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Evidence of frost-cracking inferred from acoustic emissions in a high-alpine rock-wall

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Abstract

Ice formation within rock is known to be an important driver of near-surface frost weathering as well as of rock damage at the depth of several meters, which may play a crucial role for the slow preconditioning of rock fall in steep permafrost areas. This letter reports results from an experiment where acoustic emission monitoring was used to investigate rock damage in a high-alpine rock-wall induced by natural thermal cycling and freezing/thawing. The analysis of the large catalog of events obtained shows (i) robust power-law distributions in the time and energy domains, a footprint of rock micro-fracturing activity induced by stresses arising from thermal variations and associated freezing/thawing of rock; (ii) an increase in AE activity under sub-zero rock-temperatures, suggesting the importance of freezing-induced stresses. AE activity further increases in locations of the rock-wall that are prone to receiving melt water. These results suggest that the framework of further modeling studies (theoretical and numerical) should include damage, elastic interaction and poro-mechanics in order to describe freezing-related stresses.

Keywords: frost-cracking, acoustic emission, mechanical weathering, scaling properties

1. Introduction

2 The formation of ice within rock is likely to be an important driver of near-
3 surface frost weathering (Hallet et al., 1991) and rock damage at the depth of

4 several meters (Murton et al., 2006), and in steep terrain, this process may be
5 crucial for the slow preconditioning of rock fall from warming permafrost areas
6 (Gruber and Haeberli, 2007). However, the transfer of corresponding theoretical
7 insight and laboratory evidence to natural conditions characterized by strong
8 spatial and temporal heterogeneity of the rock properties (e.g. fracture state,
9 water content, thermal and hydraulic conductivity) and thermal conditions is
10 nontrivial. To examine rock fracture in natural conditions, we performed a
11 pilot experiment, monitoring acoustic emissions (AE) in a high-altitude rock-
12 face during a four days period. In such conditions, the mechanical loading
13 of rock results from the combination of a constant gravity load and fluctuating
14 loads related to (i) thermal stresses, arising from the gradient of the temperature
15 field, (ii) pressure variations in rock pores and cracks, due to water or to ice
16 formation and (iii) short-term external loading such as earthquakes. While large
17 thermal stresses can only occur close to the rock surface, ice formation in pores,
18 cracks and fractures can potentially generate large stresses at greater depths, as
19 suggested by theoretical and lab studies. Reporting a preliminary analysis of the
20 microseismic activity monitored at a high alpine ridge, Amitrano et al. (2010)
21 recently stressed the importance of ice formation in fractures as they observed
22 micro-seismic activity corresponding with particular trends of the temperature
23 that could enhance ice formation. But the lack of details in the spatial and
24 temporal distribution of the seismic events precluded the full understanding of
25 the relationship between temperature evolution and related ice formation at
26 small spatial scale and the triggering of seismic events.

27 The mechanical loading of rocks involves local inelastic processes that pro-
28 duce elastic wave propagation so called acoustic emission (AE) at small scales
29 and micro-seismicity (MS) at larger scales. Beside the common physical origin
30 of the elastic wave emission, essentially induced by the propagation or shearing
31 of cracks, these two terms denote differences in the frequency content of the
32 recorded signals corresponding to sources of different size (see Hardy (2003) for
33 a full presentation). MS relates to the range $1-10^3$ Hz whereas AE relates to
34 the range 10^4-10^6 Hz. The corresponding source size is $1-10^3$ m for MS and

35 10^{-3} - 10^{-1} m for AE. The material attenuation, that increases with frequency,
36 precludes the detection of AE after approximately 1 m of wave propagation,
37 whereas MS can be detected at larger distances (up to km).

38 Measuring AE or MS activity therefore provides a powerful technique to
39 monitor the evolution damage at different scales. Due to their wide frequency
40 range, the simultaneous recording of AE and MS currently is technically not
41 possible. The AE has been extensively used as a tool at the laboratory rock
42 sample scale [e.g. Lockner (1993)] whereas MS has been mostly used at larger
43 scales in order to study seismicity and rockburst in mines, tunnels or quarries
44 (Hardy, 2003). In all these cases, AE/MS are considered to be an indicator of
45 inelastic behavior that can be related to damage increase or to shearing of exist-
46 ing fractures (Cox and Meredith, 1993; Lockner, 1993). Several recent studies
47 report MS monitoring of slope instability (Amitrano et al., 2010; Gaffet et al.,
48 2010). The originality of our study is to apply high frequency AE monitoring,
49 a technique traditionally used in laboratory experiments, to investigate rock
50 damage during freezing, in field conditions. The main advantage of using high
51 frequency monitoring is the sensitivity to emissions of relatively small energies.
52 This allows us to obtain a large catalog of events within a short monitoring
53 period, which is crucial to perform statistical analyses. At such high frequen-
54 cies, acoustic signals are attenuated within about a meter, which determines the
55 spatial scale of our study. This is an advantage as most of the acoustic activity
56 related to freezing can be expected to occur within a meter from the surface.
57 This technique finally offers a high temporal resolution, as event rates up to 10^3
58 per second can be detected.

59 Fracturing dynamics during mechanical loading, usually displays scaling
60 properties in the domains of size, space and time (Alava et al., 2006; Sethna
61 et al., 2001). In the domain of size (magnitude) for example, the seismic events
62 induced by damage processes display a power-law (PL) distribution, $N(s) \sim s^{-b}$,
63 where s is an estimate of the event size (e.g. the maximum amplitude of the AE
64 signal or its energy), $N(s)$ is the probability distribution function (PDF) and
65 b is a constant. This distribution is equivalent to the well-known Gutenberg-

66 Richter relationship observed for earthquakes (Gutenberg and Richter, 1954).
67 Scaling properties in space and time of the events have also been reported, char-
68 acterizing their spatial and temporal clustering. The emergence of these scaling
69 properties is considered to be a universal feature of the damage dynamics in
70 heterogeneous media (Alava et al., 2006) as it is observed in a very robust man-
71 ner for various loading conditions, various materials and scales ranging from
72 the micrometer (microcracks) to thousands of kilometers (the Earth’s-crust or
73 the sea ice cover). In this letter we report an original in-situ experiment of
74 AE monitoring in high altitude thermal conditions. We show that AE activity
75 resulting from natural thermal cycling and induced freezing/thawing shows sim-
76 ilar scaling properties, suggesting that the local stress fluctuations encountered
77 are high enough to induce micro-fracturing.

