

Control of landslide retrogression by discontinuities: evidences by the integration of airborne- and ground-based geophysical information

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Abstract The objective of this work is to present a multi-technique approach to define the geometry, the kinematics and the failure mechanism of a retrogressive large landslide (upper part of the La Valette landslide, South French Alps) by the combination of airborne (ALS) and terrestrial (TLS) laser scanning data and ground-based seismic tomography data. The advantage of combining different methods is to constrain the geometrical and failure mechanism models by integrating different source of information.

Because of an important point density at the ground surface (4.1 pt.m^{-2}), a small laser footprint (0.09 m) and an accurate 3D positioning (0.07 m), ALS data are adapted source of information to analyze morphological structures at the surface. Seismic tomography surveys (P-wave and S-wave velocities) may highlight the presence of low seismic-velocity zones which characterize the presence of dense fracture networks at the sub-surface. The surface displacements measured from TLS data over a period of two years (May 2008-May 2010) allow one to quantify the landslide activity at the direct vicinity of the identified discontinuities. An important subsidence of the crown area with an average subsidence rate of 3.07 m.year^{-1} is determined. The displacement directions indicate that the retrogression is controlled structurally by the pre-existing discontinuities.

A conceptual structural model is proposed to explain the failure mechanism and the retrogressive evolution of the main scarp. Uphill, the crown area is affected by planar sliding included in a deeper wedge failure system constrained by two pre-existing fractures. Downhill, the landslide body acts as a buttress for the upper part. Consequently, the progression of the landslide body downhill allows the development of dip-slope failures and coherent blocks start sliding along

39 planar discontinuities. The volume of the failed mass in the crown area is estimated at 500,000 m³
40 with the Sloping Local Base Level method.

41

42 **Keywords:**

43 *Slope failure, ALS data, TLS data, Seismic tomography, Discontinuity, Geological model*

44 **Introduction**

45 A challenge to progress in landslide research is to define the geometry of the failed mass
46 and the possible failure mechanism precisely in order to better forecast their spatial and
47 temporal pattern of development. From a geological point of view, unstable slopes
48 affected by landslides can be broadly divided into strong rock-types and soft rock-types
49 of failure (according to the ISRM classification of rock and soil strength; ISRM, 1981;
50 Hoek and Bray, 2004) with a transitional evolution among these two broad categories. In
51 most strong rock slopes, pre-existing discontinuities control the landscape morphology.
52 As a consequence, unfavorable small-scale pre-existing fractures are often the main
53 predisposing factors of large landslides (Cruden, 1976; Agliardi et al., 2001; Sartori et al.,
54 2003; Hoek and Bray, 2004; Eberhardt et al., 2005; Jaboyedoff et al., 2009). In soft rock
55 slopes (e.g. weakly cemented sedimentary units such as highly weathered and fractured
56 rocks, conglomerates, sandstones and clays), the heritage of pre-existing structures can
57 also play an essential role on the failure mechanisms. For example, Irfan (1998) showed
58 that the slope behaviour in saprolitic soil is controlled not only by the weathered material
59 itself but also by relict discontinuities, particularly when these are unfavorably oriented
60 with respect to the slope face. Undercut slopes affected by buttress removal are among
61 the typical failure mechanisms that occur in soft rocks, leading to the development of
62 shear zones at depth (Leroueil 2001; Cruden and Martin 2004).

63 In landslide investigations, a combination of several direct and indirect techniques is very
64 often used, and several complementary ground-based and airborne-based technologies
65 have been developed in the last decade to provide spatially-distributed information about
66 the structure. In combination with field observations and classical geotechnical
67 investigation, the ground-based techniques are mainly 2D and 3D electrical resistivity and
68 seismic tomographies (Jongmans and Garambois 2007) and the airborne-based techniques
69 are mainly radar interferometry techniques (InSAR), Light Detection and Ranging
70 techniques (LiDAR) and correlation of optical imageries (Travelletti et al., in press;
71 Jaboyedoff et al., 2009). Terrestrial Laser Scanning (TLS) and Airborne Laser Scanning
72 (ALS) are very efficient techniques for characterizing the morpho-structure (Feng and
73 Röshoff, 2004; Slob and Hack, 2004; Jaboyedoff et al., 2009) and the kinematics of
74 landslides (Rosser et al., 2007; Monserrat and Crosetto 2008; Travelletti et al., 2008;
75 Prokop and Panholzer 2009) because they provide a rapid collection of field
76 topographical data with a high density of points within a range of several hundreds of
77 meters. Possible mechanisms affecting the slope can then be estimated from the
78 displacement vectors at the ground surface (Crosta and Agliardi, 2003; Jaboyedoff et al.,
79 2004a) such as the geometry of the slip surface (Casson et al., 2005; Travelletti et al.,
80 2008; Oppikofer et al., 2009).

81 In strong rock-types of failure, the morpho-structures identified at the ground surface
82 often reflect the internal geometry of the deformation (Agliardi et al., 2001; Eberhardt et
83 al., 2005). The extensions of persistent structures in depth are more difficult to identify in
84 soft rocks because these lithologies are very often affected by low persistence, closely
85 spaced joints that occur in a wide variety of orientations. The landslide kinematics in soft
86 rock can be both controlled by regional discontinuities and recent internal failure surface
87 under development in the rock mass (Irfan, 1998). The development of new circular or
88 planar failures which partly encompass the intact rock is therefore possible as observed,
89 for example, in weathered basalts or sandstones (Hoek and Bray, 2004). Another example

hal-00649682, version 1 - 8 Dec 2011

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90 of weak rocks is evaporites lithology that is particularly exposed to dissolution processes.
91 They generally form smooth topographies at small scale which make the identification of
92 regional discontinuities difficult from morpho-structure analyses at the ground surface or
93 through DEMs (Travelletti et al., 2010). Additional surveys are thus necessary to
94 complement this lack of information in depth.
95 In the last decade, the applications of seismic tomographies for landslide investigations
96 showed that S and P-wave velocities are of interest techniques to characterize properties
97 such as the layering, the degree of fracturing and the stiffness of the material (Grandjean
98 et al., 2006; Jongmans et al., 2009). The internal strain affecting soft-rock landslides
99 usually induces a velocity contrast between the unstable mass and the stable bedrock
100 (Caris and van Asch, 1991, Méric et al., 2007; Jongmans and Garambois, 2007).
101 Grandjean et al. (2006) showed that seismic velocities are much more sensitive to the
102 degree of fracturing than the electrical resistivity tomography techniques (ERT) which
103 complement geological and geomorphological assessments.
104 Still, a major difficulty consists in interpreting and integrating all the available data in a
105 coherent framework to provide a complete picture of the landslide structure.
106 This work presents a multi-technique approach to characterize the structure of the upper
107 part of the La Valette landslide (South French Alps) by combining high-resolution
108 seismic tomographies, airborne and terrestrial LiDAR surveys (ALS, TLS) and
109 geomorphological analyses. First, the geomorphological and historical setting of the
110 landslide is presented; then the methodology used to analyse the multi-source data is
111 detailed and the Sloping Local Base Level (SLBL) method is applied to estimate the
112 landslide volume. Finally a kinematic model explaining the failure mechanism in the
113 upper part of the landslide is proposed and discussed.
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115 **Geomorphological and historical features of the** 116 **La Valette landslide**

