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1 **The effect of the Martian 2018 global dust storm on**
2 **HDO as predicted by a Mars Global Climate Model**

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8 **Key Points:**

- 9 • We reimplemented the HDO cycle in a more recent version of the LMD Mars GCM
10 • We reproduce the observed gradient of the D/H ratio between cold and warm re-
11 gions
12 • The Global Dust Storm of MY34 has a strong effect on the vertical distribution
13 of HDO

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Abstract

The D/H ratio is commonly used to investigate the history of water on Mars, yet the mechanisms controlling present-day HDO behavior are poorly understood. Significant variations of the D/H ratio were first predicted on the basis of a 3D global climate model, which were later confirmed by ground-based observations. This behaviour, consisting of lower HDO/H₂O ratios in the colder regions of Mars, is related to the isotopic fractionation occurring at condensation. We leverage this previous effort and present an updated implementation, using the modern version of the model, that remains in agreement with the older version. We explore the impact of the Global Dust Storm that occurred during Martian Year 34 on HDO. Our simulations indicate that HDO is on average 40% more abundant at 100 km during the MY34 GDS year than during a regular year, with likely large consequences for the escape flux of water that year.

Plain Language Summary

HDO, the semi-heavy isotope of water, when compared to water, is a good indicator of how much water has been escaping from the atmosphere of Mars over the ages. Ultimately, it can be used to estimate the past reservoir of water available on Mars in its early youth. Because HDO has a slightly higher molecular mass compared to H₂O, condensation induces an enrichment of HDO in the ice phase compared with the vapour phase. This subsequently causes spatial and temporal variations of the deuterium to hydrogen ratio (D/H). We use a global circulation model to simulate the HDO cycle in the atmosphere. Our model is an upgrade of the previous model presented in Montmessin et al. (2005). We then simulate the effect of the Global Dust Storm that affected Mars during the summer of 2018, and show that it should have had a strong impact on the vertical distribution of HDO, allowing it to reach higher altitudes. Such simulations are intended to be compared with observations from the Trace Gas Orbiter, currently in orbit around Mars.

1 Introduction

Mars is known to have had a significant liquid water reservoir on the surface (Carr & Head III, 2003; Bibring et al., 2006) and the D/H ratio derived from the HDO/H₂O abundance ratio is a sensitive tool to constrain the primordial abundance of the water reservoir on Mars and its evolution with time. The current ratio is at least five times that of the Vienna Standard Mean Ocean Water (SMOW) reference derived from Earth's oceans (Owen et al., 1988; Encrenaz, T. et al., 2018; Krasnopolsky, 2015; Villanueva et al., 2015). Henceforth, D/H ratios presented in the rest of this study are expressed with respect to SMOW.

H and D atoms in the upper atmosphere come from H₂O and HDO, their sole precursor in the lower atmosphere. The lower mass of H over D atoms and the fact that H₂O is preferentially photolysed over HDO (Cheng et al., 1999) explain the differential escape of these two elements. Also, the heavier isotope, HDO, has a lower vapor pressure than H₂O, which results in an enrichment of the deuterated isotope in the solid phase of water. This effect is known as the Vapor Pressure Isotope Effect (VPIE) which can reduce the D/H ratio above the condensation level to values as low as 10% of the D/H ratio near the surface (Bertaux & Montmessin, 2001; Fouchet & Lellouch, 2000).

Fractionation should affect the amount of HDO depending on latitude, longitude, altitude and season. In particular, a previous model (Montmessin et al., 2005, hereafter referred to as M05) has predicted that an isotopic gradient should exist between the cold regions where condensation depletes the atmosphere in HDO and the warmer, condensation-free, regions where D/H is found to be maximum. This led M05 to predict a latitudinal gradient of D/H (with variations greater than a factor of 5) between the warm and

63 moist summer hemisphere and the cold and dry winter hemisphere. This gradient was
 64 later confirmed by several Earth ground-based and Mars orbiting assets (Krasnopolsky,
 65 2015; Villanueva et al., 2015; Encrenaz, T. et al., 2018; Aoki et al., 2019). Yet some ob-
 66 servations (Villanueva et al., 2015; Khayat et al., 2019) have revealed longitudinal vari-
 67 ations of D/H ratios in appearance correlated to topography that were not reproduced
 68 by any model.

