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10           3D Modelling of the climatic impact of  
11 outflow channel formation events on Early  
12           Mars.

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22

**Abstract**

23 Mars was characterized by cataclysmic groundwater-sourced surface flooding  
24 that formed large outflow channels and that may have altered the climate for  
25 extensive periods during the Hesperian era. In particular, it has been speculated  
26 that such events could have induced significant rainfall and caused the formation  
27 of late-stage valley networks. We present the results of 3-D Global Climate Model  
28 simulations reproducing the short and long term climatic impact of a wide range  
29 of outflow channel formation events under cold ancient Mars conditions. We  
30 find that the most intense of these events (volumes of water up to  $10^7$  km<sup>3</sup> and  
31 released at temperatures up to 320 Kelvins) cannot trigger long-term greenhouse  
32 global warming, regardless of how favorable are the external conditions (e.g.  
33 obliquity and seasons). Furthermore, the intensity of the response of the events is  
34 significantly affected by the atmospheric pressure, a parameter not well constrained  
35 for the Hesperian era. Thin atmospheres ( $P < 80$  mbar) can be heated efficiently  
36 because of their low volumetric heat capacity, triggering the formation of a convective  
37 plume that is very efficient in transporting water vapor and ice at the global scale.  
38 Thick atmospheres ( $P > 0.5$  bar) have difficulty in producing precipitation far  
39 from the water flow area, and are more efficient in generating snowmelt. In any  
40 case, outflow channel formation events at any atmospheric pressure are unable to  
41 produce rainfall or significant snowmelt at latitudes below 40°N. As an example,  
42 for an outflow channel event (under a 0.2 bar atmospheric pressure and 45° obliquity)  
43 releasing  $10^6$  km<sup>3</sup> of water heated at 300 Kelvins and at a discharge rate of  
44  $10^9$  m<sup>3</sup> s<sup>-1</sup>, the flow of water reaches the lowest point of the northern lowlands  
45 (around ~ 70°N, 30°W) after ~ 3 days and forms a 200m-deep lake of  $4.2 \times 10^6$  km<sup>2</sup>  
46 after ~ 20 days; the lake becomes entirely covered by an ice layer after ~ 500 days.  
47 Over the short term, such an event leaves  $6.5 \times 10^3$  km<sup>3</sup> of ice deposits by precipitation  
48 (0.65% of the initial outflow volume) and can be responsible for the melting of  
49 ~ 80 km<sup>3</sup> (0.008% of the initial outflow volume; 1% of the deposited precipitation).  
50 Furthermore, these quantities decrease drastically (faster than linearly) for lower  
51 volumes of released water. Over the long term, we find that the presence of the  
52 ice-covered lake has a climatic impact similar to a simple body of water ice located  
53 in the Northern Plains.

54 For an obliquity of ~ 45° and atmospheric pressures > 80 mbar, we find that  
55 the lake ice is transported progressively southward through the mechanisms of  
56 sublimation and adiabatic cooling. At the same time, and as long as the initial

57 water reservoir is not entirely sublimated (a lifetime of  $10^5$  martian years for the  
58 outflow channel event described above), ice deposits remain in the West Echus  
59 Chasma Plateau region where hints of hydrological activity contemporaneous with  
60 outflow channel formation events have been observed. However, because the high  
61 albedo of ice drives Mars to even colder temperatures, snowmelt produced by  
62 seasonal solar forcing is difficult to attain.

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## 1 Introduction

During the Late Hesperian epoch of the history of Mars (about 3.1-3.6 Gyrs ago; Hartmann and Neukum (2001)), the large outflow channels observed in the Chryse Planitia area are thought to have been carved by huge water floods caused by catastrophic and sudden release of groundwater (Baker, 1982; Carr, 1996). It has been speculated that such events could have warmed the climate and possibly explain the contemporaneous formation of dendritic valley networks observed in the nearby Valles Marineris area and on the flanks of Alba Patera, Hecates Tholus, and Ceraunius Tholus, and that have been interpreted to be precipitation-induced (Gulick and Baker, 1989, 1990; Baker et al., 1991; Gulick et al., 1997; Gulick, 1998, 2001; Mangold et al., 2004; Quantin et al., 2005; Weitz et al., 2008; Santiago et al., 2012). Although the Late Hesperian epoch is thought on the basis of geology and mineralogy to have been cold (Head et al., 2004; Bibring et al., 2006; Ehlmann et al., 2011), the characteristics of these dendritic valleys suggest that the valleys were formed under persistent warm conditions (.e.g, Mangold et al. 2004). First, their high degree of branching is interpreted to indicate formation by precipitation. Second, their high drainage densities - evidence of their high level of maturity - and the presence of inner channels favor the presence of stable liquid water for geologically long periods of time (Craddock and Howard, 2002). Third, sedimentary morphologies observed in the region of Valles Marineris (Quantin et al., 2005) suggest a fluvial and lacustrine environment. Under this hypothesis, the warm liquid water floods that formed the outflow channels would inject water vapor into the atmosphere, a powerful greenhouse gas that could trigger a significant warming period possibly leading to long lasting pluvial activity (rainfall).

In this paper, we use a 3-Dimensional Global Climate Model (LMD GCM) to explore the global climatic impact of outflow channel water discharge events on a Late Hesperian Mars over a range of temperatures and atmospheric pressures. These bursts of warm liquid groundwater outflows onto the surface can trigger strong evaporation, possibly leading to global climate change. How warm and how wet was the atmosphere of Late Hesperian Mars after such major outflow channel events? The climatic effect of relatively small and cool groundwater discharges has been studied on a regional scale (Kite et al., 2011a) and localized precipitation is indicated. In this contribution, we investigate the climatic impact at a global scale of a wide range of possible outflow channel events, including the case of the most intense outflow events ever recorded on Mars (Carr, 1996). Our work focuses on both (1) the direct short-term climate change due to the initial strong evaporation of water vapor and (2) the long-term change of the water cycle

100 due to the presence of liquid water and ice at non-stable locations.

101 When a warm liquid water flow reaches the surface, strong evaporation occurs  
102 and the total evaporation rate increases with the temperature and the surface area  
103 of the flow. In term of energy budget, a 300K warm liquid water flow can potentially  
104 convert  $\sim 5\%$  of its total mass into water vapor before freezing starts. The injected  
105 water vapor will have a major role on the radiative budget of the planet. First,  
106 water vapor is a greenhouse gas that can absorb ground thermal infrared emission  
107 efficiently. Second, water vapor can condense to form clouds. In the process, large  
108 amounts of latent heat can be released in the atmosphere. The clouds can reflect  
109 the incoming solar flux as well as contribute to an additional greenhouse effect,  
110 depending on their height and the opacity of the background atmosphere, which  
111 depends on the total atmospheric pressure.

112 To study the global climatic effect of localized outflow channel events, 3D  
113 Global Climate Models are particularly relevant because they not only model the  
114 physical processes described above, but also the 3D dynamical processes that play  
115 a major role in climatic evolution. In particular, we show in this paper that 3D  
116 dynamical processes (horizontal advection, in particular) are key to understanding  
117 the relaxation timescale of the Late Hesperian martian atmosphere immediately  
118 following major outflow channel events.

## 119 **2 Background**

### 120 **2.1 Outflow channels**

#### 121 **2.1.1 Description**

122 Outflow channels are long (up to  $\sim 2000$  km) and wide (up to  $\sim 100$  km) valleys  
123 that were sculpted by large-scale debris-laden water flows (Baker, 1982; Baker  
124 et al., 1992; Carr, 1996). The most prominent martian outflow channels are  
125 located in the circum-Chryse area and debouch in the Chryse Planitia itself (Carr,  
126 1996; Ivanov and Head, 2001).

127 Several processes have been suggested to have caused such outburst floods  
128 (Kreslavsky and Head, 2002). It is likely that the water that was released during  
129 these events come from subsurface aquifers (Clifford, 1993; Clifford and Parker,  
130 2001). In this scenario, the temperature of the extracted subsurface water is  
131 controlled by the geothermal gradient and thus would depend on its initial depth of  
132 origin. During the Late Hesperian, when outflow channel events largely occurred,  
133 this gradient could have been locally higher (Baker, 2001), because the circum-Chryse

134 area is close to the volcanically active Tharsis region (Cassanelli et al., 2015).  
135 Therefore, the discharged water could have reached the surface at a maximum  
136 temperature of tens of degrees above the freezing point (Kreslavsky and Head,  
137 2002). We note that the run-away decomposition of CO<sub>2</sub> clathrate hydrate (Milton,  
138 1974; Baker et al., 1991; Hoffman, 2000; Komatsu et al., 2000), proposed as a  
139 possible mechanism for the origin of the outflow water, cannot produce water  
140 temperature greater than 10K above the freezing point. To a first approximation,  
141 and from a climatic point of view, the only difference between these two processes  
142 of liquid water discharge is the temperature of the water. Thus, we considered in  
143 this paper various cases ranging from 280 Kelvins to 320 Kelvins (see section 6.1.1).

144 Whatever the physical process operating, large amounts of water released at  
145 very high rates are needed at the origin of the water flow in order to explain the  
146 erosion of the circum-Chryse outflow channels. The quantity of water estimated  
147 to erode all the Chryse basin channels is  $\sim 6 \times 10^6 \text{ km}^3$  assuming 40% by volume  
148 of sediment (Carr, 1996) but could possibly be much more if one assume lower  
149 sediment loads (Kleinhans, 2005; Andrews-Hanna and Phillips, 2007), which is,  
150 for example the case on Earth ( $\sim 0.1\%$  of sediment by volume).

151 The different estimates of outflow channel single-event volumes, discharge  
152 rates and durations lead to a wide range of results, but two endmember scenarios  
153 can be defined and explored. On the one hand, some researchers estimated that  
154 only a limited number of very intense (volume up to  $3 \times 10^6 \text{ km}^3$ , discharge rates  
155 up to  $10^9 \text{ m}^3 \text{ s}^{-1}$ ) outflow channel formation events actually occurred (Rotto and  
156 Tanaka, 1992; Baker et al., 1991; Komatsu and Baker, 1997).

157 On the other hand, more recently, other researchers argued that outflow channels  
158 were formed by numerous individual small events (Andrews-Hanna and Phillips,  
159 2007; Harrison and Grimm, 2008). This latter work implies water volumes from  
160 hundreds to thousands of  $\text{km}^3$ , discharge rates of  $10^6\text{-}10^7 \text{ m}^3 \text{ s}^{-1}$  for individual  
161 events and a minimum period between successive single events of  $\sim 20$  martian  
162 years. These endmember estimates differ by several orders of magnitude, but in  
163 this paper, we explored the full range.

### 164 **2.1.2 Fate of the outflow channel liquid water flow**

165 In this section, we provide a description of the possible fate, and calculations of  
166 the possible velocities, of the outflow channel water; these will serve as input for  
167 the description of the liquid water runoff under various conditions in the GCM  
168 simulations.

169 The ejected liquid water flows from the circum-Chryse area all inevitably

170 debouch into the basin of Chryse. However, Chryse Planitia is not a closed basin  
 171 and if the total amount of water released in a single event is high enough, the  
 172 water will spill into the Northern Plains (Ivanov and Head, 2001), flowing down  
 173 on slopes inclined at  $\sim 0.03^\circ$  for more than 2000km. This is an important point  
 174 because, as the wetted area of the flow increases, the total rate of evaporation  
 175 rises. The fate of the outflow channel liquid water flow can be subdivided into  
 176 two steps:

177 1. First, the ground-surface liquid water flows 'inside' the outflow channels.  
 178 The Reynolds Number  $Re$  of such flows is given by

$$Re = \rho U_c R_c / \mu, \quad (1)$$

179 with  $U_c$  the mean water flow velocity in the channel,  $R_c$  the hydraulic radius (see  
 180 below) of the channel,  $\rho$  the density and  $\mu$  the viscosity of the flow. For most of the  
 181 outflow channel events, this number must have been orders of magnitude higher  
 182 than 500 (Wilson et al., 2004), meaning that the released ground water flows were  
 183 turbulent.

184 The most accurate way (Bathurst, 1993; Wilson et al., 2004) to calculate the  
 185 mean velocity of such flows is to use the Darcy-Weisbach equation:

$$U_c = (8gR_c \sin \alpha / f_c)^{1/2}, \quad (2)$$

186 with  $g = g_{\text{mars}} = 3.71 \text{ m s}^{-2}$  the gravity on Mars,  $\alpha$  the slope angle of the channel  
 187 and  $f_c$  a dimensionless friction factor which mostly depends on the bed roughness  
 188  $z_c$  and the water depth  $h$  of the flow. This factor can be expressed as follows  
 189 (Wilson et al., 2004):

$$(8/f_c)^{1/2} = a \log_{10}(R_c/z_c) + b, \quad (3)$$

190 with  $a$  and  $b$  two empirical coefficients, which are respectively equal to 5.657 and  
 191 6.6303 if the bed roughness  $z_c$  ( $z_c = 10^{-2}$  m here) is fixed (Knudsen and Katz,  
 192 1958): This leads equation (2) to the following equation:

$$U_c = (gR_c \sin \alpha)^{1/2} (a \log_{10}(R_c/z_c) + b). \quad (4)$$

193 The hydraulic radius  $R_c$  is defined as the cross-sectional area of the channel divided  
 194 by its wetted perimeter:

$$R_c \sim (W_c h)/(W_c + 2h), \quad (5)$$

195 with  $W_c$  the channel width and  $h$  the flow depth. Because outflow channels are  
 196 wider than deep ( $W_c \sim 10\text{-}100$  km wide but  $h \leq 1$  km deep), the hydraulic radius  
 197  $R_c$  can be replaced by the depth of the water flow  $h$ .

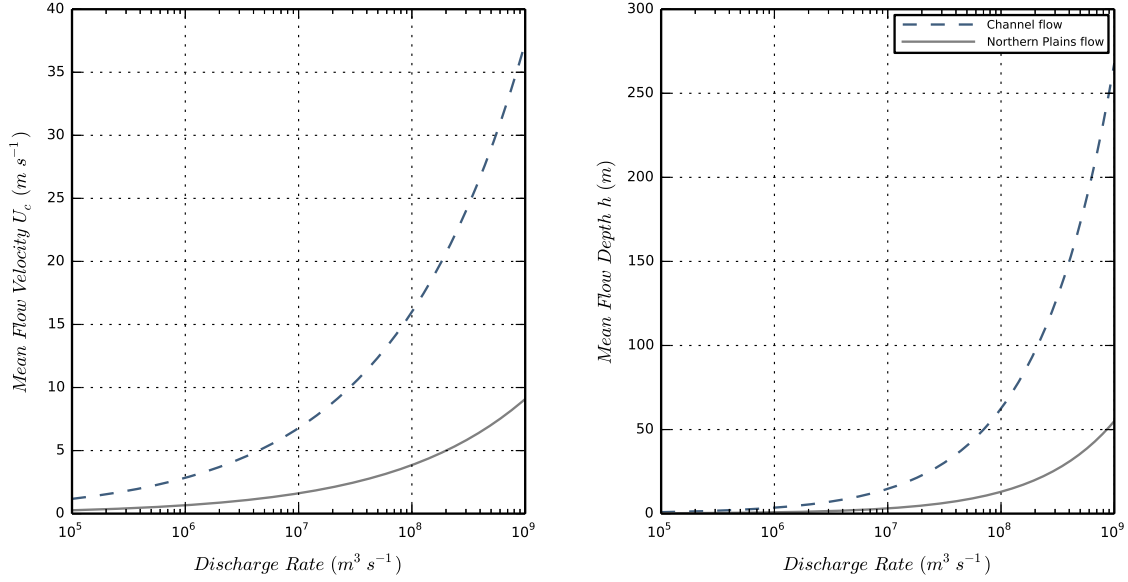


Figure 1: Estimates of mean flow velocity (left) and mean flow depth (right) for  
 1. (blue) the case of a 100 km-wide channel flow on an  $\sim 0.1^\circ$  slope angle and  
 2. (grey) the case for the same flow spilling onto the Northern Plains of Mars  
 ( $\sim 2000$  km-wide and slope angle  $\sim 0.03^\circ$ ). These quantities were calculated for  
 a wide range of water discharge rates, using the Darcy-Weisbach equation.

198 To estimate the velocity of the flow according to its discharge rate  $Q = U_c W_c h$ ,  
 199 we solve equation (4) using the Lambert special function  $W$  defined by  $x = W(xe^x)$ .  
 200 We obtain:

$$h = \left( \frac{3 \ln 10}{2a W_c (g \sin \alpha)^{\frac{1}{2}}} \frac{Q}{W\left(\frac{3 \times 10^{\frac{3b}{2a}} \ln 10}{2a z_c^{\frac{3}{2}} W_c (g \sin \alpha)^{\frac{1}{2}}} Q\right)} \right)^{\frac{2}{3}} \quad (6)$$

201 and

$$U_c = \left( \frac{2a (g \sin \alpha)^{\frac{1}{2}} W\left(\frac{3 \times 10^{\frac{3b}{2a}} \ln 10}{2a z_c^{\frac{3}{2}} W_c (g \sin \alpha)^{\frac{1}{2}}} Q\right)}{3 \ln 10 W_c^{\frac{1}{2}}} \right)^{\frac{2}{3}} Q^{\frac{1}{3}}. \quad (7)$$

202 The high concentrations of sediments in the flows (up to 40% of the volume)  
 203 can increase the volumetric mass density  $\rho$  (initially of  $\rho_{\text{water}} \sim 1000 \text{ kg m}^{-3}$ ) by a

204 factor of 2 and the viscosity  $\mu$  (initially of  $\mu_{\text{water},300\text{K}} \sim 8 \times 10^{-4}$  Pa s) by a factor of  
205 16 (Andrews-Hanna and Phillips, 2007), reducing by almost 10 the corresponding  
206 Reynolds Number. Nonetheless, since both the sediment load (from 0.1 to 40 %) and  
207 the dependence of the friction factor  $f_c$  on the Reynolds Number  $Re$ , are  
208 poorly known (Andrews-Hanna and Phillips, 2007), their effects were not taken  
209 into account in the flow depth/velocity calculations.

210 2. As soon as the water flow leaves its channel and reaches Chryse Planitia, the  
211 width of the flow strongly increases (up to 2000 km) and the slope angle decreases  
212 down to  $0.03^\circ$ . The mean flow velocity and height both decrease (Figure 1)  
213 whereas the wetted area increases significantly, leading to even more evaporation.  
214 The water will eventually end up in the main topographic depression of Vastitas  
215 Borealis (around  $-30^\circ/70^\circ$  in longitude/latitude) building up with time. If the  
216 volume of water released by the outflow channel event is higher than  $\sim 2.6 \times 10^6$  km<sup>3</sup>,  
217 the water will spill from the North Polar basin to the Utopia Basin, filling it  
218 potentially up to  $1.1 \times 10^6$  km<sup>3</sup> (Ivanov and Head, 2001). If the volume of water  
219 exceeds  $3.7 \times 10^6$  km<sup>3</sup>, the two basins become connected. They can be filled up to  
220 few tens of millions of km<sup>3</sup>.

221 Once the flow stops, some water will possibly remain in local topographic  
222 depressions such as impact craters or tectonic basins, thereby contributing to  
223 extended evaporation.

224 If the volume of water or the temperature of the flow are too low, the liquid  
225 water flow can potentially freeze before reaching the lowest points of the northern  
226 lowlands. This would likely occur only for the weakest outflow channel events  
227 (low volumes/discharge rates/temperatures), and we do not discuss this possibility  
228 further in this work.

## 229 2.2 Late Hesperian Climate

230 Late Hesperian Mars was likely to have been cold and dry globally, as suggested  
231 by the weak occurrence of well-developed valley networks (Carr, 1996; Fassett and  
232 Head, 2008; Harrison and Grimm, 2005), the absence of observed phyllosilicates  
233 within layered deposits (Bibring et al., 2006; Chevrier et al., 2007; Carter et al.,  
234 2013), and the low erosion rates inferred from impact craters morphologies (Craddock  
235 and Maxwell, 1993; Golombek et al., 2006; Quantin et al., 2015).

