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1 The RESOLVE project: a multi-physics experiment with a temporary
2 dense seismic array on the Argentière Glacier, French Alps

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13

14

15 **ABSTRACT**

16 Recent work in the field of cryo-seismology demonstrates that high frequency (>1 Hz) seismic
17 waves provide key constraints on a wide range of glacier processes such as basal friction,
18 surface crevassing or subglacial water flow. Establishing quantitative links between the
19 seismic signal and the processes of interest however requires detailed characterization of the
20 wavefield, which at high frequencies necessitates the deployment of large and particularly
21 dense seismic arrays. Although dense seismic array monitoring has recently become
22 increasingly common in geophysics, its application to glaciated environments remains limited.
23 Here we present a dense seismic array experiment made of 98 3-component seismic stations
24 continuously recording during 35 days in early spring 2018 on the Argentière Glacier, French
25 Alps. The seismic dataset is supplemented with a wide range of complementary observations
26 obtained from ground penetrating radar, drone imagery, GNSS positioning and in-situ
27 measurements of basal glacier sliding velocities and subglacial water discharge. We present
28 first results through conducting spectral analysis, template matching, matched-field
29 processing and eikonal wave tomography. We report enhanced spatial resolution on basal

30 stick slip and englacial fracturing sources as well as novel constraints on the heterogeneous
31 nature of the noise field generated by subglacial water flow and on the link between crevasse
32 properties and englacial seismic velocities. We outline in which ways further work using this
33 dataset could help tackle key remaining questions in the field.

34

35 **INTRODUCTION**

36 The deployment of large and dense seismic arrays is becoming increasingly common in various
37 geophysical contexts thanks to new technological developments of autonomous wireless
38 seismographs and increased computational power. Spatially dense arrays enhance the
39 characterization of high frequency (>1 Hz) body and surface waves propagating in the
40 subsurface, such as for example in near-surface fault systems exhibiting hundreds to few tens
41 of meters long structures (e.g., the Newport-Inglewood Fault, see Lin *et al.* (2013), and the
42 San Jacinto Fault, see Roux *et al.* (2016)). The improved resolution provided by dense arrays
43 increases the completeness of impulsive seismic event catalogs (Vandemeulebrouck *et al.*,
44 2013), thus allowing source spatio-temporal dynamics and subsurface structure to be studied
45 in greater details (Meng and Ben-Zion, 2018; Chmiel *et al.*, 2019). Dense arrays also help to
46 detect other sources of radiation (e.g. tremor and anthropogenic sources) compared to what
47 is possible with single stations or regional networks (Inbal *et al.*, 2016; Li *et al.*, 2018; Meng
48 and Ben-Zion, 2018). Despite its strong potential, dense array monitoring is still limited in the
49 study of glaciers, although it could be used to address some of the key open questions in the
50 field of glaciology.

51

52 Glaciers exhibit damage zones created by surface and/or basal crevasses (Walter *et al.*, 2015;
53 Lindner *et al.*, 2019; Zhan, 2019; Sergeant *et al.*, 2020) as well as complex three dimensional

54 structures associated with firn/debris layers, bedrock topography or englacial water conduits
55 (Cuffey and Paterson, 2010). These features are known to undergo large spatial and temporal
56 changes and to play an important role in glacier dynamics and thermo-dynamics (Durand *et*
57 *al.*, 2011; Scherler *et al.*, 2011; Gilbert *et al.*, 2020). However, conventional geophysical
58 techniques such as radar sounders capable of resolving englacial structural features (e.g.,
59 Evans and Robin, 1966; Church *et al.*, 2019) are not suited for evaluating detailed spatial and
60 temporal changes as well as their effects on the overall glacier behavior. Instead, active and
61 passive surveys using dense seismic arrays may enable accurately monitoring these changes
62 (e.g. using englacial seismic velocities, see Lindner *et al.* (2019) and Preiswerk *et al.* (2019))
63 and thus yield unprecedented constraints on glacier structure and temporal evolution.

64

65 Glaciers and ice sheets generate a large variety of seismic signals, from impulsive transients
66 to emerging and sustained tremors (Podolskiy and Walter, 2016; Aster and Winberry, 2017).
67 Impulsive arrivals from basal stick-slip events have been observed in numerous glaciological
68 contexts (Weaver and Malone, 1979; Allstadt and Malone, 2014; Helmstetter, Nicolas, *et al.*,
69 2015; Lipovsky and Dunham, 2016; Lipovsky *et al.*, 2019; Walter *et al.*, 2020) and may yield
70 crucial information on glacier basal motion, which exerts a primary control on glacier and ice-
71 sheet dynamics and the associated eustatic sea level rise (Ritz *et al.*, 2015; Vincent and
72 Moreau, 2016). The mechanisms giving rise to stick-slip sliding as well as its effect on large-
73 scale ice flow, however, remain poorly constrained (Lipovsky *et al.*, 2019). Dense array
74 monitoring could allow improving the detection of stick-slip events, yield more accurate
75 inversions of locations and show if and how fast stick-slip asperities migrate.

76

77 Impulsive events from englacial fracturing are also commonly observed (Neave and Savage,
78 1970; Roux *et al.*, 2010; Mikesell *et al.*, 2012; Podolskiy *et al.*, 2018; Garcia *et al.*, 2019), and
79 may help elucidate the role of crevassing in iceberg calving, the disintegration of ice-shelves
80 and the occurrence of serac falls (Faillettaz *et al.*, 2008; Krug *et al.*, 2014; Lipovsky, 2018). The
81 improved detection and resolution provided by dense arrays could provide novel constraints
82 on crevasse depth and rupture propagation rates, which are needed to test models (van der
83 Veen, 1998; Weiss, 2004; Tsai and Rice, 2010) and thus better understand ice sheet integrity.

84

85 Recent seismic investigations have also reported widespread emergent and sustained tremor
86 signals generated by resonances in moulins or water-filled crevasses (Helmstetter, Moreau, *et*
87 *al.*, 2015; Roeoesli *et al.*, 2016; Aso *et al.*, 2017), subglacial water flow (Bartholomaus *et al.*,
88 2015; Eibl *et al.*, 2020; Lindner *et al.*, 2020; Nanni *et al.*, 2020) and subglacial sediment
89 transport (Gimbert *et al.*, 2016). The possibility to calculate physical characteristics of
90 subglacial water flow as well as of subglacial sediment transport from seismic tremor
91 observations (Tsai *et al.*, 2012; Gimbert *et al.*, 2014, 2016; Bakker *et al.*, 2020) is particularly
92 appealing. These processes play an important role in ice sliding speeds (Zwally *et al.*, 2002;
93 Schoof, 2010; Tedstone *et al.*, 2015) and bedrock erosion (Beaud *et al.*, 2016), and yet
94 hydraulic measurements at the ice bed are notoriously difficult, with traditional approaches
95 such as borehole techniques providing point measurements, only (e.g., Iken and Bindschadler,
96 1986; Iken *et al.*, 1993). Theoretical links between discharge, pressure regime, sediment
97 transport rates and geometry of the subglacial drainage system and seismogenic hydraulic
98 noise sources remain to be more fully tested and dense seismic arrays may provide the
99 necessary spatial extent and resolution for doing so.

100

101 Properly evaluating the knowledge gain that dense seismic arrays may provide to address the
102 above-mentioned challenges requires (i) monitoring a glacier that gathers the processes of
103 interest, (ii) deploying instrumentation that covers scales and durations over which significant
104 changes operate, and (iii) acquiring complementary observations to test the seismically-
105 derived findings and incorporate these into a wider glaciological context. Here we present
106 data and first analysis from a 98 sensor array deployed over 35 days during early spring 2018
107 on an Alpine Glacier, the Argentière Glacier in the French Alps (Fig. 1). We also provide and
108 analyze key complementary observations from Ground Penetrating Radar (GPR), drone
109 imagery, Global Navigation Satellite System (GNSS) positioning and in-situ instrumentation of
110 basal glacier sliding velocities and subglacial water discharge. We argue that the selected
111 glacier, the time period of investigation as well as the completeness of the present dataset
112 satisfies all of the three above-mentioned conditions. Through application of spectral analysis,
113 template matching, matched-field processing and eikonal wave tomography, we demonstrate
114 that use of the present dataset enhances spatial resolution of basal stick slip activity and near
115 surface crevassing. We further provide novel constraints on the degree of heterogeneity of
116 the seismic noise field generated by subglacial water flow and the variations of englacial
117 seismic velocities. We finally outline how future work using this dataset could help overcome
118 classical observational limitations and address key challenges in the field.

119

120 **EXPERIMENT DESIGN**

121

122 **FIELD SITE**

123

124 The Argentière Glacier is located in the Mont Blanc Massif (French Alps, 45°55' N, 6°57'E, Fig.
125 1a) and is the second largest French glacier. It is about 10 km long, covers an area of about 12

126 km², extends from an altitude of 1700 m above sea level (a.s.l.) up to about 3600 m a.s.l. with
127 an equilibrium line altitude at about 2900 m a.s.l. (Vincent *et al.*, 2018). Over the past three
128 decades, glacier total mass balance has been negative (0.7 meter water equivalent loss per
129 year on average over the period 1976-2019), glacier snout has been retreating by several
130 hundreds of meters (815 m retreat between 1990 and 2019), glacier surface elevation has
131 decreased by several tens of meters and glacier surface velocities have decreased by about a
132 factor of two in the ablation zone (Vincent *et al.*, 2009). The upper part of the glacier is
133 constricted in a typical U-shaped narrow valley where ice sits on granite. The lower part of the
134 glacier is characterized by a sharper incised, V-shaped valley where ice sits on metamorphic
135 rocks (Vallon, 1967; Hantz and Lliboutry, 1983; Vincent *et al.*, 2009). The glacier generally
136 exhibits temperate bed conditions (Vivian and Bocquet, 1973), i.e. basal ice temperature is at
137 the pressure melting point and water flow occurs at the interface as a result of year-round
138 basal melt and summer surface melt (Cuffey and Paterson, 2010).

139

140 The monitored site is located in the lower part of the glacier (about 2 km from the glacier
141 front) and at about 2400 m a.s.l. (Fig. 1b). In this area the surface slope is gentle (1-2%) and
142 crevasses are restricted to an area of about 200 m from the glacier sides. The glacier flows at
143 a yearly average velocity of about 60 m.yr⁻¹ in its center, about half of which is due to sliding
144 at the ice-bed interface and the other half to internal ice deformation (Vincent and Moreau,
145 2016). Internal ice deformation likely occurs primarily through ice creep except near the
146 glacier sides where crevasses are large and potentially deep, such that englacial fracturing
147 could also play a role. A strong seasonality is observed in glacier dynamics, with summer
148 (typically May to September) velocities equal to about 1.5 times winter velocities (Vincent and

149 Moreau, 2016). This behavior is known to result from melt water input lubricating the ice-bed
150 interface and enhancing basal sliding (Lliboutry, 1959, 1968; Cuffey and Paterson, 2010).
151 Previous seismic observations at this site report various seismogenic sources associated with
152 surface and intermediate depth crevassing (Helmstetter, Moreau, *et al.*, 2015), basal stick-slip
153 (Helmstetter, Nicolas, *et al.*, 2015), subglacial water flow (Nanni *et al.*, 2020), and serac
154 instabilities in the glacier front (Roux *et al.*, 2008).

155

156 SEISMIC INSTRUMENTATION AND GEOPHYSICAL CHARACTERISATION OF GLACIER STRUCTURE 157 AND DYNAMICS

158

159 **Seismic instrumentation**

160

161 Sensors of the dense seismic array (red dots in Fig. 1b) are Fairfield ZLand 3-component nodal
162 seismographs with a sampling frequency of 500 Hz (hereafter referred to as nodes). These
163 sensors have a low-corner cut-off frequency of 5Hz, a sensitivity of $76.7 \text{ V.m}^{-1}.\text{s}^{-1}$ (see Ringler
164 *et al.* (2018) for a detailed laboratory analysis of sensor response characteristics) and a typical
165 power autonomy of about 35 days. We deployed the nodes on 24 April 2018 when the glacier
166 was entirely covered by a snow layer of about 3 m thick. We placed the sensors about 40 m
167 apart from each other in the along-flow direction and about 50 m apart in the across-flow
168 direction in order to enable subwavelength analysis in the 4-50 Hz frequency range of interest.
169 Given that snow melt occurs at an average rate of $2\text{-}4 \text{ cm.day}^{-1}$ water equivalent at this
170 location (Vincent and Moreau, 2016), we decided to bury the nodes into snow about 30 cm
171 below the surface to ensure the sensors be (i) well coupled to their surroundings and
172 maintained levelled over a week-long time period until snow melt uncovered them and (ii)

173 shallow enough for the GNSS signal to pass through the snow layer and ensure proper
174 reception for time synchronization. Given the limited melt that occurred over the first half of
175 the monitoring period, we had to re-bury the sensors only once over the monitored period,
176 on 11 May 2018. This strategy ensured that little data was lost due to melt-out-induced tilt.

177

178 We supplemented the seismic array with one three-component borehole seismic station
179 placed at 5 m below the ice-surface (orange dot in Fig. 1b). This Geobit-C100 sensor was
180 connected to a Geobit-SRi32L digitizer and provides higher sensitivity ($1500 \text{ V.m}^{-1}.\text{s}^{-1}$), higher
181 frequency sampling (1000 Hz) and a lower low-corner cut-off frequency (0.1 Hz) compared to
182 the nodes. This seismic station is the same as the one used for the two-year long seismic study
183 of Nanni *et al.* (2020).

184

185 **Recovery of surface and bed digital elevation models from structure from motion surveys** 186 **and ground penetrating radar**

187

188 We construct a digital surface elevation model based on a drone geodetic survey that we
189 conducted on September 5, 2018 when the glacier surface was snow free and crevasses could
190 be identified. We used a senseFly eBee+ Unmanned Aerial Vehicle and acquired a total of 720
191 photos using the onboard senseFly S.O.D.A. camera (20 Mpx RGB sensor with 28 mm focal
192 lens). We generate a digital elevation model of 10-cm resolution (see white contours in Fig.
193 1b) using differential Global Positioning System (GPS) measured ground control points (see
194 green stars in Fig. 2a) and the Structure for Motion algorithm implemented in the software
195 package Agisoft Metashape Professional version 1.5.2. A detailed description of the
196 processing steps can be found in Brun *et al.* (2016) and Kraaijenbrink *et al.* (2016).

197

198 We calculate a crevasse map (black dots in Fig. 2a) based on the surface digital elevation
199 model, which has been shown to be more reliable and precise than using optical/radar images
200 (Foroutan *et al.*, 2019). We first apply a 2D highpass filter with a low cut-off wavelength of 10
201 m and then define any location with elevation lower than -50 cm as being part of crevasses.
202 Finally, we apply a 2D median filter with a 1 by 1 m kernel in order to remove artifacts from
203 boulders and moraines.

204

205 To establish a digital elevation model of the glacier bed we primarily use Ground Penetrating
206 Radar (GPR) data acquired using a system of two transmitting and receiving 4.2 MHz antennas
207 connected to a time triggered acquisition developed especially for glacial applications by the
208 Canadian company Blue System Integration Ltd (Mingo and Flowers, 2010). The GPR signal
209 processing consists of correcting for source time excitation. We use both dynamic corrections
210 to reproduce a zero-incidence acquisition from data acquired with a 20 m offset between
211 source and receiver (Normal Moveout correction) and static corrections to highlight elevation
212 variations along a profile. We do so using a constant wave velocity of $0.168 \text{ m}\cdot\text{ns}^{-1}$ that is
213 typical for ice (Garambois *et al.*, 2016). We then apply a [1-15 MHz] Butterworth band-pass
214 filter followed by a squared time gain amplification to the signal in order to increase signal-to-
215 noise ratio. We show an illustration of the processed GPR data in Fig. 2b, where the direct air-
216 wave first arrival is followed by a large reflectivity V-shape pattern reaching 3000 ns around
217 the center of the profile. This latter profile corresponds to the ice/bedrock interface, although
218 its apparent shape is biased by reflections off the closest point on the ice-bed interface rather
219 than off the bed portion directly below the instrument. We correct for this bias by applying a
220 frequency-wavenumber Stolt migration technique (Stolt, 1978) and convert time into distance

221 using the constant wave velocity of $0.168 \text{ m}\cdot\text{ns}^{-1}$. We note that prior to migration we add null
222 traces (i.e. with null amplitudes) in places where harsh glacier surface conditions (mainly
223 crevasses) prevented us from acquiring data. As illustrated in Fig. 2c the migration process is
224 effective in correcting the artefacts due to the geometrical variation of the interface along the
225 profile, which now appears smooth and continuous. We then pick the ice-bed reflection
226 (yellow line in Fig. 2c) over all GPR profiles, such that a three-dimensional bed DEM can be
227 reconstructed.

228

229 We reconstruct a three-dimensional bed DEM over a larger area than that covered by GPR
230 surveys by incorporating additional constraints like glacier edge elevation as measured from
231 drone imagery (blue area in Fig. 2a) and bed elevations obtained through rock drilling to the
232 ice-bed interface from bedrock excavated tunnels located further down-glacier (purple area
233 in Fig. 2a). Furthermore, we interpolated all data using a kriging method onto a 10 by 10 m
234 grid. From different first onset pickings we estimate that the recovered depth uncertainty is
235 of about 5 m below the seismic array and likely on the order of a few tens of meters outside
236 of the array where observations are sparser.

237

238 In Fig. 2a we show the ice thickness map (using 25-m bin contours) as reconstructed based on
239 subtracting the bed DEM from the surface DEM. The glacier bed exhibits a gently dipping
240 valley, with a maximum ice thickness of about 255 m at the center of the seismic array. Glacier
241 thickness decreases relatively sharply on the glacier margins where surface crevasses are
242 observed. We also observe that bed elevation significantly increases down glacier, which
243 results in a decrease by more than 150 m in glacier thickness (Fig. 2d). Beyond these generic
244 characteristics we identify two interesting reflectivity features in the migrated GPR images

245 (yellow ellipses in Fig. 2c) that correspond to localized scattering observed near the surface
246 and a large reflectivity pattern observed just above the deepest portion of the interface. The
247 near surface scattering feature could be caused by deep crevasses, and the deeper feature
248 could be caused by englacial and/or subglacial water conduits as recently proposed by Church
249 *et al.* (2019), who made similar GPR observations in a temperate glacier and were able to
250 bolster their interpretation with in-situ borehole observations.

251

252 **Meteorological and water discharge characteristics**

253

254 We use air temperature and precipitation measurements obtained at a 0.5 h time step with
255 an automatic weather station maintained by the French glacier-monitoring program
256 GLACIOCLIM (Les GLACIers un Observatoire du CLIMat; <https://glacioclim.osug.fr/>), which is
257 located on the moraine next to the glacier at 2400 m a.s.l. (green diamond in Fig. 1b).
258 Precipitation is measured with an OTTPluvio weighing rain gauge. Subglacial water discharge
259 is monitored at a 15 min time step in tunnels excavated into bedrock by the Emossons
260 hydraulic power company, which are located 600 m downstream of the array center (at 2173
261 m a.s.l.) near the glacier ice fall (see blue star in Fig. 1b).