69 Since HDO and H₂O dictate eventually how many D and H atoms will populate
 70 the Martian exosphere, studying their escape and extrapolating back in times to infer
 71 the amount of water lost to space requires modeling their present-day behavior correctly
 72 in the lower atmosphere. It is therefore relevant to address the HDO cycle and its as-
 73 sociated fractionation with respect to water in a 3D Mars climate model (GCM) akin
 74 to M05. The present work leverages on this past effort, and implements important changes
 75 that occurred between the 2005 GCM version and the one currently in use. Upgrading
 76 the HDO model enabled the possibility to test the theoretical reaction of HDO during
 77 the most extreme meteorologic event that can happen on Mars, namely a Global Dust
 78 Storm (GDS). The Martian Year 34 (henceforth MY34, corresponding to the 2017-2019
 79 time frame) was the siege of a GDS that started in the middle of the year ($L_s=176^\circ$) and
 80 stopped shortly before perihelion ($L_s=235^\circ$) (Guzewich et al., 2018). This event profoundly
 81 affected the behavior of water vapor in the Martian atmosphere (Fedorova et al., 2020)
 82 with hints of a comparable affect for HDO (Vandaele et al., 2019).

83 The present study describes a 3D climate model projection of the behavior of HDO
 84 in the context of a dust annual scenario mimicking the dust seasonal and spatial evo-
 85 lution observed during MY34 which include the occurrence of the GDS. The simulations
 86 reported here emphasize the contrast in HDO behavior between a perturbed year such
 87 as MY34 and a more traditional MY where the seasonal evolution of dust obeys a re-
 88 current and more quiet, so-called "climatological" scenario.

89 2 Model overview

90 The simulations presented here were conducted with the Martian Global Climate
 91 Model (MGCM) developed at the Laboratoire de Météorologie Dynamique (LMD) in
 92 collaboration with several European teams (LATMOS, IAA Granada, University of Ox-
 93 ford, The Open University), with support from ESA and CNES. A general description
 94 of the model is given in Forget et al. (1999). A first implementation of the water cycle
 95 was presented in Montmessin et al. (2004) (used in M05) and was then improved in Madeleine
 96 et al. (2011) and subsequently in Navarro et al. (2014), with the implementation of the
 97 radiative effect of clouds and the microphysics accounting for the fine processes of cloud
 98 formation as nucleation, ice growth, scavenging of dust by condensing ice, and supersat-
 99 uration. However, those were not activated for the simulations conducted in this study.
 100 This is motivated by the desire to track changes added to the GCM since M05 before
 101 increasing model complexity. The simpler cloud approach of M05 employed here is nev-
 102 ertheless augmented by the predictive dust representation of Madeleine et al. (2011), which
 103 better constrains the availability of dust nuclei for water ice clouds to condense onto, even
 104 if the details of cloud-dust interactions (nucleation, scavenging and supersaturation as
 105 mentioned above) are not yet introduced. Observations from Earth (Khayat et al., 2019;
 106 Villanueva et al., 2015; Novak et al., 2011) are in agreement with predictions of M05 re-
 107 garding the latitudinal variability of the D/H ratio over the column of atmosphere, sup-
 108 porting the idea that a simple cloud model carries enough physics to capture the effect
 109 of condensation-induced fractionation on HDO. We also mention that since Colaïtis et
 110 al. (2013) the model integrates a thermal plume model to better represent the sub-grid
 111 scale convection occurring in the Planetary Boundary Layer (PBL). This parametriza-
 112 tion has been shown to significantly improve the wind and temperature structures near
 113 to the surface in comparison to observations and LES studies, which has a direct impact
 114 on the transport of water and dust above the PBL.