78 **2. Measurement site and instrumental setting**

79 The measurement site is a south-facing cliff of granitic gneiss (Wegmann and
80 Keusen, 1998) that is situated at an elevation of 3500 m a.s.l. in the Swiss Alps.
81 The local mean annual air temperature is about -7.3°C (1961–1990), whereas
82 mean annual rock temperatures near the surface are between -2 and -3°C
83 in this south face (Hasler et al., 2011). The site is directly next to the high-
84 altitude research station Jungfrauoch and thus can be measured with standard
85 AE equipment housed inside a heated building.

86 A six-channel high-frequency acquisition board (Mistras, Euro Physical Acous-
87 tic) with 16-bit resolution and 10 V maximum amplitude was used. The AE
88 piezo-electrical sensors (EPA R6I) had an operating frequency range of 10–
89 150 kHz with a peak sensitivity at 60 kHz. They included a pre-amplifier of
90 40 dB and were connected with 20 meters coaxial cables. The system was con-
91 tinuously sampling with a frequency of 1 MHz and basic signal characteristics
92 such as time, amplitude, energy, duration, spectral content were calculated in
93 real time for events over 35 dB amplitude. The maximum noise amplitude has
94 been measured to be near 1 mV (30 dB), indicating that recorded events were

95 not induced by noise fluctuations. Full waveforms of 2 ms including a pre-trigger
96 of 0.4 ms were recorded for events over 40 dB (4 mV). Using an ultrasonic cou-
97 pling gel (Sofranel), sensors were pressed on a steel plate with rubber bands.
98 We verified in the lab that the coupling gel behaves similarly for various tem-
99 perature and do not generate AE when freezing (Weber et al., 2012). The steel
100 plates were screwed onto extension bolts (10 mm diameter) anchored about 5 cm
101 deep in the rock (Figure 1A). Each sensor installation was protected from water
102 with a plastic sleeve (Figure 1B). We verified that the crumpling of the plastic
103 cover due to wind or other factors do not cause AE. This has been tested during
104 the installation by moving by hand the plastic cover and no AE were generated.
105 Moreover, the windy periods we observed during the experiment have not been
106 recognized to generate AE.

107 Figure 2 shows an example of AE trace recorded at channel 1, with the trigger
108 level, the maximal and minimal amplitude and the method for calculating the
109 signal energy.

110 The site shows some heterogeneities in terms of the spatial distribution of
111 fractures (coexistence of compact and fractured zones), microtopography (con-
112 vex and concave zones) and hydraulic conditions (dry and wet zones). Sensor
113 locations (Figure 1C), referred to as AE1–AE6, have been chosen to investigate
114 different configurations of fractured or compact rock and wet or dry conditions.
115 AE1 is installed in compact and homogeneous rock that is barely fractured in
116 a radius of about one meter. The rock surface bulges out slightly in this area.
117 It has been uncovered from snow one hour before installation, with some snow
118 remaining about 30 cm below the sensor. AE2 has compact, lightly fractured
119 rock that is slightly concave outward and subject to melt water flow from above.
120 AE3 is located in fractured, concave but rather dry rock. AE4 is in fractured
121 rock that is convex and apparently dry in surface. But a gully containing snow
122 is located 0.5 m above it that could supply water in the vicinity. AE5 is installed
123 in compact rock in an overhang underneath a gully and could receive melt water
124 through fractures. AE6 is installed in compact rock in a gully receiving melt
125 water from above. AE5 and AE6 are located near a fracture zone in the deepest

126 part of the gully; the rock mass to their left is slightly overhanging and may
127 be prone to movement. AE1–AE4 were installed on 6 April, AE5 and AE6 on
128 7 April 2010. All sensors were uninstalled on 10 April 2010. The Permasense
129 measurement site (Hasler et al., 2008; Beutel et al., 2009) at the same location
130 provides rock temperature data, measured close to sensor AE3, at depths of 10,
131 35, 60 and 85 cm (Hasler et al., 2011).

132 **3. Results**

133 The system has been operated continuously for four days for AE1–AE4 and
134 three days for AE5–AE6. Air temperatures fluctuated between -2 and -10°C
135 during this time. There were virtually no clouds and radiative diurnal cycles
136 caused near-surface rock temperatures to rise to 10°C during the day and cool
137 to -5°C during nights. Diurnal thawing penetrated about 20 cm deep into the
138 rock wall which remained continuously frozen at greater depths. Due to snowfall
139 in the days before, small snow patches in concave portions of the rock wall locally
140 provided melt water flow during the day. Several thousand events were recorded
141 at each sensor, allowing a robust statistical analysis (Table 1). Because of the
142 large sensor spacing, attenuation precluded the detection of individual events
143 on several sensors. Given the frequency range at which the sensors operate,
144 the detected events can be expected to have their source within the meter scale
145 around the sensor. Since event source localization is impossible, detected events
146 are thus considered to be close to the receiving sensor.

147 Figure 3 reports the time evolution of AE activity and temperature of the
148 rock at 10, 35 and 60 cm depth as a function of the hour of the day. It high-
149 lights the connection between the daily fluctuations of AE activity and that
150 of rock temperature during the monitoring period. For the entire duration of
151 the experiment, the temperature at 35 cm depth remains below 0°C whereas it
152 remains under 0°C at 60 cm depth. The temperature at 10 cm depth exhibits
153 freeze-thaw cycles with temperature ranging from -5°C to 10°C . At this depth,
154 the refreezing can be seen by the plateau characterizing the temperature near

155 zero in the evening (i.e. at $h \simeq 20$ on days 1–2) when temperature decreases
156 more slowly due to latent heat of water. This time period corresponds to a
157 small peak of the AE activity visible on AE1, AE3, AE5 and AE6. We recall
158 that our analysis is based on a single point measurement of rock temperature.
159 The spatial variability of temperature at different locations could thus partly
160 explain the shift seen in AE activity observed for the different sensors.

