



HAL
open science

Retrogressive thaw slumps on ice-rich permafrost under degradation: Results from a large-scale laboratory simulation

François Costard, L Dupeyrat, A Séjourné, F Bouchard, A Fedorov, B Saint-Bézar

► **To cite this version:**

François Costard, L Dupeyrat, A Séjourné, F Bouchard, A Fedorov, et al.. Retrogressive thaw slumps on ice-rich permafrost under degradation: Results from a large-scale laboratory simulation. *Geophysical Research Letters*, 2021. hal-03065348v2

HAL Id: hal-03065348

<https://hal.science/hal-03065348v2>

Submitted on 14 Dec 2020 (v2), last revised 14 Oct 2021 (v3)

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

1 **Retrogressive thaw slumps on ice-rich permafrost under degradation:**

2 **Results from a large-scale laboratory simulation**

3 **F. Costard¹, L. Dupeyrat¹, A. Séjourné¹, F. Bouchard¹, A. Fedorov², and B. Saint-**
4 **Bézar¹**

5
6 ¹Géosciences Paris-Saclay (GEOPS), CNRS/Université Paris-Saclay, Orsay, France,

7 ²Melnikov Permafrost Institute, Russian Academy of Sciences, Yakutsk, Russia

8
9 Corresponding author: François Costard (francois.costard@universite-paris-saclay.fr)

10 **Key Points:**

- 11 • Retrogressive thaw slumps (RTS) result from the thermal destabilization of ice-
12 rich permafrost as a consequence of increasing subsurface temperature
- 13 • Experimental RTS were designed to simulate the thawing of ice-rich permafrost
14 with vertical and horizontal ice layers
- 15 • Our laboratory simulations show how ground ice heterogeneities influence RTS
16 development

17

18 **Abstract**

19 In the ice-rich permafrost of the Arctic regions, thermokarst erosion on slopes induces the
20 formation of large-scale retrogressive thaw slumps (RTS). They have significant
21 geomorphological, hydrological and biogeochemical impacts on the landscape. Further research
22 is thus needed to better understand the respective effect of ice content and permafrost
23 heterogeneities on the dynamics of these erosional features. Here we present results of a full-scale
24 physical modelling of RTS development in a cold room. The experimental setup was designed to
25 simulate and compare two ground-ice settings (ice wedges, icy layers) with the thawing of ice-
26 poor permafrost (i.e.,reference model). Our results show that the melting of the icy layers induces
27 a loss of decohesion of the overlapping frozen soil. The heterogeneous frozen soil with ice
28 wedges needs a longer time until degradation, but undergoes a stronger and faster decohesion of
29 its structure during the thawing phase.

30

31 **1 Introduction**

32 Over the last few years, various studies have documented significant impacts of recent
33 global warming across the Arctic since the mid-twentieth century, with a preferential thermal
34 degradation of ice-rich permafrost and the related release of previously sequestered carbon (e.g.,
35 Grosse et al., 2011; Romanovsky et al., 2010). These studies indicate a sensitivity of cold Arctic
36 permafrost to climate-driven thermokarst (thaw) initiation (Olefeldt et al., 2016). The
37 development of thermokarst results from the thermal destabilization of ice-rich permafrost as a
38 consequence of the increase in subsurface temperature (Soloviev, 1973; French, 2017). Over the
39 last decades, this ice-rich permafrost has been highly vulnerable and prone to extensive
40 degradation within the continuous permafrost zone of Eurasia and North America (Olefeldt et al.,
41 2016; Lewkowicz and Way, 2019).

42 Retrogressive thaw slumps (RTS) represent a dynamic form of thermokarst. They expand
43 inland by melting of exposed ground ice in the headwall to form landslide-like U-shaped scars
44 (Lantuit and Pollard., 2008). In the western Canadian Arctic, widespread permafrost degradation
45 accompanied by active RTS development is mostly observed in ice-rich formerly glaciated
46 landscapes (Olefeldt et al. 2016, Rudy et al., 2017; Murton et al., 2017; Kokelj et al., 2017; Nitze
47 et al., 2018). These ice-rich areas contain buried glacier ice that has been preserved under
48 permafrost conditions (Rudy et al., 2017). They are present in the form of horizontal massive ice

49 layers. In Eastern Siberia, RTS are mostly developed in the Yedoma (Pleistocene-aged) ice
50 complex area (Opel et al., 2019). In central Yakutia (CY), the upper section of permafrost
51 contains Pleistocene ice-rich sediments, up to 20–50 m in thickness, containing ~70-80 % of ice
52 by volume (Soloviev, 1973; Schirrmeister et al., 2020). The Yedoma ice complex is dominated
53 by huge ice wedges up to a few meters wide and 15-25 meters high within sandy deposits (Opel
54 et al., 2019; Shepelev et al., 2020). In some places, these large-scale ice wedges can be
55 continuous and form a massive horizontal ice layer at the base (Opel et al., 2019). In CY, RTS
56 are smaller than in the High Arctic. Amphitheater-shaped slump hollows are 30-50 m wide and
57 are characterized by (1) their concave profile of collapsed banks and (2) highly-degraded conical
58 polygons (Figure 1) with wide troughs referred to as “baydjarakhs” (Soloviev, 1979). This ice-
59 rich permafrost is characterized by its medium and/or fine sand (D_{50} 200 – 300 μm) and silty
60 material (D_{50} 18 – 30 μm) of Quaternary lacustrine or aeolian (loessic) origin (Schirrmeister et
61 al., 2020; Soloviev, 1973). These sediments contain ~30–40% of ice by volume, in addition to
62 large syngenetic ice wedges (Ulrich et al. 2014; Strauss et al., 2017).