236 As suggested by the stability of liquid water, and as supported by using the  
237 size distribution of ancient craters (Kite et al., 2014), the atmosphere of Mars  
238 at the end of the Noachian epoch was likely to have been thicker than the  $\sim$   
239 8 mbar present day atmosphere. From the Noachian-Hesperian transition to the

240 Late Hesperian era, magmatism may have been responsible for the build up of  
241 up to 400 mbar of CO<sub>2</sub> in the atmosphere (Grott et al., 2011, Figure 4). In  
242 fact, it is during the period of formation of the outflow channels that the release  
243 of gaseous CO<sub>2</sub> could have been at its maximum (Baker et al., 1991; Gulick  
244 et al., 1997): 1. Up to 100 mbars of CO<sub>2</sub> could have been released by the  
245 contemporaneous Tharsis volcanism; 2. up to 60 mbars of CO<sub>2</sub> per volume of  
246 10<sup>6</sup> km<sup>3</sup> of outflow waters if produced by clathrate destabilization; and 3. up  
247 to 20 mbars of CO<sub>2</sub> per volume of 10<sup>6</sup> km<sup>3</sup> of outflow waters if coming from  
248 highly pressurized groundwater reservoirs saturated in CO<sub>2</sub>. However, most recent  
249 estimates of the several CO<sub>2</sub> loss processes (photochemical escape, effect of solar  
250 wind, sputtering, impact erosion, loss to carbonates, etc.; summarized in Forget  
251 et al. (2013, section 3)) suggest that, in spite of the previously mentioned high  
252 estimates of CO<sub>2</sub> outgassing amounts, it is very unlikely that the atmosphere of  
253 Late Hesperian Mars was thicker than 1 bar. In other words, there are currently  
254 no known physical/chemical processes that could accommodate the loss of an  
255 atmosphere at pressures of more than 1 bar.

256 To summarize, the Late Hesperian atmosphere was probably thicker than 8 mbar  
257 and thinner than 1 bar, but the actual surface pressure is still a matter of debate.  
258 In this paper, we find that the thickness of Late Hesperian Mars atmosphere plays  
259 an important role in relation to the climatic impact of outflow channel formation  
260 events. We chose to explore a wide possibility of atmospheric surface pressures,  
261 ranging from 40 mbar to 1 bar.

## 262 **3 Model description**

### 263 **3.1 The Late Hesperian Global Climate Model**

264 In this paper we use the 3-Dimensions LMD Generic Global Climate Model,  
265 specifically developed for the study of the climate of ancient Mars (Forget et al.,  
266 2013; Wordsworth et al., 2013), and adapted here for the study of the influence of  
267 outflow channel events on Mars climate during the Late Hesperian.

268 This model is originally derived from the LMDz 3-dimensional Earth Global  
269 Climate Model (Hourdin et al., 2006), which solves the basic equations of geophysical  
270 fluid dynamics using a finite difference dynamical core on an Arakawa C grid.

271 The same model has been used to study many different planetary atmospheres  
272 including Archean Earth (Charnay et al., 2013), a highly irradiated 'future' Earth  
273 (Leconte et al., 2013a), Pluto (Forget et al., 2014), Saturn (Guerlet et al., 2014);

274 Spiga et al., 2015) and exoplanets (Wordsworth et al., 2011; Leconte et al., 2013b;  
275 Forget and Leconte, 2014; Bolmont et al., 2016; Turbet et al., 2016a,b).

276 Most of the simulations presented in this paper were performed at a spatial  
277 resolution of  $96 \times 48$  (e.g.  $3.75^\circ \times 3.75^\circ$ ; at the equator, this gives in average  
278  $220 \text{ km} \times 220 \text{ km}$ ) in longitude / latitude. This corresponds approximately to  
279 twice the horizontal resolution used and eight times the calculation time needed  
280 in the work done by Forget et al. (2013) and Wordsworth et al. (2013). For  
281 this reason, a parallelized version of the GCM was used to deal with the long  
282 computation times. We explored the influence of the horizontal resolution (up to  
283  $1^\circ \times 1^\circ / 360 \times 180$  grid in longitude / latitude) and did not find any significant  
284 discrepancy compared with the  $96 \times 48$  lower resolution simulations.

285 In the vertical direction, the model is composed of 15 distinct atmospheric  
286 layers, generally covering altitudes from the surface to  $\sim 50 \text{ km}$ . Hybrid  $\sigma$  coordinates  
287 (where  $\sigma$  is the ratio between pressure and surface pressure) and fixed pressure  
288 levels were used in the lower and the upper atmosphere, respectively. The lowest  
289 atmospheric mid-layers are located around [18, 40, 100, 230, ..] meters and the  
290 highest at about [.., 20, 25, 35, 45] kilometers.

291 We used the present-day MOLA (Mars Orbiter Laser Altimeter) Mars surface  
292 topography (Smith et al., 1999; Smith et al., 2001), and we considered that most of  
293 the Tharsis volcanic load was largely in place by the end of the Hesperian epoch  
294 (Phillips et al., 2001).

295 We set the obliquity of Mars at  $45^\circ$  to be consistent with both the most likely  
296 obliquity ( $41.8^\circ$ ) for ancient Mars calculated by Laskar et al. (2004) and one of  
297 the reference obliquities ( $45^\circ$ ) used in Wordsworth et al. (2013). The sensitivity  
298 of obliquity (and more generally of the seasonal effects) is discussed in section 4.

299 To account for the thermal conduction in the subsurface, we use an 18-layer  
300 thermal diffusion soil model that originally derives from Hourdin et al. 1993 and  
301 has been modified to take into account soil layers with various conductivities. The  
302 mid-layer depths range from  $d_0 \sim 0.1 \text{ mm}$  to  $d_{17} \sim 18 \text{ m}$ , following the power law  
303  $d_n = d_0 \times 2^n$  with  $n$  being the corresponding soil level, chosen to take into account  
304 both the diurnal and seasonal thermal waves. We assumed the thermal inertia of  
305 the Late Hesperian martian regolith to be constant over the entire planet and equal  
306 to  $250 \text{ J m}^{-2} \text{ s}^{-1/2} \text{ K}^{-1}$ . This is slightly higher than the current Mars global mean  
307 thermal inertia in order to account for the higher atmospheric pressure.

308 Subgrid-scale dynamical processes (turbulent mixing and convection) were  
309 parameterized as in Forget et al. (2013) and Wordsworth et al. (2013). The planetary  
310 boundary layer was accounted for by the Mellor and Yamada (1982) and Galperin  
311 et al. (1988) time-dependent 2.5-level closure scheme, and complemented by a



Physical parameters	Values
Mean Solar Flux	465 W m <sup>-2</sup> (79% of present-day)
Obliquity	45°
Bare ground Albedo	0.2
Liquid water Albedo	0.07
H <sub>2</sub> O and CO <sub>2</sub> ice Albedos	0.5
Surface Topography	Present-day
Surface Pressure	0.2 bar
Surface roughness coefficient	0.01 m
Ground thermal inertia	250 J m <sup>-2</sup> s <sup>-1/2</sup> K <sup>-1</sup>

Table 1: Physical Parameterization of the GCM.

312 convective adjustment which rapidly mixes the atmosphere in the case of unstable  
313 temperature profiles (see section 3.1.3 for more details).

314 In the simulations that include outflow channel events, the dynamical time step  
315 is ~ 45 seconds (respectively ~ 184 s for the control simulations). The radiative  
316 transfer and the physical parameterizations are calculated every ~ 15 minutes  
317 and ~ 4 minutes (respectively every ~ 1 hour and ~ 15 minutes for the control  
318 simulations).

### 319 3.1.1 Radiative Transfer in a CO<sub>2</sub>/H<sub>2</sub>O mixed atmosphere.

320 The GCM includes a generalized radiative transfer for a variable gaseous atmospheric  
321 composition made of a mix of CO<sub>2</sub> and H<sub>2</sub>O (HITRAN 2012 database, Rothman  
322 et al. (2013)) using the 'correlated-k' method (Fu and Liou, 1992)) suited for fast  
323 calculation. For this, we decomposed the atmospheric Temperatures / Pressures /  
324 Water Vapor Mixing Ratio into the following respective 7 x 8 x 8 grid: Temperatures  
325 = {100,150, .. ,350,400} K; Pressures = {10<sup>-6</sup>,10<sup>-5</sup>, .. ,1,10} bar; H<sub>2</sub>O Mixing Ratio  
326 = {10<sup>-7</sup>,10<sup>-6</sup>, .. ,10<sup>-2</sup>,10<sup>-1</sup>,1 } mol of H<sub>2</sub>O / mol of air (H<sub>2</sub>O+CO<sub>2</sub> here).

327 CO<sub>2</sub> collision-induced absorptions (Gruszka and Borysow, 1998; Baranov et al.,  
328 2004; Wordsworth et al., 2010) ) were included in our calculations as in Wordsworth  
329 et al. (2013), as well as the H<sub>2</sub>O continuums. For this, we used the CKD model  
330 (Clough et al., 1989) with H<sub>2</sub>O lines truncated at 25 cm<sup>-1</sup>.

331 For the computation, we used 32 spectral bands in the thermal infrared and  
332 35 in the visible domain. 16 non-regularly spaced grid points were used for  
333 the g-space integration, where g is the cumulative distribution function of the  
334 absorption data for each band. We used a two-stream scheme (Toon et al., 1989)  
335 to take into account the radiative effects of aerosols (CO<sub>2</sub> ice and H<sub>2</sub>O clouds) and  
336 the Rayleigh scattering (mostly by CO<sub>2</sub> molecules), using the method of Hansen  
337 and Travis (1974).

338 In summary, compared to the radiative transfer calculation used in Wordsworth  
339 et al. (2013), we utilized here a more recent spectroscopic database (HITRAN2012  
340 instead of HITRAN2008) and built new correlated-k coefficients suited for wet  
341 atmospheres (water vapor VMR up to 100%). In practice, the maximum water  
342 vapor Mass Mixing Ratio that was reached in our simulations (in the case of low  
343 surface pressure simulations) was ~ 20%.

344 In addition, we chose a mean solar flux of 465 W.m<sup>-2</sup> (79% of the present-day  
345 value of Mars; 35% of Earth's present-day value; and 105% of the flux used in the  
346 Wordsworth et al. (2013) work), corresponding to the reduced luminosity from  
347 standard solar evolution models (Gough, 1981) 3.0 Byrs ago, during the Late  
348 Hesperian era. During this epoch, the Sun was also 1.5 % cooler (Bahcall et al.,  
349 2001); we did not, however, include in our model the resulting shift in the solar  
350 spectrum.

351 It is worth nothing anyway that absolute ages are based here on crater counting  
352 and are therefore not well constrained. For instance, the valley networks observed  
353 in West Echus Chasma Plateau are 2.9 to 3.4 billion years old (Mangold et al.,  
354 2004).

### 355 3.1.2 CO<sub>2</sub> and Water cycles

356 Both CO<sub>2</sub> and H<sub>2</sub>O cycles are included in the GCM used in this work.

357 1. Carbon Dioxide is here the dominant gaseous species. In our model, CO<sub>2</sub>  
358 can condense to form CO<sub>2</sub> ice clouds and surface frost if the temperature drops  
359 below the saturation temperature. Atmospheric CO<sub>2</sub> ice particles are sedimented  
360 and thus can accumulate at the surface. The CO<sub>2</sub> ice layer formed at the surface  
361 can sublime and recycle the CO<sub>2</sub> in the atmosphere. The CO<sub>2</sub> ice on the surface  
362 contributes to the surface albedo calculation: if the CO<sub>2</sub> ice layer overpasses a  
363 threshold value of 1 mm thickness, then the local surface albedo is set immediately  
364 to the albedo of CO<sub>2</sub> ice (0.5 in this work).

365 2. A self-consistent H<sub>2</sub>O water cycle is also included in the GCM. In the  
366 atmosphere, water vapor can condense into liquid water droplets or water ice

367 particles, depending on the atmospheric temperature and pressure, forming clouds.  
368 At the surface, because the range of surface pressures modeled in this work are  
369 well above the triple point 6 mbar pressure, liquid water and water ice can coexist.  
370 Their contributions are both taken into account in the albedo calculation as in  
371 Wordsworth et al. (2013).

372 The stability of liquid water / ice / CO<sub>2</sub> ice at the surface is governed by the  
373 balance between radiative and sensible heat fluxes (direct solar insolation, thermal  
374 radiation from the surface and the atmosphere, turbulent fluxes) and thermal conduction  
375 in the soil. Melting, freezing, condensation, evaporation, sublimation and precipitation  
376 physical processes are all included in the model.

### 377 3.1.3 Convective Adjustment

378 Outflow channel events result in the emplacement of warm liquid water, which  
379 leads to the sudden and intense warming of the atmosphere. Global Climate  
380 Models (~ 200 km grid size for our simulations) are not suited to resolve the  
381 convection processes as is done in the case of mesoscale models, which have a  
382 typical km-size resolution (Kite et al., 2011a,b).

383 Moist convection was taken into account following a moist convective adjustment  
384 that originally derives from the 'Manabe scheme' (Manabe and Wetherald, 1967;  
385 Wordsworth et al., 2013). In our scheme, relative humidity is let free and limited to  
386 100%, since it is inappropriate here to use an empirical value for relative humidity  
387 (versus altitude) that comes from Earth observations, as proposed in the original  
388 scheme.

389 This scheme has been chosen instead of more refined ones because it is: 1.  
390 robust for a wide range of pressures; 2. energy-conservative; and 3. it is the  
391 most physically consistent scheme for exotic (non Earth-like) situations such as  
392 the ones induced by outflow channel events. In practice, when an atmospheric  
393 grid cell reaches 100% saturation and the corresponding atmospheric column has  
394 an unstable temperature vertical profile, the moist convective adjustment scheme  
395 is performed to get a stable moist-adiabatic lapse rate.

396 In our simulations, after major outflow channel events, large amounts of water  
397 vapor can be released into the atmosphere and the water vapor can easily become  
398 a dominant atmospheric species. In fact we recorded up to 20% water vapor Mass  
399 Mixing Ratios following intense outflow channels (in the case of low surface  
400 pressure). Thus, we used a generalized formulation of the moist-adiabatic lapse  
401 rate developed by Leconte et al. (2013a) (Supplementary Materials) to account  
402 for the fact that water vapor can become a main species in our simulations.

403 In our model we also used the numerical scheme proposed by Leconte et al.  
404 (2013a) to account for atmospheric mass change after the condensation or the  
405 evaporation of gases (water vapor in our case); this calculation is usually neglected  
406 in most of the well-known Global Climate Models. More details on the scheme  
407 can be found in Leconte et al. (2013a) (Supplementary Materials). This scheme  
408 comes from previous work for the CO<sub>2</sub> cycle on present-day Mars (Forget et al.,  
409 1998), where there is some observational validation.

#### 410 **3.1.4 Parameterization of the precipitation events**

411 H<sub>2</sub>O precipitation events were parameterized using a simple cloud water content  
412 threshold scheme (Emanuel and Ivkovi-Rothman, 1999) as in Wordsworth et al.  
413 (2013). If the cloud water content overpasses a threshold  $l_0$  in a given atmospheric  
414 grid cell, precipitation occurs. We chose  $l_0$  to be constant and equal to 0.001 kg/kg  
415 as in Wordsworth et al. (2013). Wordsworth et al. (2013) examined the influence  
416 of  $l_0$  and found it to be very low (1K difference between  $l_0=0.001$  and 0.01 kg/kg).

417 We note that the reevaporation of the precipitation is also taken into account  
418 in our numerical scheme.

### 419 **3.2 Control Simulations without outflow events**

420 We performed control simulations in the conditions described above for 5 different  
421 surface pressures (40 mbar, 80 mbar, 0.2 bar, 0.5 bar, 1 bar) and we obtained  
422 results which are consistent with Wordsworth et al. (2013) and Forget et al. (2013).  
423 For these control runs, the three main differences between our work and Wordsworth  
424 et al. (2013) were: 1. the updated absorption coefficients (now HITRAN 2012);  
425 2. an increase of the solar luminosity (now 79% of Mars present-day value); and  
426 3. the increase of the horizontal model resolution (from 32 x 32 to 96 x 48 in  
427 longitude x latitude).

428 Figure 2 shows the mean annual surface temperatures and the position of the  
429 stable ice deposits for the reference case (0.2 bar) and the two surface pressure  
430 endmembers (40 mbar and 1 bar). The mean annual surface temperatures are  
431 slightly lower than in Figure 3 in Wordsworth et al. (2013) which were obtained  
432 for a fixed 100% relative humidity. It is also perhaps due to a slightly reduced CO<sub>2</sub>  
433 ice cloud warming effect at high spatial resolution. The stable surface ice deposit  
434 locations were calculated using the ice equilibration algorithm of Wordsworth  
435 et al. (2013). Starting from a random initial surface ice distribution, (1) we run  
436 the GCM for two martian years then (2) we extrapolate the ice layer field  $h_{ice}$

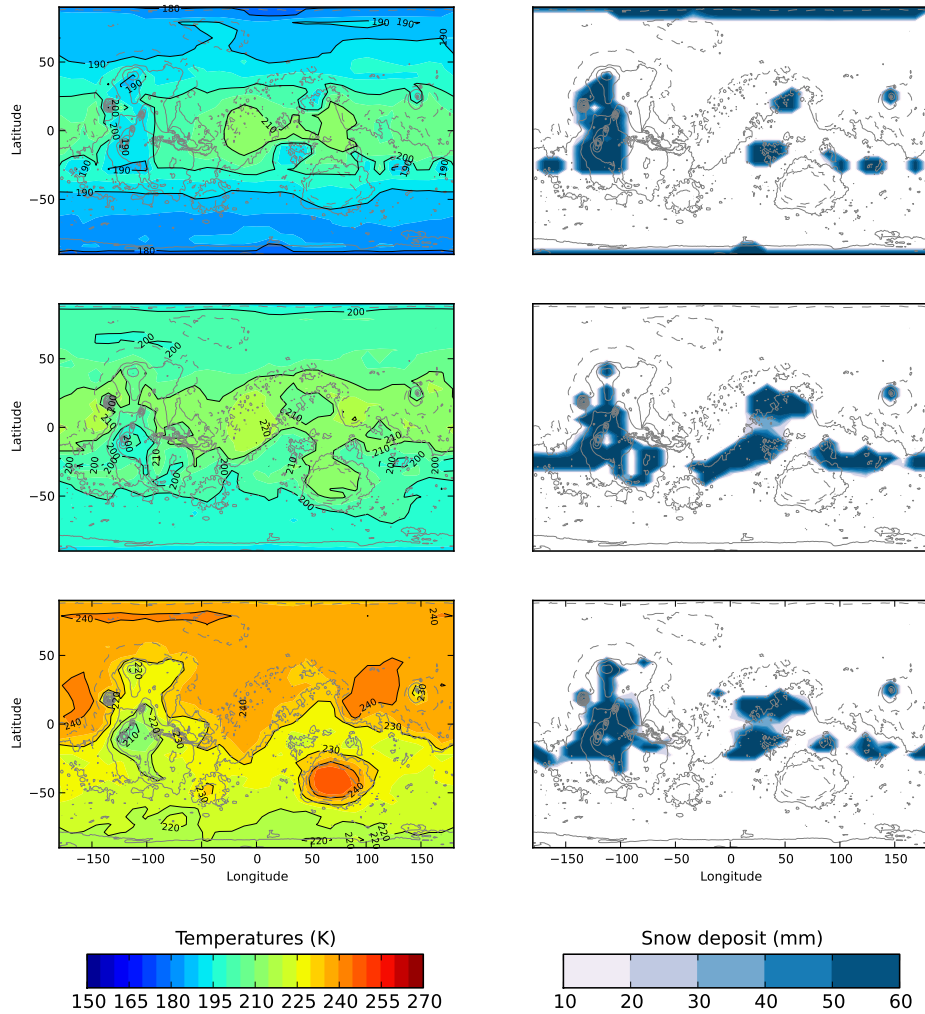


Figure 2: Surface Temperatures (left) and H<sub>2</sub>O ice deposit (right) annual means for the control simulations (at a 96 x 48 horizontal resolution) after ~ 800 martian years (or 30 loops of the ice iteration scheme). Top, middle and bottom plots correspond respectively to control simulations performed at 40 mbar, 0.2 bar and 1 bar. Grey contours show the topography used in the simulation. Ice iteration was performed every 2 years, with a 100-year timestep used for the first five iterations and 10-year timesteps used thereafter.

437 evolution calculation using:

$$h_{\text{ice}}(t + n_{\text{years}}) = h_{\text{ice}}(t) + n_{\text{years}} \times \Delta h_{\text{ice}}, \quad (8)$$

438 with  $\Delta h_{\text{ice}}$  the annual mean ice field change of the one-martian-year previous  
 439 simulation and  $n_{\text{years}}$  the number of years requested for the extrapolation. Then,  
 440 (3) we eliminate of seasonal ice deposit and (4) we normalize the extrapolated ice  
 441 field by the initial ice inventory to conserve the total ice mass. Eventually, (5) we  
 442 repeat the process.

443 This algorithm has been shown (Wordsworth et al., 2013) to be insensitive to  
 444 the proposed initial ice field location at the beginning of the simulation, at least  
 445 assuming that the scheme has been repeated a sufficient number of times.

446 In total, for our control simulations, we performed the scheme 30 times, with  
 447  $n_{\text{years}}=100$  for the first 5 loops and  $n_{\text{years}}=10$  for 20 more loops for a resolution of  
 448  $32 \times 32$ . Then, we ran the algorithm 5 more times at the increased resolution of  
 449  $96 \times 48$  to obtain a stable initial state necessary for the implementation of outflow  
 450 channel events.