262

263 Temperature generally increases over the instrumented period, from a multi-daily average of
264 about 0° C at the beginning of the measurement period to about 5 °C at the end (Fig. 3a). This
265 drives the general increase in water discharge, which varies from few tenths of $\text{m}^3\cdot\text{s}^{-1}$ to
266 several $\text{m}^3\cdot\text{s}^{-1}$ over the period. Episodic rain events also occur during the instrumented period,
267 but have little to no effect on subglacial discharge likely as a result of the snow cover acting as
268 a water storage buffer (Fountain and Walder, 1998).

269

270 **Glacier dynamics instrumentation and general features**

271

272 We evaluate changes in glacier dynamics over the instrumented period by means of two
273 observational methods. The first one is unique to the present site, and consists of basal sliding
274 velocity measurements made continuously in the down glacier serac fall area (see red star in
275 Fig. 1b) by means of a bicycle wheel placed directly in contact with the basal ice at the
276 extremity of an excavated tunnel (Vivian and Bocquet, 1973; Moreau, 1999). The wheel is
277 coupled with a potentiometer that retrieves its rotation rate, which is then recorded digitally
278 and converted back to a sliding velocity at a 1-s sampling time with a displacement increment
279 resolution of 0.07 mm. The second type of measurements consists of 4 glacier surface and 1
280 reference bedrock GNSS stations (yellow stars in Fig. 1b) of type Leica GR25 acquiring the GNSS
281 signals every second. This temporary array is supplemented by a permanent ARGR GNSS
282 station from the RESIF-RENAG network (<http://renag.resif.fr>) on the bedrock close to the
283 glacier 3 km uphill (yellow star in Fig. 1a). The GNSS antennas on the glacier are installed on
284 8-m long aluminum masts anchored 4-m deep in the ice and thus emerging about a meter
285 above the snow surface at the beginning of the measurement period. The temporary station
286 placed next to the glacier side provides a reference for validating kinematic GNSS processing
287 approaches, evaluating station positions from every single set of GNSS signal recordings (i.e.
288 every second, as opposed to static processing, which cumulates GNSS signals over a much
289 longer time). We conduct such kinematic processing using the TRACK software ((Herring *et al.*,
290 2018), <http://geoweb.mit.edu/gg/docs.php>). Our processing chain includes the on-line tool
291 SARI (<https://alvarosg.shinyapps.io/sari/>) for the removal of outliers that arise from low
292 satellite coverage in the glacier valley and to perform a de-trend and re-trend analysis to

293 estimate and correct for offsets due to manual antenna mast shortening as snow melt
294 progresses. We also correct for multi-path effects induced by GNSS signal reflections from the
295 ground, although we find that those are attenuated by the combination of GPS and GLONASS
296 signals thanks to their different sidereal periods (~ 24 h for GPS and ~ 8 days for GLONASS). We
297 finally calculate position time series at a 30-s time step sufficient to capture glacier dynamics
298 and subsequently evaluate three-dimensional velocities by the linear trends of the position
299 components. The horizontal velocity is calculated as $v_h = \sqrt{v_N^2 + v_E^2}$ where v_N and v_E are
300 the North and East components, respectively.

301

302 To facilitate comparison of basal sliding and surface velocity here we smooth both timeseries
303 at a 36-hr timescale (Fig. 3b), since daily down to sub-daily fluctuations in basal sliding
304 velocities are largely affected by unconstrained variations in the local ice roughness in contact
305 with the wheel, as for example, when an entrained rock passes over the wheel. Although basal
306 sliding velocity is to be lower than surface velocity, here both quantities have similar absolute
307 values because the sliding velocity is measured at a place where the glacier is much steeper
308 (25% slope as opposed to 1-2%) and thus driving stress is much larger than at the GNSS
309 locations. We observe an increase in basal sliding velocity from 4.5 mm/h to more than 6
310 mm/h at the very beginning of the monitored period. Such an acceleration is commonly
311 observed in spring on alpine glaciers (Iken and Bindschadler, 1986; Mair *et al.*, 2002; Vincent
312 and Moreau, 2016) and is known to correspond to water pressurization caused by an increase
313 in water input at the bed due to surface melt water supply, which causes the reduction of
314 friction and thus the enhancement of sliding (Lliboutry, 1959, 1968; Iken, 1981; Schoof, 2005;
315 Gagliardini *et al.*, 2007). This acceleration is not seen in the GNSS observations, which could
316 be due to the glacier seasonal acceleration occurring earlier at this location. We also observe

317 one major acceleration event in the location of the dense seismic array occurring between 4
318 May and 8 May 2018 likely due to the large concomitant increase in water discharge (see
319 blue line in Fig. 3a) causing basal water pressurization (Cuffey and Paterson, 2010).

320

321

322 **FIRST RESULTS**

323

324 SEISMIC NOISE CHARACTERISTICS

325

326 We investigate the spatial and temporal variability of seismic power P (in dB) across a wide
327 range of frequencies by applying Welch's method (Welch, 1967) over 4 seconds-long vertical
328 ground motion timeseries (with 50 % overlap) prior to averaging power (in the decibel space)
329 over 15 minutes-long time windows. This two-step strategy allows limiting the influence of
330 impulsive events (which are studied in more details in the next sections) on the seismic power
331 while enhancing that of the continuous background noise (Bartholomäus *et al.*, 2015; Nanni
332 *et al.*, 2020). In Fig. 4 we present 1-100 Hz spectrograms (i.e. seismic power at any given
333 frequency and time) over the first half of the instrumented period (15 April to 14 May 2018)
334 together with timeseries of 2-20 Hz frequency median seismic power at five different stations
335 of the array, four stations located on the four array sides and one located in the array center
336 (see node numbers in Fig. 1b and Fig. S1 for spectrograms across all stations and over the
337 entire frequency range and experimental period). Time periods when sensors tilted beyond
338 their specifications (and thus were no longer deemed functional) as a result of snow melt
339 causing them to be no longer buried are manifested by drastically reduced seismic power
340 across the whole frequency range (see node 6 from 8 May to 11 May 2018). Fortunately,

341 sensor tilt only occurred at a small number of seismic stations (11 out of 98) and during a
342 restricted time duration (less than 2 days on average, see Fig. S1). We also observe that
343 spectrograms do not undergo significant change from prior to after sensor reinstallation on
344 11 May 2018. This suggests that these are not significantly affected by potential changes in
345 sensor coupling to snow, which is pleasant given that a previous study found that coupling can
346 strongly affect nodes recorded signals (Farrell *et al.*, 2018).

347

348 All stations generally experience similar multi-day (e.g. four days' average, see black lines in
349 Fig. 4) variations in seismic power that are highly correlated with multi-day discharge
350 variations, although seismic power precedes discharge variations by about a day or two. The
351 likely reason is that seismic power is controlled by the hydraulic pressure gradient, which is
352 highest during periods of rising discharge (Gimbert *et al.*, 2016; Nanni *et al.*, 2020). Although
353 shorter term (e.g. diurnal) variations in seismic power are also similar across stations when
354 discharge is low (from 24 April to 28 April and from 30 April to 4 May) and anthropogenic noise
355 dominates (Nanni *et al.*, 2020), the picture is different at higher discharges when seismic
356 power is dominated by subglacial water flow (Nanni *et al.*, 2020). On 29 April and from 5 May
357 to 10 May seismic power exhibits pronounced (up to 10 dB) and broad frequency (1-100 Hz)
358 short time scale (sub-diurnal to diurnal) variations that are particularly marked at certain
359 stations (e.g. node 6 (Fig. 4a), node 44 (Fig. 4c) and node 50 (Fig. 4d)) and not at others (e.g.
360 node 38 (Fig. 4b) and 95 (Fig. 4e)). We also observe that at certain stations seismic power
361 appears to be continuously or intermittently enhanced within narrow frequency bands. For
362 instance node 38 systematically presents higher seismic power above 20 Hz. These
363 discrepancies suggest that measurements of ground motion amplitude are sensitive to
364 heterogenous and intermittent subglacial water flow, although certain features discussed

365 here could be due to extraneous noise sources associated with sensor coupling (Farrell *et al.*,
366 2018), to localized sources other than subglacial fluid flow or to site effects.

367

368 DETECTING AND LOCATING BASAL STICK SLIP IMPULSIVE EVENTS USING TEMPLATE 369 MATCHING

370

371 We detect high-frequency (>50 Hz) basal stick-slip events using template matching. This
372 follows a two-step analysis as in Helmstetter *et al.* (2015). We first build a catalog of events
373 through applying a short-term-average over long-term-average (STA/LTA) detection method
374 (Allen, 1978) to the continuous high-pass filtered signal (>20 Hz) using a STA time window of
375 0.1 s and a LTA time window of 1 s. We identify an event when the STA/LTA ratio exceeds a
376 factor of 2. We then manually select all events with short duration (<0.2 s) and high average
377 frequency (>50 Hz) and define groups of events referred to as clusters when their correlation
378 with each other exceeds 0.8. For each cluster, we compute the average waveform to define
379 its “template” signal (using a time window of 0.25 s). We take the sum of seismograms
380 normalized by their peak amplitude and weighted by the square correlation between each
381 event and the template event, iterating this procedure several times until convergence. We
382 visually check that events present distinct P and S wave arrivals and use a polarization analysis
383 to ensure that they are not associated with surface waves (Fig. 5a). We then use the template
384 matching filter method (Gibbons and Ringdal, 2006) to further detect smaller amplitude
385 events not picked with the above strategy but exhibiting a correlation higher than 0.5 with the
386 template signal.

387

388 We first conduct the analysis using the borehole station ARG, which has a higher sensor
389 sensitivity, signal-to-noise and sampling rate compared to the nodes. We identify 31 active
390 clusters during the dense array experiment. Interestingly, these clusters constitute a large part
391 of the 46 clusters identified on a much longer period (from December 2017 to June 2018, using
392 the borehole sensor which ran almost continuously, see Fig. 6). Although the amplitude of
393 these signals varies strongly through time (from $1 \cdot 10^{-7} \text{ m.s}^{-1}$ to $4 \cdot 10^{-6} \text{ m.s}^{-1}$ Fig. 6a),
394 waveform characteristics remain strikingly similar (Fig. 6b and 6c). All 46 identified clusters
395 exhibit similar characteristics to that shown in Fig. 6, and their activity does not appear to be
396 temporally correlated with each other, nor with external drivers related to meteorology,
397 hydrological or glacier dynamics.

398

399 We also apply the template matching algorithm on a subset of 10 nodes covering the whole
400 study area, using the same 31 clusters as previously identified using station ARG. At each time
401 step we compute the correlation coefficient between each template signal and the continuous
402 signal averaged over all nodes and all components (using a time window of 0.35 s instead of
403 0.25 s in order to match signal duration at all selected nodes). The large number of sensors
404 allows us to lower the correlation threshold from 0.5 to 0.2 while reducing the number of false
405 detections. Indeed, events belonging to different clusters with very different locations can end
406 up being correlated above the detection threshold when using one station, while using several
407 stations this scenario is much more unlikely. We detect 79% more events using the nodes
408 compared to using the station ARG. The newly detected events are mostly smaller amplitude
409 events. Most (83%) of the events detected using ARG are also detected using the nodes. This
410 shows that increasing the number of sensors allows detecting more events and reducing the
411 number of false detections despite signal-to-noise ratio and sampling rate not being optimum.

412

413 We determine the position of the 31 identified clusters by first manually picking on each node
414 the P and S arrival times associated with the event in each cluster that is associated with the
415 largest correlation with the template event, and then inverting for the location of each event
416 and the associated P and S wave velocities assuming velocities are homogeneous and identical
417 for all events. We only consider first arrivals that are usually geometrically predicted to be
418 direct (as opposed to refracted) waves for most sensors and most events. Assuming a simple
419 1D velocity model with $V_p=3620$ m/s in the ice and $V_p=4300$ m/s in the bedrock (see Fig 10b)
420 and a glacier thickness of 200 m, the direct wave is faster for epicentral distances shorter than
421 306 m. Moreover, even when the refracted wave is faster, it is usually less impulsive and has
422 a smaller amplitude than the direct wave. We estimate P and S wave velocities using a grid
423 search inversion with a step of $10 \text{ m}\cdot\text{s}^{-1}$ and the Nonlinloc software (Lomax *et al.*, 2000) to
424 locate clusters. We assume a standard error of arrival times of $2 \cdot 10^{-3}$ s for P waves, $4 \cdot 10^{-3}$ s
425 for S waves and of $3.5 \cdot 10^{-3}$ s for calculated travel times. We can see in Fig. 5a that the picked
426 arrival times (black circles) are consistent with the computed travel times (green lines). The
427 root-mean-square error for this event is 2.4 ms, which corresponds to about one sample (2
428 ms).

429

430 We show the locations of basal icequakes versus depth in an average transverse section in Fig.
431 7a and on a two-dimensional map in Fig. 8. They are mainly located in the down-glacier part
432 of the array and in the central part of the glacier or near the right side, while there is no event
433 observed towards the left side. Icequake depths range between 80 m and 285 m, and are in
434 good agreement with the bedrock topography estimated from the radar profiles. Uncertainty
435 on absolute source depth is on the order of 10 m (see errorbars in Fig. 7a), and the estimated

436 seismic wave velocities of $V_p=3620 \text{ m.s}^{-1}$ and $V_s=1830 \text{ m.s}^{-1}$ (Fig. 7b) are in good agreement
437 with velocities measured on other alpine glaciers (Podolskiy and Walter, 2016). V_s is much
438 better constrained by the data compared to V_p (Fig. 7b).

439

440 SYSTEMATIC LOCATION OF EVENTS USING MATCHED-FIELD PROCESSING

441

442 Contrary to in the previous section where a priori constraints on waveform characteristics and
443 wave velocity are used to target basal stick-slip events, we next test location of a wide range
444 of seismic events generated by impulsive or emergent sources with no a priori knowledge on
445 waveform characteristics and minimal a priori knowledge on medium properties. The
446 rationale is that the limited a-priori knowledge for source identification is balanced by the high
447 spatial and temporal resolution provided by array processing techniques, which may provide
448 spatial or temporal characteristics facilitating source identification (Vandemeulebrouck *et al.*,
449 2013; Chmiel *et al.*, 2019).

450

451 We conduct Matched-Field Processing (MFP), which consists of recursively matching a
452 synthetic field of phase delays between sensors with that obtained from observations using
453 the Fourier transform of time-windowed data. We obtain the synthetic field from a source
454 model with a frequency-domain Green's function that depends on 4 parameters, which are
455 the source spatial coordinates x , y and z and the medium phase speed c . MFP output is
456 normalized to range from 0 to 1 with higher values corresponding to better matches between
457 modelled and observed signal phases, and therefore a higher confidence in true source
458 location. Here we use a spatially homogeneous velocity field within the glacier, which the
459 advantage of a fast-analytical computation, although it also results in a higher degree of

460 ambiguity between z and c . Contrary to classical beamforming techniques in which a planar
461 wave front is often assumed, our MFP approach considers spherical waves and allows locating
462 sources closer to and within the array. To build a large catalog of events, we apply MFP over
463 short time windows of 1-s with 0.5-s overlap, across 16 frequency bands of ± 2 Hz width equally
464 spaced from 5 to 20 Hz and over the entire study period. Calculating source locations over
465 such a large number of windows requires minimizing computational cost. We do so by using
466 a minimization algorithm that relies on the downhill simplex search method (Nelder-Mead
467 optimization) of Nelder and Mead (1965) and Lagarias *et al.* (1998) instead of using a multi-
468 dimensional grid search approach. As the exploration of the solution space is characterized by
469 a certain level of randomness, we maximize the likelihood that our minimization technique
470 finds a global minima and thus the dominant source over the considered time window through
471 (i) starting the optimized algorithm from a set of 29 points located at a depth of 250 m inside
472 and near the array (see black crosses in Fig. 9d) with a starting velocity $c=1800 \text{ m}\cdot\text{s}^{-1}$ and (ii)
473 taking the highest MFP output out of the 29 inversions found after convergence.

474

475 In Fig. 9b,c we present two examples of events located inside and outside the array and
476 associated with a high MFP output of 0.92. The half-size of the focal spot in the MFP output
477 field gives a measure of the location uncertainty (Rost and Thomas, 2002), which is about 10
478 m for events located inside the array and can increase up to 40 m when for events up to 100
479 m away from the array edges. Gathering all sources over one continuous day of record, we
480 find that the associated MFP output distribution exhibits a heavy tail towards high values (red
481 area in Fig. 9a for an example at 13 Hz). Such a heavy tail is not obtained for a random field,
482 in which case MFP output exhibits a distribution shifted towards almost one order of
483 magnitude lower values. This suggests that most identified sources correspond to real and

484 detectable seismic events. Well resolved seismic events with MFP outputs higher than 0.8 are
485 located near the surface and delineate crevasse geometries, such that they likely correspond
486 to englacial fracturing (red dots in Fig. 8). Few (less than one percent) of these events are
487 however located outside of the glacier and likely correspond to rock falls. Typical waveforms
488 associated with englacial fracturing events are dominated by surface waves arrivals (Fig. 5b),
489 although P waves arrivals as well as arrivals showing hyperbolic moveout (black arrows in Fig.
490 5b) are also distinguishable. Although P waves arrivals associated with surface crevassing
491 events are not commonly observed (Walter *et al.*, 2009; Helmstetter, Moreau, *et al.*, 2015),
492 their observation here may result from improved detection thanks to the dense seismic array.
493 Arrivals showing hyperbolic moveout likely correspond to reflected waves at the
494 glacier/bedrock interface.

495

496

497 USING EVENT CATALOGS FOR STRUCTURE INVERSION

498

499 Dense-array techniques for seismic imaging often involve interferometry analysis on
500 continuous seismic noise. Such techniques however require an equipartitioned wavefield
501 inherited directly from homogenously distributed noise sources and/or indirectly from
502 sufficiently strong scattering (Lobkis and Weaver, 2001; Fichtner *et al.*, 2019). These
503 conditions strongly limit the applicability of such techniques on glaciers where sources are
504 often localized and waves in ice are weakly scattered (Sergeant *et al.*, 2020). An alternative
505 way is to use localized and short-lived sources with known positions (Walter *et al.*, 2015) as
506 those previously identified using our systematic MFP technique, which are numerous and
507 more evenly distributed in space (Fig. 8).