63 Several studies estimated the development of RTS in the High Arctic with extended photo
64 time series (30 years). Headwall retreat rates range from 0.5–1 m yr^{-1} along thermokarst lakes in
65 the tundra uplands of the Mackenzie Delta, Canada (Lantz and Kokelj, 2008) to an average up to
66 0.68 m.yr^{-1} and 29.0 $\text{m}^3.\text{m}^{-1}.\text{yr}^{-1}$ on the coast of Herschel Island, Canada, next to the Beaufort Sea
67 (Lantuit and Pollard, 2008; Obu et al., 2016). Headwall retreat rates of up to 13 m yr^{-1} were
68 recently observed in the western coast of Kolguev Island, Russia (Kizyakov et al., 2013). In CY,
69 the maximal average headwall retreat of these RTS can reach 3.16 m yr^{-1} for the 2012-2013
70 period (Séjourné et al., 2015), although maximal values of 30 m yr^{-1} were recorded at the
71 Batagay megaslump in Yakutia (Opel et al., 2019). With a length of more than 1000 m, a width
72 of 800 m and 60 m high headwalls in 2019, this mega-RTS is representative of three typical
73 cryolithological types of permafrost at a single site. It is characterized by its ice-rich Yedoma
74 permafrost close to the surface, underlain by ice-poor sands, while at its base a pure ground-ice
75 layer is present (Shepelev et al., 2020). In close association with the erosion of RTS, small mud-
76 flows settling on gentle slopes are observed from melted ice wedge network and contribute to the
77 transport of sediments, dissolved elements and meltwater from the melting of ice wedges (Figure
78 1a). These RTS liberate both soluble materials and organic carbon accumulated in the
79 sedimentary deposits that are transported by runoff (Vonk et al. 2015; Shakil et al., 2020). The

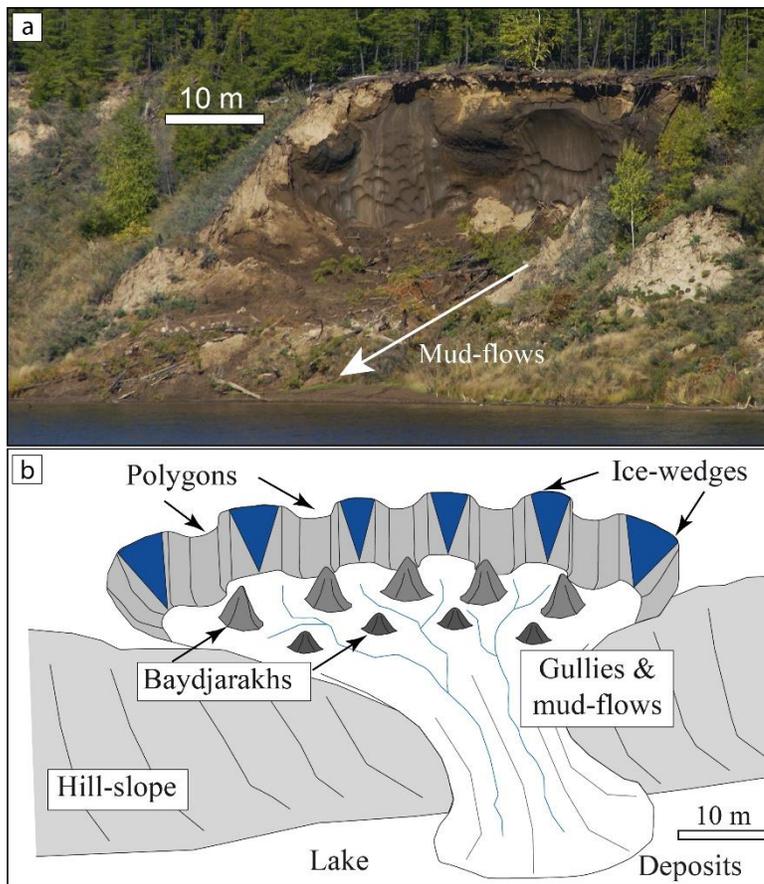
80 abrupt thawing of ice-rich permafrost is thus a major issue because of the positive feedback on
81 global climate (Schuur et al., 2008; Turetsky et al., 2020).

82 The possible causes of present-day initiation of thermokarst are numerous but their
83 interrelations are not well understood. According to Ward Jones et al. (2019), record summer
84 warmth in 2011 and 2012 in the Canadian Arctic promoted thermokarst initialization in
85 previously unaffected terrain. Permafrost degradation is always initiated by deepening of the
86 active-layer as a consequence of an increase in subsurface temperature (Fedorov et al., 2014;
87 Fedorov and Konstantinov, 2009). Retrogressive thaw slumps are initiated by a variety of
88 mechanisms that expose ice-rich sediments, including mechanical erosion by fluvial processes,
89 wave action, thermally driven subsidence along the banks, and mass-wasting (Burn and
90 Lewkowitz 1990; Wolfe et al., 2001; Lantuit and Pollard, 2008; Kokelj et al., 2017; Lacelle et
91 al., 2010). According to Segal et al. (2016), climate change, in conjunction with increases in air
92 temperature and precipitation, induces an acceleration of RTS activity. According to Séjourné et
93 al. (2015), RTS in CY mainly occur along the south- to southwest-facing banks of thermokarst
94 lakes due to solar insolation and increased air temperature. However, the timing of this recent
95 increase over the last few decades and its links to climate have not been fully established (Lantuit
96 and Pollard, 2005; Lantz and Kokelj, 2008; Wolfe et al., 2001). Local differences in recent
97 erosion rates across the circumpolar Arctic may be due to a difference in local climate, headwall
98 height, and ice content (Kokelj and Jorgenson, 2013). Other authors suggest that the presence of
99 massive ground ice (ice wedges, massive segregated ice, buried glacier ice) is a precondition for
100 RTS development. (Leibman et al., 2008; Lantuit et al., 2012). While it was observed that RTS in
101 CY develop in Yedoma ice complex terrain with huge syngenetic ice wedges (Figure 1b), the
102 heterogeneous structure of ice-rich permafrost and its subsequent influence on thermokarst
103 degradation has not been analyzed. The relative importance of the main parameters affecting RTS
104 are, however, difficult to assess in the field. Numerous factors work simultaneously, and their
105 interdependence makes their separate analysis difficult. Physical modelling provides a unique
106 tool for a more detailed monitoring compared to field observations.