451 We note that 3D climate modeling under conditions similar to those described  
 452 above (Forget et al., 2013; Wordsworth et al., 2013) have not yet been able to  
 453 produce liquid water or at least significant precipitation by climatic processes  
 454 anywhere on the planet, even when maximizing the greenhouse effect of  $\text{CO}_2$   
 455 ice clouds.

### 456 3.3 Experiment - Modeling of Outflow Channel Events

#### 457 3.3.1 Description of the parameterization

458 Outflow channel events can be modeled to a first approximation by the sudden  
 459 release, and then the spread of warm liquid water over the surface of Mars. In our  
 460 simulations, this was accomplished by the emplacement of a fully mixed layer of  
 461 warm liquid water at the surface. The fate of this water depends on the following  
 462 processes (summarized in Figure 3):

463 1. The liquid water layer loses some energy by thermal conduction to the  
 464 initially cold ground. For this, we fix the uppermost of the 18th martian regolith  
 465 layers at the temperature of the water, and calculate the heat flux lost (or gained)  
 466 by the warm water to the downward layers.

467 2. The warm liquid water layer cools by emitting thermal infrared radiation at  
 468  $\sigma T_{\text{surf}}^4$ . This emission contributes to the radiative transfer budget.

469 3. The liquid water evaporates and loses some latent heat. The evaporation  
 470  $E$  at the location of the warm water was computed within the boundary-layer  
 471 scheme, using the following bulk aerodynamic formula:

$$E = \rho_1 C_d V_1 [q_{\text{sat}}(T_{\text{surf}}) - q_1], \quad (9)$$

472 where  $\rho_1$  and  $V_1$  are the volumetric mass of air and the wind velocity at the first  
 473 atmospheric level,  $q_{\text{sat}}(T_{\text{surf}})$  is the water vapor mass mixing ratio at saturation  
 474 at the surface, and  $q_1$  is the mixing ratio in the first atmospheric layer. The  
 475 aerodynamic coefficient is given by  $C_d = (\kappa / \ln(1 + z_1/z_0))^2 \sim 2.5 \times 10^{-3}$ , where  
 476  $\kappa = 0.4$  is the Von Karman constant,  $z_0$  is the roughness coefficient and  $z_1$  is the  
 477 altitude of the first level ( $\sim 18$  meters).

478 We modeled the sensible heat exchanged between the surface and the first  
 479 atmospheric layer using a similar formula:

$$F_{\text{sensible}} = \rho_1 C_p C_d V_1 [T_{\text{surf}} - T_1], \quad (10)$$

480 with  $T_1$  the temperature of the first atmospheric level and  $C_p$  the mass heat capacity  
 481 assumed equal to  $850 \text{ J K}^{-1} \text{ kg}^{-1}$  in case of a  $\text{CO}_2$ -only atmospheric composition.

482 4. Depending on the volume of water modeled, liquid water will flow from  
 483 the Circum-Chryse outflow channel sources to Chryse Planitia, then to Acidalia  
 484 Planitia, and eventually to the Northern Plains. First, we modeled the displacement  
 485 of the flow calculated from its height and its velocity. The velocity of the flow  
 486 mostly depends on its width but also on the slope of the terrain. For each grid,  
 487 we used the subgrid mean slope and the subgrid mean orientation of the slope to  
 488 evaluate (using equations (4) and (7)) the velocity and the direction of the flow.  
 489 Second, we used a simple bucket scheme to model the progressive filling of the  
 490 topographic depressions.

491 Warm waters flowing on the Northern Plain slopes can also encounter  $\text{H}_2\text{O}$   
 492 ice (it can be either stable at a particular latitude, or related to previous outflow  
 493 channel events, but from the point of view of latent heat exchange and climate, it  
 494 does not change anything) or seasonal  $\text{CO}_2$  ice (typically present for atmospheres  
 495 thinner than 1 bar). We modeled the interaction of  $\text{H}_2\text{O}$  and  $\text{CO}_2$  ices with warm  
 496 liquid water using energy conservation. If the liquid water is warm and in a  
 497 sufficient amount, all the  $\text{CO}_2$  ice sublimates and is added to the atmosphere.  
 498 Similarly, all the water ice encountered by the warm flow is melted and converted  
 499 at the resulting equilibrium temperature.

500 Once the flow has reached a stable position (e.g. forming a lake), in reality  
 501 some water may be trapped in local topographic depressions (impact craters,

502 tectonic basins, ...); it is difficult, however, to estimate adequately how much water  
 503 might be sequestered in this manner. First, the detailed topography of the terrains  
 504 is unknown prior to resurfacing by the outflow channel events. Second, the water  
 505 outflows themselves modified (and probably smoothed) the topography. Thus, to  
 506 take into account not only the effect of the trapped water but also the role of the  
 507 wet ground, we arbitrarily placed a minimum 20 cm layer of liquid water in all  
 508 the locations where the liquid water flow passed through. This assumption may  
 509 also be representative of the fact that in reality the discharge rate does not have a  
 510 rectangular shape (in time) as we assumed in our parameterizations.

511 5. As time goes on, the liquid water flow cools. If its temperature reaches the  
 512 273.15 K freezing temperature (assuming no salts), the water starts to freeze. On  
 513 Earth, salinity drives the freezing point of oceans to  $\sim 271\text{K}$  and assuming similar  
 514 salt rates in outflow waters would not change much our results. To account for  
 515 this process, we developed a multiple layer modified version of the soil thermal  
 516 conduction model already included in the GCM. We have in total 100+ layers,  
 517 with mid-layer depths ranging from  $d_0 \sim 0.1\text{ mm}$  to  $d_{14} \sim 2\text{ m}$ , following the  
 518 power law  $d_{n,n \leq 14} = d_0 \times 2^n$  with  $n$  being the corresponding soil level and the  
 519 linear law  $d_{n,n > 14} = d_{14} \times (n - 13)$  for the deepest layers. The layers are separated  
 520 into two parts: the ice cover above and the liquid water below. For the water ice  
 521 layers, we use a thermal conductivity of  $2.5\text{ W m}^{-1}\text{ K}^{-1}$  and a volumetric heat  
 522 capacity of  $2 \times 10^6\text{ J m}^{-3}\text{ K}^{-1}$ . For the liquid water, we use, respectively, a thermal  
 523 inertia of  $20000\text{ J m}^{-2}\text{ K}^{-1}\text{ s}^{-1/2}$  (artificially high to account for convection) and  
 524 a volumetric heat capacity of  $4 \times 10^6\text{ J m}^{-3}\text{ K}^{-1}$ . At each physical timestep, we  
 525 estimate the thermal diffusion flux lost by the liquid water layer to the water ice  
 526 layer and calculate (using the conservation of energy) the amount of liquid water  
 527 to freeze. If the depth of the ice - initially going down to  $d = d_n$  - overpasses the  
 528 layer  $d_{n+1}$ , we convert the  $n + 1$  layer into ice.

529 We note that the use of a multi-layer soil model is important to describe  
 530 the sea-ice formation, evolution and its impact on possible cold early martian  
 531 climates. Such refined models are better suited to represent the temperature profile  
 532 evolution within the ice layer (that may evolve with seasonal forcing or as the ice  
 533 layer thickens) and thus the surface temperature that controls the sublimation rate.  
 534 In particular, our simulations show that up to 95% of the annual sublimation rate  
 535 can be produced during the summer seasons. This requires a good estimate of the  
 536 seasonal variations of the surface temperature above the ice.

537 Simultaneously, as the ice layer forms, we also linearly increase the surface  
 538 albedo from  $A_{\text{liq}} = 0.07$  (if no ice) to  $A_{\text{ice}} = 0.5$  (if the ice layer thickness  $h$   
 539 overpasses the threshold value of  $h_* = 3.5\text{ cm}$ ; Le Treut and Li 1991) as follows:



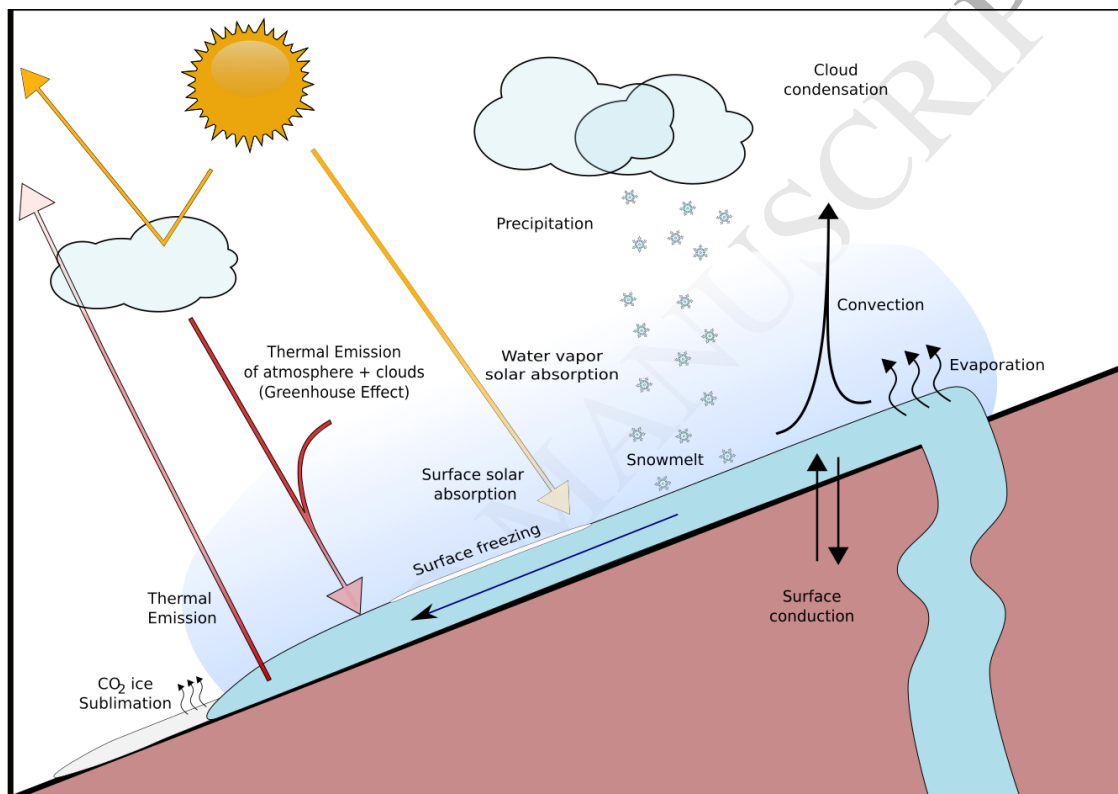


Figure 3: Schematic drawing of the physical processes taken into account during outflow channel events in our GCM simulations.

$$A = A_{\text{liq}} + (A_{\text{ice}} - A_{\text{liq}}) \frac{h}{h_*}. \quad (11)$$

540 6. The amount of water delivered by outflow events can be very large and thus  
 541 lead to the accumulation of large quantities of liquid water. The timing expected  
 542 for this water to freeze can be evaluated using a combination of the thermal  
 543 conduction flux in the ice layer  $F = \lambda_{\text{ice}} \frac{(T_{\text{surf}} - T_{\text{bottom}})}{h}$  and the conservation of  
 544 energy. Assuming that the temperature in the frozen layer varies linearly between  
 545  $T_{\text{bottom}} = 273.15$  K and  $T_{\text{surf}}$  (assumed constant) as hypothesized in classical 2-layers  
 546 thermodynamical models (Codron, 2012), we have:

$$\rho_{\text{ice}} (L_m - C_{\text{ice}} \frac{(T_{\text{bottom}} - T_{\text{surf}})}{2}) \frac{\partial h}{\partial t} = \lambda_{\text{ice}} \frac{(T_{\text{bottom}} - T_{\text{surf}})}{h}, \quad (12)$$

547 where  $\rho_{\text{ice}}$  is the volumetric mass of the ice ( $9.2 \times 10^2$  kg m<sup>-3</sup>),  $C_{\text{ice}}$  is the specific  
 548 heat capacity of the ice ( $2.1 \times 10^3$  J kg<sup>-1</sup> K<sup>-1</sup>),  $\lambda_{\text{ice}}$  is the conductivity of the ice  
 549 ( $2.5$  W m<sup>-1</sup> K<sup>-1</sup>) and  $L_m \sim 3.34 \times 10^5$  J kg<sup>-1</sup> is the latent heat of fusion of water  
 550 ice.

551 This leads after integration over time to an expression of  $t(h)$ , the timing  
 552 required to freeze a layer of depth  $h$ :

$$t(h) = \frac{\rho_{\text{ice}}}{2 \lambda_{\text{ice}}} \left( \frac{L_m}{(T_{\text{bottom}} - T_{\text{surf}})} - \frac{C_{\text{ice}}}{2} \right) h^2. \quad (13)$$

553 For example, the outflow event presented in section 4 leads to the accumulation  
 554 of up to 600 meters of liquid water. A typical timescale (for  $T_{\text{surf}} \sim 200$  K) for this  
 555 water to freeze, according to equation 13, is  $\sim 4 \times 10^3$  martian years.

556 To account for such long timescales, we developed a modified version of the  
 557 ice iteration scheme presented above. (1) First, we run the GCM for a few years  
 558 then (2) every 2 years, we extrapolate the amount of ice that has locally condensed  
 559 and sublimed in the simulations by an arbitrary factor  $n_{\text{years}}$ . Simultaneously, (3)  
 560 we proceed to a linear extrapolation of the amount of frozen water/of the growth of  
 561 the ice layer thickness by the same factor  $n_{\text{years}}$ , using the conservation of energy.  
 562 We actually fit the  $t = f(h)$  function by straight lines of sizes multiple of  $n_{\text{years}}$ . In  
 563 the reference simulation presented in section 4, we performed first 5 martian years,  
 564 then we extrapolated every 2 years using  $n_{\text{years}} = [5, 5, 20, 20, 50, 50, 100, 100, 500, 500]$ .  
 565 After the extrapolation of the ice field/the ice layer depth is completed, (4) we  
 566 arbitrarily set the ground temperature profile (where liquid water remains) to be  
 567 linear, between  $T_{\text{bottom}} = 273.15$  K and  $T_{\text{surf}}$  calculated using the conservation  
 568 of energy. This is a way to take into account (at first order) the evolution of the

569 deepest ground layers that require very long timescales to stabilize their temperature  
 570 profiles. The year following the extrapolation is thus also useful to get back a  
 571 consistent temperature profile in the first layers (up to 15 meters typically).

572 7. Once the outflow water is completely frozen, we use again the ice iteration  
 573 scheme (see section 3.2) to get estimates of the timing required for the ice to reach  
 574 its stable positions.

## 575 **4 Results - the reference simulation**

576 We present in this section the results of simulations of outflow channel formation  
 577 events occurring in the largest of the Circum Chryse channels: Kasei Vallis. We  
 578 chose this particular location because 1. The Kasei Vallis outflow channel begins  
 579 in Echus Chasma, which is close to the West Echus Chasma Plateau valley networks;  
 580 and 2. Kasei Vallis is one of the largest outflow channels on Mars (Carr, 1996).

581 We focus first on a discharge of  $10^6 \text{ km}^3$  (6.9 meters of GEL - Global Equivalent  
 582 Layer) of liquid water heated at 300 Kelvins. Water is released at a constant rate  
 583 of  $10^9 \text{ m}^3 \text{ s}^{-1}$  in the region of Echus Chasma (see Figure 4 for the associated  
 584 flow). This event is an upper estimate (in volume, discharge rate and temperatures)  
 585 of the characteristics of outflow channel formation events (see section 2.1.1 for  
 586 references).

587 As explained in section 2.2, surface atmospheric pressure in the Late Hesperian  
 588 epoch is poorly constrained. Thus, we focus first on the case of a surface pressure  
 589 of 0.2 bar.

### 590 **4.1 Description of the flow**

591 A volume of  $10^6 \text{ km}^3$  of liquid water is released at the discharge rate of  $1 \text{ km}^3 \text{ s}^{-1}$ .  
 592 It takes approximately 1.1 martian days for the liquid water to travel from the  
 593 source of the flow (in Echus Chasma, at  $\sim 4^\circ \text{N}, -79^\circ \text{E}$ ) to the end of Kasei Vallis (at  
 594  $\sim 30^\circ \text{N}, -45^\circ \text{E}$ ), and 1.5 more days for the same flow to reach the main topographic  
 595 depression of the northern plains (at  $\sim 70^\circ \text{N}, -30^\circ \text{E}$ ). This corresponds, respectively,  
 596 to mean flow speeds of  $\sim 30 \text{ m s}^{-1}$  and  $\sim 16 \text{ m s}^{-1}$ , which are consistent with the  
 597 two endmembers values shown in Figure 1.

598 After  $\sim 11$  days, the source of ground water (located in Echus Chasma) becomes  
 599 inactive. Eventually, it takes approximately 20 martian days in this scenario for the  
 600 liquid water that has erupted in Echus Chasma to form a stable lake in the lowest  
 601 part of the Northern Plains. This lake extends over an area of 4.2 millions of  $\text{km}^2$

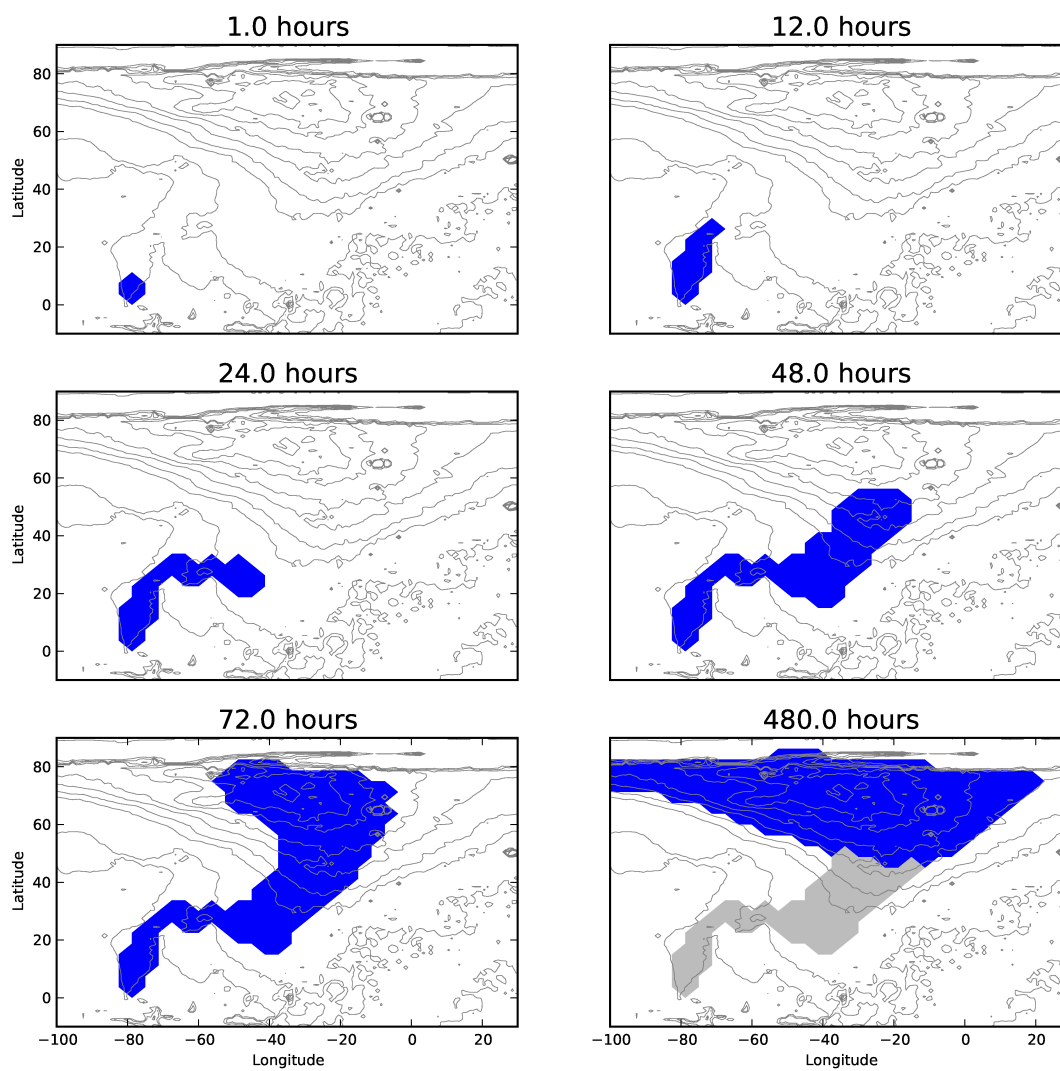


Figure 4: Time lapse of the runoff of the outflow channel event occurring in Echus Chasma, and flowing from Kasei Vallis down to the Northern Plains main topographic depression. The blue area corresponds to the position of the flow. The grey color was used to represent the 'wash' regions where the flow passed through but did not accumulate.

602 (~ 2.9% of the global surface area of Mars), has a mean depth of ~ 240 meters and  
603 a peak depth of ~ 600 meters. Some water (~ 20 centimeters) is left at locations  
604 with latitude < 50°N to account for the wet ground and the water possibly trapped  
605 in the topographic depressions.

