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Aymeric Spiga, Isaac Smith. Katabatic jumps in the Martian northern polar regions. *Icarus*, inPress, 10.1016/j.icarus.2017.10.021 . hal-01643458

HAL Id: hal-01643458

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Katabatic jumps in the Martian northern polar regions

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October 2, 2017

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1 Abstract

2 *Martian polar regions host active regional wind circulations, such as the downslope katabatic winds*
3 *which develop owing to near-surface radiative cooling and sloped topography. Many observations*
4 *(stratigraphy from radar profiling, frost streaks, spectral analysis of ices) concur to show that aeolian*
5 *processes play a key role in glacial processes in Martian polar regions. A spectacular manifestation*
6 *of this resides in elongated clouds that forms within the polar spiral troughs, a series of geological*
7 *depressions in Mars’ polar caps. Here we report mesoscale atmospheric modeling in Martian*
8 *polar regions making use of five nested domains operating a model downscaling from horizontal*
9 *resolutions of twenty kilometers to 200 meters in a typical polar trough. We show that strong*
10 *katabatic jumps form at the bottom of polar troughs with an horizontal morphology and location*
11 *similar to trough clouds, large vertical velocity (up to +3 m/s) and temperature perturbations*
12 *(up to 20 K) propitious to cloud formation. This strongly suggests that trough clouds on Mars*
13 *are caused by katabatic jumps forming within polar troughs. This phenomena is analogous to*
14 *the terrestrial Loewe phenomena over Antarctica’s slopes and coastlines, resulting in a distinctive*
15 *“wall of snow” during katabatic events. Our mesoscale modeling results thereby suggest that trough*
16 *clouds might be present manifestations of the ice migration processes that yielded the internal cap*
17 *structure discovered by radar observations, as part of a “cyclic step” process. This has important*
18 *implications for the stability and possible migration over geological timescales of water ice surface*
19 *reservoirs – and, overall, for the evolution of Mars’ polar caps over geological timescales.*

20 1 Introduction

21 Studying the meteorology of the Martian polar regions is a means to address key questions related
22 to the Martian climate from the global to the local scales. Planetary-scale “flushing” storms
23 originate from the Martian polar regions and transport dust particles in lower latitudes, thereby
24 impacting the global climate of Mars (Cantor et al., 2002; Toigo et al., 2002). Modeling studies to

25 prepare the landing of the Phoenix polar lander (Kauhanen et al., 2008; Tyler et al., 2008; Michaels
26 and Rafkin, 2008) detailed how the atmospheric variability in the Martian northern polar regions is
27 controlled by a combination of the mean meridional circulation (Wilson, 1997; Forget et al., 1999)
28 responsible for the characteristic polar warming in Mars’ lower mesosphere (McCleese et al., 2008),
29 the circumpolar jet (Toigo et al., 2012; Mitchell et al., 2015; Guzewich et al., 2016), baroclinic waves
30 (Barnes et al., 1993; Collins et al., 1996), and regional circulations (Toigo et al., 2002; Tyler and
31 Barnes, 2005). The variability of regional (so-called mesoscale) circulations in the polar regions is
32 also a matter of active research to disentangle the combined influence of slope acceleration (Spiga
33 et al., 2011; Smith et al., 2013), “sea-breeze” circulations caused by the direct cap-edge thermal
34 contrasts (Toigo et al., 2002; Smith et al., 2015), and polar transient eddies (Tyler and Barnes,
35 2005). Understanding the interplay of global and mesoscale circulations in the Martian polar
36 regions is of primary importance to characterize the seasonal source / sink those regions represent
37 for Mars’ water cycle (Tyler and Barnes, 2014; Navarro et al., 2014).

38 Katabatic winds are a salient component of the atmospheric variability in Martian polar regions,
39 just as they are on the Earth (Parish and Waight, 1987; Gallée and Schayes, 1992; Bromwich
40 et al., 2001; Nylen et al., 2004). Katabatic winds are drainage atmospheric flows that form when
41 cooled dense air is accelerated down sloping terrains by gravity, overcoming the opposing along-
42 slope pressure gradient (Mahrt, 1982). The combination of sloping terrains, near-surface radiative
43 cooling, and surface ice cover (either CO₂ ice in winter or H₂O deposits apparent in spring),
44 makes the Martian polar regions particularly prone to the development of katabatic winds over an
45 extended period of time (Spiga, 2011; Smith et al., 2015). This is evidenced by numerous surface
46 morphologic features in the northern polar regions, from frost streaks to dune fields, thought to
47 be caused by the polar katabatic flow deflected by the Coriolis force (Massé et al., 2012). The
48 variability of spectral signatures (H₂O or CO₂) over the polar slopes during the seasonal retreat
49 have been plausibly ascribed to katabatic winds too (Appéré et al., 2011), as well as small-scale
50 sedimentation undulations over the northern polar plateau (Herny et al., 2014). The formation of

51 geological structures through time has also been possibly linked to the action of katabatic winds
52 (e.g., Abalos Mensae, Brothers et al., 2013). The impact of katabatic winds on polar geological
53 features is not only aeolian, but also thermal: especially above steep slopes, they induce a significant
54 downward sensible heat flux which acts either to warm the surface or to increase sublimation rates
55 (Spiga et al., 2011).

56 Radar stratigraphic observations of the Martian polar caps strongly suggest polar wind circu-
57 lations, especially katabatic winds, have been instrumental in shaping the northern polar cap, and
58 the related northern polar layered deposits, over geological timescales (Holt et al., 2010; Smith and
59 Holt, 2010; Smith and Holt, 2015). Radar measurements renewed the interest for the idea that
60 katabatic winds, deflected by the Coriolis force, may explain the spiral organization of the polar
61 troughs, the deep depressions (400 – 1000 m) incised in the northern polar cap of Mars (Howard
62 et al., 1982; Howard, 2000; Pathare and Paige, 2005). This idea has been pushed forward by Smith
63 et al. (2013) who combined radar measurements, visible imaging, and numerical modeling, to show
64 that 1. the layering below troughs evidenced by radar stratigraphy supports a migration of trough
65 in the upslope direction as part of a “cyclic step” process (Kostic et al., 2010); 2. the fluid current
66 associated with this “cyclic step” mechanism is the katabatic wind flow where the occurrence of
67 “hydraulic jumps” enables ice migration to occur from the upstream to the downstream part of
68 the trough; 3. ice migration process could be observed directly in the form of elongated “trough
69 clouds” (Figure 1) occurring at the bottom of the northern polar troughs in early summer. This
70 scenario has been extended to the southern polar cap of Mars (Smith et al., 2015), with the addi-
71 tional result that “sea-breeze” circulations caused by the receding seasonal CO₂ cap modulate the
72 intensity of katabatic winds and account for the poleward progression of clouds in southern polar
73 regions.

74 The arguments developed by Smith et al. (2013) on hydraulic jumps within Martian polar
75 katabatic flows (what is named hereafter “katabatic jumps”, see diagram in Figure 2) only relied on
76 simple analogy with the terrestrial elongated “wall-of-snow” that results from katabatic jumps over

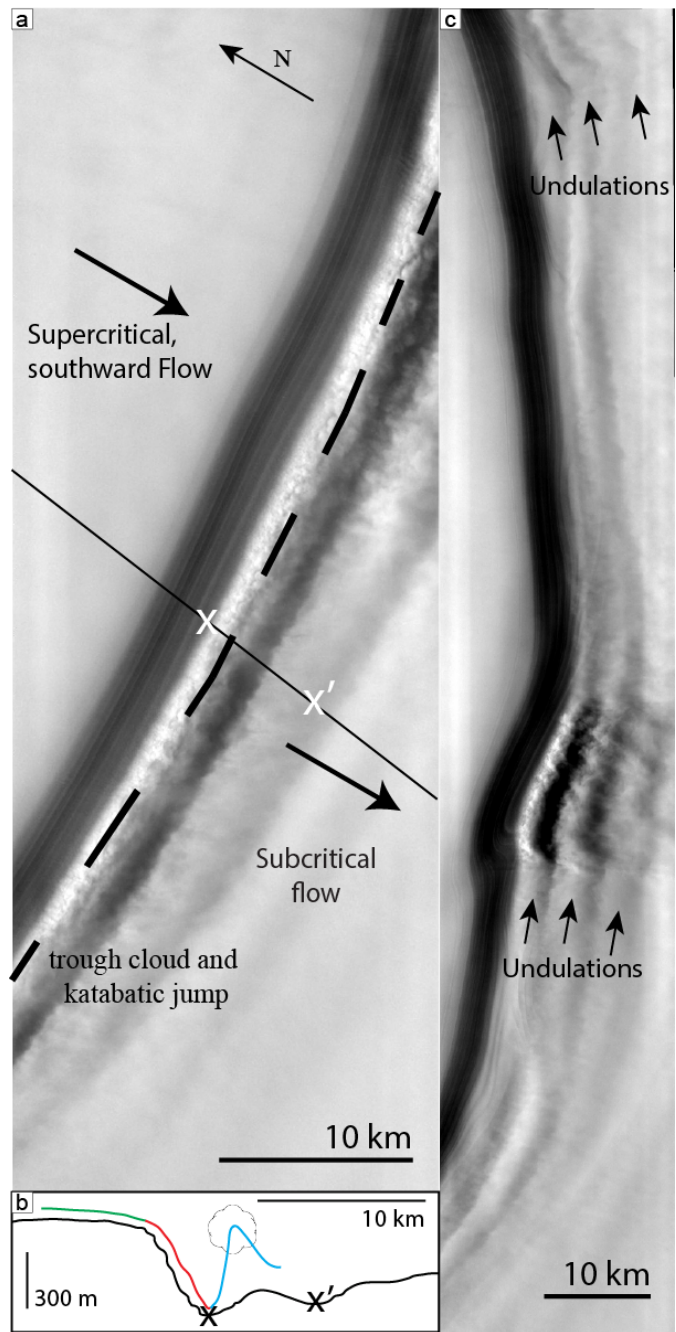


Figure 1: *Annotated Mars Odyssey THEMIS images of trough clouds (see Smith et al. (2013) for details on those images). Left image (reference V28743004) shows a distinctive elongated cloud with additional undulations downstream. This image has been obtained in the same trough as the one simulated in domain #5 (see Figure 7). Right image (reference V28744006) shows a similar phenomena, albeit with smaller extent, in a neighboring trough.*