115 In this upgraded version of the HDO model, the vertical distribution of dust obeys
 116 a semi-interactive scheme (Madeleine et al., 2011) based on an order 2 moment repre-
 117 sentation of the dust tracer. Dust is injected everywhere and at every numerical time-
 118 step (15 minutes in our simulations) at the surface of the model by a constant lifting rate.
 119 Both the mass and the number of dust particles are transported, enabling the interac-
 120 tive derivation of the dust particle size in every model cell to better constrain dust sed-
 121 imentation. While the vertical profile of the dust evolves freely, sustaining changes due
 122 to winds and gravitational settling, the resulting opacity profile is multiplied by a scal-
 123 ing factor so that the column dust opacity of the model matches the observed column
 124 dust opacity as compiled by Montabone et al. (2015, 2020). The latter being the main
 125 parameter controlling the radiative balance of the atmosphere and thus the background
 126 thermal state. It also linearly scales the number of dust nuclei available for water ice cloud
 127 formation and thus ice crystal size and subsequent gravitational fall.

128 The MGCM used for our HDO simulations relies on a 64×48 longitude-latitude grid
 129 with 32 vertical levels, covering the atmospheric column from the surface up to 120 km
 130 (10^{-3} Pa). An extension to the thermosphere up to the exobase around 250 km (Angelats i
 131 Coll et al., 2005; González-Galindo et al., 2009) and a photochemical module with 15
 132 species (Lefèvre et al., 2004; Lefèvre et al., 2008) are available in the model. However
 133 these modules have not been activated in the simulations used for the present paper, since
 134 their integration will be part of a future study analysing the whole HDO and D/H cy-
 135 cle.

136 2.1 Simulation of HDO

137 HDO is tracked in both its vapour and ice phases in the atmosphere and also on
 138 the surface as ice deposit. It is treated as a tracer separate from H_2O , but undergoes the
 139 same condensation and transport processes. Isotopic fractionation is introduced during
 140 condensation following the approach described by Fouchet and Lellouch (2000), Bertaux
 141 and Montmessin (2001) and later by M05.

142 The fractionation factor α has been traditionally derived from the formula estab-
 143 lished by Merlivat and Nief (1967). This factor describes the relative enrichment of HDO
 144 in the ice with respect to the surrounding vapour phase at equilibrium (i.e. no net flux
 145 between the two phases), for a temperature T , and can be expressed as:

$$\alpha(T) = \frac{(HDO/H_2O)_{ice}}{(HDO/H_2O)_{vap}} = \exp\left(\frac{16288}{T^2} - 9.34 \times 10^{-2}\right) \quad (1)$$

146 We note that Eq. (1) is based on experiments performed at temperatures above
 147 233 K, warmer than typical temperatures encountered on Mars, leading M05 to extrap-
 148 olate Eq. (1) down to temperatures as low as 100 K. However, recent experimental mea-
 149 surements from Lamb et al. (2017) were conducted between 234 and 194 K. In doing so,
 150 they updated the fractionation law and proposed a new formula:

$$\alpha(T) = \exp\left(\frac{13525}{T^2} - 5.59 \times 10^{-2}\right) \quad (2)$$

151 The difference between the two formulas increases exponentially with decreasing
 152 temperature, from 3% at 200 K, to 8% at 150 K and 21% at 100 K (Fig. 1). The Lamb
 153 formula yields a fractionation factor systematically smaller than the classical Merlivat
 154 formula at Martian temperatures, which suggests the studies conducted so far on Mar-
 155 tian HDO have slightly overestimated the fractionation factor. Lamb's formula is used
 156 as a reference for this study.

157 We assume that the condensation flux is in isotopic equilibrium with the vapour
 158 phase. This is justified by the fact that diffusivity of HDO inside ice is too slow to ho-