161 The AE activity appears to be significantly larger when the temperature
162 is below zero. The largest AE activity takes place during colder (night-time)
163 periods, when the near-surface of the rock refreezes. This is observed for all
164 the sensors with the notable exception of AE5. Sensors AE1, AE3 and AE5
165 show activity peaks significantly smaller than AE2, AE4 and AE6. AE1 and
166 AE3 are located in dry areas whereas AE2, and AE6 are located in wet areas
167 with melting water coming from upslope. AE4 was located in an apparently
168 dry area but the presence of snow 0.5 m above could provide melting water that
169 was not visible on the surface. In contrast to the other sites, any meltwater
170 reaching AE5 would have to percolate through fractures. No water comes from
171 the surface due to its situation under an overhang. These observations express
172 the large spatial variability of the near-surface AE activity. They also suggest
173 that the availability of water has a strong control on AE activity. In order
174 to verify the dependence between AE activity and temperature we calculate
175 the distribution of events as a function of the temperature (Fig. 4 left). The
176 relationship between negative temperatures and AE activity appears clearly for
177 all the sensors except AE5. Between 70% and 95% of the events are recorded
178 during negative temperature periods, whereas for AE5, this proportion is only of
179 37% (see Tab. 1 for details). One may note also that a slight AE peak is visible
180 around temperature zero. The difference in the amplitude of the AE activity
181 between dry and wet areas is confirmed. In order to take into account the fact
182 that much more time is spent in each temperature increment below zero than
183 above, we normalized the event number by the time spent into each temperature
184 interval (Fig. 4 right). the temperature dependence appears stronger for sensors
185 AE2 AE4 and AE6, and weaker for sensors AE1 and AE3. The particularity of

186 the sensor AE5 appears stronger.

187 We analyzed the scaling properties of the AE focusing on the domains of
188 energy and time and described other striking behavior of the measured data.
189 For each sensor we calculated the distribution of event energy for the entire
190 duration of the experiments. The PDF's of event energy (Fig. 5) show a power-
191 law (PL) behavior spanning several orders of magnitude with an exponent $b =$
192 1.55 ± 0.05 . For E smaller than 100, the pdf shows a clear departure from PL
193 trend characterized by a decrease of the slope toward the smallest events. This
194 effect is commonly related to the completeness of the catalog (see e.g. Wiemer
195 and Wyss, 2000). Under this value, due to the signal attenuation, the smallest
196 events are detected only on the vicinity of the sensor and so the sampling of these
197 events is incomplete. On the contrary, above this value there is no statistical
198 sampling bias on the PL trend can be estimated with confidence.

199 We verified that the completeness and the PL trend remain unchanged when
200 selecting the events recorded during positive or negative temperature periods.
201 We also verified the effect of selecting only events of energy larger than the
202 completeness. The trends observed on Figures 3 and 4 remain the same, in
203 particular with respect to the temperature dependence of AE activity. The
204 only notable effect is the reduction of the amplitude of AE peaks.

205 As the PL distribution is characteristic of the rupture processes (Alava et al.,
206 2006), the later observation is the first direct evidence published showing that
207 natural thermal cycling and associated freezing/thawing induce near-surface
208 damage in a rock wall.

209 In order to investigate scaling properties of the temporal distribution of AE
210 events we use the correlation integral (Grassberger and Procaccia, 1983),

$$C(\Delta t) = 2\mathcal{N}(\Delta t)/(N(N-1)) \quad (1)$$

211 where N is the total number of damage events, $\mathcal{N}(\Delta t)$ is the number of pairs
212 of events separated by a time smaller than Δt . This integral expresses how the
213 events are distributed in time. If the correlation integral exhibits a PL $C(\Delta t) \sim$
214 Δt^{D_2} , the population can be considered as a fractal set, i.e. characterized by a

215 scaling invariance in time domain, with a correlation dimension D_2 . A value of
 216 D_2 smaller than 1 indicates a time clustering, i.e. the probability of observing
 217 an event is larger when a previous one has occurred within a short time. In other
 218 words, an event is more likely to occur within a short separating time from the
 219 previous one. Departure from power-law trend or slope changes indicate the
 220 existence of characteristic time scale limiting the extent of scale invariance.

221 Figure 6 shows the correlation integrals obtained for each sensor. At small
 222 time scales, a first PL trend is identified over two to three orders of magnitude
 223 in time, for all sensors except AE5, with a correlation dimension D_2 between
 224 0.75 and 0.9. The value $D_2 < 1$ expresses the temporal clustering of AE events,
 225 i.e. the detected events are strongly correlated over this temporal scale range.
 226 The extent of the PL trend towards small time scales is limited by the duration
 227 of the recording (1 ms) so no event can be detected with a lower separating
 228 time. The upper limit of PL extends to about 0.5 s for AE1 and AE4, 1 s for
 229 AE2 and AE6, and 5 s for sensor AE3. This mean that the events are strongly
 230 correlated in time within this time scale. So this value can be interpreted as
 231 the duration of correlated events series that we may consider as sequences of
 232 cascading events, i.e. the temporal correlation length, i.e. the duration for
 233 which a pair of events is more correlated in time than in a random serie. For
 234 larger separating time the correlation integral displays a rather flat shape until
 235 recovering a secondary PL trend. In this region, AE4 and AE5 are characterized
 236 by a slope $D_2 \simeq 0.9$ indicating a slightly clustered temporal distribution. The
 237 duration of the correlated sequences is about 2-3 hours. For AE3 the secondary
 238 power-law trend appears of poor quality, with a low D_2 value corresponding
 239 to highly clustered events. On the other hand, a clear secondary power law
 240 trend of exponent $D_2 \simeq 0.75$ is observed for AE1, AE2 and AE6, a temporal
 241 correlation dimension similar to the one observed at smaller time scales. This
 242 power law trend extends up to about $\Delta t = 10 - 12$ hours, indicating scaling
 243 properties spreading over the duration of the freezing period.

244 4. Discussion and conclusions

245 The first striking aspect of the measurements presented is the relationship
246 between negative temperatures and AE activity, suggesting that damage is re-
247 lated to freezing-induced stresses. As ice is a better wave transmitter than liquid
248 water one may ask if this observation could be an artefact induced by the lack
249 of detection instead of a lack of AE at temperatures $> 0^\circ$. In order to verify
250 this, we compare the two sets of events distinguishing positive and negative
251 temperature periods. A first evidence is that the completeness is the same for
252 the two sets (see section 3). This indicates that there is no significant changes
253 in the event detection. As the attenuation is known to affect the frequency
254 content (Hardy, 2003), we analyse the evolution of the mean frequency of the
255 events. Surprisingly, we observed a slight increase of the mean frequency for
256 warmer periods (Fig. 7), particularly for AE2, AE5 and AE6. This could be
257 interpreted as an effect of the thermal dilation of cracks. The rock material in
258 between cracks expands inducing a closure of cracks and then a reduction of
259 the attenuation. The fact that variations of the frequency are limited could be
260 explained by the resonant nature of the sensors we used. The cracks closure
261 can also induce AE, but this has been shown to be limited compared to the
262 crack shearing process (e.g. Moradian et al., 2010). We also examine the rela-
263 tionship between the signal amplitude and duration, considering that a higher
264 attenuation should correspond to a shorter duration for a given amplitude. We
265 observed no (or slight) variations regarding the temperature. Consequently, the
266 increase of AE activity in sub-zero temperatures is unlikely do be an artefact in
267 the AE detection.