77 e.g. Antarctica slopes and coastlines (Lied, 1964; Pettré and André, 1991). What is alternatively
 78 named the Loewe phenomenon has been reproduced in simulations of the katabatic flow in the
 79 vicinity of coastal slopes in the terrestrial polar caps (Gallée and Schayes, 1992; Pettré et al., 1993;
 80 Gallée et al., 1996; Yu and Cai, 2006). It remains to be demonstrated that katabatic jumps do
 81 occur in the environmental conditions of the Martian polar regions – and what the characteristics
 82 of these katabatic jumps are. It also remains to be elucidated how katabatic jumps cause the
 83 elongated clouds observed in Martian polar troughs.

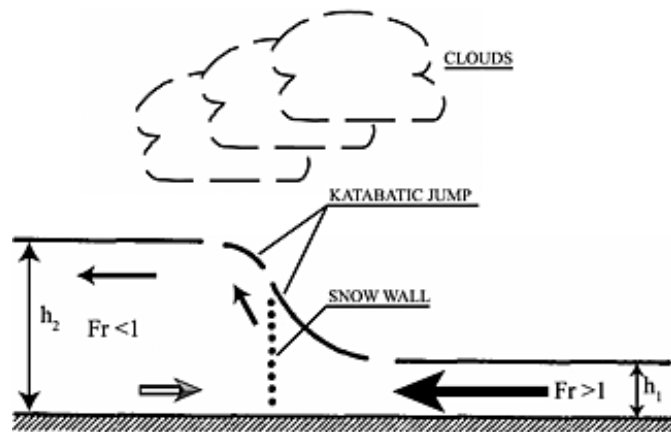


Figure 2: *Cartoon depiction of katabatic jump in Antarctica – reproduced with permission from Pettré and André (1991). Arrows indicate atmospheric flow directions. Right to left, the incoming flow, resulting e.g. from katabatic acceleration along a slope, is supercritical (“shooting” flow, according to the terminology in Ball, 1956). Flow depth increases at the katabatic jump. Ice forms at snowline (wall of snow), and clouds form at site of katabatic jump. Downstream of the katabatic jump, flow is subcritical (“tranquil” flow), with a return flow close to the surface in opposite direction than the incoming flow.*

84 In this paper, we propose to address those open questions with mesoscale numerical modeling
 85 of the Martian atmosphere. Mesoscale modeling aims at resolving the vast and diverse population
 86 of phenomena of smaller extent than a few hundreds of kilometers – in other words, the plethora of
 87 atmospheric phenomena left unresolved by Global Climate Models [GCM] (Wilson, 1997; Forget
 88 et al., 1999; Navarro et al., 2014). The technique of mesoscale modeling has been employed to
 89 obtain many of the aforementioned results on the Martian polar meteorology, especially on the

90 regional scale (Toigo et al., 2002; Tyler and Barnes, 2005; Kauhanen et al., 2008; Tyler et al.,
91 2008; Michaels and Rafkin, 2008; Spiga et al., 2011; Smith et al., 2013; Tyler and Barnes, 2014;
92 Smith et al., 2015). Mesoscale models are well-suited to get insights into atmospheric and sur-
93 face processes in polar regions. Contrary to GCMs, mesoscale models integrate the atmospheric
94 dynamics at high resolution in a specific region of interest on the planet with an adapted map
95 projection. Polar mesoscale domains are defined through stereographic projections, hence devoid
96 of the “pole singularity” present in many GCMs. In addition, high-resolution surface topograph-
97 ical and thermophysical properties (albedo, thermal inertia, CO₂ ground ice cover) are used in
98 mesoscale modeling.

99 Section 2 contains a technical description of the novel mesoscale simulations we carried out
100 to address the formation of katabatic jumps in Martian polar troughs. Section 3 comprises a de-
101 scription of the results we obtained with those simulations and a discussion of the properties of
102 the simulated katabatic jumps – including an exploration of the analogy with terrestrial phenom-
103 ena, and the possibility for cloud formation within those katabatic jumps. Section 4 contains a
104 summary of both our conclusions and the perspectives our work opens for future studies.

105 The companion paper Smith and Spiga (2017) addresses the variability of regional winds over
106 the northern polar cap; this paper addresses the behavior of the atmospheric flow at the local trough
107 level. The two companion papers complement one another and can be consulted independently.

108 **2 Model**

109 This study is based on simulations performed with the “Laboratoire de Météorologie Dynamique”
110 (LMD) Martian Mesoscale Model (MMM) (Spiga and Forget, 2009; Spiga et al., 2011). Details
111 about the LMD-MMM and typical test simulations can be found in Spiga and Forget (2009).
112 The hydrodynamical solver (dynamical core) of the LMD-MMM is borrowed from the three-
113 dimensional, fully compressible, non-hydrostatic Weather Research and Forecasting (WRF) model,

114 capable to resolve fine-scale circulations on the Earth (Skamarock and Klemp, 2008). The physical
115 parameterizations for the phenomena left unresolved by the dynamical core (radiative transfer,
116 small-scale mixing) in the LMD-MMM are similar to those developed for the LMD Martian GCM
117 (MGCM) (Forget et al., 1999). Turbulent vertical diffusion (small-scale mixing) is parameterized
118 by a “2.5-order” Mellor and Yamada approach (Mellor and Yamada, 1982), including improvements
119 from Galperin et al. (1988), suitable for the Martian atmosphere prone to strong variability in at-
120 mospheric stability in the near-surface; horizontal diffusion is handled by the built-in ARW-WRF
121 scheme based on horizontal deformation (Smagorinsky, 1963). The transport of water tracers and
122 the formation of clouds are not activated in our nested mesoscale simulations, for it significantly
123 raises the computational cost of those (already expensive) simulations. Moreover, reproducing the
124 water vapor and ice fields in the northern polar regions is a notoriously difficult task (Tyler and
125 Barnes, 2014), which is far beyond the scope of the present paper.

126 The LMD-MMM and LMD-MGCM simulations performed in this paper do not include the
127 recent improvements of the physical parameterizations developed at LMD, namely the interactive
128 dust scheme (Spiga et al., 2013), the radiative transfer of water-ice clouds (Madeleine et al., 2012),
129 and the thermal plume model for boundary-layer turbulence (Colaitis et al., 2013). We rely instead
130 on a less up-to-date version of the LMD-MMM (Spiga and Forget, 2009; Spiga et al., 2011), akin to
131 the one used in Smith et al. (2013) and similar to the published polar mesoscale models (Kauhanen
132 et al., 2008; Michaels and Rafkin, 2008; Tyler and Barnes, 2014), which proved to satisfyingly
133 reproduce the near-surface wind directions observed by frost-streak mapping (Massé et al., 2012).
134 The reproduction of this wind regime above the northern polar cap is an essential basis of the
135 present work, which focuses on local-scale phenomena arising within the regional katabatic flow
136 over the cap. The interactions in polar regions between water-ice clouds and regional circulations
137 through the radiative impact by water-ice particles remain to be investigated in future work.

138 The mesoscale domains employed for polar simulations are centered on the northern pole of
139 Mars and make use of polar stereographic map projection. To downscale our mesoscale simulation