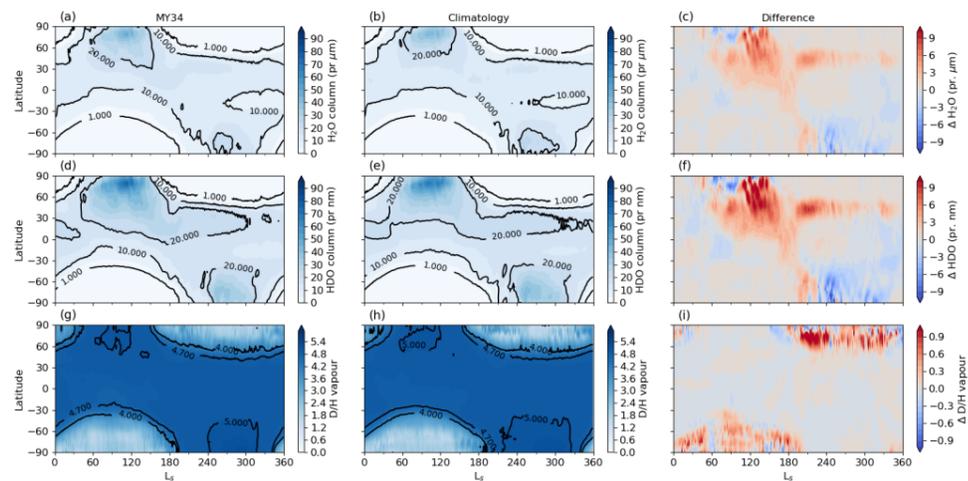


Figure 1. (Left) Value of the fractionation factor α as derived from Lamb et al. (2017) or Merlivat and Nief (1967). (Right) Change on the zonally averaged seasonal variation of the D/H ratio. The absolute difference is showed between the use of the formula from Lamb et al. (2017) and that of Merlivat and Nief (1967). Contour lines indicate where the D/H ratio reaches 3 and 5 in the Lamb simulation.

159 mogenize HDO inside the ice particle. Therefore, the newly added icy layer is indepen-
 160 dent of the existing ones. In that context, the condensing mass dM_D of HDO can be ex-
 161 pressed as:

$$dM_D = \alpha(T)dM_H \times \frac{M_D}{M_H} \quad (3)$$

162 where M indicates the mass in the vapour phase, dM the mass condensing, with sub-
 163 script H referring to water and D to HDO.

164 In the model, the flux of water during cloud condensation is computed assuming
 165 that any supersaturated excess of water vapor is turned into atmospheric water ice (Montmessin
 166 et al., 2004). At the surface, the flux of water is computed with the surface turbulent
 167 flux (Forget et al., 1999; Montmessin et al., 2004). In both cases, once the flux of wa-
 168 ter is known, the flux of HDO is computed using equation 3. The temperature used to
 169 compute $\alpha(T)$ is that of the atmospheric layer condensing (for the particular case of the
 170 surface flux, it corresponds to the temperature of the first layer above the surface).

171 A significant difference with M05 is the implementation of the "tracer genealogy"
 172 scheme developed by Risi et al. (2010) to study the HDO atmospheric cycle in the LMDZ
 173 Earth GCM. M05 had an older version of the dynamical core of the GCM that did not
 174 include or require this specific transport of isotopes in the tracer scheme. In this scheme,
 175 the isotopic ratio is transported by the dynamical transport scheme instead of HDO it-
 176 self (more details about this transport scheme can be found in the appendix of Risi et
 177 al., 2010, and in the Supplementary Information). The implementation of the "tracer
 178 genealogy" scheme has significantly improved the GCM results by eliminating the un-
 179 physical isotopic ratios initially observed.

180 2.2 Dust scenarios

181 We employ two dust scenarios for this study. We are calling "dust scenario" the
 182 prescription in the model of the seasonal evolution and the spatial variation of the dust

183 column opacity. The scenarios for each MY have been compiled by Montabone et al. (2015),
 184 and the dust scenario relevant to MY34 that encompasses the 2018 GDS was recently
 185 added by Montabone et al. (2020) to the existing database.

186 The first one is the Climatology scenario, which corresponds to an average of non-
 187 GDS martian years, and that is used as our reference run. The other one is the dust sce-
 188 nario of Martian Year 34 (MY34), which includes the dust storm occurrence between $L_s=176^\circ$
 189 and $L_s=235^\circ$.

190 The MY34 GDS started in May 2018 in the northern hemisphere and progressively
 191 expanded to eventually enshroud the entire Martian globe by June 2018 ($L_s=186^\circ$). Then
 192 the GDS progressively faded and let the Mars atmosphere return to a nominal (clima-
 193 tological) state from August to September 2018 ($L_s=270^\circ$) (Snchez-Lavega et al., 2019;
 194 Montabone et al., 2020).