268 We now discuss the different mechanisms that could induce damage through
269 thermal cycling. A possible mechanism for explaining the relationship between
270 AE and temperature is the differential thermal dilation inducing thermal stress-
271 ing. In the present case, the amplitude of thermal gradients (in space and
272 time) is limited compared to the one needed for inducing damage within in-
273 tact rock, as observed at the laboratory or estimated analytically considering

274 thermo-mechanical coupling (e.g. Fredrich and Wong, 1986; Wai et al., 1982).
275 In the case of already damaged or fractured material, it has been shown that
276 daily thermal cycling can induce shearing along existing fractures (Gunzburger
277 et al., 2005) even in absence of freeze-thaw cycles. The most favorable peri-
278 ods are the ones corresponding to high thermal gradient in time ($\delta T/\delta t$) and
279 in space ($\delta T/\delta z$, z representing the depth). Referring to our experiment, this
280 mechanism could induce AE events when the absolute value of the spatial tem-
281 perature gradient is high, independently of positive or negative temperature.
282 We estimated the temperature gradient in depth as the difference between tem-
283 perature measured at 10 and 35 cm depth, divided by their separating distance
284 $\delta T/\delta z = (T_{35cm} - T_{10cm})/0.25m$. Figure 8 shows that most of the AE activity
285 occurred when $\delta T/\delta z > 20^\circ C/m$, when the rock 10 cm below the surface is
286 much cooler than 35 cm below.

287 Thermo-elastic stresses arise when non-uniform (spatial) gradients in tem-
288 perature develop in elastic materials, which tend to result from rapidly vary-
289 ing temperatures. Note that, somewhat non-intuitively, thermo-elastic stresses
290 would no be expected in unconfined elastic media if the temperature gradient
291 is uniform even if temperatures vary considerably in space. So this mechanism
292 can be evoked only when the temporal gradient is also high.

293 Figure 8 shows the temperature gradient in time at 10 cm depth. The
294 maximum absolute value of the temporal gradient occurs at approximatively 13
295 and 18 hours. These two periods corresponds to AE activity peaks visible on
296 Figures 3 and 8, suggesting that AE could be related to fast contraction/dilation
297 of the rock. As the temperature in depth is stable, the temperature gradient is
298 highly correlated with the shallow temperature (Fig. 8). The highest absolute
299 value of the gradient corresponds to the minimum of the shallow temperature.
300 Consequently it is difficult to distinguish between the impacts of the temperature
301 and of the gradient of temperature on the AE activity.

302 AE activity was shown to be more intense at locations with melt-water,
303 where large bursts of events were recorded during night-time refreezing. This
304 suggests that water freezing plays an important role in the mechanical load-

305 ing. A possible origin for the damage is the volume expansion of freezing water
306 contained in fractures, that is often evoked for explaining cryo-fracturing. The
307 volume change provokes considerable ice pressure in the fractures that can prop-
308 agate and release the induced stresses. This could explain the AE activity peak
309 observed when temperature decreases near zero (Fig. 4). Note that the tempo-
310 ral gradient is very small at this time precluding the effect of fast contraction
311 evoked before. When occurring within in a porous medium, the phase change
312 of water spreads over a temperature range that depends on the pore dimension
313 (Coussy, 2005) and not only for $T \approx 0^\circ C$. This could explain why increased
314 AE is sustained across the entire observed range of negative temperatures and
315 not restricted to temperatures near $0^\circ C$.

316 Another possible mechanism related to ice formation is the cryo-suction or
317 ice segregation (Hallet et al., 1991; Coussy, 2005). Ice in pores or cracks is
318 surrounded by an unfrozen water film due to disjoining (intermolecular) forces.
319 Disjoining forces between the ice and the rock can then cause ice-filled cracks to
320 widen as water is drawn in from the surrounding medium by a free-energy gra-
321 dient. This phenomenon operates at temperatures below zero in a temperature
322 range depending mostly on the pore size distribution, permeability and fracture
323 mechanical properties of the rock (Hallet et al., 1991). This could explain why
324 AE activity increases for temperatures several degrees below $0^\circ C$. Moreover,
325 one may keep in mind that the surface was probably much cooler than at a
326 depth of 10 cm during the freezing period.

327 The former mechanism is related to an increase of the ice pressure within
328 fractures. One may reasonably ask if the AE could be induced by the cracking
329 of ice itself instead of the embedding rock. Ice growth is supposed to induce
330 compression stress in the ice and, due to the reaction of the embedding material,
331 tensile stress at the cracks tips. The ice, as rock do, behaves according the
332 Coulomb failure criterion (Weiss and Schulson , 2009). So its strength is larger
333 in compression than in tension. The symmetrical case applies for the rock
334 around the cracks: its strength is smaller in tension than in compression, more
335 over the crack tips acts as a stress concentrator. This suggests that the rupture

336 is more likely to occur in the rock rather than in the ice. To test this, one may
337 realize waves velocity measurement during periods without ice, to verify that
338 the damage induced in the rock is increasing with time.