117 The La Valette landslide, originated in 1982, is one of the most important large and
118 complex slope movements in the South French Alps. The landslide associates two styles
119 of activity: a mudslide type of behavior with the development of a flow tongue in the
120 medium and lower part, and a slump type of behavior with the development of several
121 rotational and planar slides in the upper part at the main scarp. The landslide extends over
122 a length of 2 km for a variable width of 0.2 km in the lower and medium parts, to 0.5 km
123 in the upper part (Fig. 1a). The maximum depth, estimated by seismic and electrical
124 resistivity tomography and geotechnical boreholes, varies from 25 m in the lower and
125 middle parts (Evin 1992; Travelletti et al., 2009) to 35 m in the upper part (Le Mignon
126 2004). The mean slope gradient is ca. 30° in the scarp area and ca. 20° in the mudslide
127 area (Fig. 1b). The volume of the landslide body is estimated at $3.5 \times 10^6 \text{ m}^3$.
128 The landslide affects a hillslope located uphill of the municipality of Saint-Pons (Alpes-
129 de-Haute-Provence), and poses a significant threat for the 170 community housings
130 located downstream (Le Mignon and Cojean 2002). The occurrence of rapid mudflows
131 triggered from the landslide body and in the scarp area in the 1980s and 1990s conducted
132 to the development of an early-warning system since 1991 composed of a survey network
133 of benchmarks, optical and infra-red camera monitoring and the installation debris height
134 detection sensors in the run-out channel, and the drainage of the lower part of the
135 landslide.

136 **Geological setting**

137 From a geological viewpoint, the La Valette landslide is located at the overthrust fault
138 between two major lithologies outcropping in the geologic window of the Barcelonnette
139 basin (South-east France; Fig. 2a):

- 140 - An autochthonous formation represented by the closely stratified Callovo-Oxfordian
141 black marls (e.g. “Terres Noires”) and characterized by a typical landscape of
142 badlands. This formation is located in the middle and the lower parts of the slope. The
143 bedding plane is characterized with a decametric alternance of carbonate beddings
144 within the marls. This formation dips constantly towards 083° with an inclination of
145 23° in the landslide surroundings;
- 146 - An allochthonous formation represented by two nappes and in which the upper part of
147 the landslide has developed. The basal nappe is a tectonic wedge belonging to the
148 Pelat Nappe and is composed of highly fractured flysch and planctonic carbonates of
149 the Turonian and Paleocene Superior age (BRGM 1974). This formation has a few
150 dozen of meters of thickness at the location of the main scarp with an average dip
151 direction and dip of 135°/30°. The Pelat Nappe is overlaid by the upper Autapie
152 Nappe composed of highly fractured Helminthoid flyschs, sandstones, marls and
153 schists. This formation is dated at the Upper Cretaceous–Upper Eocene (BRGM
154 1974). The tectonic discordance between the autochthonous and the allochthonous
155 (052°/16°) materializes the major thrust fault delimiting the Barcelonnette basin, and
156 constitutes a weak zone where many landslide source areas are located (Le Mignon
157 2004). Generally, the bedding plane, the fold axes and the schistosity are very difficult
158 to identify with certainty at the outcrop scale due to the high variability of the
159 orientation measurements and the bad rock mass quality which does not ensure that
160 the rock outcrop is in place. Therefore, the small-scale geological observations of
161 persistent structures carried out in the field indicate only the regional structural and
162 tectonic patterns and cannot directly be integrated in a local geological model of
163 discontinuities.

164 **Hydro-geological setting**

165 From a hydro-geological viewpoint, the tectonic discordance has an important role on the
166 landslide hydrology. Due to the high heterogeneity of the landslide material and of the
167 highly dislocated texture of the flysch, both materials are considered as aquifers at the
168 scale of the landslide. At the opposite, the black marls formation is considered as an
169 aquitard (Dupont and Taluy 2000; Le Mignon 2004). Consequently, the contrast of
170 permeability between the black marls and the flysch controls the spatial occurrence of
171 several springs and marshy areas observed in the direct vicinity of the overthrust fault
172 between the elevation 1870 m and 1950 m (Fig. 2a, 2b). According to Le Mignon (2004),
173 a spring (the “Rocher Blanc” spring at 1900 m) is currently partially buried by the
174 landslide. Consequently a deep water circulation affects the hydrological regime of the
175 upper part of the landslide but the characteristics of the water flows (fluxes, quality) are
176 unknown. Near the North West boundary of the landslide, springs are remarkably aligned
177 at the elevation 1880 m. They are likely connected to the Rocher Blanc spring through a
178 discontinuity buried by the landslide. Remediation works were built by the local
179 stakeholders in charge of the prevention (Service of “Restauration des Terrains de
180 Montagne”, RTM) in order to drawdown and buffer the hydraulic heads within the
181 landslide. Sub-horizontal drains below the major overthrust fault were installed in the
182 1990s, but their maintenance was too difficult due to the rapid shearing of the tubes. The
183 most efficient mitigation solution has been the installation of shallow drainage systems in
184 the middle and lower parts of the landslide to impede streaming water to infiltrate the
185 landslide.

