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1 **Amazonian Northern Mid-Latitude**
2 **Glaciation on Mars:**
3 **A Proposed Climate Scenario**

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30 **ABSTRACT**

31 Recent geological observations in the northern mid-latitudes of Mars show
32 evidence for past glacial activity during the late Amazonian, similar to the
33 integrated glacial landsystems in the Dry Valleys of Antarctica. The large ac-
34 cumulation of ice (many hundreds of meters) required to create the observed
35 glacial deposits points to significant atmospheric precipitation, snow and ice
36 accumulation, and glacial flow. In order to understand the climate scenario
37 required for these conditions, we used the LMD (Laboratoire de Météorologie
38 Dynamique) Mars GCM (General Circulation Model), which is able to re-
39 produce the present-day water cycle, and to predict past deposition of ice
40 consistent with geological observations in many cases. Prior to this analysis,
41 however, significant mid-latitude glaciation had not been simulated by the
42 model, run under a range of parameters.

43 In this analysis, we studied the response of the GCM to a wider range of or-
44 bital configurations and water ice reservoirs, and show that during periods of
45 moderate obliquity ($\epsilon = 25\text{-}35^\circ$) and high dust opacity ($\tau_{dust} = 1.5\text{-}2.5$), broad-
46 scale glaciation in the northern mid-latitudes occurs if water ice deposited on
47 the flanks of the Tharsis volcanoes at higher obliquity is available for subli-
48 mation. We find that high dust contents of the atmosphere increase its water
49 vapor holding capacity, thereby moving the saturation region to the northern
50 mid-latitudes. Precipitation events are then controlled by topographic forc-
51 ing of stationary planetary waves and transient weather systems, producing
52 surface ice distribution and amounts that are consistent with the geological
53 record. Ice accumulation rates of $\sim 10 \text{ mm yr}^{-1}$ lead to the formation of a
54 500-1000 m thick regional ice sheet that will produce glacial flow patterns
55 consistent with the geological observations.

56 *Keywords:* MARS, ATMOSPHERE, DYNAMICS, CLIMATE, GEOLOGI-
57 CAL PROCESSES

58 **1 Introduction: Evidence for northern mid-latitude ice presence**
59 **and glaciation on Mars**

60 Evidence for the influence of non-polar ice deposition on geomorphic features
61 and processes became available as a result of the comprehensive global cov-
62 erage provided by the Viking Orbiter imaging system. For example, Squyres
63 (1978, 1979) attributed a variety of landforms (e.g., lobate debris aprons, lin-
64 eated valley fill, concentric crater fill, terrain softening) to the creep of the
65 martian regolith aided by the deformation of ground ice at latitudes higher
66 than $\sim 30^\circ$. Other workers (e.g., Lucchitta (1981)) noted that many of these
67 features appeared to represent not just ice-assisted creep, but rather more sub-
68 stantial glacial-like flow. More recently, new high-resolution data have shown
69 the presence of deposits interpreted to represent the remnants of extensive
70 glacial landsystems that formed in the parts of the northern mid-latitudes dur-
71 ing the Amazonian (e.g., Head et al. (2006b,a); Head and Marchant (2006);
72 Dickson et al. (2008)). These recent analyses show the widespread development
73 of valley glaciers, piedmont glaciers, plateau glaciation, and the development
74 of extensive glacial landsystems across the northern mid-latitudes (see Fig. 1).
75 Detailed examination of these deposits shows that ice may have reached thick-
76 nesses of up to 2-2.5 km in some regions along the dichotomy boundary (e.g.,
77 Head et al. (2006b,a); Dickson et al. (2008)). Clearly, the current atmosphere
78 and climate do not permit the accumulation of snow and ice at the level nec-
79 essary to produce such deposits. This raises the question: Under what past
80 climate conditions could the accumulation of snow and ice occur to produce
81 the types of glacial deposits seen in the northern mid-latitudes?

82 [Fig. 1 about here.]

83 Despite several climate modeling studies (Haberle et al., 2000; Mischna et al.,
84 2003; Levrard et al., 2007; Forget et al., 2006; Montmessin et al., 2007), the
85 origin of the northern mid-latitude glaciation has remained an enigma. Here,
86 we extend this previous work to a wider range of climate parameters, and show
87 that this broad-scale glaciation occurs if we assume that atmospheric dust con-
88 tent is higher than today (Newman et al., 2005), and that water ice deposited
89 on the flanks of the Tharsis volcanoes is available for sublimation (Forget
90 et al., 2006). Using the LMD (Laboratoire de Météorologie Dynamique) Mar-
91 tian Global Climate Model (Forget et al., 1999), we thus address the following
92 questions:

- 93 (1) What climatic mechanism can explain the formation of water-ice deposits
94 of hundreds of meters thickness in the northern mid-latitudes of Mars?
- 95 (2) What accounts for the regionally heterogeneous longitudinal distribution
96 of the deposits in the 30-50°N band?
- 97 (3) How are these glaciations related to orbital variations and can we deter-
98 mine the probable geologic periods of activity?
- 99 (4) What are the impacts of these deposits on the recent history of the mar-
100 tian water cycle?
- 101 (5) Is there any evidence for ice sequestration (removal of water ice from the
102 system) during these glacial phases?

103 After a short review of the recent climatic history and description of our
104 methods, we analyze in the following sections the climate of the northern
105 mid-latitude glaciation. Then, we study its sensitivity to climate parameters,
106 and finally discuss an updated climatic scenario for late Amazonian ice ages.

107 2 Geological evidence for orbital-driven climate change on Mars

108 2.1 Geomorphological settings

109 Accumulations of snow and ice, and glacial and periglacial landforms on Mars
110 exhibit a range of morphologies typical of different types of deposits, glaciers
111 and glacial subenvironments. Many of these show a stratigraphy which has
112 been interpreted to record climate shifts due to orbital variations. In addition
113 to the mid-latitude glacier deposits described in section 1, numerous other
114 examples of ice accumulation and glacial morphologies have been reported
115 (Head and Marchant, 2008), the major ones being 1) the North and South
116 Polar Layered Deposits, 2) the Latitude Dependent Mantle and 3) the Tropical
117 Mountain Glaciers.

118 - Polar layered deposits consist of alternating dark and bright layers of ice
119 mixed in different proportion with dust. They are visible on the walls of
120 the north polar cap, and form a thick stratigraphic sequence seen in outcrop
121 (Milkovich and Head, 2005) and in the subsurface with the SHARAD radar
122 instrument on board Mars Reconnaissance Orbiter (Phillips and 26 colleagues,
123 2008). Recent detailed analysis by the High-Resolution Imaging Science Ex-
124 periment (HiRISE) on board MRO revealed layers whose true thickness is as
125 low as 10 cm, and whose apparent brightness is not only the result of layer
126 composition, but also of surfacial frost and roughness (Herkenhoff et al., 2007),
127 explaining why the interpretation of the polar layered deposit frequency sig-
128 nals is so difficult (Laskar et al., 2002; Milkovich and Head, 2005; Levrard
129 et al., 2007).

130 - The north and south latitude dependent mantles are meters thick layered
131 deposits draped on both hemispheres above 50°, and present in partially de-
132 graded states from 30° to 50° latitude. They are revealed in MOLA data by
133 a latitudinal trend of roughness and concavity at 0.6 km baseline (Kreslavsky
134 and Head, 2000, 2002), and in MOC images by various latitude-dependent
135 geomorphologies (Mustard et al., 2001; Milliken et al., 2003). These results
136 led to the conclusion that the latitude dependent mantle was an ice and dust
137 cover of atmospheric origin, deposited during recent ice ages and currently
138 undergoing desiccation at lower latitudes (Head et al., 2003).

139 - Tropical mountain glaciers refer to large mountain glacial systems on the
140 western flanks of the Tharsis Montes and Olympus Mons (Head and Marchant,
141 2003). The largest of these, at Arsia Mons, covers an area of $\sim 170,000 \text{ km}^2$.
142 Exploration of cold-based glaciers in the Antarctic Dry Valleys, one of the
143 most Mars-like environments on Earth, has led to an understanding of the
144 cold-based nature of most Mars glaciers and the interpretation of cold-based
145 glacial deposits such as drop moraines, sublimation tills and debris-covered
146 glaciers (Marchant and Head, 2007). The identification of deposits interpreted
147 to result from cold-based glaciation in high-resolution images has permitted
148 the reconstruction of these tropical mountain glaciers (Head and Marchant,
149 2003; Shean et al., 2005, 2007; Milkovich et al., 2006; Kadish et al., 2008),
150 identification of the climatic conditions necessary for their formation (e.g.,
151 Forget et al. (2006)), and the formulation of glacial flow models consistent
152 with the geological features and settings (e.g., Fastook et al. (2008)). Mul-
153 tiple arcuate ridges have been interpreted as drop moraines, lobate deposits
154 represent debris-covered glaciers, and knobby terrain is interpreted to repre-
155 sent sublimation tills formed as the glaciers collapsed. These deposits show

156 numerous episodes of advance and retreat during the late Amazonian.

157 Ages obtained through crater size-frequency analyses span the period from
158 less than 10 Myr for the latitude dependent mantle (Head et al., 2003) to 10-
159 200 Myr for the tropical mountain glaciers (Shean et al., 2005). These data,
160 together with the distribution of several other latitudinally distributed ice-
161 related deposits (e.g., Head and Marchant (2008)) suggest long-term glacial
162 activity during the Amazonian.

163 *2.2 Mars orbital variations*

164 Climate changes on Mars are driven by insolation variations comparable to
165 terrestrial Milankovitch cycles. Spin-axis and orbital parameter variations of
166 Mars are much larger than on Earth, and their evolution can only be calculated
167 over a few millions of years (Laskar and Robutel, 1993) due to the strongly
168 chaotic nature of the solutions prior to this time. A robust solution for the last
169 10 Myr, however, has been derived by Laskar et al. (2004), and is currently
170 used as a guideline to explore recent climate changes. Variations of obliquity
171 and eccentricity are given on Fig. 2. Insolation varies with a short 51 kyr
172 period due to climatic precession, a 120 kyr period in obliquity, two 95 and
173 99 kyr periods in eccentricity, with the whole signal being finally modulated
174 with a 2.4 Myr period (Laskar et al., 2002).

175 [Fig. 2 about here.]

177 These geomorphological and orbital analyses suggest that under past orbital
178 configurations, atmospheric mechanisms were available and operating to de-
179 posit significant amounts of ice in widespread regions of Mars, from the lat-
180 itude dependent mantle covering at least 23% of the planet (Kreslavsky and
181 Head, 2002) to the regional mid-latitude valley and tropical mountain glaciers.

182 Concurrently, developments in climate modeling are providing robust tools
183 to explore this past water cycle (Haberle et al., 2000; Mischna et al., 2003;
184 Levrard et al., 2007; Forget et al., 2006; Montmessin et al., 2007). For example,
185 using the orbital calculations of Laskar et al. (2004) and the LMD/GCM
186 (Forget et al., 1999), Levrard et al. (2004) came to the conclusion that the
187 tropical mountain glaciers and the latitude dependent mantle can be formed
188 during periods of high (35-40°) and low (15-25°) mean obliquity respectively by
189 atmospheric water exchange between tropical and polar reservoirs. Indeed, the
190 north polar cap becomes unstable for obliquities higher than $\sim 35^\circ$, resulting in
191 a total water column of ~ 3000 pr. μm during the northern summer, compared
192 to a present-day observed value of ~ 60 pr. μm (Fouchet and 10 colleagues,
193 2007). Under low dust opacity conditions and 45° obliquity, precipitation and
194 deposition of ice on the western flanks of Tharsis Montes and Olympus Mons
195 by adiabatic cooling of strong westerly winds is then possible (Forget et al.,
196 2006), with accumulation rates up to 60 mm yr^{-1} . Given the known duration
197 of high obliquity excursions, such an accumulation rate can create ~ 3 km
198 thick glaciers on the volcanoes consistent with geological observations (e.g.,
199 Shean et al. (2005); Fastook et al. (2008)). Under the same orbital conditions
200 and by switching the water ice source from the north pole to the south pole,

201 a significant accumulation of ice is also predicted in eastern Hellas (Forget
202 et al., 2006) where evidence for glacial flow has been identified (Crown et al.,
203 1992; Head et al., 2005). When returning to lower obliquities and assuming
204 that tropical mountain glaciers are new sources of atmospheric water vapor,
205 Levrard et al. (2004) showed that an accumulation of ice of several millimeters
206 per year occurs above 60° in both hemispheres, providing a possible origin for
207 the latitude dependent mantle. In the same study, complete desiccation of the
208 tropical mountain glaciers finally leads to a decreasing amount of atmospheric
209 water vapor and retreat of the latitude dependent mantle to the poles, to form
210 the polar layered deposits. Montmessin et al. (2007) went one step further by
211 exploring the effect of a reversed perihelion 25 kyr ago, and found a possible
212 origin for the south residual water ice cap (Titus et al., 2003; Bibring et al.,
213 2004).