339 Beside the results common to all sensors we also observed differences be-
340 tween sensors. The amplitude of the AE activity seems to depend both on the
341 fracture state and the water availability. The relationship between AE activity
342 and temperature displays also some variations from sensor to sensor. This could
343 be attributed to the fact that we use a single temperature measurement point
344 that we considered as representative of the whole monitoring area. Spatial vari-
345 ability probably exists (cf Gubler et al., 2011) that we were not able to take into
346 account in more detail here. The sensor AE5 behaves differently than others
347 in the sense that the AE crisis occurred during the day whereas the night time
348 was relatively quiet. In contrast to all others, this sensor was sensitive to AE
349 activity associated with fractures supplied with meltwater from above. The AE
350 activity could then be related to water pressure increase in the fractures that
351 has been shown to be able to produce seismicity even with limited water table
352 fluctuations (Guglielmi et al., 2008). Here we based our discussion on qualita-
353 tive considerations of water availability. For a more quantitative discussion, it
354 is clear that further measurements are necessary, in particular for assessing the
355 amount of liquid water available for the formation of ice, and the water pres-
356 sure. This could help us to define the respective roles played by thermal dilation,
357 ice formation and cryosuction in the generation of AE. Such measures should
358 be spatially distributed for better understand the spatial heterogeneity of AE
359 activity.

360 The spatial variability we observed concerning AE activity, and consequently
361 damage, has major consequences on the potential erosion induced by frost-
362 cracking. Considering the frost-cracking as homogeneous and deriving the po-
363 tential of damage from calculation based on the air temperature only (e.g. Hales
364 and Roering, 2007; Delunel et al., 2010), could lead to poor estimates of weather-
365 ing. Another important point for the evolution of the morphology in mountain
366 areas is that fractured zones appear more prone to frost-cracking. This en-

367 hances the localization of damage that tends to concentrate in already damaged
368 zones. As a consequence, the spatial variability of frost-cracking should increase
369 with time.

370 The second striking aspect is the PL distribution of the AE event energy,
371 which is clearly identified on all sensors with an exponent $b = 1.55 \pm 0.05$ and
372 verified over six orders of magnitude. This scaling relationship is the signature of
373 the rock micro-fracturing activity. Beside the claim of the possible universality
374 of the b -value, this parameter has been proposed to be dependent on various
375 parameters, in particular on the strength heterogeneity of the material, the
376 applied stress and the proximity of failure (see Amitrano , 2012, and references
377 therein for an extended discussion).

378 The b -value of 1.55 for the energy pdf (corresponding to 0.55 for the cdf), we
379 observed for our data, is in the range expected for rocks experiencing uniaxial
380 or triaxial compression stress state in laboratory experiences (Lockner, 1993)
381 and is very close to that observed prior to the peak load (Amitrano, 2003;
382 Lockner, 1993) and for creep of compression tests after the onset of tertiary
383 creep (Grgic and Amitrano, 2009) and for seismic forerunners recorded in a cliff
384 before its collapse (Amitrano et al., 2005) although the loading mode is very
385 different. This provides an indication that the stress induced by thermal cycling
386 and/or freezing/thawing of water in rock pores and cracks reaches values close
387 to the rock strength. The fact that the b -value is found to be similar for all
388 sensors indicates that strength heterogeneity and stress, which are the two main
389 factors influencing this parameter, are comparable in the different locations we
390 investigated.

391 The time distribution of AE also reveals power-law distribution, which is
392 a supplementary indication of the complex behavior the frost induced damage
393 and of the presence of strong interaction between damage events. The damage
394 activity appears to be clustered at two different time scales, for $\Delta t < 5 - 10s$
395 and for $100s < \Delta t < 10h$. This could be interpreted as the effect of two loading
396 mechanisms for which the interaction operates at different time scale. The first
397 could correspond to cascading events related to the elastic redistribution of stress

398 when damage occurs. The second could be related to the reloading induced by
399 temperature changes and/or water migration that operates at longer time scales.

400 In order to investigate these questions more precisely, further measurements
401 estimating the source depth of AE events, the evolution of the liquid water
402 content and of the temperature at depth are crucial. Moreover, the mechanisms
403 through which freeze/thaw-induced stresses can occur also need to be better
404 differentiated under in-situ conditions. differentiated under in-situ conditions.

405 The experiments of Hallet et al. (1991), revealing sustained microfracturing
406 activity throughout a 3-day period during which temperature and temperature
407 gradients were held constant in the sample, demonstrated the ability of ice
408 segregation to fracture rock. However, in the field, such conditions are never
409 achieved and it may be difficult to robustly distinguish the role of volumetric
410 expansion (as water turns into ice) from that of ice segregation. A number of
411 factors such as solutes, pressure, pore size and pore material can depress the
412 freezing point of water contained in rock (Krautblatter et al, 2010) down to
413 several degrees below 0°C. This explains why ice formation in rock, such as in
414 any porous media, is progressive (Coussy and Fen-Chong, 2005) occurring over
415 a whole range of sub-zero temperatures. Volumetric expansion-induced damage
416 could potentially occur over this whole range of temperatures. Using theoretical
417 arguments, Vlahou and Worster (2010) reported that volumetric expansion can
418 only develop significant pressures ($\sim 10\text{MPa}$) in water saturated confined (spher-
419 ical) cavities larger than 1cm in diameter of very low permeability (10^{-15}cm^2).
420 This prediction basically rules out the role of volumetric expansion on rock
421 fracture, since such conditions are very seldom (if never) achieved in nature.
422 Contrastingly, a different body of work from the cement and concrete research,
423 has shown theoretically that both crystallization and micro-cryosuction mech-
424 anisms can induce pressures of several tens of MPa in a single water-saturated,
425 micrometer-size pore embedded in a porous material (Coussy and Fen-Chong,
426 2005). While the pore structure of rocks certainly does contain such small fea-
427 tures (Fredrich et al, 1995), a partial saturation of pore space, which is often
428 achieved in natural conditions, might give a completely different picture.

429 The results we have reported in this letter show the feasibility of studying
430 rock damage under natural conditions of thermal cycling and freezing using the
431 AE technique. AE activity was shown to significantly increase in sub-zero tem-
432 peratures, especially in locations of the rock-wall that are prone to receiving
433 melt water, suggesting that freezing-induced stresses contribute to rock dam-
434 age. The robust PL distribution of AE event in the domains of energy and
435 time distributions, a common observation in rupture processes dynamics, is an
436 indication that damage is acting in such conditions. These results finally sug-
437 gest that the framework of further modeling studies (theoretical and numerical)
438 should include damage, elastic interaction and poro-mechanics in order to de-
439 scribe freezing-related stresses.

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Sensor	AE1	AE2	AE3	AE4	AE5	AE6
Nb. AE events	1.2×10^4	1.5×10^4	0.2×10^4	1.8×10^4	0.3×10^4	1.6×10^4
R	0.80	0.86	0.70	0.94	0.37	0.75
Water supply	Snow cover	From above	Dry	From above	Through fractures	From above
Fractured	No	Slightly	Yes	Yes	Yes	No
Topography	Slightly convex	Slightly concave	Concave	Convex	Concave	Concave

Table 1: Number of AE events recorded and configuration at each sensor location. R is the proportion of events recorded during negative temperature periods.

