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1 Combined use of remote-sensing and ground geophysical
2 techniques to investigate geological control on landslides
3 (Trièves area, Western Alps, France)

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17 **Abstract**

This paper presents a study combining remote sensing and seismic prospecting for investigating two clayey landslides located in the Trièves area (Western French Alps). Although affecting similar slopes made of similar clay layer under the same meteorological conditions, the two adjacent Avignonet and Harmalière landslides exhibit, since the 1980s, major differences in morphology, displacement rate magnitudes and motion directions. These observations suggest the control of at least one additional internal factor on the landslide characteristics. Combination of geological mapping, airborne Light Detection and Ranging (LiDAR) data, aerial photographs, Global Positioning System (GPS) and seismic noise measurements allows characterizing the landslide

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morphology and the thickness of the soft layer down to the seismic substratum, providing a 3D view of the soft layer bottom. GPS measurements and digital photographs reveal that the difference in kinematics between the two earthslides can be tracked back to 60 years ago at least. The Avignonet slide is mainly directed towards the East while Harmalière is mainly oriented towards SE. The LiDAR scan map illustrates this differential motion and morphology between the two slides and highlights that the Harmalière slide is still presently much more active than the Avignonet one. A ground geophysical prospecting based on ambient noise measurements allowed to map the base of the clays and indicates the presence of a N-S ridge of hard sediments (Jurassic bedrock and/or compact alluvial layers) on the eastern side of the Avignonet landslide. This ridge disappears when approaching the Harmalière landslide and makes place to what can be interpreted like a NW-SE oriented palaeovalley of the river Drac. It is proposed that the ridge could act as a butress which could mechanically prevent the Avignonet landslide from evolving as fast as the Harmalière one. Furthermore the NW-SE palaeovalley located under the Harmalière landslide corresponds to the motion direction of the slide. It is then finally proposed that the slides different behaviours are partly controlled by the palaeotopographic setting of lake Trièves. These results suggest a major influence of the bedrock palaeotopography on the kinematics of the two landslides.

¹⁸ *Keywords:* Clay landslides, LiDAR, GPS, aerial photographs,
¹⁹ microtremors, differential kinematics, palaeotopographic control

²⁰

21 1. Introduction

22 . This paper presents a study combining remote sensing and seismic prospect-
23 ing for investigating two clayey landslides (Avignonet and Harmalière) lo-
24 cated in the Trièves area (Western Alps; Fig. 1). Although affecting similar
25 slopes made of similar clay layer under the same meteorological conditions,
26 these two adjacent landslides exhibit, since the 1980s, major differences in
27 morphology, displacement rate magnitudes and motion directions. These ob-
28 servations suggest the control of at least one additional internal factor on the
29 landslide characteristics. Such control by lithological variations, hydrogeolog-
30 ical drainage or structural features have been regularly depicted in clay slides
31 (Bonci et al., 2004; Lapenna et al., 2005; Eilertsen et al., 2008; Bozzano et al.,
32 2008). This study aims to explain the significant differences in geometry and
33 kinematics observed between the landslides of Avignonet and Harmalière.
34 Combination of geological mapping, airborne Light Detection and Ranging
35 (LiDAR) data, aerial photographs, Global Positioning System (GPS) and
36 seismic noise measurements allows characterizing the landslide morphology
37 and the thickness of the soft layer down to the seismic substratum, providing
38 a 3D view of the soft layer bottom. This reveals the palaeotopography upon
39 which settled the clays and helps to assess the geological control on the two
40 landslides behaviour.

41 . In the last two decades, remote sensing techniques and geophysical prospect-
42 ing methods have been increasingly used to image landslide structures at the
43 surface and at depth, respectively. Remote sensing techniques, such as radar
44 interferometry, high-resolution optical images correlation and multi-temporal
45 laser scanning, allow the surface displacement field to be measured, which

46 is a key parameter to understand the landslide mechanics (for a review, see
47 [Metternicht et al., 2005](#); [Delacourt et al., 2007](#)). In complement to conven-
48 tional ground-based geodetic techniques and GPS surveys that provide only a
49 limited number of discrete measurements, multi-temporal remote sensing im-
50 agery has the potential to measure nearly continuous displacement rate field
51 over landslides (among others, [Fruneau et al., 1996](#); [Rott et al., 1999](#); [Kimura
52 and Yamaguchi, 2000](#); [Squarzoni et al., 2003](#); [Metternicht et al., 2005](#); [Strozzi
53 et al., 2005](#); [Corsini et al., 2007](#)). Nonetheless, performances of the remote
54 sensing techniques strongly depend on the site conditions (slope steepness
55 and orientation, vegetation, size, slide velocity) and on weather conditions.
56 These techniques that can be borne on various platforms (space, aerial or
57 ground) and combined together, can achieve decimetric to centimetric res-
58 olution and accuracy in favorable conditions ([Delacourt et al., 2007](#)). In
59 the last few years, both airborne and ground LiDAR techniques have been
60 increasingly applied for mass movement studies, particularly in steep and
61 rugged terrain (e.g. [Thoma et al., 2005](#); [Abellan et al., 2006](#); [Corsini et al.,
62 2007](#); [Deparis et al., 2008](#); [Oppikofer et al., 2008](#)). Recently, airborne LiDAR
63 images were successfully used to map recent and historical landslides in gentle
64 slope areas ([Schulz, 2007](#); [Van den Eeckhaut et al., 2007](#)). Major advantages
65 of LiDAR technique are the flexibility and the quickness of the acquisition as
66 well as the relatively simple data processing, allowing multi-temporal Digi-
67 tal Elevation Models (DEMs) to be generated ([McKean and Roering, 2004](#);
68 [Rosser et al., 2005](#); [Thoma et al., 2005](#); [Oppikofer et al., 2008](#)).

69 In parallel, shallow geophysics has also considerably evolved with the
70 emergence of 2D and 3D spatial imaging, allowing the study of the spatial

71 and temporal variations inside landslides (for a recent review, see [Jongmans](#)
72 [and Garambois, 2007](#)). Geophysical imaging has the major advantages to
73 give continuous information on the studied body and to be non-invasive.
74 However, resolution generally decreases with depth. Geophysical prospect-
75 ing applied to landslides encompasses a large number of techniques: seismic
76 reflection, seismic refraction, electrical resistivity tomography (ERT), seismic
77 noise measurements, spontaneous potentials, electromagnetic, Ground Pen-
78 etrating Radar (GPR) and gravimetry. Flexible and quick to deploy, these
79 techniques have been applied on various types of landslides for slope varying
80 from a few degrees (earth slide) to vertical (rock fall), with a penetration
81 depth of the surveys ranging from 3 m to 400 m ([Green et al., 2007](#); [Heincke](#)
82 [et al., 2006](#); [Jongmans and Garambois, 2007](#)).

83 . Although remote sensing and geophysical techniques are complementary
84 for landslide imaging purposes, they have been rarely associated. [Roch et al.](#)
85 [\(2006\)](#) and [Deparis et al. \(2008\)](#) combined remote and ground imaging tech-
86 niques for determining the geometry and the 3D fracture pattern of poten-
87 tially unstable cliff sites. A dense digital surface model of the rock face
88 was measured from laser scanning (LiDAR) and/or photogrammetry, while
89 the GPR performed on the cliff allowed the discontinuity pattern inside rock
90 mass to be obtained. On low slope made of clay-rich sediments, [Perrone et al.](#)
91 [\(2006\)](#) presented a joint analysis of SAR interferometry and ERT surveys for
92 investigating and understanding complex ground deformation of different ori-
93 gin. [Brückl et al. \(2006\)](#) utilized photogrammetric, GPS and seismic data
94 to derive the kinematics of the Gradenbach deep-seated landslide (Austria)
95 affecting crystalline rocks.

96 . In the following, the geological setting of the study area is first assessed.
97 Then, the use of LiDAR data, aerial photos and GPS measurements will
98 lead to better constrain the two landslides activity and extent as well as
99 their kinematics. The easy-to-deploy and fast H/V prospecting will allow to
100 estimate the soft layer thickness. The original combination and integration
101 of these techniques will help to build a 3D view of the soft layer and assess
102 a possible geological control on the landslides.

103 **2. Geological structure and mapping**

104 . The Trièves area is located 40 km south of the city of Grenoble in the ex-
105 ternal French Alps (Fig. 1). This plateau region with a maximum altitude of
106 800 m above sea level (asl) corresponds to a large depression of about 300 km²
107 drained by the Drac river and its tributaries. It is bordered, to the West and
108 to the South by the Vercors and Dévoluy carbonate massifs, respectively.
109 To the East, it is limited by the southern end of the crystalline Belledonne
110 range (Fig. 1). This area is the main outcrop of quaternary glaciolacustrine
111 clays. Many landslides affect this zone among which 15 % are supposed to
112 be sliding (Lorier and Desvarreux, 2004). These slides might affect surfaces
113 as large as 1 km². They can develop over several slip surfaces ranging from
114 superficial (5 m to 15 m) to rather deep (down to 50 m; Blanchet, 1988; Jong-
115 mans et al., 2009). Slide velocities are generally low (a few cm/year) but can
116 reach several m/year in certain places. In some cases, generally after a long
117 wet period accompanied by quick snowmelt, the slides can evolve into a mud-
118 flow and velocities can reach several m/h. This lead to dramatic events in
119 Harmalière in 1981 (Moulin and Robert, 2004) and in La-Salle-en-Beaumont

120 in 1994 (Moulin and Chapeau, 2004). This important gravitational instabil-
121 ity is mainly related to the Quaternary geological history of the region. It
122 was controlled by several glacier fluctuations, which resulted in alternating
123 deposition and erosion phases. The particular geometrical setting and the
124 sediments recorded these major climatic fluctuations.

125 . The substratum on which rely the quaternary formations is made of early
126 Jurassic carbonate strata which were folded and faulted during the alpine oro-
127 genesis. Ancient glacial (Riss) and interglacial phases (Riss-Würm) carved
128 the substratum and generated valleys partly filled with Riss-Würm alluvial
129 deposits. This lead to an irregular shape of the basement prior to the last
130 glacial phase (Würm; -80 to -12ky BP). During the last glacial maximum
131 extension (LGM, Würm period, -22 to -18 ky BP; Clark et al., 2009), the
132 Isère glacier, coming from the North, blocked downstream the torrential flows
133 from the Drac river and its tributaries, generating an ice-dammed lake (lake
134 Trièves, Fig. 1; Monjuvent, 1973). This lake was progressively filled during
135 thousands of years mainly by millimetric to decimetric rythmic alternations
136 of clay and silt layers originating from nearby Mesozoic marls and crystalline
137 massifs (Huff, 1974). These laminated clays rest either on carbonate or al-
138 luvial, locally cemented, compact layers from the interglacial Riss-Würm
139 period. The irregular shape of the basement induces strong lateral thickness
140 variations of the palaeolake infill, from 0 to more than 250 m (see Figs. 2 and
141 3; Monjuvent, 1973; Antoine et al., 1981). The top of the clay is generally
142 found at an elevation of about 750 m asl (Antoine et al., 1981). Morainic
143 deposits which cap the clays are found as far as the south of the village
144 of Sinard, indicating the southward limit of the würmian glacier extension

145 (Figs. 1 and 2). Also, these moraines are not present in the downslope parts
146 of landslides, where clays outcrop (Fig. 2). At the study site, their thickness
147 evolves, around Sinard, from 50 m to the West to a few meters to the East.
148 At the end of the LGM in Europe, the Isère glacier withdrew, allowing the
149 rivers to cut deeply into the formations. This last erosion phase created the
150 actual Drac river valley. This favoured the conditions for landslide develop-
151 ment in the clay with a general Eastward motion in the study area (Fig. 2;
152 Brocard et al., 2003; Jongmans et al., 2009).

153 . The geological map and the cross-section of figures 2 and 3, respectively,
154 show the geometrical arrangement of the geological formations in presence.
155 Two palaeovalleys, labelled v1 and v2 on figure 3, depict previous interglacial
156 incision phases. Palaeovalley v1 is the oldest one (Monjuvent, 1973; Brocard
157 et al., 2003). Both palaeovalleys are filled with locally cemented alluvial lay-
158 ers. They have been recognized by field observations and, under the clay
159 cover, by geophysical investigations conducted in the 1950s for the construc-
160 tion of the Monteynard dam (location in Fig. 2; Crosnier-Leconte et al., 1953;
161 Monjuvent, 1973). In the study area (Fig. 1), these erosion and deposition
162 phases resulted in an irregularly shaped substratum top (Fig. 3). The last
163 erosion phase, at the end of the LGM, resulted in the actual Drac river valley.
164 A general view from the opposite side of the lake is presented on figure 4a.
165 Jurassic bedrock, quaternary alluvial layers and laminated clays are shown
166 on outcrop pictures of figures 4b, 4c and 4d, respectively.

167 3. Landslide geomorphology and history

168 3.1. Techniques

169 In this part we will focus on the comparison between the Avignonet and
170 Harmalière landslides. Three techniques have been used to analyze and com-
171 pare the geomorphology and kinematics differences of these slides. To obtain
172 a DEM for geomorphological interpretation a LiDAR laser scan, covering the
173 two landslides, was performed. Aerial photographs and GPS measurements
174 were used to analyse the kinematics back to 1948.

175 . The LiDAR scan was performed in November 2006 using the handheld air-
176 borne mapping system Helimap[®] (Vallet and Skaloud, 2004) mounted on a
177 helicopter flying about 300 m above the terrain. The landscape is character-
178 ized by forest, agriculture and grassland. There are no artificial structures
179 on Harmalière but a few small roads, 3 piles of an electric landline and 54
180 buildings on the Avignonet landslide. The time for the acquisition was cho-
181 sen to be in November, so that the leaf coverage has already decreased and
182 snowfall has not started yet. The system has recorded 36 million last pulses
183 of the surface reflection resulting in an average of 6 points per m² with an
184 accuracy of 10 cm in vertical and horizontal directions. In order to derive the
185 bare earth model excluding trees and houses the point cloud was filtered and
186 interpolated to a 2 m raster grid with the software SCOP++[®] IPF (2004)
187 using the *robust interpolation* method (Briese et al., 2002). The number of
188 points classified as ground reflections and therefore used for the gridding of
189 the DEM is 21 million, which is equivalent to an average of 3 points per m².
190 The resulting shaded DEM is shown in figure 5. Shaded representations with

191 different light angles were used for geomorphologic interpretation.

192 . GPS campaign measurements have been performed biannually (April and
193 November) since 1995 by RTM (*Restauration des Terrains en Montagne*, a
194 french public survey) at 25 points on the Avignonet landslide (Fig. 5), rel-
195 atively to several reference points located on nearby stable bed-rock. The
196 average standard deviation for all measured points is 6 mm. No GPS mea-
197 surements have been done so far inside the Harmalière landslide, because of
198 the quick surface evolution making benchmark installation difficult. Average
199 horizontal velocity vectors are shown in figure 5. No clear temporal pattern
200 of velocity changes emerges from the data certainly due to the low temporal
201 sampling.