214 **3 Numerical simulations under new paleoclimatic conditions**

215 *3.1 Method of analysis*

216 Insolation variations result from changes in the obliquity ϵ , the eccentricity
217 e , and the areocentric longitude of the Sun at perihelion L_p (see for instance
218 Fig. 1 of Armstrong et al. (2004)). These three parameters are able to produce
219 major changes in insolation, atmospheric circulation, and thus in dust lifting
220 rates and distribution of surface water-ice. Consequently, to define a climatic
221 scenario, we first need to assign a given orbital configuration (ϵ, e, L_p) . Be-
222 cause the dust cycle is not predicted by the climate model, we also need to
223 assign a dust content of the atmosphere under these past orbital conditions,

224 given by the dust optical depth τ_{dust} at $0.67 \mu\text{m}$ (Pollack et al., 1979). τ_{dust} is
 225 typically equal to 0.2-0.4 under common conditions, but can reach values of
 226 1-2.5 during local dust storms, and 5 during global dust storms. Finally, the
 227 location of surface water-ice reservoirs (WIR) must be specified to generate a
 228 water cycle. These locations are given by the acronyms N/SPC (North/South
 229 polar cap), LDM (latitude-dependent mantle) and TMG (tropical mountain
 230 glaciers). A climatic scenario is then a point in the parameter space defined by
 231 $\underline{x} = \{\epsilon, e, L_p, \tau_{dust}, \text{WIR}\}$. We ran a set of 5.625×3.75 -degree resolution simu-
 232 lations limited to the lower atmosphere (from the ground to $\simeq 50 \text{ km}$) in the
 233 parameter space defined by $\epsilon = (15, 25, 35, 45^\circ)$, $e = (0, 0.1)$, $L_p = (90, 270^\circ)$,
 234 $\tau_{dust} = (0.2, 1, 2.5)$, and $\text{WIR} = (\text{NPC}, \text{SPC}, \text{TMG})$.

235 Given the large number of combinations, studying the specific climatic origin
 236 of the mid-latitude glaciation presents an imposing task. However, it is pos-
 237 sible to approach solutions by systematically comparing the predicted water
 238 cycle to geological observations. Among the two dozen simulations performed
 239 with this approach, the best conditions for development of the mid-latitude
 240 glaciation was obtained at point $x_{ref} = (35^\circ, 0.1, 270^\circ, 2.5, \text{TMG})$. Coordinates
 241 of the other simulations of interest will further be noted as x_i , and analyzed
 242 in the sensitivity study of section 5.2. x_0 refers to present-day conditions (i.e.
 243 $\epsilon = 25.19^\circ$, $e = 0.093$, $L_p = 251^\circ$) with a time-varying dust opacity that follows
 244 the observations of Mars Global Surveyor Thermal Emission Spectrometer ac-
 245 quired during the martian year 24 (1999-2000, Smith (2004)), and a water-ice
 246 reservoir corresponding to the north polar cap. All the results will be shown
 247 for the sixth year of simulation, once an equilibrium is reached.

249 The terrestrial GCM of the Laboratoire de Météorologie Dynamique (LMD,
250 Paris) has been adapted to Mars in 1989 by changing the radiative transfer and
251 adding the CO₂ cycle (Forget et al., 1998). It was the first model to reproduce a
252 self-working Mars, and pressure variations consistent with the Viking Landers
253 observations of transient weather systems (Hourdin et al., 1993). Since then,
254 two teams from the LMD and the department of Atmospheric, Oceanic and
255 Planetary Physics (AOPP), Oxford, have been developing two grid-point and
256 spectral models in parallel, with a full description of the planetary boundary
257 layer, the propagation of gravity waves, the dust cycle (Newman et al., 2002),
258 the water cycle (Montmessin et al., 2004), the role of the regolith (Böttger
259 et al., 2005), the thermosphere (Angelats i Coll et al., 2005), the ozone photo-
260 chemistry (Lefèvre et al., 2004), and the HDO cycle (Montmessin et al., 2005).
261 A detailed description of the model can be found in Forget et al. (1999), and
262 we only review here the relevant processes of the Mars water cycle, such as
263 the dust and water radiative effects, the formation of water-ice clouds, and
264 the stability of surface ice deposits.

265 3.2.1 *Dust radiative effects:*

266 From the observed or assigned values of τ_{dust} , a dust mixing ratio is deduced
267 at a reference pressure level of $p_0 = 700$ Pa, the vertical profile being then
268 set to constant under this pressure level, and decreasing above (Forget et al.,
269 1999). Visible and near-infrared radiative effects of dust are taken into account
270 by calculating the radiative transfer in two different bands : 0.1-0.5 μm and
271 0.5-5 μm . Thermal IR absorption and emission of dust is also computed in the

272 silicate band (5-11.5 μm) and the rest of the IR band (20-200 μm). The ratio
273 between the visible and the 9 μm dust opacity ($\tau_{0.67\mu\text{m}}/\tau_{9\mu\text{m}}$) is essential to
274 predict the energy distribution between the atmosphere and the surface, and
275 a value of 2 is assigned in the GCM, following a method described in Forget
276 (1998).

277 *3.2.2 Water radiative effects:*

278 At the moment, water vapor and cloud radiative effects are neglected in the
279 GCM. Water vapor column can be increased by an order of magnitude under
280 past conditions (Mischna et al., 2003; Forget et al., 2006), and broad scale
281 cloud cover can reach significant optical thickness. Clouds change the lower
282 atmosphere temperature by both reflecting more sunlight in the shortwave
283 domain, and increasing the thermal infrared opacity at cloud altitude (Wilson
284 et al., 2007). Resulting temperature change feeds back on nucleation rates,
285 water vapor amount and dust scavenging by ice particles (Rodin et al., 1999).
286 Development of a new model, accounting for these complex processes, is un-
287 derway.

288 *3.2.3 Water transport and cloud formation:*

289 Water vapour and ice crystals are taken into account in the model by adding
290 two (radiatively passive) tracers that are advected conservatively by the Global
291 Climate Model. A Van Leer scheme I (van Leer, 1977; Hourdin and Armen-
292 gaud, 1999) accounts for the global transport of these tracers, whereas subgrid
293 processes, i.e. turbulent mixing and convective adjustment, are parameterized
294 with a diffusion equation and an energy conserving scheme, respectively (Hour-

295 din et al., 1993; Forget et al., 1999). The crystal size is predicted by the cloud
 296 microphysics described in Montmessin et al. (2002). When a parcel of the
 297 atmosphere is supersaturated, the increasing mass of water-ice is predicted
 298 by taking into account the number of nuclei, which is a function of the dust
 299 content of the atmosphere. Consequently the model accounts for ice-crystal
 300 microphysics and interaction with dust nuclei. The model also accounts for
 301 sedimentation of ice crystals (Montmessin et al., 2004), but scavenging of at-
 302 mospheric dust and subsequent feedbacks are not yet taken into account.

303 3.2.4 *Stability of surface ice deposits:*

304 Soil and surface temperatures depend on the balance between incoming fluxes
 305 and thermal conduction in the soil (Hourdin et al., 1993). The conduction
 306 equation is calculated by using the thermal inertia deduced from IRTM and
 307 TES observations. Sublimation of surface ice deposits is controlled by the
 308 surface turbulent flux:

$$309 \quad F_w = \rho_1 C_d U_1 (q_{sat} - q_{wv}), \quad (1)$$

310 where ρ_1 is the atmospheric density in the first layer of the GCM (up to 5 m
 311 high), U_1 the wind speed in the same layer, and C_d the drag coefficient (Forget
 312 et al., 1999; Montmessin et al., 2004). q_{sat} is the saturation mass mixing ratio
 313 of water vapor at the ground temperature, and q_{wv} is the actual mass mixing
 314 ratio of water vapor in the first layer. In other words, the surface ice sublimates
 315 if the first layer of the model is not saturated, i.e. $q_{sat} > q_{wv}$. On the contrary,
 316 ice can build up if $q_{sat} < q_{wv}$, or if the atmosphere itself is supersaturated, in
 317 which case precipitation leads to deposition of ice crystals onto the surface. It
 318 is worth noting that $q_{sat} \propto p_{sat}/p_s$, where p_{sat} is the saturation pressure and p_s

319 the actual surface pressure. Given the fact that saturation pressure exponen-
320 tially increases with temperature (Clausius-Clapeyron equation), sublimation
321 is dependent, to first order, on summer temperatures.

322 The positive ice-albedo feedback is also taken into account by setting the
323 surface albedo to 0.4 when deposits are more than $5 \mu\text{m}$ thick. Thermal inertia
324 feedback is not activated until section 5.3, where its impact on surface water
325 ice stability is analyzed. Finally, water vapor is not allowed to diffuse into the
326 regolith.

327 **4 Results: From present-day climate to the northern mid-latitude** 328 **glaciation**

329 To explore the meteorological conditions under which a regional ice sheet
330 may have accumulated and grown in the northern mid-latitudes, we describe
331 in the subsequent sections the manner in which the Martian climate system
332 can evolve from its present-day interglacial state to an ice age by gradually
333 changing the orbital parameters, the water-ice sources and the dust content
334 of the atmosphere. Our analysis will be based on Fig. 3, 4 and 5, that re-
335 spectively show the annual evolution of the water cycle, the zonal structure
336 of the northern winter atmosphere, and the relationship between stationary
337 planetary waves and cloud distribution at this season, each for three different
338 simulations. The first one corresponds to present-day climate, while the sec-
339 ond and third ones correspond to 35° obliquity and equatorial water source
340 under clear ($\tau_{dust} = 0.2$) and dusty ($\tau_{dust} = 2.5$) conditions, respectively. We
341 will particularly focus on the onset of the mid-latitude precipitation activity
342 during late northern fall ($L_S = 240\text{-}270^\circ$).

343 *4.1 The present-day water cycle*

344

[Fig. 3 about here.]

345 Following the mapping of water vapor by the Viking Mars Atmospheric Wa-
346 ter Detector (MAWD) in the late 1970s (Jakosky and Farmer, 1982), infrared
347 spectrometer data from the Mars Global Surveyor, Mars Express and Mars
348 Reconnaissance Orbiter missions are providing a detailed picture of the Mars
349 water cycle from global to regional scale. Annual zonal-mean evolution of the
350 water cycle, as simulated by the LMD/GCM, is illustrated in Fig. 3.a, with
351 contours corresponding to the precipitable water vapor column in microme-
352 ters (pr. μm), and shaded regions to water-ice clouds (same unit). The main
353 source of the water cycle comes from the release of up to 60 pr. μm of water
354 vapor during late northern spring and summer, as observed by Smith (2002);
355 Fouchet and 10 colleagues (2007). Visible and near-infrared imaging of the
356 northern polar regions (see Fig. 11 of Wang and Ingersoll (2002); Bibring and
357 10 colleagues (2005); Langevin et al. (2005)) during this period revealed the
358 recession of a seasonal CO_2 frost (James and Cantor, 2001; Kieffer and Titus,
359 2001) followed by a fine-grained H_2O ice edge, whose sublimation finally ex-
360 poses the perennial H_2O cap to summer insolation. All this results in the slow
361 increase of the water vapor column observed in Fig. 3.a, between $L_S = 50^\circ$
362 and $L_S = 120^\circ$.

363 *4.1.1 Aphelion cloud belt and the “Clancy” effect:*

364 Then water vapor reaches the subsolar point, where advection into the over-
365 turning Hadley cell occurs (see the descending 25 pr. μm isocontour in Fig. 3.a,

366 and previous work by Houben et al. (1997); Richardson and Wilson (2002)).
367 Adiabatic expansion and cooling of the rising parcels increases their relative
368 humidity, until condensation and formation of the Aphelion cloud belt oc-
369 curs (Kahn, 1984; Clancy et al., 1996; Wolff et al., 1999; Wang and Ingersoll,
370 2002). These clouds appear during mid-spring and summer around the north-
371 ern tropics, as predicted in Fig. 3.a (see the water-ice rich shaded region in the
372 $\pm 30^\circ$ latitude band around $L_S = 90^\circ$). Sublimation of the Aphelion cloud belt
373 around $L_S = 150^\circ$ is mainly due to a warming of the atmosphere as we leave
374 the Aphelion season (Richardson et al., 2002), during which the condensation
375 level (known as the hygropause) was low, and thus favorable to cloud forma-
376 tion. The existence of low hygropause conditions in the rising branch of the
377 Hadley cell tends to retain water vapor in the summer hemisphere, a key phe-
378 nomenon first described in Clancy et al. (1996), and since called “the Clancy
379 effect”. Its paleoclimatic implications when aphelion occurred during southern
380 summer 25 kyr ago have been studied by Montmessin et al. (2007), who used
381 the LMD/GCM to explain the possible origin of the south residual cap (see
382 section 2). We will see that the Clancy effect is also key to understanding the
383 northern mid-latitude glaciation.