195 2.3 Initial conditions

196 The initial conditions of the simulation impose a constant D/H ratio of 5 every-
 197 where in the atmosphere and in the surface ice. The perennial ice in the northern po-
 198 lar cap acts as an infinite reservoir of water ice, which is also prescribed to have a D/H
 199 ratio of 5.

200 In order to reproduce the column of water vapour, in particular the peak of the sub-
 201 liming north pole in northern spring and summer, we had to adjust the albedo of the peren-
 202 nial surface ice, as described in Navarro et al. (2014), to a value of 0.32 yielding a max-
 203 imum column abundance of 70 pr.- μm similar to observations (Trokhimovskiy et al., 2015).

204 We start using the Climatology scenario and run the model for a couple of Mars
 205 years, allowing us to reach a stable state. The resulting model state, after this two-year
 206 spin-up phase, supplies the new initial conditions of two following simulations: one based
 207 on the "climatology" dust scenario, and one based on the MY34 dust scenario.

208 3 Results

209 3.1 Reference run: the "climatological" scenario

210 Figure 2 (center column) shows the latitudinal and seasonal distribution of the zon-
 211 ally averaged column abundance of HDO vapour along with the corresponding D/H ra-
 212 tio. In essence, HDO vapour and ice reproduce the global cycle of water vapor and ice
 213 with a peak of vapor and atmospheric ice abundance in the Aphelion period where the
 214 north polar cap, fully exposed to the Sun, rejects massive amounts of water and heavy
 215 water in the atmosphere. At the same time, water ice clouds are building up in the north-
 216 ern tropics where air masses ascend and adiabatically cool as a consequence of a global
 217 convergence of the atmosphere in the lower atmosphere, causing the recurrent appari-
 218 tion of the "Aphelion" cloud belt.

219 In summary, the polar night in both hemispheres, which encompasses the $L_s =$
 220 180° to $L_s = 360^\circ$ period in the north and the $L_s = 0^\circ$ to $L_s = 180^\circ$ period in the
 221 south, extends equatorward up to 45° near solstices. These regions host the coldest at-
 222 mospheric temperature overall and as a result the most active water ice condensation
 223 mechanism. Likewise, HDO condenses a lot in these regions and is subsequently heav-
 224 ily fractionated, even more because of the exponential increase of the fractionation fac-
 225 tor with decreasing temperature (α doubles between 150 and 100 K). This results in a
 226 5:1 gradient between the polar night regions and the regions elsewhere in both hemispheres.
 227 Remarkably, the edge of the polar night corresponds to a sharp gradient in D/H, indi-
 228 cating fractionation is very effective there and rather ineffective in regions equatorward.
 229 Therefore, the latitudinal gradient suggested by M05 between the cold regions and the

230 warm regions, and whose existence has been confirmed later by several observers, is also
 231 present in our simulation.

232 Clearly, an event such as the Aphelion Cloud belt has no strong D/H signature in
 233 a column-integrated perspective since (1) the cloud belt results from moderately cold tem-
 234 perature (180 K) condensation and (2) forms above 10 to 15 km where atmospheric den-
 235 sity has already dropped by a factor 3 to 10, fractionating HDO over a small portion of
 236 the column. Altogether, this explains why the global effect of fractionation on Mars is
 237 so much dominated by the polar regions.

238 **3.2 The Martian Year 34 run: impact of the Global Dust Storm**

239 Figure 2 (left column) shows the latitudinal and seasonal evolution of the HDO vapour
 240 column and D/H ratio, this time for the MY34 dust scenario. The difference with the
 241 climatology scenario is shown in the right panels. One can note that the D/H ratios over
 242 the column are quite similar, suggesting that the dust storm does not affect the isotopic
 243 ratio of the atmosphere as a whole.