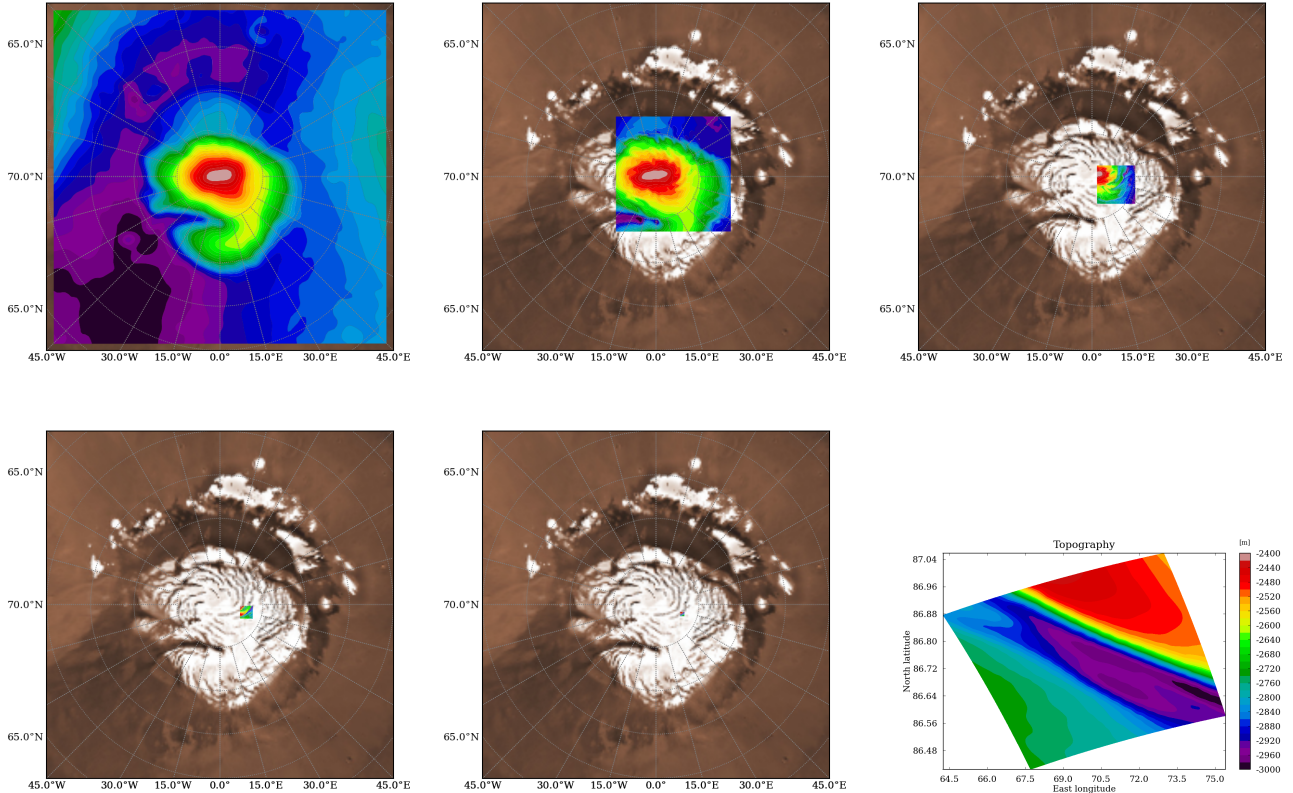


Figure 3: *The five nested LMD-MMM domains set for northern polar trough simulations (#1-5 from left to right and top to bottom). These horizontal domains comprise 121×121 horizontal grid points with grid spacing being 20 km (domain #1 “mother domain”), 6.7 km (domain #2), 2.2 km (domain #3), 740 m (domain #4), 250 m (domain #5). Color shading shows topography and the extent of the LMD-MMM domains. An albedo map of the Martian northern polar region is included in the background to provide context. The bottom-right panel shows the detailed topography of the fifth nest located within a single northern polar trough with an fine horizontal resolution of 250 m.*

140 from the complete extent of the Martian northern polar cap (about 15° latitude wide) towards
 141 a single polar trough (about 10 – 25 km wide, with a characteristic spacing over the polar cap
 142 of 20–70 km, cf. Howard (2000); Pathare and Paige (2005); Smith et al. (2013)), we use 5 “nested”
 143 domains comprising 121×121 horizontal grid points, with grid spacing being respectively 20 km,
 144 6.7 km, 2.2 km, 740 m, 250 m (the factor 3 in grid nesting is the one recommended for most
 145 terrestrial applications with the WRF model). The location and extent of the mesoscale nested
 146 domains are shown in stereographic projection in Figure 3. The nested domains #1, #2, #3 are

147 identical to the mesoscale simulations presented in Smith et al. (2013) (cf. Figure 14) and Smith
148 and Spiga (2017); the present study adds the two unprecedented fine-resolution nested domains
149 #4 and #5 to resolve atmospheric winds within a polar trough. Note that the mother domain
150 #1 is not wide enough to capture the variability imposed by polar transients evidenced in Tyler
151 and Barnes (2005); here we emphasize the major properties of katabatic jump events on Mars,
152 and their variability over two to three days, but our simulations are not tailored to investigate any
153 longer-duration transient effects, which is left as future work.

154 The meteorological fields in nested domain # n are impacted by those predicted in the wider
155 domain # $n - 1$. This configuration is named “one-way nesting”; the possibility of “two-way
156 nesting” (predictions in domain # n also influencing back predictions in wider domain # $n - 1$)
157 does exist in the WRF model, but is not activated here since we do not aim to study the impact of
158 small-scale circulations on the large-scale flow. Initial and boundary conditions for the domain #1
159 in the LMD-MMM are provided by LMD-MGCM simulations (Forget et al., 1999) which use similar
160 physical parameterizations, thereby reducing inconsistencies in physics. The WRF dynamical core
161 can be employed either with or without the hydrostatic assumption: in the first three nested
162 domains #1, #2, #3 hydrostatic equilibrium is assumed, while non-hydrostatic integrations are
163 performed in the two higher-resolution nested domains #4 and #5 where strong local vertical
164 acceleration (namely, katabatic jumps) are expected to be resolved. LMD-MMM integrations are
165 carried out with timesteps of 60 s, 30 s, 10 s, 3 s, 1 s in the respective nested domains; radiative
166 transfer computations are performed every 300 s (about 1/12th of a Martian hour) in all 5 nested
167 domains.

168 The topography resolved in the latest (fifth) nest is also shown in Figure 3 to illustrate how our
169 mesoscale simulations is unprecedented in that it resolves the atmospheric flow at high horizontal
170 resolution within a given polar trough. The resolution used in this latest nest is akin to the kind
171 of resolution used in turbulence-resolving Large Eddy Simulations employed to study the daytime
172 convective boundary layer (Rafkin et al., 2001; Michaels and Rafkin, 2004; Tyler et al., 2008; Spiga

173 et al., 2010). Both the resolutions of the fourth (740 m) and fifth (250 m) nests are within the
174 “grey zone” (or “Terra Incognitae”, Wyngaard, 2004) for resolved / parameterized convection: this
175 is not, however, an issue here since our modeling domains are located at polar latitudes where the
176 Martian atmosphere is characterized by high stability and the daytime boundary-layer convection
177 is weak, if not absent.

178 Along the vertical dimension (Figure 4), 61 levels are set from the surface to a pressure level
179 of about 1 Pa (about 60 km), with a refined spacing close to the surface (first level at an altitude
180 of 8 m above the surface and 10 levels for the first kilometer above the surface), suitable to study the
181 strong near-surface gradients of temperature and winds putatively occurring in katabatic jumps.
182 We carried out simulations with a distinct, alternate, refinement of the vertical grid close to the
183 surface which yield very similar (i.e., almost identical) results to the ones obtained with the vertical
184 grid shown in Figure 4.

185 Our 5-nest mesoscale simulations are carried out close to northern summer solstice ($L_s = 85^\circ$)
186 which is a season propitious to the occurrence of trough clouds according to Smith et al. (2013)
187 (in the companion paper, Smith and Spiga (2017) discuss the possible reasons for this strong
188 seasonal trend). The outputs from LMD-MGCM simulations at this season are directly used
189 as initial and boundary conditions for domain #1: no modification are introduced to enhance
190 the likelihood of katabatic jump occurrences in our LMD-MMM simulations. The LMD-MMM
191 simulations are carried out for 3 Martian days, with the first day serving as a spin-up for the
192 mesoscale circulations (Rafkin et al., 2001; Spiga and Forget, 2009). The dust scenario used in the
193 LMD-MMM simulations is derived from Thermal Emission Spectrometer observations for Martian
194 Year 24 (typical of any Martian Year devoid of global dust storms) and interpolated using a kriging
195 technique (Montabone et al., 2015). Topography in the 5 mesoscale domains is interpolated from
196 the 64-pixel-per-degree laser altimetry (MOLA) dataset (Smith et al., 2001) available on the NASA
197 Planetary Data System. We use the surface thermophysical properties (albedo and thermal inertia)
198 tailored for the northern polar regions published and carefully validated in Tyler and Barnes (2014).