508

509 We consider the catalog of sources associated with MFP outputs larger than 0.6, located near
510 the surface ($z < 10\text{m}$) and close to the array (within a radius of 400 m from the array center).
511 With these criteria our catalog includes about 10^6 sources gathered over the 35 days of
512 continuous recordings. In order to further demonstrate that our MFP calculations yield
513 reliable velocities (i.e. the ambiguity between z and c is limited for these sources), we use the
514 velocities given from our MFP calculation (which for shallow events recover the dominant
515 surface waves) to construct dispersion curves, as opposed to classical f-k analysis (Capon,
516 1969). We infer surface wave phase velocity at each frequency between 3.5 Hz and 25 Hz by
517 fitting a Gaussian function to the probability density distributions of velocities in each
518 frequency bin, and taking the center of the Gaussian function as the most representative
519 velocity in that frequency bin (see Fig. 10a (inset) for an example at 13 Hz). We note that the
520 presently constructed dispersion curve is similar to the one that would be obtained using a
521 classical f-k analysis (not shown). We find that surface wave velocity increases gently from
522 $1560\text{ m}\cdot\text{s}^{-1}$ to $1630\text{ m}\cdot\text{s}^{-1}$ as frequency decreases from 25 Hz down to 7 Hz, and then increases
523 sharply up to $2300\text{ m}\cdot\text{s}^{-1}$ as frequency decreases down to 3.5 Hz. These observations can be
524 reproduced using a three-layer one-dimensional elastic model (using the Geopsy package,
525 Wathelet *et al.* (2020)) that incorporates a gentle velocity increase (from 1670 to $1720\text{ m}\cdot\text{s}^{-1}$
526 for V_s) at 40 m depth and a drastic velocity increase (from 1720 to $2800\text{ m}\cdot\text{s}^{-1}$ for V_s) located
527 between 200 and 220 m depth (Fig. 10b). These values were obtained by trial and error tests.
528 The slightly slower velocities and density within the first 40-m deep layer may be due to
529 surface crevasses, and are consistent with surface events being associated with smaller P wave
530 velocities than those associated with stick-slip events at the ice/bedrock interface (Fig. 5). The

531 200- to 220-m deep drastic discontinuity results from the ice/bedrock interface, consistent
532 with the radar-derived average glacier thickness beneath the seismic network (Fig. 2a).

533

534 We go one step further and perform two-dimensional surface wave inversions from eikonal
535 wave tomography (Roux *et al.*, 2011; Lin *et al.*, 2013; Mordret *et al.*, 2013). We first extract
536 ~200,000 Rayleigh wave travel times using the best (associated with MFP outputs larger than
537 0.9) seismic events and then perform a simple linear inversion for the slowness (starting from
538 a homogeneous initial model with a phase velocity of $1580 \text{ m}\cdot\text{s}^{-1}$, see Fig. 10a) assuming
539 straight rays as propagation paths and an *a-priori* error covariance matrix that decreases
540 exponentially with distance over 10 m. The weight of the spatial smoothing is chosen at the
541 maximum curvature of the standard trade-off analysis (L-curve) based on the misfit value
542 (Hansen and O’Leary, 1993), and the inversion produces a residual variance reduction of ~98%
543 relative to the arrival times for the homogeneous model. In Fig. 8 we show the Rayleigh wave
544 phase velocity maps obtained as a result of the travel-time inversion on a regular horizontal
545 grid with steps of 5 m and using 13-Hz Rayleigh waves, which have largest sensitivity between
546 20 m and 60 m depth (Fig. 10c) according to kernel sensitivity computations performed on the
547 three layer elastic model (Fig. 10b) using the code of Herrmann (2013). We observe that
548 locations with higher crevasse density are generally associated with lower phase velocities, as
549 observed in the left glacier side and in the down glacier part of the array. This observation is
550 however not systematic, since high velocities are also observed in the right glacier side and in
551 the up glacier part of the array where crevasses are also present. This could be explained by
552 shallower crevasses or by crevasse orientations, which affect different wave propagation
553 directions in these regions. This latter potential source of bias could be investigated by

554 explicitly accounting for anisotropy in the tomography inversion scheme (Mordret *et al.*,
555 2013).

556

557

558 **DISCUSSION**

559

560 INTERPRETING SPATIAL AND TEMPORAL VARIATIONS IN GROUND MOTION AMPLITUDES

561

562 Although our seismic array observations generally exhibit spatially homogenous multi-day
563 changes in seismic power, there exist specific times when changes in seismic power are
564 spatially heterogeneous (Fig. 4). A surprising observation is that these heterogeneous changes
565 are observed down to the lowest frequencies (3 to 10 Hz) associated with wavelengths larger
566 than the inter-station spacing, such that the observed spatial heterogeneity is unlikely solely
567 caused by wave attenuation. It remains to be investigated as to which processes mainly cause
568 the observed spatial variability in seismic power. Punctual sources identified from the MFP
569 analysis could be used to investigate the respective control of wave attenuation, wave
570 scattering and site effects on amplitude field heterogeneity and its potential dependency on
571 site attributes like crevasse density, glacier thickness or snow layer thickness. Full waveform
572 modelling combined with wave polarity analysis could also be conducted in order to further
573 understand how wave focusing in the near field domain as well as source heterogeneity and
574 directivity may cause heterogenous amplitude wavefields. Incorporating these constraints
575 into an improved model describing the control of both source and wave propagation physics
576 on the seismic wave amplitude field (Gimbert *et al.*, 2016) could allow using our dense array

577 observations to infer the spatial variability in subglacial water flow parameters such as
578 subglacial channel size and pressure.

579

580

581 PHYSICS OF STICK SLIP EVENTS

582

583 We demonstrate that dense array observations provide enhanced resolution on stick-slip
584 motion. Applying template matching on an array of sensor as opposed to on a single station
585 enables detecting many more events within clusters. Using the whole array for location
586 inversions also allows significantly reducing location uncertainties. Future studies may focus
587 on applying template matching across all sensors of the array in order to detect and locate
588 more events within clusters and potentially more clusters. With our present analysis we find
589 that events are all located in the down-glacier part of the array and in the central part of the
590 glacier or near the right side, while there is no event observed towards the left side (Fig. 7 and
591 8). This provides further observational support that specific bed conditions (e.g. water
592 pressure, bed shear stress, bed roughness, bed topography, carried sediments) are necessary
593 for these events to occur (Zoet *et al.*, 2013; Lipovsky *et al.*, 2019). Further insights into the
594 physics controlling the spatio-temporal dynamics of these events could be gained by
595 performing relative event location within each cluster using double-differences (Waldhauser
596 and Ellsworth, 2000) instead of simply inferring single cluster locations as presently done.
597 These improvements could allow identifying whether or not stick-slip asperities migrate.

598

599 USING MFP TO RETRIEVE SOURCES AND STRUCTURAL PROPERTIES

600

601 Systematic MFP analysis with adequate parametrization opens a route to continuous,
602 automatic, and statistics-based monitoring of glaciers. A wide diversity of seismic sources may
603 be identified and studied separately with this technique by scanning through different values
604 of MFP outputs. High MFP outputs may be used to study the dynamics of crevasse propagation
605 with particularly high spatio-temporal resolution. Such observations may allow to better
606 understand the underlying mechanisms associated with crack propagation, in particular
607 through providing an opportunity to better bridge the gap between laboratory and theoretical
608 material physics of crack propagation (van der Veen, 1998; Weiss, 2004) and crevasse
609 propagation under realistic glacier conditions in which water is expected to play an important
610 role (van der Veen, 2007). Lower MFP outputs may be used to locate spatially distributed
611 sources generating coherent signals over only a limited spatial extent. These distributed
612 sources may include tremor sources (e.g. water flow) or various glacier features (e.g.
613 crevasses, englacial conduits) acting as scatterers. One could also combine MFP with
614 eigenspectral decomposition to reveal weaker noise sources that would otherwise be hidden
615 within the background noise (Seydoux *et al.*, 2016). Additional constraints for seismic imaging
616 may also be provided through identifying specific events generating indirect arrivals of
617 particular interest for structural analysis, such as in bed-refracted waves shown in Fig. 5 (black
618 arrows).

619

620 **SUMMARY**

621

622 We present a dense seismic array experiment made of 98 3-component seismic stations
623 continuously recording during 35 days in early spring 2018 on the Argentière Glacier, French
624 Alps. The seismic dataset is supplemented by complementary observations obtained from

625 ground penetrating radar, drone imagery, GNSS positioning and in-situ instrumentation of
626 basal glacier sliding velocities and subglacial water flow discharge. We show that a wide range
627 of glacier sources and structure characteristics can be extracted through multiple seismic
628 processing techniques such as spectral analysis, template matching, matched-field processing
629 and eikonal wave tomography. Future studies focusing more specifically on each aspect of the
630 herein presented observations may yield novel quantitative insights into spatio-temporal
631 changes in glacier dynamics and structure.

632

633 **DATA AND RESOURCES**

634 Raw seismic data can be found at:

635

636 Roux, P., Gimbert, F., & RESIF. (2021). *Dense nodal seismic array temporary experiment on*
637 *Alpine Glacier of Argentière (RESIF-SISMOB)* [Data set]. RESIF - Réseau Sismologique et
638 géodésique Français. <https://doi.org/10.15778/RESIF.ZO2018> (see also link
639 http://seismology.resif.fr/#NetworkConsultPlace:ZO%5B2018-01-01T00:00:00_2018-12-31T23:59:59%5D).

641

642

643 Processed data used in this paper can be found at:

644

- 645 - <https://doi.org/10.5281/zenodo.3701519> for meteorological, subglacial water flow
646 discharge and glacier sliding speed data
- 647 - <https://doi.org/10.5281/zenodo.3971815> for bed thickness, surface elevation, nodes
648 positions, crevasses positions, surface velocity, noise PSDs, event occurrences and
649 locations derived from template matching for stick-slip events and MFP for englacial
650 fracturing events
- 651 - <https://doi.org/10.5281/zenodo.3556552> for drone orthophotos

651

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662

663

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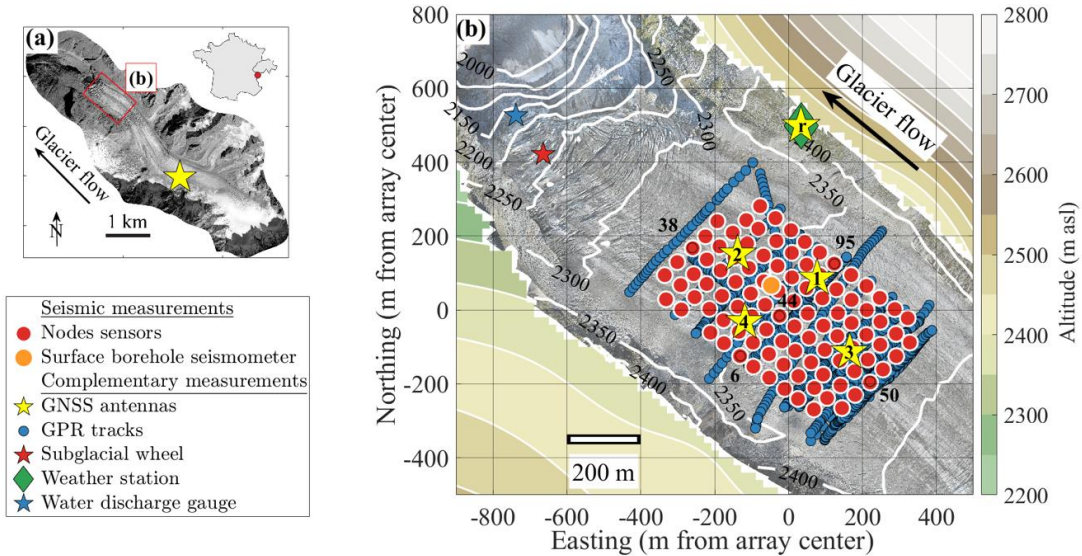
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1 **FIGURES**

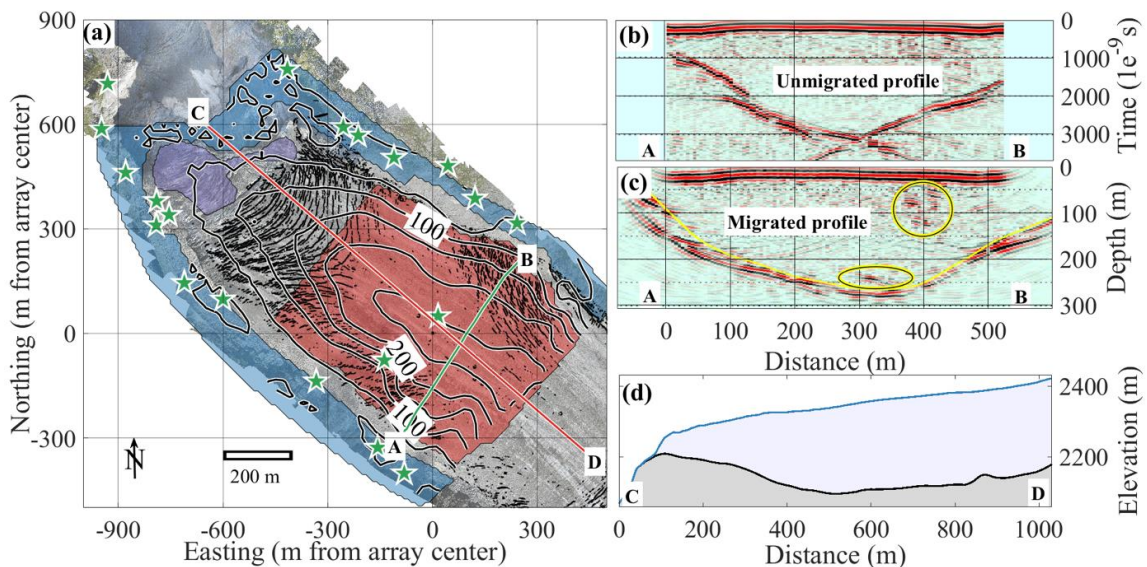
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3

4 Fig. 1: Maps of the Argentière Glacier and of the instruments deployed during the dense array
 5 experiment (see associated legend for symbols correspondence). (a) Aerial picture of the
 6 Argentière Glacier taken in 2003. The red rectangle indicates the area shown in Fig. 1(b), which
 7 we focus on in this study. The yellow star refers to a permanent GNSS station and the red dot
 8 in the inset shows the location of the glacier with respect to French and Swiss borders. (b) Map
 9 showing the lower part of the Argentière Glacier along with instrument positions. White
 10 contours indicate glacier surface topography as retrieved from structure from motion, and color
 11 contours indicate topography outside of the glacier. Symbols refer to instruments as specified
 12 in the legend. Numbers associated with red circles indicate nodes that are used for illustrative
 13 examples in Fig. 4.

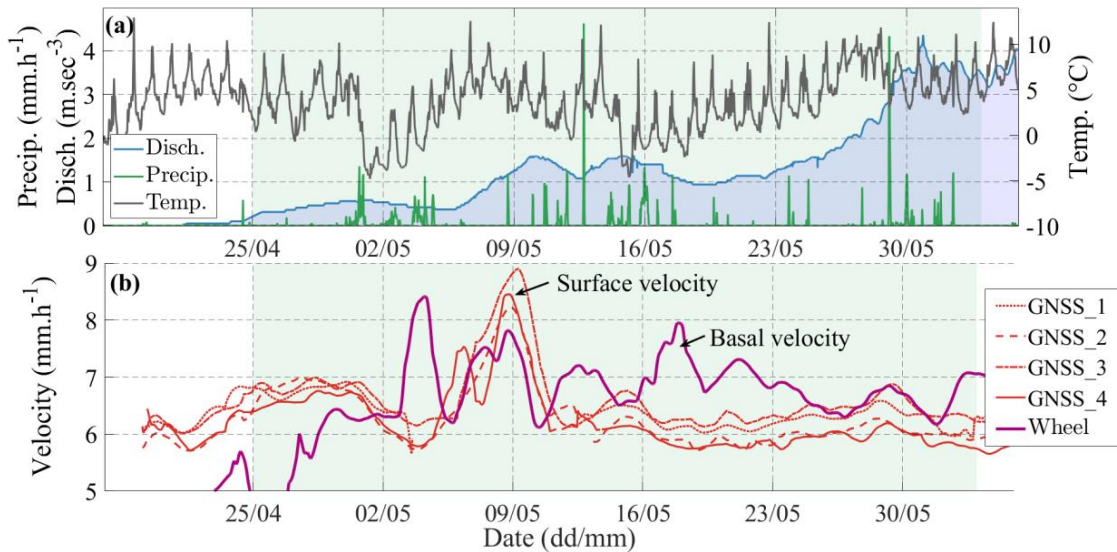
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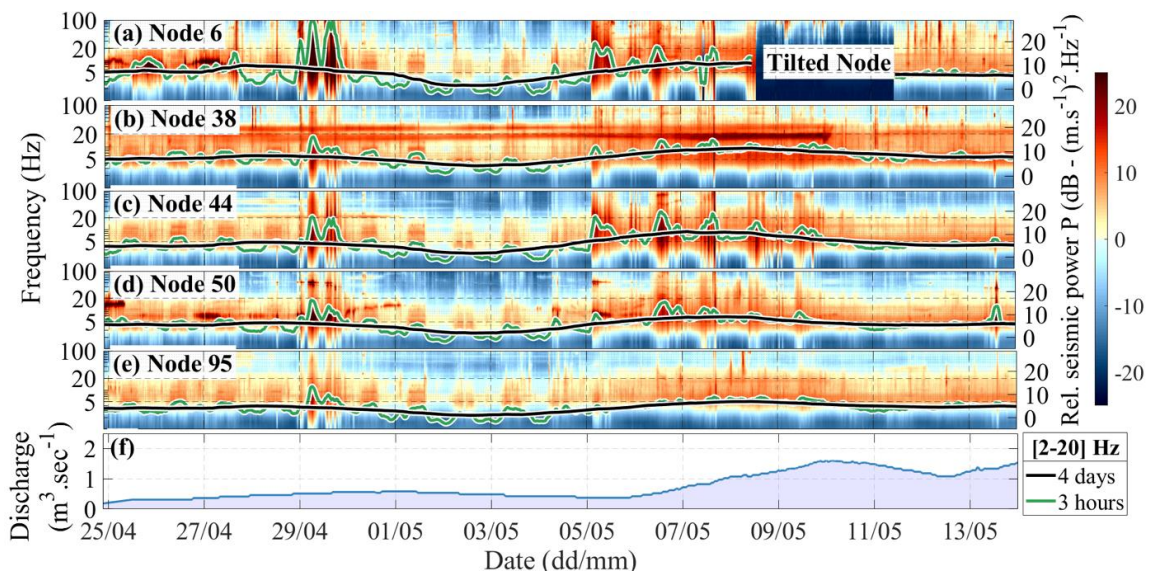
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16 Fig. 2: (a) Ice thickness (black contours) and surface crevasse (black dots) maps. Green stars
 17 correspond to the GNSS measured ground control points, while colored areas differentiate
 18 between observations used to constrain the bed DEM: the blue area is from a 2018 surface
 19 DEM, the purple area corresponds to regions where ice-bed coordinates are known from in-situ