107 Here, we present results of a full-scale physical modelling of RTS development in a cold
108 room. For this purpose, the experimental RTS were designed to simulate the thawing of ice-rich
109 permafrost with various heterogeneities (ice wedges, icy layers) that were supposed to be
110 representative of RTS development in CY. In a first experimental set-up, we simulated artificial

111 ice wedges (vertical ice layers), which are common in large parts of continuous permafrost across
 112 the Northern Hemisphere (i.e., polygonal landscapes and Yedoma ice complex). In a second set-
 113 up, horizontal ice layers mimic a second type of ice-rich permafrost that might find its natural
 114 expression in formerly glaciated Arctic landscapes with soft sediments and large masses of
 115 segregated ice or buried glacier ice. Each experimental set-up was compared to reference frozen
 116 ground with no excess ice. Therefore, this study is representative of three typical cryolithological
 117 types of permafrost in unconsolidated material, which cover several hundreds of thousands of
 118 km² in the Arctic.

119



120
 121 Figure 1: (a) Headwall of retrogressive thaw slumps (RTS) along a thermokarst lake in central
 122 Yakutia (Eastern Siberia). Permafrost with alluvial sandy loam containing up to 40-50% of ice by volume
 123 and syngenetic ice-wedges of ~7 m in thickness. (b) The localized melting of ice wedges along the banks
 124 of thermokarst lakes forms highly-degraded conical polygons with wide troughs (baydjarakhs). These
 125 typical structures are indicative of heterogenous ice-rich permafrost with a preferential erosion along ice
 126 wedges.

127

128

2 Experimental setup

129

130

131

132

133

134

135

136

137

138

139

140

141

142

143

144

145

146

147

148

149

150

151

152

153

154

155

156

157

Physical modelling of RTS was undertaken using large-scale frozen soils in a temperature-controlled chamber. The major purpose of the experiment was to examine the respective effects of different ground-ice content and type affecting the ground thermal regime and vulnerability to subsidence. We used the cold-room at the GEOPS laboratory (University Paris Saclay, France) dedicated to the physical modelling in periglacial geomorphology. The specificity of our experiments was to simulate the thawing of ice-rich permafrost, compared to the thawing of ice-poor permafrost (i.e., reference conditions). In a first experimental setup (experiment n°1), we worked with artificial ice wedges (vertical ice layers, see SI1), which are common in large parts of continuous permafrost across the Northern Hemisphere (i.e., polygonal landscapes). In a second setup (experiment n°2), we worked with horizontal ice layers mostly typical of formerly glaciated Arctic landscapes with soft sediments and large masses of segregated ice or buried glacier ice (see SI2). Our experiment is composed of a rectangular box of 2.5 m x 2.5 m wide and 0.5 m deep in which fine sand ($D_{50} = 200 \mu\text{m}$) was saturated with water (Figure 2). Here, we used the grain size distribution of Bayeux sand as an analog to sands from the banks of most RTS in CY. The basal slope was 5° for the release of meltwater overflow during the thawing phase. A foam rubber was placed all around the experiment to ensure thermal insulation. For each experiment, the sediment structure was homogeneous and had a water content of 20% by volume. A well-proven technique was used to create a homogeneous porous medium of more than one ton of saturated soil (Costard et al., 2003). To obtain the optimal porosity at water saturation, we first mixed the sand with water to a wet density of 1.81 g.cm^{-3} corresponding to the “Proctor compaction” technique (Costard et al., 2003). We manually compacted layers of 10 cm each and then put them on top of each other until 0.5 m thickness was reached. Such technique allows a parametric control on a homogeneous block of about 1 ton of already saturated material with a volumetric water content of 20 %. There might be some small interfaces (thin unconformities), but they remain insignificant. In order to evaluate the effect of permafrost heterogeneities on the development of RTS, we artificially built vertical and horizontal ice layers to respectively mimic ice wedges and icy beds (volumetric ice content of 100 %). Before the freezing phase, each icy block was installed within the right side of the experiment (see Figures SI1 and SI2). Finally, all the material was frozen in the cold room at -

158 25°C and then progressively reheated at -10°C (Figures SI3 and SI4). This technique allowed us
 159 to prepare homogeneous samples with reproducible densities and without ice segregation
 160 (Costard et al., 2003). Figures 2, SI1 and SI2 show the homogeneous model S1 (or reference
 161 model) on the left, and the heterogeneous model S2, which corresponds to a ice rich permafrost
 162 with a total of 30 % ice by volume, on the right. The size, geometry and spacing of these icy
 163 blocks within the frozen soil represent a simplification of the natural setting, but the main
 164 objective here was to understand the influence of heterogeneities on permafrost degradation from
 165 a strictly thermal point of view.



166
 167 Figure 2: RTS experiment n°1 (with ice wedges) of 2.5 m x 2.5 wide and 0.5 m height showing on
 168 the left side a homogeneous frozen soil and on the right side the same frozen soil (saturated fine sand)
 169 with additional artificial ice wedges. The thick vertical brown line on the top of each photo separates left
 170 from right part. Active erosion due to the presence of ice wedges is observed on the right side of the
 171 experiment. White cables correspond to temperature sensors.