606 The fate of the outflow channel formation event can be divided into two main  
607 parts:

608 1. During the first ~ 500 days following the event, the 'Warm Phase', an  
609 intense hydrological cycle takes place. The end of this phase approximately  
610 coincides with the time when the Northern Plains lake becomes fully covered  
611 by an ice layer.

612 2. During the following ~ 10<sup>5</sup> martian years, the martian climate is controlled  
613 by a weak and cold water cycle. It takes approximately the first  $4 \times 10^3$  years (as  
614 predicted by simple energy-balanced models; Kreslavsky and Head 2002) for the  
615 lake to be entirely frozen, and the rest to sublimate the lake completely and move  
616 the ice to its positions of equilibrium, assuming no ice gets buried below a lag  
617 deposit or gets transported through glacier flows.

## 618 4.2 The Warm Phase

619 As soon as the simulation starts, the warm 300 K liquid water released in Echus  
620 Chasma evaporates efficiently following equation 9, while flowing over the Northern  
621 Plains slopes. At the locations reached by the flow, which represent ~ 11 million  
622 km<sup>2</sup> (~ 7.5% of the global surface area of Mars), the evaporation rate can reach  
623 ~ 10<sup>-3</sup> kg m<sup>2</sup> s<sup>-1</sup> for tens of days. Figure 5 (left) shows the mean evaporation rate  
624 for the 4.2×10<sup>6</sup> km<sup>2</sup> Northern Plains stable lake formed by the outflow channel  
625 flood accumulation.

626 During the 500 days following the event, a global precipitable water amount  
627 of ~ 23 centimeters is evaporated by the liquid water flow. Evaporation of the  
628 lake accounts for 96 % of this amount (blue region in Figure 4, after 480 hours)  
629 and 4 % by the evaporation of the transient flow (grey region in Figure 4, after  
630 480 hours). This amount of cumulative evaporation corresponds to ~ 3.4 % of  
631 the initial volume of water ejected by the outflow event, which is approximately  
632 0.7 times the amount of evaporated water that would be expected if the extra  
633 thermal heat (compared to 273 K) of the 300 K flow was simply converted into  
634 latent heat.

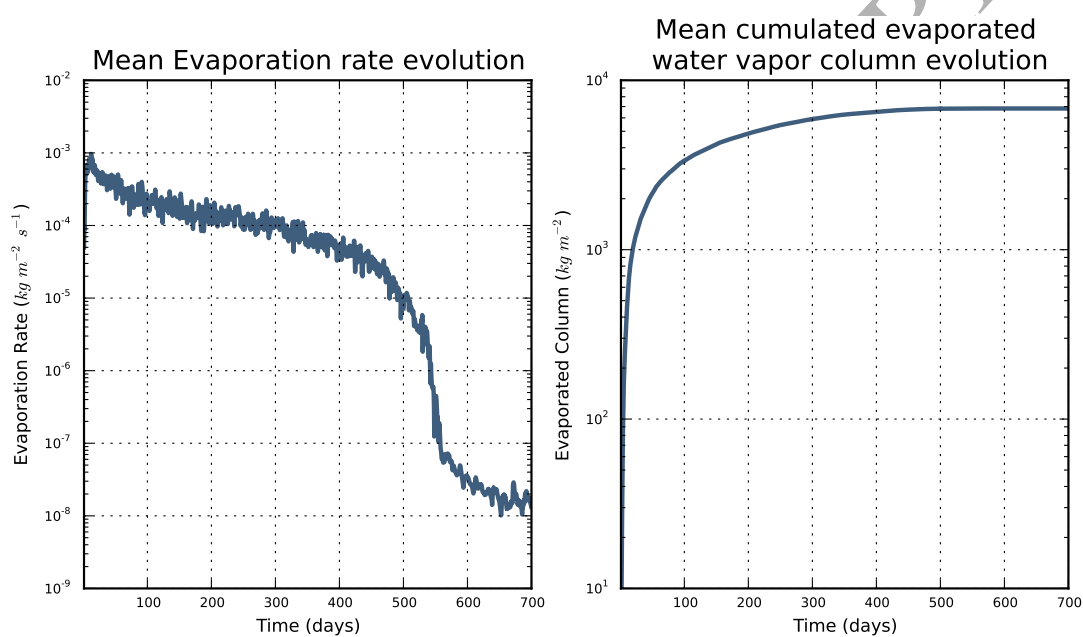


Figure 5: Mean lake evaporation rate evolution (left) and mean cumulative evaporated liquid water from the lake (right). The right curve is the cumulative integral over time of the left curve.

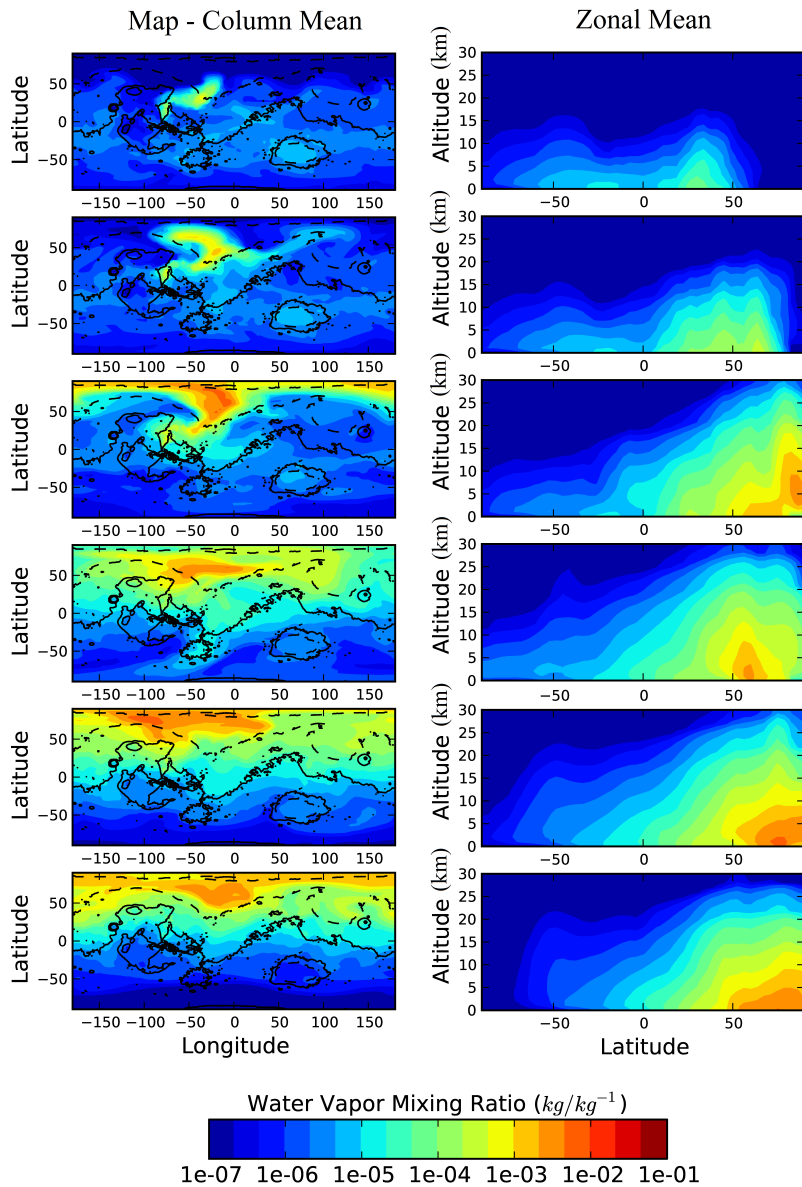


Figure 6: Time-lapse of the water vapor mixing ratio after (from the top to the bottom) 2.5/5/10/20/40/80 days. The left panels show the map of the water vapor distribution (column mean mixing ratio); the right panels show the corresponding zonal mean distribution (water vapor mixing ratio) as a function of latitude ( $^{\circ}N$ ) and altitude (km).

### 635 4.2.1 Mechanisms warming the atmosphere

636 As the water vapor starts to accumulate above the flow, the initially cold martian  
 637 lower atmosphere soon reaches the water vapor saturation pressure. For instance,  
 638 at 210 Kelvins, which is typically the mean surface temperature expected for a  
 639 0.2 bar atmosphere (Figure 2), the water vapor saturation pressure is  $\sim 1.4$  Pascals  
 640 and the mass mixing ratio at saturation in a 0.2 bar atmosphere is thereby  $\sim$   
 641  $7 \times 10^{-5}$  kg/kg $^{-1}$ . This situation leads to the early condensation of the water vapor,  
 642 latent heat release and thus to the warming of the atmosphere. We identified this  
 643 process as the dominant mechanism responsible for the warming of the atmosphere  
 644 after an outflow event.

645 As the atmospheric temperatures increase, the capability of the atmosphere to  
 646 retain water vapor also increases. The mass mixing ratio at saturation, namely  
 647  $Q_{\text{sat}}$ , can be written as follows:

$$Q_{\text{sat,H}_2\text{O}} = \frac{P_{\text{sat,H}_2\text{O}}}{P_{\text{CO}_2} + P_{\text{sat,H}_2\text{O}}}, \text{ with } P_{\text{sat,H}_2\text{O}}(T) = P_{\text{ref}} e^{\frac{L_v M_{\text{H}_2\text{O}}}{R} \left( \frac{1}{T_{\text{ref}}} - \frac{1}{T} \right)}, \quad (14)$$

648 with  $P_{\text{sat,H}_2\text{O}}$  the water vapor saturation pressure and  $P_{\text{CO}_2}$  the CO $_2$  partial pressure,  
 649 with  $P_{\text{ref}}$  and  $T_{\text{ref}}$  the pression/temperature of the triple point of water, respectively  
 650 equal to 612 Pascals/273.16 Kelvins,  $M_{\text{H}_2\text{O}} \sim 1.8 \times 10^{-2}$  kg mol $^{-1}$  the molar mass  
 651 of water, and  $L_v \sim 2.26 \times 10^6$  J kg $^{-1}$  the latent heat of vaporization of liquid water.  
 652 For low amounts of water, this relation simply becomes:

$$Q_{\text{sat,H}_2\text{O}}(T) \sim \frac{P_{\text{ref}}}{P_{\text{CO}_2}} e^{\frac{L_v M_{\text{H}_2\text{O}}}{R} \left( \frac{1}{T_{\text{ref}}} - \frac{1}{T} \right)}. \quad (15)$$

653 Therefore, as the atmospheric temperatures increase, the atmosphere is also  
 654 able to transport more and more water upwards. Thus, as time goes on, the  
 655 atmosphere becomes more and more warm and wet. As the atmospheric water  
 656 vapor content increases, the absorption of the atmosphere in the infrared wavelength  
 657 range (essentially due to the thermal emission of the warm outflow waters) increases  
 658 and thus contributes to an additional warming of the atmosphere.

659 In total, during the warm phase (the first 500 days), the atmosphere (above  
 660 the flow/lake) is directly warmed by the following processes (in decreasing order  
 661 of importance): 1. the condensation of the water vapor produced by the warm  
 662 flow ( $\sim 56$  %); 2. the sensible heat exchanged between the flow/lake and the  
 663 lowest atmospheric layer ( $\sim 22$  %); 3. the thermal infrared emission of the flow  
 664 absorbed by the mixture of gaseous CO $_2$ /H $_2$ O ( $\sim 13$  %); and 4. the extra solar  
 665 absorption resulting from the presence of water vapor excess, which has strong



666 absorption lines in the solar domain ( $\sim 9\%$ );. The atmospheric solar absorption  
 667 is particularly important in this scenario, because we chose the outflow channel  
 668 event to start at  $L_s = 5^\circ$  and thus to occur during the northern hemisphere spring  
 669 and summer. Of course, all these processes reinforce and strengthen each other.

670 Figure 6 shows the spatial evolution of the water vapor atmospheric content.  
 671 Initially, water vapor accumulates at low altitudes, in the regions where the liquid  
 672 water flow is located. After a few days, the water vapor has reached much higher  
 673 altitudes (up to  $\sim 30$  km) through the aforementioned warming mechanisms and  
 674 the convective adjustment scheme. Eventually, once the upper part of the atmosphere  
 675 has become wet enough (typically after  $\sim 10$  days in this scenario), the high  
 676 altitude horizontal winds (around  $\sim 15$  km) advect the water vapor into the neighbouring  
 677 regions. After  $\sim 50$  days, all the martian regions located above  $\sim 50^\circ N$  have  
 678 become more or less wet, with a typical water vapor mean mass mixing ratio of  
 679  $0.3\%$ .

680 Similarly, the impact of  $H_2O$  condensation (and other additional warming  
 681 sources) on atmospheric temperatures is shown in Figure 7. After  $\sim 100$  days,  
 682 at the peak of the outflow channel event, the atmospheric temperatures in the  
 683 lower atmosphere (0-5 km) almost reach 280 K, +90 Kelvins above the regular  
 684 temperature (peak above the lake) as calculated in the control simulation; the  
 685 atmospheric temperatures in the higher parts of the atmosphere typically extend  
 686 up to 230 Kelvins (at 10 km) and to 170 Kelvins (at 25 km), which are respectively  
 687 +50 K and +25 K above the temperatures prescribed by the control simulation.

#### 688 4.2.2 The mechanisms cooling the flow

689 After  $\sim 500$  days, which corresponds to the complete surface freezing of the  
 690 outflow channel event water, the evaporation  $E$  produced by the stable lake (see  
 691 Figure 5) suddenly reduces (by almost 3 orders of magnitude). To a first order, we  
 692 have in fact:

$$E \propto Q_{\text{sat}}(T) \propto e^{-\frac{\alpha}{T}}, \quad (16)$$

693 with  $\alpha = \frac{L_{\text{sub}} M_{H_2O}}{R}$  and  $L_{\text{sub}}$  the latent heat of sublimation of water ice. The  
 694 evaporation rate  $E$  has thereby a strong dependence on temperature. This is why  
 695 the drop in temperature associated with the surface freezing of the Northern Plains  
 696 lake is responsible for the sudden decrease of evaporation visible in Figure 5 (also  
 697 seen through the latent heat surface flux in Figure 8). This drop in evaporation  
 698 defines the end of the 'warm phase', which includes the decrease of the water  
 699 vapor content, the atmospheric temperatures and the precipitation events (see

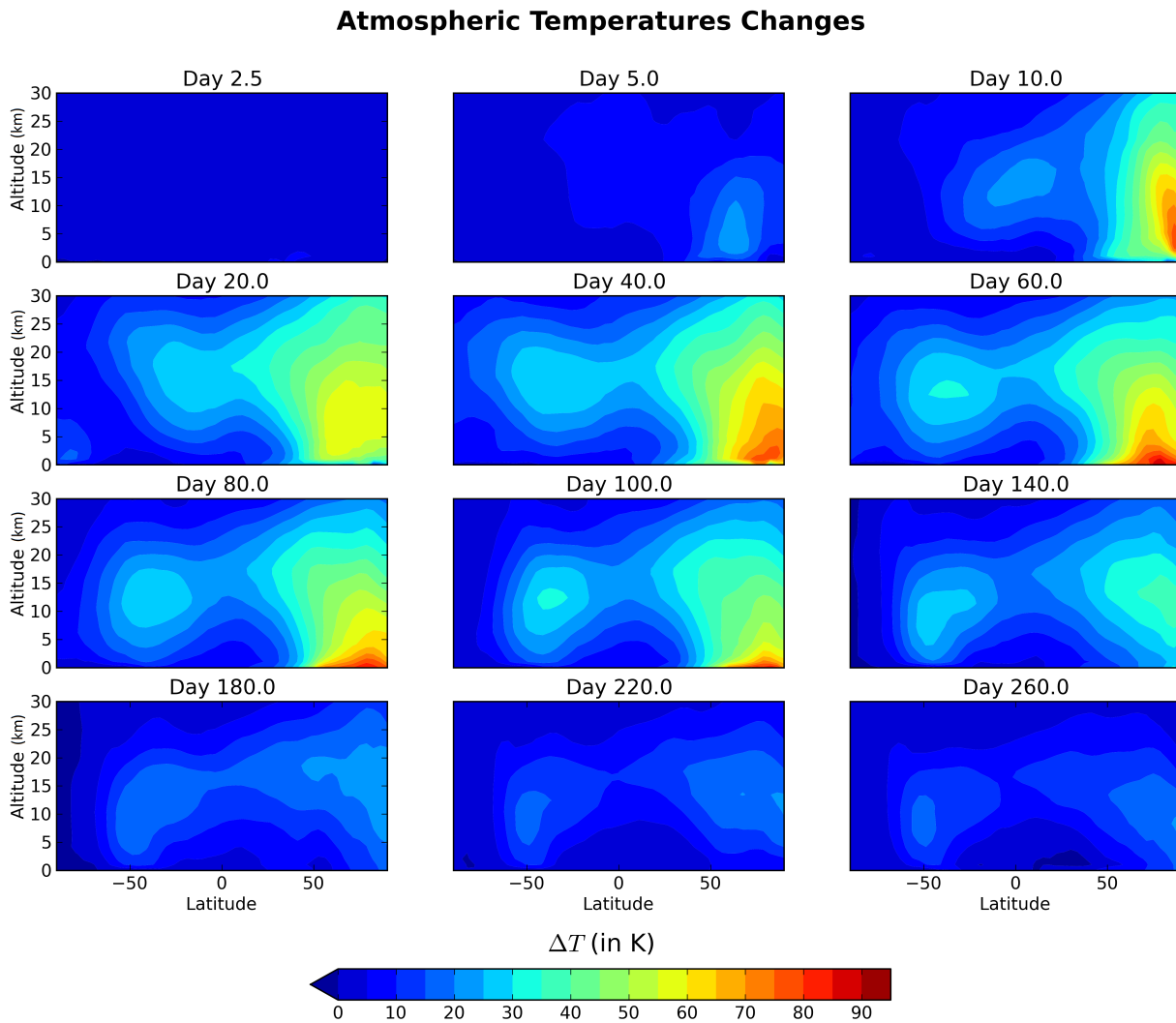


Figure 7: Time-lapse of the zonal mean cross-section atmospheric temperature difference between the reference simulation (with outflow) and the control simulation (without outflow) for the same surface pressure of 0.2 bar.

700 Figure 12).

701 There are several physical processes that are responsible for the cooling of the  
 702 flow, leading to its solidification as ice. Figure 8 shows the relative importance of  
 703 the different thermal heat losses by the Northern Plains lake, from the beginning  
 704 of the event to one martian year later. For the first 500 days, the main cooling  
 705 surface fluxes are the latent heat loss ( $420 \text{ W m}^{-2}$ , 43.3 %), the sensible heat loss  
 706 ( $190 \text{ W m}^{-2}$ , 19.6 %), the radiative thermal emission loss ( $280 \text{ W m}^{-2}$ , 28.8 %)  
 707 and the ground conduction loss ( $8 \text{ W m}^{-2}$ , 0.8 %). Some other surface fluxes  
 708 related to the  $\text{CO}_2$  ice sublimation by the warm flow ( $13 \text{ W m}^{-2}$ , 1.3 %) and the  
 709 cooling of the lake by the melting of the falling snow ( $60 \text{ W m}^{-2}$ , 6.2 %) also  
 710 contribute to the cooling of the outflow waters. In total, the average cooling flux  
 711 of the outflow waters for the warm phase (first 500 days) is  $\sim 970 \text{ W m}^{-2}$ .

712 For large outflow channel formation events like the one described in this section,  
 713 the sublimation of the seasonal carbon dioxide ice deposit represents a small  
 714 fraction of the heat loss. Nonetheless, smaller outflow channel events ( $5 \times 10^3 \text{ km}^3$   
 715 for example (Andrews-Hanna and Phillips, 2007)) flowing on the Northern Plains  
 716 slopes may be deeply affected by the energy gap required to sublimate the  $\text{CO}_2$   
 717 ice seasonal deposit. For a 0.2 bar atmosphere, the control simulations show, for  
 718 example, that the  $\text{CO}_2$  ice seasonal deposit reaches a yearly average of  $\sim 300 \text{ kg m}^{-2}$   
 719 from the North Pole down to  $30^\circ\text{N}$  latitudes.

720 Two radiative processes may counteract the cooling of the flow: 1) the absorption  
 721 of solar radiation and 2) the greenhouse effects (of the atmosphere and of the  
 722 clouds).