20 borehole measurements and from excavated tunnels, and the red area corresponds to a region
 21 where glacier depth is inferred from the GPR measurements. The green line shows the track
 22 associated with the selected GPR profile shown in (b) and (c). The red line shows the profile
 23 shown in (d). (b) and (c) Examples of processed (b) unmigrated and (c) migrated GPR data
 24 acquired along the AB profile shown in (a). The yellow curve corresponds to the picked
 25 interface and the yellow ellipses highlight local reflectivity anomalies. (d) Surface elevation
 26 and bed elevation along the CD profile shown in (a).
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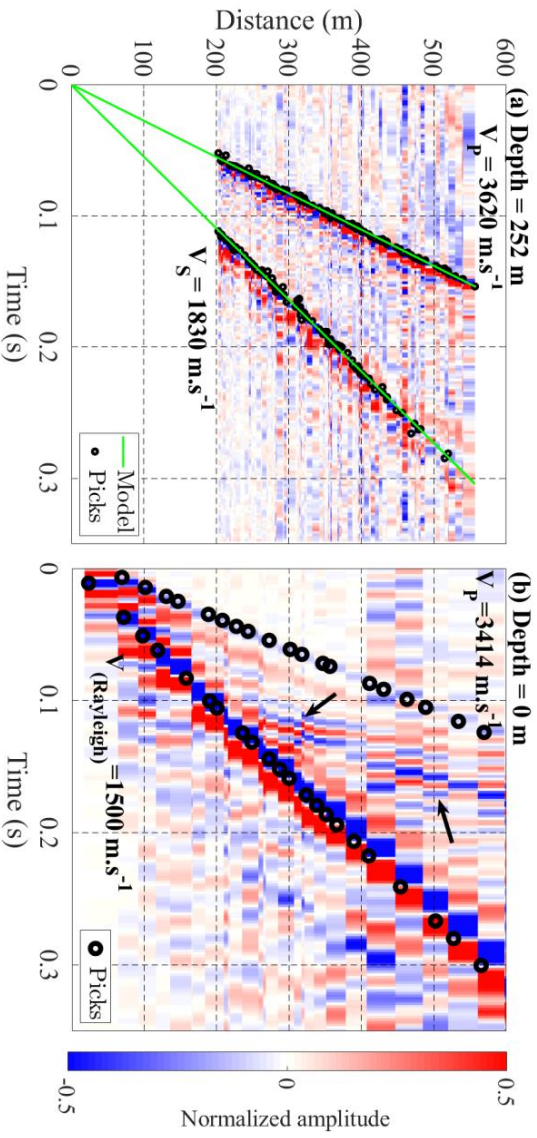


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 31 Fig. 3: Time series of meteorological, hydrological and ice dynamic parameters during the
 32 dense array experiment (from 25 April to 6 June 2018, shaded in green). (a) Proglacial water
 33 discharge (blue), surface temperature (grey) and precipitation (green). (b) Horizontal ice surface
 34 measured with GNSS stations (orange lines) and basal sliding measured with at the ice-bed
 35 interface (thick purple line). See Nanni *et al.* (2020) for longer time series over the 2016-2018
 36 period.
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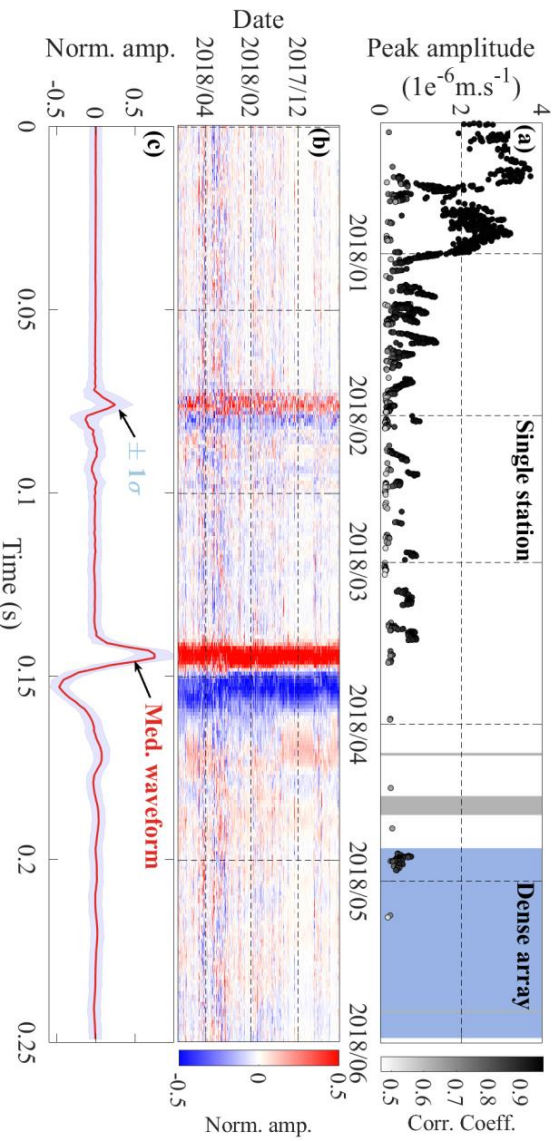


39

40 Fig. 4: (a-e) Vertical-component spectrograms calculated at five selected stations across the
 41 array (see corresponding numbers in Fig. 1) from 25 April to 5 May 2018. Curves indicate 2-
 42 20 Hz frequency-averaged seismic power as smoothed over short (3 hours, green lines) and
 43 long (4 days, black lines) periods. See Fig. S1 for spectrograms over the whole period and all
 44 stations. (f) Proglacial water discharge timeseries (same as in Fig. 3).
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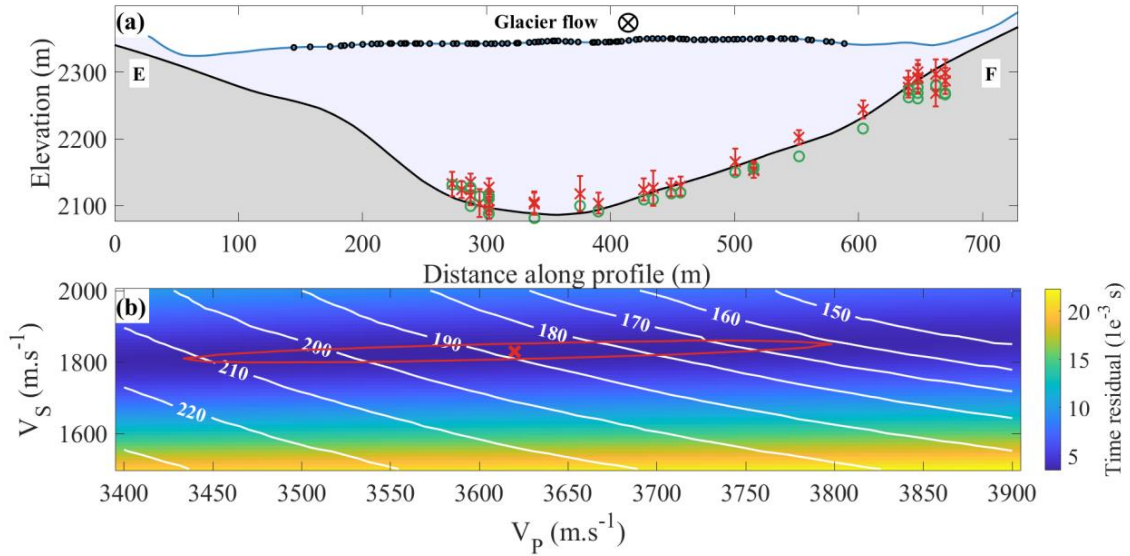


46 Fig. 5: Vertical-component seismicograms of (a) a basal event identified from template matching
 47 and (b) a surface event identified from match-field-processing. Corresponding event locations
 48 are shown in Fig. 8. Black circles correspond to picked P, S and Rayleigh arrival times and
 49 green lines on (a) correspond to predicted arrival times using a P-wave velocity of $3620 \text{ m}\cdot\text{sec}^{-1}$
 50 and an S-wave velocity of $1830 \text{ m}\cdot\text{sec}^{-1}$. A hyperbolic arrival is also visible at large offsets on
 51 panel b, likely corresponding to a refracted or reflected wave (black arrows). Time origin
 52 corresponds to the inferred source time.
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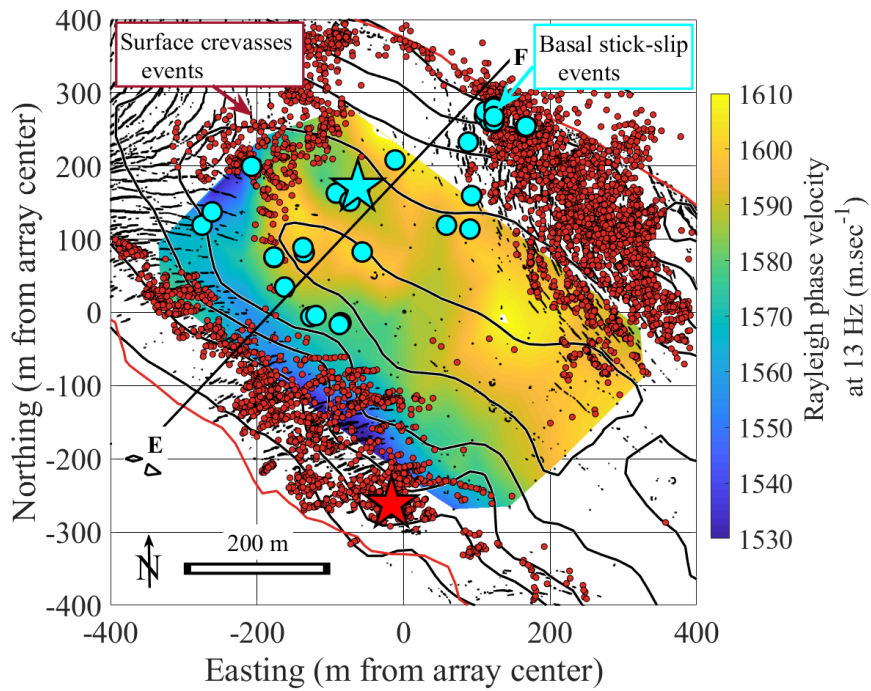
56 Fig. 6: a) Vertical-component time series of peak amplitude for one cluster of repeating basal
 57 events. The grey scale indicates correlation with the template signal. Grey areas indicate gaps
 58 in the data and the blue area highlights the time period spanned by our dense-array experiment.
 59

60 (b) Waveforms of all events of a cluster normalized by peak amplitude (using the North
 61 component of the borehole station). The color bar indicates normalized waveform amplitude.
 62 Each horizontal line represents one event. Time origin corresponds to the source time. (c)
 63 Median seismogram of all events shown in panel (b).
 64

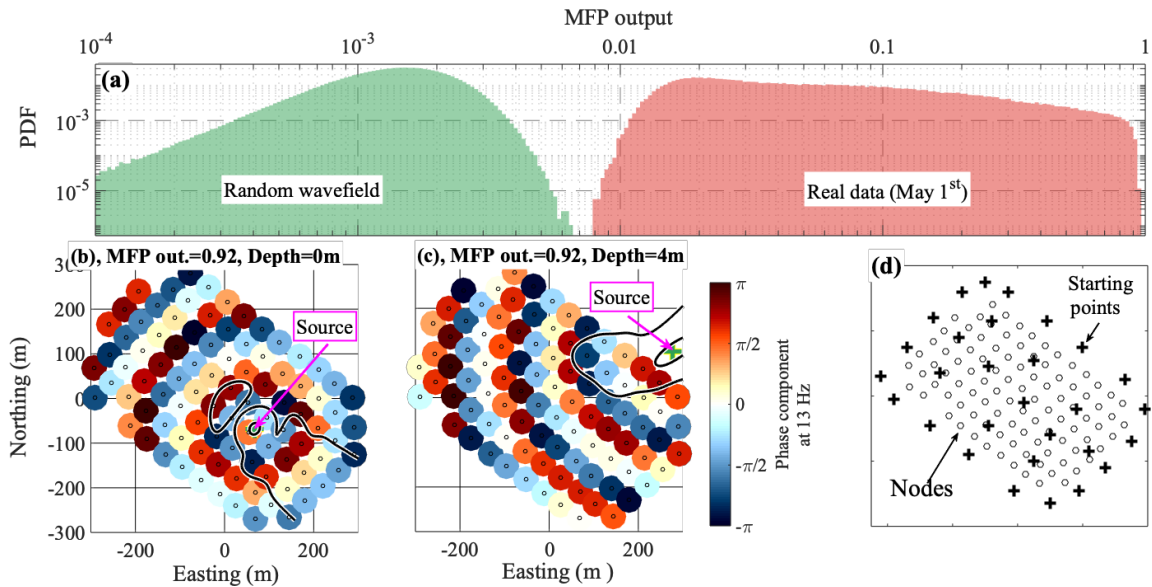


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 67 Fig. 7: (a) Two-dimensional representations of stick-slip event locations (red crosses) along the
 68 EF profiles shown in Fig. 8. Red error bars show the 95% confidence interval. Green circles
 69 indicate the projected depth at the exact location of each event. Black dots show node positions.
 70 (b) Average time residuals (background color coded image) and average icequake depth (white
 71 contours) as a function of the seismic wave velocities V_P and V_S used to locate basal icequakes.
 72 The red cross indicates the velocities $V_P=3620 \text{ m}\cdot\text{sec}^{-1}$ and $V_S=1830 \text{ m}\cdot\text{sec}^{-1}$ that minimize the
 73 average time residuals. The red line delineates the range of V_P and V_S associated with an average
 74 residual that is smaller than 105% of the minimum value.

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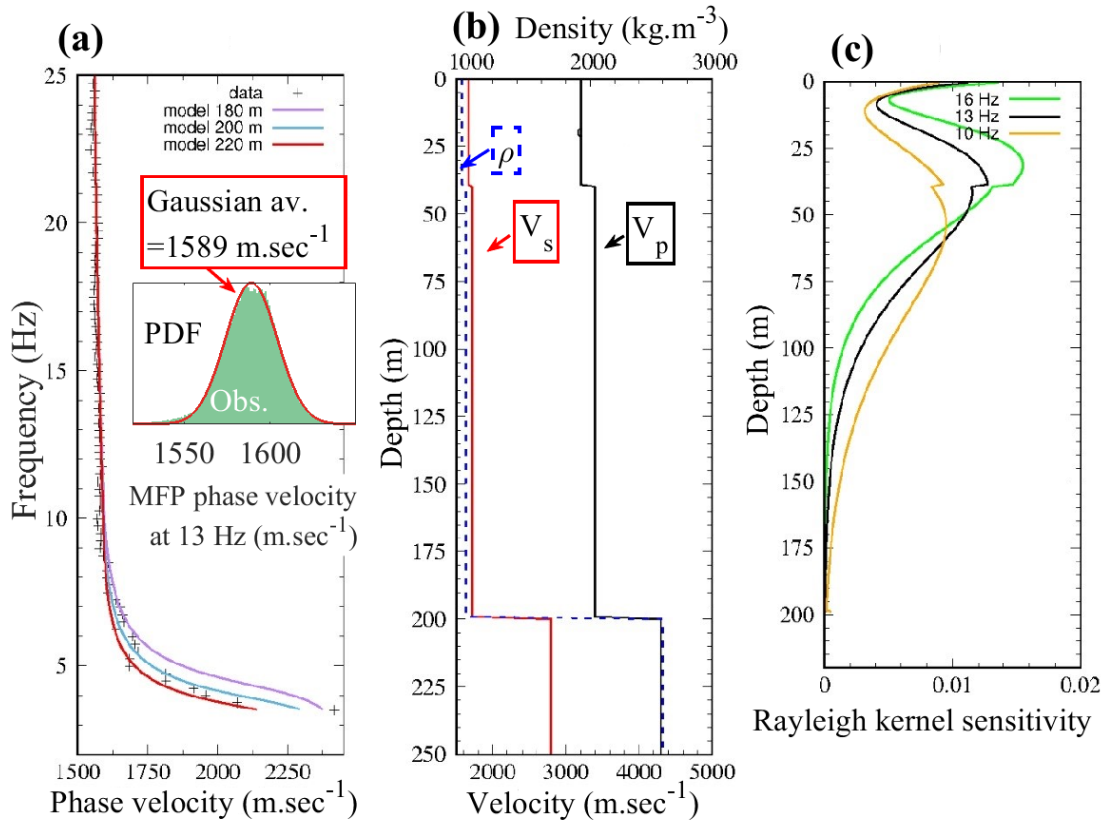