384 *4.1.2 Onset of the northern winter season:*

385 As summer comes to the southern uplands and cold polar night arrives in
386 the northern high latitudes, CO_2 and H_2O ice start to sublime in the south
387 polar regions, while the northern atmosphere approaches saturation, to form
388 the polar hood.

389 Considering the departures from zonally symmetric flow is essential to capture

390 the behaviour of the south seasonal cap and north polar hood. It is here
391 instructive to further describe this present-day northern winter climate, before
392 analyzing the changes that occurred during glacial excursions.

393 [Fig. 4 about here.]

394 Figure 4.a illustrates the zonal mean circulation predicted by the LMD/GCM
395 for the $L_S = 240-270^\circ$ period. It is characterized by 1) an intense meridional
396 temperature gradient that results in, and from, a westerly (west \rightarrow east) jet
397 in the winter hemisphere and an easterly jet in the southern tropics, 2) a
398 cross-equatorial Hadley circulation that adiabatically heats the atmosphere in
399 its descending branch, explaining the comma-shaped temperature inversion in
400 the northern mid-latitudes (Haberle et al., 1993; Forget et al., 1999), and 3) a
401 surface westerly jet around 30°S , resulting from eastward acceleration of the
402 flow in the return branch of the Hadley cell through momentum conservation
403 (Joshi et al., 1997). The zonal-mean water vapor distribution during the same
404 period is illustrated in Fig. 4.b along with the ice mixing ratio in Fig. 4.c. Two
405 regions of enhanced water vapor can be distinguished, around 60°S and 30°N ,
406 that correspond respectively to the sublimation of the southern seasonal cap
407 and northward transport by the upper-branch of the Hadley cell. These general
408 dynamics are clearly highlighted by the $300 \cdot 10^{-6} \text{kg kg}^{-1}$ line in Fig. 4.b. An
409 equatorial cloud belt appears in Fig. 4.c at much higher altitudes than the
410 Aphelion one, and thus allows water vapor to be advected in the northern
411 hemisphere, and to finally reach the winter polar vortex.

412 This water vapor condenses at the edge of the polar night, where poleward
413 advection of water vapor by the mid-latitude non-axisymmetric circulation
414 plays a major role in the formation of the polar hood (see the enhanced ice

415 mixing ratio around 50°N in Fig. 4.c). These departures from zonal-mean
416 circulation are due to traveling and stationary waves. These waves result in
417 poleward or equatorward advection of tracers and horizontal mixing.

418 [Fig. 5 about here.]

419 A more detailed understanding of the polar hood structure can thus be gained
420 through the analysis of the stationary vortex asymmetries and the resulting
421 cloud pattern. Figure 5.a is a polar stereographic map of the mean zonal
422 wind at the 5.6-km level (shaded colors, m s^{-1}) and total water-ice column
423 (contours, $\text{pr. } \mu\text{m}$). Three intense jet cores appear in the Alba Patera region
424 (left shaded area in Fig. 5.a), the Deuteronilus-Protonilus Mensae area (lower-
425 right), and Arcadia Planitia (upper-right), and are comparable to the western
426 Pacific, Atlantic and north African jets on Earth. Theoretical analysis by
427 Hollingsworth and Barnes (1996) and Nayvelt et al. (1997) demonstrated that
428 at mid-to-high latitudes, stationary waves were mainly mechanically forced by
429 the Tharsis, Arabia Terra and, to a lesser extent, Elysium ridges, explaining
430 their wavenumber-2 and 3 dominant pattern. Based on momentum conserva-
431 tion, regions where westerlies are accelerated (decelerated) in Fig. 5.a corre-
432 spond to poleward (equatorward) advection of warm (cold) air masses across
433 the polar front.

434 Consequently, air parcels are lifted to saturation as they enter and leave the jet
435 cores, and stationary waves thus favor cloud formation in the Acidalia, Utopia
436 and Arcadia basins, as indicated by the arrows in Fig. 5.a. The contribution
437 of the waves to polar hood formation has been quantified by Montmessin
438 et al. (2004), and is apparent from many observations, for instance by James
439 et al. (1996), Wang and Ingersoll (2002) (Fig. 2.g) or Tamppari et al. (2008).

440 This same mechanism will be central to the development of the mid-latitude
441 glaciation.

442 *4.2 Excursion to 35° obliquity under clear conditions*

443 We now step back in geological time and suppose that obliquity is leaning
444 towards an intermediate value of 35° following an excursion to higher values.
445 As explained in section 2, tropical mountain glaciers can be formed at around
446 45° obliquity on the western flanks of the Tharsis volcanoes (Forget et al.,
447 2006). Here we assume that they are the only source of water on the planet,
448 with both polar caps being exhausted. The simulation is thus initialized by
449 placing water-ice reservoirs on the volcanoes (as indicated by red squares in
450 Fig. 7), and by setting the eccentricity and solar longitude of perihelion L_p to
451 0.1 and 270°, respectively. In the first approach, we assume low dust conditions,
452 i.e. $\tau_{dust} = 0.2$.

453 Under such a scenario, the water cycle is fed by sublimation of the tropical
454 mountain glaciers. This occurs especially when Mars is at perihelion, as illus-
455 trated in Fig. 3.b by the appearance of two symmetric 100 pr. μm lobes and
456 increased water vapor column as we approach the northern winter solstice.
457 Many characteristics of the present-day water cycle are conserved, for exam-
458 ple the two polar water vapor maxima near solstices, or the Aphelion cloud
459 belt at $L_S = 60\text{-}120^\circ$. However, a general one order of magnitude increase in
460 water vapor and ice column results from the significant and constant supply
461 by the tropical mountain glaciers, whereas water can only be provided during
462 northern summer under current conditions, and at lower rates. As a result,
463 saturation is reached at much higher temperature, and the hygropause is low-

464 ered by several kilometers. A tropical cloud belt is thus formed all year long,
465 as already noted by Mischna et al. (2003), who analyzed the same orbital con-
466 figuration, but using simplified cloud microphysics and the north polar cap as
467 the only source of water.

468 This effect is clearly apparent if we focus on the northern winter season, whose
469 zonal-mean atmospheric structure is illustrated in Fig. 4.d. Cross-equatorial
470 Hadley cells are known to be intensified at higher obliquity, due to the in-
471 creased meridional gradient of surface temperature (Fenton and Richardson,
472 2001; Haberle et al., 2003). This appears in Fig. 4.d through acceleration of
473 the high-level easterly and low-level westerly jets. Water vapor, as shown in
474 Fig. 4.e, is transported in the ascending branch of the cell, but trapped by
475 early saturation of the parcels in the lowered hygropause region. A thick trop-
476 ical cloud belt, obvious in Fig. 4.f, thus forms 20 km below the condensation
477 level we observe today at the same period (Smith, 2002).

478 Despite the trapping of water vapor in the southern hemisphere, sublimation
479 of the tropical mountain glaciers allows water vapor to be still available at the
480 edge of the polar vortex, as indicated by the bulge of the $2000 \cdot 10^{-6} \text{kg kg}^{-1}$
481 contour towards the northern tropics in Fig. 4.e. This creates a secondary
482 cloud belt centered at 30°N (see Fig. 4.f), characterized by a water-ice mixing
483 ratio of $300 \cdot 10^{-6} \text{kg kg}^{-1}$, i.e. two times more than what the model predicts
484 in the Aphelion cloud belt today. Further analysis of this region highlights
485 a wavenumber-3 structure of the cloud belt, linked to the entrance and exit
486 regions of the polar jets and resulting eddy heat fluxes. This relationship is
487 illustrated in Fig. 5.b that shows increased water-ice columns in the lee of
488 the Tharsis, Arabia and Elysium topographic barriers. This resembles the
489 well-known structure of the polar hood borders (Fig. 5.a), that has the same

490 dynamical origin.

491 However, under these past conditions, the latitude at which condensation oc-
492 curs is moved equatorward by the increased water vapor content of the atmo-
493 sphere, and advection of water vapor across the polar front by the stationary
494 waves dominates over the simple condensation that occurs in the polar night.
495 A new climate system is thus emerging, in which the main active regions, and
496 possible precipitation, are located in the mid-latitudes, instead of the high-
497 latitudes. Consequently, surface condensation and deposition of fine water-ice
498 crystals extend down to 30°N in the regions of enhanced cloud formation, form-
499 ing seasonal deposits that are up to 1 mm thick in some areas of the northern
500 mid-latitudes. These paleoclimatic conditions probably represent quiet peri-
501 ods of slow deposition between dusty episodes, thought to be frequent under
502 such obliquity. One further step is thus required to formulate a more realistic
503 scenario.

504 *4.3 Increasing the dust content of the atmosphere*

505 As explained earlier, Hadley and monsoon circulation are strengthened as the
506 obliquity increases, probably resulting in frequent dust storms at both solstices
507 (Haberle et al., 2003). Dust cycle modeling under past orbital conditions led
508 Newman et al. (2005) to conclude that "huge amounts of lifting" are already
509 produced at 35° obliquity. Based on these previous studies, we performed a
510 reference simulation x_{ref} with a dust opacity of 2.5, chosen to portray the
511 effect of frequent dust storms on the water cycle.

512 *4.3.1 Dust-induced global change:*

513 The water cycle under dusty conditions is illustrated in Fig. 3.c, and shows
514 water vapor and ice column values that both increase by a factor of ~ 6
515 outside the polar regions compared to the cycle given in Fig. 3.b. A first order
516 explanation lies with the net radiative effect of dust, which is generally to
517 warm the atmosphere above 10 km, because of the large dust opacity at visible
518 compared to infrared (the anti-greenhouse effect). During northern winter,
519 comparison of clear and dusty simulations in the northern mid-latitudes thus
520 reveals a ~ 10 K cooling in the first 10 km of the atmosphere, whereas overlying
521 layers are warmed by up to 40 K. Consequently, in a dusty atmosphere, upper
522 layers and summer hemisphere, through increased saturation vapor pressure,
523 can hold much more water vapor, while the cold lower atmosphere, especially
524 in the winter hemisphere, favors ice nucleation and cloud formation. Activity
525 in the winter hemisphere is thus characterized by a thick cloud belt in the
526 northern mid-latitudes that contrasts with the thin polar night hazes. This
527 mid-latitude cloud belt is revealed in Fig. 3.c by intense condensation between
528 $L_S = 210^\circ$ and $L_S = 300^\circ$, during which the ice column can be 2 orders of
529 magnitude higher than what exists today on Mars.

530 What can explain such a large amount of ice in the northern mid-latitudes?
531 Zonal-mean winter circulation, shown in Fig. 4.g, gives further details on the
532 changes induced by dust storm conditions. Warming of the atmosphere is
533 clearly apparent, especially in the southern hemisphere, and results in an even
534 more intense meridional circulation than before, when it was already acceler-
535 ated by the high obliquity and perihelion conditions. The polar warming, due
536 to adiabatic compression of air parcels in the subsiding branch of the Hadley
537 cell (Wilson, 1997; Forget et al., 1999), produces a large inversion that ex-

538 tends up to the north pole. The low-level westerly jet is two times faster than
539 today, and is thought to play the main role in lifting dust and warming the
540 southern hemisphere, as shown for example by Newman et al. (2005) or Basu
541 et al. (2006). Finally, the polar vortex is significantly weakened under dusty
542 conditions, as already noticed by Newman et al. (2002), for dynamical reasons
543 that are beyond the scope of this paper, but possibly related to the intense
544 polar warming. These changes in atmospheric thermodynamics have a wide
545 range of impacts on the water cycle.

546 Warming of the southern hemisphere elevates the hygropause 20 km higher
547 than the previous simulation, and allows water vapor to be transported from
548 the equatorial reservoir to the intense upwelling zone of the Hadley cell, and up
549 to the upper atmosphere (see the first water vapor maximum around 50°S on
550 Fig. 4.h). Then, a large part of this water vapor crosses the equator and builds
551 up along the polar vortex (second maximum around 30°N on Fig. 4.h), while
552 another part condenses out to form a thin high-altitude cloud belt centered
553 on the equator, and apparent in Fig. 4.i. Around 10 g kg⁻¹ of water vapor
554 thus reaches the polar front, and is available for cloud formation.

555 [Fig. 6 about here.]