202 . To analyze the kinematics further in the past, digital photogrammetric
203 scans of aerial photographs from IGN (*Institut Géographique National, France*)
204 of the years 1948, 1956, 1978, 1985, 1993, 2001 and 2003 at scales between
205 1:20 000 and 1:30 000 covering the two landslides have been used. The photos
206 were orthorectified with the software Geomatica[®] using the DEM BD-ALTI[®]
207 from IGN with a resolution of 50 m (the LiDAR DEM was not covering a
208 large enough area to be used), the calibration certificates of the utilized cam-
209 eras and 56 ground control points measured on the field with differential
210 GPS. The ortho-photos have been analyzed to investigate the landslide ac-
211 tivity through time. The mapping of the denuded area and of the headscarps
212 was found to be useful indicators of landslide activity.

213 In the studied area, bare soil (mostly clays) can be observed in sev-
214 eral places, over significantly large surface (from about 250 m² to more than
215 50000 m² in Harmalière). Where field observation has been done, it turns out

216 that these bare soil surfaces result from erosion and weathering processes in
217 relation to superficial landslide activities. In the following the term denuded
218 area is used to refer to these surfaces. To quantify the spatial and temporal
219 distribution of denuded area over the whole studied area, aerial photos have
220 been used.

221 When comparing field observations at Harmalière and Avignonet land-
222 slides with ortho-photos, denuded area appears as very bright pixels in aerial
223 photography compared to the surrounding forested or agricultural landscape
224 except for some field crops, the river and some human constructions (road,
225 buildings) that can also be bright (see Fig. 6, top). Consequently, for each
226 ortho-photo, pixels have been classified as denuded area, using as reference
227 the level of brightness of areas where bare soil has been observed in the
228 field, and excluding the river (or the lac), bright fields, roads and buildings
229 identifiable by their characteristic geometric shapes (see Fig. 6, bottom).

230 *3.2. Analysis*

231 Using the three sets of data, as described in the previous section, several
232 features can be analyzed to highlight the difference in kinematics between the
233 Avignonet and Harmalière landslides: the general orientation of the landslide,
234 surface velocity, denuded area, headscarp evolution and surface roughness.

235 . The clay area affected by sliding is larger on Avignonet (1.8 km^2) than on
236 Harmalière (1.2 km^2). The headscarps of Harmalière and Avignonet mapped
237 from the 1956 aerial photo (before the construction of the Monteynard Dam),
238 show that the maximum distance between the headscarps and the Drac Valley
239 is about 1.5 km in both cases. Both headscarps follow a curve that have a

240 symmetry axis oriented NNW-SSE, perpendicular to the Drac Valley, as it is
241 expected in a homogeneous clay mass deeply cutted by a valley (see Fig. 6 on
242 the bottom left). In Avignonet, the global sliding direction follows this axis
243 as expressed by the fan-shape pattern of the GPS velocity vectors and of the
244 slope directions of the main scarps. The 11-years average velocity vectors
245 from GPS on the Avignonet landslide are oriented from N 70 ° E to N 130 ° E
246 (Fig. 5), with a global N 100 ° E movement.

247 However, in Harmalière, according to the evolution of denuded area, the
248 most active areas of the landslide develop along a NNW-SSE axis (Fig. 6)
249 making a 30 ° angle with the Avignonet global sliding direction. The percent-
250 age of denuded surface is much larger in Harmalière than in Avignonet since
251 the first photos in 1948 (13 % versus 5 % in 1948, 20 % versus 3 % in 1984
252 and 19 % versus 3 % in 2003). Only a very small active area in the south
253 of Avignonet can be identified. The higher activity of Harmalière certainly
254 goes back, at least, to the end of the 19th century: at that time, according to
255 [Moulin and Robert \(2004\)](#), because of intense gully erosion occurring on the
256 NW part of the Harmalière slide, a significant reforestation work has been im-
257 plemented by the state authorities. Since the catastrophic Harmalière event
258 in 1981, denuded area are more concentrated in the eastern part of the Har-
259 malière landslide and has grown up-slope since then. Morphological changes
260 due to the 1981 event can be seen in the 1985 sketch of figure 6 forming
261 a NNW-SSE elongated body of denuded areas and also in figure 5 where,
262 at the same place, a roughness contrast with the surroundin terrain can be
263 noticed. The body itself can be divided into a steeper upstream part where
264 erosion dominates and a smoother lower part corresponding to an accumu-

265 lation zone, which has been conquered by trees again since the 1981 event
266 (Fig. 6 on the right). It is noteworthy that the recent up-slope evolution
267 is not directed straight to the NW, but has rotated clockwise to the N-NE
268 toward the Avignonet landslide.

269 The main headscarp of Harmalière, which has developed since the catas-
270 trophic event of March 1981, has mainly regressed through several brutal
271 events (biggest events were in 1988, 1996 and 2001; Moulin and Robert,
272 2004). This evolution is traceable from aerial photos analysis, providing a
273 mean regression rate of about 10 m/year between 1981 and 2003. These val-
274 ues are in agreement with ground observations (Moulin and Robert, 2004).
275 Since 2001, the headscarp regression has started to affect the southern limit
276 of the Avignonet landslide. A GPS measurement on the crest between Har-
277 malière and Avignonet indicates mass displacement towards the Southwest,
278 which is consistent with a faster regression of the Harmalière scarp toward
279 the Northeast (Fig. 6) relative to Avignonet. To the North of Harmalière
280 the limits with another landslide show a convexity toward the south (Fig. 5
281 and 6). On the LiDAR DEM, in the northern part of Avignonet slide, the
282 four uppermost major scarps (spaced by about one hundred of meter) seems
283 to be cut by the northern limit. This suggests that the northern landslide
284 has been the last to be the most active with respect to the Avignonet one.

285 . Surface roughness is also an indicator for landslide activity (Glenn et al.,
286 2006). Analysis of surface roughness from the shaded LiDAR DEM (Fig. 5)
287 shows differences between the Harmalière and Avignonet landslides in terms
288 of wavelengths and amplitudes along the slope direction. In Harmalière, a
289 characteristic small-scale roughness with wavelengths of 5-15 m and ampli-

290 tudes of 0.5-5 m can be observed, as well as roughness at a larger scale with
291 wavelengths of 80-200 m and amplitudes of 5-20 m. Two representative el-
292 evation profiles along the slope direction are shown in Fig. 5b. They are
293 divided into an upper and a lower part according to the presence of signifi-
294 cant changes of slope angle and roughness. On these profiles, the roughness
295 has been estimated using the root-mean-square deviations as described by
296 [Shepard et al. \(2001\)](#) with step-sizes of 10 m and 100 m along with the mean
297 slope angles (see table 1) corresponding small-scale roughness and large-scale
298 roughness at the wavelengths of 10 m and 100 m respectively. In Avignonet
299 the roughness is higher at the toe (1.6 m) than in the upper part (1.1 m)
300 for the small-scale roughness, which can be explained by the higher activity
301 down-slope, as suggested by the GPS, associated to a higher slope angle and
302 may be a higher erosion rate. In Harmalière, the roughness is lower in the
303 lower part than in the upper part for both wavelengths. Indeed, since 1981,
304 Harmalière has shown intense activity in its upper part including collaps-
305 ing at the headscarp and several minor scarps below, that are responsible in
306 a higher large-scale roughness (5.4 m). In the lower part, it evolves into a
307 more fluent mudslide (large-scale roughness at 2 m) related to the develop-
308 ment of an accumulation zone which presents a lower slope angle. Similar
309 large-scale roughness values are found for the upper parts of Avignonet and
310 Harmalière. Assuming that features related to large-scale roughness are more
311 robust through time than the small-scale one, this observation suggests that
312 Avignonet and Harmalière have experienced similar landslide processes in
313 the past. It must be stated that using roughness as an indicator of land-
314 slide activity can be biased by several factors that should be discussed. First

315 of all the density of data points after filtering when building the DEM can
316 influences the roughness. Data density decreased in forested area because
317 of the filtering and could lead into lower roughness. In figure 5, the con-
318 structed DEM is resampled at 2 m resolution: beside a few exceptions and
319 after filtering the forest, the laser points were still dense enough to avoid
320 smoothing effects due to interpolation. For instance, the south-western part
321 of the Harmalière landslide, which is largely forested, show a higher small-
322 scale roughness than the farming area of the Avignonet slide. On the other
323 hand the effect of farming altering the surface roughness should also be con-
324 sidered. On a very small scale farming can decrease the surface roughness
325 in a very short time, but also high scale roughness could be decreased by
326 farming over decades. On the other hand if the activity of the landslide and
327 with that the roughness is too high, the field will become abandoned and
328 the roughness will be not further decreased. Therefore it can be said that
329 farming amplifies the trend: smooth areas will be more smoothed and rough
330 areas will stay rough above a certain threshold. If there are other causes to
331 abandon a field then the landslide activity, the interpretation of roughness
332 can be misleading. In the case of Harmalière, it can be observed from the
333 aerial photos that a lot of fields become abandoned after being affected by
334 the past 1981 events. Today farming is completely stopped on Harmalière.
335 Also the active part in the south of Avignonet was partly farmland in the
336 past and become abandoned recently.

337 Taking into account these limitations, the roughness comparison between
338 the Avignonet and Harmalière slide still suggests a significant difference of the
339 recent sliding activity: the Avignonet slide do not show major recent active

340 movements in agreement with GPS data and aerial photographs analysis, in
341 contrast to the Harmalière slide. Only some areas in the lower part and in
342 the South of Avignonet exhibit small-scale roughness similar to Harmalière.
343 This is consistent with the location of denuded area in the most recent aerial
344 photos, which are correlated to a higher landslide activity.

345 . Regarding slide velocity, values measured by GPS at the surface of the
346 Avignonet landslide (Fig. 5) increase from less than 20 mm/year at the top
347 to more than 130 mm/year in the most active parts at the toe. The aerial
348 photos show no major signs of activity in Avignonet for the last 60 years with
349 no significant evolution of the headscarp and very small denuded areas. In
350 Harmalière no GPS measurements are available but, by tracking morpholog-
351 ical features through different dates in the aerial photos, one can get a rough
352 estimate of the average velocity of several meters per year for some parts of
353 the main landslide body, which is significantly higher than in Avignonet.

354 . Comparison of the two landslides by LiDAR scan, GPS data and aerial
355 photos has shown that the recent and former landslide kinematics are highly
356 different for the two earth slides. Harmalière seems to be much more active
357 than Avignonet, today and in the past 60 years. A major factor for this could
358 be the underlying bedrock topography. This possible influence parameter is
359 investigated in the following section.

360 **4. Geophysical investigation**

361 . The objective of the geophysical investigation is to map the thickness of
362 the soft layer over the two landslides. This thickness ranges from 0 to more

363 than 250 m from the East to the West, respectively (Antoine et al., 1981).
364 The site to be characterized encompasses a surface of 5 km². Geological
365 investigations conducted in the 1950s for the study of the Monteynard dam
366 (location in Fig. 2) were limited to the Avignonet area. No information is
367 available about the geological setting of the Harmalière area. Furthermore,
368 geophysical campaigns which aimed to characterize the geological setting
369 in the immediate vicinity of the dam were not published and only synthetic
370 data are available. This poses the problem of the reliability of the established
371 model of figure 3.

372 Recent works (Jongmans et al., 2009) have shown the existence of a strong
373 shear wave (S-wave) velocity contrast (more than 3 on average) between the
374 soft upper layers (clays and moraines; 250 m/s < Vs < 600 m/s) and the
375 substratum made of compact cemented alluvial layers (Vs = 1250 m/s) and
376 Jurassic limestone (Vs = 2000 m/s). With such characteristics, microtremor
377 processing (H/V technique) has been proven to be one of the most robust
378 and easy exploration tool for mapping the thickness of alluvial or lacustrine
379 sediments (Ibs-von Seht and Wohlenberg, 1999; Delgado et al., 2000; Guéguen
380 et al., 2007; Méric et al., 2007; Le Roux et al., 2008) including the Trièves
381 area (Jongmans et al., 2009).

382 *4.1. Method*

383 The H/V technique is a single station method consisting in calculating the
384 horizontal to vertical spectral ratios (H/V) of seismic noise records. For a sin-
385 gle homogeneous soft horizontal layer (1D geometry) overlying the bedrock,
386 the H/V curve exhibits a peak at a frequency f_{HV} that is the shear-wave res-
387 onance frequency f_0 of the soft layer (Bard, 1998). This theoretical resonance

388 frequency is given by (Haskell, 1960):

$$f_0 = \frac{V_s}{4T} \quad (1)$$

389 where V_s is the soft layer S-wave velocity (in m/s) and T is the layer
390 thickness (in m). The resonance frequency decreases with a decrease of V_s
391 and an increase of thickness. If V_s is known, this equation allows the layer
392 thickness to be determined. For a layered medium overlying a halfspace, the
393 resonance frequency can be computed from the thickness and the dynamic
394 characteristics of each soil layer (Haskell, 1960) and the total thickness of
395 the soil layers can be deduced if the vertical velocity profile is known. When
396 the seismic impedance contrast Z between the soft layer and the bedrock is
397 high enough ($Z > 4$), the H/V peak was shown to result from a change in the
398 ellipticity of the fundamental mode of the Rayleigh waves (Bonnetfoy-Claudet
399 et al., 2006). The corresponding ellipticity frequency f_{ell} is then equal to the
400 resonance frequency f_0 .

401 Measurements were made with a single three-component 5 s sensor giving
402 a flat response between 0.2 and 50 Hz and connected to a light seismic ac-
403 quisition system (Chatelain et al., 2000). Seismic noise was recorded during
404 15 minutes with a 200 Hz sampling frequency. Data were processed with the
405 Sesarray package (www.geopsy.org; Wathelet et al., 2004). Microtremor
406 records were cut into 30 s time windows, for which Fourier spectra were com-
407 puted and smoothed using the technique proposed by Konno and Ohmachi
408 (1998). H/V spectral ratios were computed for all time windows and the
409 mean H/V curve is given with standard deviations at each site.

410 The H/V method offers the advantages of being easy to deploy (one

411 station-one people) and quick (about 10 sites per day), depending on field
412 conditions. Limits are weather conditions, likely to influence the H/V curve
413 (Koller et al., 2004), and the assumption of horizontally layered medium,
414 which cannot be valid in case of strong lateral seismic contrasts (Uebayashi,
415 2003). In the study site conditions, the 5 km² area was covered within a two
416 weeks delay by a single operator.