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 79 Fig. 8: Map showing the positions of basal stick-slip clusters (filled blue circles) and of icequake
 80 events (filled red circles, using signal at 13 Hz and sources with MFP output higher than 0.8).
 81 The colored area shows phase velocities from Rayleigh-wave travel-time tomography at 13 Hz.
 82 The CD profile refers to the profile used in Fig. 7 and the blue and red stars refer to the events
 83 shown in Fig. 5(a) and 5(b), respectively. Black dots show crevasses, contour lines show ice
 84 thickness (m) and the red lines delineate the glacier extent.
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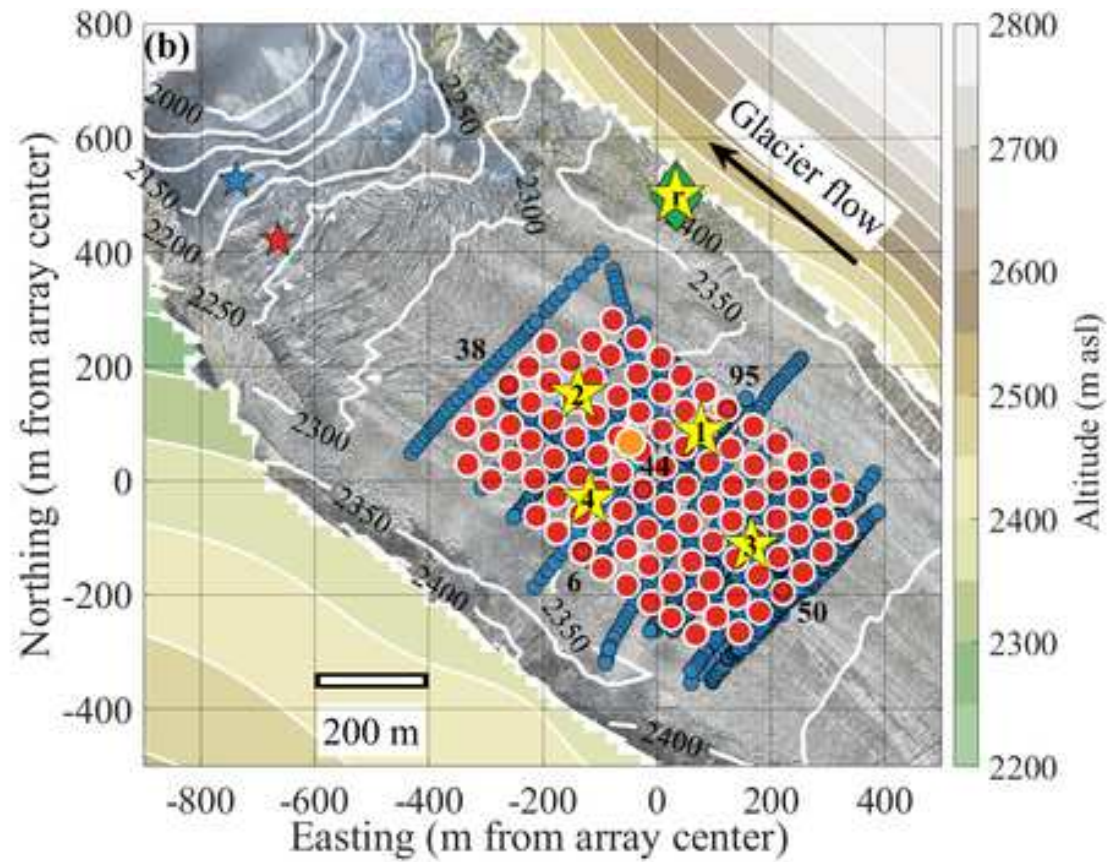
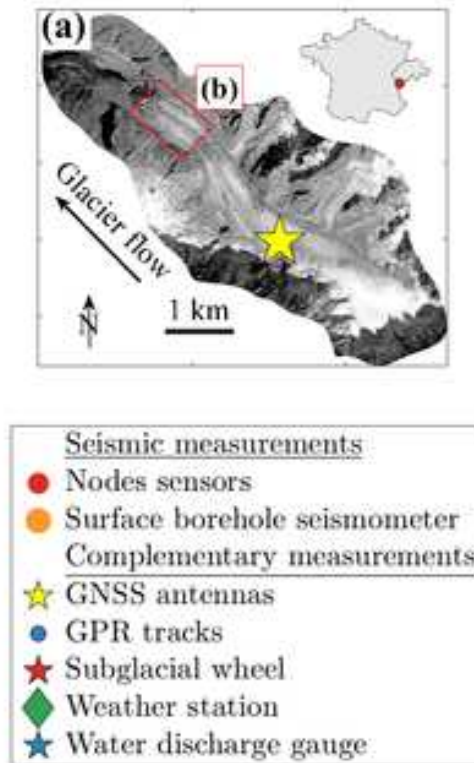
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 88 Fig. 9: (a) Probability distribution functions (PDF) of MFP outputs obtained at 13 Hz when
 89 applying MFP on one day (1 May 2018) of real data (red) and on a numerically-generated
 90 random wavefield (green). The bottom panels (b) and (c) show the phase fields observed over
 91 a 1-s time window at 13 Hz for two selected events. Locations obtained from MFP using our
 92 minimization process are shown by the pink arrow/green crosses, while the contour lines show
 93 0.1 and 0.8 MFP outputs iso-contours calculated by applying a grid search over the glacier

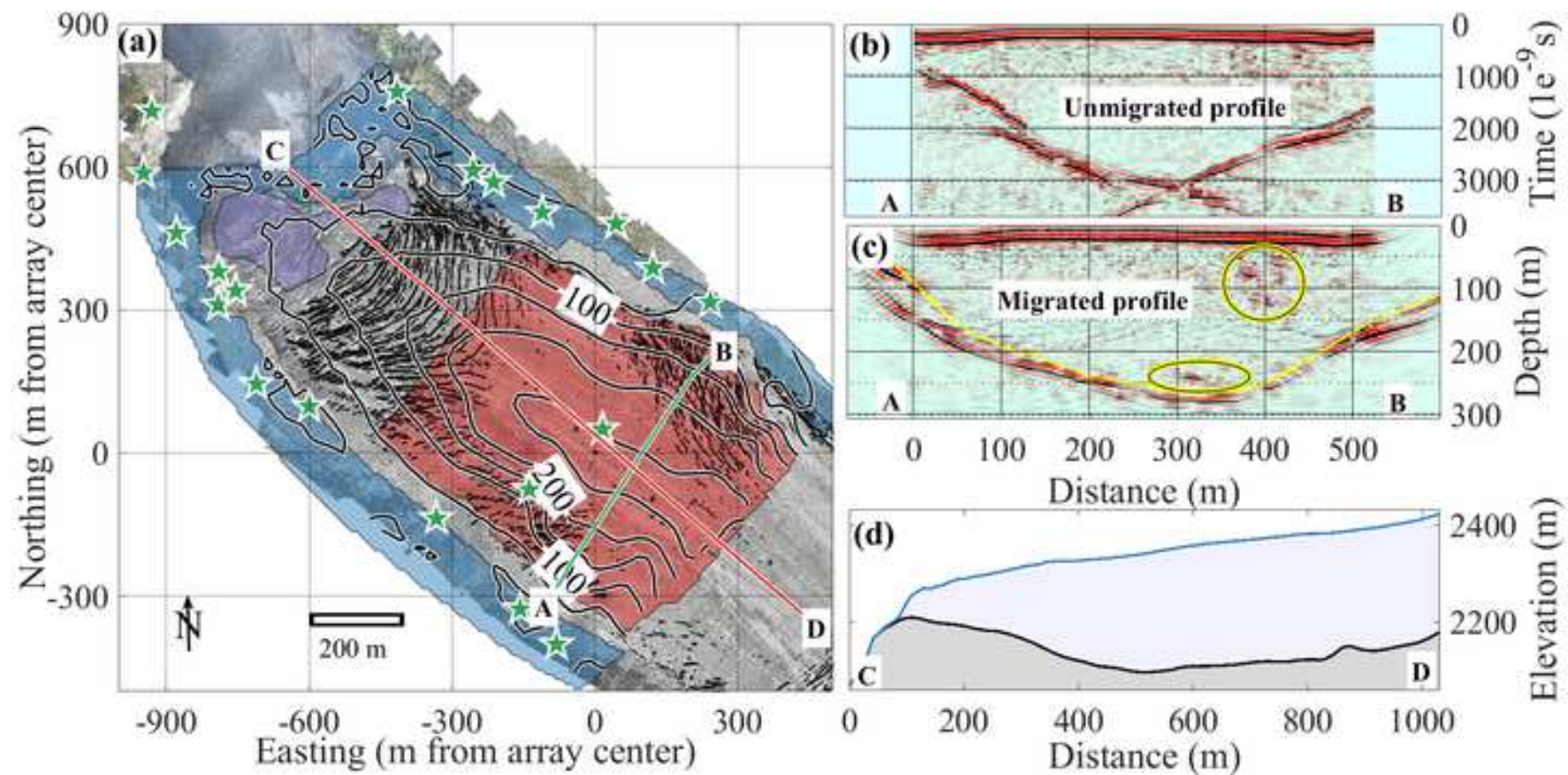
94 surface. Panel (d) shows the locations of the 29 starting points (black crosses) used for MFP
 95 along with node positions (black circles).
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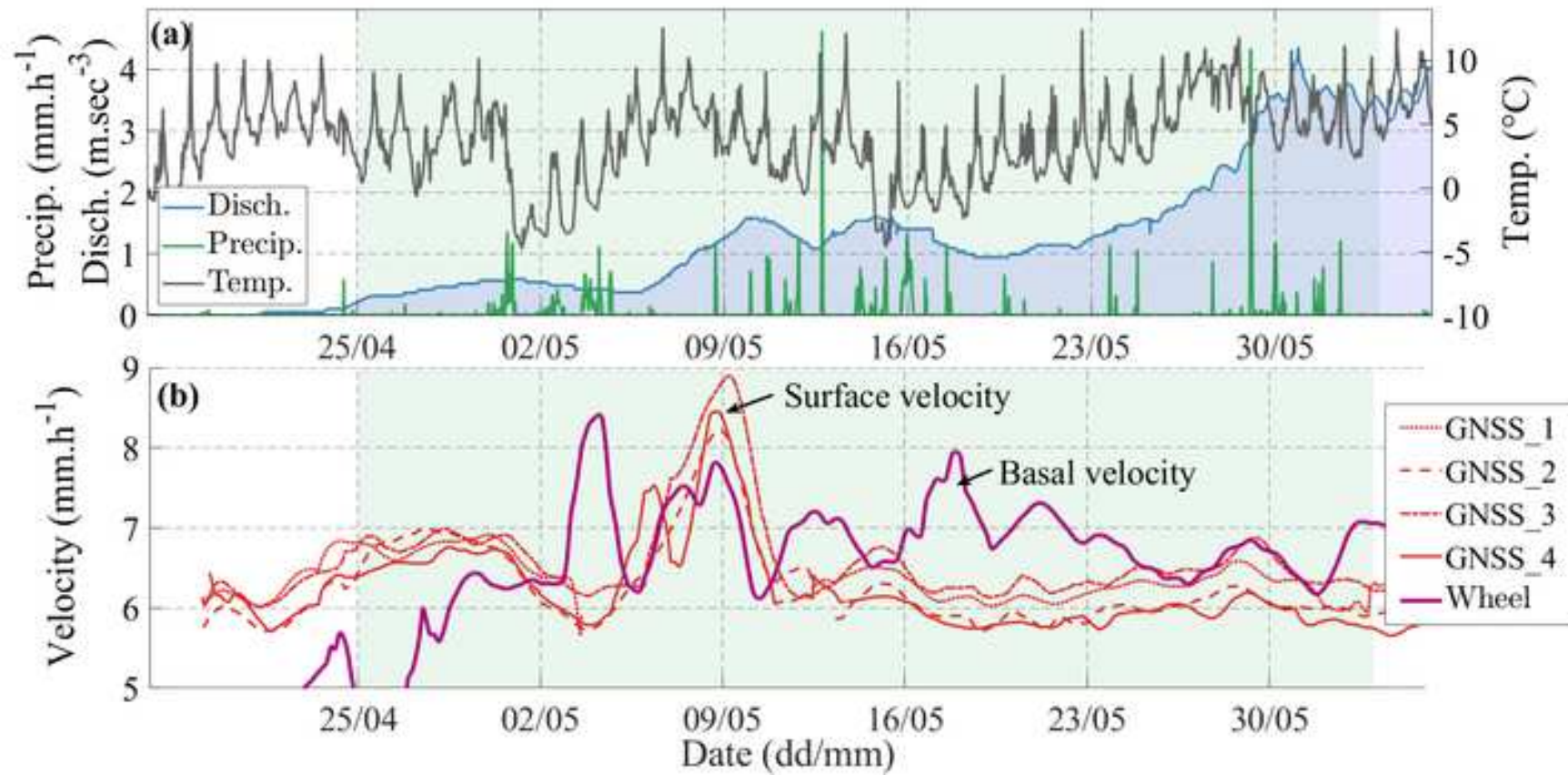


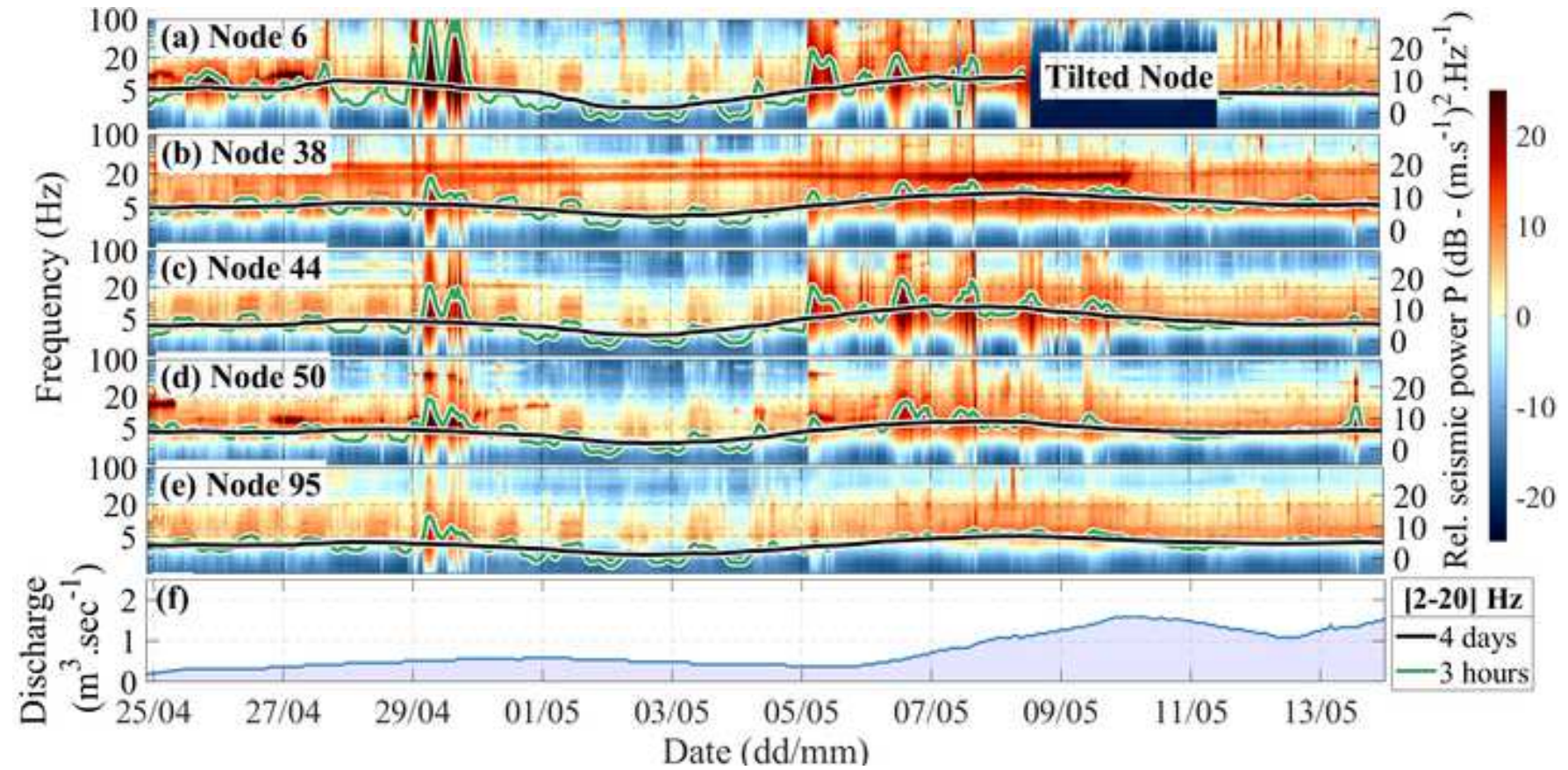
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 98 Fig. 10. Inversion of an average one-dimensional structure using an average surface
 99 dispersion curve. (a) Comparison between the observationally-derived dispersion curve (black
 100 crosses) and synthetic Rayleigh wave dispersion curves computed using the elastic model
 101 displayed in (b) using glacier thicknesses of 180 m (purple), 200 m (blue) or 220 m (red). The
 102 inset shows the distribution of phase velocity obtained from match-field-processing at 13Hz
 103 (green) along with a Gaussian fit (red). The central value of the gaussian fit is used to establish
 104 the dispersion curve. (b) Synthetic model used to predict the observed dispersion curve. (c)
 105 Sensitivity kernels of Rayleigh waves as a function of depth for 16 Hz (green), 13 Hz (black)
 106 and 10 Hz (orange) associated with the glacier model shown in (b).
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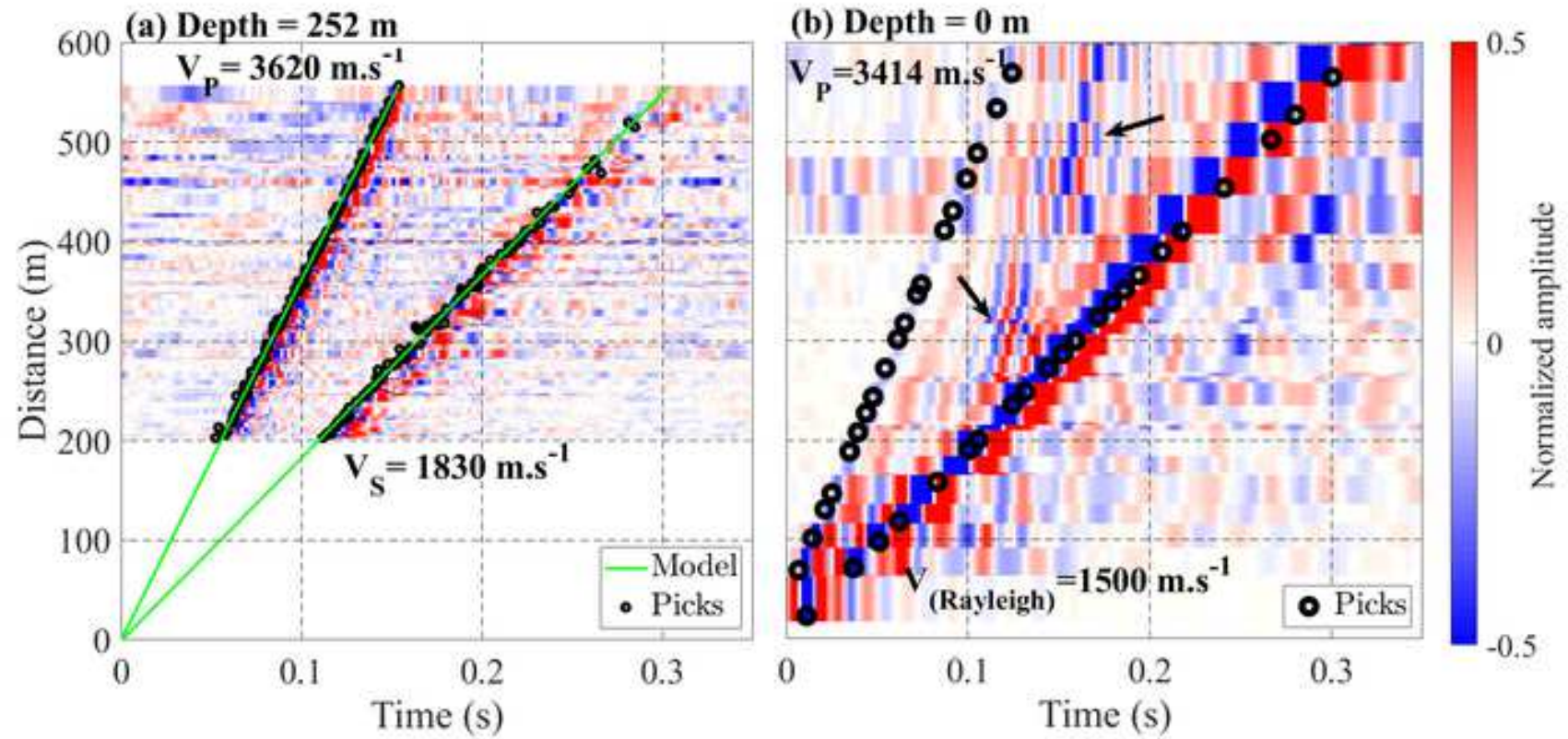
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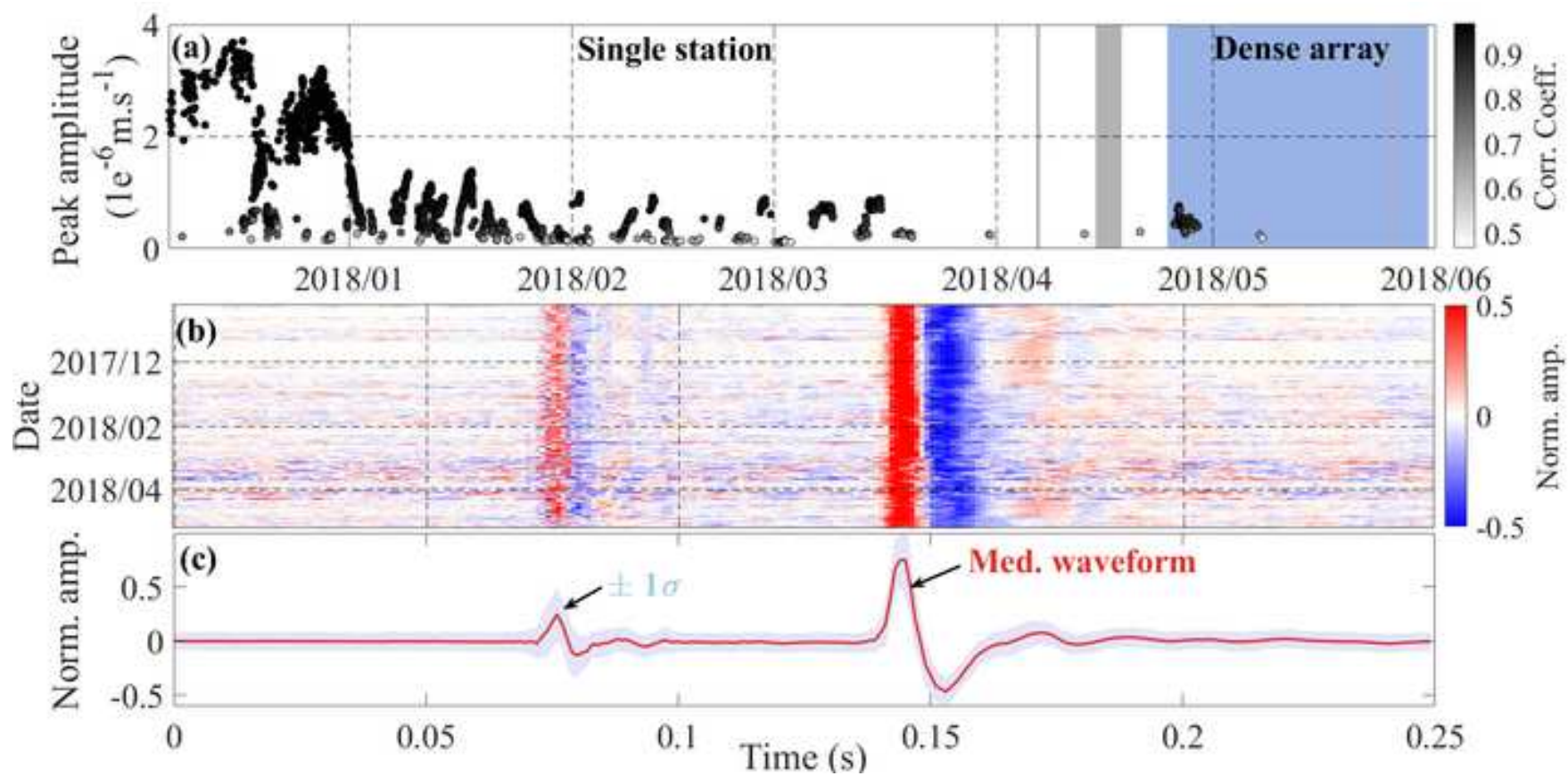