244 In order to assess the effect that the dust storm may potentially have on the ver-
 245 tical transport of H₂O and HDO and its evolution with time, we analyze zonally aver-
 246 aged meridional profiles of these quantities (Figure 3). This figure shows a comparison
 247 of the computed meridional profiles between the Climatology and the MY34 scenarios.
 248 The influence of the dust storm is evident on the temperature field with a strong (> 20 K)
 249 increase of the temperatures in the whole atmosphere due to heating by dust.

250 Increased dust content has a direct effect on fractionation: higher temperatures re-
 251 strain condensation, and the subsequent cloud formation in the middle atmosphere ex-
 252 hibited in the Climatological run is pushed to higher altitudes in the MY34 run. Con-
 253 densation being the main source of isotopic fractionation for water, the deuteropause,
 254 that is the level above which D/H declines due to condensation-induced fractionation
 255 (Bertaux & Montmessin, 2001), is found around 100 Pa in the Climatology run through-
 256 out the same L_s period as the GDS run. In the MY34 run, clouds are forming above 10 Pa,
 257 moving the deuteropause higher in altitude, leaving the D/H ratio mostly unchanged ver-
 258 tically below that level with values of 4.5 reached at 10 Pa, instead of 3.5 in the Clima-
 259 tology run. This implies that the vertical barrier of deuterium becomes 30% more porous
 260 in a MY34 configuration, likely letting the same excess of deuterium atoms accessing the
 261 upper atmosphere where they can escape to space.

262 The GDS therefore affects the D/H profile in two ways: (1) higher temperature sup-
 263 presses the vertical confinement imposed by condensation with associated fractionation,
 264 (2) intensified atmospheric circulation conveys higher amount of water and its isotope
 265 to much higher altitude. This effect on D/H, predicted here by our HDO model, has been
 266 already highlighted for water by concomitant observations from the instruments ACS
 267 and NOMAD onboard TGO (Fedorova et al., 2020; Aoki et al., 2019; Vandaele et al.,
 268 2019) and also reproduced by a GCM (Neary et al., 2020).

269 Figure 4 shows vertical profiles of the planetary-averaged amount of H₂O and HDO
 270 vapour and the related D/H ratio, averaged over the second half of the year (L_s = 180° –
 271 360°). The effect of the GDS is quite visible on the water and HDO vapour with a sharp
 272 increase in the mixing ratio at most altitudes above 100 Pa in the case of the MY34 sce-
 273 nario. The amount of water is increased in the upper atmosphere by about 10%, but the
 274 effect is even stronger for HDO with an increase by 40% which directly scales with the
 275 escaping flux of deuterium and can be attributed to the reduced efficiency of the deuteropause.
 276 On average over the dust-storm season, the D/H at high altitude is greater by 25% com-
 277 pared to a regular year represented by the Climatology run.