417 4.2. H/V data

418 H/V measurements were performed at 104 sites (Fig. 7) and were located
419 with a GPS with a horizontal accuracy ranging from one to a few meters
420 under forest. Elevation values were extracted from the LiDAR DEM. H/V
421 curves (spectral ratio versus frequency) at two sites (S1 and S2) are presented
422 in figure 7a. Both curves exhibit a peak with amplitude over 8 at 0.58 Hz
423 for S1 and at a 3.63 Hz for S2, corresponding to the resonance frequency of
424 the site. At S2, a second peak appears at 20 Hz, which could correspond
425 to the resonance of a superficial layer or to a higher resonance mode. The
426 great majority of the 104 measured H/V curves fit the criteria proposed in
427 the SESAME guideline (Koller et al., 2004) for a 1D resonance phenomenon,
428 with well-individualized peaks and H/V amplitudes greater than 2. For some
429 measurements, located in the southeastern part of the Harmalière landslide
430 (location in Fig. 7b), however, curves exhibit a plateau-like shape, suggesting
431 2D or 3D effects. This point will be discussed below.

432 . The frequency field was gridded with a kriging algorithm (Kitanidis, 1997),
433 using an exponential variogram model with a N-S anisotropy for the search
434 radius. The gridded surface fits the experimental data with an absolute error

435 of 3%. The results are presented on figure 7c.

436 . To the West, the frequency map (Fig. 7c) reveals a NNE-SSW 500 m wide
437 elongated low-frequency zone, with values ranging from 0.4 to 1 Hz. To the
438 East, frequencies increase rapidly with distance, from 1 to 4 Hz. This east-
439 ward evolution of the resonance frequency is consistent with the thinning of
440 the clay layer and the corresponding rise of stiff layers shown in previous
441 works (Fig. 3; Crosnier-Leconte et al., 1953; Lambert and Montjuvent, 1968;
442 Blanchet, 1988; Jongmans et al., 2009). Applying equation 1 with a mean
443 S-wave velocity of 600 m/s in the clay layer (Méneroud et al., 1995) yields a
444 clay thickness between 375 m to the west and 37 m to the east. These thick-
445 ness values are however approximate, owing to the vertical S-wave velocity
446 variation in the clay layer. This results from the effects of compactness and
447 landslide activity. The dynamic characteristics (P-wave and S-wave veloci-
448 ties, density) within the different layers (from top to bottom: morainic col-
449 luvium, moraine, disturbed clays, undisturbed clays, alluvium and bedrock)
450 were obtained from previous seismic studies (Méneroud et al., 1995; Renalier
451 et al., 2007; Jongmans et al., 2009) and are presented in table 2. Thick-
452 ness values in the layers were estimated at each station from the available
453 geophysical and geotechnical data (Monjuvent, 1973; Antoine et al., 1981;
454 Blanchet, 1988; Lorier and Desvarreux, 2004; Jongmans et al., 2009). Then,
455 the only free parameter left is the thickness of the undisturbed clay layer.

456 Preliminary theoretical sensitivity tests were conducted using the Sesar-
457 ray package (Wathelet et al., 2004). They have shown that the clay thickness
458 is the main parameter controlling the resonance frequency. Furthermore,
459 these tests also showed that the impedance contrast, using the parameters

460 of Table 2, between clays and cemented alluvial layers was sufficient enough
461 to generate a peak corresponding to the resonance frequency. It implies that
462 the measured frequency corresponds in each case to the base of the clays.
463 No change in the f_0 value has been found when passing the landslide head-
464 scarp (cf. Fig. 7c). Furthermore, within the slide, no high-frequency peak
465 that could sign a shear plane has been recorded nor computed. These results
466 support the use of resonance frequency measurements for determining the
467 base of the clays. On an other hand, they indicate that this technique might
468 not be suitable for detecting and mapping the slide itself (on the contrary of
469 [Méric et al., 2007](#)). This could be due to an insufficient seismic impedance
470 contrast between the disturbed clays and the undisturbed laminated clays
471 (cf. Table 2).

472 The soft layer thickness was computed at each station by fitting the the-
473 oretical resonance frequency to the measured one using a trial and error
474 method. The plot of computed thicknesses versus experimental resonance
475 periods (inverse of the frequency) for the whole data set shows a good cor-
476 relation (Fig. 8a), corroborating the determined velocities in the layers. The
477 H/V deduced geometry of the seismic bedrock top is compared to the sec-
478 tion previously established from a long refraction profile conducted for the
479 study of the Monteynard dam (profile P1, location in Fig. 7c; [Blanchet, 1988](#))
480 which depicts the top of the Jurassic bedrock. A general good agreement is
481 observed between the two sections (Fig. 8b) in terms of relative way of the
482 palaeotopography. Observed discrepancies range from a very few m to the
483 East to some 40 m to the West. These disparities may come from strong
484 changes in layer velocities which are not known. This also may come from

485 the reference geological model, established from a refraction study and which
486 reliability is not known. This prevents this H/V mapping from being exhaus-
487 tive but allows the relative palaeotopography to be estimated and analyzed
488 in terms of geometry. Errors may also arise from 2D and/or 3D effects as well
489 as from varying cementation of the alluvial layers, as observed at outcrop.
490 These points will be discussed further.

491 . The thickness values obtained over the whole area were then kriged with
492 an exponential variogram model to produce a map depicting the distribution
493 of the thickness of the soft layer (moraines, morainic colluvium and clays)
494 over the seismic substratum (alluvial compact layers and Jurassic bedrock;
495 Fig. 9a). Outcrops of stiff alluvium and carbonate bedrock (0 m of soft cover
496 thickness) were added in order to better constrain the interpolation. The
497 thickness map is in agreement with the frequency map of Figure 7, showing a
498 significant westward increase of the clay thickness from 0 m on the valley flank
499 to more than 300 m below the village of Sinard. These values are consistent
500 with field observations and the outcropping of the alluvial layers to the East
501 (Fig. 3). This is also consistent with previous estimations (Monjuvent, 1973;
502 Antoine et al., 1981) that give a soft layer thickness from 0 to 300 m on the
503 West (moraines and clays; see Table 2).

504 . Thickness values were subtracted from elevation values given by the LiDAR
505 DEM at each measurement point. These points were again spatially interpo-
506 lated using a kriging method with an exponential variogram model. The relief
507 of the bottom of the clay layer is given in Figure 9b. The palaeotopography
508 upon which clays have been deposited is very irregular, with elevation varia-
509 tions of more than 150 m. The major feature is the presence of a depression

510 striking NNE-SSW, which is bordered to the East by a N-S ridge culminating
511 at an elevation of about 600 to 620 m (cross-section AB on Fig. 10). This
512 depression probably coincides with palaeovalley v2 of the palaeodrac river, as
513 suggested by the presence of alluvial outcrops along the lake (Figs. 2 and 3)
514 and by previous works (Crosnier-Leconte et al., 1953; Monjuvent, 1973). The
515 N-S ridge appearing to the east and shown on cross-section AB corresponds
516 to the presence of carbonate bedrock irregularly covered with compact allu-
517 vial layers and sporadically outcropping along the lakeshore. To the South,
518 this ridge disappears (cross-sections AC and AD; Fig. 10). In figure 9 are
519 superimposed the limits of the landslides of Avignonet and Harmalière. Be-
520 low the southern part of the Avignonet slide, on its eastern side, the ridge of
521 compact layers continuously extends perpendicularly to the global slide mo-
522 tion and could act as a buttress. On the contrary, the Harmalière landslide
523 clearly developed over the lower elevation zone (Fig. 9b) and its motion can
524 be explained by its orientation which changes from N-S to NW-SE in the
525 South.

526 5. Discussion

527 . Geophysical investigation with ambient noise measurements allows to com-
528 pute a frequency map (Fig. 7c) that can be turned into a soft sediments
529 thickness map which is used to derive a paleotopography map (Fig. 9). Con-
530 sidering the thicknesses that have been computed, one may argue that 2D
531 and/or 3D effects are present and that measured f_{HV} might be biased. The
532 presence of non-1D cases is highlighted by strong lateral variations observed
533 on the frequency map (Fig. 7a). Consequently, calibration based on the el-

534 lipticity of the fundamental mode of Rayleigh waves may be irrelevant and
535 cannot provide the correct thickness for the soft layers. Some of these effects
536 have been noticed on the signals located in the southeastern part of the Har-
537 malière landslide (Fig. 7b) by plateau-like signals. By picking the frequency
538 at the plateau cut-off, the resonance frequency tends to be overestimated
539 (Guillier et al., 2006). Similar side effects have already been reported by
540 studies conducted within small-apex valleys (Uebayashi, 2003; Koller et al.,
541 2004; Guéguen et al., 2007; Le Roux et al., 2008). This is illustrated on
542 cross-section AC of figure 9. Between abscissa 800 and 850 m, a bulge of
543 substratum is visible. It corresponds to a zone with pleatau-like H/V curves.
544 This strong side effect tends to provide an overestimated measured f_{HV} and,
545 thus, a lowered clay thickness. As a consequence, the observed bulge on
546 cross-section AC is likely to result from side effects and not to represent a
547 real structure.

548 Another problem arising from calibrations is the eventual presence of
549 lateral Vs variations in the soft sediments. Vertical and W-E variations are
550 known to be present within clays (Renalier et al., 2007; Jongmans et al., 2009)
551 and were taken into account for the frequency to thickness computations. On
552 an other hand, since the Würm glacier extension was limited to the village of
553 Sinard (Figs. 1 and 2), the presence of such an ice-cap and of the 50 m thick
554 moraines may influence the compactness and, then, the S-wave velocity of the
555 clays across this boundary. Nevertheless, N-S velocity variations within the
556 soft layer have not been reported hitherto and were not eventually integrated
557 to this work.

558 As a consequence of these uncertainties, the soft layer thickness map com-

559 puted from H/V data may contain more or less important error bars. These
560 errors are estimated to a few meters to the East (were soft layer thickness is
561 of a few tens of meters) to some 40 m to the West, were the soft layer reaches
562 300 m. These uncertainties do not question the reliability of the first order
563 of observations since the aim of this work is to map the relative variations of
564 the seismic substratum. The basement of the former Trièves lake on which
565 were deposited the glaciolacustrine laminated clays is irregularly shaped and
566 shows strong lateral variations. At most, this introduces an uncertainty in
567 the exhaustive mapping of the clay thickness.

568 . There is a N-S ridge to the East of the investigated area, culminating at
569 about 620 m, which disappears to the south. Cross-sections AC and AD on
570 figure 10 illustrate the progressive vanishing of this hard ridge marking a
571 depression to the South at about 460 m, which continues toward the North-
572 East and then runs along the west side of the ridge. This depression allowed
573 the deposition of 300 m of clays and moraines over a seismic substratum lo-
574 cated at an elevation less than 520 m asl. Over the ridge only a very few
575 tens of meters of clay remain because of a lower thickness of sediment and
576 of erosion. This analysis confirms previous works done by other geophysical
577 methods on Avignonet (Crosnier-Leconte et al., 1953; Blanchet, 1988), and
578 also extends consistently the knowledge of the palaeotopography to the south
579 up to Harmalière. The new map also confirms the geological observations and
580 interpretation inferring that several glacial and interglacial erosion phases,
581 that took place during the Quaternary (Monjuvent, 1973; Brocard et al.,
582 2003), created a palaeotopography marked by the incision of at least two
583 paleovalleys of the Drac, called "Drac de Sinard" and "Drac de Cros", filled

584 with alluvium and running below the laminated clay of the Sinard plateau,
585 notably below Harmalière and Avignonet (Lambert and Montjuvent, 1968;
586 Monjuvent, 1973). However, Lambert and Montjuvent (1968) and Monju-
587 vent (1973) indicate that the top of the "Drac de Sinard" alluvium is at
588 about 500-520 m asl in the Harmalière-Avignonet area, and that its bottom
589 is at about 410 m asl, whereas our results show, west to the ridge, interme-
590 diate values ranging between 440 and 500 m asl. This could suggest that the
591 method used is not sensible to the contrast between alluvial deposits and
592 clay. Preliminary theoretical sensitivity tests revealed that the impedance
593 contrast between the clays and the compact, cemented, alluvial layers was
594 large enough to generate a resonance frequency peak. On an other hand,
595 field observations, along with previous works (e.g. Monjuvent, 1973) indicate
596 that these alluvial layers are only locally cemented. Where they are not,
597 their mechanical parameters, especially Vs, would dramatically decrease and
598 approach values within clays (for Vs values within alluvial sediments, see
599 for example, Pugin et al., 2009). In such cases, the measured resonance fre-
600 quency and, consequently, the base of the soft layer, would correspond to
601 the interface between the base of non-cemented alluvial layers and the top of
602 cemented alluvial layers and/or the Jurassic bedrock. Here again, this aspect
603 is pointless since the aim of this work is to map the seismic substratum. Fur-
604 thermore, it could partly explain the important discrepancies between the
605 computed thicknesses and the reference cross-section (Fig. 8).

606 The other palaeovalley, called "Drac de Cros", recognized by geophysics
607 and geology (Crosnier-Leconte et al., 1953; Lambert and Montjuvent, 1968;
608 Monjuvent, 1973) is supposed to run east to the ridge, below the North-

609 Eastern part of the Avignonet landslide. The top of its alluvial deposits is
610 expected to be at about 600 m asl. The location of this palaeovalley is not
611 clear on our map. Geological observations suggest that this palaeovalley is
612 narrower with steeper flanks than the "Drac de Sinard" one, which could ex-
613 plain the difficulty to see it. Here again, our estimations of the clay basement
614 give a substratum at 530-560 m asl at the location of the expected "Drac de
615 Cros", that is lower than geological estimate (600 m). This bias could again
616 be explained by the fact that, the alluvial deposits are not always cemented
617 (Monjuvent, 1973), leading to a localization of the basement that is inter-
618 mediate between the top of the alluvial deposits and the top of the Jurassic
619 bedrock. One can also notice a small East-West depression across the ridge
620 in the middle of the Avignonet landslide that has not been reported before.
621 However, because of the possible uncertainties of the map mentioned here-
622 before, we do not try to interpret it as a paleomorphological feature. In the
623 following, we will retain only the robust features of our geophysical investi-
624 gations, that are the N-S ridge and the depression related to the palaeovalley
625 of the "Drac de Sinard".