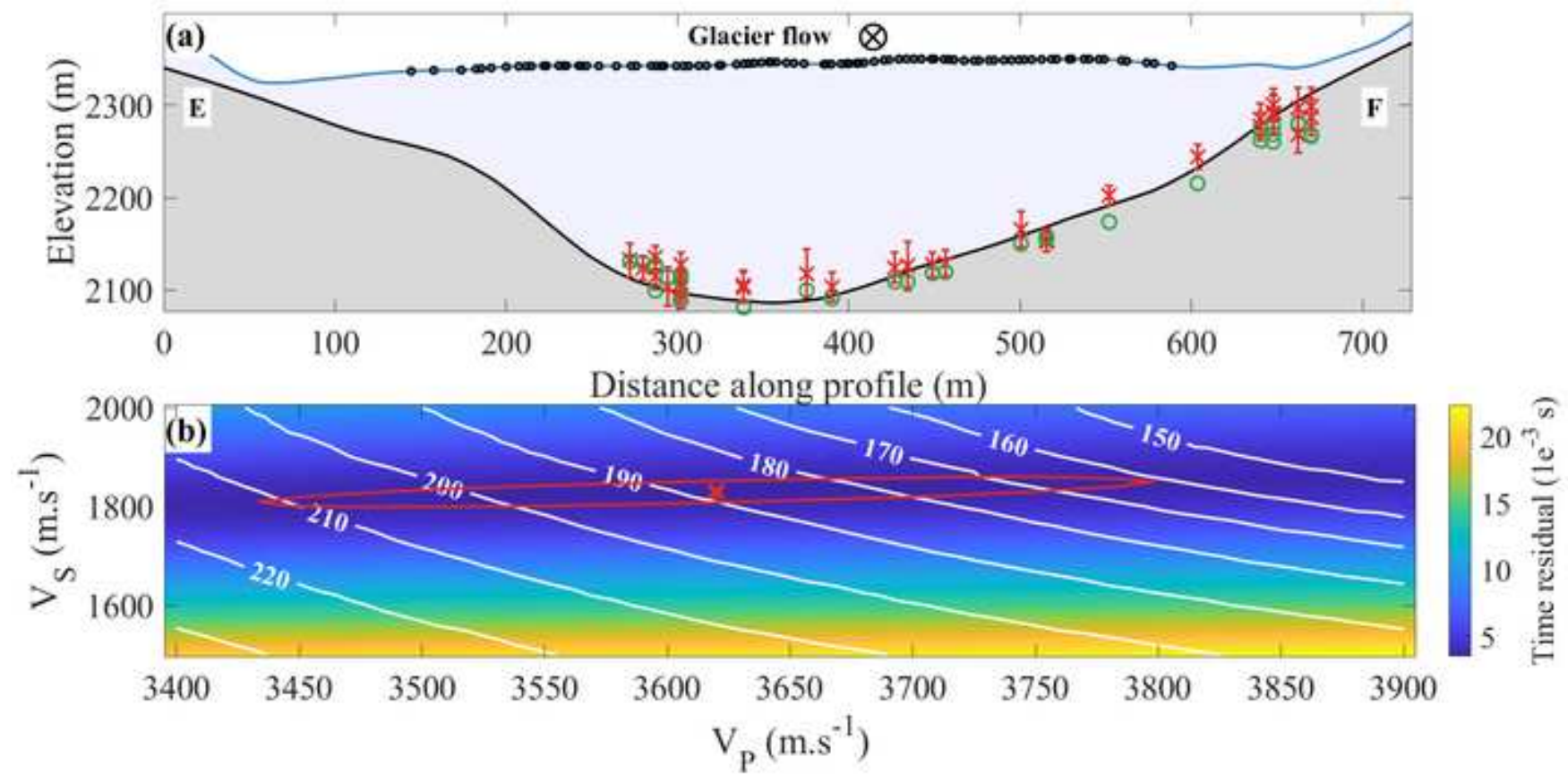


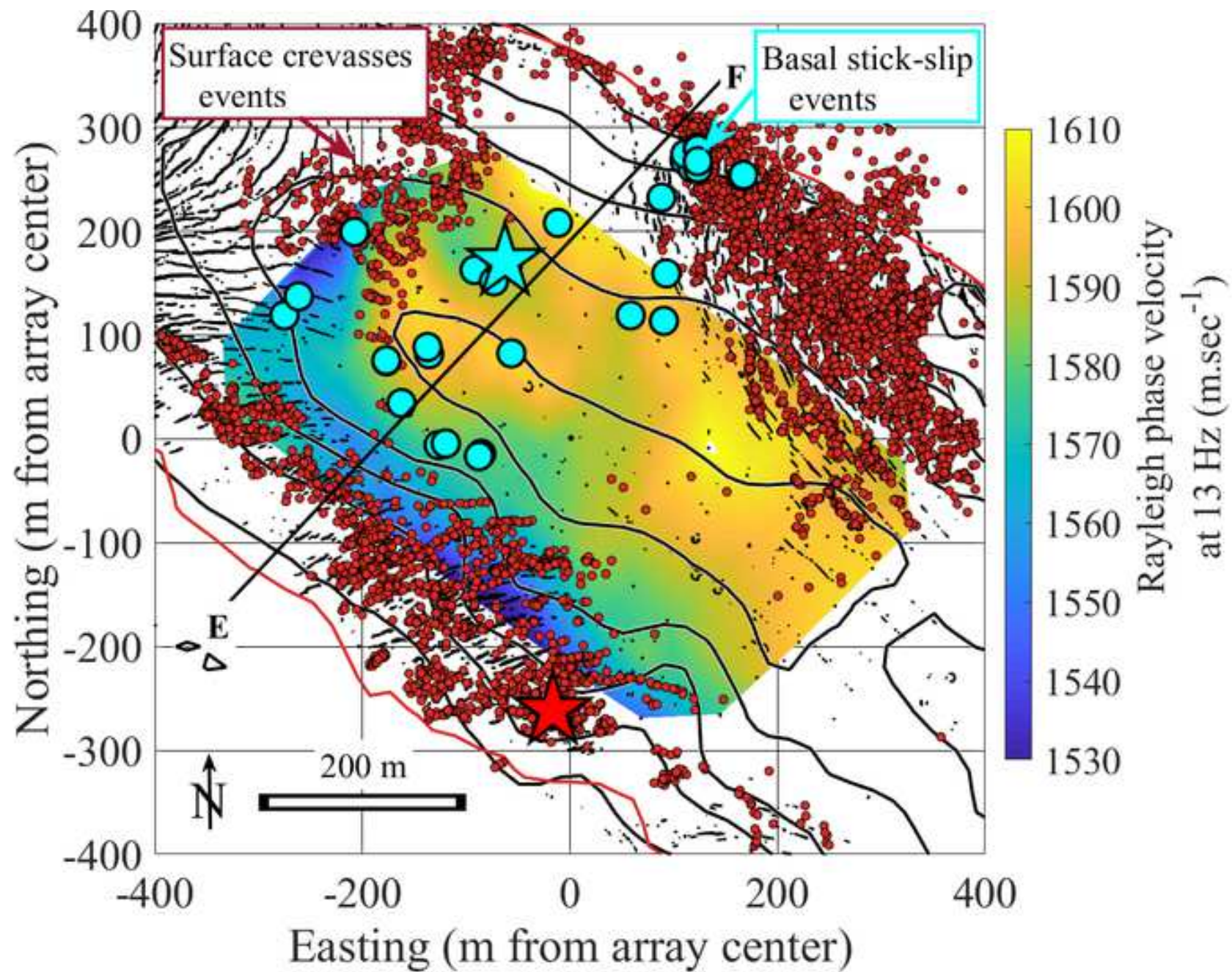


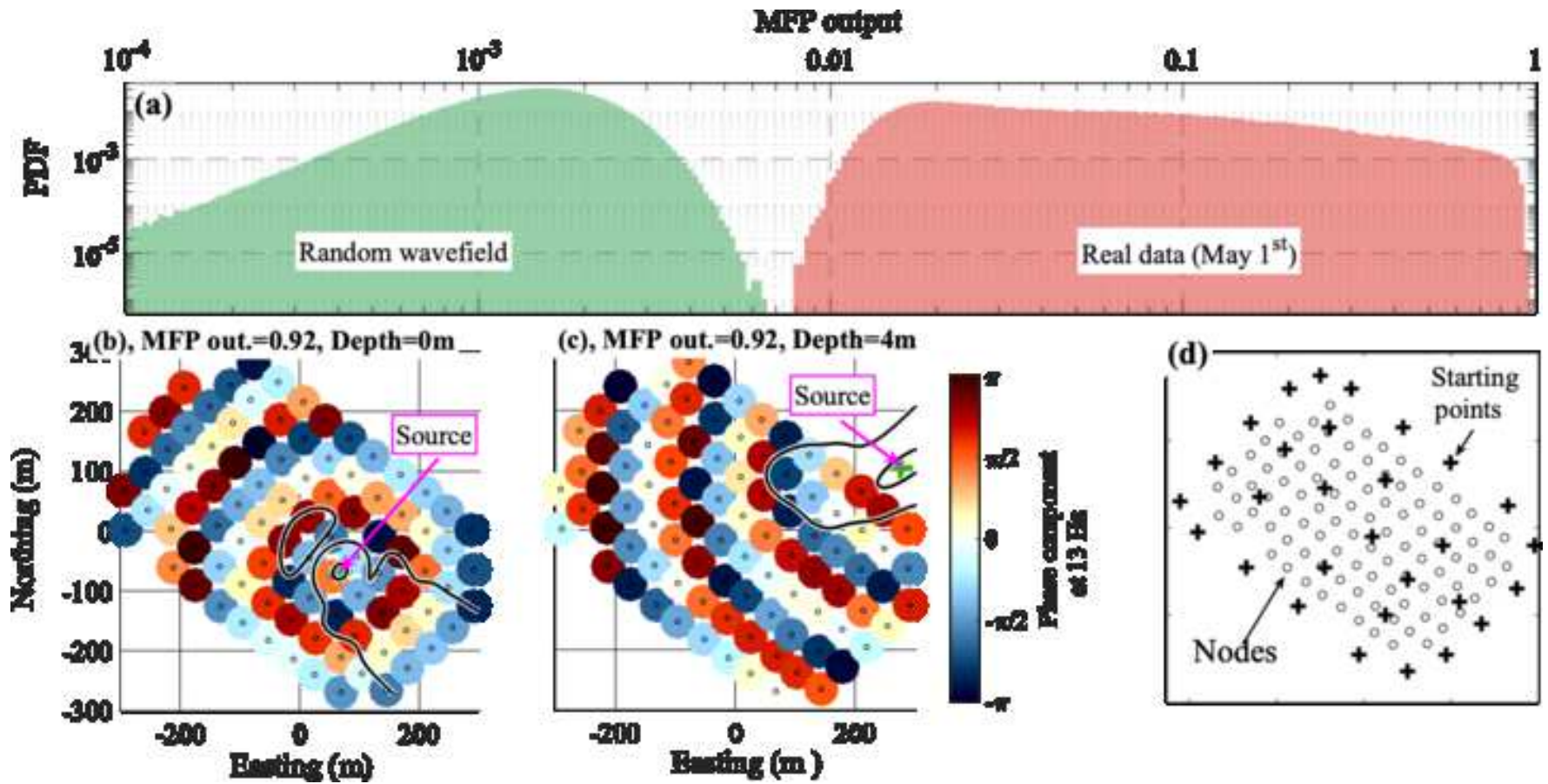


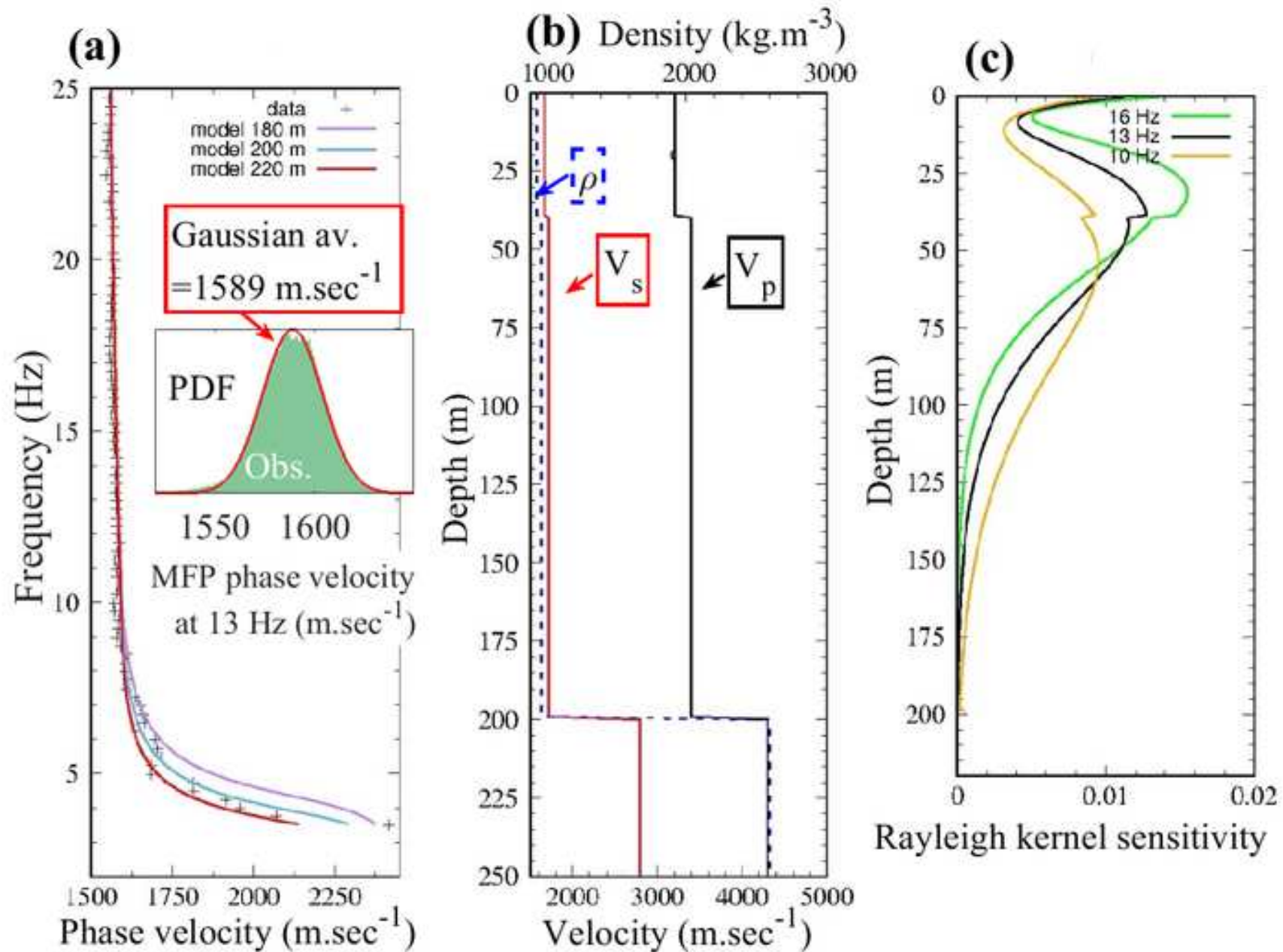


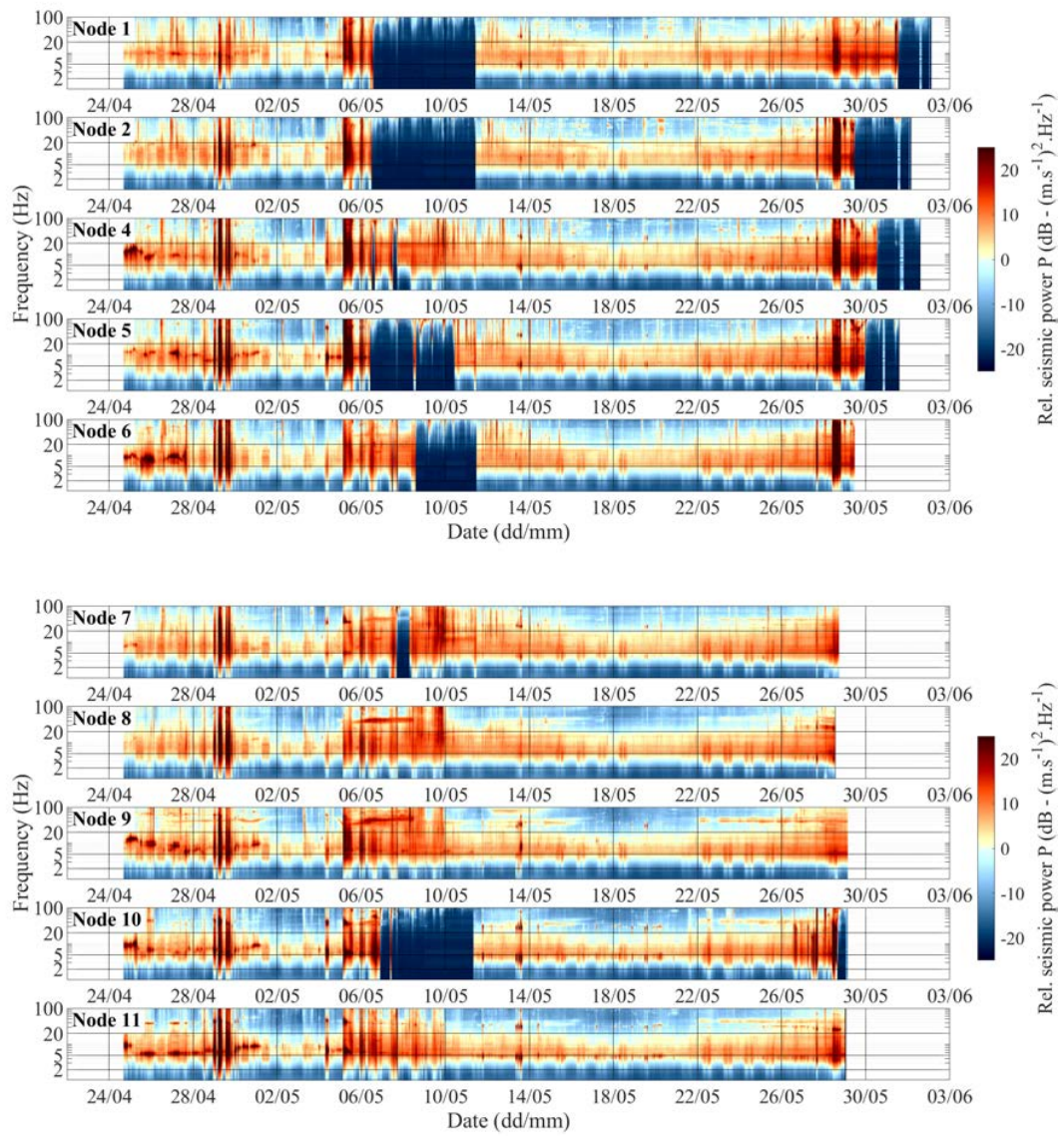


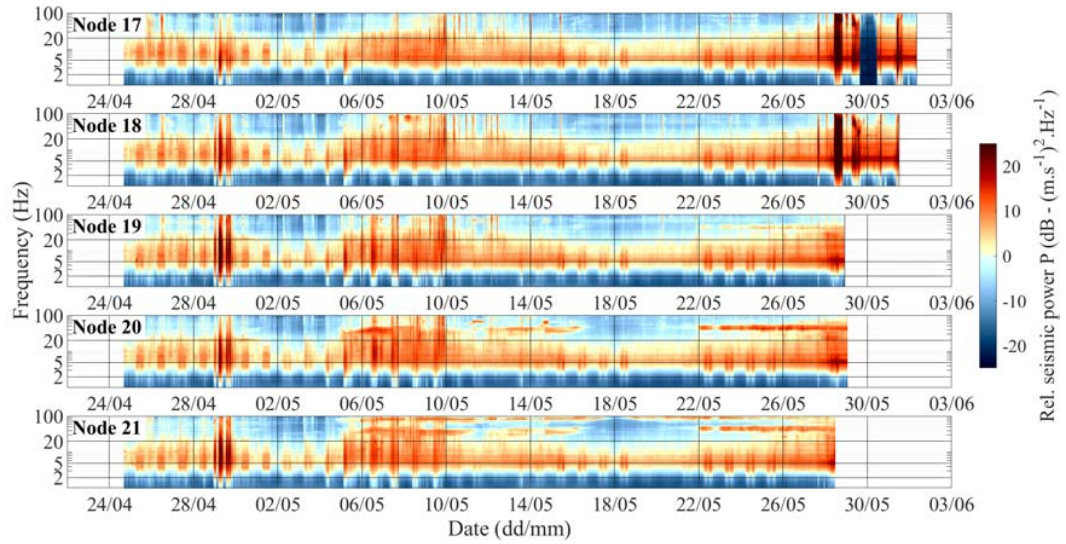