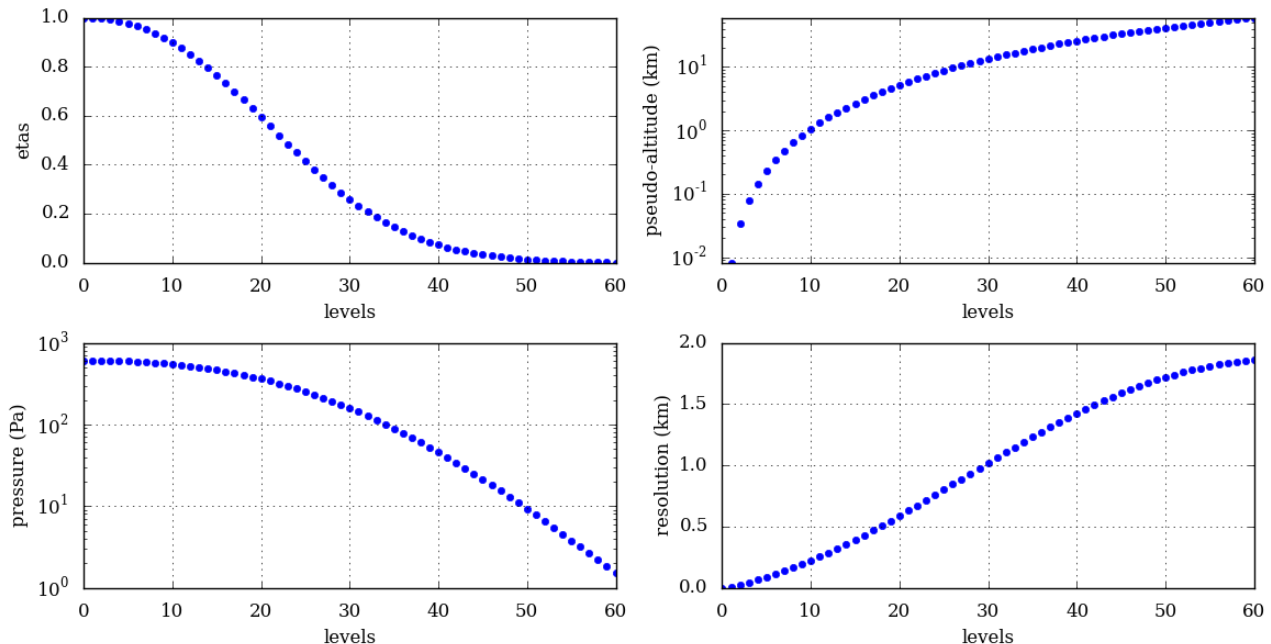


Figure 4: *Vertical discretization in the LMD-MMM domains. Pressure at the top of the mesoscale domains is 1 Pa. x axis denotes model vertical levels. y axis corresponds respectively to (top-left panel) WRF terrain-following mass-based coordinates ($\eta = (p - p_t)/(p_s - p_t)$, where p denotes here the hydrostatic component of pressure, and the t and s subscript denotes respectively the top and surface boundaries), (top-right panel) altitude in km, (bottom-left panel) pressure in Pa, (bottom-right panel) vertical resolution in km. Indicative values shown here are computed with standard surface pressure 610 Pa and scale height 10 km ; the actual model top in simulations is at ~ 57 km altitude.*

199 Similarly to what is described in Smith et al. (2015), and in the companion paper Smith and Spiga
 200 (2017), we use in our LMD-MMM simulations a prescribed CO₂ seasonal deposit that evolves
 201 by L_s date according to infrared measurements of the surface temperature during three typical
 202 Mars years (Titus, 2005), using analytical functions to obtain the “crocus line” i.e. the external
 203 boundary of seasonal CO₂ ice deposits (Kieffer et al., 2000). This accounts for the influence of
 204 the ice-soil thermal contrasts in driving near-surface winds in an analogous phenomenon as “sea-
 205 breeze” on the Earth (Siili et al., 1999; Toigo et al., 2002). Outside the seasonal CO₂ deposits,
 206 surface temperature is calculated in our model by a surface energy balance model (as in Spiga and
 207 Forget, 2009).

208 **3 Results**

209 Results obtained in the domains 1 to 3 with the LMD-MMM are in line with results previously
210 published in Massé et al. (2012) and Smith et al. (2013). A typical near-surface wind field obtained
211 in domain #2 is shown in Figure 5. These wind circulations over the northern polar cap are mostly
212 controlled by katabatic acceleration and Coriolis force, with wind directions in agreement with those
213 obtained by frost streak mapping (Howard, 2000; Massé et al., 2012) and other mesoscale models
214 (Tyler and Barnes, 2005; Kauhanen et al., 2008). Model results in domain #3 show that polar
215 troughs cause a local reinforcement of katabatic winds above the steepening slope on the upstream
216 part of the trough, before katabatic winds are severely weakened downstream, within the bottom
217 of the trough (similar to Figure 14 in Smith et al., 2013). Katabatic winds undergo a diurnal
218 cycle and day-to-day variability under the influence of varying regional atmospheric conditions.
219 The full variability imposed by polar transient eddies described in Tyler and Barnes (2005) is not
220 represented in our mesoscale simulations since we ran our mesoscale model for much less than
221 the 25 simulated days required for those transients to develop and propagate through the model.
222 While this does not jeopardize the main results discussed here, the longer-duration (i.e., weekly)
223 variability of the resolved atmospheric phenomena is not captured by our simulations, which will
224 require a dedicated study in the future to explain why trough clouds did not persist throughout
225 northern summer (Smith et al., 2013; Smith and Spiga, 2017).

226 Analyzing the results from the LMD-MMM in domain #5 (horizontal resolution 250 m) demon-
227 strates that katabatic jumps take place within troughs in the Martian northern polar regions. This
228 is evidenced by Figure 6 where potential temperature is displayed, with wind vectors superimposed,
229 for the second day of simulation at local time 1730 (i.e. 31 hours after the mesoscale simulation was
230 started). Potential temperature is temperature corrected for the impact of adiabatic compression
231 / depression on the temperature (see e.g. Holton, 2004), meaning that potential temperature is
232 conserved for adiabatic motions and represents material contours for the flow. h is the thickness
233 of the katabatic layer, calculated as the distance between the surface and a reference potential

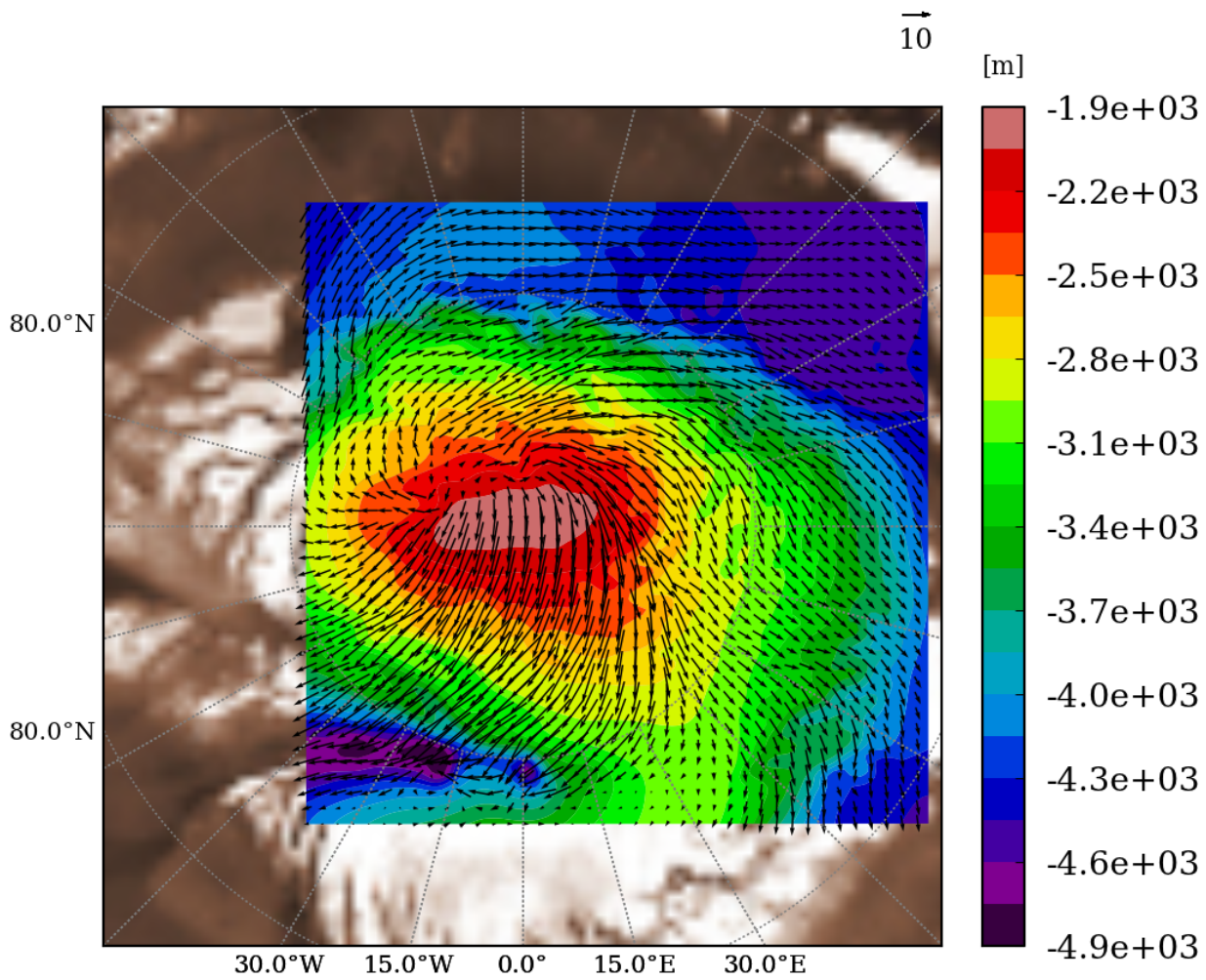


Figure 5: *LMD-MMM results in nested domain #2 ($\Delta x = 6.7$ km). Horizontal wind vectors 50m above the local surface (the reference wind vector with a value in m s^{-1} is included in the top right side of the plot). Topography is in shaded colors. An albedo map of the Martian northern polar region is included in the background to provide context.*