723 1. We chose in this scenario to start the outflow channel event at  $L_s = 5^\circ$  in  
 724 order to maximize the role of solar absorption. The peak of the event (between  
 725  $\sim 0$ -300 days,  $L_s \sim 5$ - $165^\circ$ ) was therefore chosen to overlap with the peak of  
 726 insolation in the Northern hemisphere, which is a maximum of  $\sim 170$  days after  
 727 the event ( $L_s = 90^\circ$ ). There are three factors that need to be taken into account  
 728 in the solar absorption processes: absorption by water vapor, albedo changes and  
 729 clouds. For this reference simulation, compared to the control simulation, these  
 730 three effects more or less compensate at the location of the flow. The increase of  
 731 the solar absorption due to the low albedo of liquid water (0.07 compared to 0.2  
 732 for the bare ground and 0.5 for the remaining  $\text{CO}_2$  ice seasonal cover) and due  
 733 to the absorption by water vapor are more or less balanced by the reflection of  
 734 the cloud cover, which can reach on average a coverage of 80 % during the first  
 735 500 days above the lake (Figure 11). Most of these water clouds are located at  
 736 low altitude (Figure 11). During the warm phase, the lake absorbs a solar flux  
 737 of  $\sim 67 \text{ W m}^{-2}$  ( $\sim 16 \text{ W m}^{-2}$  less than the control run, see Figure 9) and the

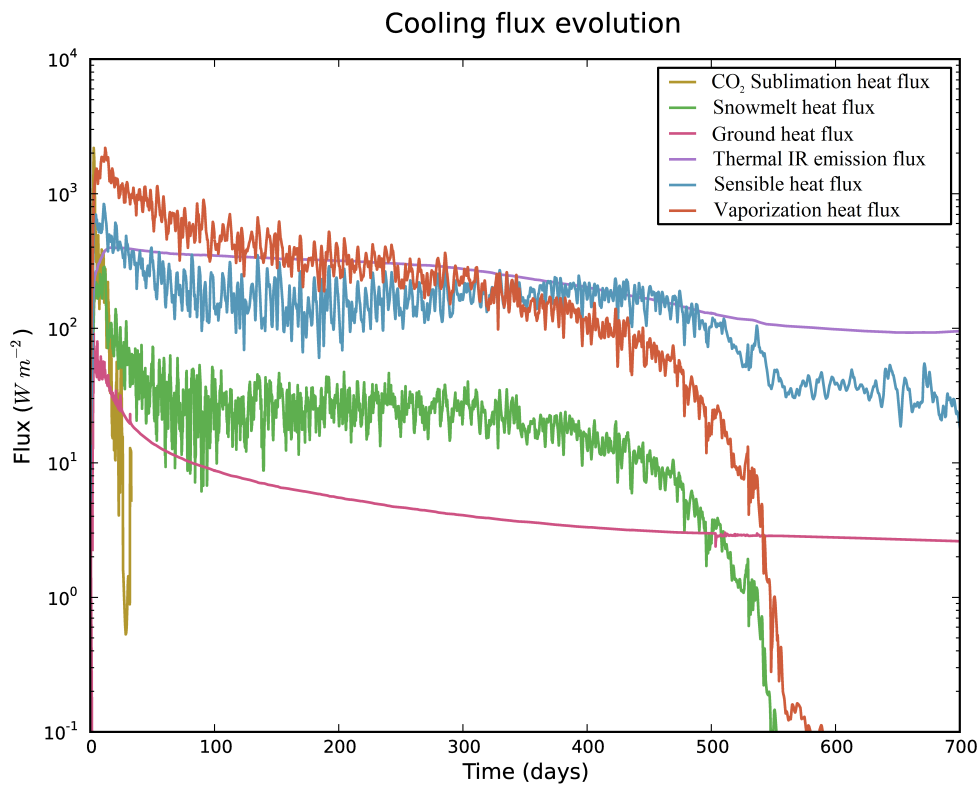


Figure 8: Surface cooling heat fluxes evolution averaged over all the Northern Plains lake grid cells in the  $P_{\text{surf}} = 0.2$  bar reference simulation. We note here that, depending on the nature and the intensity of a given outflow channel formation event, each of these fluxes can potentially become dominant.

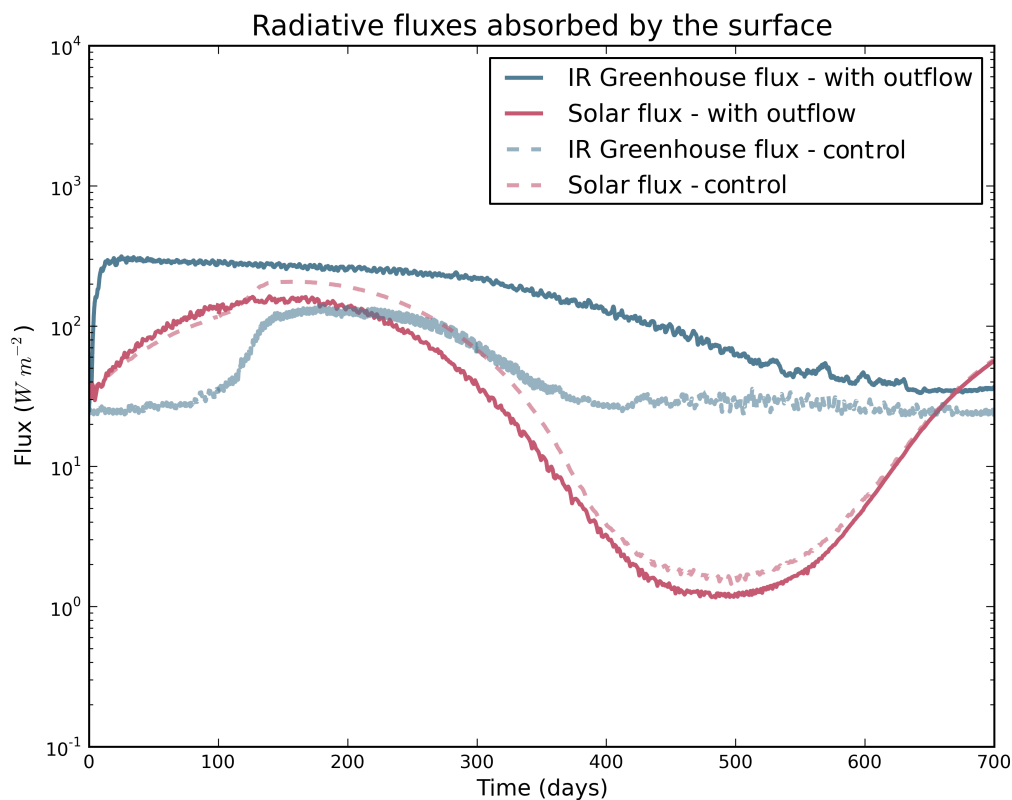


Figure 9: Evolution of the radiative fluxes absorbed by the surface and averaged over all the Northern Plains lake grid cells. Solid lines refer to the solar flux (in red) and the thermal infrared (in blue) for the reference simulation. Dashed lines correspond to the control simulations. For better visibility, we filtered diurnal waves from the absorbed solar fluxes using a 1 day running average.

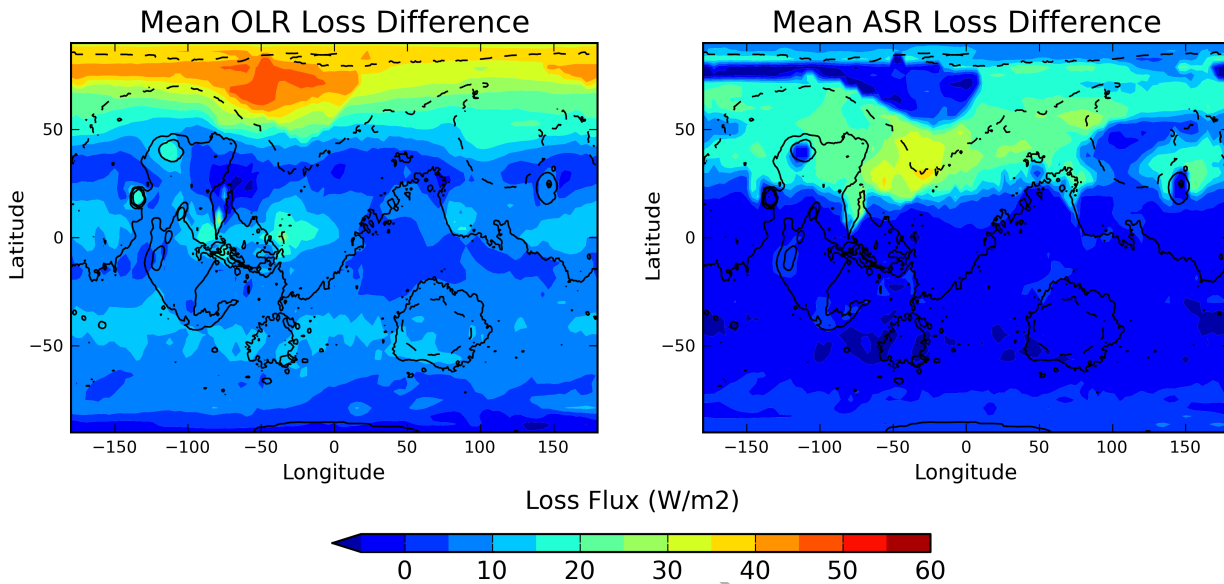


Figure 10: Mean Outgoing Longwave Radiation (OLR, left) and Absorbed Solar Radiation (ASR, right) loss during the warm phase, for the reference simulation, and relative to the control simulation performed for the same surface pressure.

738 atmosphere (essentially the troposphere)  $\sim 20 \text{ W m}^{-2}$  ( $\sim 12 \text{ W m}^{-2}$  more than the  
 739 control run). This corresponds to an average absorption of 65 % of the available  
 740 incoming solar flux ( $\sim 135.6 \text{ W m}^{-2}$  for the first 500 days).

741 2. The downward thermal infrared emission from the atmosphere and the  
 742 clouds is the dominant warming flux (see Figure 9). On average, during the warm  
 743 phase, this greenhouse effect brings  $\sim 210 \text{ W m}^{-2}$  to the lake ( $+ 150 \text{ W m}^{-2}$   
 744 more than the control run). The main source of thermal infrared emission surface  
 745 heating comes from the gaseous atmosphere itself, which can reach up to  $\sim 280 \text{ K}$   
 746 (above the lake) for the first 5 km, at the peak of the event.

747 In total, both solar and infrared heating counterbalance only  $\sim 30\%$  of the  
 748 cooling of the flow, and are thus unable to sustain the perturbation generated by  
 749 the outflow channel. We note here that the radiative effect of  $\text{H}_2\text{O}$  clouds during  
 750 the warm phase is approximately neutral or at least very limited (only  $+7 \text{ W m}^{-2}$ )  
 751 above the Northern Plains lake, with  $+23 \text{ W m}^{-2}$  of greenhouse warming and  
 752  $-17 \text{ W m}^{-2}$  due to the reflection of the sunlight.

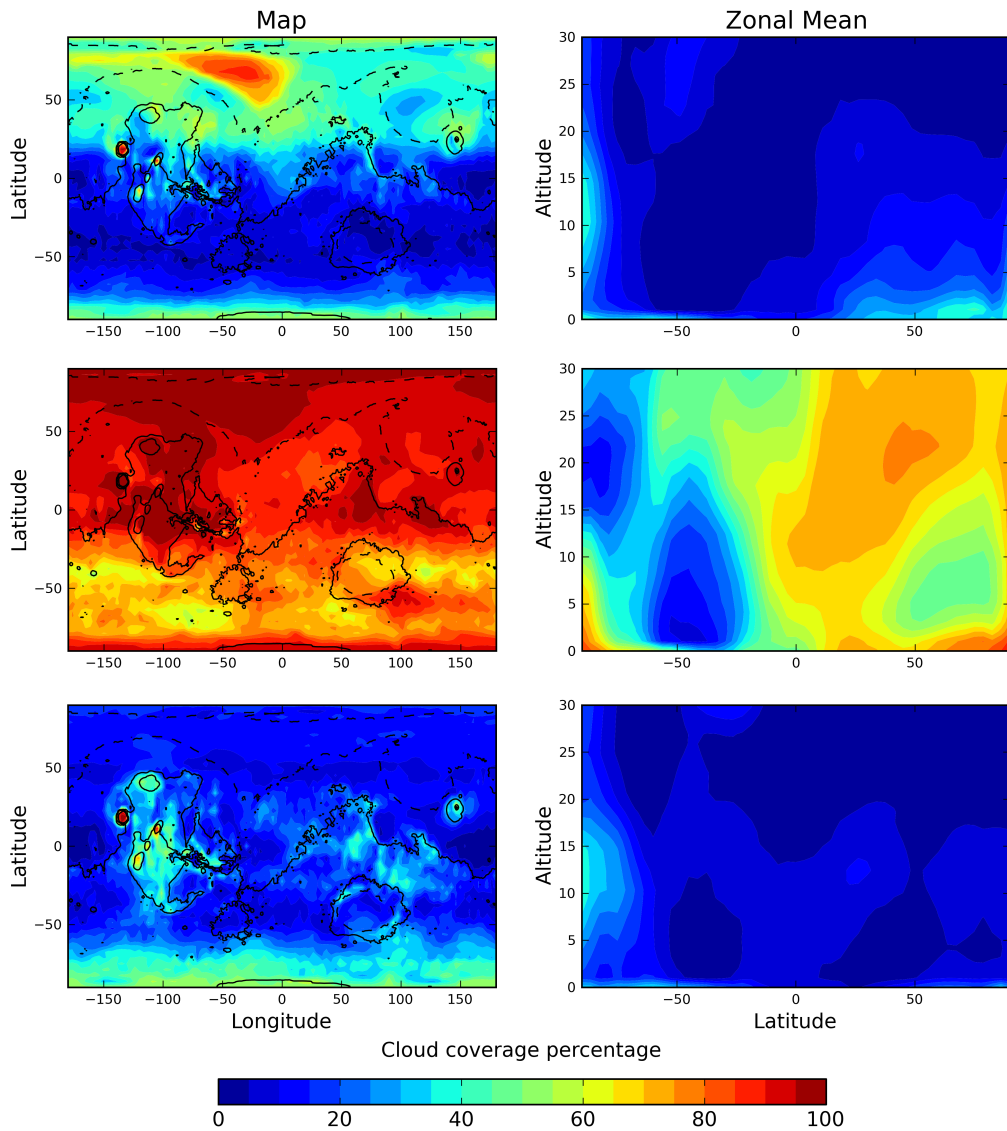


Figure 11: One year average of the cloud coverage following the outflow event. The first row corresponds to the map and the zonal mean cross-section of the reference simulation. In the second row, precipitation was removed (see section 7.2). The third row is for the control simulation.

### 753 4.2.3 The mechanisms cooling the atmosphere

754 One of the main results of our work is that outflow channel events are not able  
755 to sustain warm conditions. We present here the two processes that act efficiently  
756 together to cool down the atmosphere after outflow events.

757 1. In the time following catastrophic outflow channel events like the one  
758 described in this section, the atmosphere above the flow warms very quickly. In  
759 our reference simulation, 10 days after the beginning of the event, the temperature  
760 in the lower atmosphere (0-5 km) above the lake increases by almost 90 Kelvins.  
761 During the first 500 days after the event, because of this significant warming,  
762 the flow and the atmosphere just above it contribute to an extra thermal infrared  
763 emission loss to space of  $38 \text{ W m}^{-2}$  compared to the control simulation. Yet the  
764 amount of energy lost by the lake and the atmosphere above represent only  $\sim 11\%$   
765 of the extra total cooling to space. Figure 7 shows that, as the atmosphere gets  
766 warmer in the regions of the flow, high altitude winds around  $\sim 15 \text{ km}$  advect  
767 the heat to the neighbouring areas (in particular into the Northern Plains). This  
768 increases the surface of the emissions and therefore strengthens the cooling.

769 Figure 10 (left) shows the regions of the planet responsible for the extra thermal  
770 emission to space. Globally, during the warm phase (the first 500 days), the planet  
771 loses  $\sim 10 \text{ W m}^{-2}$ . One third of the emissions are due to the regions of latitude  
772  $> 50^\circ\text{N}$ . During the warm phase, the most important mechanism of cooling is the  
773 thermal infrared emission, enhanced by the advection processes.

774 2. Interestingly, another important cooling mechanism is the decrease of solar  
775 absorption due to the increase of surface albedo that follows the outflow channel  
776 event. In fact, the precipitation caused by the event, essentially in the form of  
777 snowfall (see Figure 12), leaves ice (see Figures 13 and 10) over an area of  
778  $\sim 30 \times 10^6 \text{ km}^2$  that reflect an important part of the sunlight ( $\sim 21.5 \text{ W m}^{-2}$ ). In  
779 total, during the warm phase and compared to the control simulation, the decrease  
780 of solar absorption contributes to a global equivalent extra cooling of  $\sim 4.5 \text{ W m}^{-2}$ ,  
781 which represents half of the infrared emission loss to space.

782 The large amount of water vapor released after the outflow channel event  
783 condenses very quickly in the atmosphere, forming clouds that are mostly located  
784 in the area of the flow and of the resulting lake (see Figure 11). In total, for  
785 the reference simulation, the clouds have a slight positive effect of  $+1.3 \text{ W m}^{-2}$   
786 ( $+ 2.3 \text{ W m}^{-2}$  of greenhouse effect and  $- 1.0 \text{ W m}^{-2}$  of solar reflection).



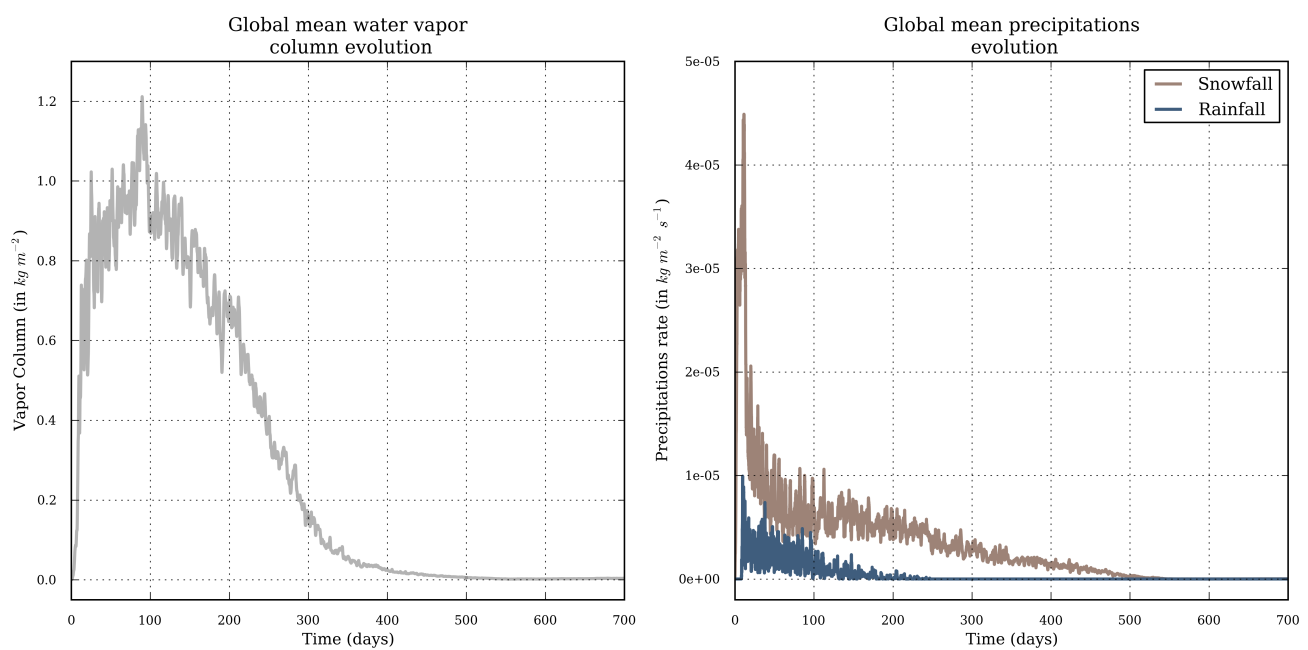


Figure 12: Evolution of the global mean water vapor column (left) and the precipitation (right), during the year following the outflow channel reference event.

#### 787 4.2.4 Consequences on the water cycle and the precipitation

788 The maximum total amount of water vapor that is carried by the atmosphere  
 789 during the event (GEL of 1.2 mm at the peak) remains limited by comparison  
 790 to the cumulative total amount of precipitable water generated (GEL of 230 mm).  
 791 It represents only  $\sim 0.5\%$  of the cumulative evaporated water vapor produced by  
 792 the entire outflow channel event during the first 500 days. Figure 12 (left) shows  
 793 the global mean atmospheric water vapor content (column mass in  $\text{kg m}^{-2}$ , and  
 794 also GEL in mm). It peaks at  $\sim 100$  days and considerably decreases from  $\sim 200$   
 795 days to  $\sim 500$  days.