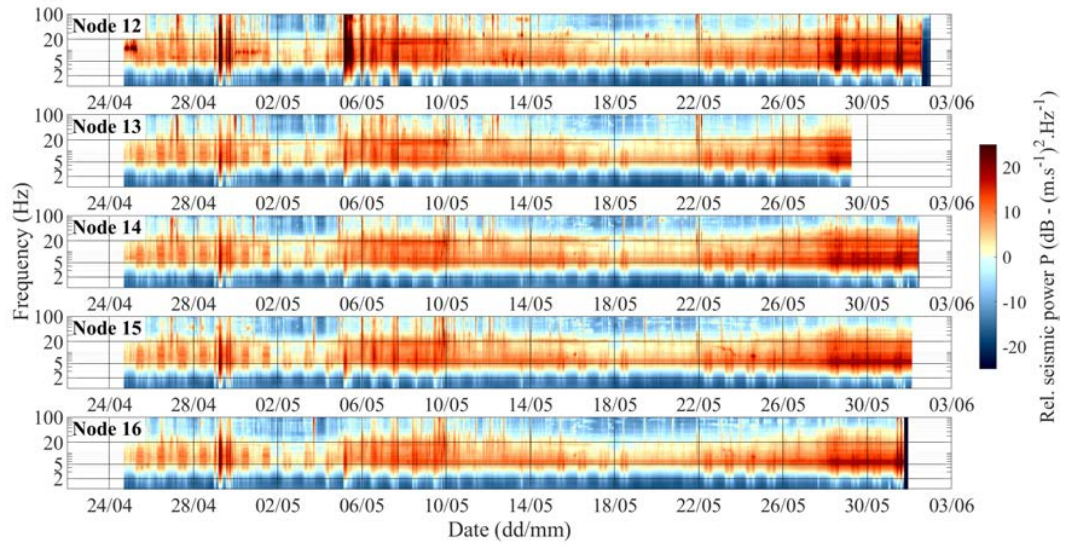


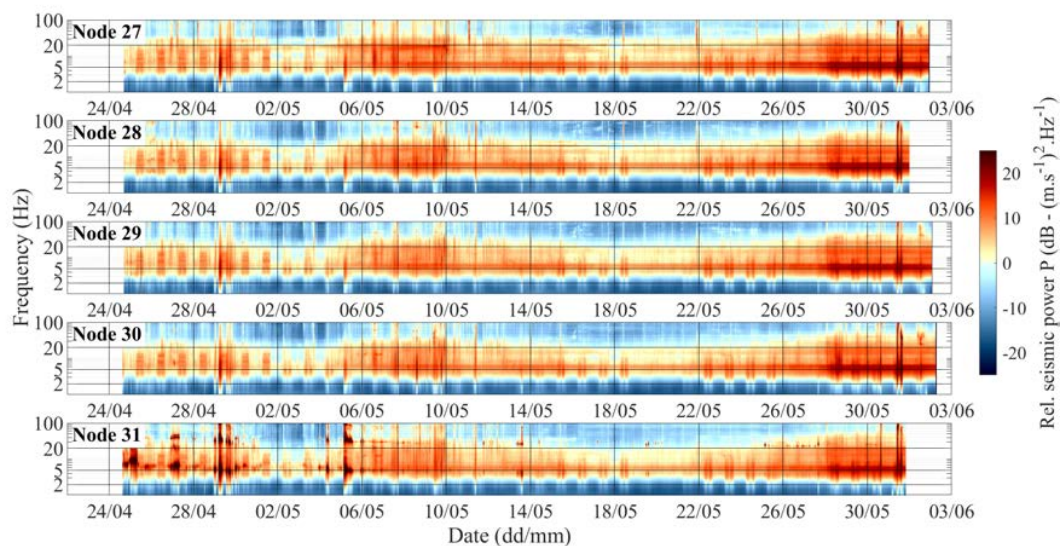
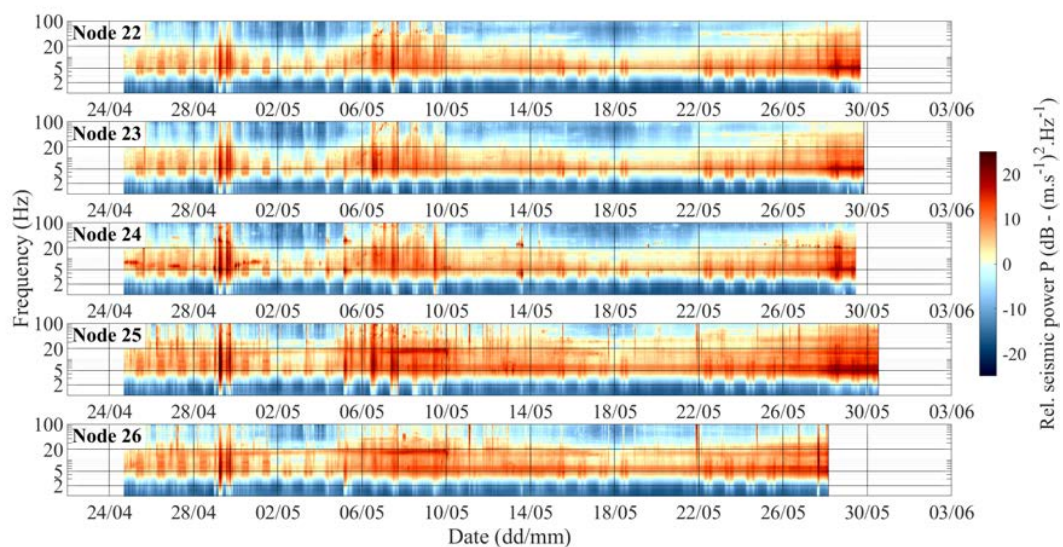


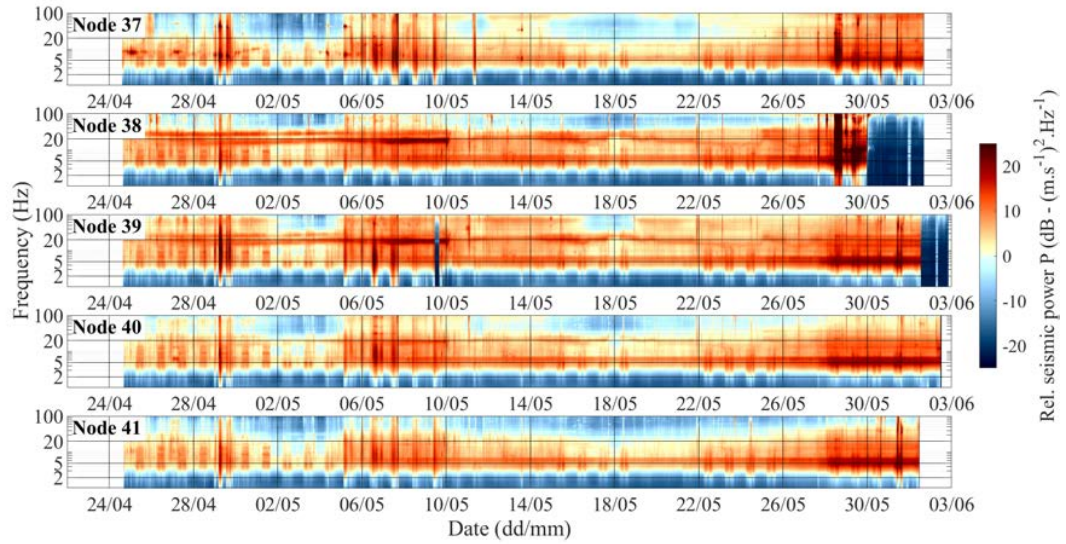
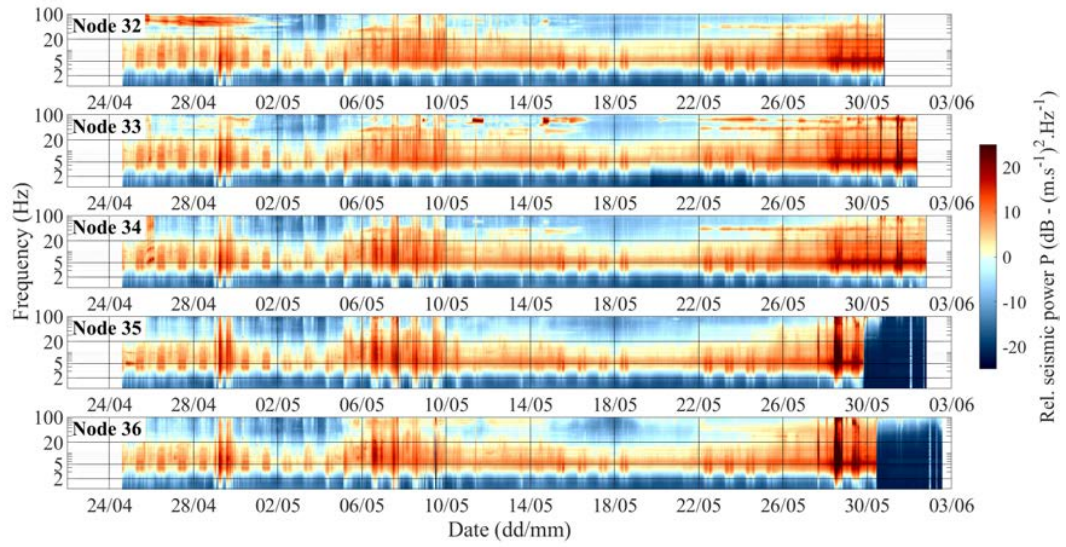


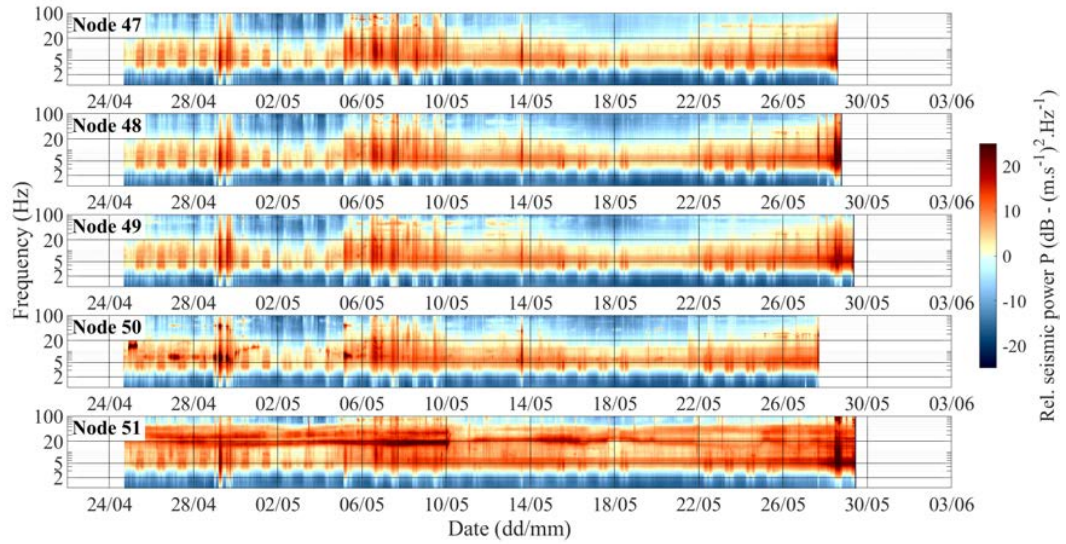
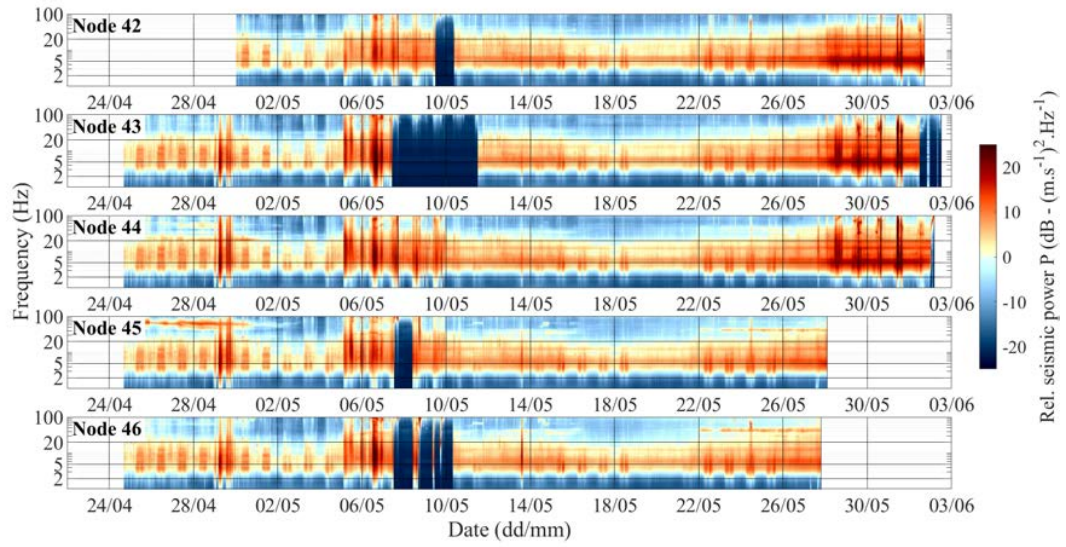