626 . Beside the geophysical maps of soft layer thickness and of palaeotopogra-
627 phy, a comparison of the Harmalière and Avignonet landslides kinematics
628 has been done based on GPS, aerial photo and LiDAR DEM analysis. It
629 turns out that two main differences between the two landslides are observed
630 for the recent time: the orientation of the main sliding direction, and the
631 level of sliding activity. Regarding the orientation it has been shown in sec-
632 tion 3 that the Harmalière landslide is developing since 1948 along a NW-SE
633 direction, which is significantly oblique with the expected NNW-SSE direc-

634 tion toward the Drac valley (before the lake filling) that can be observed, for
635 instance, in Avignonet. The NW-SE axis also corresponds to the orientation
636 of the "Drac de Sinard" palaeovalley below the Harmalière slide, suggesting
637 that the orientation of the Harmalière landslide might be controlled by the
638 palaeotopography. Regarding the level of sliding activity, a first observation
639 is that the northern part of the Harmalière headscarp is moving to the North
640 starting to erode the southern flank of the Avignonet slide. Secondly, GPS
641 measurements and analysis of digital photographs, back to 1948, reveal that
642 Harmalières main slide body is far more active than the Avignonet one: there
643 are regression of the headscarp of several m per year being accompanied by
644 rapid evolution of the denuded area (that are not observed on Avignonet
645 since 60 years) and sliding velocity than can exceed 1 m/year, whereas, in
646 Avignonet, the 11 years GPS data do not show velocity exceeding 15cm/year.
647 These observations are consistent with the morphology of the landslides, es-
648 pecially the roughness, shown by the Lidar DEM which reflects more long-
649 term activity. In Avignonet there is a gradient of displacement from the
650 uppermost sliding clays, showing displacements of about 1-2cm/year, up to
651 the toe of the clay mass with 10-13cm/year where the clay thickness are less
652 than a few tens of meters (around 30m). Downward, outcrops of alluvial
653 layer or of Jurassic bedrock do not show sliding processes. On a W-E profile
654 along the Avignonet slide, [Jongmans et al. \(2009\)](#) have observed a negative
655 correlation between the sliding velocity and the Vs values of the first 5 m sug-
656 gesting a downward increasing deformation state of this clay material. They
657 also found evidence for a slip surface at about 40 m depth within the clay
658 confirming previous hypothesis done from inclinometric data ([Blanchet, 1988](#);

659 [Lorier and Desvarreux, 2004](#)). When comparing these observations with the
660 position of the ridge in the palaeotopography, one can be inclined to think
661 that, here also, there is a palaeotopographic control over the kinematics: the
662 ridge of hard layers along the eastern part of the Avignonet landslide acts as
663 a buttress that could mechanically prevent the slide from rapid evolution as
664 observed in Harmalière (Fig. 5).

665 . However, the hypothesis of a palaeotopography control over the landslide
666 kinematics has to be discussed taking into account a longer time interval
667 than the last 60 years of observation on which it is based. This perspective
668 arises several questions: is the present-day kinematics, in progress since at
669 least 1948, representative of the long term evolution of the two landslides
670 (i.e. over several thousands of years)? If the palaeotopography controls the
671 kinematics, what is its long term impact on the landslide morphology? One
672 may argue that each landslide has its own evolution that consists in successive
673 phases of slow movement (like in Avignonet today), followed by rapid events
674 (like in Harmalière today). In this case, the present-day situation could have
675 been the reverse at some periods in the past: an active Avignonet slide and a
676 quiet Harmalière slide. Then, the supposed relationships between kinematics
677 and palaeotopography would be just a coincidence or negligible.

678 . To address this issue, morphological parameters can be considered that may
679 reflect the relative long term level of activity of the two landslides. The area
680 of the Avignonet landslide is larger than in Harmalière. However, as noticed
681 in section 2, the maximum distance to the Drac river is similar. This suggests
682 a similar average speed of regression of the headscarps, if we do the reasonable

683 hypothesis that erosion processes start at the same time for the two nearby
684 landslides, when the Drac started to incise into the glaciolacustrine deposits.

685 Note that, in this discussion, we will not try to use absolute time to
686 estimate incision or erosion rates, because the timing of the beginning of the
687 Drac incision is still debated. Some authors favor a last Würm maximum
688 at 50-40 Ka (Monjuvent, 1973; Nicoud et al., 2002) and others at 30-20 Ka
689 coincident with the LGM (Brocard, 2003; Brocard et al., 2003). The volume
690 of eroded clays can be also considered. As described in section 2, it can be
691 assumed that, at the end of the last Würm maximum, the glacio-lacustrine
692 clay material formed a flat plateau culminating at about 750 m asl on top
693 of which were deposited an additional 50 m of moraines as far as the south
694 of the village of Sinard (Figs. 1 and 2; Monjuvent, 1973; Antoine et al.,
695 1981). The eroded volume of material can be calculated by subtracting the
696 Lidar DEM from this 800 m asl surface over the area below which clay are
697 present for each landslide. This gives about 436 million of m³ for Avignonet
698 and 340 million of m³ for Harmalière (for comparison, during the 1981 main
699 Harmalière mudflow, 250000 m³ went into the lake (Blanchet, 1988) but more
700 materials were mobilized that accumulated at the bottom of the slide above
701 the lake). If we divide theses values by the landslide area, it gives an average
702 of 157 m of eroded clay material per m² for Avignonet and 195 m per m² for
703 Harmalière, suggesting a higher average long-term activity for Harmalière.
704 However, one cannot say if, for instance, these values correspond to very
705 large events closely related in a short period of time (with the rest of the
706 time a lower activity of the slide) or, in contrast, if they are representative
707 of continuous evolution of the slides. The DEM morphology analysis has

708 shown that there is a contrast in roughness between the two landslides that
709 suggests (with the mapping of the evolution of the denuded area) that the
710 Harmalière slide is more active than the Avignonet one. However it can be
711 observed within the Harmalière slide that a short term event like the 1981
712 mudflow and its subsequent events can change the small-scale roughness and
713 vegetation cover of the slopes: the 1981 active body shows a visible roughness
714 contrast with the surrounding areas of the slide. Regarding the Harmalière
715 headscarp regression over the Avignonet slide, it seems to reflect only a recent
716 evolution because it does not significantly modify the regular headscarp curve
717 of the Avignonet slide. Furthermore, the observed speed of the Harmalière's
718 headscarp regression since 1981, at about 10m/year, is not sustainable over
719 the long term. In brief, all these morphological criteria do not seem very
720 reliable for reflecting significant differences in kinematics of the long term
721 (thousands of years) evolution of the slides.

722 . Physical mechanisms responsible of the laminated clays instability, has
723 also to be taken into account in the analysis. It has been shown by several
724 authors that, for the Trièves laminated clays, natural slope become instable
725 when exceeding 6 to 8° ([Antoine et al., 1981, 1991](#); [Giraud et al., 1991](#)).
726 According to [Antoine et al. \(1981\)](#), three main types of sliding are frequently
727 observed: **1**) sliding of the vegetative cover (thickness less than 0.5 m) over
728 the clays; **2**) creeping of a whole top clay layer (thickness 0-6 m) associated
729 with localized and superficial mudflow; **3**) sliding of mass of clay over a
730 slip surface (thickness up to 50 m), than can evolve into a mudflow like in
731 Harmalière in 1981 with velocities that can reach several m/h.

732 . According to these processes, the palaeotopography should have an effect
733 onto landslide activity when the clays thickness is lower than 50 m except
734 if the material of the substratum has a significant impact on the hydrogeo-
735 logical conditions, for instance by draining water and influencing the water
736 table. In Avignonet, analysis of piezometric data done so far, do not provide
737 evidence of such an effect. Considering **1)** the present day topographic profile
738 along the main slope direction of the Avignonet slide, **2)** the position of the
739 ridge at about 700m west to the Drac valley with a highest point at about
740 620m and **3)** the initial level surface, at around 800m asl, in which the Drac
741 started to incise, one can say that the possible effects of the palaeotopog-
742 raphy on the landslide activity did not start at the beginning of the slide's
743 life and may be relatively recent in the slide histories. That could explain
744 why there is no clear evidence of long term difference of activities between
745 the two landslides. When the deep slip surfaces started to reach the clay
746 basement, the development of new deeper slip surface has been influenced
747 because the basement prevents them to go deeper, as it has been expected
748 if no ridge were present. This certainly causes a slowing of their evolution.
749 In Avignonet, [Blanchet \(1988\)](#) and [Jongmans et al. \(2009\)](#) proposed several
750 hypotheses about the geometry of a deep slip surface at about 50m identified
751 from inclinometric data. These geometries are influenced by the basement
752 in their lowest part, suggesting that the Avignonet landslide activity may
753 have been recently slow down by the palaeotopography. The fact that the
754 GPS velocities are the highest in the lowest part near the ridge is interpreted
755 as being due to more intensive superficial creep at the toe of the slide. In
756 the Harmalière slide, no inclinometric data are available. However, applying

757 similar reasoning, one can say that the situation where the deep slip surfaces
758 reach the clay basement has not started on the NW-SE profiles along which
759 the slide is developing (Fig. 9). If, instead, we consider an E-W profile in
760 the middle of the Harmalière slide, the situation looks more similar to the
761 Avignonet case. This could explain why the Harmalière slide is developing
762 along an NW-SE direction and why it has recently shown a highest activity
763 than in Avignonet.

764 . Consequently, we propose that palaeotopography is a significant control-
765 ling factor of the Harmalière and Avignonet landslide evolutions, acting as a
766 mechanical buttress against the development of slip surfaces in the clays. Pa-
767 leotopography may also have another impact (not investigated in this study),
768 that is an influence on the drainage condition of the two landslides because
769 paleotopography is responsible of large permeability heterogeneities due to
770 the presence of alluvial deposits in the palaeovalleys.

771 **6. Conclusion**

772 . The two adjacent landslides of Avignonet and Harmalière have been studied
773 following an original multidisciplinary approach based on geodetic (GPS, dig-
774 ital photographs), remote sensing (LiDAR), and ground geophysics (ambient
775 noise measurements). The aim was to understand the differential kinematics,
776 motion directions and morphology that characterize these landslides located
777 in the same geotechnical setting.

778 GPS measurements and digital photographs reveal that the difference in
779 kinematics between the two earthslides can be tracked back to 60 years ago
780 at least. The Avignonet slide is mainly directed towards the East (N 100 ° E)

781 while Harmalière is mainly oriented towards SE. The LiDAR scan map il-
782 lustrates this differential motion and morphology between the two slides and
783 highlights that the Harmalière slide is still presently much more active than
784 the Avignonet one.

785 A ground geophysical prospecting based on ambient noise measurements
786 allowed to record the resonance frequencies at different locations. These mea-
787 sured resonance frequencies were turned into soft sediments thicknesses. Fi-
788 nally, a map depicting the base of the clays was computed. It indicates that
789 the basement is very irregularly shaped with strong lateral E-W variations
790 over 150 m. This map confirms previous field and geophysical observations
791 that revealed a westward thickening of the clays. This map also reveals the
792 presence of a N-S ridge of hard sediments (Jurassic bedrock and/or compact
793 alluvial layers) on the eastern side of the Avignonet landslide. This ridge
794 disappears when approaching the Harmalière landslide and makes place to
795 what can be interpreted like a NW-SE oriented palaeovalley of the river Drac.
796 It is proposed that the ridge could act as a butress which could mechanically
797 prevent the Avignonet landslide from evolving as fast as the Harmalière one.
798 Furthermore the NW-SE palaeovalley located under the Harmalière landslide
799 corresponds to the motion direction of the slide. It is then finally proposed
800 that the slides different behaviours are partly controlled by the palaeotopo-
801 graphic setting of lake Trièves.

802 This approach reveals to be a quick and relatively low-cost way to charac-
803 terize the geomorphological setting of this sedimentary basin over an impor-
804 tant area of 5 km² by building a geometrical framework for landslide charac-
805 terization within fine-grained soft-sediments.

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811 **7. References**

812 **References**

- 813 Abellan, A., Vilaplana, J. M., Martinez, J., 2006. Application of a long-
814 range terrestrial laser scanner to a detailed rockfall study at Vall de Nuria
815 (Eastern Pyrenees, Spain). *Engineering Geology* 88, 136–148.
- 816 Antoine, P., Giraud, A., Montjuvent, G., 1981. Les argiles litées du Trièves
817 (Isère) ; conditions de gisement et exemples de propriétés géotechniques.
818 *Bulletin de la Société Géologique de France* (7), t. XXIII (2), 117–127.
- 819 Antoine, P., Monnet, J., Rai, N. E., Moulin, C., Mériaux, P., 1991. Résultats
820 de cinq années d’auscultation sur un glissement dans les argiles glacio-
821 lacustres du Trièves (Sud-Est de la France). In: 6th International Sympo-
822 sium on Landslides. Balkema, Rotterdam.
- 823 Bard, P.-Y., 1998. Microtremor measurements: a tool for site effect esti-
824 mation? In: Irikura, K., Kudo, K., Okada, H., Sasatani, T. (Eds.), *The*
825 *Effects of Surface Geology on Seismic Motion*. Balkema, Rotterdam, pp.
826 1251–1279.
- 827 Blanchet, F., 1988. Etude géomécanique de glissements de terrain dans les
828 argiles glaciolacustres de la vallée du Drac. Ph.D. thesis, Université Joseph
829 Fourier, Grenoble, France, Grenoble, France.
- 830 Bonci, L., Bozzano, F., Calcaterra, S., Eulilli, V., Ferri, F., Gambino, P.,
831 Manuel, M. R., Martino, S., Scarascia Mugnozza, G., 2004. Geological
832 control on large seismically induced landslides: the case of Cerda (Southern

- 833 Italy). In: Proceedings of the 9th International Symposium on Landslides.
834 pp. 985–991.
- 835 Bonnefoy-Claudet, S., Cornou, C., Bard, P.-Y., Cotton, F., Moczo, P., Kris-
836 tek, J., Fäh, D., 2006. H/V ratio: a tool for site effects evaluation. Results
837 from 1-D noise simulations. *Geophysical Journal International* 167, 827–
838 837.
- 839 Bozzano, F., Lenti, L., Martino, S., Paciello, A., Scarascia Mugnozza, G.,
840 2008. Self-excitation process due to local seismic amplification responsible
841 for the reactivation of the Salcito landslide (Italy) on 31 October 2002.
842 *Journal of Geophysical Research* 113, B10312.
- 843 Briese, C., Pfeifer, N., Dorninger, P., 2002. Applications of the robust inter-
844 polation for DTM determination. In: IAPIS XXXIV. Vol. 3A. pp. 55–61.
- 845 Brocard, G. Y., Van Der Beek, P. A., Bourlès, D. L., Siame, L. L., Mugnier,
846 J.-L., 2003. Long-term fluvial incision rates and postglacial river relaxation
847 time in the French Western Alps from ^{10}Be dating of alluvial terraces
848 with assessment of inheritance, soil development and wind ablation effects.
849 *Earth and Planetary Science Letters* 209, 197–214.
- 850 Brocard, G. Y., 2003. Origine, variabilité spatio-temporelle et signature mor-
851 phologique de l’incision fluviale dans les Alpes dauphinoises (SE France).
852 *Géologie Alpine, Mémoire H.S.* 43.
- 853 Brückl, E., Brunner, F. K., Kraus, K., 2006. Kinematics of a deep-seated
854 landslide derived from photogrammetric, GPS and geophysical data. *En-
855 gineering Geology* 88 (3-4), 149–159.