556 4.3.2 *The mid-latitude cloud belt:*

557 Meanwhile, winter eddy circulation starts in the northern mid-latitudes, and
558 controls the cross-front mixing of moisture. The structure of the stationary
559 waves is clearly modified compared to non-dusty conditions, as illustrated in
560 Fig. 5.c. While the Tharsis jet strength remains nearly unchanged with more
561 than 30 m s⁻¹ zonal wind speed, Arabia and Elysium jets damp down to 25 and

562 20 m s^{-1} respectively. The acceleration pattern of Elysium is mainly preserved,
563 whereas a southwest-northeast oriented jet now covers the northern part of
564 Arabia Terra, before being abruptly decelerated in Utopia Planitia. Figure 6
565 gives details on this winter circulation, and represents a horizontal wind field
566 at the 5.6 km level, superposed on the atmospheric water-ice column in μm .

567 The wavenumber-3 jet structure is clearly pronounced in the northern mid-
568 latitudes, and the jet is far from zonal, as opposed to clear conditions. Poleward
569 motion of the flow peaks over Deuteronilus Mensae, Phlegra Montes and the
570 western flank of Alba Patera, reaching a meridional velocity of $\sim 8 \text{ m s}^{-1}$ at
571 the 5.6 km level, whereas Amazonis, Chryse and Utopia basins are regions
572 of equatorward motion, with meridional speed up to 6, 10 and 16 m s^{-1} , re-
573 spectively. The largest modification occurs in Deuteronilus Mensae, where the
574 flow was almost exclusively zonal under non-dusty conditions. Barnes et al.
575 (1996) noticed similar trends under current orbital conditions by increasing
576 dust opacity from 0.3 to 2.5, and found strong meridional acceleration over
577 Deuteronilus and Elysium at the $\sim 18 \text{ km}$ level (Barnes et al. (1996), Fig. 19),
578 with an increase in wavenumber 2 amplitude. Interestingly, TES observed the
579 same increased wave 2 amplitude at 45°E (Banfield et al. (2003), Fig. 10) dur-
580 ing winter, which corresponds again to the Deuteronilus-Protonilus region.
581 Hollingsworth and Barnes (1996) also observed a tendency of wavenumber 2
582 to be "squashed" in mid-latitudes as dust content is increased, with "larger
583 poleward momentum flux" south of 45°N . According to Nayvelt et al. (1997),
584 low-level eddy circulation north and south of 45°N is dominated by mechanical
585 and radiative forcing of topography, respectively (Fig. 10.a of Nayvelt et al.
586 (1997)). Comparing this with Fig. 6, it appears that the eddy flux north of the
587 $150 \mu\text{m}$ white line is mainly mechanically forced, while the largest water

588 vapor advection south of this line is due to the radiative forcing response, for
589 instance in Amazonis and Chryse Planitia, north of Syrtis Major, as well as
590 on each side of Elysium. Heat flux and resulting condensation thus seem to
591 be mainly controlled by radiative forcing of stationary waves in mid-latitudes.
592 Furthermore, this forcing is enhanced by the stronger inversion that occurs in
593 the lower dusty atmosphere, which results in larger lateral temperature gradi-
594 ents, with low-lying regions up to 20 K colder at 45°N than the surrounding
595 plateaus.

596 Clouds are thus located along the zonally symmetric water vapor gradient
597 (white lines in Fig. 6) in these regions of large eddy activity, and appear in
598 both Fig. 5.c and 6 as opaque cloud covers in Tempe Terra, west and east
599 of Amazonis Planitia, and in a southwest-northeast oriented zone, from west-
600 ern Arabia to Utopia. The water-ice column attains 800 pr. μm , which is two
601 orders of magnitude thicker than the current Aphelion cloud belt. Low-level
602 condensation processes create ice crystals more than 7 μm in radius in the
603 northern mid-latitudes, similar to the largest particle predicted in the Polar
604 Hood on present-day Mars (Montmessin et al., 2004). Under these conditions,
605 mean annual precipitation in some regions of the mid-latitudes is of the order
606 of 10 mm yr⁻¹, with surface temperature around -75°C. This is clearly differ-
607 ent from the non-dusty simulation, where only transient snow deposits of a
608 few micrometers were present. Here, ice is accumulating at a pace that could
609 lead to regional ice sheet formation, with precipitation and temperature that
610 correspond to inferred conditions of the last glacial maximum in Antarctica
611 (Marchant and Head, 2007).

612 5 Accumulation rates and dependence on climate parameters

613 5.1 Annual water-ice budget

614 Glaciers represent an equilibrium between winter precipitation and summer
615 sublimation. Precipitation and accumulation of ice during winter are not suf-
616 ficient to produce glaciers, whose distribution thus depends on summer subli-
617 mation, and to first order on summer peak temperatures (the dominant term
618 of Eq. 1). The upper panel of Fig. 7 illustrates the net annual snow accumu-
619 lation (color shaded in mm yr^{-1}), along with the winter accumulation and
620 summer sublimation rates in the lower left and lower right panel, respectively.
621 Model predictions are superposed on the map by Squyres (1979), which shows
622 the specific location of several different types of ice-related features (see also
623 Fig. 1).

624 [Fig. 7 about here.]

625 Three main regions of winter precipitation clearly appear in Fig. 7.b. One area
626 of enhanced winter precipitation occurs around Alba Patera (west, north and
627 east), then a second one north of Arabia Terra in the Deuteronilus-Protonilus
628 Mensae region and, finally, a third one between the Tharsis and Elysium rises.
629 All three of these regions correspond to the largest concentration of ice-related
630 landforms noticed by Squyres (1979), and concentrations of glacial landforms
631 mapped by Head and Marchant (2006) (see Fig. 1).

632 Seasonal variations of surface ice deposits in four different regions pointed out
633 in Fig. 7 are represented in Fig. 8, and clearly show constant precipitation
634 during winter, with a period of enhanced accumulation (represented by the

635 second shaded region) at the beginning of this season. A detailed analysis
636 of this period reveals the passing of low pressure systems due to baroclinic
637 waves, with precipitation events of around $10 \mu\text{m}$. Associated cold fronts and
638 surface temperature variations are seen in the lower panel of Fig. 8. These
639 waves, stabilized by the topographic high of the dichotomy boundary, could
640 play a major role in the mid-latitude glaciation, because they also result in
641 the formation of dust storms. Large weather systems could therefore arise
642 from these local dust storms, propagating along the dichotomy boundary and
643 creating ideal conditions for nucleation of ice crystals on dust nuclei over
644 glacial regions.

645 [Fig. 8 about here.]

646 Winter deposits are only preserved on the northern flank of Alba Patera and
647 along the scarp of the Deuteronilus-Protonilus region, while the others, north
648 of Olympus Mons and on the uplands of DPM, are lost by summer sublima-
649 tion (see Fig. 7.c). This sublimation is also visible in Fig. 8, and highlighted
650 by the first shaded region. Interestingly, ice builds up even during summer in
651 Nilosyrtris Mensae and Phlegra Montes (lines # 3 and 4), where precipitation
652 is brought by the dry western boundary currents that meet humid air masses
653 of the northern summer. Otherwise, summer sublimation explains why the net
654 annual accumulation regions in Fig. 7.a do not only reflect winter precipita-
655 tion (see the cloud cover of Fig. 6), but also summer peak temperatures and
656 resulting sublimation. These surface temperatures are controlled by, in order
657 of decreasing impact, thermal inertia, albedo, local slope, atmospheric dust
658 and altitude. In this simulation, our model only accounts for dust and alti-
659 tude effects, with an arbitrary change in surface albedo when ice is present.
660 Thermal inertia is equal to current observed values, and its changes induced

661 by deposition of ice on the surface is considered in section 5.3. Atmospheric
662 dust reduces diurnal temperature variations by producing lower short-wave
663 fluxes during the day and increased long-wave fluxes during the night, and
664 induces in most cases mean surface cooling of the summer mid-latitudes. Con-
665 sequently, dust storm conditions not only favor winter precipitation, but also
666 reduce summer sublimation.

667 Altitude has a very small impact on surface temperature due to the low den-
668 sity of the atmosphere. This is in sharp contrast to the terrestrial case, in
669 which glaciers flow down into lower and warmer environments, and undergo
670 melting or sublimation. Mars glaciology might be significantly different than
671 that of the Earth, with an unexpected glacier mass balance due to changes in
672 equilibrium line approach, as mentioned in Fastook et al. (2008). Sublimation
673 dependence on elevation not only involves temperature, but also water vapor
674 circulation and resulting humidity found in glacial valleys. These questions are
675 being addressed using mesoscale atmospheric modeling and glacial modeling
676 (Fastook et al., 2009).

677 It is worth noting that ice also builds up in Valles Marineris, Terra Meridiani
678 and southern Elysium (see Fig. 7.a). However, these regions have a relatively
679 high thermal inertia, which often results in lower summer sublimation, and
680 this forcing could bias our results, as thermal inertia has probably varied
681 through the Amazonian. To assess this effect, we set the thermal inertia and
682 albedo to $217 \text{ J s}^{-1/2} \text{ m}^{-2} \text{ K}^{-1}$ and 0.23 everywhere on Mars, which corre-
683 spond to planetary mean values, and reran the reference simulation. Interest-
684 ingly, the ice distribution of Fig. 7 remains broadly unchanged in the northern
685 mid-latitudes, confirming that winter atmospheric precipitation controlled by
686 topography is the main accumulation process. However, accumulation rates

687 in Terra Meridiani and southern Elysium are reduced to a few millimeters,
688 and ice disappears in Valles Marineris. This suggests that present-day ther-
689 mal inertia was the main cause of preferential accumulation in these regions.
690 Consequently, despite potentially interesting, we cannot further comment on
691 these deposits in the present state of our model.

692 5.2 *Sensitivity to climate parameters*

693 The way glaciers are affected by perturbations in climate parameters is crucial
694 to understanding their geologic history, and the relationship between orbital
695 forcing and glacial stratigraphy. Following the orbital calculations of Laskar
696 et al. (2004), we address this issue by disturbing the reference conditions
697 (i.e. the ones leading to the glaciation of Fig. 7) and analyzing changes in
698 precipitation activity and ice preservation. We will also study the impact of
699 changing thermal inertia as the surface is being covered with ice.

700 Results are summarized in Table 1, that will be used to monitor water-ice
701 reservoir and atmospheric response to different forcings, and in Fig. 9, that
702 illustrates the different accumulation rates obtained for each sensitivity exper-
703 iment.

704 [Table 1 about here.]

705 [Fig. 9 about here.]

706 5.2.1 *The dust cycle: Main actor in the glaciation*

707 As underlined in section 4, increased atmospheric dust content is the key factor
708 allowing the initiation of the northern glaciation, by increasing water vapor
709 holding capacity, moving the saturation region to lower latitudes, and favoring
710 stationary waves. Consequently, when we increase the dust opacity in simula-
711 tions x_3 , x_4 and x_{ref} , source sublimation rises and the meridional component
712 of the 5.6 km wind, which is a good indicator of stationary wave activity,
713 strengthens. Atmospheric water vapor and ice columns at 45°N, which reflect
714 precipitation activity, reveal the same tendency. Interestingly, once dust opac-
715 ity has reached 1.5, few changes are observed in *winter* accumulation rates
716 compared to maximum dust conditions. However, large changes in *annual*
717 accumulation rates are observed between Fig. 9.e ($\tau_{dust} = 1.5$) and Fig. 9.a
718 ($\tau_{dust} = 2.5$) because the higher dust opacity cools the daytime lower atmo-
719 sphere during summer, and favors ice preservation. But we have to remember
720 that clouds are radiatively inactive in these experiments, and might play the
721 same role by cooling daytime surface temperatures (Wilson et al., 2007). Work
722 is underway to assess this effect.

723 5.2.2 *Obliquity range: effect on ice preservation*

724 If we maintain optimal conditions and only change the obliquity, glaciation
725 occurs for an obliquity between 15° and 35°, as seen in Fig. 9.a,b,c. As obliquity
726 is decreased, a larger fraction of the insolation goes to the equatorial region,
727 and sublimation of the equatorial reservoirs is enhanced, as clearly seen in
728 Table 1. Finally, stationary waves and precipitation remain active, even though
729 a decrease of the water-ice column is observed at 15° obliquity, due to much

730 higher atmospheric temperatures in northern mid-latitudes during winter.

731 However, 45° obliquity leads to a sharp change in the winter circulation pattern
732 and to a large increase of summer surface temperatures. At this obliquity, the
733 largest fraction of insolation falls in the mid-latitudes during summer, and
734 prevents any ice preservation, as seen in Fig. 9.f. Attenuation of stationary
735 planetary waves and baroclinic activity also reduces winter precipitation, as
736 revealed by a slight decrease in the wind speed, water vapor and ice columns
737 in Table 1.