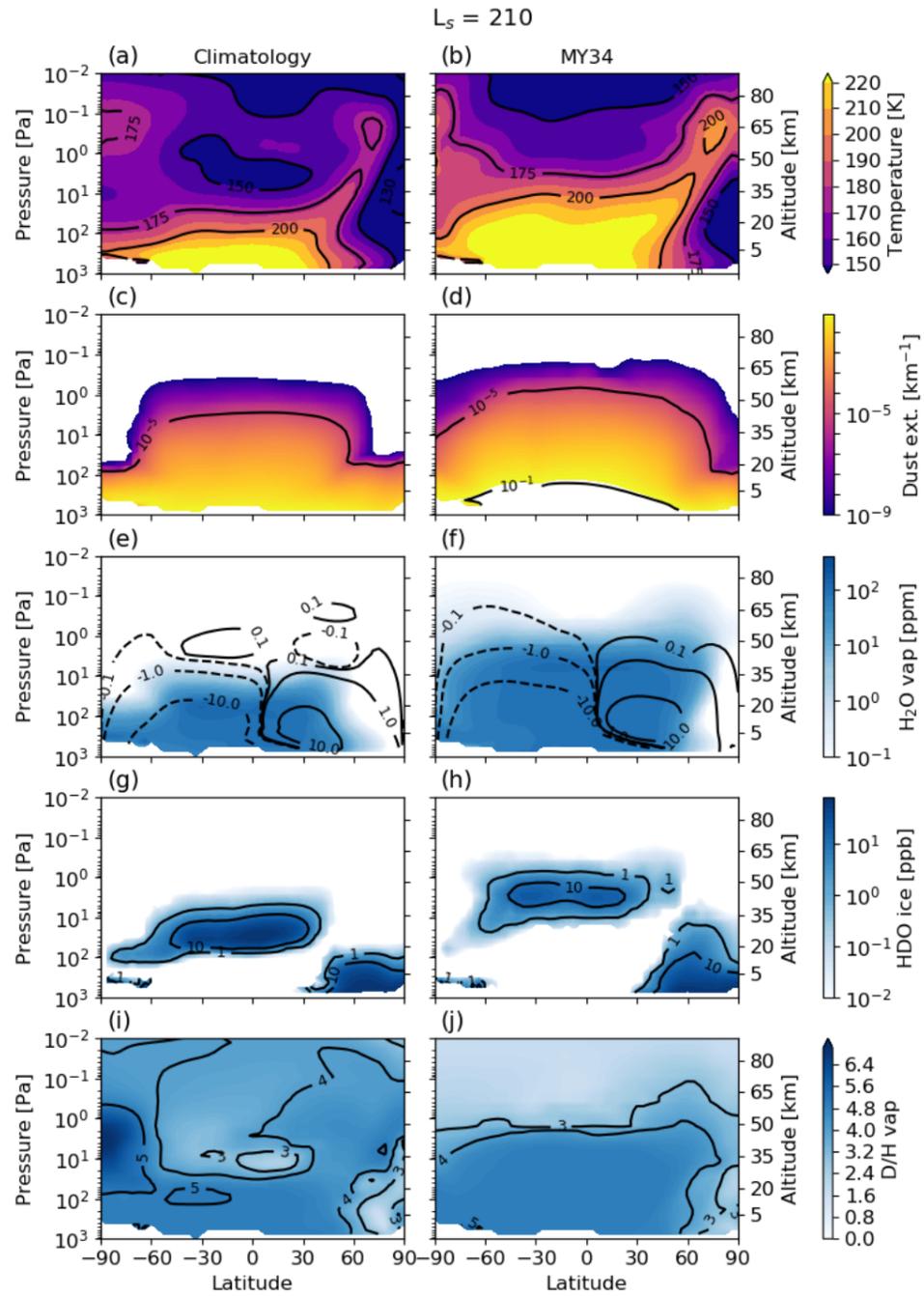


Figure 2. Seasonal evolution of the zonally averaged column of H_2O (panel a and b) and HDO (panel d and e) in the vapour phase, for the climatology and MY34 scenarios. Below is the corresponding D/H ratios in terms of SMOW (panels g and h), for the vapour phases. Units are precipitable microns or nanometers. Left column is for MY34, center column for the Climatology scenario and right column (c, f and i) is the absolute difference between MY34 and Climatology. A similar figure for the ice phase can be found in the Supplementary Information.

Figure 3. Zonally averaged meridional profiles of (from top to bottom) temperature (a and b), dust extinction at 500 nm (c and d), H₂O vapour (e and f), HDO ice (g and h) and D/H ratio of the vapour (i and j), at a local-time of 14h and at solar longitude $L_s = 210^\circ$. The left column is the climatology scenario and the right column is MY34. The contours on panels (e) and (f) are the zonally averaged streamfunction, with positive values indicating a clockwise circulation.

Figure 4. Planetary averaged profiles of H₂O vapour (a), HDO vapour (b) and D/H ratio of the vapour (c) averaged over the period $L_s = 180^\circ - 360^\circ$ for the climatology and the MY34 scenarios. The relative change with respect to the climatology scenario is shown with a dashed curve.

278 4 Discussion

279 Our scenario of the MY34 GDS is not fully representative of the observed condi-
 280 tions. In particular regarding the dust profile and the altitude of the hygropause. In our
 281 simulations, the dust remains mostly below 70 km, while observations from MCS show
 282 that the dust reached altitudes up to 80 km during the storm (Kass et al., 2020). And
 283 the height of the hygropause is limited to 60-65 km in our simulation, while observations
 284 show that it extended up to 75-80 km during the GDS (Heavens et al., 2019).