172
 173 Figure 2 presents the case of RTS development in an ice-wedge setting with pure ice
 174 blocks of 5 cm thick. In experiment n°2, we added horizontal heterogeneities (pure icy layers)
 175 instead of ice-wedges (Figure SI2). In that experiment, we artificially built icy layers of 10 cm in
 176 thickness and set them up in the right part of the experiment in the cold room (Figure SI2). We
 177 used the same protocol as for the experiment n°1 for preparing the frozen soil. The left side of the
 178 experiment corresponds to the reference model with saturated and homogeneous frozen ground.

179 All the models were instrumented using ten temperature sensors (platinum resistance
180 thermometers Pt100 with $\pm 0.1^\circ\text{C}$ accuracy) to survey the freezing and thawing front vs. time. A
181 first series of 5 sensors, spaced every 5 cm, was placed in the homogeneous frozen soil at a depth
182 of 20 cm. A similar series of sensors was placed in the frozen soil with ice heterogeneities
183 (Figures SI1 and SI2). During each simulation, we also analyzed the development of associated
184 degradation landforms using time-lapse photography with a video camera (Figures SI5 and SI6).
185 This allows later analysis of the repeated observation of permafrost degradation. For each
186 experiment, we compared ice-rich permafrost degradation (with a total of 30 % ice by volume)
187 with a reference model (homogeneous permafrost with 20% ice by volume). In these
188 experiments, we assumed that the scale effect was not a limiting factor. Here, we restricted our
189 approach to the relative importance of the heterogeneities (ice wedges, icy layers) and their
190 thermal influence on permafrost degradation.

191

192 **3 Results and discussion**

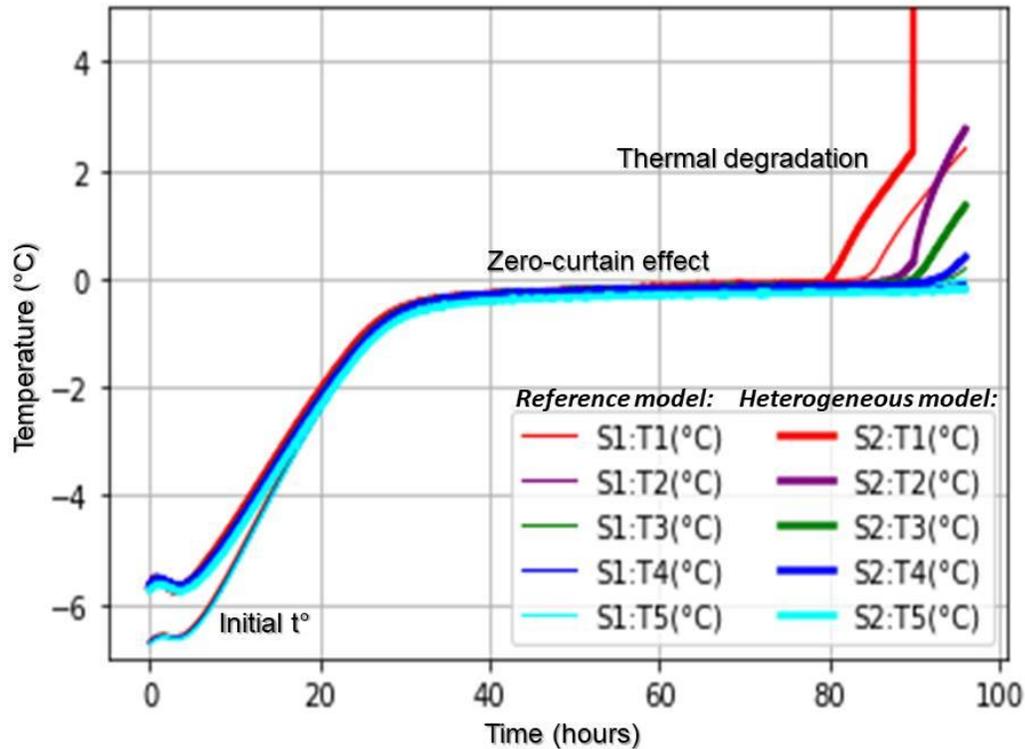
193 Our objective was to test not only different scenarios (different types of heterogeneities)
194 but also different environments. Therefore, we tested permafrost thawing with different initial
195 temperatures (-6°C and -10°C) and different warming conditions (slight and severe). Figure
196 3 shows the typical case of a slight warming, whereas Figure 4 presents a more severe scenario
197 (i.e. with a steeper warming phase).

198 For experiments n°1 and n°2, we started to analyze the data recording when all
199 temperatures in the frozen ground were stabilized at around -6°C and -10°C respectively. We
200 performed a systematic measurement of the thermal wave propagation from the ten Pt100 sensors
201 during the warming phase (Datasets SI3 and SI4). Three main stages within that warming phase
202 were recorded (Figures 3 and 4): (i) The first stage corresponds to a rapid increase in the
203 temperature monitored for all sensors, within approximately 24 hours, with all sensors reaching
204 0°C . (ii) The second stage corresponds to a period of relatively stable temperatures around 0°C
205 due to the zero-curtain effect (latent heat release). A considerable amount of heat has to warm the
206 ground up to the melting point before ablation can proceed, in agreement with recent
207 investigations on temperature-dependent RTS development (Zwieback et al., 2018). (iii) The
208 third stage starts when sensors record positive temperatures.