738 The 15° obliquity simulation (x_1) may be unrealistic, because the atmospheric
739 dust opacity during periods of low obliquity is not expected to be so high
740 ($\tau_{dust} = 2.5$), even though dust lifting is favored by the baroclinic activity
741 (Haberle et al., 2003). However, knowing that under current orbital condi-
742 tions, a dust opacity of 2.5 is possible at regional scales during perihelion
743 dusty season, simulation x_2 might be a reasonable scenario. Indeed, we as-
744 sume that Mars returns from a high obliquity excursion at $\sim 45^\circ$, during
745 which tropical mountain glaciers have been formed, and approaches orbital
746 conditions similar to current ones. In this case, regional dust storms, coupled
747 with increased moisture levels due to the equatorial water sources, might have
748 been sufficient to bring precipitation in the northern mid-latitudes. Finally,
749 x_{ref} conditions were probably common in the past. Therefore, we suggest that
750 geological periods favorable for glacial activity were probably intermediate
751 ($25\text{-}35^\circ$) obliquity periods.

752 *5.2.3 Changing the perihelion: an interglacial period*

753 The mid-latitude glaciation is also very sensitive to a change of eccentricity,
754 and a decreasing eccentricity results in a retreat, or suppression of the de-
755 posits (see Fig. 9.g). The next to last simulation of Table 1 shows the effect
756 of a null eccentricity on the 35° obliquity simulation. Precipitation is still ac-
757 tive, despite the weakening of the atmospheric circulation, but warmer surface
758 temperatures result in the sublimation of most of the mid-latitude ice during
759 summer. However, some deposits remain stable near Nilosyrtris Mensae and
760 Phlegra Montes. All in all, high eccentricity (~ 0.1) seems necessary for a
761 large extension of the mid-latitude glaciation, and periods of low eccentricity
762 might be interglacial periods, where a retreat of the glaciers is likely to occur.

763 The argument of perihelion is also essential, and a reversed argument $L_p = 90^\circ$
764 (see Fig. 9.h) also results in a retreat of the northern mid-latitude deposits.
765 However, as noticed by Mischna et al. (2003), ice then tends to be stable in
766 aphelion summer hemisphere. Figure 9.h thus reveals many interesting depo-
767 sition patterns in the southern mid-latitudes, further represented in Fig. 10.
768 Net accumulation of ice predicted by the LMD/GCM is compared to the ice-
769 related landforms mapped by Squyres (1979). Two deposition regions of the
770 model are located south of Argyre Planitia (arrow 1.) and around western Hel-
771 las (arrow 2.), and are in good agreement with the observations. Interestingly,
772 no accumulation is predicted in the glacial region of eastern Hellas, which is
773 discussed in Forget et al. (2006), and results from other orbital conditions (see
774 section 2.3). A large deposition belt is also present at 50°S, between 120°E and
775 120°W, in regions where no glacial landforms have been mapped by Squyres
776 (1979). A further analysis is beyond the scope of this paper, even though this
777 high precipitation belt might be geologically relevant.

779 This broad deposition pattern of the southern hemisphere gives support to
780 the idea that mid-latitude glaciation occurs in both hemispheres under dusty
781 and intermediate obliquity conditions ($\sim 35^\circ$), assuming an equatorial water
782 source. Glaciation is then favored in the northern or southern hemisphere,
783 depending on the solar longitude of perihelion.

784 *5.3 Impact of thermal-inertia feedback*

785 Thermal inertia is a composite quantity defined as $I = \sqrt{\lambda C}$ in $\text{J s}^{-1/2} \text{m}^{-2} \text{K}^{-1}$,
786 where C is the volumetric heat capacity, and λ the thermal conductivity, the
787 latter undergoing the largest variations among different geological materials.
788 In all the simulations presented above, thermal inertia was based on the IRTM
789 and TES observations (Forget et al., 1999), and we assumed that the surface
790 thermal inertia was not affected by the presence of ice. This assumption holds
791 for a seasonal micrometer-sized frost that is too thin to have an impact on
792 surface and subsurface temperatures. Consequently, under current climatic
793 conditions, this assumption is always valid. However, it does not hold for cm-
794 sized ice layers stable at depth or draped over the regolith.

795 To assess this effect, we have taken advantage of a new conduction model that
796 solves the unsteady heat diffusion equation on 18 fixed vertical grid points
797 of varying thermal inertia. As a first approach, we use the 1-D version of
798 the LMD/GCM to assess the impact on surface temperatures generated by
799 an ice layer of different thicknesses (x-axis of Fig. 11) deposited on top of a
800 regolith of varying thermal inertia (y-axis). The model is run at 45°N with

801 the x_{ref} climate parameters (see Table 1), assuming a volumetric heat capac-
802 ity C of $10^6 \text{ J m}^{-3} \text{ K}^{-1}$, an ice albedo of 0.4, and an ice thermal inertia of
803 $1000 \text{ J s}^{-1/2} \text{ m}^{-2} \text{ K}^{-1}$, which is an intermediate value between the 600 and
804 $2000 \text{ J s}^{-1/2} \text{ m}^{-2} \text{ K}^{-1}$ water-ice deposits observed in the north polar region
805 by Paige et al. (1994).

806 [Fig. 11 about here.]

807 Figure 11 represents the obtained annual mean and maximum surface tem-
808 peratures. Increasing thermal inertia decreases the diurnal temperature range,
809 resulting in a drop of maximum temperatures in Fig. 11 when the ice thick-
810 ness exceeds 1 cm. Such an abrupt collapse of annual maximum temperatures
811 would prevent the ice from sublimating during summer, and would give rise
812 to a positive feedback. However, since daily temperature changes are reduced,
813 less infrared radiation is lost during the day and on annual mean basis, the
814 surface is warmed up by several degrees, as seen in Fig. 11 and already noticed
815 by Paige (1992). Does this feedback accelerate or decelerate the growth of the
816 ice sheet?

817 To answer this question, we allow the thermal inertia of the subsurface to vary
818 in the GCM as a function of ice thickness, and rerun the reference simulation
819 (x_{ref} in Table 1). As observed in previous sections, ice is deposited in the
820 northern mid-latitudes, forming a layer of high thermal inertia that is more
821 than 1 cm thick after 3 years of simulation. From then on, summer peak tem-
822 peratures begin to cool, thereby decreasing sublimation until the ice thickness
823 reaches ~ 5 cm after 10 years of simulation, at which point summer sublima-
824 tion almost disappears and peak temperatures stabilize at ~ 230 K (compared
825 to ~ 250 K without the ice) in Deuteronilus Mensae. Consequently, thermal

826 inertia feedback does accelerate the growth of the ice layer, but only up to a
827 certain point where a maximum accumulation rate is reached. In the glacial
828 regions, the annual accumulation rate is in this case only about 20% higher
829 than for the reference simulation (see Fig. 12.a).

830 However, extending the thermal inertia feedback to the equatorial water-ice
831 sources has a significant impact in our simulations on the accumulation rates.
832 Indeed, peak temperatures of the sources are decreased by their high thermal
833 inertia, and annual sublimation plummets from dozens of cm (see Table 1) to
834 ~ 5 mm, providing less water to the climate system. Consequently, the atmo-
835 spheric water vapor amount goes down from ~ 300 pr. μm to ~ 100 pr. μm
836 on average, bringing less precipitation to the mid-latitudes and a winter snow
837 cover of only ~ 3 mm thickness, that almost completely sublimates away dur-
838 ing summer (see Fig. 12.b). If we now suppose that albedo of the tropical
839 mountain glaciers is lower than the previously used value of 0.4 due to the
840 debris cover (Marchant and Head, 2007), and equal to 0.2, atmospheric water
841 vapor is restored, and a seasonally persistent snow cover is again predicted,
842 especially in glacial regions where accumulation can reach ~ 10 mm yr $^{-1}$ (see
843 Fig. 12.c).

844 These results show that thermal inertia feedback has a large impact on surface
845 temperatures and ice sublimation, and may be able to prevent sublimation
846 during interglacial periods, thereby sequestering the ice in the mid-latitudes
847 over a significant amount of time. They also emphasize the large dependence
848 of the water cycle on physical processes that occur in the multiple sources.
849 For example, refined understanding of the TMG sublimation process, as seen
850 above, will be crucial to improve the modeling of martian ice ages (Kowalewski
851 et al., 2006).

853 **6 Discussion**

854 The good agreement of our simulations with the geological observations, the
855 orbital constraints given by the sensitivity studies of section 5.2, and the ap-
856 proximate ages of the glacial landforms allow us to propose an updated sce-
857 nario for the Martian Ice Ages. We suppose for clarity that we start from
858 a past Martian climate similar to the one observed today that would occur
859 during a high mean obliquity period (25-45°).

860 A first increase in obliquity results in the sublimation of the north polar cap,
861 and formation of tropical mountain glaciers with an accumulation rate up to
862 $\sim 60 \text{ mm yr}^{-1}$ under 45° (or more) obliquity (Levrard et al., 2004; Forget
863 et al., 2006). As we neglect the effect of a dust lag on the sublimation of the
864 sources (Mischna and Richardson, 2005), these accumulation rates can be seen
865 as optimal. This also applies to the mid-latitude glaciation predicted here.

866 After this excursion, obliquity falls to around 35°. Given the high dust content
867 of the atmosphere under these conditions (Newman et al., 2005) and the pres-
868 ence of tropical mountain glaciers, an increased water cycle results in large
869 mid-latitude precipitation controlled by strong stationary planetary waves.
870 Wavenumber 3 favors the accumulation of snow in glacial regions, including
871 the Deuteronilus-Protonilus Mensae fretted terrains (Head et al., 2006a,b),
872 and results in the formation of a regional ice sheet. The climate system can
873 sustain mid-latitude glaciation by staying in the 25-35° obliquity range given
874 in section 5.2, until a new high obliquity ($\sim 45^\circ$) excursion occurs. The cor-

875 responding period of time of approximately 50 ky creates, if we suppose ac-
876 cumulation rates around 10-20 mm yr⁻¹ (see Table 1), an ice sheet of up to
877 500-1000 m thickness. At the end of this period, if we consider that a lower
878 dust opacity is likely to occur when the obliquity is only 25°, some simulations
879 suggest that the precipitation is still active, but with decreased accumulation
880 of $\sim 1\text{-}2$ mm yr⁻¹ and increased sublimation during the summer. This lower
881 regime of glaciation might explain the formation of superposed lobate debris-
882 covered glaciers described by Levy et al. (2007) and Dickson et al. (2008). The
883 mid-latitude glaciation, based on the age of the lobate debris aprons, might
884 have lasted for several hundreds of Myr, and some deposits may be as young
885 as ~ 10 Myr (Mangold, 2003). The remnants we see today thus represent the
886 signature of a recent glacial event, whose ice has been preserved under a sub-
887 limation till, as confirmed by radar sounding (Plaut et al., 2009). Age of the
888 tropical mountain glaciers is estimated to be 10-200 Myr (see section 2), con-
889 firming that the mid-latitude and tropical glacial activities occurred during
890 the same geological period.

891 The 25-35° obliquity period is also favorable for the formation of a mantle
892 that is draped over the high latitudes, and visible in Fig. 9.a,b,c,e. These
893 deposits may have contributed to the formation of the latitude dependent
894 mantle (see section 2) under higher obliquity conditions than those observed
895 by Levrard et al. (2004). However, increasing number of observations point
896 to an extremely young age of the latitude dependent mantle (~ 0.1 Myr, see
897 Head et al. (2003); Kostama et al. (2006)), that might have formed during
898 the current low mean obliquity period, at epochs where the equatorial water
899 sources were likely depleted. Consequently, its climatic origin is not yet clearly
900 established.

901 As we start a new high obliquity excursion ($\sim 45^\circ$), the mid-latitude deposits,
902 as well as the high-latitude mantle, might sublimate away and feed the tropical
903 mountain glaciers again. In that case, the water cycle would be closed, and
904 a high mean obliquity period would be characterized by the oscillations of
905 the ice reservoirs between the tropical mountain glaciers and the mid-to-high
906 latitude deposits, producing consistent stratigraphy in both deposits.

907 It is worth noting that the mid-latitude glacial activity might also be sustained
908 by the south polar cap, or any other source able to provide enough water
909 vapor to the winter atmosphere. At the moment, from a modeling point of
910 view, an equatorial water source is necessary, but improvements of the GCM
911 may change this result.