186 **Landslide historical and recent development**

187 The landslide exhibits a complex style of activity in space and time. It has developed first
188 as a rotational slide affecting the Autapie Nappe in relation to a major overthrust fault
189 following important rainfalls favoring fast snow melting (Colas and Locat 1993; Le
190 Mignon 2004). The failed mass has progressively loaded the underlying black marls
191 formation, and the landslide has developed by a series of rapid mudflows triggered in the

hal-00649682_versjon 1 - 8 Dec 2011

192 marls such as in March 1982, April 1988, March 1989 and March 1992. The most
193 important acceleration occurred in 1988 when a mudflow of 50.000 m³ triggered at the
194 elevation of 1400 m propagates over a runout distance of ca. 500 m (Colas and Locat
195 1993). Up to now, these mudflows did not mobilize the complete failed mass. The
196 displacements are monitored with topometric benchmarks since 1991 (Squarzoni et al.,
197 2005), differential dual-frequency GPS (Déprez et al., submitted) and an extensometer
198 since 2008, and at regular periods by digital correlation of satellite images (Le Prince
199 2008) and satellite radar interferometry (Squarzoni et al., 2003).
200 Two main aspects can be pointed out from these studies and from the observations by the
201 local stakeholders. The first one is the decrease of velocity (from 0.4 m.day⁻¹ to about
202 0.01 m.day⁻¹) in the middle and lower part of the landslide caused by the local
203 groundwater drawdown since the installation of a drainage system in the 1990s. The
204 second is the important activity since the year 2000 of the upper part at the Soleil Boeuf
205 crest, which is characterized by a rapid retrogression of the main scarp towards the North-
206 East and an enlargement of the landslide towards the North-West (Fig. 3). In response to
207 this worrying situation, the RTM Service has installed several additional benchmarks
208 along profiles both in the unstable and stable parts of the Soleil-Boeuf crest to monitor the
209 displacements in the crown area (Fig. 1b, Fig. 4). Actually, an accumulation of material
210 and a steepening of the slope are observed in the upper part of the landslide because of
211 the retrogression of the scarp. Consequently, the possible hazard scenario consists in the
212 undrained loading of underlying black marls formation and the triggering of new rapid
213 and mobile mudflows.
214

215 Methodology

216 The recent increase of activity in the crown and in the main scarp areas has motivated
217 further investigations to define the possible volume of material still able to fail, and to
218 better understand the failure mechanisms in that part of the slope. Geological
219 observations, small-scale morpho-structural analysis with ALS data, large-scale
220 kinematical analysis with TLS data and a morpho-structural analysis in depth with
221 seismic tomographies have been carried out since 2008 (Fig. 1b, Fig. 4).

222 Geological observations of the discontinuities in the main scarp and in the 223 crown areas

224 Field investigations in the scarp area are focused on the identification and the orientation
225 of the major morpho-structures and discontinuities measured with a geological compass
226 and mapped with a differential GPS. The landslide scarp is characterized by slopes
227 ranging from 25° to 55° over a maximum height of 80 m and a crown width of 170 m.
228 The crown area is characterized by a complex morphology formed by a dense network of
229 tension tracks and shear fissures in the weathered flysch formation of the Autapie Nappe
230 forming a graben-like morphology. Counter-slopes with accumulated water are also
231 observed. The tension cracks present a sub-vertical dip distributed along the circular
232 shape of the crown. This spatial distribution is typical for failures in soft rocks (Cornforth
233 2005). The complex morphology is cut by three main persistent discontinuities D1, D2
234 and D3 visible over a distance of several hundred meters (Fig. 5).
235 The D1 fracture coincides with the direction of the Soleil Boeuf crest which represents
236 the upper boundary of the landslide. D1 is characterized with a dip direction and a dip of
237 247°/42° ± 6°/4°. Because striations identified over the entire height of the crest present
238 an orientation nearly parallel to the dip and dip direction of D1 (calcite recrystallization,
239 pitch of S86°), the landslide is sliding above D1 without shearing component. The
240 observed striations are a direct mechanical consequence of the sliding along D1.

hal-00649682_versjon 1 - 8 Dec 2011

241 The fractures D2 and D3 are characterized with dip directions and dips of $287^{\circ}/55^{\circ}$
242 $\pm 4^{\circ}/5^{\circ}$ and $166^{\circ}/44^{\circ} \pm 10^{\circ}/4^{\circ}$ respectively. D2 is located in the North-East of D1 and D3
243 is the conjugate fracture of D2. The landslide is sliding along D2 and D3 with a shearing
244 component. The most active part of the landslide is actually constrained between these
245 two discontinuities.

246 **Laser scanning surveys**

247 *Acquisition of Airborne laser scanning (ALS) data*

248 In order to determine the spatial extent of the main morpho-structural features, small-
249 scale analysis of ALS has been carried out. The ALS survey was performed in July 2009
250 with the handheld airborne mapping system of the *Helimap* company) at a constant
251 elevation of 300 m above the ground topography inducing a laser footprint at the ground
252 surface of about 0.09 m (Vallet and Skaloud, 2004). The measurement device is
253 composed of a GPS receiver (Topcon Legacy GGD with a record frequency of 5Hz), an
254 inertia measurement unit (IMU, record frequency of 500 Hz) which provides the
255 orientation of the laser beam in space and a scanner unit (Laser Riegl Q240i) configured
256 to record last pulses of the ground surface with an acquisition frequency of $10^{\circ}000$ pts.s⁻¹.
257 Table 1 summarizes the specification of the ALS survey. The orientation of the system is
258 obtained in real-time with an accuracy estimated at 0.07 m. The maximum scanner range
259 is about 850 m with an aperture angle of 60°. After vegetation filtering, an average point
260 density of 4.1 pt.m⁻² is obtained. A 0.5 m-mesh DEM from the ground surface elevation
261 points has been generated with a Delaunay triangulation. The DEM was then used to
262 calculate a shaded relief map and a difference map with a 10 m-mesh DEM interpolated
263 from topographic contour lines before the landslide event (maximal elevation error of
264 10 m).

265 *Acquisition of Terrestrial Laser Scanning (TLS) data*

266 Displacement monitoring of the upper part of the landslide has been carried out by
267 repeated TLS data acquisitions. The displacement monitoring device consists of a long-
268 range terrestrial laser scan Optech ILRIS-3D based on the time-of-flight distance
269 measurements using an infrared laser (Slob and Hack 2004). Mirrors inside the scanner
270 allow the acquisition of a 40° wide and 40° high field of view in a single acquisition with
271 about 2500 pts.s⁻¹ with an effective range up to 800 m in field conditions (Table 1).
272 Seven TLS datasets were acquired over the period 18 May 2008 to 27 May 2010 from the
273 same base position (Fig. 5); the scanned area was orientated in the direction of the
274 discontinuity D2 at a distance of 130 m from the base. At that distance the beam width
275 diameter is estimated at 0.03 m. The discontinuity D1 (Soleil Boeuf crest) along the main
276 scarp was systematically included in the scanning. The TLS datasets comprise 9 to 12
277 million points and the resulting mean point density on the ground surface is about
278 150 pts.m⁻² at a distance of 130 m. Only the last return pulse is registered to maximize the
279 number of points at the ground surface.