- 856 Chatelain, J.-L., Guéguen, P., Guillier, B., Fréchet, Bondoux, F., Sarrault,
857 J., Sulpice, P., Neuville, J. M., 2000. Cityshark: A user-friendly instrument
858 dedicated to ambient noise (microtremor) recording for site and building
859 response studies. *Seismological Research Letters* 71, 698–703.
- 860 Clark, P. U., Dyke, A. S., Shakun, J. D., Carlson, A. E., Clark, J., Wohlfarth,
861 B., Mitrovica, J. X., Hostetler, S. W., McCabe, A. M., 2009. The Last
862 Glacial Maximum. *Science* 325 (5941), 710–714.
- 863 Corsini, A., Borgatti, L., Coren, F., Vellico, M., 2007. Use of multitemporal
864 airborne LiDAR surveys to analyse postfailure behaviour of earthslides.
865 *Canadian Journal of Remote Sensing* 33 (2), 116–120.
- 866 Crosnier-Leconte, J., Bordet, C., Duffaut, P., 1953. Séparation de deux an-
867 ciens lits successifs dans la vallée du Drac à Monteynard (Isère). *Comptes-*
868 *Rendus Sommaires de la Société Géologique de France* 12, 221–223.
- 869 Debemas, J., 1967. La Chapelle-en-Vercors. In: *Carte géologique de la*
870 *France au 1/50000*. BRGM Éditions, Orlans, France.
- 871 Delacourt, C., Allemand, P., Bertier, E., Raucoules, D., Casson, B., Grand-
872 jean, P., Pambrun, C., Varel, E., 2007. Remote-sensing techniques for
873 analysing landslide kinematics: a review. *Bulletin de la Société Géologique*
874 *de France* 178 (2), 89–100.
- 875 Delgado, J., López Casado, C., Estévez, A., Giner, J., Cuenca, A., Molina,
876 S., Aug. 2000. Mapping soft soils in the Segura river Valley (SE Spain);
877 a case study of microtremors as an exploration tool. *Journal of Applied*
878 *Geophysics* 45 (1), 19–32.

- 879 Deparis, J., Fricourt, B., Jongmans, D., Villemin, T., Effendiantz, L., Mathy,
880 A., 2008. Combined use of geophysical methods and remote techniques for
881 characterizing the fracture network of a potential unstable cliff site (the
882 'Roche du Midi', Vercors massif, France). *Journal of Geophysical Engi-*
883 *neering* 5, 147–157.
- 884 Eilertsen, R. S., Hansen, L., Bargel, T. H., Solberg, I.-L., 2008. Clay slides
885 in the Milselv valley, northern Norway: Characteristics, occurrence, and
886 triggering mechanisms. *Geomorphology* 93 (3-4), 548–562.
- 887 Fruneau, B., Achache, J., Delacourt, C., 1996. Observation and modelling of
888 the Saint-Étienne-de-Tinée landslide using SAR interferometry. *Tectono-*
889 *physics* 265, 181–190.
- 890 Giraud, A., Antoine, P., Van Asch, T. W. J., Nieuwenhuis, J. D., 1991.
891 Geotechnical problems caused by glaciolacustrine clays in the French Alps.
892 *Engineering Geology* 31, 185–195.
- 893 Glenn, N. F., Streutker, D. R., Chadwick, D. J., Thackray, G. D., Dorsch,
894 S. J., Jan. 2006. Analysis of LiDAR-derived topographic information for
895 characterizing and differentiating landslide morphology and activity. *Geo-*
896 *morphology* 73 (1-2), 131–148.
- 897 Green, A. G., Maurer, H., Spillmann, T., Heincke, B., Willenberg, H., 2007.
898 High-resolution geophysical techniques for improving hazard assessments
899 of unstable rock slopes. *The Leading Edge* 25 (3), 311–316.
- 900 Guéguen, P., Cornou, C., Garambois, S., Banton, J., Jan. 2007. On the Lim-
901 itation of the H/V Spectral Ratio Using Seismic Noise as an Exploration

902 Tool: Application to the Grenoble Valley (France), a Small Apex Ratio
903 Basin. *Pure and Applied Geophysics* 164 (1), 115–134.

904 Guillier, B., Cornou, C., Kristek, J., Bonnefoy-Claudet, S., Bard, P., Fah,
905 D., , Moczo, P., 2006. Simulation of seismic ambient vibrations: does the
906 H/V provide quantitative information in 2D-3D structure ? In: ESG2006.
907 Grenoble.

908 Haskell, N. A., 1960. Crustal reflexion of plane Sh waves. *Journal of Geo-*
909 *physical Research* 65, 4147–4150.

910 Heincke, B., Maurer, H., Green, A. G., Willenberg, H., Spillmann, T.,
911 Burlini, L., 2006. Characterizing an unstable mountain slope using shallow
912 2D and 3D seismic tomography. *Geophysics* 71 (6), B241–B256.

913 Huff, W. D., 1974. Mineralogy and provenance of Pleistocene lake clay in an
914 alpine region. *Geological Society of America Bulletin* 85, 1455–1460.

915 Ibs-von Seht, M., Wohlenberg, J., 1999. Microtremor measurements used to
916 map thickness of soft sediments. *Bulletin of the Seismological Society of*
917 *America* 89 (1), 250–259.

918 IPF, 2004. Software SCOP++, <http://www.ipf.tuwien.ac.at/products/produktinfo/scop/scop.dtm>.

919 Jongmans, D., Bièvre, G., Schwartz, S., Renalier, F., Bearez, N., 2009. Geo-
920 physical investigation of the large Avignonet landslide in glaciolacustrine
921 clays in the Trièves area (french Alps). *Engineering Geology* 109, 45–56.

922 Jongmans, D., Garambois, S., 2007. Geophysical investigation of landslides
923 : a review. *Bulletin de la Société Géologique de France* 178 (2), 101–112.

- 924 Kimura, H., Yamaguchi, Y., 2000. Detection of landslide areas using satellite
925 radar interferometry. *Photogrammetric engineering and remote sensing* 66,
926 337–344.
- 927 Kitanidis, P. K., 1997. *Introduction to Geostatistics: Applications in Hydro-*
928 *geology*. Cambridge University Press, Cambridge.
- 929 Koller, M. G., Chatelain, J.-L., Guillier, B., Duval, A.-M., Atakan, K., La-
930 cave, C., Bard, P.-Y., 2004. Practical user guidelines and software for the
931 implementation of the H/V ratio technique: measuring conditions, pro-
932 cessing method and results interpretation. In: *Proceedings of the 13th*
933 *world conference in earthquake engineering*, Vancouver, Canada.
- 934 Konno, K., Ohmachi, T., 1998. Ground-motion characteristics estimated
935 from spectral ratio between horizontal and vertical components of mi-
936 crotremor. *Bulletin of the Seismological Society of America* 88 (1), 228–241.
- 937 Lambert, A., Montjuvent, G., 1968. Quelques vues nouvelles sur l’histoire
938 quaternaire de la vallée du Drac (note préliminaire). *Géologie Alpine* 44,
939 117–138.
- 940 Lapenna, V., Lorenzo, P., Perrone, A., Piscitelli, S., Rizzo, E., Sdao, F.,
941 May 2005. 2D electrical resistivity imaging of some complex landslides in
942 Lucanian Apennine chain, southern Italy. *Geophysics* 70 (3), B11–B18.
- 943 Le Roux, O., Schwartz, S., Gamond, J.-F., Jongmans, D., Tricart, P., Se-
944 brier, M., 2008. Interaction between tectonic and erosion processes on the
945 morphogenesis of an Alpine valley: geological and geophysical investiga-

- 946 tions in the lower Romanche valley (Belledonne massif, western Alps).
947 International Journal of Earth Sciences, 1437–3254.
- 948 Lorier, L., Desvarreux, P., 2004. Glissement du Mas d’Avignonet, commune
949 d’Avignonet. In: Proceedings of the workshop Ryskhydrogeo, Program
950 Interreg III, La Mure (France). p. 8 p.
- 951 McKean, J., Roering, J., 2004. Objective landslide detection and surface
952 morphology mapping using high-resolution airborne laser altimetry. Geo-
953 morphology 57 (3-4), 331–351.
- 954 Méneroud, J.-P., Duval, A.-M., Vidal, S., Fréchet, J., Gamond, J.-F., Beck,
955 C., Tardy, M., Bard, P.-Y., Barnichon, E., Gaboriaud, J.-M., 1995.
956 Franchissement de l’Ebron, Etude de l’aléa sismique local. Tech. Rep.
957 93/95666/74, CETE Méditerranée.
- 958 Méric, O., Garambois, S., Malet, J.-P., Cadet, H., Gueguen, P., Jongmans,
959 D., 2007. Seismic noise-based methods for soft-rock landslide characteriza-
960 tion. Bulletin de la Societe Geologique de France 178 (2), 137–148.
- 961 Metternicht, G., Hurni, L., Gogu, R., 2005. Remote sensing of landslides:
962 An analysis of the potential contribution to geospatial systems for hazard
963 assessment in mountainous environments. Remote Sensing of Environment
964 98, 284–303.
- 965 Monjuvent, G., 1973. La transfluence Durance-Isère. Essai de synthèse du
966 Quaternaire du bassin du Drac (Alpes françaises). Géologie Alpine 49, 57–
967 118.

- 968 Moulin, C., Chapeau, C., 2004. Le glissement de la Salle en Beaumont (Isère).
969 In: Proceedings of the workshop Ryskhydrogeo, Program Interreg III, La
970 Mure (France). p. 9 p.
- 971 Moulin, C., Robert, Y., 2004. Le glissement de l'Harmalière sur la commune
972 de Sinard. In: Proceedings of the workshop Ryskhydrogeo, Program Inter-
973 reg III, La Mure (France). p. 11 p.
- 974 Nicoud, G., Royer, G., Corbin, J.-C., Lemeille, F., Paillet, A., 2002. Creuse-
975 ment et remplissage de la vallée de l'Isère au Quaternaire récent. Apports
976 nouveaux du forage GMB1 (1999) dans la région de Grenoble (France).
977 Géologie de la France 4, 39–49.
- 978 Oppikofer, T., Jaboyedoff, M., Keusen, H.-R., 2008. Collapse at the eastern
979 Eiger flank in the Swiss Alps. *Nature Geoscience* 1, 531–535.
- 980 Perrone, A., Zeni, G., Piscitelli, S., Pepe, A., Loperte, A., Lapenna, V.,
981 Lanari, R., 2006. Joint analysis of SAR interferometry and electrical re-
982 sistivity tomography surveys for investigating ground deformation: the
983 case-study of Satriano di Lucania (Potenza, Italy). *Engineering Geology*
984 88 (3-4), 260–273.
- 985 Pugin, A. J.-M., Pullan, S. E., Hunter, J. A., Oldenborger, G. A., 2009.
986 Hydrogeological prospecting using P- and S-wave landstreamer seismic re-
987 flection methods. *Near Surface Geophysics* 7, 315–327.
- 988 Renalier, F., Jongmans, D., Bièvre, G., Schwartz, S., Orengo, Y., 2007.
989 Characterisation of a landslide in clay deposits using Vs measurements.

- 990 In: Near Surface 2007, EAGE meeting, 03-04 September 2007, Istanbul,
991 Turkey.
- 992 Roch, K. H., Chwatal, E., Brückl, E., 2006. Potential of monitoring rock
993 fall hazards by gpr: considering as example of the results of Salzburg.
994 Landslides 3, 87–94.
- 995 Rosser, N. J., Petley, D. N., Lim, M., Dunning, S. A., Allison, R. J., 2005.
996 Terrestrial laser scanning for monitoring the process of hard rock coastal
997 cliff erosion. Quarterly Journal of Engineering Geology 38, 363–375.
- 998 Rott, H., Scheuchl, B., Siegel, A., Grasemann, B., 1999. Monitoring very
999 slow slope motion by means of SAR interferometry: A case study from
1000 a mass waste above a reservoir in the Ötztal Alps, Austria. Geophysical
1001 Research Letters 26, 1629–1632.
- 1002 Schulz, W. H., 2007. Landslide susceptibility revealed by LiDAR imagery
1003 and historical records, Seattle, Washington. Engineering Geology 89 (1-2),
1004 67–87.
- 1005 Shepard, M. K., Campbell, B. A., Bulmer, M. H., Farr, T. G., Gaddis, L. R.,
1006 Plaut, J. J., 2001. The roughness of natural terrain: A planetary and
1007 remote sensing perspective. Journal of Geophysical Research 106 (E12),
1008 32777–32795.
- 1009 Squarzoni, C., Delacourt, C., Allemand, P., 2003. Nine years of spatial and
1010 temporal evolution of the La Valette landslide observed by SAR interfer-
1011 ometry. Engineering Geology 68, 53–66.

- 1012 Strozzi, T., Farina, P., Corsini, A., Ambrosi, C., Thüning, M., Zilger, J., A.,
1013 W., Wegmüller, U., Werner, C., 2005. Survey and monitoring of landslide
1014 displacements by means of L-band satellite SAR interferometry. *Landslides*
1015 2, 193–201.
- 1016 Thoma, D. P., Guptab, S. C., Bauerc, M. E., Kirchoff, C. E., 2005. Air-
1017 borne laser scanning for riverbank erosion assessment. *Remote Sensing of*
1018 *Environment* 95, 493–501.
- 1019 Uebayashi, H., 2003. Extrapolation of Irregular Subsurface Structures Us-
1020 ing the Horizontal-to-Vertical Spectral Ratio of Long-Period Microtremors.
1021 *Bulletin of the Seismological Society of America* 93 (2), 570–582.
- 1022 Vallet, J., Skaloud, J., 2004. Development and experiences with a fully-digital
1023 handheld mapping system operated from helicopter. *The International*
1024 *Archives of the Photogrammetry, Remote Sensing and Spatial Informa-*
1025 *tion Sciences XXXV (Commission 5), Part B.*
- 1026 Van den Eeckhaut, M., Poesen, J., Verstraeten, G., Vanacker, V., Moeyer-
1027 sons, J., Nyssen, J., Van Beek, L. P. H., Vandekerckhove, L., 2007. Use
1028 of LiDAR-derived images for mapping old landslides under forest. *Earth*
1029 *Surface Processes and Landforms* 32, 754–769.
- 1030 Wathelet, M., Jongmans, D., Ohrnberger, M., 2004. Surface-wave inversion
1031 using a direct search algorithm and its application to ambient vibration
1032 measurements. *Near Surface Geophysics* 2, 211–221.

Table 1: Root-mean-square deviations from the elevation profiles seen in Figure 5b for 2 different wavelengths (step sizes) representing the roughness at different scales. See text for details. * This value is not reliable due to the short profile length.

	A1-A2 upper Avignonet	A2-A3 lower Avignonet	H1-H2 upper Harmalière	H2-H3 lower Harmalière
Profile length (m)	577	247	874	624
Mean slope angle (°)	9.8	16.3	11.6	6.9
RMS deviation (m) small scale (10 m)	1.1	1.6	1.6	1
RMS deviation (m) large scale (100 m)	4.4	4.4*	5.4	2

Table 2: Dynamic characteristics (compressional wave velocity V_p , shear wave velocity V_s and density) used for the calculation of the theoretical resonance frequency. See text for details.