912 **7 Conclusions**

913 A climatic origin for the Amazonian northern mid-latitude glaciation is ex-
914 plored using the LMD Global Climate Model, and the main findings can be
915 summarized as follows:

916 (1) Formation of an ice sheet is predicted by the LMD/GCM in the northern
917 mid-latitudes during periods of moderate obliquity by 1) sublimation of
918 an equatorial water source, 2) increased dust content and water vapor
919 holding capacity of the atmosphere, 3) resulting migration of the satu-
920 ration zone from high to mid-latitudes, 4) poleward flux of water vapor
921 controlled by strong stationary planetary waves and transient weather
922 systems, 5) condensation and precipitation of water ice to form a thick
923 cloud belt in the northern mid-latitudes, 6) deposition of ice in glacial

924 regions, 7) differential sublimation and erosion of the deposits during
925 northern summer and preservation in glacial regions;

926 (2) Longitudinal distribution of the deposits in the 30-50°N band results
927 from the wavenumber-3 structure of the stationary planetary waves and
928 topographic forcing induced by the Tharsis, Arabia, and Elysium barriers;

929 (3) Based on climate sensitivity experiments, glaciation is found to occur dur-
930 ing moderate obliquity periods, and requires an equatorial water source
931 (here assumed to be the tropical mountain glaciers), an obliquity of 25-
932 35°, a high eccentricity (~ 0.1), an aphelion season corresponding to
933 northern summer ($L_p = 270^\circ$), and a high dust opacity ($\tau_{dust} > 1.5$).
934 Such conditions probably occurred many times during the estimated pe-
935 riod of glaciations (~ 10 -300 Myr, Mangold (2003));

936 (4) Predicted accumulation rates of ~ 10 -20 mm yr⁻¹ are consistent with
937 the formation of regional ice sheets, and have large implications for the
938 recent history of the water cycle. Glacial activity in the mid-latitudes
939 involves a significant amount of water and should thus appear as a major
940 component of the global stratigraphy. GCM simulations point to the im-
941 portance of dust for controlling the water vapor holding capacity of the
942 atmosphere, and suggest a complex coupling between the dust content of
943 the atmosphere and the glacial activity of the planet, as also observed on
944 Earth (Harrison et al., 2001; Lambert et al., 2008).

945 (5) Thermal inertia feedback caused by surface water-ice results in a progres-
946 sive lowering of summer peak temperatures in the glacial regions, and can
947 lead to disappearance of the summer sublimation phase. This result of
948 the climate model suggests that ice might be preserved and sequestered
949 over long periods of time by the impact of ice thermal inertia on surface
950 temperatures.

[Fig. 13 about here.]

Figure 13 illustrates the different components of the climate system controlling the northern mid-latitude glaciation. Dashed arrows represent major feedbacks that still need to be constrained and implemented into the LMD/GCM. These include:

- Radiative effect of clouds, today but also under past conditions where clouds reach significant optical depth and spatial coverage;
- Radiative effect of water vapor, that is neglected on present-day Mars, but may influence our results under past conditions, given the high water vapor holding capacity of the atmosphere;
- Dust lifting and coupling with the cloud microphysics, as well as scavenging of dust by water-ice particles;
- Coalescence of ice crystals induced by high precipitation events;
- Physics of the ice deposits: latent heat exchange induced by sublimation or melting of the deposits, heating within the ice layer by absorption of solar radiation (Clow, 1987), and the protective effect of a dust lag (Mischna and Richardson, 2005).

The detailed analysis of the mid-latitude geomorphology, in light of terrestrial analogs and new data from the Mars Reconnaissance Orbiter mission, is continuing to provide major insights into the microclimate and water budget of the glacial systems. A comparison of these detailed observations with GCM and mesoscale predictions (Spiga and Forget, 2009) will certainly improve our knowledge of the glacial climate.

We are just beginning to understand the main components of this complex climate signal. Hundreds of millions of years of glacial-interglacial climate is

976 recorded in the extremely well-preserved martian stratigraphy. Mars is the
977 only opportunity at hand to explore a climate system similar to the terres-
978 trial glacial ages, and to test our understanding of the fundamental feedbacks
979 controlling climate changes on Earth.

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1281 **List of Tables**

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1288		finally the annual accumulation rate in three regions indicated	
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#	Climate parameters				Source (cm/yr)	Wind speed (m/s)	Water/Ice column		Deposits (mm/yr)		
	ϵ	e	L_p	τ_{dust}			at 45°N (pr. μm)		1.	2.	4.
x_{ref}	35°	0.1	270°	2.5	37	8.8	186	512	11.3	13.5	12.1
x_1	15°	0.1	270°	2.5	92	9.8	165	248	6.7	10.6	5.3
x_2	25.2°	0.09	251°	2.5	51	10.8	212	503	6.3	12.5	8.9
x_3	35°	0.1	270°	0.2	3.4	1.4	3.4	6.3	0.	0.	0.
x_4	35°	0.1	270°	1.5	25	6.3	118	260	14.1	0.	11.1
x_5	45°	0.1	270°	2.5	26	4.4	154	491	0.	0.	0.
x_6	35°	0.		2.5	28	4.5	77	303	0.	0.	6.2
x_7	35°	0.1	90°	2.5	43	3.6	32	157	0.	0.	0.

Table 1

Results of the sensitivity tests. Climate parameters are listed on the left, and results are summarized by giving the annual sublimation rate of the equatorial sources, the meridional wind speed at 5.6 km over Deuteronilus Mensae, along with the zonal mean water vapor and water ice columns at 45°N (in pr. μm) both averaged over the $L_S = 240\text{-}270^\circ$ period, and finally the annual accumulation rate in three regions indicated in Fig. 7.

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1367 13 Major components of the climate system related to the
1368 northern mid-latitude glaciation. Orbital parameters represent
1369 the only external forcing, whereas dust content of the
1370 atmosphere and changing surface properties are sources
1371 of internal oscillations. These three factors control lower
1372 atmosphere temperature, the meridional temperature gradient,
1373 and surface temperature. Lower atmosphere temperature
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1375 temperature gradient controls the cloudiness of the northern
1376 mid-latitudes, that is dependent on the eddy heat flux
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1378 temperature gradient also changes the Hadley cell and dust
1379 lifting activity, the latter providing condensation nuclei to the
1380 water cycle. This all leads to precipitation in the storm-track
1381 region and accumulation of ice, whose preservation is finally
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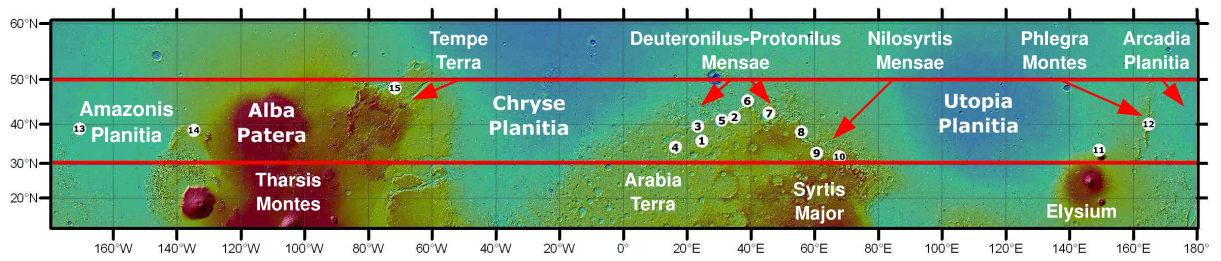


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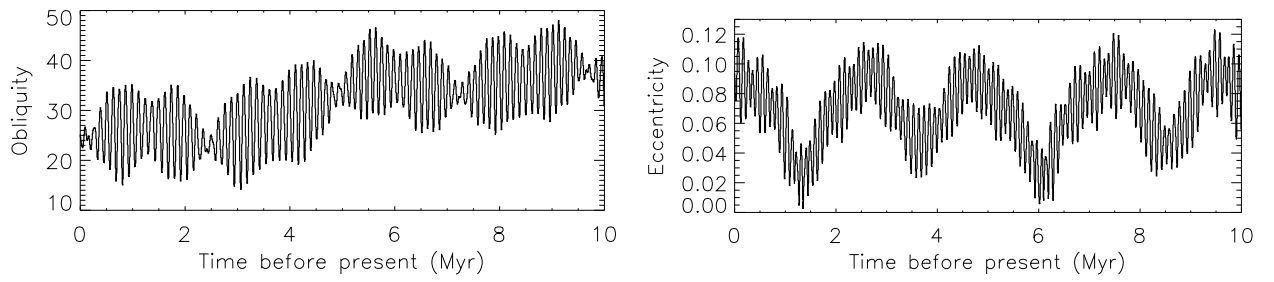


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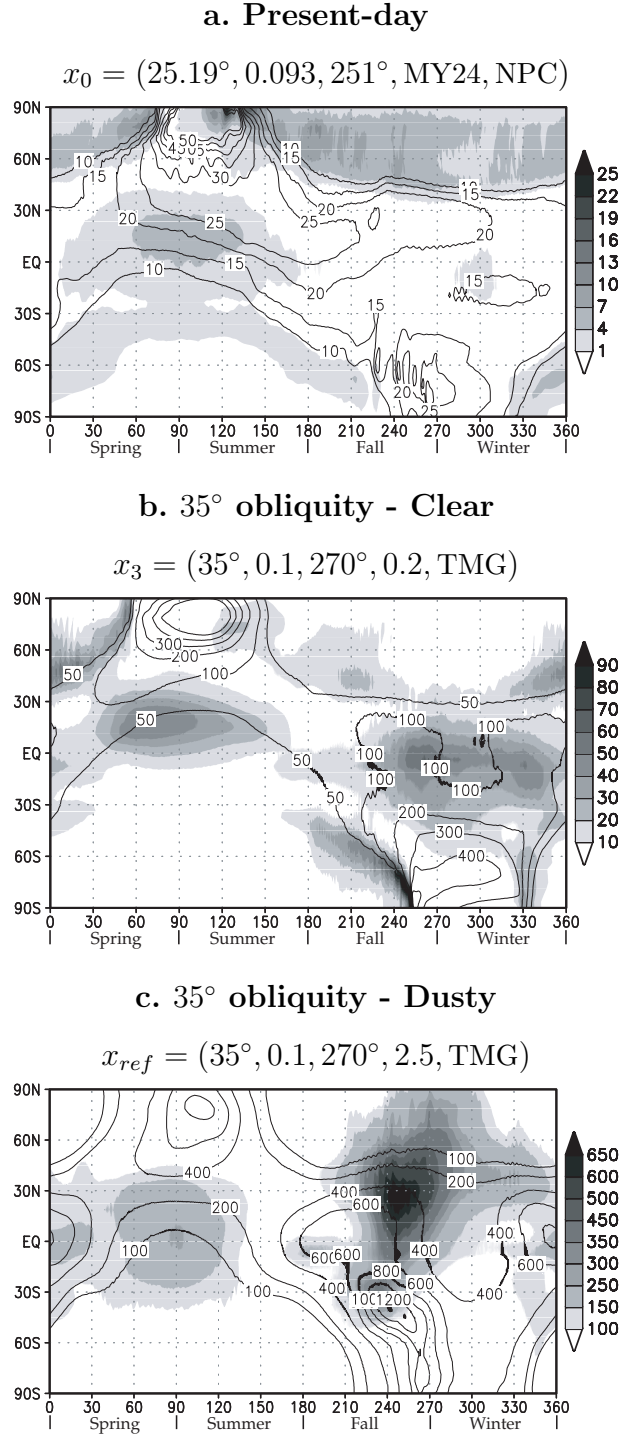