209 In experiment n°1, the reference model (left side of the experiment on Figure 2) showed a
210 relatively slow rise of temperatures (Figure 3). On the contrary, the heterogeneous frozen soil
211 (right side of the experiment on Figure 2) showed a rapid increase of soil temperature, notably
212 faster than the reference model (Figure 3). From a morphological point of view (Figures 2 and
213 SI5), the homogeneous block (reference model) remains stable (without any deformation, even
214 after thawing during the warming phase), while the heterogeneous one shows dramatic evolutions
215 (slumping, subsidence). In experiment n°2 with horizontal heterogeneities (icy layers), again the
216 reference model (homogenous frozen soil) remained mechanically stable without subsidence,
217 even after complete thawing (Figure SI6). On the contrary, the heterogeneous frozen soil with icy
218 layers showed a clear subsidence due to the melting of the excess ice. From a thermal point of
219 view, we can observe a relatively faster warming of the heterogeneous frozen soil and a relatively
220 colder temperature for the reference model (Figure 4). The analysis of the time-lapse
221 photography (Figure SI5) clearly shows the evolution of the individual blocks of permafrost in
222 between ice wedges (i.e., modeled polygon centers) towards a hummocky morphology similar to
223 the development of high-centered polygons (i.e., baydjarakhs). Our experiments clearly show the
224 detachment and vertical subsidence of individual blocks along ice wedges and the subsequent
225 formation of ‘mud’ flow at the basal slope due to the basal concentration of oversaturated
226 sediments (Figure SI5), in agreement with previous laboratory work focused on rainfall-induced
227 slope failures (Tohari et al., 2007).

228

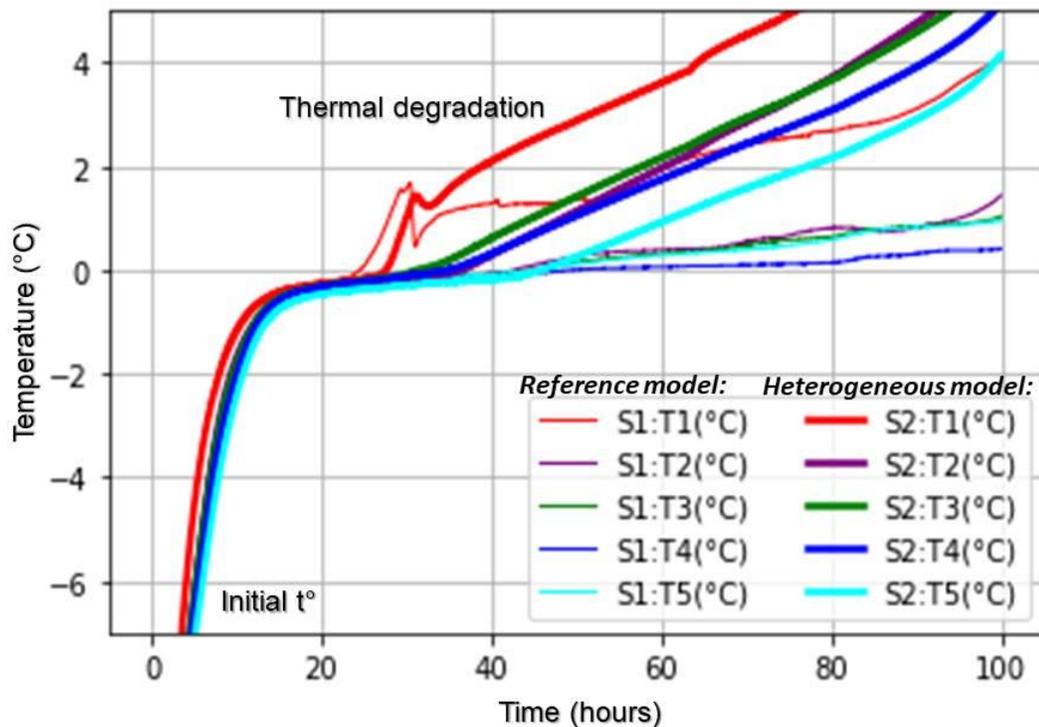
229



230
 231 Figure 3: Experimental setup n°1: evolution of temperature in homogeneous frozen soil (thin
 232 lines: S1) and heterogeneous frozen soil with artificial ice wedges (bold lines: S2). A first series of 5
 233 sensors (S1:T1 to S1:T5), spaced every 5 cm, was placed in the homogeneous frozen soil at a depth of 20
 234 cm. A similar series of sensors was placed in permafrost with ice heterogeneities (S2:T1 to S2:T5). In both
 235 blocks, the temperature is similar in the first heating stage and in the following zero-curtain effect stage.
 236 After melting, greater temperatures are observed in the heterogeneous permafrost compared to the
 237 homogeneous one.

238
 239 In experiment n°1, the efficient warming in heterogeneous soil can be related to the
 240 melting of ice wedges (Figure 2) and the formation of associated voids along the ice wedges that
 241 allow for a better heat diffusion within the thawing soil. In fact, during the thawing phase, the
 242 progressive melting of ice wedges (excess ice) induces a larger heat capacity of liquid water that
 243 enhances temperature increase of the soil. All the experiments showed that ice wedges increase
 244 the preferential erosion during the warming phase and supply meltwater to the thawing frozen
 245 interface, which accelerates basal slumping. These observations are in agreement with field
 246 observations of RTS from Lantuit et al. (2012) and Séjourné et al. (2015). For experiment n°2,
 247 the melting of the icy layers induces a loss of cohesion of the overlapping frozen soil (Figure SI6:

248 video material) and a better warm air circulation within the soil. These results indicate that
 249 discontinuities (ice wedges, icy layers) in frozen soils are much more prone to intense thermal
 250 degradation (warming and thermokarst development) compared to homogeneous and ice-poor
 251 permafrost. Our laboratory results are in agreement with recent investigations along the Batagay
 252 thaw slump in Yakutia, where the presence of a basal horizontal ice layer of merged ice wedges
 253 is the precondition for its deep incision, while the continuous presence of large ice wedges close
 254 to the surface seems to be decisive for its large areal extend and exponential growth (Opel et al.,
 255 2019).
 256