280 *Vegetation filtering, co-registration and georeferencing of the sequential TLS*
281 *datasets*

282 The TLS datasets were processed and analyzed using the *Polyworks v.11* software
283 (InnovMetric 2009). The vegetation filter consists of an automatic selection of the points
284 localized beyond a minimum height relative to a low-resolution square-grid DEM surface
285 computed on the sequential point clouds. In this study, the mesh size of the low resolution
286 DEM was fixed at 0.5 m and the minimum height at 0.1 m. The filtering result is
287 systematically controlled and manually refined. A co-registration procedure is then used

hal-00649682, version 1 - 8 Dec 2011

288 for aligning the sequential TLS point clouds in the same coordinate system. The co-
289 registration applied in this study follows the methodology proposed by Teza et al. (2007)
290 and Oppikofer et al. (2009). The sequential point clouds alignment is limited to a stable
291 part in the image corresponding to the Soleil Boeuf crest. First a manual alignment
292 procedure is used; then an automated Iterative Closest Point (ICP) algorithm is applied to
293 minimize (least square method) the distance between the points belonging to the different
294 sequential datasets. The co-registration procedure mainly depends on the accuracy of the
295 ICP algorithm which is very sensitive to the roughness of the terrain and the accuracy of
296 the measurements (Lee et al., 1999). Rough terrains yield higher reliability in the co-
297 registration. Therefore a high point density was systematically acquired in this area to
298 increase the topographic resolution ($\sim 0.3 \text{ pt.cm}^{-2}$). A good confidence is given to the co-
299 registration quality because of the large size of the stable area of the image used for the
300 co-registration (1000 m^2) in reference to the size of the moving area not introduced in the
301 co-registration procedure (4300 m^2). The direct proximity of both areas in the datasets is
302 also an advantage for an accurate co-registration.
303 For the absolute georeferencing, the ALS point clouds were used as a reference. The
304 sequential TLS datasets were aligned as single point clouds on the ALS point cloud. The
305 co-registration accuracy of the sequential point clouds is thus not affected by the
306 georeferencing accuracy of the ALS point cloud, and is estimated at 0.07 m for the planar
307 and vertical accuracy (Vallet and Skaloud, 2004).

308 *Accuracy of the TLS point clouds*

309 In order to assess the accuracy associated with the TLS measurements, repetitivity
310 measurements were realized on a planar stratum of the main scarp, and corresponding to
311 the black marls formation (9 m^2). This test zone is located approximately perpendicular to
312 the laser beam direction at a distance of 120 m from the TLS base. The average point
313 spacing was fixed at 0.15 m. The point cloud acquisition was compared to a reference
314 (first acquisition, May 2008) using the ICP algorithm to calculate the misfit between each
315 pair of points in both acquisitions theoretically located at the same position. The
316 repetitivity analysis indicates that the measurement error of the TLS used in this study
317 follows a normal distribution characterized with an average error μ of $1.0 \cdot 10^{-3} \text{ m}$ and a
318 standard deviation σ of $1.2 \cdot 10^{-2} \text{ m}$. This calculated error is in agreement with the range of
319 error given by the manufacturer.
320 The accuracy of co-registration procedure is given by the residual 3D misfit computed on
321 the stable part of the TLS of the 18 May 2008 taken as the reference (Table 2). The same
322 procedure is applied to assess the error of the absolute positioning relative to the ALS
323 survey. The higher error in the absolute positioning is mainly related to the lower point
324 density and accuracy of the ALS datasets providing less geomorphological details than
325 the TLS survey (Table 2).

326 *Displacement characterization and quantification*

327 The displacements are calculated by comparing the TLS datasets with the reference. Two
328 methods are used to quantify the displacements from the original point clouds. The first
329 method is based on shortest distance comparison of point clouds. The second uses
330 displacements of Specific Points (SPs)
331 The shortest distance (SD) comparison consists of computing for each point the distance
332 to its nearest neighbor in the reference point cloud. This method is particularly useful to
333 detect spatially distributed changes if the direction of movement is unknown and to define
334 zones with different displacement directions (Oppikofer et al., 2009). In order to
335 determine the vertical displacement affecting the upper part of the landslide, the SD is
336 constrained to compute displacement only along the vertical direction (SD_v) assuming a
337 tolerance angle for the vertical direction of $\pm 10^\circ$. The results are therefore comparable to
338 elevation changes computed with differential DEMs (Bitelli et al., 2004). The accuracy of
339 the vertical displacement depends on two independent factors: (i) the co-registration

hal-00649682_v331-27 Dec 2011

340 accuracy, and (ii) the computed distance D according to the tolerance angle which gives a
341 maximal error E_{\max} ($E_{\max} = D * \sin(10^\circ)$).

342 The observed movement of Specific Points (SPs) allows one to define the direction of
343 displacements. In this study, natural SPs were chosen (Fig. 6). They consist of tree
344 stumps recognizable in the unfiltered sequential point clouds. In order to assess the
345 displacements of the SPs in the crown area, a method based on the roto-translation
346 technique is used (Montserrat and Crosetto, 2008; Oppikofer et al., 2009). It takes into
347 account both translation and rotation of individual objects and uses the very high density
348 available in the point clouds. Eleven SPs were triangulated in the plane normal to the
349 laser viewing direction in order to minimize the effect of shadow zones in the
350 interpolation (Fig. 6). To calculate the true displacement field, the center points of the SPs
351 in the first acquisition is determined by averaging the X, Y Z positions of the points
352 forming the SPs. Then the triangulated SPs of the reference are aligned on their
353 corresponding triangulated SPs in the sequential point clouds using the ICP algorithm
354 implemented in *Polyworks* (Fig. 6). Finally, the displacement vectors of the SPs are given
355 by the initial and the final position of the center points of the first acquisition. Because the
356 SPs are very well defined, the error mainly depends on the co-registration accuracy of the
357 sequential point clouds.

358 **Seismic Tomography investigation**

359 As the laser scanning data provide only information on the structure and the kinematics
360 visible at the ground surface, additional geophysical data to obtain information in depth
361 were acquired in order to better constrain the interpretation. Seismic tomography has
362 proven to be an efficient technique to detect the contact between a landslide body (highly
363 fractured) and a stable bedrock in the same type of geological setting as for La Valette
364 (Jongmans et al., 2009, Grandjean et al., 2006; Grandjean et al., 2007). Three seismic
365 tomography surveys of P and S-waves velocities (L1, L2, L3) have been carried out in the
366 upper part of the landslide to characterize the seismic velocities (V_p , V_s) of the failed
367 material and determine the extension in depth of the discontinuities D1 and D2 (Fig. 5).
368 Two cross-sections were installed along the direction of the main slope and one cross-
369 section was installed perpendicular to the previous line in order to cross the discontinuity
370 D2.

