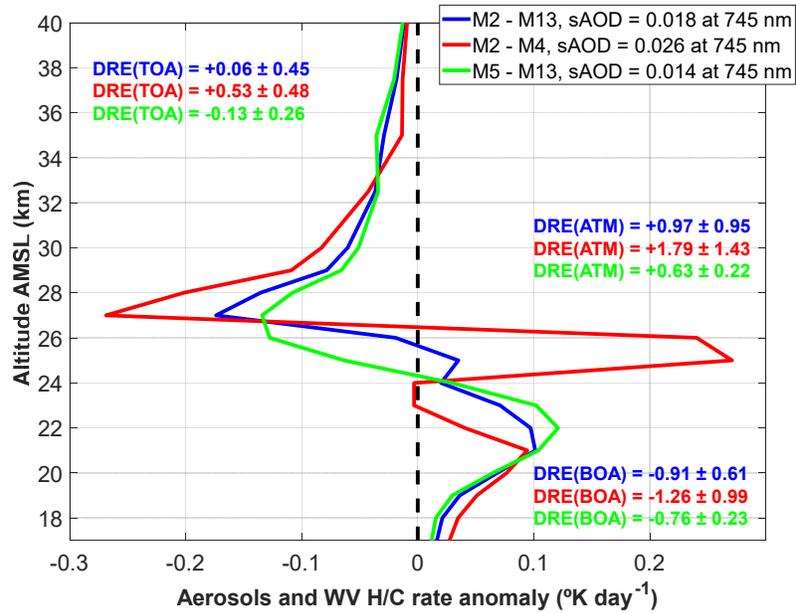
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481 **Figure 6.** Idem as Figure 4 for the aerosols and water vapor. In (a), the gray shaded areas represent
 482 the short- and mid-term periods: February to April 2022 (M2 – M4) and May 2022 to February
 483 2023 (M5 – M13).

484 $\pm 0.61 \text{ W m}^{-2}$ for the entire period), with a decreasing tendency with time ($DRE(BOA) = -1.26 \pm$
 485 0.99 W m^{-2} for M2 – M4 and $-0.76 \pm 0.23 \text{ W m}^{-2}$ for M5 – M13). (Zuo et al., 2022) modeled the
 486 global surface temperature in the first year after the HTHH eruption and found a negative anomaly
 487 of $-0.004 \text{ }^\circ\text{K}$ but recognized that it is “within the amplitude of internal variability at the interannual
 488 time scale and thus not strong enough to have significant impacts on the global climate”. In
 489 contrast, by extension of our results to the southern tropical Indian Ocean region, our analysis
 490 shows that the eruption of HTHH might have had a clear cooling impact on the regional climate
 491 at the surface in this region of the Earth.

492 5 Conclusions

493 Thirteen months after the eruption of HTHH volcano, aerosols and water vapor are still present in
 494 the stratosphere of the southern tropical Indian Ocean region. During the first three months after
 495 the eruption the stratospheric aerosol optical depth increases and reaches a peak at 0.035 (~13
 496 times the background sAOD) in April 2022, the highest in the last decade, and the third highest in
 497 the last 40 years (after Pinatubo and El Chichón). From April to November 2022 the sAOD
 498 decreases and then stabilizes at a value of 0.012 (~5 times the background sAOD). Unusually high
 499 water vapor concentrations are also observed in the stratosphere. On a monthly basis, the water



500

501 **Figure 7.** Profiles of aerosol and water vapor daily heating/cooling rate anomaly averaged over
 502 the entire period (M2 – M13), the short-term period (M2 – M4) and the mid-term period (M5 –
 503 M13), all excluding week 1 and 2 after the eruption.