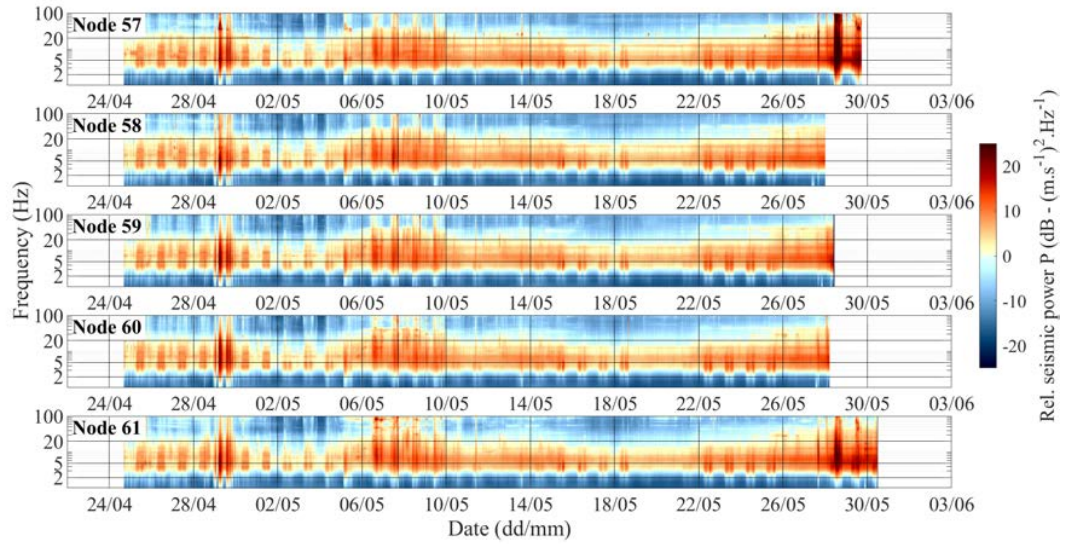
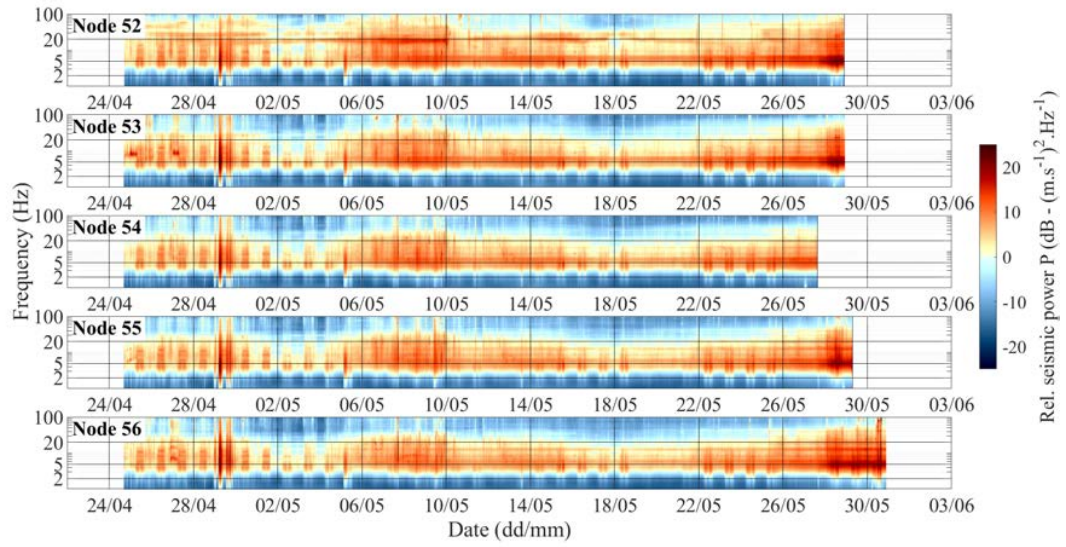


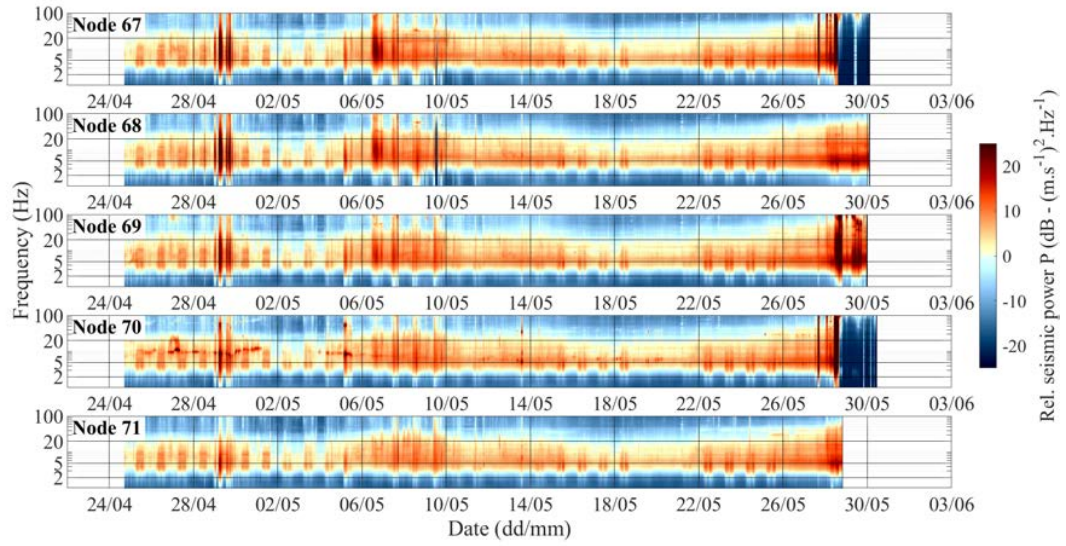
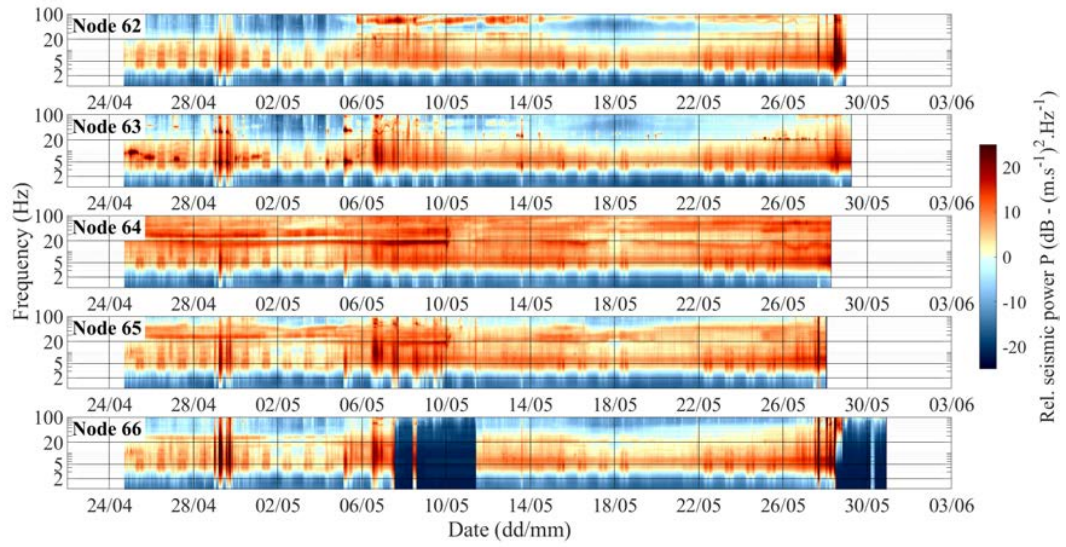


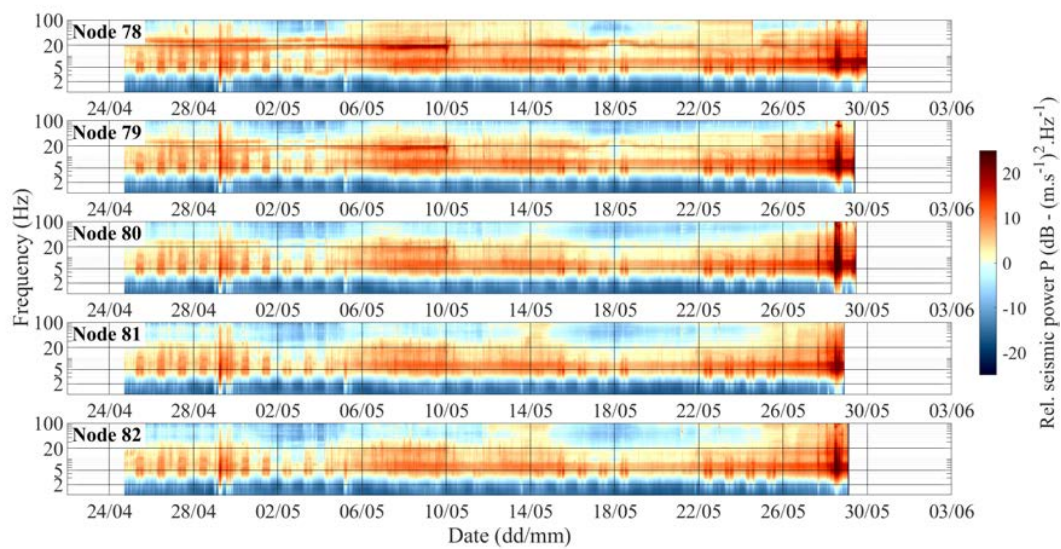
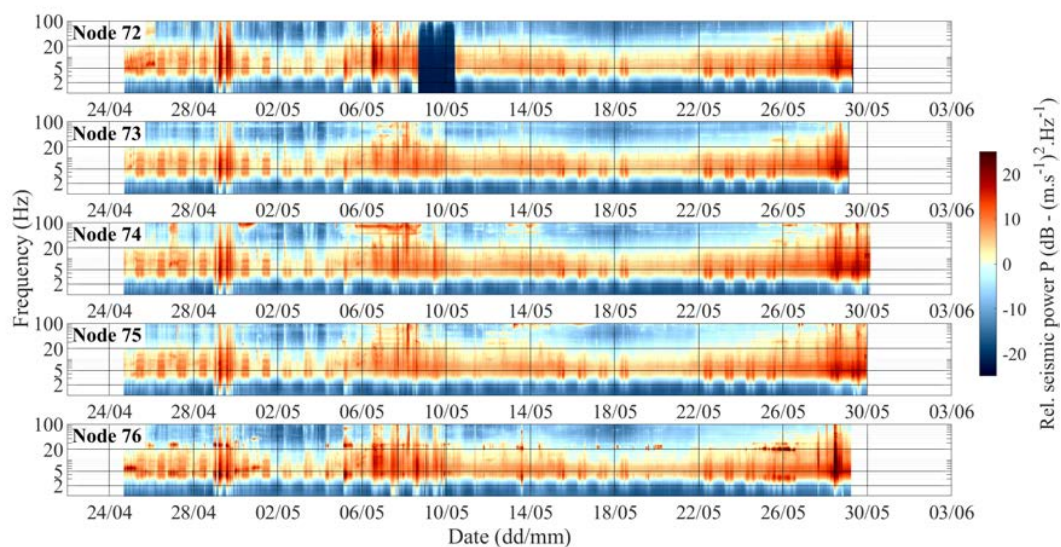


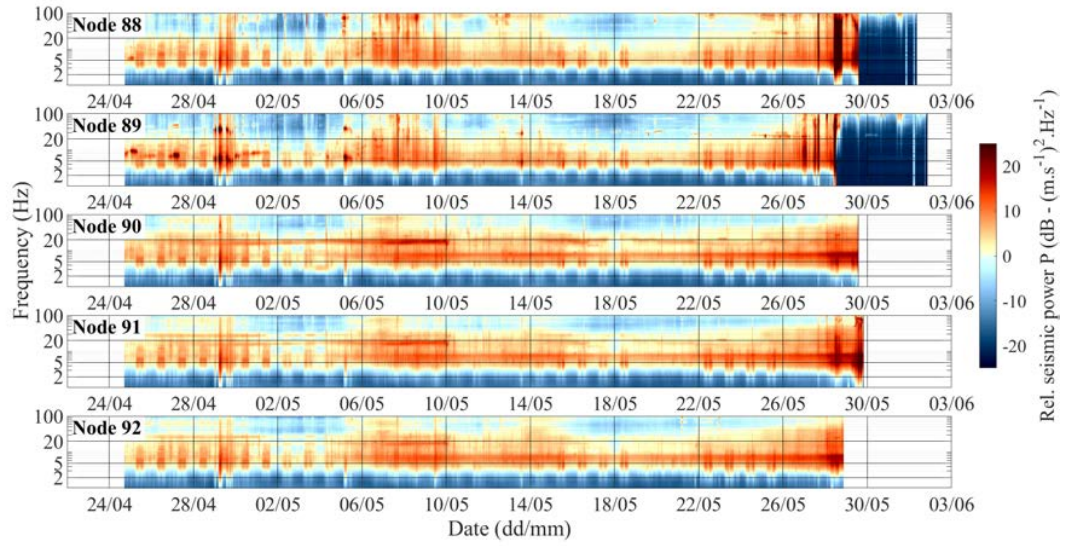
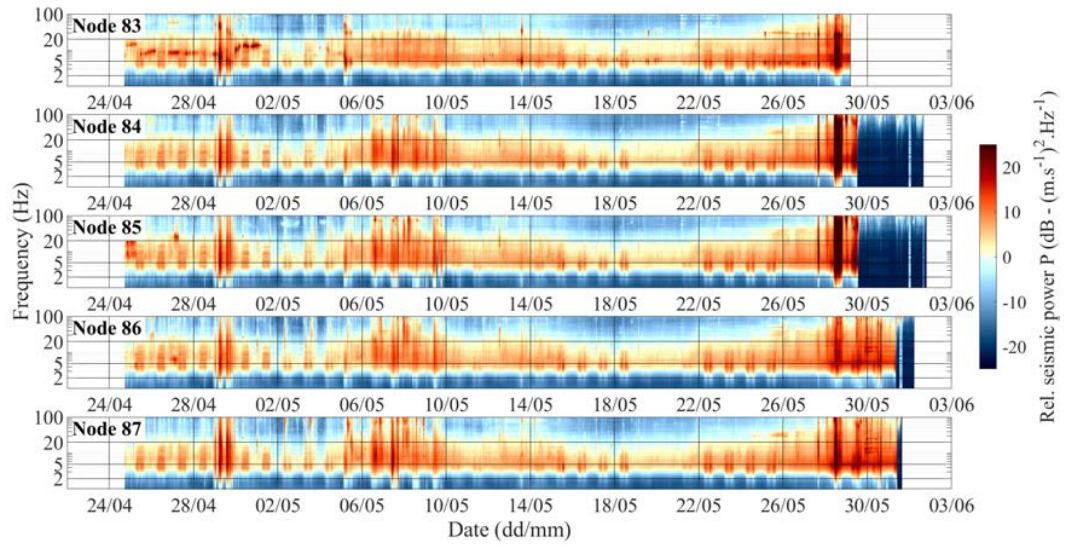












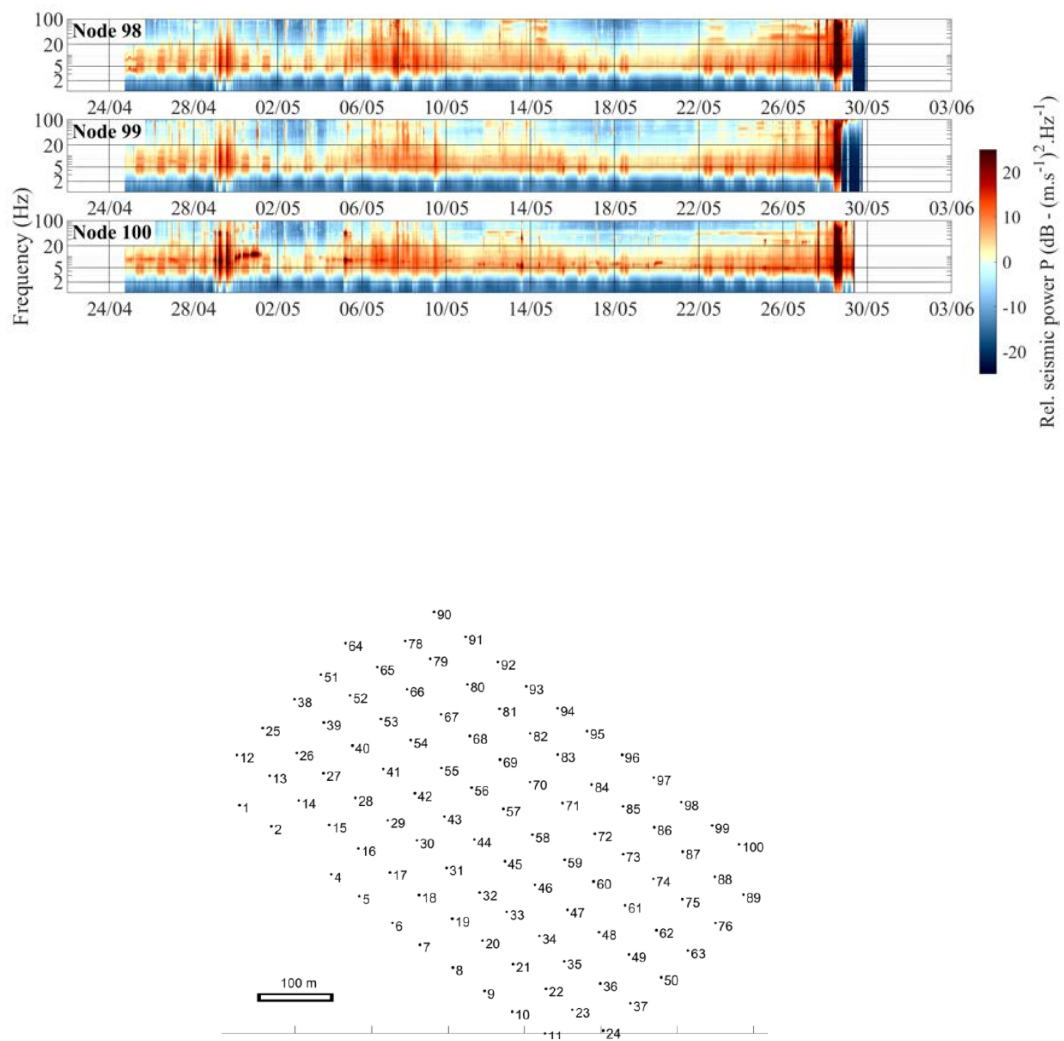


Fig. S1: Spectrograms calculated at all stations over the whole period. The lower panel shows a map view of the node number nomenclature. The reader should refer to Fig. 1 for absolute positioning of the dense array. The color scale represents seismic power, as calculated in Fig. 4, and is identical across all panels.