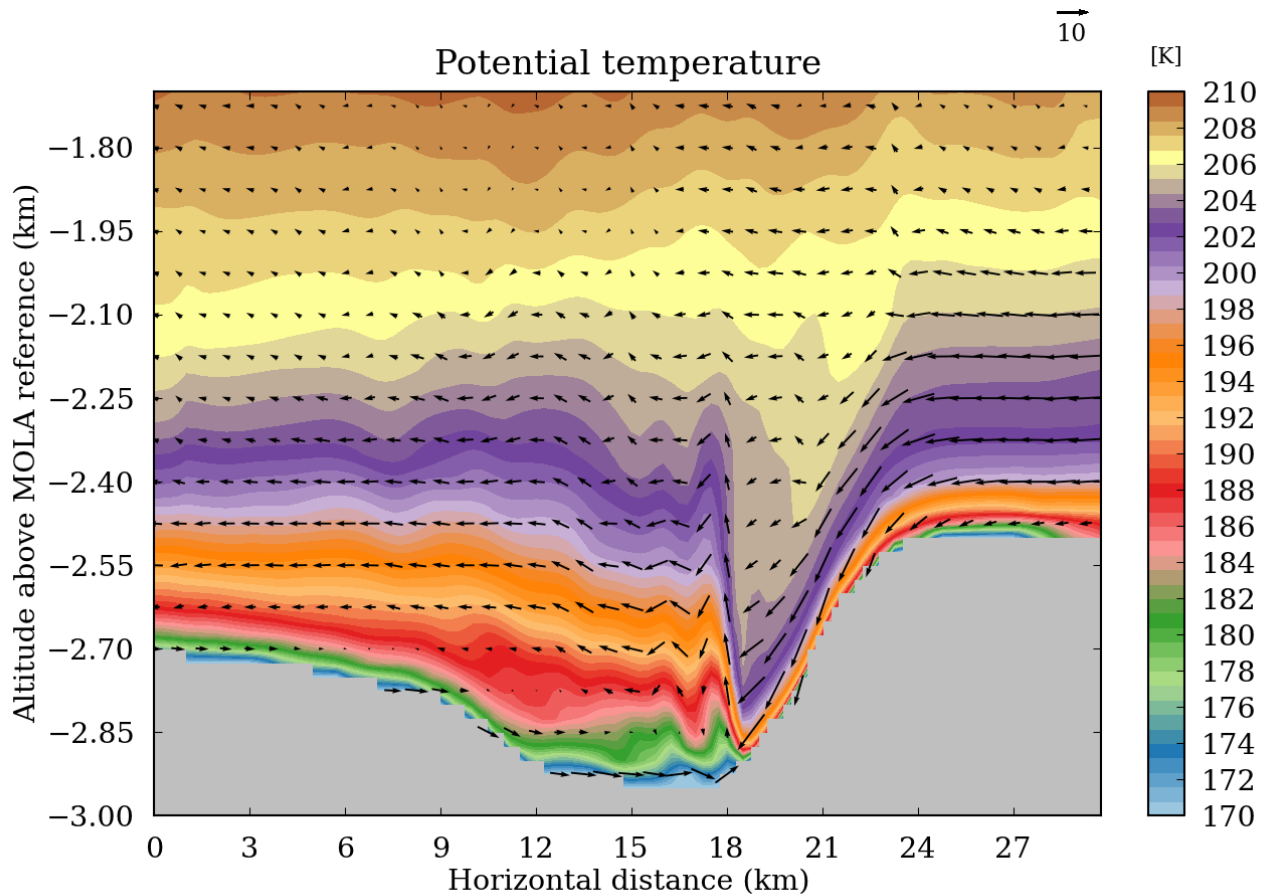


Figure 6: *LMD-MMM results in nested domain #5 ($\Delta x = 250$ m). Horizontal-vertical cross-section of potential temperature with wind vectors superimposed (the reference wind vector with a value in $m s^{-1}$ is included in the top right side of the plot). Vectors plotted every three grid points. This cross-section in the south-north direction (north is on the rightside) is extracted in the middle of domain #5 in the west-east direction. Figure 8 complements this figure by providing a detailed view on horizontal velocities and temperature.*

234 temperature level of 202 K (this value of potential temperature is chosen to enclose the near-
 235 surface atmospheric layers where the incoming katabatic wind velocity is high). Figure 6 shows
 236 that the event predicted in our nested simulations share all the characteristics of katabatic jumps
 237 as described in the terrestrial literature and summarized in Figure 2: acceleration of the atmo-
 238 spheric flow on the slope, intense jump diagnosed by abruptly increasing vertical velocity and flow
 239 thickness, and return flow with negative horizontal velocity. Katabatic jumps simulated by our

240 mesoscale model extends over a height $h \sim 600$ m (cf. also Figure 7 bottom-right panel) and
241 develop mostly at the bottom of the upstream slope of the polar trough, exactly where trough
242 clouds are observed to form (Figure 1). Katabatic jumps are also resolved in domain #4 but
243 both the topographic structure of the trough and the horizontal extent of the katabatic jump are
244 insufficiently resolved in this domain, which makes the inclusion of domain #5 necessary to fully
245 resolve the structure with an appropriate horizontal resolution.

246 Figures 7, 8, 9 represent the properties of the katabatic jump (vertical velocity, horizontal
247 velocity, temperature) respectively in the horizontal, in the vertical, and profiling the flow at
248 constant height above the surface. Our mesoscale model predicts that the vertical wind velocity
249 is very large in the katabatic jump. Ascending motions within the katabatic jump reach 3 m s^{-1}
250 (Figure 8 and 9, top panels) and maintain at those high values during ~ 10 Martian hours. This is
251 a significantly high value in a polar environment where the atmosphere is often stable and devoid
252 of strong vertical motions. The large, positive values of vertical velocity in Figures 7, 8, 9 delimit
253 horizontally the katabatic jump to less than 1 km across (about 4 – 5 horizontal grid points, which
254 justifies *a posteriori* the need to use a modeling downscaling towards a 250 m horizontal resolution).
255 The elongated structure of the katabatic jump in Figure 7 (top-left panel) mirrors the elongated
256 structure of trough cloud evidenced through orbital imagery by Smith et al. (2013) (Figure 7
257 thereby confirms that two-dimensional sections such as Figures 6, 8, 9 are valuable to obtain the
258 properties of the katabatic jump). Furthermore, the presence of trapped gravity waves in the
259 wake of the katabatic jump (cf. vertical velocity as vectors in Figure 6 and contours in Figures 7
260 and 8) provides an explanation for the frequent occurrence of undulations in the morphologies of
261 the trough clouds evidenced through imagery (see Figure 1 in this paper and Figure 3-5 in Smith
262 et al., 2013).

263 The season chosen for our 5-nest mesoscale simulation is $L_s \sim 85^\circ$, known as the peak season
264 for trough clouds (Smith et al., 2013). Smith and Spiga (2017) show through mesoscale modeling
265 that this season is propitious to strong katabatic winds over the northern polar cap, reinforced

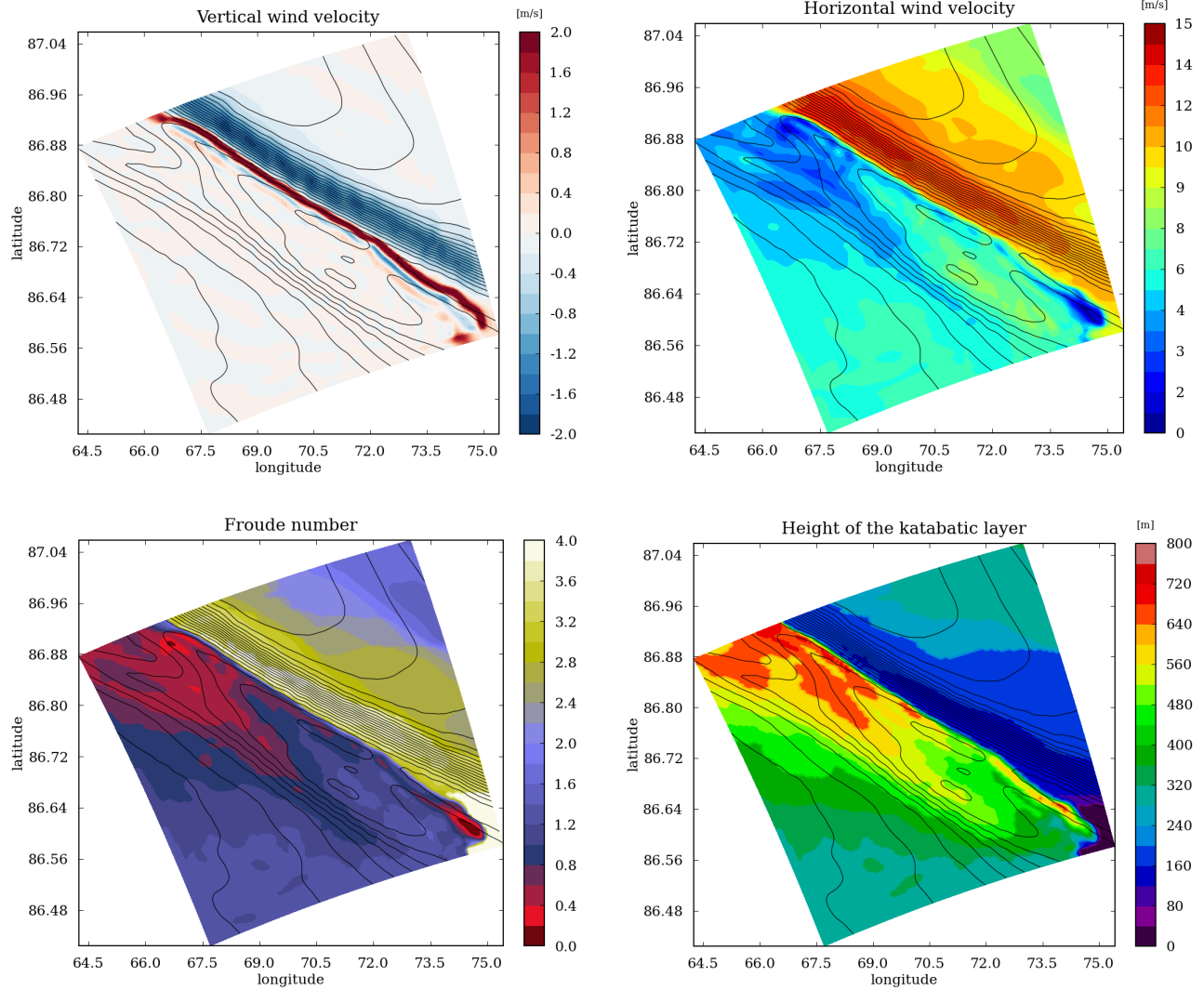


Figure 7: *LMD-MMM results in nested domain #5 ($\Delta x = 250$ m) at local time 1730 on the second day of simulation. Horizontal maps of vertical wind velocity (top-left panel), horizontal wind velocity (top-right panel), Froude number Fr (bottom-left panel), and height of katabatic layer h (bottom-right panel). The first two quantities are shown at a constant altitude of 130 m above the surface; details on computations of the last two quantities h and Fr are provided in the text. Topography is superimposed as contours (see Figure 3, bottom-right panel). North is to the upper part of the plot; katabatic flow is coming from the upper right and accelerate downward perpendicular to the trough. The periodic features seen in the vertical velocity field over the steepest slope of the trough are caused by small artefacts in the MOLA topographical datasets used for our mesoscale simulations. This does not adversely affect the results in this paper.*