504

	Aerosols	Water vapor	Aerosols and water vapor
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Daily DRE (W m^{-2}) for the entire period M2 - M13

TOA	$+0.10 \pm 0.47$	-0.04 ± 0.04	$+0.06 \pm 0.45$
ATM	$+1.01 \pm 0.97$	-0.04 ± 0.04	$+0.97 \pm 0.95$
BOA	-0.90 ± 0.61	< 0.01	-0.91 ± 0.61

Daily DRE (W m^{-2}) for the short-term period M2 – M4

TOA	$+0.60 \pm 0.49$	-0.07 ± 0.05	$+0.53 \pm 0.48$
ATM	$+1.86 \pm 1.44$	-0.07 ± 0.05	$+1.79 \pm 1.43$
BOA	-1.26 ± 0.99	< 0.01	-1.26 ± 0.99

Daily DRE (W m^{-2}) for the mid-term period M5 – M13

TOA	-0.11 ± 0.27	-0.03 ± 0.02	-0.13 ± 0.26
ATM	$+0.65 \pm 0.22$	-0.03 ± 0.02	$+0.63 \pm 0.22$
BOA	-0.76 ± 0.23	< 0.01	-0.76 ± 0.23

505 **Table 2.** SW+LW DRE at BOA, TOA and in the atmosphere produced by aerosols only, WV only
 506 and both aerosols and WV. These values are the average over the entire period M2 – M13, the
 507 short-term period M2 – M4, and the mid-term period M5 – M13, all excluding week 1 and 2 after
 508 the eruption.

509 vapor stratospheric peak reaches a maximum in February 2022 which is approximately a factor 5
510 above the climatological reference. In February 2023, this ratio has decreased down to almost 2.

511 In all moments, the water vapor plume is located at a higher altitude than the aerosol plume. The
512 height of the peak of the aerosol and water vapor layers have opposite tendencies as of April 2022:
513 the aerosol plume is slowly descending by gravitational settling whereas the moist layer is
514 ascending slowly until October 2022 and at a higher rate afterwards. The upward transport of the
515 moist layer is due to the Brewer-Dobson circulation. Both aerosol and WV plumes are still present
516 on 15 April 2023. The aerosol plume is located at 18.5 – 23.5 km height and the moist layer is
517 above 30 km. As far as aerosols are concerned, the plume peak height decreases since April 2022
518 at an average steady rate of $\sim -0.008 \text{ km day}^{-1}$. Assuming this rate constant in time, the remaining
519 life time of the volcanic plume in the stratosphere is estimated to be between 2 and 2.5 years after
520 15 April 2023.

521 The radiative impact of both aerosol and water vapor layers is estimated at our site and assumed
522 representative of the southern tropical Indian Ocean. Averages are made over 3 different periods
523 of time in order to explain the temporal evolution: the first thirteen months, the short-term (M2 –
524 M4) and the mid-term (M5 – M13) periods. During the first thirteen months after HTHH eruption,
525 the overall aerosol and water vapor impact on the Earth's radiation budget is positive (warming,
526 $+0.06 \pm 0.45 \text{ W m}^{-2}$) and dominated by the aerosol impact. However, the decreasing rate with time
527 of the aerosol warming effect is larger than that of the water vapor cooling effect, so that, in the
528 long run, the impact on the Earth's radiation budget might reduce to quasi-neutral, or even become
529 slightly negative. At least two factors are at play here: between April and November 2022, sAOD
530 is quantitatively decreasing with time faster than the slow decaying stratospheric water vapor; the
531 volcanic aerosol plume decreases in height while the moist layer ascends. Both effects reinforce
532 the radiative efficiency at TOA of water vapor against volcanic aerosols. At the Earth's surface,
533 aerosols are the main driver and produce a negative (cooling, $-0.91 \pm 0.61 \text{ W m}^{-2}$) radiative impact
534 with also a decreasing tendency with time. Heating/cooling rate profiles show a clear vertical
535 difference in the stratosphere between the aerosol warming impact (17 to 25 km) and the water
536 vapor cooling one (25 to 40 km). Although the aerosol warming and the WV cooling rates are not
537 drastically different in intensity, because the first one occurs at lower altitudes than the second
538 one, what in the end drives the stratospheric radiative effect is that of the aerosols. During the first
539 thirteen months after HTHH eruption, aerosols and water vapor produce a warming of the
540 stratosphere ($+0.97 \pm 0.95 \text{ W m}^{-2}$) with also a decreasing tendency with time. This study shows that
541 the eruption of HTHH has had, so far, a clear impact on the regional climate of the Earth-
542 Atmosphere system in the southern tropical Indian Ocean region.

543 **Acknowledgments**

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546 **Open Research**

547 The lidar data and the parametrization data used in the three simulations of the radiative transfer
548 model will be available by the time of publication. These data will receive a doi from the library of
549 Université de la Réunion that will be cited in the paper and listed in the Reference section.

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