371 The two longest devices are composed of 24 geophones (resonance frequency of 10 Hz)
372 spaced each five meters in order to obtain a sufficient investigation depth for a large scale
373 characterization of the landslide structure. The shortest devices, composed of 24
374 geophones spaced each two meters, bring information on the fracturing between the failed
375 material and the bedrock. For the seismic source, one hundred grammes (100 gr) of
376 pentrite were used for each shot. The processing of the first arrival travel time P wave
377 was carried out with the Rayfract seismic tomography software based on the wavepath
378 Eikonal travelttime inversion algorithm (Schuster and Quintus-Bosz 1993). Figure 7a
379 shows an example of the arrivals of the P waves and Surface waves. Dispersion of surface
380 waves is closely related to the structure and properties of the landslide material and, in
381 particular, to shear wave velocity. The vertical distribution of the shear wave velocity can
382 be estimated on the basis of the dispersion analysis of different kinds of surface waves
383 contained in the P-wave seismic records (Fig. 7b). The dispersion analysis results in the
384 generation of a dispersion curve (frequency vs. phase velocity) for each geophone
385 location. The shear waves velocity sections were therefore obtained using a spectral
386 analysis of the surface waves (SASW method) with the surf96-CPS program (Hermann
387 1987) which allows one to analyse the dispersive character of surface waves (McMechan
388 and Yedlin 1981) and to obtain a S-wave vertical velocity profile by 1D inversion of the
389 dispersion curves (Tarantola 2005) (Fig. 7b). To derive a 2D section, the 1D shear-wave
390 velocities inverted from each local dispersion curve are then interpolated along each
391 seismic line.

392 In order to georeference the 2D tomographies, the location of each geophones was
393 measured with a dGPS (horizontal and vertical accuracy of 0.04 m and 0.07 m). In each

394 profile, the geophone locations are then projected on a straight line in the local coordinate
395 system calculated with a linear regression. The equations of the regression line allow one
396 to allocate a 3D position in the local coordinate system for all nodes of the tomography.

397 **Estimation of the volume of the failed mass**

398 To estimate the volume of the failed mass, the SLBL (Sloping Local Base Level) method
399 has been used. It is a generalization of the base level concept proposed by Mills (2003)
400 and is defined as a surface above which the rocks are assumed to be erodible by
401 landslides due to the absence of buttress (Golts and Rosenthal 1993; Jaboyedoff et al.,
402 2004b). This method was successfully applied both on hard and soft rock slopes
403 (Jaboyedoff et al., 2009; Travelletti et al., 2010). The SLBL method is used in this study
404 by integrating the information from the seismic survey and the TLS interpretation.

405 The SLBL algorithm is based on an iterative routine that replaces the elevations of any
406 mobile point of a digital elevation model (DEM) by the mean value of the altitude of its
407 neighbours, allowing a certain tolerance. Points with an altitude (z_i) larger than the mean
408 of their neighbours are replaced by the mean value of the two neighbours $(z_{i-1} + z_{i+1})/2$ or
409 by this value plus a tolerance C . Explicitly, in 2D, the procedure can be formalized as
410 follows (Jaboyedoff et al., 2004b):

$$411 \quad 412 \quad \text{If } z_i > (z_{i-1} + z_{i+1})/2 \text{ then } z_i = (z_{i-1} + z_{i+1})/2 \quad (1)$$

413
414 The result is a straight line between z_{i-1} and z_{i+1} . The introduction of the tolerance value C
415 leads to a second-degree curve:

$$416 \quad 417 \quad \text{If } z_i > (z_{i-1} + z_{i+1})/2 - C \text{ then } z_i = (z_{i-1} + z_{i+1})/2 \pm C \quad (2)$$

418
419 The tolerance can produce holes between points, e.g with an altitude smaller than the
420 altitude of the surroundings points. To avoid this, two additional conditions can be added
421 to ensure that the new point has a higher altitude than the lowest of its neighbours:

$$422 \quad 423 \quad \text{If } z_i > (z_{i-1} + z_{i+1})/2 - C \text{ and } (z_{i-1} + z_{i+1})/2 \pm C > z_{i-1} \text{ or } (z_{i-1} + z_{i+1})/2 \pm C > z_{i+1} \text{ then } z_i = (z_{i-1} \\ 424 \quad + z_{i+1})/2 \pm C \quad (3)$$

425
426 If it is not the case, the altitude z_i of the point i is replaced by the lower altitude of its
427 neighbours. The procedure is iterative, and is stopped once the change between two
428 iterations is near a zero value and a failure surface nearly circular is obtained if the
429 tolerance used is very different from zero. In order to take into account planar sliding
430 along discontinuities, a small tolerance has to be used to fit the calculated failure surface
431 on the observed discontinuities. In 3D, the procedure is similar, but the test is then
432 performed using the highest and lowest values among the four closest neighbours. Some
433 points must be fixed during the computation, to avoid the calculation of a flat topography.
434 The discontinuities determined in this study from the field observations, the seismic
435 tomographies and the TLS surveys are used to constrain the SLBL computations.
436

437 **Results**

438 **Morpho-structural analysis**

439 The combined analysis of the geological field observations, the shaded relief map and the
440 differential DEM map allows one to propose a kinematic model of the landslide
441 retrogression.

hal-00649682_v331 - 8 Dec 2011

442 Although the highly dislocated flysch formation can be considered as a relatively soft
443 rock, the shaded relief analysis with the differential DEM clearly demonstrates that, at
444 small scale, the key factor controlling the failure geometry and the overall stability of the
445 mass is not the flysch formation itself, but the spacing and the orientation of the
446 discontinuities composed by D1, D2 and D3. Consequently, the upper part of the La
447 Valette landslide can be divided in four terrain units (Fig. 8a).

448 Since the triggering date in March 1982, the terrain unit 1 has been confined between the
449 steep discontinuities D2 and D3 that constrained the landslide retrogression to the North-
450 East. The failed mass is composed of coherent blocks (up to 50 m wide) which are sliding
451 towards the main slope direction. These blocks form minor counter-slopes affected by
452 multiple open tension cracks (up to 1 m in opening and in spacing) favoring water
453 infiltration. The blocks are progressively dislocated and incorporated in the mudslide
454 body downhill. The negative elevation difference developing along D1 indicates that the
455 landslide retrogression to the North-East is limited by D1 which forces the retrogression
456 to develop laterally to the North (Fig. 8b). As a consequence of the loss of buttress given
457 by the terrain unit 1, the terrain units 2 and 3 are progressively destabilized. The terrain
458 unit 4 (North-West side of D3) is characterized with a hummocky morphology indicating
459 a lower destabilization of the slope due to the loss of buttress provided by the terrain
460 unit 3 located downhill.