257
 258 Figure 4: Experimental setup n°2: evolution of temperatures of modelled heterogeneous
 259 permafrost (bold lines: S2) with horizontal heterogeneities (icy layers). Thin lines (S1) correspond to the
 260 homogeneous frozen soil (reference model).
 261
 262

263 **4 Conclusions**

264 The underpinning question of our study was to evaluate the main controlling factors of the
265 formation of retrogressive thaw slumps. Cryogeological (granulometry, ice-content and type) and
266 climatological conditions differ greatly across regions where RTS can be observed. We simulated
267 two major ground-ice settings (ice wedges and buried massive ice) compared to a reference
268 model (low ice content). Therefore, this study is representative of three typical cryolithological
269 types of permafrost, which can be found at one site together (Batagay megaslump in Central
270 Yakutia). Our study allows a normalized view on RTS development assuming equal thermal,
271 volumetric ground-ice and grain size conditions and contributes to a better understanding of the
272 relative contribution of heterogeneities, during permafrost degradation processes. Our laboratory
273 simulations attest the efficiency of heterogeneities that control thermokarst and RTS development
274 by two main mechanisms: (i) the strong decohesion of its structure during the thawing phase and
275 an easier circulation of warm air (Figures 3 and 4), and (ii) the excess of water from the
276 preferential melting in ice-rich permafrost, which increases the slumping effect (Figures SI5 and
277 SI6).

278 We calculated the onset of degradation for each experimental setup using the duration of the
279 zero-curtain effect. The moderate initial temperature and the slight warming phase of model 1
280 (ice wedges, Figure 3) induce a longer onset of degradation than the one for model 2 (icy layers
281 under a severe warming condition, Figure 4). The ice-wedge type needs more time before
282 degradation, but then this degradation occurs more rapidly compared to model 2 (icy layers). Our
283 experimental approach suggests that the ice-wedge setting strongly increases the efficiency of
284 RTS development.

285 This study provides a better understanding of how ground-ice heterogeneities influence
286 RTS development, hence serving as a point of reference for future in situ studies of periglacial
287 geomorphic processes and their impacts on the carbon cycle in the Arctic.

288

289 **Acknowledgments**

290 Authors are funded by the Labex IPSL, the GDR2012 *Arctique : Enjeux pour*
291 *l'Environnement et les Sociétés* and the Agence Nationale de la Recherche (ANR) through the
292 Make Our Planet Great Again (MOPGA) initiative (Programme d'investissements d'avenir –
293 project No. ANR-17-MPGA-0014). Assistance of the Melnikov Permafrost Institute of Yakutsk

294 for field studies is gratefully acknowledged. All data from our cold chamber at Orsay (Université
295 Paris-Saclay) are presented in the supporting information S1 to S5. These data are also available
296 on the repository PANGAEA Data Archiving & Publication:
297 <https://doi.pangaea.de/10.1594/PANGAEA.921498>. We thank the Editor, M. Fritz and an
298 anonymous reviewer for their constructive comments and questions, which substantially
299 improved the quality of our manuscript.

300

301 **Authors contributions:**

302 F.C. conceived, designed and carried out the cold-room experiment. A.S. and F.B.
303 developed the hypothesis and its implications. L.D. created and performed calculations of the
304 thermal database, and B.S.B. and A.F. guided the research effort and discussions. The manuscript
305 was collectively written by F.C., A.S. and F.B. All authors provided input on the manuscript and
306 the broader implications of this work.

307

308 **References**

309 Burn C. R. & A.G. Lewkowicz (1990), Retrogressive thaw slumps. *The Canadian*
310 *Geographer*, 34: 273–276. doi.org/10.1111/j.1541-0064.1990.tb01092.x

311

312 Costard, F., L. Dupeyrat, E. Gautier & E. Carey-Gailhardis (2003), Fluvial thermal
313 erosion investigations along a rapidly eroding river bank: application to the Lena river (central
314 Yakutia). *Earth Surface Processes and Landforms*, 28, 1349-1359. doi.org/10.1002/esp.592

315

316 Fedorov, A.N. & P. Y. Konstantinov (2009), Response of permafrost landscapes of
317 central Yakutia to current changes of climate, and anthropogenic impacts. *Geograph. Nat.*
318 *Resour.*, 30. 146-150. doi.org/10.1016/j.gnr.2009.06.010

319

320 Fedorov, A.N., A. Ivanova, R. Park, H. Hiyama, & Y. Iijima (2014), Recent air
321 temperature changes in the permafrost landscapes on northeastern Eurasia. *Polar Science*, 8(2),
322 114-128. doi.org/10.1016/j.polar.2014.02.001

323

324 French, H. M. (2017), *The Periglacial Environment*, 4th ed. Wiley. ISBN: 978-1-119-
325 13278-3

326
327 Grosse, G., J. Harden, M. Turetsky, A. D. McGuire, Ph. Camill, C. Tarnocai, S. Frolking,
328 E. A. G. Schuur, T. Jorgenson, S. Marchenko, V. Romanovsky, K. P. Wickland, N. French, M.
329 Waldrop, L. Bourgeau-Chavez. and R. G. Striegl (2011), Vulnerability of high-latitude soil
330 organic carbon in North America to disturbance. *J. Geophys. Res.*, 116, G00K06,
331 doi:10.1029/2010JG001507

332
333 IPCC, (2019), *IPCC Special Report on the Ocean and Cryosphere in a Changing Climate*
334 [H.-O. Pörtner, et al. (eds.)].