550 Figure 1 : Sensor installation at rock surface (A) and weatherproofing (B)
551 shown for AE4. Sensor locations (C) are referred to as AE1-AE6 in the text.

552 Figure 2 : Example of an acoustic emission trace recorded on channel 1. The
553 dotted lines indicate the trigger above which the trace is recorded. The signal
554 energy, E , is calculated by summing the squared amplitude over the duration of
555 the signal. The maximal and minimal amplitude are shown for information.

556 Figure 3: AE event rates and rock temperature at 10, 35 and 60 cm depth
557 as a function of the time. Time corresponds to the hour in the day.

558 Figure 4: AE event rates as a function of the temperature at 10 cm depth.
559 Left: Number of events detected in each temperature interval. The bins are
560 $0.5^\circ C$ width. Right: Event number detected in each temperature interval nor-
561 malized by the time spent in each interval. Sensors 1 3 and 5 (Top), which
562 are in relatively dry sites, have an AE activity smaller than sensors 2 4 and
563 6. Except sensor 5, they all show a clear dependence on the temperature, with
564 a huge increase of the AE for negative temperatures and reduced activity for
565 positive temperatures. When normalizing by the time spent in each bin (right
566 draft) the temperature dependence appears stronger for sensors AE2 AE4 and
567 AE6, and weaker for sensors AE1 and AE3. The particularity of the sensor AE5
568 appears stronger.

569 Figure 5: AE event energy distribution for all channels for the complete
570 experiment. The Power-law trend with $b = 1.55$ is shown as guideline. The
571 completeness energy is also shown to correspond to 100 (Arbitrary unit).

572 Figure 6: Temporal correlation integral of AE events for all channels. The
573 dashed lines provide guidelines corresponding to exponents $D_2 = 0.75$ and $D_2 =$
574 0.9 .

575 Figure 7 Mean frequency of the AE event for each sensors. Blue dots corre-
576 spond to the individual values whereas red crosses give the moving average for
577 100 successive events and 90% of overlap.

578 Figure 8: Top: AE activity for all the sensors. Center: In depth gradient of
579 the temperature. Bottom: rate of change temperature. All values are plotted
580 as a function of hours.

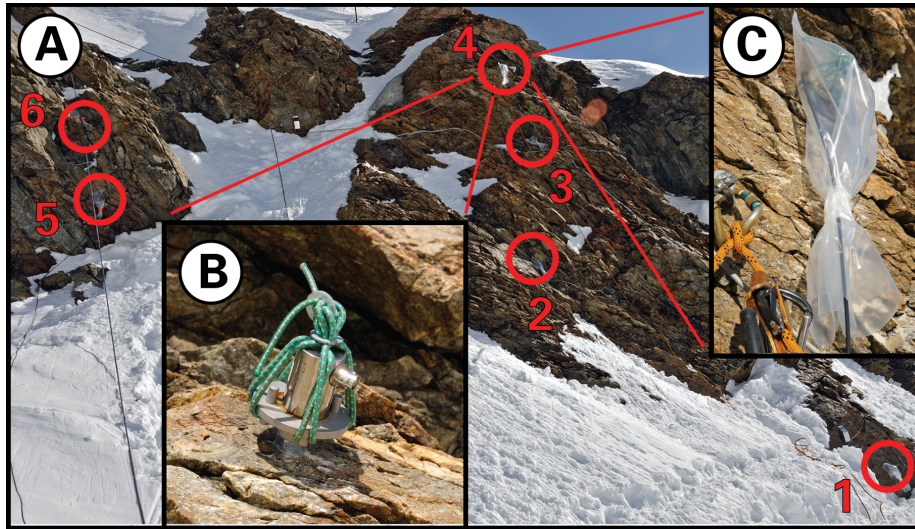


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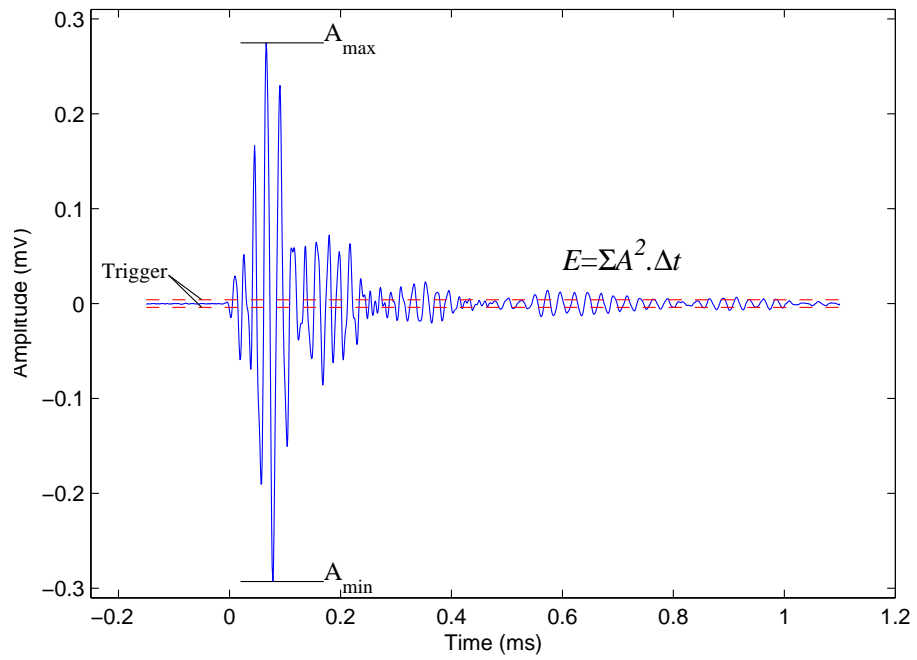