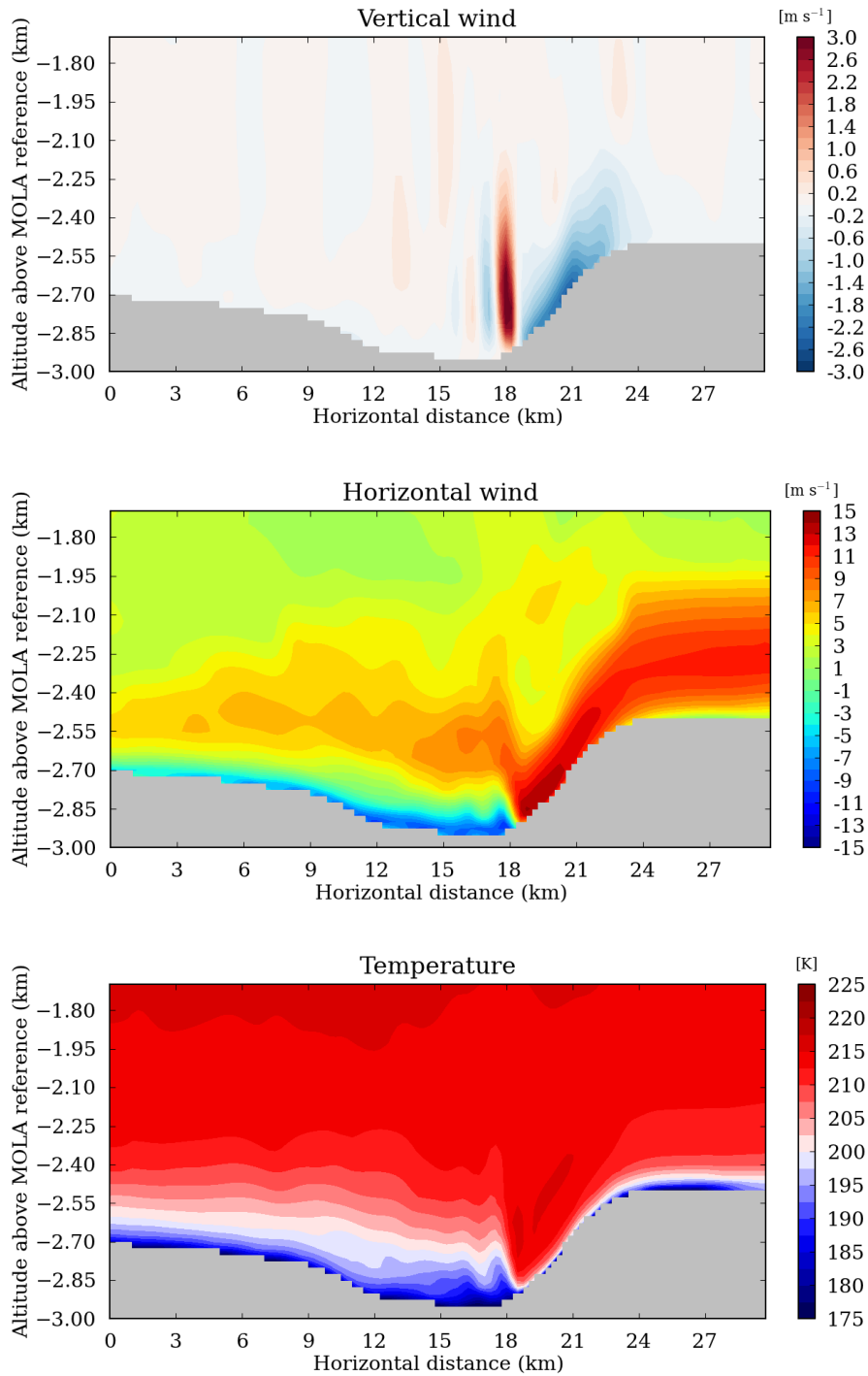


Figure 8: *LMD-MMM results in nested domain #5 ($\Delta x = 250$ m). Horizontal-vertical cross-section of vertical wind velocity (top), horizontal wind velocity (middle), temperature (K). Horizontal dimension is along the south-north direction (north is on the rightside). Horizontal wind is positive in the southward direction (i.e. downstream, following the katabatic flow incoming on a polar trough). Regular temperature is shown in this figure, in contrast with Figure 6 where potential temperature is shown.*

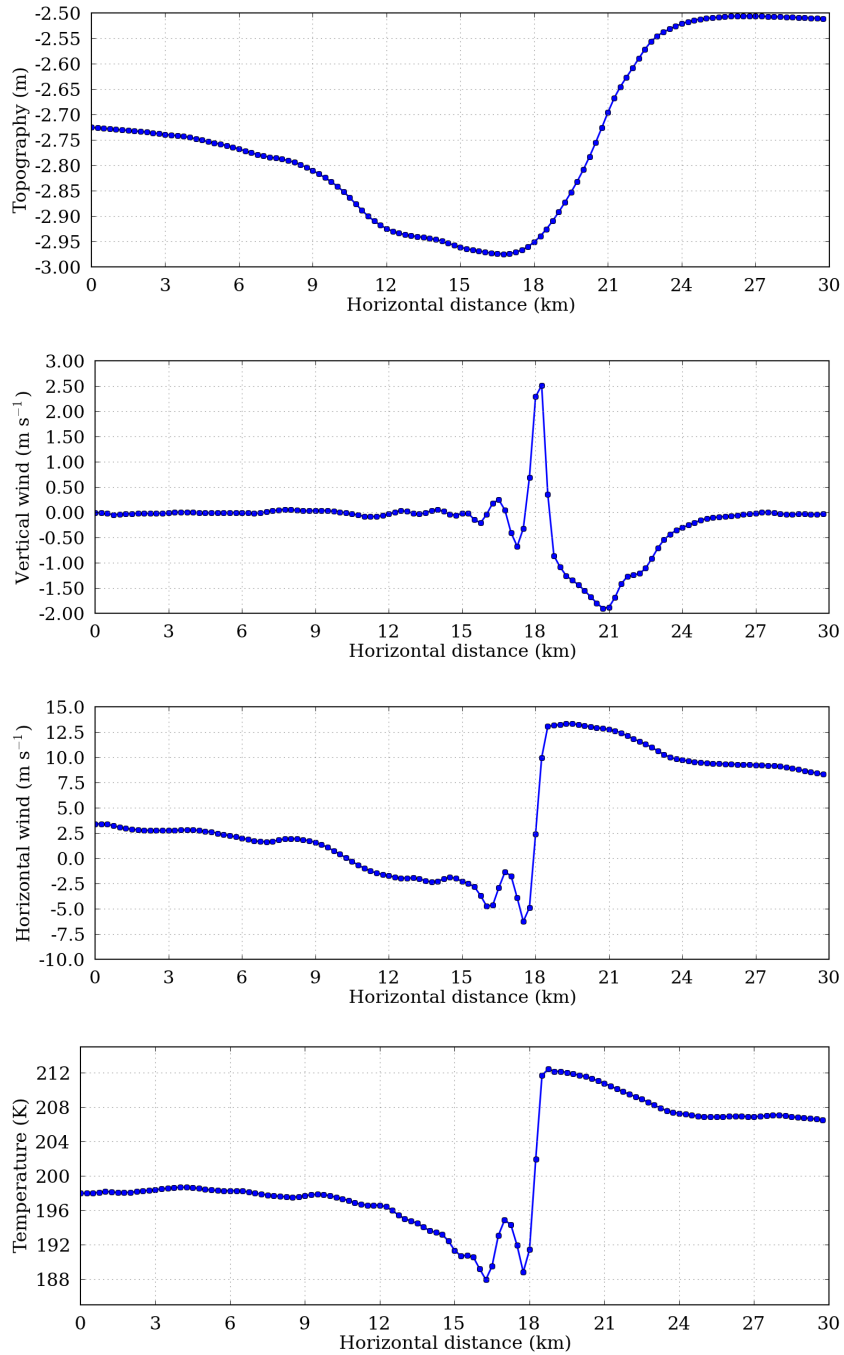


Figure 9: *LMD-MMM* results in nested domain #5 ($\Delta x = 250$ m). From top to bottom: topography, vertical wind velocity, horizontal wind velocity, (regular) temperature. Those fields are shown at the constant altitude of 100 m above the local surface. Horizontal dimension is along the south-north direction, with north on the rightside, as in Figure 8. Horizontal wind is positive in the southward (downstream) direction.