Geological unit	Thickness (m)	V_p (m/s)	V_s (m/s)	Density
Morainic colluvium	5	500	250	1.9
Moraines	0-50	1850	150-450	2
Disturbed clays	0-45	1850	150-450	2
Undisturbed clays	0-250	1850	600-650	2
Compact alluvium	0-100	2350	1250	2
Carbonate bedrock	Halfspace	3000	2000	2.6

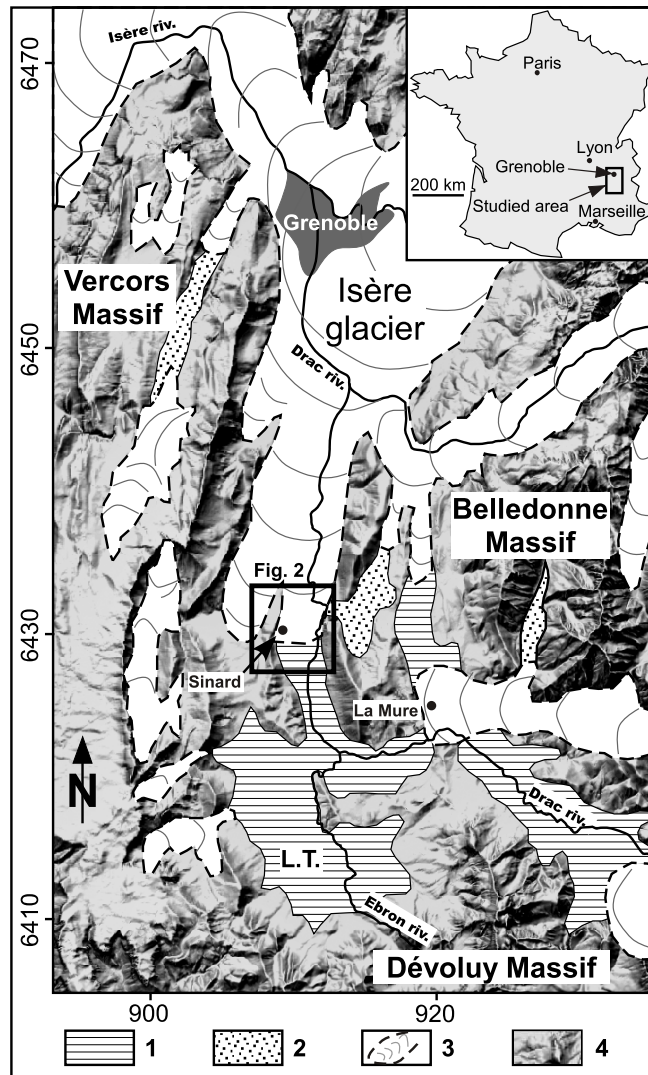


Figure 1: Location of the area and palaeogeographical map at the end of the Würm age (adapted from Monjuvent, 1973). Coordinates are kilometric and expressed in the French system Lambert-93. The black thick box shows the location of the study area and the extent of Fig. 2. 1: Laminated clay deposits in the Trièves area; 2: glaciofluvial deposits; 3: extension of the Isère Glacier at the end of the Würm age; 4: present-day topography; L.T.: lake Trièves.

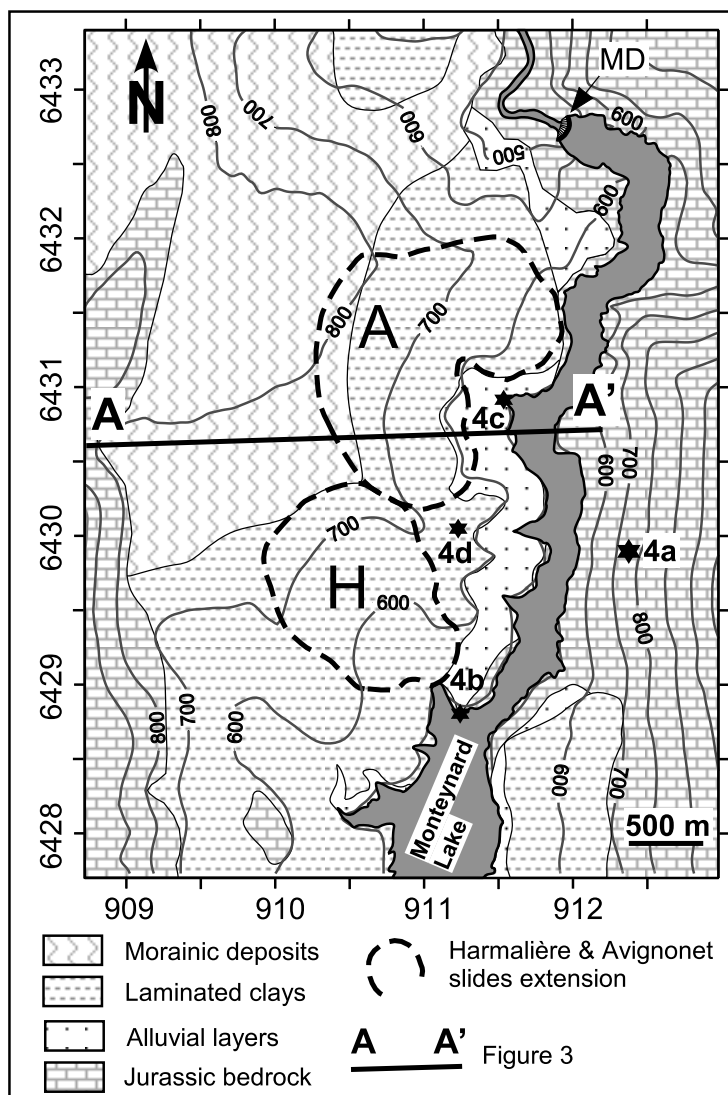


Figure 2: Geological map (location in Fig. 1) modified from [Debelmas \(1967\)](#) and [Lambert and Montjuvent \(1968\)](#). Screens and morainic colluvium have not been reported. Black stars refer to pictures in figure 4. Position of cross-section AA' of figure 3 is indicated. A: Avignonet landslide. H: Harmalière landslide. MD : Monteynard dam.

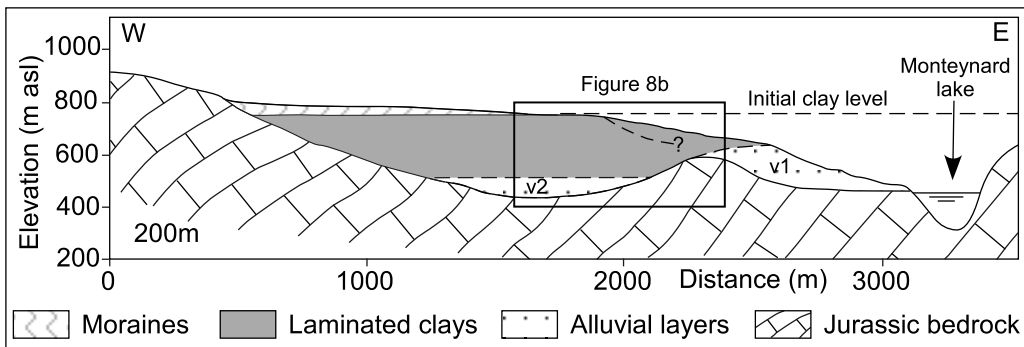


Figure 3: Geological cross section (location in Fig. 3). Screens and morainic colluvium have not been reported. v1: first palaeovalley ("Drac de Cros"). v2: second palaeovalley ("Drac de Sinard"). Initial level of laminated clays has been reported (750 m asl). Dashed line within clays depicts the lower limit of the Avignonet landslide, deduced from inclinometer data. See text for details.

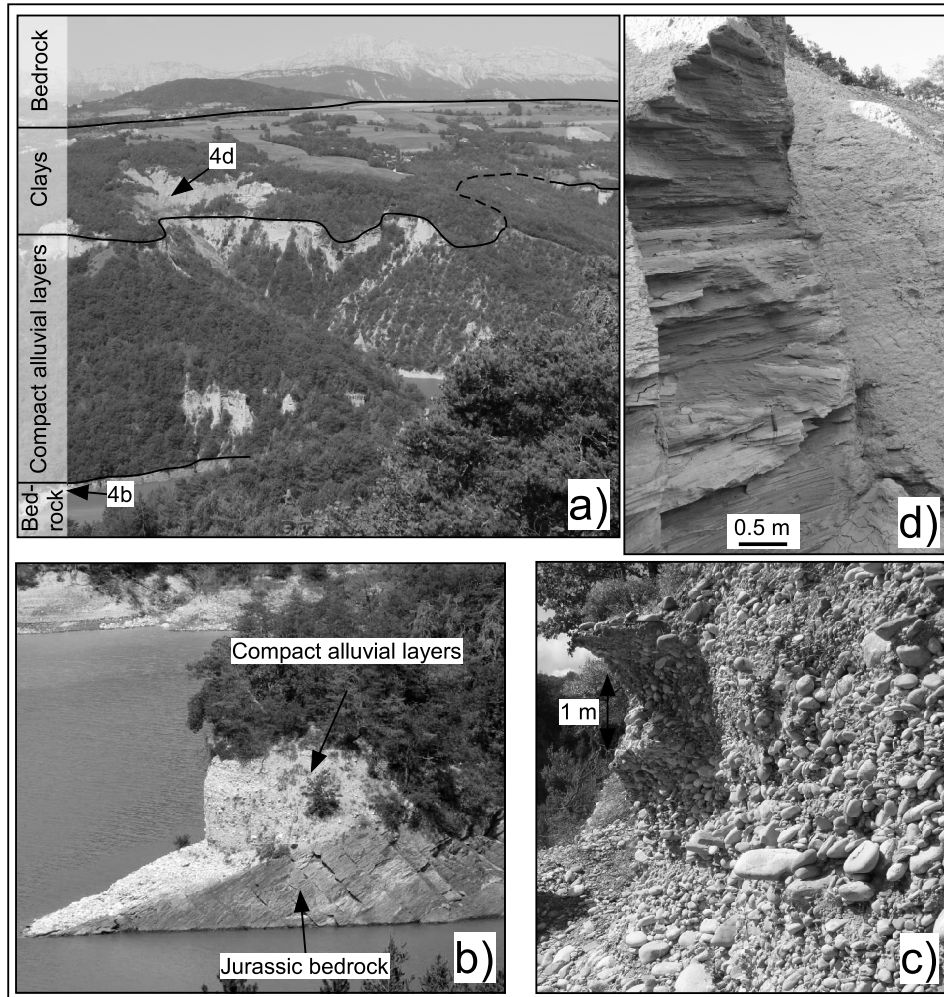


Figure 4: Outcrop pictures. **a)** General view of the Avignonet landslide from the opposite bank of lake Monteynard (picture taken from location 4a in Fig. 2); **b)** Jurassic bedrock overlain by compact, locally cemented alluvial layers constituted by heterometric and heterogeneous gravels. Present-day Drac river is at the forefront; **c)** Compact alluvial layers; **d)** laminated clays. Location of the pictures is detailed on Fig. 2

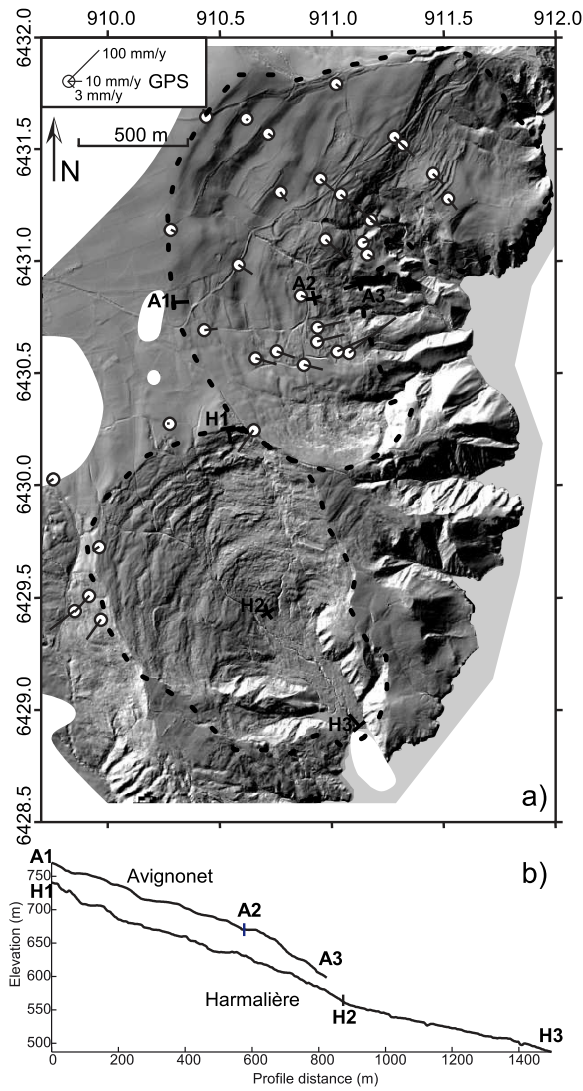


Figure 5: LidAR DEM and elevation profiles of the study area. **a)** Shaded LiDAR DEM (light direction is from NW, horizontal resolution of 2 m) covering the Avignonet (North) and Harmalière (South) landslides. Dotted lines indicate the landslide limits. White areas indicate data holes in the LiDAR coverage. The white circles show the position of the GPS points, and the thin straight lines represent the 11-years average horizontal velocity measured by GPS. A1-3 and H1-3 indicate the location of the elevation profiles. **b)** Elevation profiles through Avignonet and Harmalière with subdivisions in an upper and lower part. Profiles are 2 times vertically exaggerated.