796 The fact that the atmosphere is not able to accumulate more than  $\sim 1.2 \text{ kg m}^{-2}$   
 797 (globally) and  $\sim 50 \text{ kg m}^{-2}$  (locally, just above the warm lake) has one main  
 798 consequence: the atmosphere does not manage to carry enough water vapor far  
 799 enough from the lake to create precipitation in regions of interest (West Echus  
 800 Chasma Plateau in particular). The typical lifetime of the atmospheric water vapor  
 801 is in fact  $\sim 0.5$  days

802 Rainfall, which represents a very small fraction ( $\sim 10\%$ ) of the precipitation  
 803 (Figure 12), occurs only above the Northern Plain lake, because this is the only  
 804 location of Mars where atmospheric temperatures exceed (up to 10 km) the temperature  
 805 of the triple point. Outside the lake, the only mechanism of precipitation is  
 806 snowfall. Approximately 50% of the snow falls back directly on the flow/lake.  
 807 The rest of the precipitation (the 50% remaining) is essentially confined in the  
 808 northern regions. Figure 13 shows the map of the deposited ice field (generated  
 809 by precipitation) after a simulation of one martian year. The fraction of this ice  
 810 that is melted after an outflow event is very limited (see Figure 17), because 1)  
 811 most of the thermal perturbation has been dissipated by advection/cooling to space  
 812 processes after  $\sim 200$  days, 2) the remaining water vapour abundance after these  
 813 200 days is too low to trigger a significant greenhouse warming (as found by Kite  
 814 et al. 2011a) and 3) the ice field itself raises the albedo of the surface and thus acts  
 815 as a very efficient climatic cooling agent.

816 In summary, the short-term climatic impact of outflow channel formation events  
 817 seems very limited. For a 0.2 bar atmosphere, an outflow channel event of  $10^6 \text{ km}^3/300 \text{ K}$   
 818 leads to the formation of a lake (located in the Northern Plains main topographic  
 819 depression) that triggers a warm period that lasts for  $\sim 500$  days, which coincides  
 820 approximately with the complete surface freezing of the water in the lake. Such  
 821 events leave globally  $\sim 6.5 \times 10^3 \text{ km}^3$  of water ice/snow (0.65% of the initial  
 822 outflow reservoir) and are able to melt  $\sim 80 \text{ km}^3$  (0.008% of the initial reservoir;  
 823 1% of the deposited precipitation). Because the outflow events do not manage to

824 warm the atmosphere enough, water vapour stays confined to the regions neighbouring  
825 the lake (essentially in the Northern Plains) and therefore precipitation (mostly  
826 snowfall) and melting only occur in the lowland regions.

827 The long-term climatic impact of the ice-covered lake is discussed in the next  
828 section.

### 829 **4.3 The Cold Phase**

830 After 500 martian days, the surface of the Northern Plains lake is completely  
831 covered by ice. Temperatures, water vapor content and precipitation all decrease.  
832 Because the area of high albedo ice deposits is larger than in the control simulations,  
833 the mean surface temperatures extend even lower than before the outflow event  
834 ( $-2$  K for the global annual surface temperatures of the 0.2 bar reference simulation,  
835 and compared to the control simulation).

836 Using the extrapolation scheme presented in section 3.3.1, we estimated that  
837 the released water was completely frozen after  $\sim 4 \times 10^3$  martian years. This  
838 corresponds to the full solidification of the water to ice at the location of the main  
839 Northern Plains topographic depression (which is the deepest point of the lake).  
840 After  $\sim 500$  years, more than 70 % of the lake (in area) is frozen, from the surface  
841 to the top of the regolith. We note that the ground thermal flux (Clifford and  
842 Parker, 2001) during the Late Hesperian era was one order of magnitude too low  
843 (at best) to be able to increase the lifetime of the  $\sim 500$  m deep lake.

844 In our simulations,  $\sim 10$  years after the beginning of the lake-forming event,  
845 the mean ice thickness over the lake is  $\sim 25$  meters. The annual mean conduction  
846 heat flux for this ice thickness is  $\sim 10$  W m $^{-2}$ . The annual mean solar/IR fluxes  
847 absorbed by the ice are  $\sim 53/57$  W m $^{-2}$  by comparison, 110 W m $^{-2}$  in total. Under  
848 these conditions, the thermal conduction flux represents less than 10 % of the total  
849 heat flux received by the surface at the location of the lake. Moreover, because the  
850 temperature profile oscillates annually, in the first 5 meters (typically) of the ice  
851 cover, from positive values (summer season) to negative values (winter season),  
852 the heat conduction from the liquid water to the surface is mainly returned during  
853 the winter seasons. Yet, the water cycle in this cold phase is essentially controlled  
854 by the summer seasons, because sublimation rates are several orders of magnitude  
855 higher than during the winter seasons (see section 3.3.1 for discussion). Thus,  
856 after a few years (typically around 10), the climatic effect of the lake becomes,  
857 to a first order, the same as simply placing a comparable-sized body of ice in the  
858 Northern Plains. During these 10 years, ice transportation/water vapor cycle/precipitation  
859 is very limited by comparison to the warm phase and do not play any significant

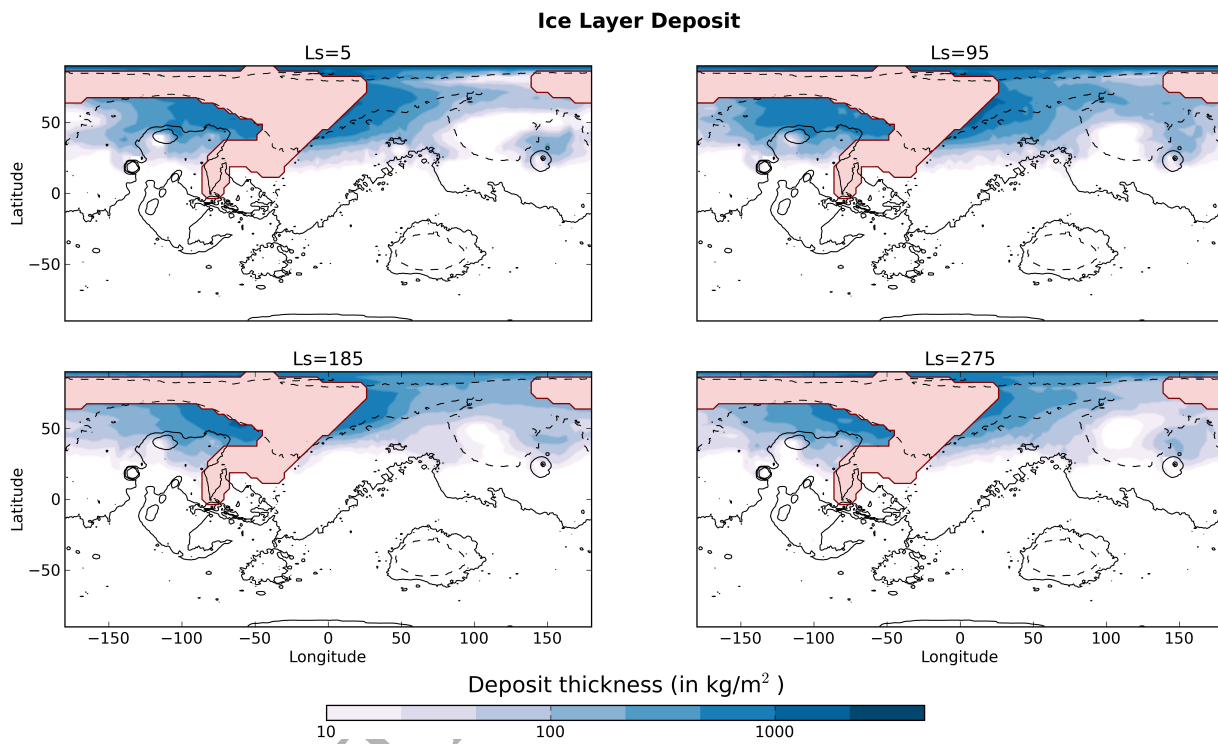


Figure 13: Ice deposit field obtained 1 martian year after the beginning of the event, for four different start dates ( $L_s = 5^\circ, 95^\circ, 185^\circ, 275^\circ$ ). The first figure (on the top left corner) corresponds to the reference simulation and starts at  $L_s = 5^\circ$ . We use the pink color for the regions where the flow passed through on its way to the lake.

860 role in the ice field position.

861 Within the lifetime of the liquid water lake, the ice field position evolution is  
862 completely controlled by the  $\sim 4 \times 10^3$  (-10) years of the water cycle forced by  
863 the sublimation of the large body of non-stable ice.

864 Each year, during Northern summer,  $\sim 20 \text{ mm year}^{-1}$  of lake ice sublimates  
865 to condense elsewhere and approximately 30 % of it is transported away from  
866 the lake. Progressively, the water vapor produced during the summers migrates  
867 southward and - through the mechanism of adiabatic cooling - condenses on  
868 the regions of high altitudes and low latitudes. The lifetime of the frozen lake  
869 predicted by our simulations is  $\sim 7 \times 10^4$  martian years.

870 The evolution of the ice field through the phases that follow the outflow channel  
871 reference event are shown in Figure 14. After  $\sim 10^5$  martian years, the outflow  
872 channel water is located more or less exclusively in the highland regions. During  
873 this cold phase ( $\sim 10^5$  martian years), some ice appears stable in the region of  
874 West Echus Chasma Plateau, due to the uninterrupted supply of ice coming from  
875 the northern parts of the planet. This snow deposit is produced by the adiabatic  
876 cooling of the ascending air masses that provoke the condensation of the water  
877 vapor initially generated by the sublimation of the Northern Plains ice field.

878 Some water ice is also transported to the drainage regions of **Alba Patera**,  
879 **Hecates Tholus** and **Ceraunius Tholus** but this might not be a critical factor  
880 since our model already predicts that ice deposits should be stable in these regions  
881 (Wordsworth et al. 2013, Figure 2 this work) and therefore available for either  
882 seasonal snowmelt or ground melting.

883 In spite of this, because the global surface albedo is increased during that  
884 period, global temperatures are much lower than before the outflow event, making  
885 snowmelt difficult.

886 We note here that we did not take into account the flow of the ice on the  
887 Northern Plains slopes. This could significantly increase the lifetime of the lake  
888 located in the main topographical depression and thus the lifespan of the snow  
889 deposited in non-stable locations (in particular in West Echus Chasma Plateau  
890 area). However, at these temperatures and over these timescales, ice is unlikely  
891 to flow significantly (Fastook et al., 2012; Fastook and Head, 2014, 2015). In  
892 addition, we did not take into account the formation of a possible lag deposit  
893 (Kreslavsky and Head, 2002; Mouginot et al., 2012) which could have decreased  
894 the sublimation rate of the ice. Both of these factors, however, appear to have  
895 minimal effects on the general processes.

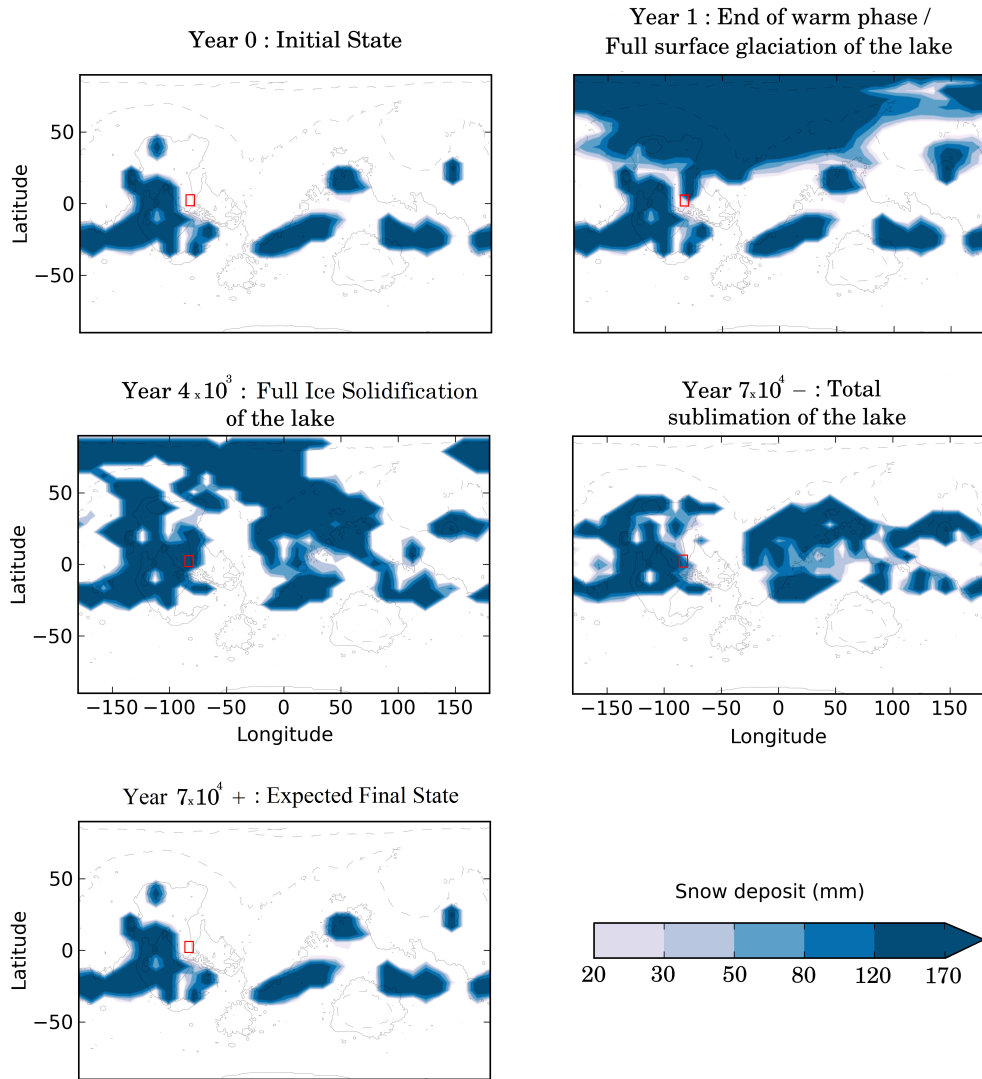


Figure 14: Ice deposit field after 0 / 1 /  $4 \times 10^3$  /  $7 \times 10^4$  martian years, corresponding to the main phases following the outflow channel reference event. The red rectangle corresponds to the West Echus Chasma Plateau area.

### 896 **4.3.1 Influence of obliquity**

897 Orbital spin-axis obliquity is a very important factor in the duration and the characteristics  
898 of the cold phase, because it controls the latitudinal distribution of the solar flux  
899 and thus the sublimation processes. We performed two simulations of the reference  
900 outflow channel event, at obliquities of  $25^\circ$  and  $65^\circ$ , to complement the  $45^\circ$   
901 obliquity case presented initially.

902 In the low obliquity simulation, the sublimated ice migrates slowly toward the  
903 coldest points of the planet: the South pole and the North pole (in agreement with  
904 Wordsworth et al. (2013), Figure 4). The water present in the northern part of  
905 the lake is stable in the long term. In this situation, ice never accumulates in the  
906 region of West Echus Chasma.

907 In the high obliquity simulation, the water cycle is much more intense because  
908 the peak of insolation at high latitudes is higher. Approximately  $\sim 55$  mm of the  
909 sublimated northern lake ice migrates southward each year. The lifetime of the  
910 lake is thereby lowered to  $\sim 9 \times 10^3$  martian years. For the same reasons as that in  
911 the reference simulation, a thick ice deposit is present in the region of West Echus  
912 Chasma Plateau. Yet, its duration,  $\sim 10^4$  years, is almost 10 times less than in the  
913 reference simulation, more or less coincident with the lifetime of its supply (the  
914 frozen lake).

915 As a result, the lifetime of the ice located in West Echus Chasma area seems  
916 to be favored at obliquity  $\sim 45^\circ$ .

## 917 **5 The effect of surface pressure**

918 For many reasons (see discussion in section 2.2), the atmospheric pressure during  
919 the Late Hesperian epoch is not well constrained. We explore in this section the  
920 role of surface pressure on the climatic impact of outflow channels.

921 For this, we performed five different simulations of the same outflow channel  
922 event ( $10^6$  km<sup>3</sup>, 300 K water released at  $1$  km<sup>3</sup> s<sup>-1</sup> in Echus Chasma) for five  
923 different surface pressures (40 mbar, 80 mbar, 0.2 bar (the reference simulation),  
924 0.5 bar and 1 bar).

### 925 **5.1 Warm Phase**

926 Atmospheric pressure is one of the key factors that control the efficiency at which  
927 the warming of the atmosphere and the transport of water occur during the warm

928 phase, as pointed out by Kite et al. (2011a).

929 1. The evaporation rate: Combining equations 9 and 15 for low amounts of  
930 water vapor, the evaporation rate  $E$  can be written:

$$E = \frac{C_d V_1 P_{\text{ref}} M_{\text{CO}_2}}{R T_1} e^{\frac{L_v M_{\text{H}_2\text{O}}}{R} \left( \frac{1}{T_{\text{ref}}} - \frac{1}{T_{\text{surf}}} \right)}. \quad (17)$$

931 Hence, the evaporation rate does not (directly) depend on the surface pressure and  
932 is mostly controlled by the temperature  $T_{\text{surf}}$  of the flow/lake. To first order (and  
933 this is confirmed by our simulations), the wind velocity  $V_1$  and the atmospheric  
934 temperatures  $T_1$  do not differ sufficiently from one atmospheric pressure to another  
935 to play a major role on the rate of evaporation.

936 2. The warming rate: The volumetric heat capacity of the atmosphere increases  
937 linearly with the volumetric mass density and thus the atmospheric pressure. For  
938 example, it takes approximately  $\frac{1.0}{0.040} = 25 \times$  more energy to warm a 1 bar  
939 atmosphere than a 40 mbar one.

940 When the outflow channel event occurs, the rate of warming of the atmosphere  
941 (in K/s) is roughly proportional to the evaporation rate (which is the main source of  
942 heating) and inversely proportional to the volumetric heat capacity of the atmosphere.  
943 In our simulations, it takes  $\sim 10/40$  martian days - respectively for the 40 mbar/1 bar  
944 case - for the atmospheric temperatures at 10 km to reach a plateau at 250 K/220 K,  
945 which correspond to a +80 K/+30 K temperature increase (for initial temperatures  
946 equal to 170 K/190 K). This corresponds approximately to a factor of 10 in heating  
947 efficiency for these two endmember situations. The difference between the factor  
948 of 25 predicted and the factor of 10 obtained in our simulations is mostly due to  
949 two processes: advection and thermal emission to space.

950 The same two processes limit the growth of atmospheric temperatures. First,  
951 the advection tends to dilute the heat perturbation horizontally. In the 1 bar case,  
952 this is the dominant process for example. Second, the thermal emission to space  
953 acts as a very efficient negative feedback. This is, in fact, the first limiting process  
954 in the 40 mbar case.

955 The capability of an atmosphere to maintain high temperatures from the surface  
956 (where evaporation occurs) to the altitude where advection occurs is in fact the  
957 most important factor in the ability to transport water vapor globally and produce  
958 precipitation far from the region of evaporation. The warmer the atmospheric  
959 column above the lake is, the more water vapor will be possibly lifted and then  
960 transported globally by the high altitude winds.

961 Thin atmospheres (such as the 40 mbar) warm efficiently above the region  
962 of the flow, allowing the formation of a persistent water vapor plume that can



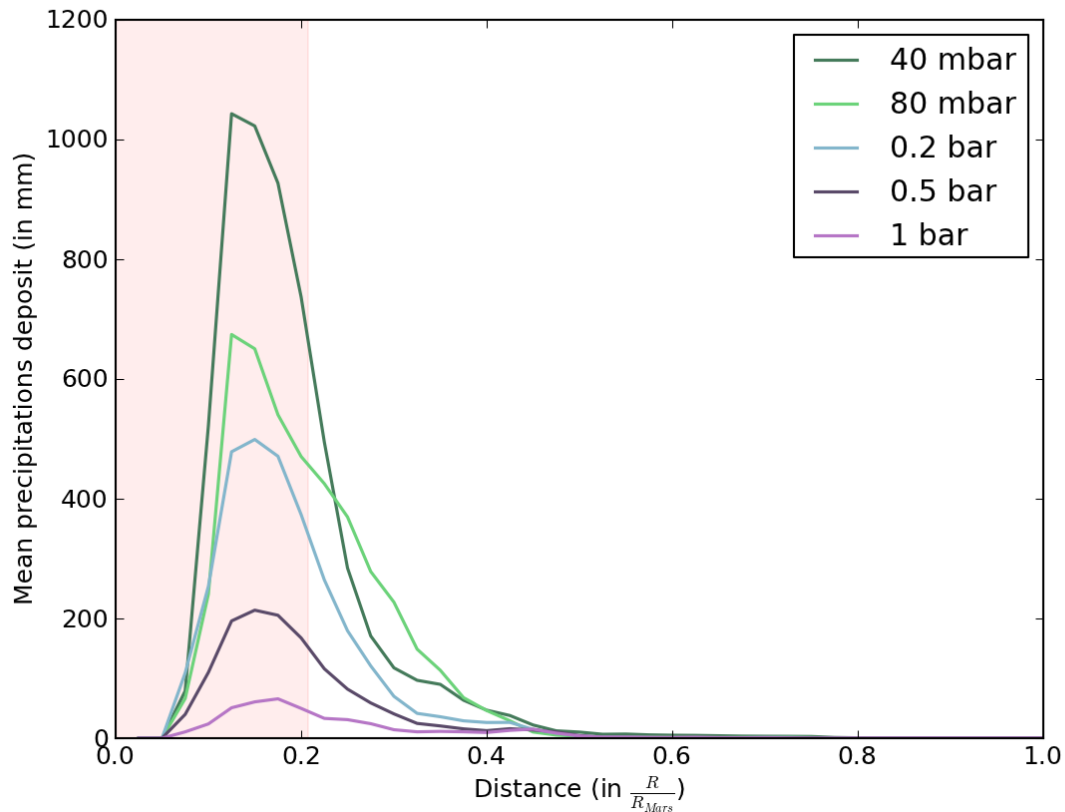


Figure 15: Radial precipitation distribution around the center of the lake (70°N, -30°E), averaged for the first 500 days following the outflow channel event, and for 5 different atmospheric pressures. The precipitation falling back on the lake/flow was removed from the plot. Because the lake is not circular, we used the pink color to represent the maximum radial extent of the lake.