266 by a sea-breeze effect caused by thermal gradients associated with the “crocus line” (the contrast
267 between the retreating seasonal CO₂ ice cap and the residual surface – bare soil or water-ice cap).
268 The location of the trough chosen as a target for our domain #5 is in the vicinity of the crocus line
269 at $L_s \sim 85^\circ$. Furthermore, the intensity of katabatic jumps is sensitive to meteorological conditions
270 around $L_s \sim 85^\circ$, following the modulation of the polar cap’s katabatic flow by diurnal and day-
271 to-day variability (through baroclinic wave activity or polar transients). Over the course of our
272 3-day mesoscale simulation, two additional katabatic jump events analogous to the reference event
273 detailed in Figures 7, 8, 9 are produced (Figure 10), although their amplitude is lower (1 m s^{-1}
274 instead of 3 m s^{-1} in vertical velocity) and undulations do not appear as clearly as in the reference
275 event. We found that the variations of the incoming katabatic wind in the considered polar trough
276 in Figure 10 is caused by transient eddies with dominant wavenumber 1, akin to the phenomena
277 described in Tyler and Barnes (2005) later in northern summer. A more complete analysis of the
278 diversity of katabatic jumps occurring over the northern polar cap as a result of polar transients
279 will require longer-duration mesoscale simulations as in Tyler and Barnes (2005).

280 As is mentioned above, the katabatic jump is associated with a sudden drop in the horizontal
281 wind speed (Figure 7, top-right panel; Figure 9, middle panel), i.e. an abrupt drop of velocity
282 within an otherwise smooth katabatic flow that continues almost uninterrupted downstream of the
283 polar trough. Figure 6 and Figure 8 (middle panel) show that the horizontal wind in the vicinity of
284 the katabatic jump is even reversing to upstream direction (with respect to the incoming katabatic
285 flow) with a significant wind velocity of $\sim 10 \text{ m s}^{-1}$. As it is the case with polar katabatic jumps
286 on the Earth, the area immediately downstream of the katabatic jump is prone to enhanced mixing
287 and/or stationary flow, which is compliant with the cloud structures observed from orbit in those
288 areas (see Figure 1 in this paper and Figures 3-5, 7-8, 11 in Smith et al., 2013).

289 Katabatic jumps borrow their name from their similarity to hydraulic jumps in open channel
290 flow, which form as the flow transitions from a supercritical “shooting” regime with $Fr > 1$ to a
291 subcritical “tranquil” regime with $Fr < 1$ (Ball, 1956), where Fr is the Froude number which can

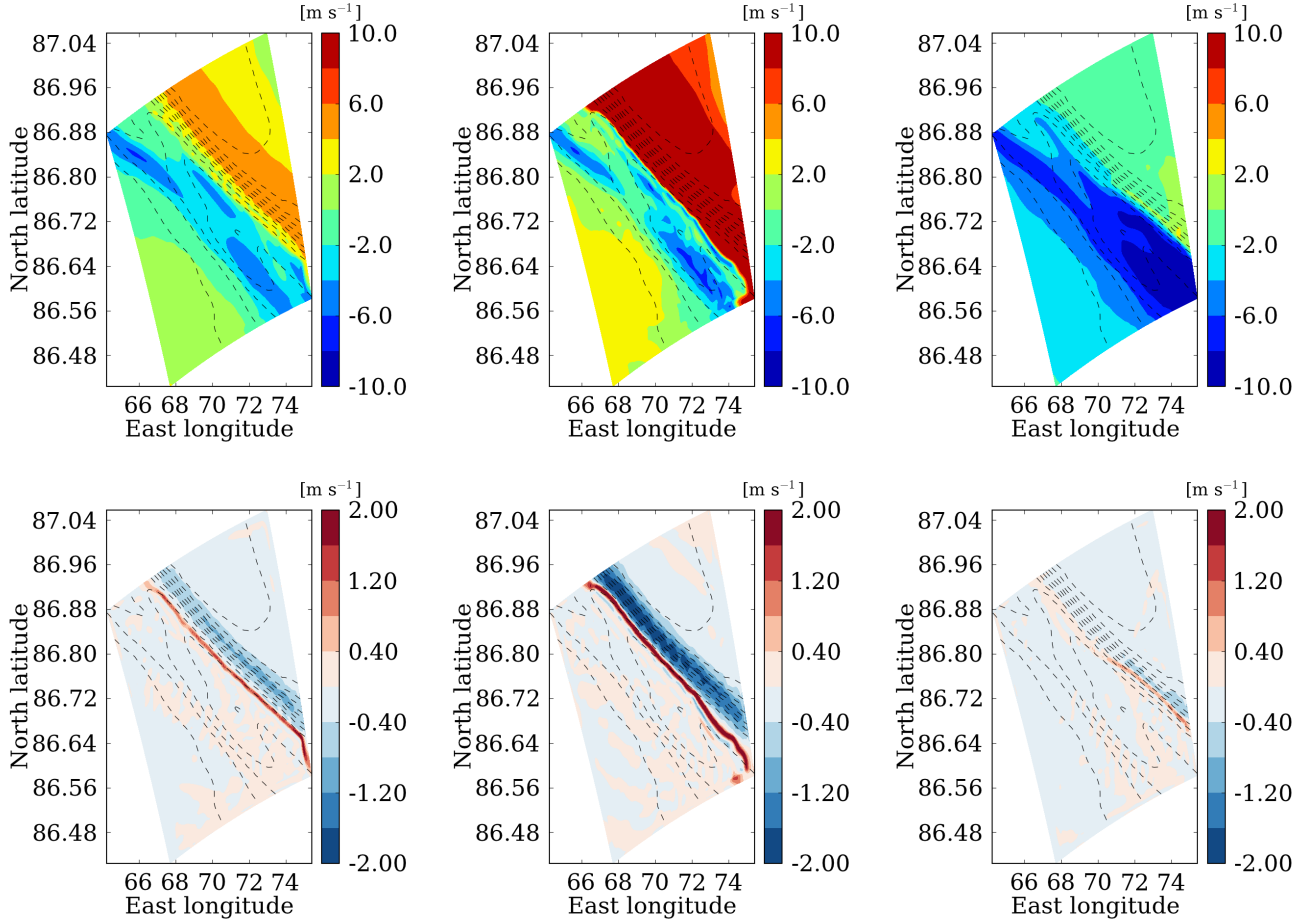


Figure 10: *LMD-MMM results in nested domain #5 ($\Delta x = 250$ m) at local time 1730 on the first, second and third day of simulation (respectively left, middle, right panels). Horizontal maps of south-north horizontal wind (top panels, with positive values for downstream direction) and vertical wind velocity (bottom panels). Topography is superimposed as contours. Middle row shows the reference katabatic jump event detailed in Figure 7.*

292 be expressed after Pettré and André (1991)

$$Fr = \frac{u}{\sqrt{\frac{\Delta\theta}{\theta} g h}}$$

293 where u is the along-slope wind speed, $\Delta\theta/\theta$ the near-surface inversion of potential temperature,
 294 g is the acceleration of gravity, and h is the thickness of the katabatic layer (as is defined above

295 and in Figure 7 bottom-right panel). We mapped the Froude number in domain #5 in Figure 7
296 (bottom-left panel). The katabatic flow is highly supercritical ($Fr > 2$ with peak values close
297 to 4.5) over the upstream slope, where the katabatic wind speed is larger and the katabatic layer
298 is thinner (two essential factors to obtain a supercritical flow), and this flow undergoes an abrupt
299 transition to subcritical values ($Fr = 0.2 - 0.4$) at the bottom of the slope. The katabatic jump
300 is found where the incoming flow transitions from a supercritical to a subcritical regime. This
301 is compliant with the katabatic jumps occurring close to the slope break between the upstream
302 trough slope and the downstream trough bottom (Figures 8 and 9), which is also where trough
303 clouds are observed (Smith et al., 2013).

304 The acceleration of the katabatic wind in the upstream slope of the trough pertains to a larger
305 class of atmospheric phenomena named “downslope windstorms” (Durrán, 1990; Magalhães and
306 Young, 1995). This phenomena occurs for instance on the leeside of a mountain when an incident
307 large-scale wind flow vanishes to zero near the surface and encounters a non-linear boundary
308 condition (Lott, 2016). Furthermore, strong downslope winds are especially favored in stable
309 conditions. Convective instability at low level in the mountain leeside might occur with a hydraulic-
310 jump-like morphology. Both the downslope windstorm and the hydraulic jump do not require
311 gravity wave breaking to form. Downslope windstorms with similar properties were shown to
312 arise as a result of a katabatic flow impinging on the Meteor Crater topographic depression on
313 the Earth, with hydraulic jumps arising shall the background wind be strong enough and the
314 crater be deep enough (Lehner et al., 2016). The katabatic jumps in polar troughs bear a strong
315 resemblance with this terrestrial analog. Besides, this terrestrial example suggests that katabatic
316 jumps shall occur downstream of the rims of a deep-enough crater on Mars on which a strong-
317 enough regional-scale katabatic flow is impinging – such as, for instance, within Gale Crater located
318 at the dichotomy boundary (Haberle et al., 2014; Pla-Garcia et al., 2016; Rafkin et al., 2016).
319 Hydraulic jumps generated by katabatic winds were also thought to cause the early morning
320 elongated clouds observed by the Viking Orbiters during late northern spring and early summer

321 (Kahn and Gierasch, 1982); mesoscale modeling provided, however, an explanation based on the
322 propagation of an atmospheric bore wave generated by a katabatic front (Sta. Maria et al., 2006),
323 which would be the propagating equivalent of the nearly-stationary katabatic jump we model here.