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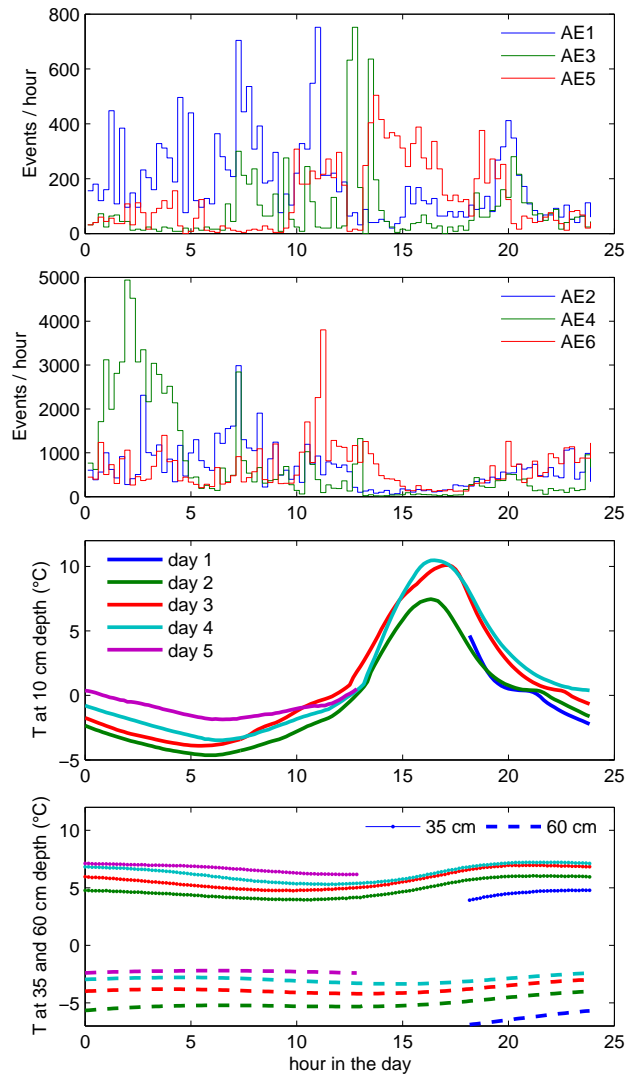


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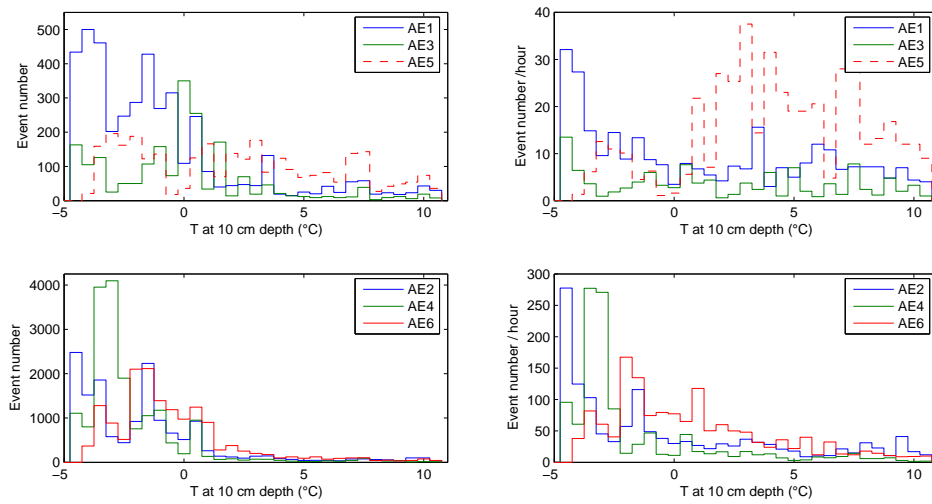


Figure 4: AE event rates as a function of the temperature at 10 cm depth. Left: Number of events detected in each temperature interval. The bins are $0.5^{\circ}C$ width. Right: Event number detected in each temperature interval normalized by the time spent in each interval. Sensors 1 3 and 5 (Top), which are in relatively dry sites, have an AE activity smaller than sensors 2 4 and 6. Except sensor 5, they all show a clear dependence on the temperature, with a huge increase of the AE for negative temperatures and reduced activity for positive temperatures. When normalizing by the time spent in each bin (right draft) the temperature dependence appears stronger for sensors AE2 AE4 and AE6, and weaker for sensors AE1 and AE3. The particularity of the sensor AE5 appears stronger.

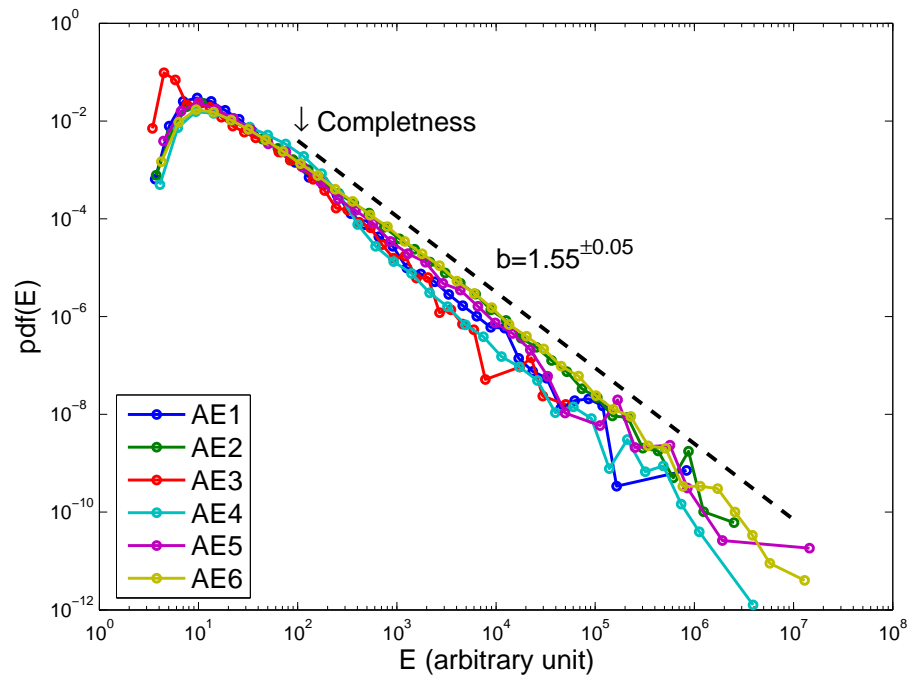


Figure 5: AE event energy distribution for all channels for the complete experiment. The Power-law trend with $b = 1.55$ is shown as guideline. The completeness energy is also shown to correspond to 100 (Arbitrary unit).