461 At a smaller scale, the La Valette landslide appears to be included in a Deep-Seated
462 Gravitational Slope Deformation (DSGSD) (Agliardi et al., 2001) which main scarp
463 coincides with the extension of D1 to the North. In addition, the Soleil Boeuf crest
464 presents the same morphology as before the triggering in 1982 when the scarp was
465 located 200 m more to the South-West (Le Mignon, 2004). Therefore, the 1982 large
466 failure is strongly suspected to be a reactivation of an older DSGSD along D1.
467 Furthermore, the dip and dip direction of the Western side of the DSGSD main scarp
468 ($166^{\circ}/38^{\circ} \pm 10^{\circ}/4^{\circ}$) is very similar to D3, suggesting that the old scarp face belongs to the
469 same discontinuity set. Consequently, D3 can thus be a pre-existing fracture which is also
470 reactivated. The difference in the fracturing degree on both sides of D2 clearly highlights
471 the preferential extension of the landslide towards the North-West. On the South-East
472 side of D2, no significant geomorphic evidence of current activity is observed except the
473 extension of D1 which progressively disappears in the flysch formation to the South-East.
474 Furthermore, no negative elevation difference is noticed in this area in reference to the
475 vertical accuracy of the differential DEM (~10 m).

476 The downhill limit of the negative elevation remarkably coincides with the tectonic
477 discordance of the Autapie Nappe and the uphill limit corresponds to the spring area. The
478 location of the triggering area of 1982 is clearly identified where the elevation difference
479 (50 m) is the maximum.

480 **Seismic tomography analysis**

481 The tomographies of P and S-waves reveal a low velocity zone ($V_p < 900 \text{ m.s}^{-1}$,
482 $V_s < 400 \text{ m.s}^{-1}$) in the terrain unit 1 affected by dense fracturing. The increase of P and S-
483 waves velocity when one gets closer to the stable rock (fracture D1) gives a good
484 confidence in the determination of the contact among the failed mass and the stable
485 bedrock in depth (Fig. 9a, 9b). The velocities in the stable part composed of the flysch
486 formation vary in the range 1200-2000 m.s^{-1} for the P-waves and 450-600 m.s^{-1} for the S-
487 waves. These velocity values are slightly lower than those expected for similar rocks
488 composed of conglomerate and sandstone with a low clay content (Gosar et al., 2001).
489 They are explained by the highly dislocated texture of the flysch (BRGM 1974). Despite
490 an important velocity contrast in V_s on both sides of D2, the extension of D2 in depth
491 cannot be precisely determined on the tomographies because of insufficient velocity
492 contrast in depth. Starting from the stable part, the iso-values of 1200 m.s^{-1} (V_p) and
493 450 m.s^{-1} (V_s) dip rapidly below the ground surface. This interface can be followed in all
494 seismic tomographies between 12 and 15 m below the ground surface. It is difficult to
495 determine the roughness of this interface knowing that the seismic tomographies vary

hal-00649682, version 1 - 8 Dec 2011

496 often tend to produce artificial undulating surfaces because of the diffractions caused by
497 small irregularities at the shallow surface of very heterogeneous media.
498 Therefore, the interface can be modeled as a plane which is an acceptable assumption for
499 a small-scale interpretation. The plane is adjusted in a least square sense on the
500 tomography nodes at the transition between the velocities characterizing respectively the
501 failed mass and the stable bedrock. A stronger weight is attributed to the V_p
502 tomographies because the velocity contrasts are more important than in the V_s
503 tomographies. The accuracy is estimated by adjusting different planes based on five
504 different interpretations. The average orientation is thus characterized with a dip direction
505 of $233^\circ \pm 8^\circ$ and a dip of $26^\circ \pm 4^\circ$ which is close to the average slope of the ground surface
506 but lower than D1. Therefore this interface is not interpreted as the extension of D1 in
507 depth, but as the limit of D4 between a highly fractured media and the less fractured
508 bedrock in flysch. D4 could possibly correspond to an internal shear zone upon which a
509 highly fractured rock is sliding in a dip-slope configuration.

510 Kinematics analysis

511 Displacements calculated from the TLS datasets between the period May 2008–May 2010
512 illustrate the landslide activity at the vicinity of D1, D2 and D3. The SDv computations
513 on the point clouds indicate a maximum absolute elevation difference of 6.14 m between
514 July 2008 and May 2010 (average error of 3.10^{-3} m with a standard deviation of 0.03 m)
515 leading to an maximum vertical displacement rate of $3.07 \text{ m}\cdot\text{year}^{-1}$ along D1 and D2 of
516 the top of the terrain unit 1 (Fig. 10). All the displacements in the terrain unit 1 are
517 concentrated between D1 and D2 where tensions cracks are developing, thus explaining
518 the low seismic velocities observed in this area (Fig. 9).
519 The elevation differences allow one to distinguish three coherent blocks 1a, 1b and 1c
520 belonging in the terrain unit 1 (Fig. 8a); these blocks are progressively separated by the
521 opening of tension cracks and the sliding along D1 and D2. Uphill, the terrain unit 2 is
522 also destabilized due to the loss of buttress provided by terrain unit 1. No displacement is
523 detected on the South-East side of D2 with reference to the accuracy of the TLS datasets
524 (less than 0.05 m). These observations are in agreement with the morpho-structural
525 analysis described previously. The displacement amplitudes of the SPs are far larger than
526 the accuracy of the TLS datasets, thus giving a good confidence in the measurements.
527 The SPs displacements allow one to determine the true 3D displacement vectors
528 characterized by an average velocity of about $4.12 \text{ m}\cdot\text{year}^{-1}$ (Fig. 11).
529 The increase of the standard deviation of SPs displacements with the elapsed days
530 between TLS acquisitions (Table 3) highlight the spatial heterogeneity of displacement in
531 the top of the terrain unit 1 due to the progressive opening of tension cracks separating
532 the blocks 1a, 1b and 1c (Fig. 10). On the opposite, the directions of the displacement
533 vectors are constant in time and in space with an average dip direction and dip of
534 $228^\circ/34^\circ (\pm 4^\circ/\pm 2^\circ)$ (Fig. 12). Consequently, the short term kinematics of the terrain unit 1
535 seems to be mainly controlled by planar failures along D1 and D4.