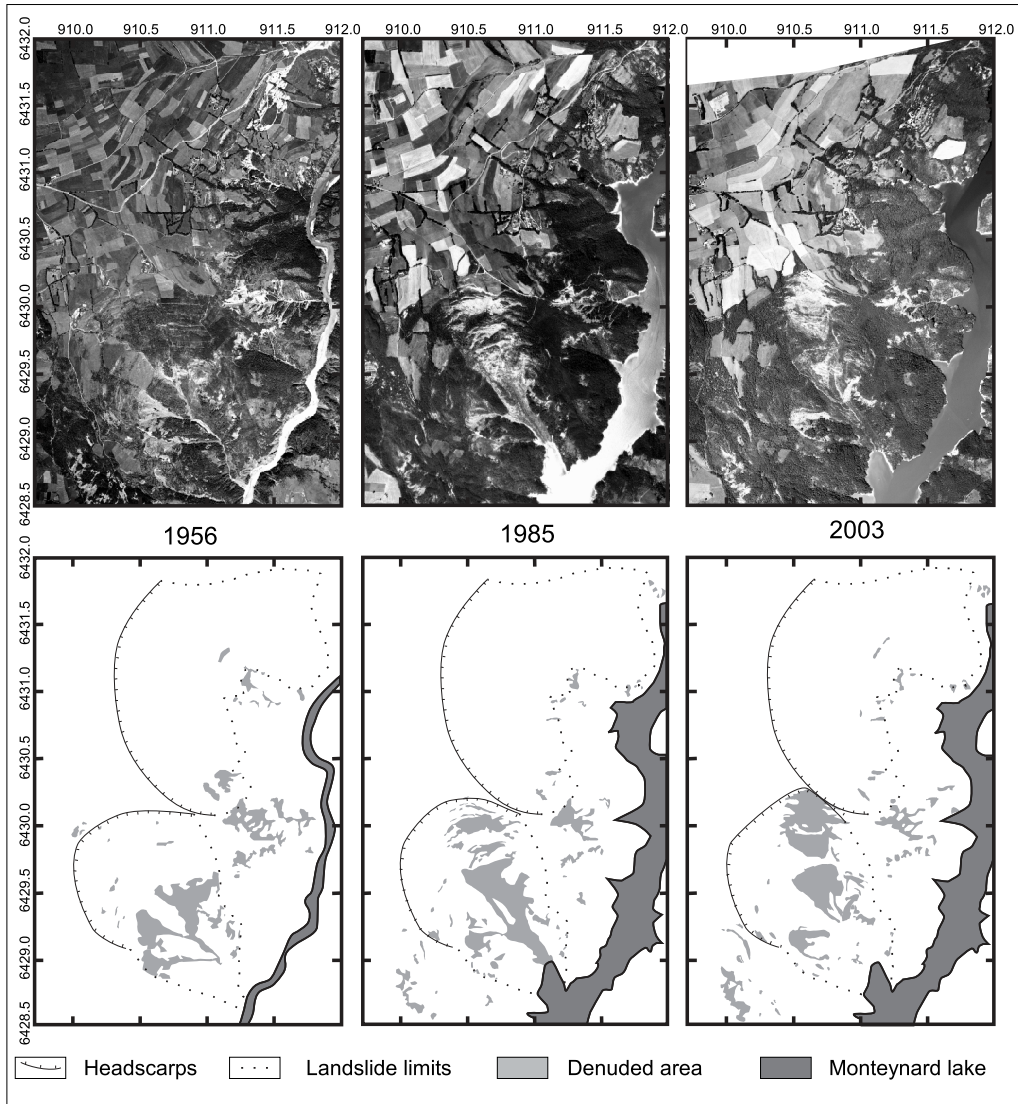


Figure 6: Top: Orthorectified aerial photos of the years 1956, 1985 and 2003. The scales of the original photos are 1:25 000, 1:30 000 and 1:25 000. Bottom: Interpretation of the corresponding aerial photo showing the denuded areas related to landslide activity. White areas are not covered by the images. In 1962 the Drac River was dammed up creating the lake Monteynard shown in the photos from 1985 and 2003.

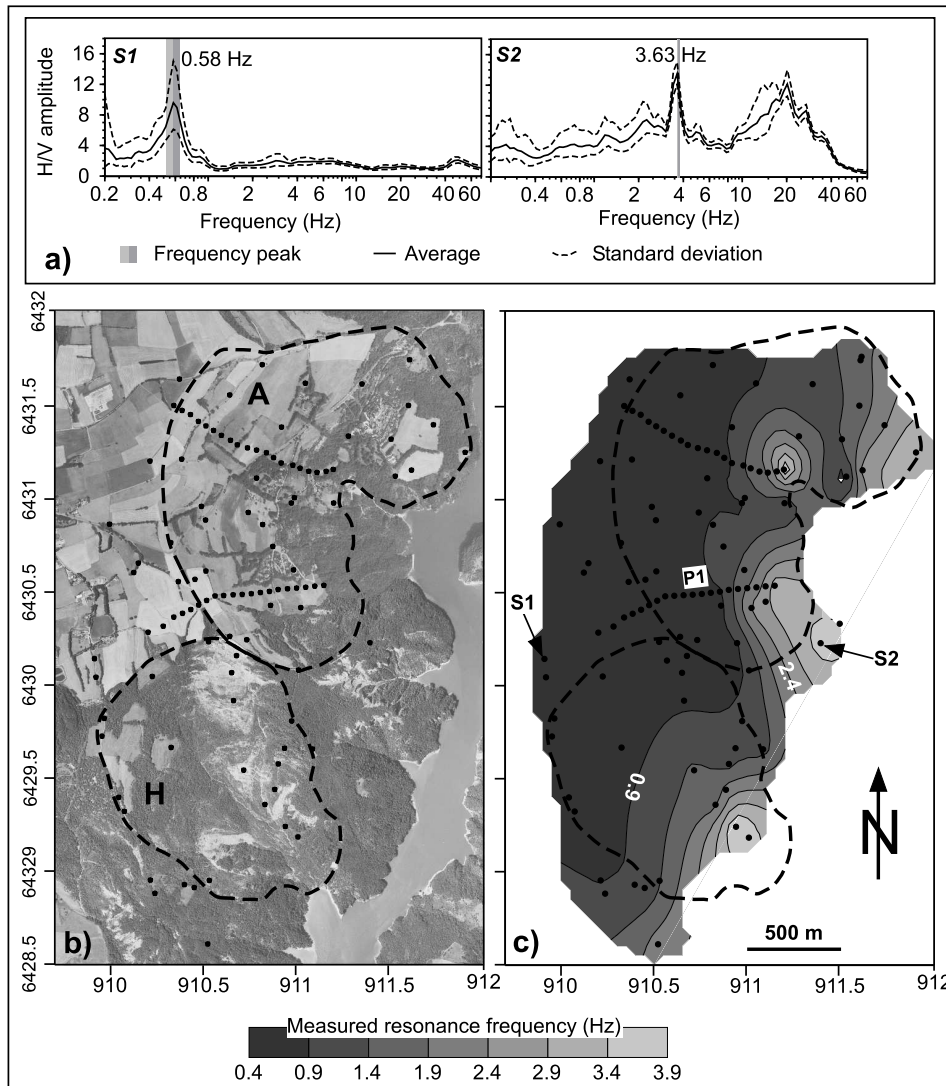


Figure 7: H/V measurements. **a)** H/V curves at points S1 and S2 (location on Fig. 7c). **b)** Location of the 104 measurement points with the landslides limits (dashed line); A: Avignonet landslide; H: Harmalière landslide. **c)** Resonance frequency map.

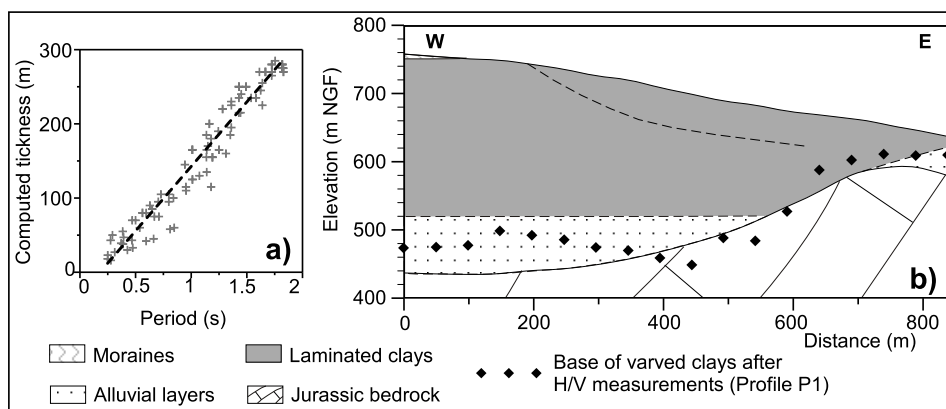


Figure 8: Frequency to thickness calibration for H/V profile P1 (location on Figs. 7b and 7c). **a)** Computed thickness as a function of the measured resonance period (inverse of the frequency) for the 104 stations. Linear regression (black dashed line) gives a determination coefficient r^2 equals to 0.95. **b)** Comparison of the geometry of the top of the Jurassic bedrock (Blanchet, 1988) and the bottom of the laminated clays after H/V measurements along profile P1 (location in Fig. 7c). Dashed line depicts the lower limit of the Avignonet landslide deduced from inclinometer data.

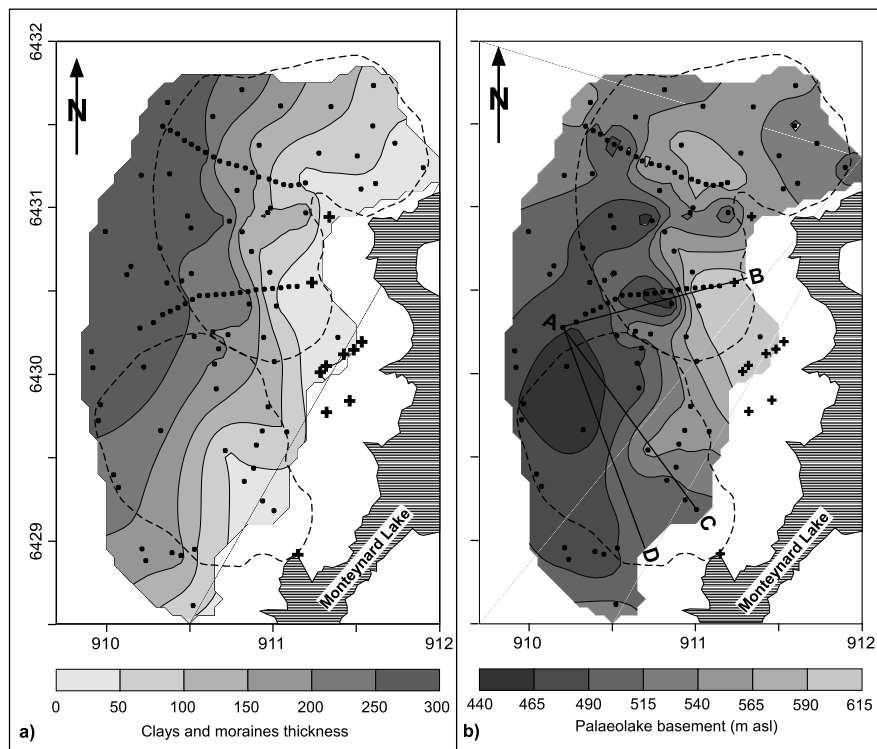


Figure 9: Soft layer thickness maps. **a)** Soft layer thickness map. The absolute error between computed and interpolated thicknesses is 5.29%. **b)** Lake Trièves palaeotopography in the study area. The absolute error between computed and interpolated palaeotopography is 0.98%. Dots refer to seismic measurements and crosses to field observations; dashed lines stand for Avignonet and Harmalière landslides limits.

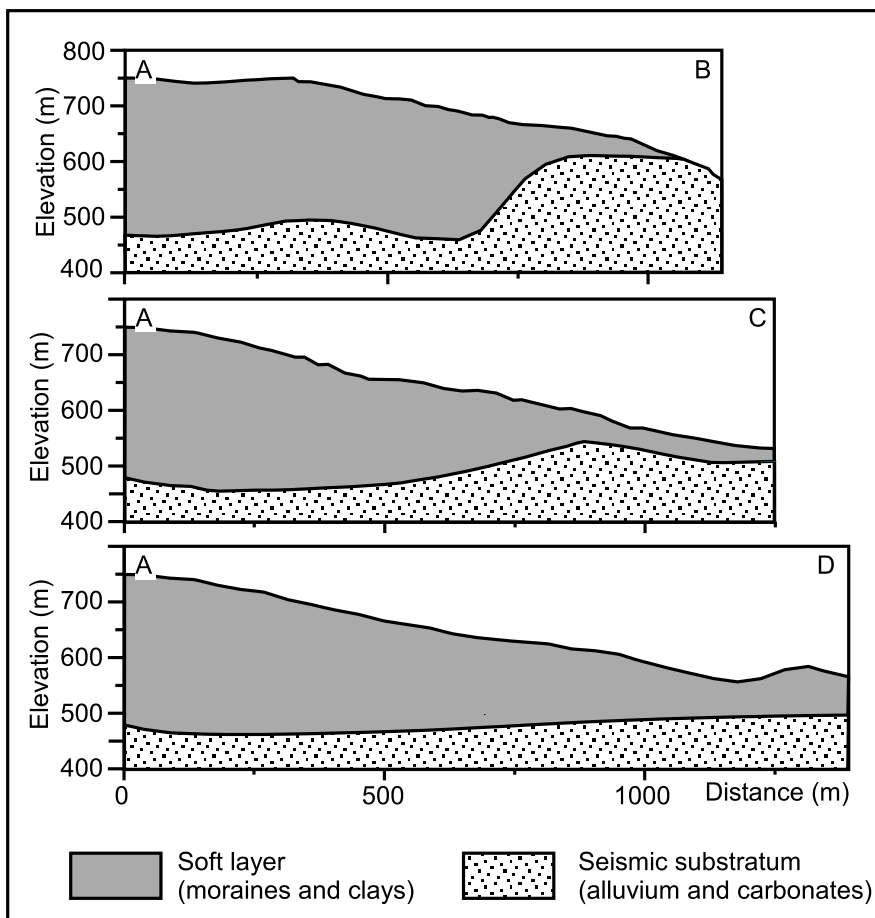


Figure 10: Cross-sections of the paleolake basement (location in Fig. 9b).