963 transport (through advection) water vapor far from the flow/lake. In contrast, thick  
964 atmospheres (such as the 1 bar case) ironically do not manage to transport water  
965 efficiently because of the advection itself. The advection prevents the atmospheric  
966 temperatures above the lake from building up and thus the water vapor from  
967 accumulating. This limits the transport of water vapor and favors local precipitation.  
968 This is summarized by Figure 15 that shows the radial mean distribution (centered  
969 above the Northern Plains lake) of precipitation for the entire warm phase (first  
970 500 days). Our experiments show that thin atmospheres are able to transport much  
971 more water and for much longer distances than thick ones.

972 We compare in Figure 16 the spatial distribution of the precipitation (only  
973 snowfall, because rainfall occurs only above the lake) for the different atmospheric  
974 pressures. Whatever the surface pressure considered, the precipitation stays confined  
975 to the Northern Plains.

976 Another important aspect concerns the role of atmospheric pressure on the  
977 ability to melt the ice initially present / transported by the outflow event itself.  
978 Thin atmospheres, while able to reach temperatures in excess of 273 K above the  
979 flow, are not able to raise global temperatures significantly. First, the relaxation  
980 timescale of the temperature field is very low in such atmospheres because of  
981 the weak infrared absorption of the atmosphere. Second, outflow channel events  
982 under thin atmospheres generate a very large ice cover that reflects sunlight efficiently.  
983 As a result, an outflow channel of  $10^6 \text{ km}^3$  that occurs under a 40 mbar atmosphere,  
984 leaves globally  $\sim 1.5 \times 10^4 \text{ km}^3$  of water ice/snow (1.5%) and is able to melt only  
985  $\sim 50 \text{ km}^3$  (0.005%).

986 Thick atmospheres are initially warmer than thin atmospheres (+ 30 K between  
987 the 1 bar and 40 mbar atmospheres). They also have a much more efficient infrared  
988 absorption and thus are better candidates to melt the deposited ice field. For  
989 example, an outflow channel of  $10^6 \text{ km}^3$  that occurs under a 1 bar atmosphere,  
990 leaves globally  $\sim 4 \times 10^3 \text{ km}^3$  of water ice/snow (0.4%) and is able to melt  
991  $\sim 110 \text{ km}^3$  (0.011%).

992 Nonetheless, this melting occurs only in the Northern Plains, in the close  
993 vicinity of the lake, because such thick atmospheres do not transport much ice  
994 anywhere on the planet in any case. In addition, ice albedo feedback (which is yet  
995 lower for thicker atmospheres) and the high volumetric heat capacity (lower heat  
996 perturbation) of such atmospheres contribute to lower the possibility of reaching  
997 melting temperatures.

998 Whatever the value of the surface pressure, the ability of the atmosphere to  
999 produce liquid water from melting is very limited.

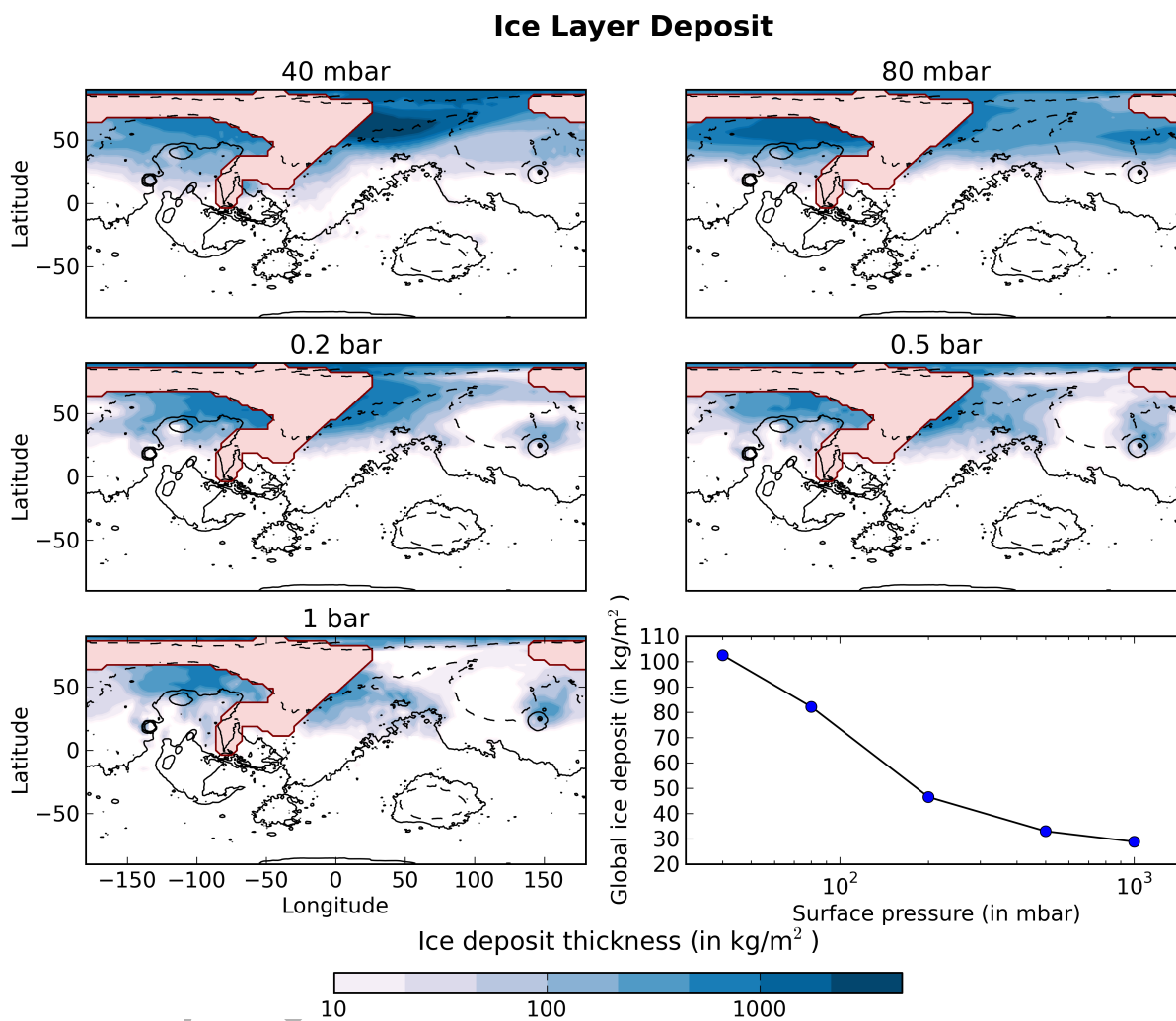


Figure 16: Final Ice layer deposit map (in  $\text{kg m}^{-2}$ ) after 1 martian year of simulations, for five different surface pressures (40 mbar, 80 mbar, 0.2 bar, 0.5 bar and 1 bar). The pink color denotes the regions where the flow passed through on the way to the lake.

### Maximum surface liquid water in the year following the event

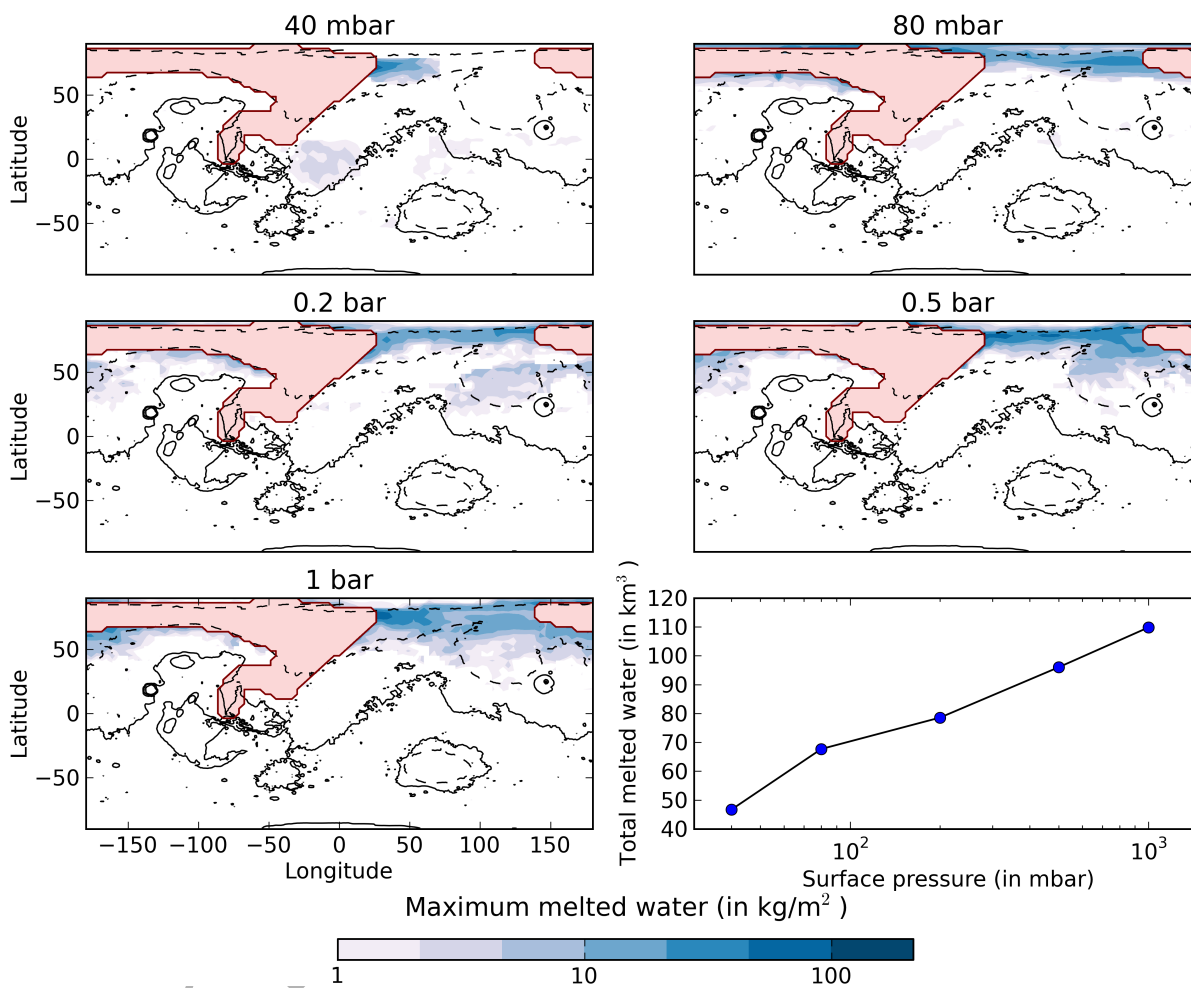


Figure 17: Maximum surface liquid water after 1 year of simulation and for five different surface pressures (40 mbar, 80 mbar, 0.2 bar, 0.5 bar and 1 bar). The pink color denotes the regions where the flow passed through on the way to the lake.

## 1000 **5.2 Cold Phase**

1001 The water cycle during the cold phase is, in contrast, more intense for thick  
1002 atmospheres than for thin ones. The sublimations rates are higher because global  
1003 temperatures (and also summer temperatures) are also higher. At the end of the  
1004 warm phase, the mean global temperatures for the 40 mbar/1 bar simulations are  
1005 respectively  $\sim 193$  K (3.5 K lower than the control simulation) and  $\sim 226$  K (1 K  
1006 lower than the control simulation). This difference is due to the increased ice  
1007 cover following the outflow event.

1008 In the 1 bar simulation (thick case), the lifetime of the frozen lake is  $\sim 5 \times 10^4$   
1009 martian years, slightly lower than in the reference simulation. The climatic response  
1010 during the cold phase behaves more or less in the same manner as in the reference  
1011 0.2 bar simulation.

1012 In the 40 mbar simulation (thin case) however, because the water cycle is  
1013 too weak (sublimation rate of the lake of 2mm/year; lifetime of the frozen lake  
1014  $\sim 2 \times 10^5$  years), the southward flux of the atmospheric water ice is not high  
1015 enough to allow the presence of stable ice in the area of the West Echus Chasma  
1016 Plateau.

1017 More generally, atmospheres with pressure higher than 80 mbar seem necessary  
1018 to produce ice deposits in the region of West Echus Chasma Plateau.

## 1019 **6 Extreme parameterizations**

1020 In this section, we study several scenarios that may deeply affect the climatic  
1021 impact of outflow events: 1. the intensity of the event and 2. the effect of clouds  
1022 and precipitation.

### 1023 **6.1 Intensity of the event**

1024 Because outflow channel events such as the one presented in Section 4 fail to  
1025 produce rainfall/transient warming, it is tempting to explore even more extreme  
1026 parameterizations of the outflow events.

#### 1027 **6.1.1 Temperature of the flow**

1028 The temperature of the groundwater released during outflow events is not well  
1029 constrained (see section 2.1.1). Hence, we used the temperature of the flow as  
1030 a tuning parameter to explore the sensitivity of our results to the intensity of

1031 the outflow event. We performed three simulations of the same outflow event  
 1032 ( $10^6 \text{ km}^3$ , released in Echus Chasma) for three different groundwater temperatures:  
 1033 280 K, 300 K (reference simulation) and 320 K.

1034 As expected, the warmer the water, the more intense the climatic effect becomes.  
 1035 For example, at the peak of the warm phase, the 320 K event is able to carry  
 1036 approximately  $8 \times$  more water vapor than in the reference simulation because  
 1037 atmospheric warming processes are amplified by the temperature (evaporation/condensation  
 1038 cycle, IR emission of the flow, ...). Consequently, 25 % of the precipitation  
 1039 following the 320 K event is rainfall (respectively 10 %/0 % for the reference/280 K  
 1040 simulations). Yet, rainfall still occurs exclusively above the lake (70 %) or in the  
 1041 northern lowlands of Mars (30 %). Snow precipitation also remains confined to  
 1042 the Northern Plains down to  $15^\circ \text{N}$  ( $25/40^\circ \text{N}$  for the 300 K/280 K simulations).

1043 The amount of water ice transported (Figure 18) and melted (Figure 19) after  
 1044 outflow channel events with 280 K/300 K/320 K water shows that in all cases,  
 1045 the mechanism of advection/cooling to space is very efficient, and as a result, the  
 1046 duration of the warm phase is approximately the same ( $\sim 500$  days) between the  
 1047 reference and the 320 K simulations.

1048 We note that, at the end of the warm phase, because the amount of ice transported  
 1049 (and the area of the deposit with it) increases with the initial temperature of  
 1050 the flow, the average surface albedo raises and the mean temperatures decrease:  
 1051 Warmer flows lead to colder states.

### 1052 **6.1.2 Magnitude of the event: from small outflows to oceans.**

1053 Recent work (Andrews-Hanna and Phillips, 2007; Harrison and Grimm, 2008)  
 1054 has suggested that outflow channels were preferentially carved by multiple events  
 1055 of reduced sizes ( $\sim 10^3 \text{ km}^3$ ) rather than by large ( $> 10^5 \text{ km}^3$ ) single outflows.  
 1056 We performed simulations for different volumes of water at 300 K and released  
 1057 in Echus Chasma at a rate of  $1 \text{ km}^3 \text{ s}^{-1}$ , from  $10^3 \text{ km}^3$  (consistent with the most  
 1058 recent estimations of outflow volumes) to  $10^7 \text{ km}^3$  (ocean case). Figure 20 shows  
 1059 the final position of the lake as a function of the initial volume of water. The  
 1060  $10^6 \text{ km}^3$  case is the reference simulation.

1061 Our results show that the large outflows, during the warm phase, transport  
 1062 much more water than the small ones (cumulative). Small outflows (typically  
 1063  $\sim 10^3\text{-}10^4 \text{ km}^3$ ) have a small wetted area (typically  $0.15\text{-}0.41 \times 10^6 \text{ km}^3$ ) and  
 1064 a small initial heat reservoir, so that they cannot warm the atmospheric column  
 1065 above the flow/lake sufficiently to transport water vapor into the neighbouring  
 1066 regions. Small outflow events inject more or less the same amount of water vapor

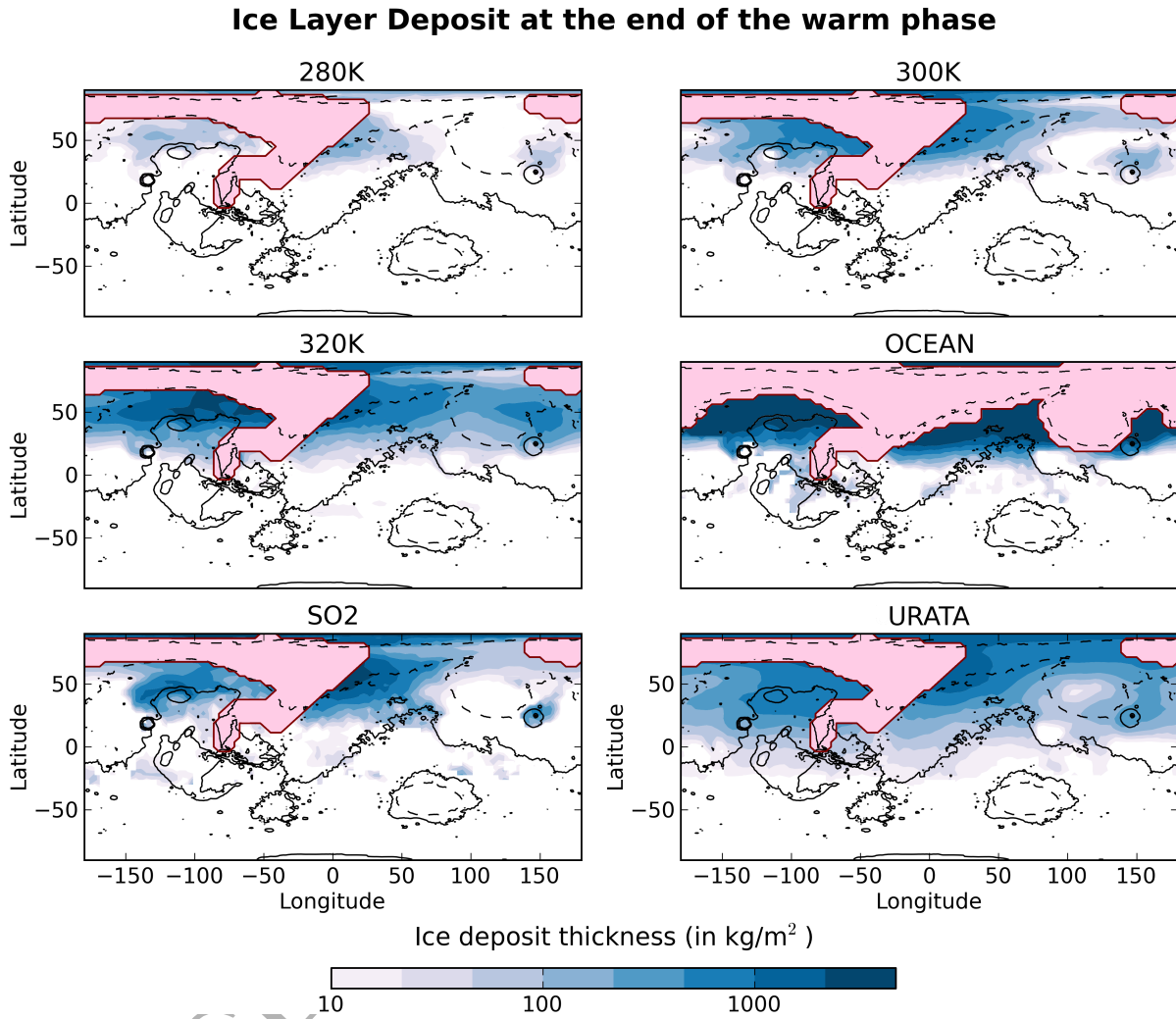


Figure 18: Final Ice layer deposit map (in  $\text{kg m}^{-2}$ ) after 1 martian year of simulations (4 martian years for the ocean case), for six different simulations: 1. 280 K outflow, 2. 300 K outflow (reference case), 3. 320 K outflow, 4.  $10^7 \text{ km}^3$  ocean case, 5. 1 %  $\text{SO}_2$  case and 6.  $l_0 = \infty$  (no precipitation case). The pink color denotes the regions where the flow passed through on the way to the lake.