335
336 Kizyakov A.I., M. V. Zimin, M. O. Leibman, and N.V. Pravikova (2013), Monitoring of
337 the rate of thermal denudation and thermal abrasion on the western coast of Kolguev Island,
338 using high resolution satellite images. *Kriosfera Zemli*, 17(4), 36-47. hdl:10013/epic.49210.d001

339
340 Kokelj, S.V. & M. T. Jorgenson (2013), Advances in thermokarst research. *PPP*, vol. 24
341 issue 2, pp. 108-119. doi.org/10.1002/ppp.1779

342
343 Kokelj, S.V., T. C. Lantz, J. Tunnicliffe, R. Segal & D. Lacelle (2017), Climate-driven
344 thaw of permafrost preserved glacial landscapes, northwestern Canada. *Geology*. V. 45, No. 2.
345 doi: 10.1130/G38626.1

346
347 Lacelle D., J. Bjornson & B. Lauriol (2010), Climatic and geomorphic factors affecting
348 contemporary (1950–2004) activity of retrogressive thaw slumps on the Aklavik Plateau,
349 Richardson Mountains, NWT, Canada. *Permafrost and Periglacial Processes*, 21: 1–15. doi :
350 10.1002/ppp.666

351
352 Lantuit, H. & W. H. Pollard (2008), Fifty years of coastal erosion and retrogressive thaw
353 slump activity on Herschel Island, Southern Beaufort Sea, Yukon Territory, Canada.
354 *Geomorphology*, 95 (1). doi: 10.1016/j.geomorph.2006.07.040

355
356 Lantuit, H., W. H. Pollard, N. Couture, M. Fritz, L. Schirrmeister, H. Meyer and H.-
357 W. Hubberten (2012), Modern and Late Holocene Retrogressive Thaw Slump Activity on
358 the Yukon Coastal Plain and Herschel Island, Yukon Territory, Canada. *PPP*, 23, 39-51.
359 doi.org/10.1002/ppp.1731

360
361 Lantz T. C. & S. V. Kokelj (2008), Increasing rates of retrogressive thaw slump activity
362 in the Mackenzie Delta region, N.W.T., Canada. *GRL*, 35: L06502. doi:10.1029/2007GL032433

363
364 Leibman, M., A. Gubarkov, A. Khomutov, A. Kizyakov, & B. Vanshtein (2008), Coastal
365 processes at the tabular-ground-ice-bearing area, Yugorsky Peninsula, Russia, in: Kane, D.L. and
366 Hinkel, K.M. (eds), *Proceedings of the Ninth International Conference on Permafrost*, University
367 of Alaska Fairbanks, June 29-July 3 2008, 1037-1042. hdl.handle.net/10013/epic.45563.d001

368
369 Lewkowicz A. G. & R. G. Way (2019), Extremes of summer climate trigger thousands of
370 thermokarst landslides in a High Arctic environment. *Nat. Commun.* 10 1329.
371 doi.org/10.1038/s41467-019-09314-7

372
373 Murton, J., M. Edwards, A. Lozhkin, P. Anderson, G. Savvinov, N. Bakulina, ... & O.
374 Zanina (2017), Preliminary paleoenvironmental analysis of permafrost deposits at Batagaika
375 megaslump, Yana Uplands, northeast Siberia. *Quat. Res.* 87, issue 2, 314–330.
376 doi.org/10.1017/qua.2016.15

377
378 Nitze, I., G. Grosse, B. M. Jone, V. E. Romanovsky, & J. Boike (2018), Remote sensing
379 quantifies widespread abundance of permafrost region disturbances across the Arctic and
380 Subarctic. *Nature Communications*, Vol. 9, 5423. doi.org/10.1038/s41467-018-07663-3

381
382 Obu, A., H. Lantuit, M. Fritz, Wayne H. Pollard, T.Sachs & F.Günther (2016), Relation
383 between planimetric and volumetric measurements of permafrost coast erosion: a case study from
384 Herschel Island, western Canadian Arctic, *Polar Res.*, 35:1, DOI: 10.3402/polar.v35.30313

385

386 Olefeldt, D., S. Goswami, G. Grosse, D. Hayes, G. Hugelius, P. Kuhry, A. D. McGuire,
387 V. E. Romanovsky, A. B. K. Sannel, E. A. G. Schuur, & M. R. Turetsky (2016), Circumpolar
388 distribution and carbon storage of thermokarst landscapes. *Nature Communications*, vol. 7,
389 13043. doi.org/10.1038/ncomms13043

390
391 Opel, T., J. B. Murton, S. Wetterich, H. Meyer, K. Ashastina, H. Günther, F. Grotheer,
392 G. Mollenhauer, P. P. Danilov, V. Boeskorov, G. N. Savvinov, & L. Schirrmeister (2019), Past
393 climate and continentality inferred from ice wedges at Batagay megaslump in the Northern
394 Hemisphere's most continental region, Yana Highlands, interior Yakutia. *Clim. Past*, 15, 1443–
395 1461, doi.org/10.5194/cp-15-1443-2019

396
397 Romanovsky, V. E., S. L. Smith, & H. H. Christiansen (2010), Permafrost Thermal State
398 in the Polar Northern Hemisphere during the International Polar Year 2007–2009: a Synthesis.
399 *PPP*, 21, 106-116. doi.org/10.1002/ppp.689

400
401 Rudy, A., C. A. Lamoureux, S. F. Kokelj, S. V. Smith, I. R., & J. H. England (2017),
402 Accelerating thermokarst transforms ice-cored terrain triggering a downstream cascade to the
403 ocean. *GRL* Vol. 44, No 21, pp. 11,080-11,087. doi.org/10.1002/2017GL074912