536 Concept for the failure mechanism

537 Because the upper part of the La Valette landslide is structurally controlled at small scale
538 by planar sliding and wedge fracture configurations, a synthesis of the structural and
539 kinematics analysis is done by the use of horizontal hemispherical projections (equal
540 angle) (Richards et al., 1978; Hoek and Bray, 2004) (Fig. 12). The fracture sets identified
541 in the terrain units 1, 2 and 4 are summarized in Table 4. Figure 13a presents the
542 conceptual model of the failure mechanism interpreted from the integration of the ALS,
543 TLS and seismic survey.
544 From a kinematical point of view, D3 and D2 define a wedge geometry with an axis
545 direction and dip of $215^\circ/30^\circ \pm 11^\circ/5^\circ$ and a maximum depth varying between 60 to 80 m
546 with a back-crack represented by D1. Because the wedge axis does not ‘daylight’ in the

hal-00649682_versjon_1 - 8 Dec 2011

547 slope face, this geometry is precluded from a strict straight-forward wedge kinematic
548 evaluation as a single homogeneous block (Hoek and Bray, 2004). In other words, the
549 wedge cannot move without a buttress breakout. Therefore the wedge geometry can only
550 constrain the landslide retrogression direction between D2 and D3 to the North. However
551 a breakout interface cannot be totally excluded at the bottom of the scarp to explain the
552 location of the source line and the subsidence along D1 in the terrain unit 1 (Fig. 13a).
553 Downhill, the mudslide body is acting as a buttress for the upper part. Consequently, the
554 progression of the mudslide allows the development of dip-slope failures and coherent
555 blocks start sliding along D4 laterally delimited by D2 and D3. Because the dip direction
556 and dip of the 2-years displacements vectors in the terrain unit 1 are located between the
557 great circles of D1 and D4 (Fig. 12), the displacement vector components are likely
558 related to a sliding along D1 and D4, thus leading to a bi-planar superimposed failure
559 mechanism. In addition, the dip direction and dip of the displacements vectors are
560 remarkably close to the intersection of the great circles of D2 and D3 which corresponds
561 to the wedge axis orientation (Fig. 12). This observation consolidates the hypothesis of
562 the predominant role of the wedge geometry on the long-term development of the
563 landslide since its triggering date while the planar sliding explain the short-term landslide
564 kinematics. D4 is probably not a pre-existing discontinuity and is likely related to a
565 plastic deformation of the highly fractured flysch formation during the development of
566 the dip-slope failure mechanism. The tension cracks observed at the ground surface are
567 certainly connected to the shear zone D4 in depth, thus leading to a listric geometry
568 which is usually observed in soft rocks and clay slopes.

569 **Volume estimation with the SLBL method**

570 To estimate the volume of the failed mass along D4 in the terrain unit 1, an interpolation
571 is carried out using the SLBL method. The discontinuities D1 (North-East limit), D2
572 (South-East limit), D3 (North-West limit) and D4 (basal limit in depth) are used to
573 constrain the calculation domain of the SLBL by assuming that D4 is continuous with a
574 slightly curved geometry. The mechanical weak zone highlighted by the spring line at the
575 lower limit of terrain unit 1 (Fig. 8; Fig. 13a) is used as the Southern limit for the SLBL
576 calculation. The unstable volume is calculated using a 2 m grid DEM interpolated from
577 the ALS data. This cell size is essentially used for computation stability and time
578 computing purposes. A tolerance of -0.3 defining the degree of curvature of the SLBL is
579 selected in such way that the SLBL surface fits at best the discontinuity D4.
580 The result gives a slightly curved surface that flattens and daylight in the spring line A
581 volume of 500'000 m³ is estimated (Fig. 13b). This volume represents the highly
582 fractured mass mobilized by D4 which is currently loading the underlying mudslide body.

583 **Discussion and Conclusion**

584 In the La Valette landslide, the morpho-structural analysis and the displacement analysis
585 indicated a structurally-controlled evolution of the landslide at small scale. The
586 retrogression failure observed in the upper part of the La Valette landslide is an
587 intermediate case study between landslide developed in soft and in strong rocks. In strong
588 rocks, the relationship between pre-existing fractures and the failure mechanism has been
589 widely observed and illustrated (Sauchyn et al., 1998; Agliardi et al., 2001; Jaboyedoff et
590 al., 2009). In soft rocks, the failure mechanism is propagated through intact materials of
591 uniform shear strength (Irfan, 1998). Therefore conventional stability analyses are not
592 fully appropriate because of the influence of both the mechanics of the discontinuities and
593 the mechanics of the deforming soil. Although this aspect is out of the scope of this study,
594 it will need specific attention for a further slope failure modeling. In addition, in such
595 slow-moving complex landslides in soft rocks, it is often difficult to determine the exact
596 location of the failure surface. The deformation may occur at more than one level or in

hal-00649682_v33-2011

597 different parts of the slope at different times (Cornforth, 2005). Therefore several
598 additional failure surfaces not depicted in the conceptual model may be present in reality
599 or may occur in the future. However, the proposed model is coherent with the ALS, TLS
600 and seismic data and the field observations.

601 The upper part of the La Valette landslide is very probably a case of reactivation of an
602 older landslide and is characterized with a succession of individual slides as it is often
603 observed in large Deep-Seated Gravitational Slope Deformation (DSGSD) (Agliardi et
604 al., 2001). A first slide occurs and, as a consequence of the movement of this mass, other
605 adjacent masses starts to move thus allowing the retrogression of the landslide to the
606 North. The opening of tension cracks in the crown area is evidence of the initiation of a
607 progressive failure. However, the flysch formation may already have undergone some
608 deformation before the discontinuities start to open. The failure may have started in the
609 weak zone D4 and along the pre-existing fractures D1, D2 and D3. Furthermore, the pre-
610 failure behavior of successive slides is probably a consequence of non-uniform stress and
611 strain conditions which prevent the upper part to a catastrophic failure. In the same way,
612 the dip of D4 extended by the SLBL calculation (internal sliding surface mobilizing
613 500'000 m³ of the landslide mass) is very close to the residual friction angle of the
614 reworked flysch (30°) (Colat and Locat, 1993; Le Mignon, 2004; Fig. 13b). Therefore the
615 stability limit is not reached simultaneously in the whole mass and it would be very
616 unlikely that the unstable mass will fail in a single event. Because the wedge axis
617 delimited by D2 and D3 does not 'daylight' in the slope face, the probability of a
618 catastrophic failure is considerably reduced to a progressive release of material through
619 shallow translational failure mechanisms controlled by D1 and D4. However the
620 existence of breakout interfaces at the bottom of the main scarp cannot be totally
621 excluded (Fig. 13a). Stresses would then be transferred by loading to the mudslide body
622 acting as a reinforced buttress since the set-up of the drainage system in the middle part
623 of the slope and the decrease of the mudslide velocity. The stress release occurs over an
624 extended period due to the continuous sliding of the mudslide body (~1 m.year⁻¹).
625 However, this buttress can become over-stressed if the activity of the upper part
626 increases, thus potentially leading to a sudden acceleration of the mudslide or the
627 development of an upper internal failure in the mudslide body. If the stress increment
628 becomes excessive, the amount of deformation becomes deviated in the shallow part of
629 the slope, thus leading to an increase of the slope inclination favoring shallow failures.
630 Sudden releases of small volumes and fluidization of the mass at the vicinity of the spring
631 line have been observed in the past and more recently in Spring 2009 when about
632 3000 m³ of reworked flysch suddenly mobilized in a mudflow over a distance of
633 250 meters. A development of the mudslide in areas where the buttress is less strong
634 typically in the borders of the mudslide through lateral spreading is also possible. As a
635 consequence of the important activity of the upper part, the lower mudslide part has to
636 evolve differently than in the past according to the drainage works installed further
637 downhill. The transition between the main scarp and the mudslide body constitutes a key
638 zone controlling the overall landslide behavior because that part is susceptible to be over-
639 stressed. Groundwater conditions of the upper part still need to be assessed in detail for a
640 better understanding of the failure mechanisms. Hydro-mechanical modelling in progress
641 will also help to better understand the failure mechanism.