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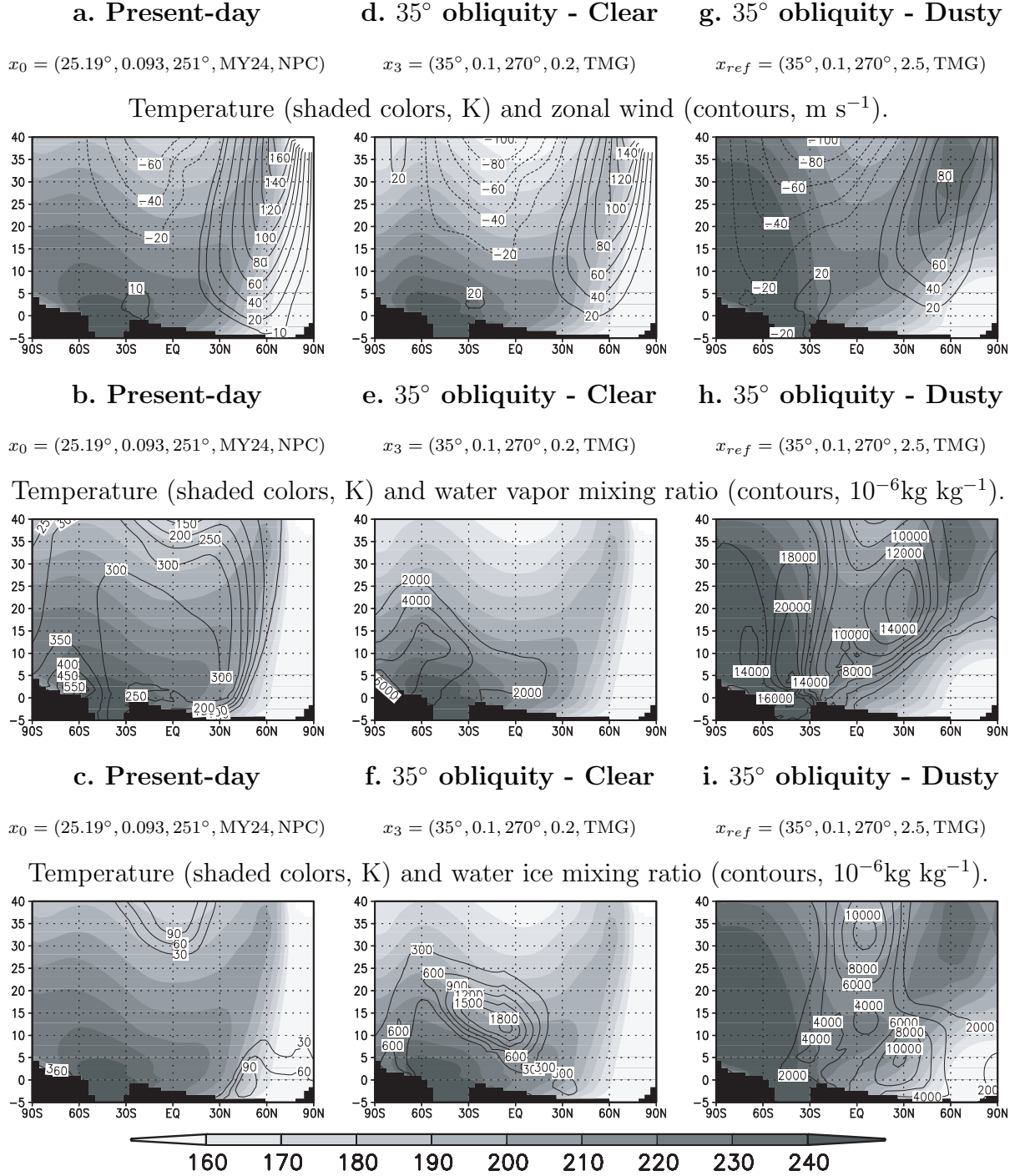


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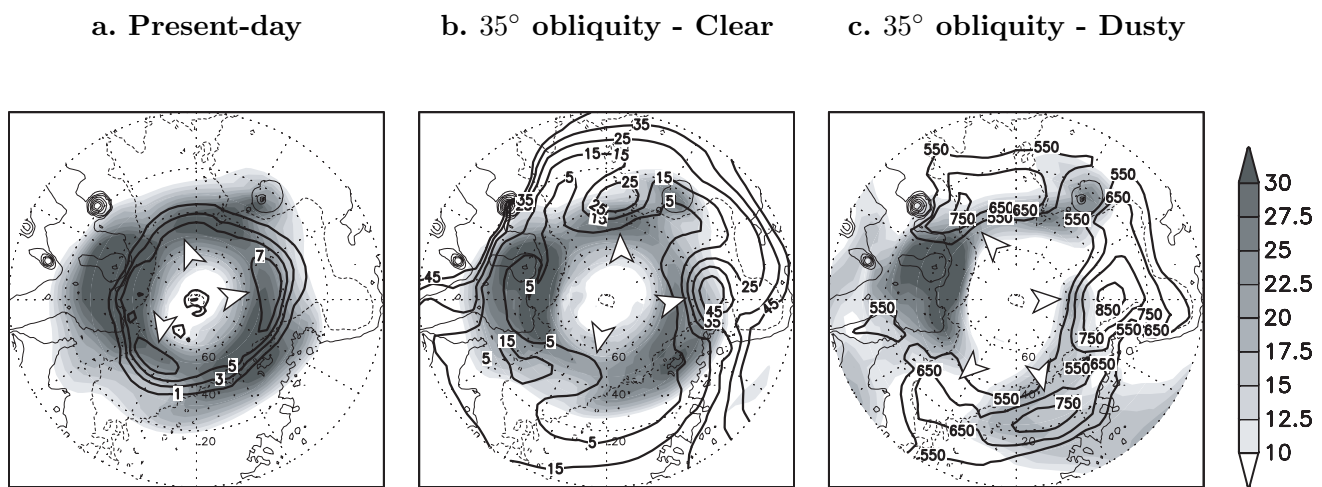


Fig. 5. Polar stereographic map of the Northern Hemisphere. Zonal winds at the 5.6-km level are depicted by shaded colors (m s^{-1}), and water-ice column by contours (pr. μm). The fields are averaged over the $L_S = 240\text{-}270^\circ$ period. Deviations from zonal symmetry are apparent in the jetstream structure, over the Tharsis, Arabia Terra and Elysium ridges. Centers of enhanced cloud formation are indicated by arrows.

Reference simulation (35° obliquity and dusty conditions) - $L_s = 240-270^\circ$

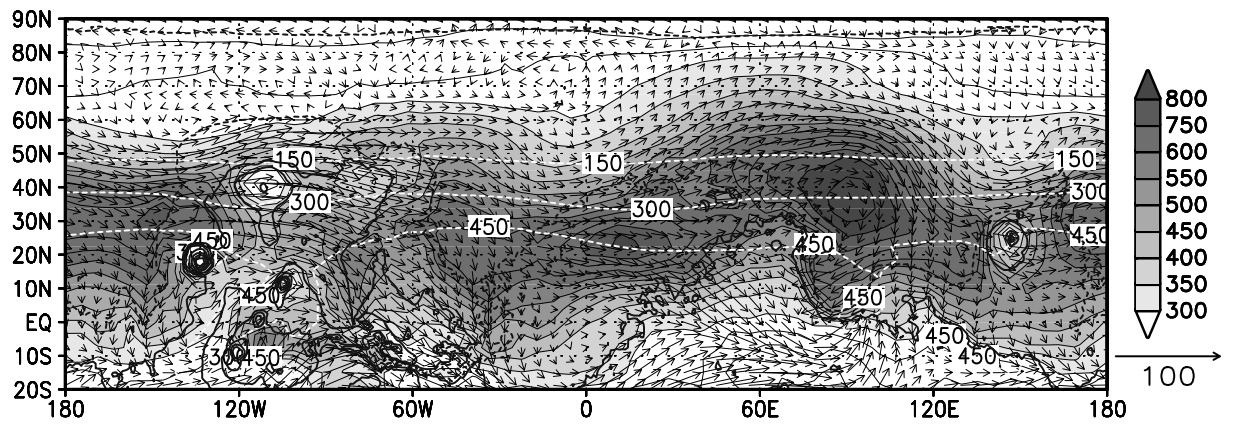


Fig. 6. Average cloud ice content (shaded regions, pr. μm) and horizontal wind field at the 5.6-km level (m s^{-1}). White lines indicate water vapor column (pr. μm).

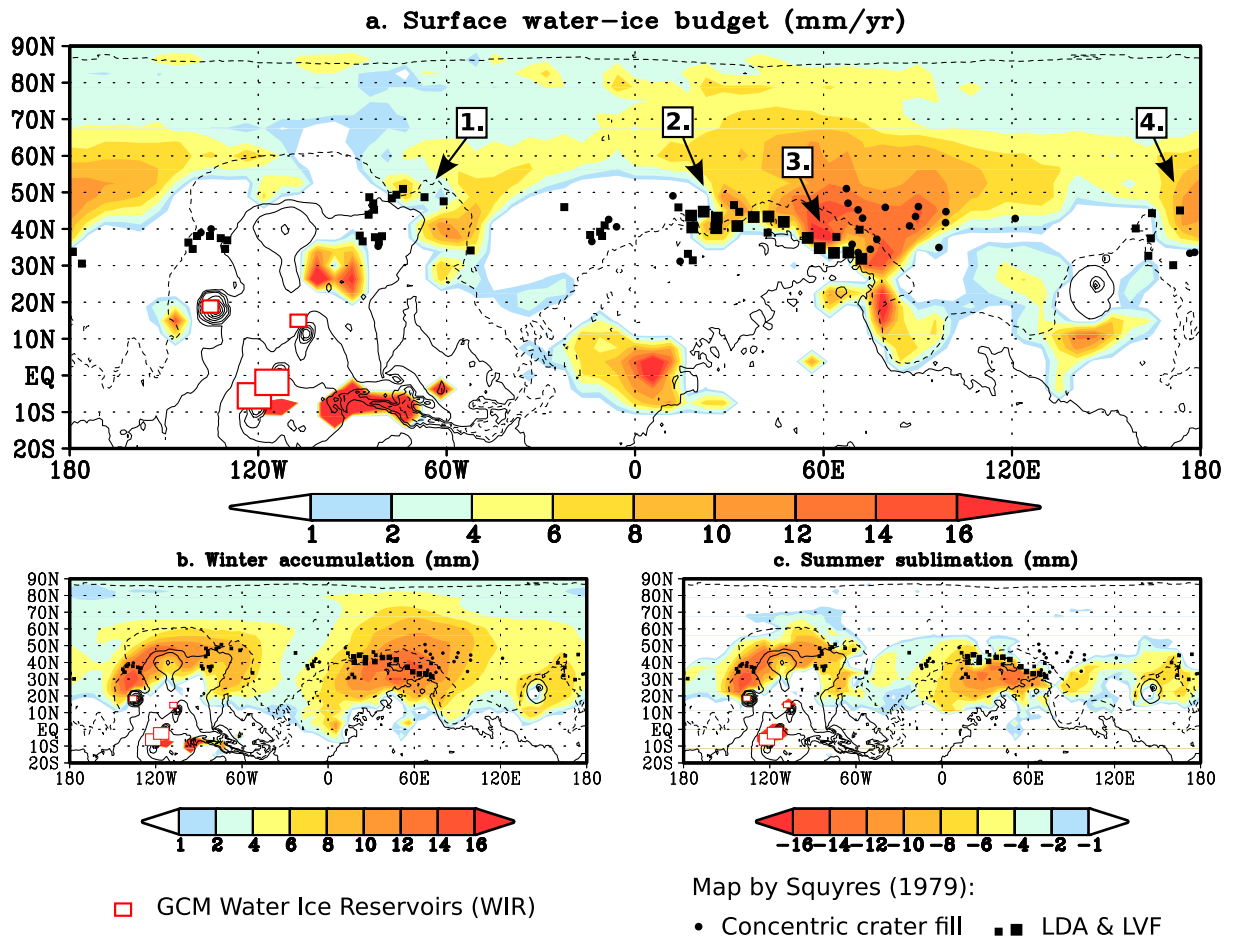


Fig. 7. **a.** Net ice accumulation (mm yr^{-1}) predicted in simulation $x_{ref} = (35^\circ, 0.1, 270^\circ, 2.5, \text{TMG})$, superposed on the map by Squyres (1979), which shows the specific location of several different types of ice-related features. LDA and LVF stand for Lobate Debris Aprons and Lineated Valley Fill. See also Fig. 1, which shows the areas of widespread glaciation documented in Head and Marchant (2006). Indicated regions: 1. Tempe Terra, 2. Deuteronilus Mensae, 3. Nilosyrtris Mensae, 4. Phlegra Montes. **b.** Water ice accumulation during the $L_S = 180\text{--}360^\circ$ period (mm). **c.** Water ice sublimation during the $L_S = 0\text{--}180^\circ$ period (mm).

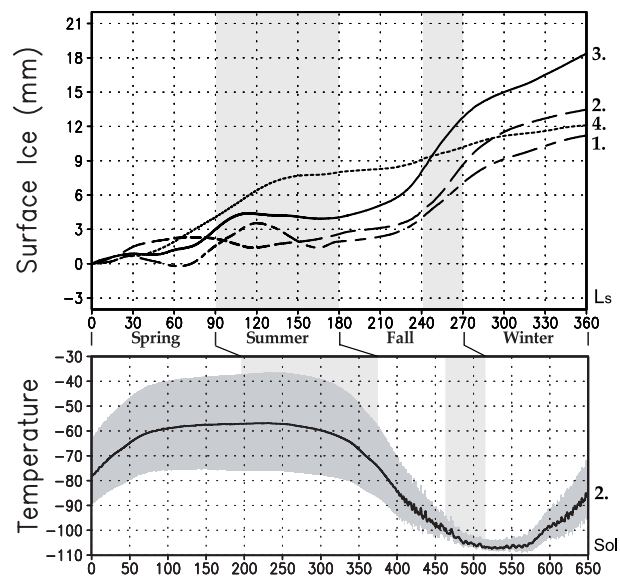


Fig. 8. **Upper panel:** Evolution of surface ice deposits in glacial regions (see the legend of Fig. 7). **Lower panel:** Maximum, minimum (shaded region) and mean (middle line) daily surface temperatures in Deuteronilus Mensae ($^{\circ}\text{C}$).