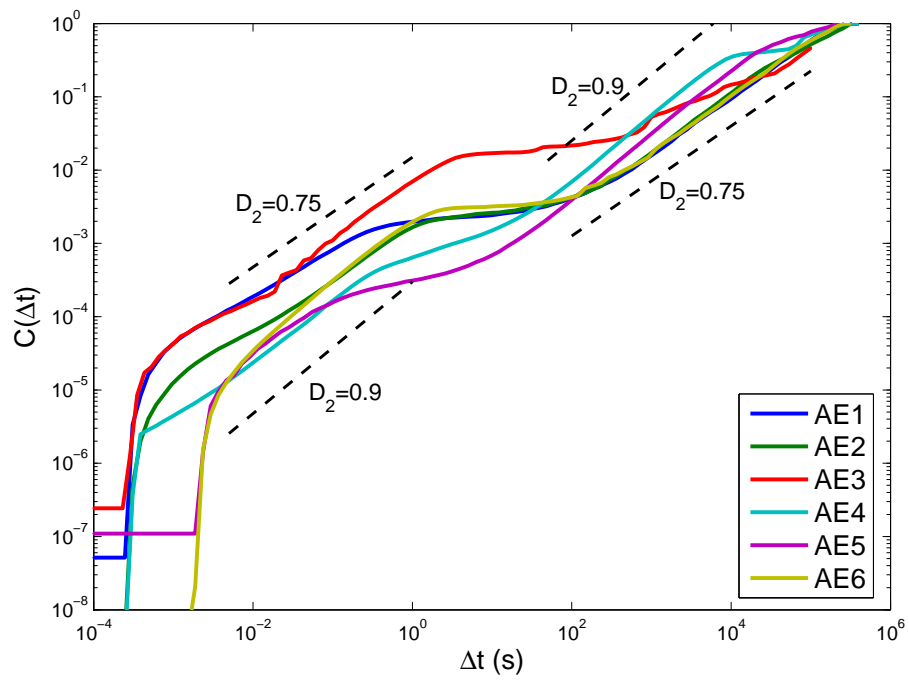


Figure 6: Temporal correlation integral of AE events for all channels. The dashed lines provide guidelines corresponding to exponents $D_2 = 0.75$ and $D_2 = 0.9$.

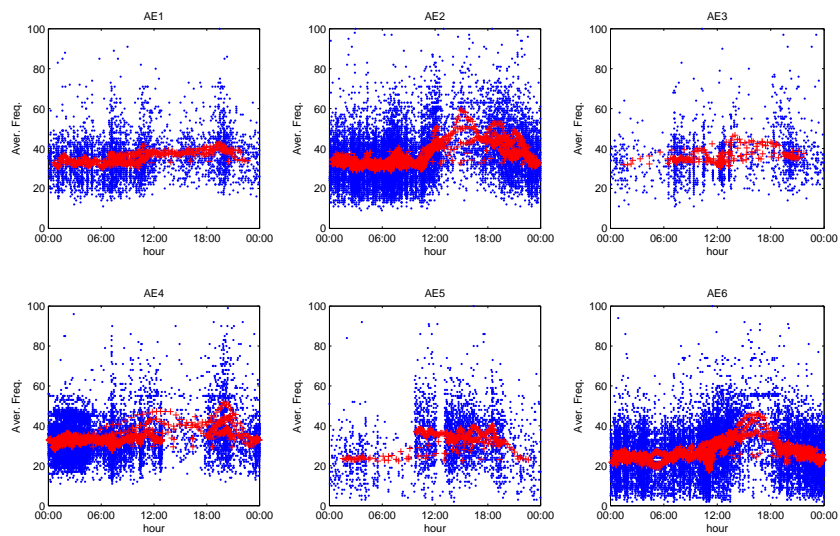


Figure 7: Mean frequency of the AE event for each sensors. Blue dots correspond to the individual values whereas red crosses give the moving average for 100 successive events and 90% of overlap.

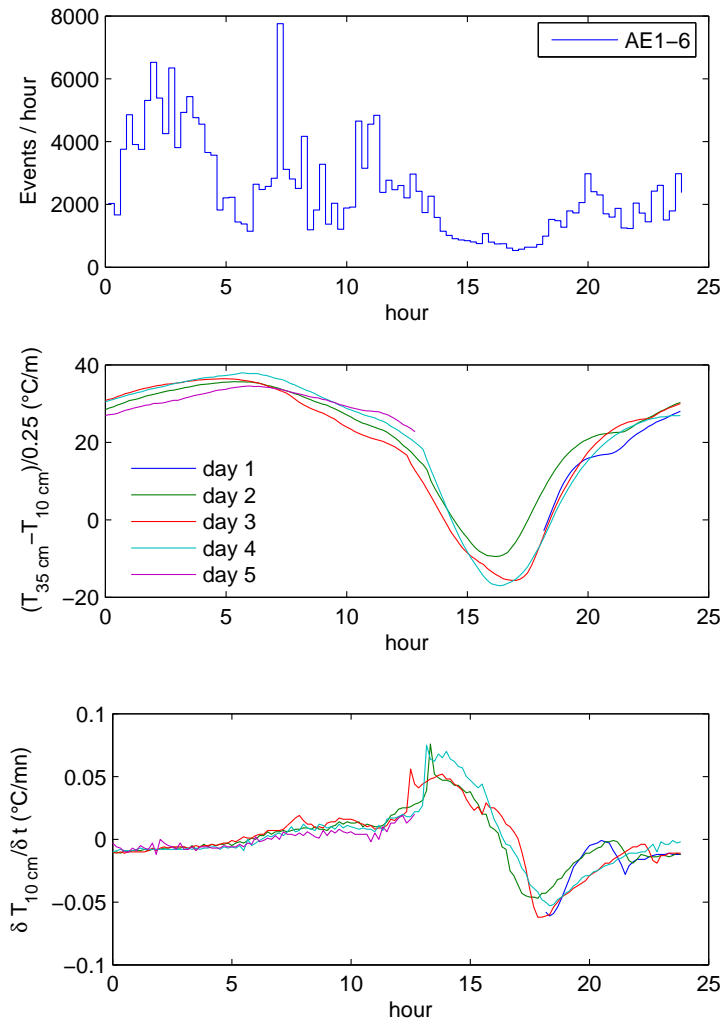


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