Figure 5 910.0 910.5 911.0 911.5 912.0

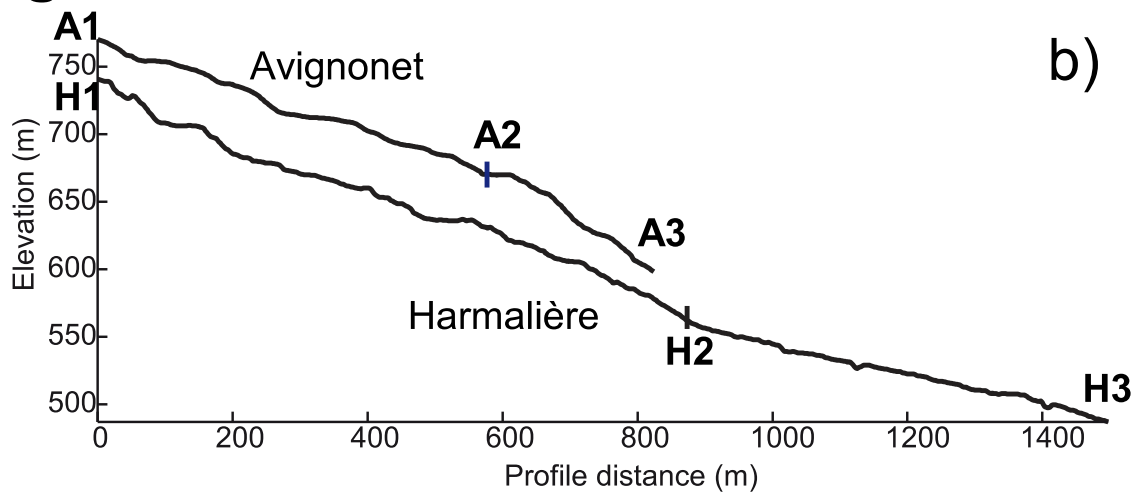
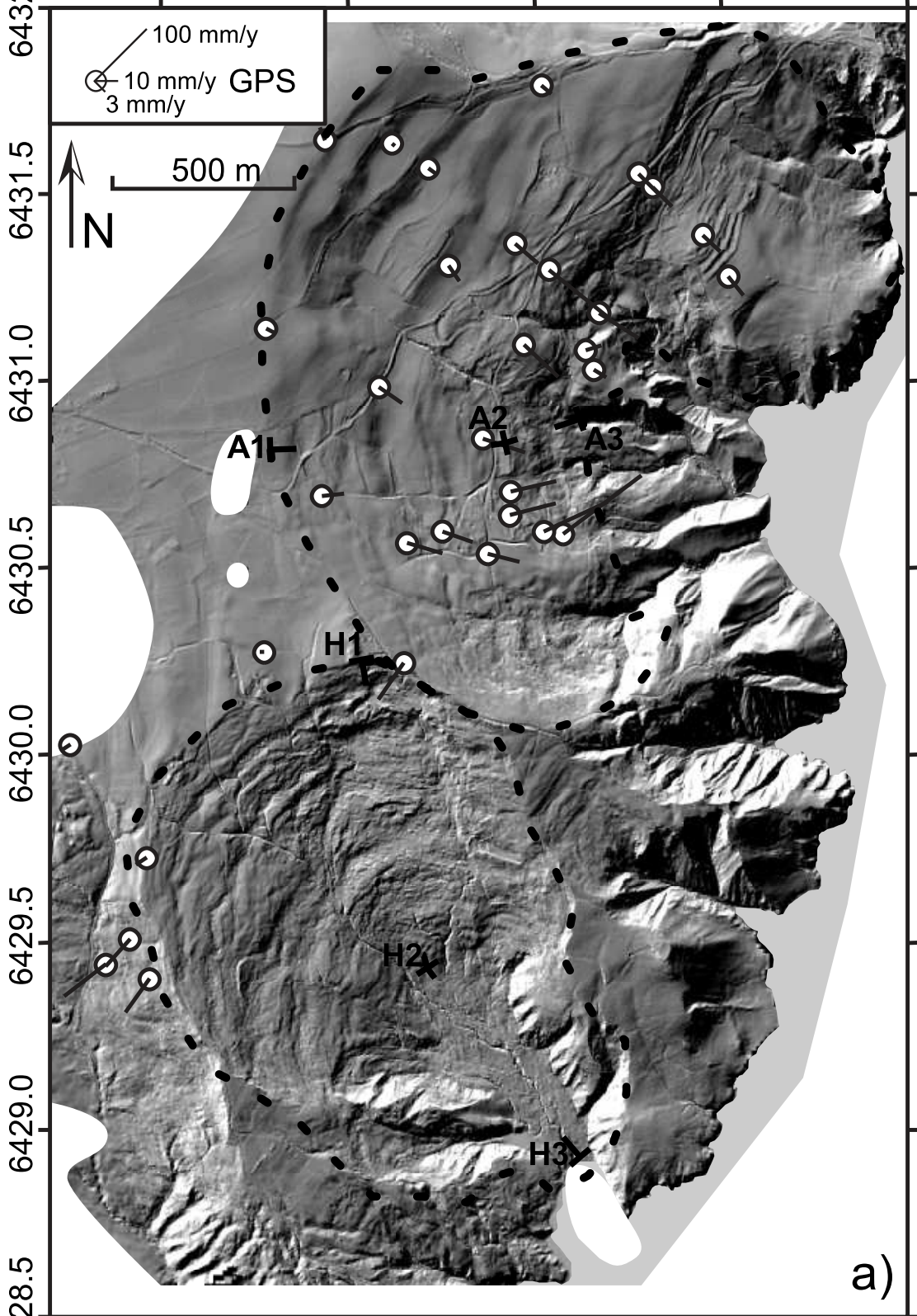


Figure 9

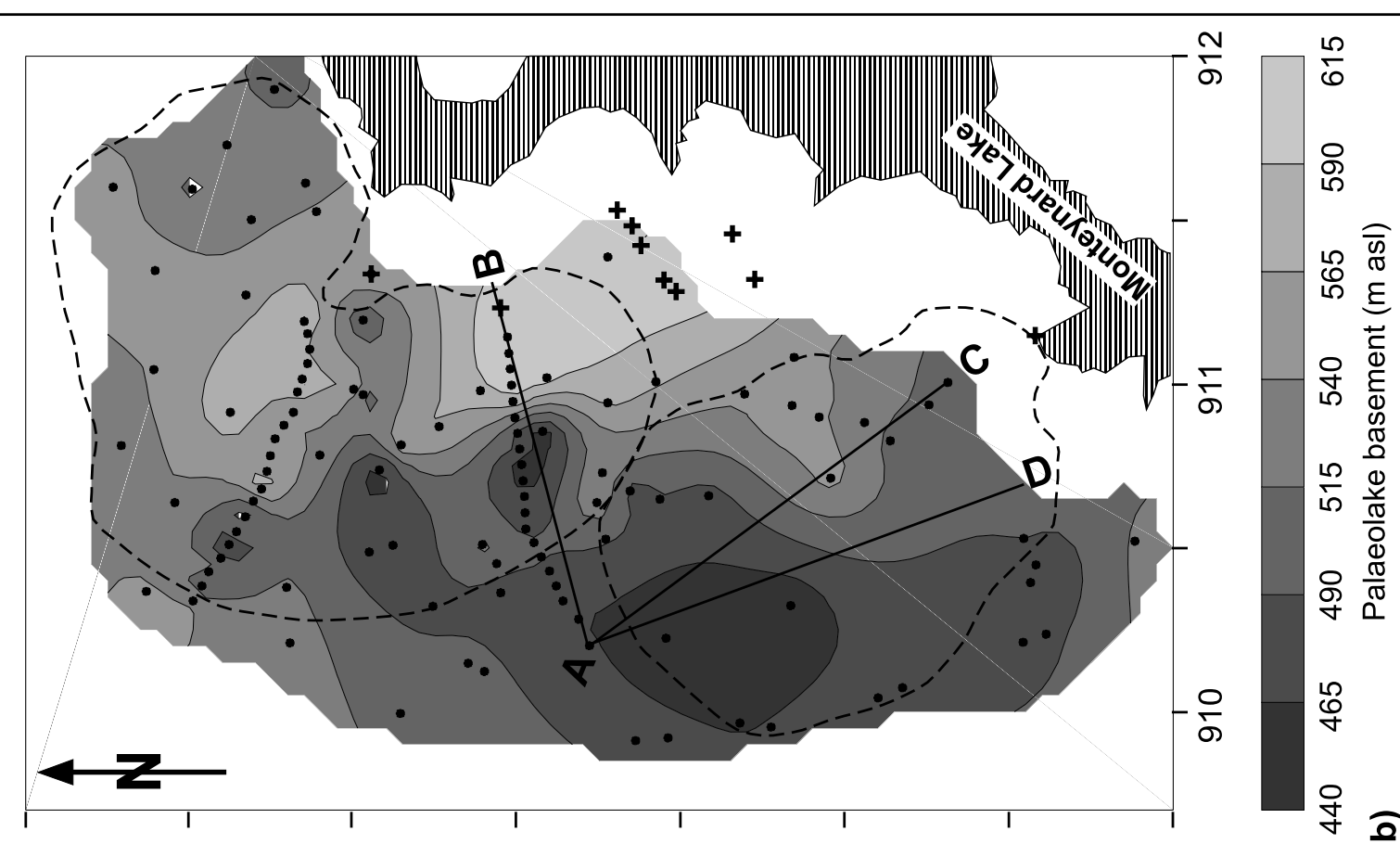
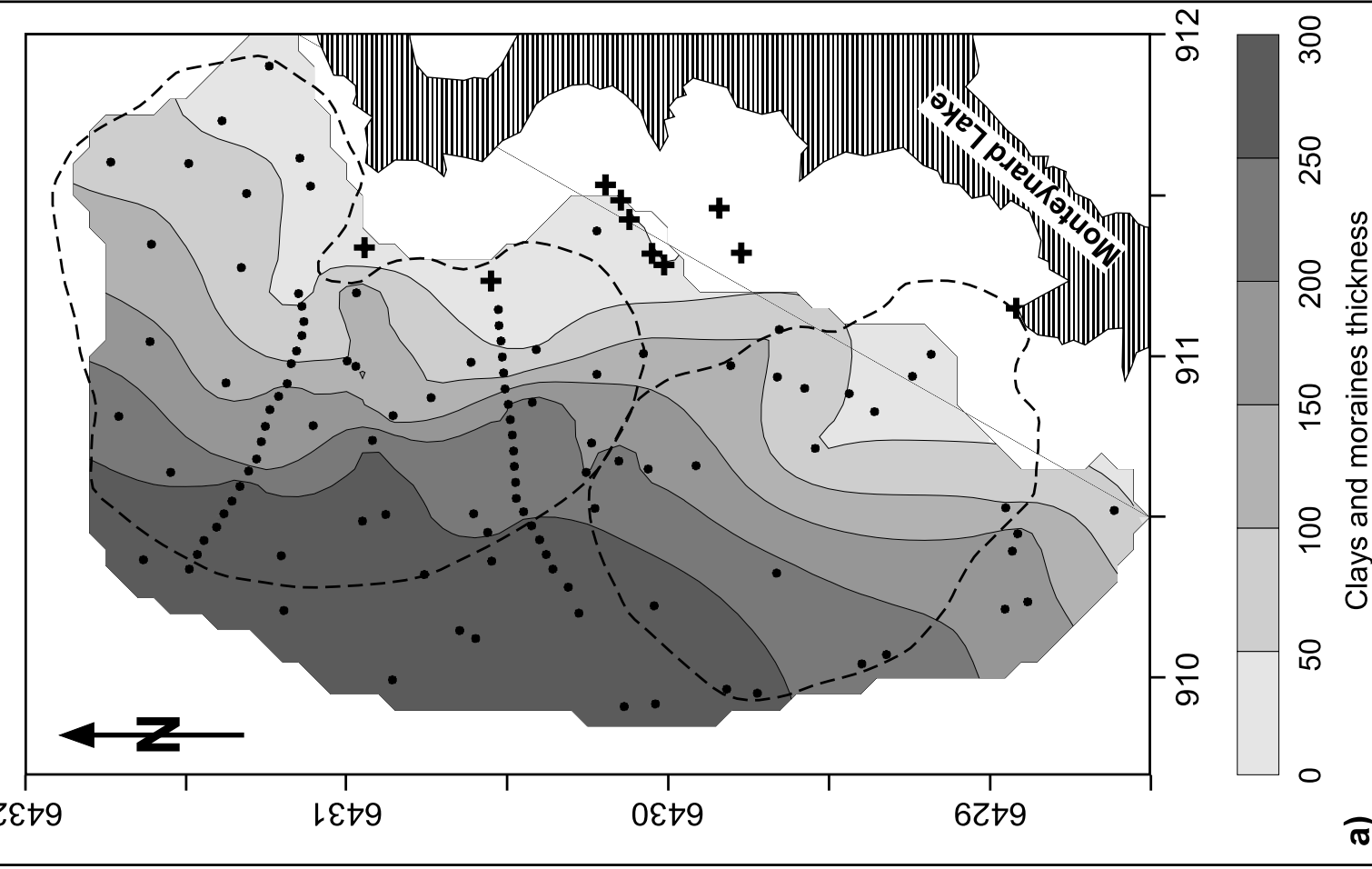
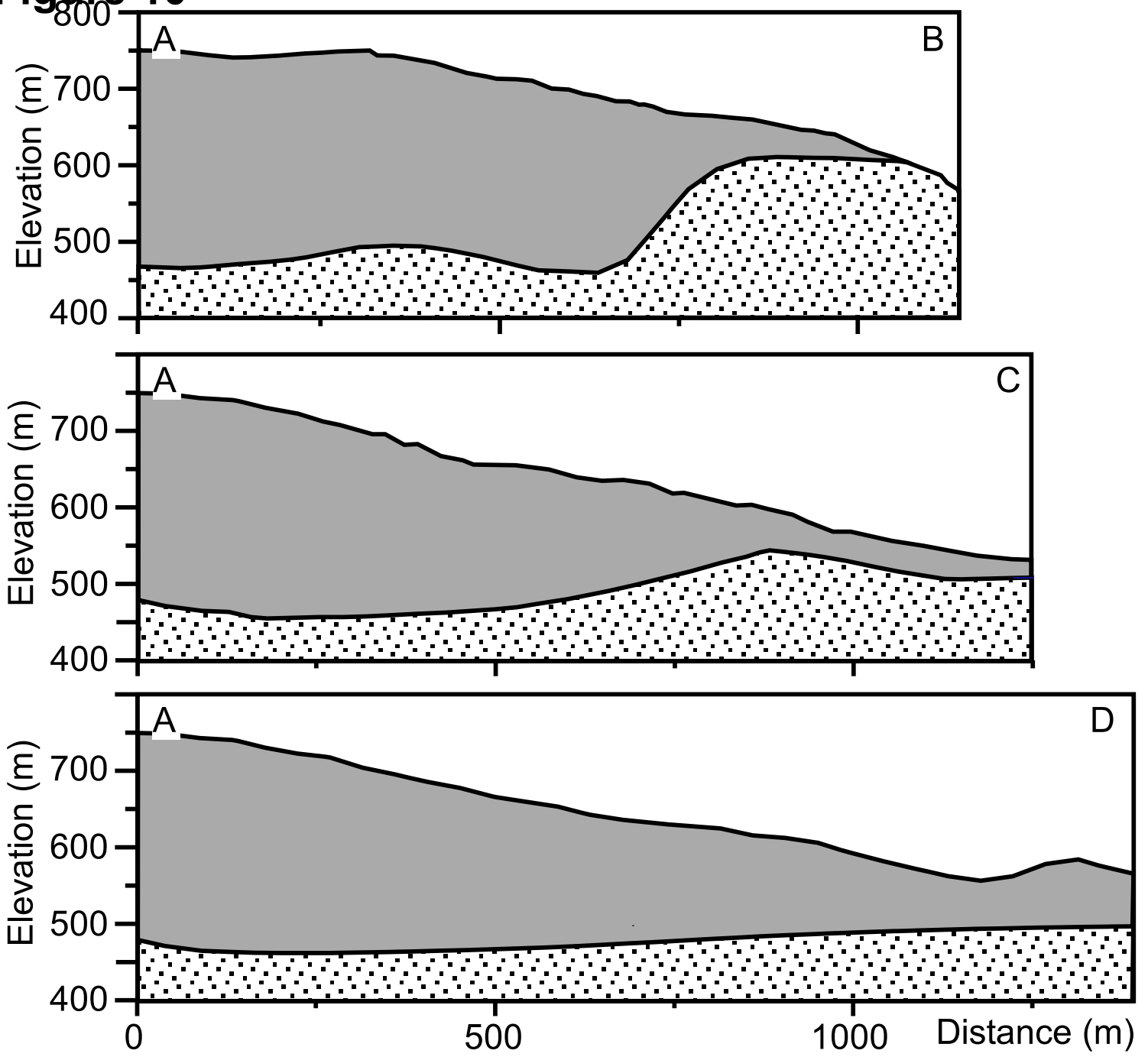

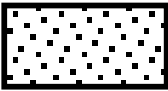


Figure 10



 Soft layer (moraines and clays)  Seismic substratum (alluvium and carbonates)