### Maximum surface liquid water in the year following the event

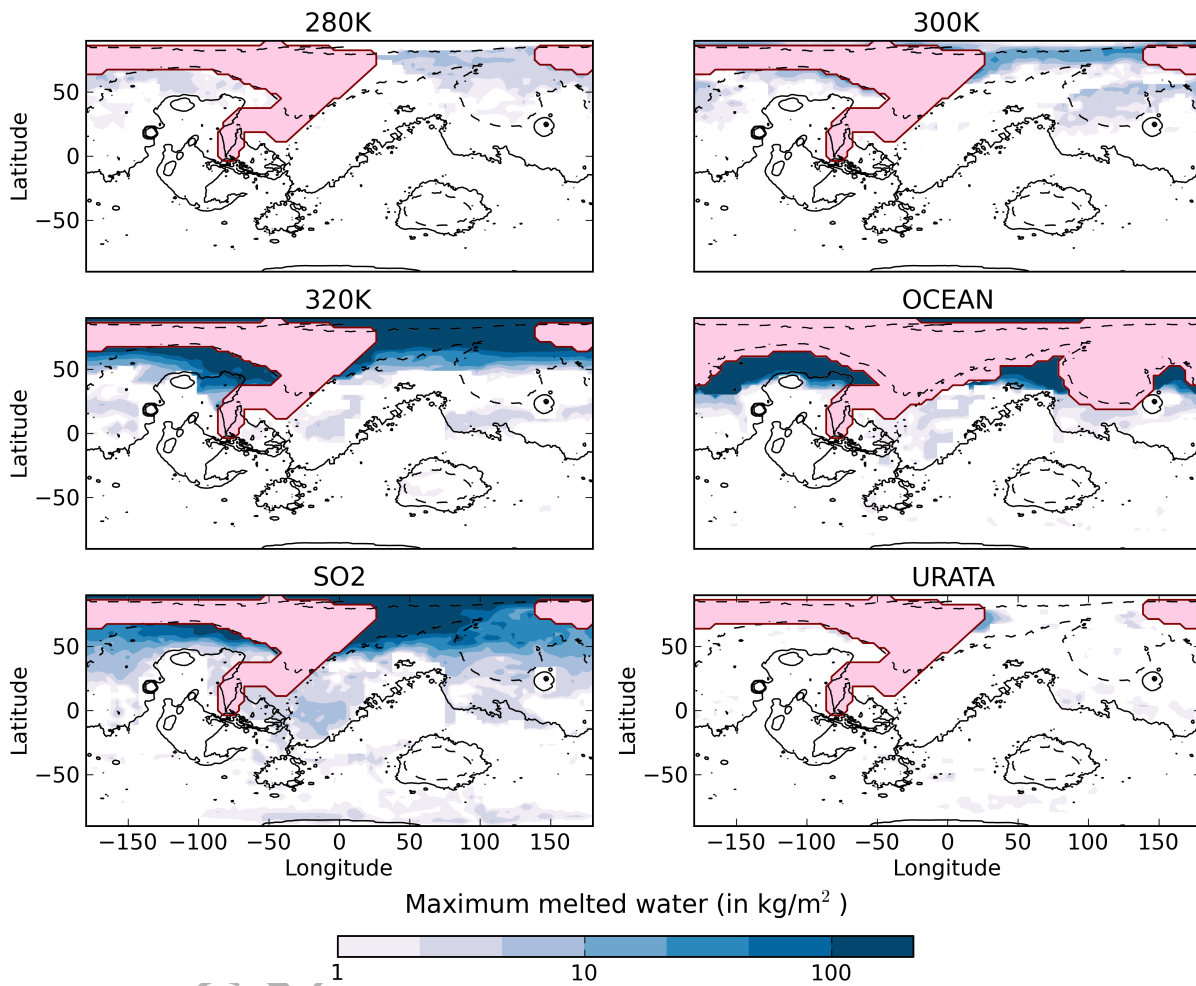


Figure 19: Maximum surface melted liquid water after 1 year of simulation (4 martian years for the ocean case) and for 6 different simulations: 1. 280 K outflow, 2. 300 K outflow (reference case), 3. 320 K outflow, 4.  $10^7$  km<sup>3</sup> ocean case, 5. 1% SO<sub>2</sub> case and 6.  $I_0 = \infty$  (no precipitation case). The pink color denotes the regions where the flow passed through on the way to the lake.



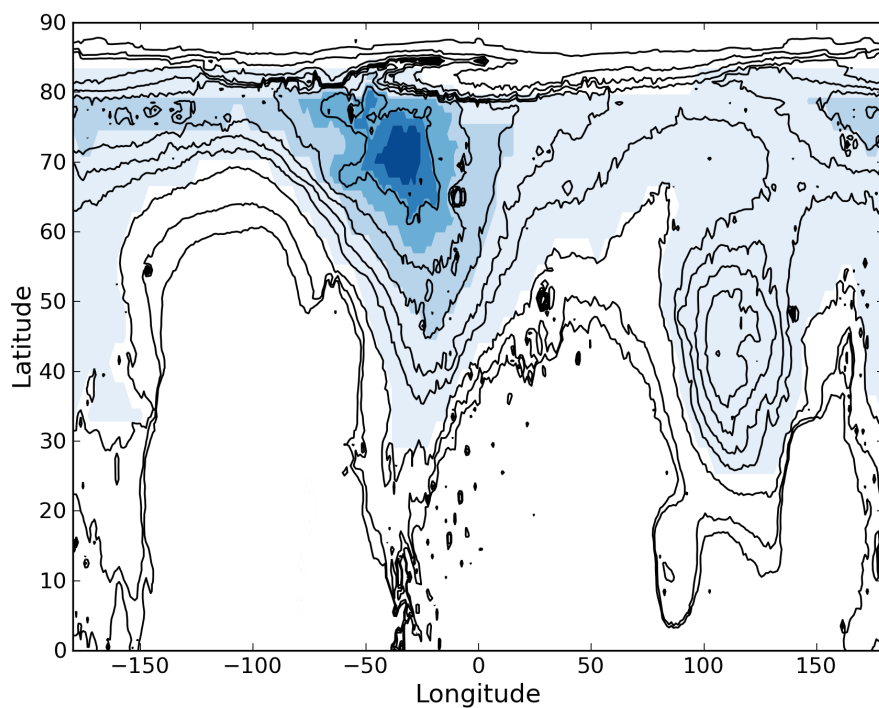


Figure 20: Stable water lake size obtained in our simulations depending on the volume of water released. Contour fills are for volumes of  $10^3, 10^4, 10^5, 10^6$  (reference case) and  $10^7$   $\text{km}^3$  (ocean case). These correspond respectively to wetted areas of 0.15, 0.41, 1.19, 4.15 and 20.4 millions of  $\text{km}^2$ .

1067 (in proportion) than large ones, but they are not able to transport it far from  
1068 the flow/lake. For example,  $2 \times 10^2$  events of  $5 \times 10^3 \text{ km}^3$  transport 2 orders  
1069 of magnitude less ice outside the flow/lake than a large  $10^6 \text{ km}^3$  one (reference  
1070 simulation). Moreover, large outflows are able to generate precipitation up to  $\sim$   
1071 5000 km from the edge of the flow/lake whereas small ones cannot produce any  
1072 precipitation at a distance greater than  $\sim 400$  km (typically the size of 2 GCM  
1073 grids).

1074 We did not explore in detail the effect of the discharge rate, which has a net  
1075 impact on the size and duration of the wetted area (and thus on the evaporation  
1076 and the albedo), but also on the intensity of the event. Nonetheless, the climatic  
1077 response to lower discharge rate events ( $< 10^9 \text{ m}^3 \text{ s}^{-1}$ ) was found to be lower,  
1078 because in such cases the temperatures and the amount of water vapor struggle to  
1079 build up above the flow/lake.

1080 Because large outflows seem to be much better candidates for generating precipitation  
1081 globally, we examined the extreme case of a catastrophic outflow event of  $10^7 \text{ km}^3$   
1082 released simultaneously by all of the circum-Chryse outflow channels (Kasei,  
1083 Ares, Tiu, Simu Vallis, etc.). This possibility, sometimes called the MEGAOUTFLO  
1084 (Mars Episodic Glacial Atmospheric Oceanic Upwelling by Thermotectonic Flood  
1085 Outburst) hypothesis (Baker et al., 1999), speculates that such events could warm  
1086 Mars during periods of  $10^4$ - $10^5$  years through a transient greenhouse effect provoked  
1087 in part by the injection of large amounts of water vapor.

1088 Our experiments show that such events cannot sustain long-term greenhouse  
1089 effects, whatever the size and the temperatures considered for the northern lake/see/ocean.  
1090 After 3.5 martian years, for the outflow event described above, the surface of the  
1091 lake/see/ocean becomes totally frozen. The thermal infrared emission to space  
1092 (enhanced by the heat horizontal advection and by the water vapor advection that  
1093 release latent heat because of adiabatic cooling; see Figure 21 for the detailed  
1094 mechanism) acts very efficiently to cool the planet. The ice deposited on the  
1095 Northern Plains slopes (Figure 18) also enhances the cooling through a depletion  
1096 of surface solar absorption. As a result, in such a scenario, rainfall/snowmelt still  
1097 only occurs in the lowest northern lowlands (see Figure 19) of the planet (far from  
1098 the region of interests).

1099 In summary, the most intense outflow channel events possible are not able to  
1100 sustain a global greenhouse warming. Such events only manage to warm up the  
1101 atmosphere regionally, in the Northern Plains, and only for a few years at best.  
1102 Consequently, rainfall (and snowmelt) occur only in the neighbourhood regions  
1103 of the final stable lake. After complete surface freezing of the lake, the climate  
1104 becomes much colder than initially (due to the increase of the surface albedo),

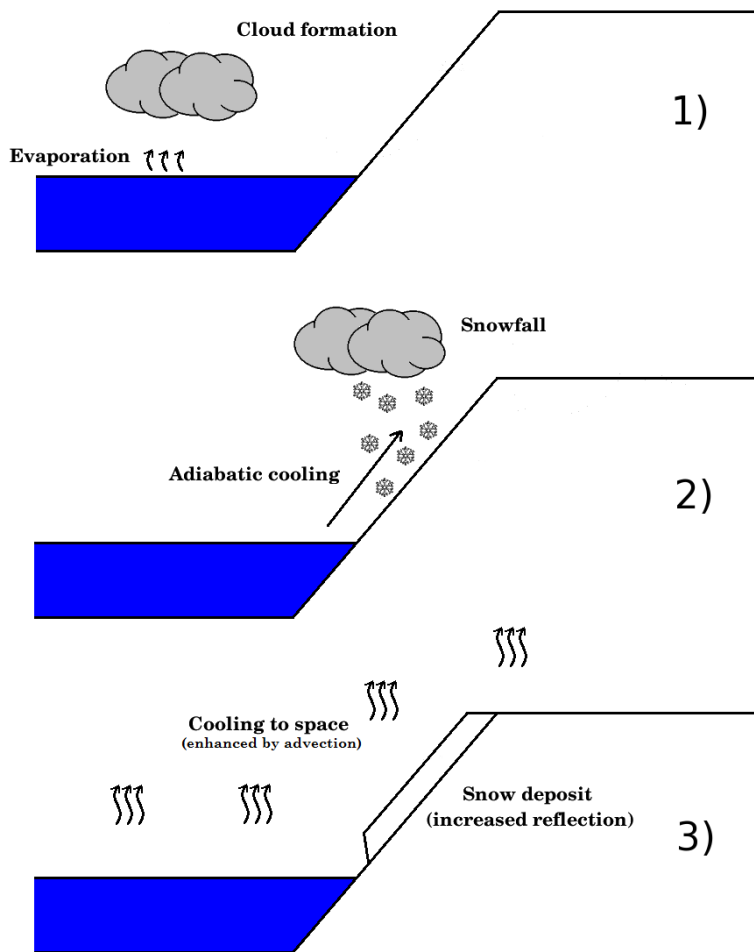


Figure 21: Why is a martian Northern ocean short-lived and unable to induce rain? The water vapor and the clouds that build up above the ocean (1) are advected southward (2). Because of adiabatic cooling, snowfall occurs from the edges of the ocean (starting from 70 °N, see Figure 20) to the highest parts of the planet (the 'Icy Highlands'). The advection of heat (increasing the surface of thermal infrared emission) and the increase of reflection (because of the snow deposit) both contribute to an intense cooling of the ocean and to reduced precipitation.

1105 making the snowmelt even more difficult.

1106 We note that we did not take into account the modification of the topography  
1107 by the presence of a lake/see/ocean, which might be a concern for very high  
1108 volumes of water ( $\geq 10^7$  km<sup>3</sup>). It could reduce significantly the role of adiabatic  
1109 cooling and thus favor the transport/deposit of water further south.

## 1110 **7 Discussion**

### 1111 **7.1 Role of the atmospheric composition.**

1112 In this analysis, we made the assumption that the Late Hesperian martian atmosphere  
1113 was made of 100% CO<sub>2</sub> (and some water vapor). Outflow channel events under a  
1114 CO<sub>2</sub> dominated atmosphere seem not to be able to provoke long-term warming or  
1115 precipitation at the global scale.

1116 Outflow channel formation events are very likely related to intense volcanic  
1117 episodes during martian history (Baker et al., 1991; Head et al., 2002). During  
1118 these periods, it is believed that volcanic gases like SO<sub>2</sub> may have been massively  
1119 released [see section 1. of Kerber et al. (2015) for more details].

1120 We performed a simulation of an outflow channel event under the same conditions  
1121 as in section 4, but this time with 1 % of SO<sub>2</sub>. Figures 18 and 19 show the  
1122 corresponding amount of water ice transported/melted after the event. Small  
1123 amounts of SO<sub>2</sub> (2 mbar here) are sufficient to raise the global atmospheric temperatures  
1124 by several tens of Kelvins and thus to favor the transport of water vapor/water ice  
1125 globally and create precipitation far from the Northern Plains stable lake.

1126 However, using the same GCM, Kerber et al. (2015) (and earlier, Tian et al.  
1127 (2010)) have shown that massive volcanic SO<sub>2</sub> outgassing cannot lead to a global  
1128 and substantial warming, because sulfur aerosols that would form at the same time  
1129 have a very strong cooling effect, even in small amounts.

1130 We also believe that, under more realistic parameterizations that would take  
1131 into account sulfur aerosols (e.g. Halevy and Head (2014)), the outflow channel  
1132 climatic impact would be also very limited.

### 1133 **7.2 The role of clouds and precipitation.**

1134 The radiative effect of clouds is one of the main sources of uncertainty in GCMs  
1135 and thus also on the consistency of our results.

1136 In particular, it has been suggested (Urata and Toon, 2013) that high altitude  
1137 ('cirrus-like') water ice clouds may trigger warm climates on Mars even under a  
1138 faint young sun. This scenario requires four assumptions: 1) Water ice particles  
1139 that have sizes  $> 10$  microns; 2) that the rate of precipitation is very low (in order  
1140 to extend the lifetime of the clouds); 3) When present, clouds need to completely  
1141 cover a grid cell (no partial cloud cover); 4) Lastly, it also requires an initial  
1142 'warm' state, for example an outflow channel event.

1143 To explore in a basic manner the role of clouds and precipitation on the climatic  
1144 impact of outflow channels, we performed a simulation of the reference outflow  
1145 channel event in which we eliminated the precipitation resulting from coalescence  
1146 ( $l_0 = \infty$ ). For this case, the vertical motion of the ice particles is governed only by  
1147 gravitational sedimentation. Figure 11 shows that the total cloud cover is near  
1148 100% over all the planet during the first year following the event, because of the  
1149 intense evaporation coupled with the increased lifetime of clouds.

1150 We found that neglecting coalescence and the subsequent precipitation led to  
1151 ice deposits that extend much more areally than in the reference case (Figure 18),  
1152 because the lifetime of ice particles increases substantially. In such a situation,  
1153 the global cloud cover (during the year following the event) has a net positive  
1154 radiative impact on the global energy balance of  $+ 12 \text{ W m}^{-2}$  ( $+ 21.3 \text{ W m}^{-2}$  of  
1155 IR warming;  $- 9.2 \text{ W m}^{-2}$  of solar absorption). This is  $\sim + 11 \text{ W m}^{-2}$  higher than  
1156 in the reference simulation.

1157 However, because the ice field produced by the event extends to a much larger  
1158 area, the global albedo increases and contributes approximately  $6 \text{ W m}^{-2}$  of cooling.  
1159 Moreover, because of advection processes, this also increases the horizontal extent  
1160 of the heat perturbation and thereby the global infrared emission to space. Under  
1161 clear sky conditions, this would lead to an extra cooling of  $\sim 5 \text{ W m}^{-2}$  compared  
1162 to the reference simulation.

1163 As a consequence, the total rate of cooling is more or less the same ( $\sim 15 \text{ W m}^{-2}$ )  
1164 as that in the reference simulation ( $l_0 = 0.001$ ). The duration of the warm phase is  
1165 also more or less the same than in the reference simulation ( $\sim 500$  days).

1166 We also note that the seasonal melting of the deposited ice (see Figure 19)  
1167 would be very limited in such scenarios, because of the increased solar reflection  
1168 by the clouds. In addition, because the ice field produced by the event extends  
1169 over a large region (Figure 18), the planet becomes much colder one year after the  
1170 event than initially.

1171 Nonetheless, we highly encourage further studies to explore in more detail the  
1172 possibility of warming early Mars through water ice clouds (as recently done by  
1173 **Ramirez and Kasting (2017)**).

### 1174 7.3 Conclusions

1175 In this analysis, we explored the climatic impact of a wide range of outflow  
1176 channel events under many possible conditions.

1177 We find that even considering outflow events with intensity (in volumes and  
1178 temperatures of water released) that exceed by far the most recent estimates, the  
1179 short term climatic response is still very limited. The duration of the 'warm' phase  
1180 that follows the outflow events is completely controlled by the total depth and  
1181 temperature of the lake that is formed and is, in practice, no more than few years  
1182 for the most extreme cases ( $10^7$  km<sup>3</sup> of water warmed at 300 K, e.g. ocean case).  
1183 In other words, outflow events fail to trigger greenhouse-sustained warm episodes.  
1184 Moreover, the precipitation (almost exclusively snowfall) produced by the events  
1185 during their warm phase is limited and confined to the Northern Plains, in the area  
1186 neighbouring the water outflow. These results are robust over a wide range of  
1187 atmospheric pressures and external conditions (e.g. obliquity and season).

1188 We also find that the intensity of outflow channel event effects can be significantly  
1189 influenced by the atmospheric pressure which is not well constrained for the  
1190 Hesperian era. Thin atmospheres ( $P < 80$  mbar), because of their low volumetric  
1191 heat capacity, can be warmed efficiently. This can trigger the formation of a  
1192 convective plume, a very efficient mechanism to transport water vapor and ice to  
1193 the global scale. Thick atmospheres ( $P > 0.5$  bar) have difficulty in producing  
1194 precipitation far from the outflow water locations but they are more suited to  
1195 generate snowmelt. Nonetheless, outflow channel formation events are unable,  
1196 whatever the atmospheric pressure, to produce rainfall or significant snowmelt at  
1197 latitudes below 40°N.

1198 During the 'cold phase' that follows the solidification to ice of the outflow  
1199 water, the body of water ice emplaced in the Northern Plains has a major contribution  
1200 to the water cycle. The ice is sublimated seasonally and transported progressively  
1201 southward toward the 'Icy Highlands' regions by the process of adiabatic cooling.  
1202 We find that under favorable conditions (obliquity  $\sim 45^\circ$ , atmospheric pressure  $\geq$   
1203 80 mbar), ice deposits can be stabilized in the West Echus Chasma Plateau area.  
1204 For an initial  $10^6$  km<sup>3</sup> body of water (0.2 bar atmospheric pressure,  $45^\circ$  obliquity),  
1205 they can be present during  $10^5$  martian years. However, seasonal melting related  
1206 to solar forcing seems difficult because 1) the West Echus Chasma Plateau is not  
1207 ideally located, and 2) the presence of (high albedo) snow at the surface has  
1208 a significant cooling effect. The global temperatures after outflow events can  
1209 thus easily be lowered by few Kelvins making the solar melting possibility even  
1210 more difficult. Therefore, in this scenario, localized warming such as geothermal

1211 activity or meteoritic impacts would be required to explain the formation of valley  
1212 networks dated to the Late Hesperian era and yet observed at this specific location.

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