285 These discrepancies may be attributed to missing physical processes in the model.
 286 Firstly, we ignore the radiative effect of clouds and the microphysics of cloud formation.
 287 These affect the temperatures and enhance the global circulation, as discussed in Madeleine
 288 et al. (2011) and Navarro et al. (2014). This could result in an overestimation of the ice
 289 content and a reduced transport of water vapour. In addition, without taking microphysics
 290 into account, it is not possible to reproduce the important supersaturation observed by
 291 ACS during the GDS, and which is thought to help water vapour reaching higher alti-
 292 tudes (Fedorova et al., 2020).

293 Secondly, Madeleine et al. (2011) particularly mention the limits of uniformly lift-
 294 ing dust in the case of strong dust storm events, as it does not account for the strong
 295 positive feedback between atmospheric dust heating and lifting occurring then. However,
 296 other approaches using finer parametrizations of dust lifting have similar difficulties re-
 297 producing the dust vertical distribution in the GDS conditions (Neary et al., 2020). Fi-
 298 nally, the model doesn't account for the convective events suspected to be responsible
 299 for the presence of the detached dust layers (Spiga et al., 2013; Daerden et al., 2015; Wang
 300 et al., 2018; Heavens et al., 2018). As discussed by Neary et al. (2020), an underestima-
 301 tion of the amount of dust above 40 km leads to insufficient heating, leaving the forma-
 302 tion of ice particles possible and therefore restraining the extent of water vapour in the
 303 upper atmosphere.

304 For these reasons, we suspect that our simulation could in fact overestimate the
 305 efficiency of the deuteropause. It is thus possible that the increased amount of HDO at
 306 high altitudes caused by the GDS as in our simulation, is in fact a lower estimate of the
 307 real effect. While we are not able to represent the full extent of the GDS of MY34, our
 308 results feature the essential processes occurring in such situation, in particular the re-
 309 duction of the condensation-induced fractionation, which defines the altitude of the deuteropause,
 310 and the increase in atmospheric circulation.

311 **5 Conclusions**

312 We have presented an updated model of the Martian HDO cycle inherited from Montmessin
 313 et al. (2005) and leveraging on the work done since then in particular by Madeleine et
 314 al. (2011). This model, which represents the 3D advection of HDO along with the phys-
 315 ical processes controlling the fate of HDO in the Martian atmosphere, has been employed
 316 to explore for the first time the theoretical impact of a global dust storm on HDO. Of
 317 particular interest is the access of HDO to the upper atmospheric layers where it is sub-
 318 sequently chemically decomposed into escaping deuterium atoms. In doing so, we find
 319 that the presence in increased amounts of dust in the atmosphere at times of the GDS
 320 restrains condensation and cloud formation, mostly removing the effect of condensation-
 321 induced fractionation. The deuteropause, that is otherwise predicted to remain at an ap-
 322 proximate altitude of 30 km in a regular year is pushed up to 50 km and appears more
 323 porous than in our climatology run representative of average climatic conditions on Mars.
 324 The GDS is predicted to enhance the presence of deuterated water at 100 km by 40%,
 325 implying that a similar excess of deuterium atoms might escape from Mars in a year of
 326 a GDS.

327 The detailed effects of GDS on the amount of HDO in the upper atmosphere will
 328 be further explored including detailed cloud microphysics and cloud radiative feedbacks,
 329 which are known to affect the cloud formation (Navarro et al., 2014), and should have
 330 an effect on condensation-induced fractionation. The details of the photochemistry and
 331 processes occurring in the thermosphere are missing from the present simulation, but are
 332 essential to represent the differential photolysis of H₂O and HDO and their subsequent
 333 escape. These processes will be the object of future improvements of the LMD MGCM,
 334 in particular in the context of detailed comparison with available spacecraft observations
 335 from ACS onboard TGO. They will help us better understand the global cycle of deu-
 336 terium from the troposphere to the exosphere, and supply a self-consistent framework
 337 to investigate its relation with the escape of water from the planet.

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