324 An important point to discuss ice migration in Martian polar troughs is to know whether the
325 downslope windstorm and the katabatic jump are conducive to the formation of trough clouds as
326 hypothesized in Smith et al. (2013). Katabatic jumps in terrestrial polar regions often lead to the
327 formation of clouds named “wall-of-snow” (Pettré and André, 1991), and the visual appearance
328 of elongated trough clouds on Mars (Figure 1) is reminiscent of those terrestrial clouds. The tem-
329 perature simulated in the polar trough is shown in Figure 8 and 9 (bottom panels). Within the
330 katabatic flow blowing downslope the trough, atmospheric temperatures are strongly increasing
331 under the influence of adiabatic compression (Spiga et al., 2011) to reach 210 K; when the katabatic
332 flow undergoes the katabatic jump, atmospheric temperatures are abruptly decreasing to 190 K.
333 Both conditions are highly propitious to the formation of water-ice clouds at the location of the
334 katabatic jump: not only the downslope windstorm implies enhanced sublimation of ground ice to
335 the atmosphere (by enhanced sensible heat flux), but this water-vapor-rich katabatic flow quickly
336 encounters a drop in saturation vapor pressure caused by the -20 K drop in atmospheric tem-
337 perature associated with the katabatic jump. Water ice at the surface is available for sublimation
338 at $L_s = 85^\circ$ in the trough modeled here; the presence of CO_2 ice at other seasons, and the seasonal
339 variability of the surface temperature of water-ice deposits, could yield a much less favorable case
340 for sublimation over the northern polar cap, thereby accounting for the strong seasonality of trough
341 clouds (cf. companion paper Smith and Spiga (2017)).

342 Using Clausius-Clapeyron equations for the Martian atmospheric conditions (e.g., Montmessin
343 et al., 2004), we compute the saturation mass mixing ratio q_{sat} for the pressure and temperature
344 conditions simulated by our model (cf. Figure 8). We then obtain specific humidity $H = q/q_{\text{sat}}$,
345 where q is a typical near-surface mass mixing ratio for water vapor in the summertime northern
346 polar regions, 3×10^{-4} kg/kg according to LMD-MGCM simulations compiled in the Mars Climate

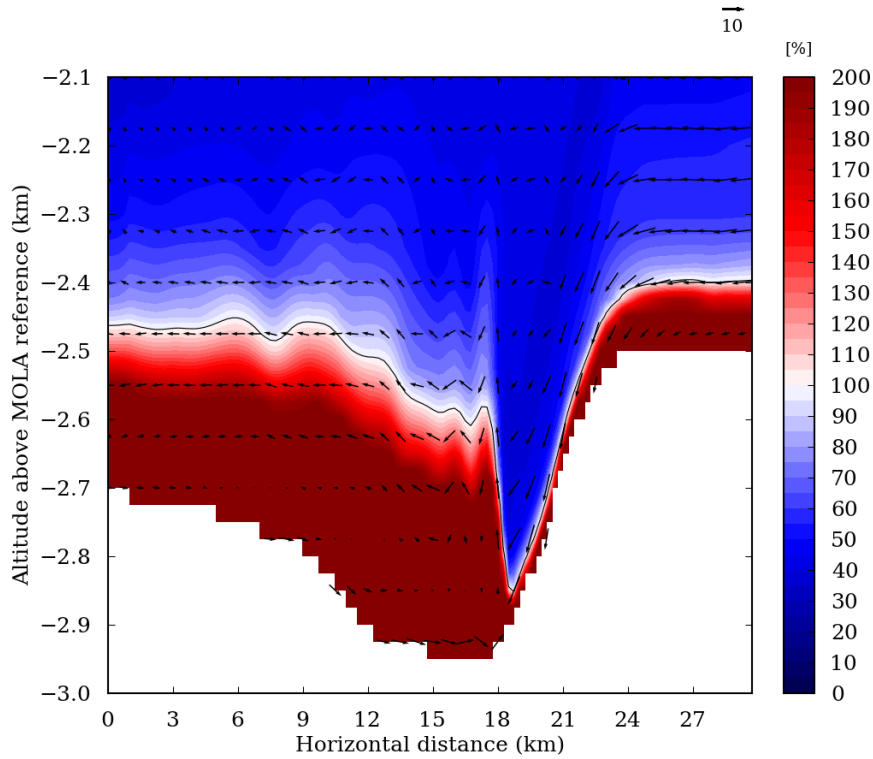


Figure 11: *LMD-MMM results in nested domain #5 ($\Delta x = 250$ m). Horizontal-vertical cross-section of specific humidity (see text for details on calculations) with wind vectors superimposed (the reference wind vector with a value in $m s^{-1}$ is included in the top right side of the plot). The 100% humidity limit is indicated by a solid line. Vectors are plotted every three grid points. This cross-section in the south-north direction (north is on the rightside) is extracted in the middle of domain #5 in the west-east direction.*

347 Database (Lewis et al., 1999; Millour et al., 2015). This quantity H is shown in Figure 11,
 348 which strongly suggests that the incoming katabatic flow undergoes subsaturated conditions when
 349 entering the polar trough, as a result of adiabatic warming, before it undergoes supersaturated
 350 conditions at the katabatic jump, conducive to the formation of water-ice clouds. The fact that
 351 in Figure 11 the strongest horizontal gradient of specific humidity coincides with the location of
 352 the katabatic jump means that the water-ice cloud forms very close to this katabatic jump. This
 353 explains that the horizontal elongated morphology of this jump (Figure 7) matches the morphology

354 of the observed trough clouds (Smith et al., 2013). Our modeling is thus in line with the putative
355 scenario described in Smith et al. (2013) for ice migration within polar troughs, with the upstream
356 slope of the trough being ablational for ice deposits (through enhanced sublimation and transport
357 away from the source) and the downstream bottom of the trough being depositional for ice deposits
358 (through water-ice cloud formation). It remains to be determined with a full microphysical model
359 (e.g., Navarro et al., 2014; Spiga et al., 2017) whether the cloud formation implies deposition of
360 ice on the ground by precipitation (if the cloud does not form immediately above the surface) or
361 by direct deposition (if the cloud forms right above the local surface, or if transported water vapor
362 directly recondenses at the surface).

363 4 Discussion

364 Our conclusions can be summarized as follows.

- 365 1. We demonstrate through high-resolution nested mesoscale modeling that powerful katabatic
366 jumps occur within the northern polar troughs with an incoming downslope (katabatic)
367 windstorm.
- 368 2. The katabatic jumps occur while the incoming flow transitions from supercritical shooting
369 conditions on the upstream slope of the trough to subcritical tranquil conditions in the
370 downstream part of the trough, making katabatic jumps analogous to hydraulic jumps in
371 open channel flows.
- 372 3. Those Martian phenomena are strikingly similar to the Loewe phenomena described in
373 terrestrial polar regions (Pettré and André, 1991).
- 374 4. The combination of the downslope windstorms and katabatic jumps make the Martian at-
375 mospheric conditions in polar troughs propitious to the formation of water-ice clouds which
376 morphology is similar to the observed trough clouds.

377 Our conclusions strengthen the scenario hypothesized by Smith et al. (2013): trough clouds are
378 the visible manifestation of katabatic jumps, which might support a mechanism for ice migration
379 over geological through cyclic steps (Kostic et al., 2010), thereby providing an explanation for
380 the stratigraphy revealed by orbital radar measurements. Despite our new conclusions obtained
381 through unprecedented mesoscale modeling, further work is needed to describe the definitive sce-
382 nario about ice migration over the Martian northern polar caps.

- 383 • Mesoscale modeling coupled to water vapor transport and water ice microphysics is needed
384 to understand how deposition of water ice on the surface occurs from precipitation or direct
385 deposition from trough clouds.
- 386 • The seasonal and spatial variability of trough clouds needs to be assessed with more extensive
387 simulations accounting for the combined influence of katabatic acceleration, baroclinic waves
388 and polar transients.
- 389 • The stability and possible migration over geological timescales of water ice surface reservoirs
390 in the Martian polar regions is left to be investigated by paleoclimatic mesoscale modeling;
391 the influence of the varying obliquity is probably an important element to future studies.
- 392 • It remains to be elucidated how troughs are initiated on the polar cap surface, and how
393 possible katabatic jumps occurring without an initial slope break could form initial surface
394 erosional features that could be reinforced through time to form polar troughs.

395 On a broader perspective, the present work emphasizes how surface-atmosphere interactions
396 are key to understand the properties and evolution of the polar regions of Mars. Additional
397 observations obtained by a dedicated polar orbiter or lander would allow for an in-depth validation
398 of the scenario built here upon numerical modeling.

399 Acknowledgments

400 We thank François Lott for helpful suggestions on mesoscale atmospheric dynamics, Jack Holt for
401 discussions on Martian polar processes, and Dan Tyler for discussions on atmospheric processes in
402 Martian polar regions. Isaac Smith acknowledges funding from the Fulbright Commission during
403 a scientific visit at LMD (Paris) in 2014, which initiated this study. Aymeric Spiga acknowledges
404 funding from Centre National d'Études Spatiales (CNES). We thank two reviewers for constructive
405 and thorough comments which helped us to improve the manuscript prior to publication.

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