404
405 Schuur, E. A. G., J. Bockheim, J. G. Canadell, E. Euskirchen, C. B. Field, S. V.
406 Goryachkin, S. Hagemann, P. Kuhry, P. M. Lafleur, H. Lee, G. Mazhitova, F. E. Nelson, A.
407 Rinke, V. E. Romanovsky, N. Shiklomanov, C. Tarnocai, S. Venevsky, J. G. Vogel, & S. A.
408 Zimov (2008), Vulnerability of permafrost carbon to climate change: Implications for the global
409 carbon cycle. *BioScience*, 58(8), 701– 714. doi.org/10.1641/B580807

410
411 Séjourné, A., F. Costard, A. Fedorov, J. Gargani, J. Skorve, & M. Massé (2015),
412 Evolution of the banks of thermokarst lakes in central yakutia due to retrogressive thaw slump
413 activity controlled by insolation. *Geomorphology* 241, 31–40.
414 doi.org/10.1016/j.geomorph.2015.03.033

415

416 Schirrmeister L., E. Dietze, H. Matthes, G. Gross, J. Strauss, S. Laboor, M. Ulrich, F.
417 Kienast, & S. Wetterich (2020), The genesis of Yedoma Ice Complex permafrost – grain-size
418 endmember modeling analysis from Siberia and Alaska. *E&G Quaternary Sci. J.*, 69, 33–53,
419 2020. doi.org/10.5194/egqsj-69-33-2020

420
421 Segal, R. A., Lantz T. C., & S. V. Kokelj (2016), Acceleration of thaw slump activity in
422 glaciated landscapes of the Western Canadian Arctic. *Environmental Research Letters*, Vol.
423 11, Number 3. doi.org/10.1088/1748-9326/11/3/034025

424
425
426 Shakil, S., S.E. Tank, S.V. Kokelj, J.E. Vonk, and S. Zolkos (2020). Particulate
427 dominance of organic carbon mobilization from thaw slumps on the Peel Plateau, NT:
428 Quantification and implications for stream systems and permafrost carbon release. *Environmental*
429 *Research Letters*. doi.org/10.1088/1748-9326/abac36.

430
431 Shepelev A.G., A. Kizyakov, S. Wetterich, A. Cherepanova, A. Fedorov, I.
432 Syromyatnikov & G. Savvinov (2020), Sub-surface carbon stocks in northern taiga landscapes
433 exposed in the Betagay Megaslump, Yanal upland, Yakutia. *Land*, 9, 305.
434 doi:10.3390/land9090305

435
436 Soloviev, P. A., (1973), Thermokarst phenomena and landforms due to frost heaving in
437 central Yakutia. *Biuletyn Peryglacjalny*, 23, 135–155.

438
439 Strauss, L., J. Schirrmeister, G. Grosse, D., Fortier, G. Hugelius, Ch. Knoblauch, V.
440 Romanovsky, Ch. Schädel, Th. Schneider Von Deimling, E. A. G. Schuur, D. Shmelev, M.
441 Ulrich & A. Veremeeva (2017), Deep Yedoma permafrost: A synthesis of depositional
442 characteristics and carbon vulnerability. *Earth Science Reviews*, vol. 172, 75-86.
443 doi.org/10.1016/j.earscirev.2017.07.007

444

445 Tohari A., M. Nishigaki, & M. Komatsu (2007), Laboratory rainfall-induced slope failure
446 with moisture content measurement. *J. Geotechnical and Geoenvironmental Engineering*, Vol. 3,
447 issue 5, 575. doi.org/10.1061/(ASCE)1090-0241(2007)133:5(575)

448
449 Turetsky, M. R., B. W. Abbott, M. C. Jones, et al. (2020), Carbon release through abrupt
450 permafrost thaw. *Nat. Geosci.* 13, 138–143. doi.org/10.1038/s41561-019-0526-0

451
452 Ulrich M, G. Grosse, J. Strauss J. & L Schirrmeister (2014), Quantifying Wedge-Ice
453 Volumes in Yedoma and Thermokarst Basin Deposits. *PPP*. Vol. 25, 151-161.
454 doi.org/10.1002/ppp.1810

455
456 Vonk, J.E., S.E. Tank, W.B. Bowden, I. Laurion, W.F. Vincent, P. Alekseychik, M.
457 Amyot, M.F. Billet, J. Canário, R.M. Cory, B.N. Deshpande, M. Helbig, M. Jammet, J. Karlsson,
458 J. Larouche, G. MacMillan, M. Rautio, K.M. Walter Anthony, and K.P. Wickland (2015).
459 Reviews and Syntheses: Effects of permafrost thaw on Arctic aquatic ecosystems.
460 *Biogeosciences*. 12: 7129-7167. doi:10.5194/bg-12-7129-2015.

461
462 Ward Jones M. K., W. H. Pollard & B. M. Jones (2019), Rapid initialization of
463 retrogressive thaw slumps in the Canadian high Arctic and their response to climate and terrain
464 factors. *Environmental Research Letters*, Vol. 14, Number 5. doi:10.1088/1748-9326/ab12fd

465
466 Wolfe, S., E. Kotler, & S. Dallimore (2001), Surficial characteristics and the distribution
467 of thaw landforms (1970-1999), shingle point to kay point, Yukon Territory. Geological Survey of
468 Canada Open File 4115. hdl.handle.net/10.4095/212842

469
470 Zwieback, S., S. V. Kokelj, F. Günther, J. Boike, J. Grosse, & I. Hajnsek (2018), Sub-
471 seasonal thaw slump mass wasting is not consistently energy-limited at the landscape scale, *The*
472 *Cryosphere*, 12, 549-564. doi.org/10.5194/tc-12-549-2018

473