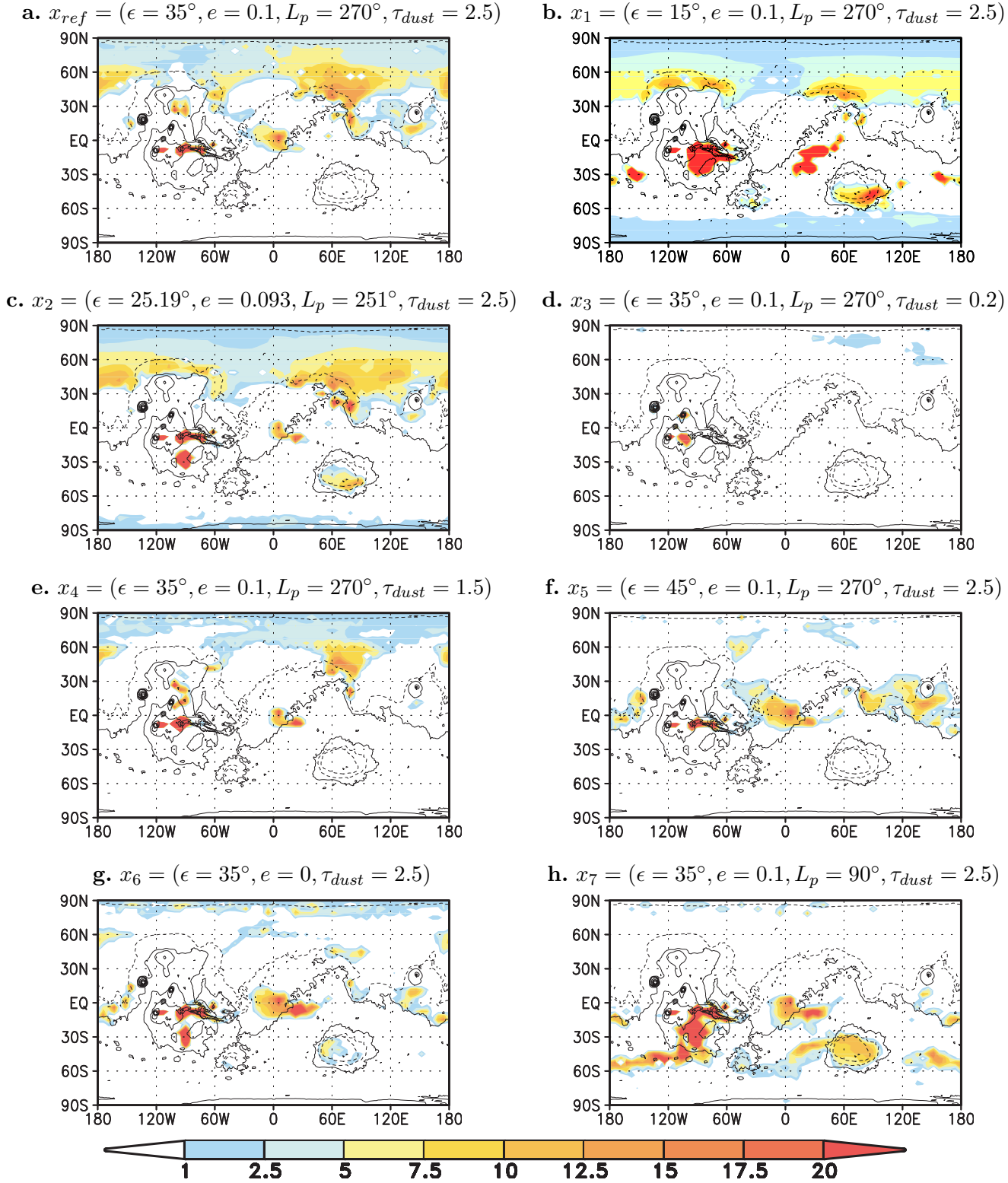


Fig. 9. Net gain of surface ice over a year (mm yr^{-1}) for each sensitivity experiment. Details can be found in Table 1.

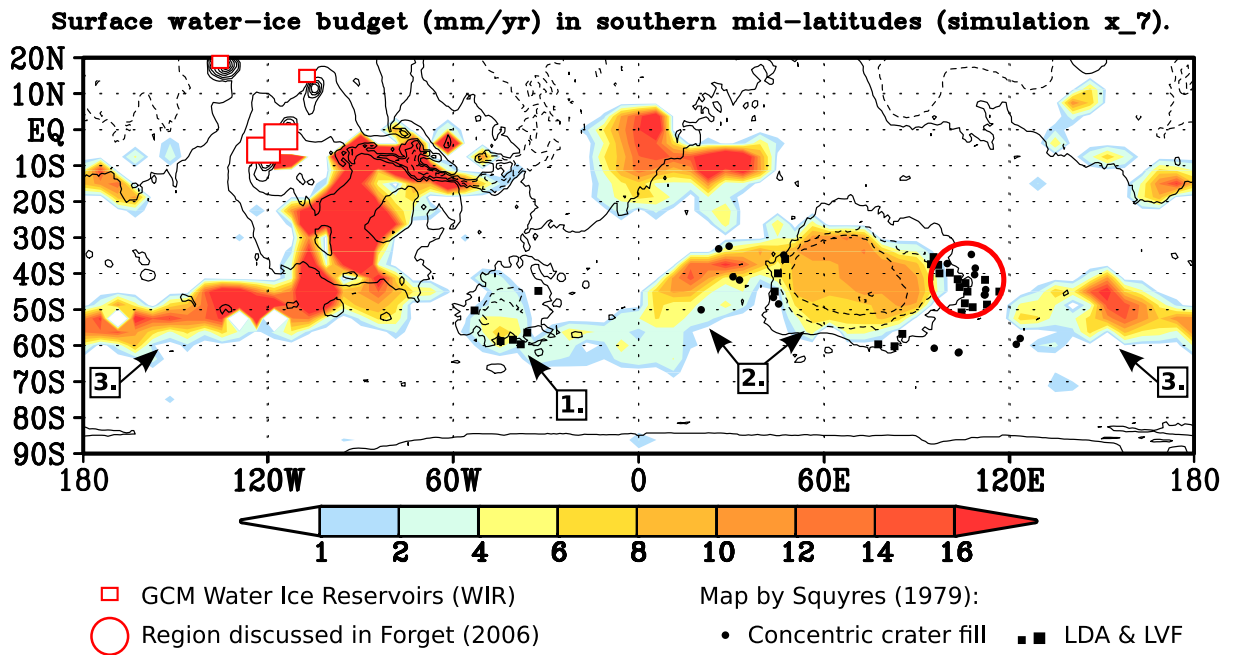


Fig. 10. Net ice accumulation (mm yr^{-1}) predicted in simulation $x_7 = (35^\circ, 0.1, 90^\circ, 2.5, \text{TMG})$, superposed on the map by Squyres (1979). LDA and LVF stand for Lobate Debris Aprons and Lineated Valley Fill. Indicated regions: 1. Argyre Planitia, 2. Western Hellas, 3. Terra Sirenum (left) and Terra Cimmeria (right).

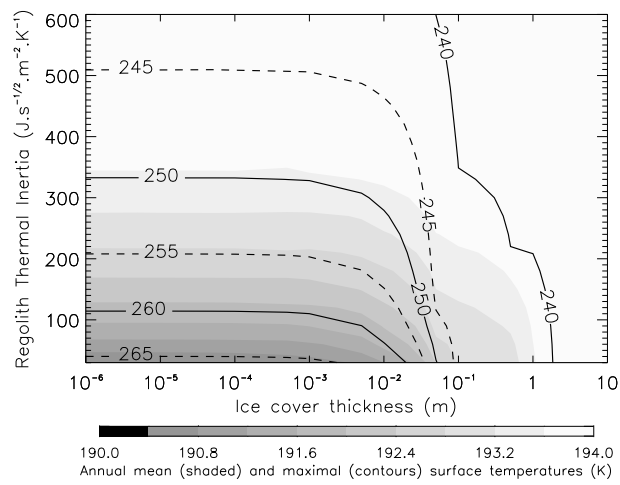
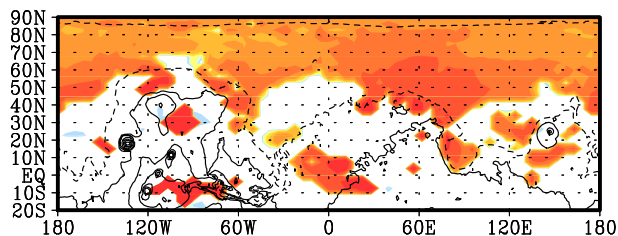
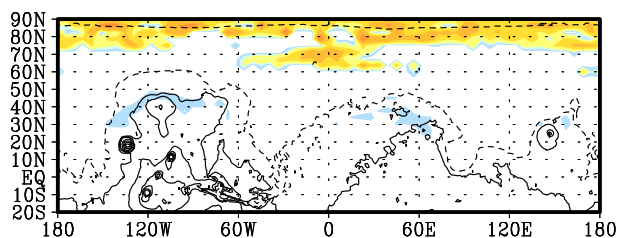


Fig. 11. Results of the 1-D model showing the impact of ice thermal inertia on surface temperatures at 45°N and under reference conditions (see Table 1). Contours and dark shades indicate annual maximum and mean temperatures, respectively. Ice thermal inertia and albedo are set to $1000 \text{ J s}^{-1/2} \text{ m}^{-2} \text{ K}^{-1}$ and 0.4, respectively.

a. TI feedback is active on deposited ice only



b. TI feedback is active also on the sources



c. Same as b., but source albedo is set to 0.2

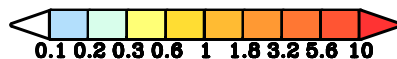
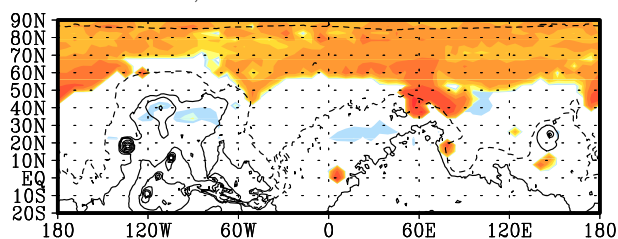


Fig. 12. Sensitivity of net ice accumulation rates (mm yr^{-1}) to different surface properties: **a.** Thermal inertia of the deposited ice is set to $\text{TI} = 1000 \text{ J s}^{-1/2} \text{ m}^{-2} \text{ K}^{-1}$. **b.** Thermal inertia of the equatorial reservoirs and deposited ice layers is set to $\text{TI} = 1000 \text{ J s}^{-1/2} \text{ m}^{-2} \text{ K}^{-1}$. **c.** Same as **b.**, but albedo of the equatorial reservoirs is set to 0.2 instead of 0.4, assuming rock glaciers being darkened by the debris cover. Climate parameters are those of the reference simulation (see x_{ref} in Table 1). All the results are shown for the tenth year of simulation.

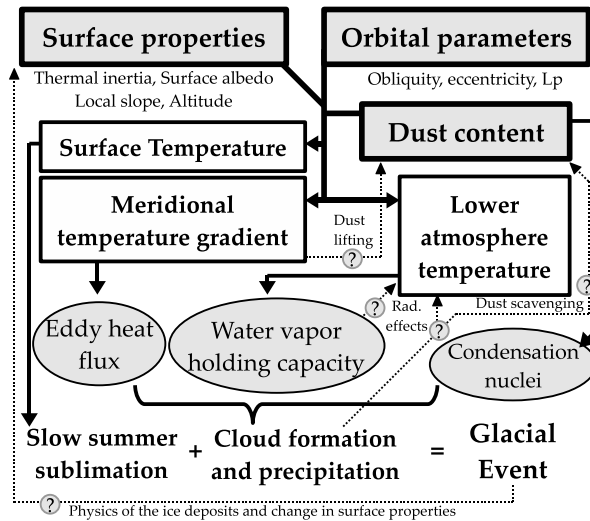


Fig. 13. Major components of the climate system related to the northern mid-latitude glaciation. Orbital parameters represent the only external forcing, whereas dust content of the atmosphere and changing surface properties are sources of internal oscillations. These three factors control lower atmosphere temperature, the meridional temperature gradient, and surface temperature. Lower atmosphere temperature acts on water vapor holding capacity, and the meridional temperature gradient controls the cloudiness of the northern mid-latitudes, that is dependent on the eddy heat flux induced by transient and stationary waves. The meridional temperature gradient also changes the Hadley cell and dust lifting activity, the latter providing condensation nuclei to the water cycle. This all leads to precipitation in the storm-track region and accumulation of ice, whose preservation is finally dictated by surface temperature. Dotted arrows with question marks indicate physical processes that are unresolved by the LMD/GCM, and also feed back on the original atmospheric response to external forcing.