642 To conclude, the efficiency of combining ground-based (LiDAR TLS, seismic
643 tomography) and airborne-based (LiDAR ALS) geophysical information to characterize
644 the landslide structure is demonstrated by the agreement observed between geological
645 field observations, morpho-structural modeling and kinematics analysis. The advantage of
646 combining different methods is to propose an interpretation adapted to the scale of the
647 landslide which is not possible when only local (e.g. punctual) measurements are used.
648 TLS data provide high resolution point clouds of the topography for large scale analysis
649 which complements ALS data which are more suitable for smaller scale analysis. The
650 seismic tomography survey provides spatially-distributed information on the geometry of
651 the fractures in depth. The integration and the interpretation of this multi-source data
652 allow one to propose possible landslide evolution scenarios. Furthermore, this study

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653 demonstrated that in soft rocks (e.g flysh formations), the morpho-structure information
654 derived from field observations and DEMs analysis does not fully reflect the internal
655 structure of the slope without complementary information on the kinematics and on the
656 internal structure provided with ground-based surveys. The proposed multi-technique
657 approach can be applied to different types of landslides. A simple plot of the
658 displacement vector in a stereonet with the observed discontinuity orientations can
659 highlight structurally controlled landslides. This approach has been illustrated in the
660 monitoring of strong rock-slope. Because multi-source data have heterogeneous qualities
661 and different spatial resolutions, a major difficulty consists in the extraction of relevant
662 information and in their integration in a coherent framework. Data georeferencing and re-
663 interpretation are among the most important steps needed to detect inconsistencies among
664 multi-source data (Caumon et al., 2009; Travelletti and Malet, 2011). A future challenge
665 to improve the proposed methodology relies on coupling of 3D Geographic Information
666 Systems (data storage and management) with 3D geometrical modeling packages
667 allowing quick re-interpretation of the conceptual model of the slope.
668

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679

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Tables

Table 1 Characteristics of the LiDAR datasets

Characteristics of the laser scanning systems	ALS	TLS
Scanner type	Riegl VQ480	Optech ILRIS-3D
Distance measurement technique	Time-of-flight	Time-of-flight
Wavelength (nm)	1500	1500
Field of view (in width and height) (°)	60	40
Laser spot diameter at 100 m (cm)	~3	~3
Frequency of measurements (Hz)	10'000	2'500
Characteristics of the field acquisitions		
Scan distance (m)	~ 300	~130
Pulse mode	Last echos	Last echos
GPS frequency (Hz)	5	Scan from a base position
Point density after vegetation filtering (pt.m ⁻²)		
Average	4.1	154.8
Standard deviation	4.3	140.6

Table 2 Average error μ and standard deviation σ of the relative and absolute co-registration of the TLS datasets

	Relative positioning error		Absolute positioning error	
	μ (m)	σ (m)	μ (m)	σ (m)
18 May 08	-	-	0.04	0.07
25 July 08	0.00	0.03	0.06	0.08
10 May 09	0.00	0.03	0.02	0.07
12 July 09	0.00	0.02	0.04	0.09
08 Oct. 09	0.00	0.03	0.00	0.07
26 Apr.10	0.00	0.02	0.01	0.07
27 May 10	0.00	0.03	0.01	0.07

Table 3 Average μ and standard deviation σ of the SPs displacements

	Elapsed days	μ (m)	σ (m)
18 May 08 - 25 July 08	68	1.69	0.07
25 July 08 - 10 May 09	289	3.3	0.34
10 May 09 - 12 July 09	63	0.52	0.05
12 July 09 - 08 Oct. 09	88	0.39	0.09
08 Oct. 09 - 26 Apr.10	200	1.93	0.17
26 Apr. 10 - 27 May 10	31	0.34	0.05

Table 4 Characteristics of the fracture sets identified in the terrain units 1, 2 and 4 of the La Valette landslide

	Number of observations	Dip direction (°)	Dip (°)
D1	52	247	42
D2	11	287	55
D3	77	166	44
D4	5	233	26

Figure captions

Fig. 1. Geomorphological setting of the La Valette landslide on the South-facing slope of the Barcelonnette basin. a) General view of the landslide to the North East. b) View of the main scarp and the crown areas; the dashed line delineates the area investigated by TLS and seismic tomographies. The displacement profiles measured by the “Restauration des Terrains de Montagne, RTM” office to monitor the retrogression of the crown are also indicated.

Fig. 2. Geological setting of the La Valette landslide. a) Extract of the regional geological map at 1:50.000 with the topography before the landslide event (adapted from BRGM, 1974) and schematic cross-section detailing the hydro-geological setting of the slope before the failure of 1982 (adapted from Colat and Locat, 2003). The interval between elevation contour lines is 20 m b) Photographs of the spring line at an elevation of ca. 1880 m along a possible weak zone above the tectonic discordance near the North West boundary and the Rocher Blanc location.

Fig. 3. Development of the La Valette scarp since 1974 (before the failure) to 2009 from the analysis of aerial orthophotograph. From 1982 to 2009 a scarp regression of 200 m towards the North East is observed.

Fig. 4. Displacement monitoring of some profiles along the crown (monitoring carried out by the “Restauration des Terrains de Montagne, RTM” office). The monitoring indicates an acceleration of the displacement since 2000. The displacement profiles measured by RTM are also indicated on the orthophotograph of 2009..

Fig. 5. Photographs of the crown area presenting three major discontinuities D1, D2 and D3 and the associated stereonet (equal angle, lower hemisphere). Striations are observed on D1. The location of the terrestrial laser scanning (yellow square) and the seismic profiles (L1, L2 and L3) are also indicated.

Fig. 6. Location of the eleven SPs in the crown area used to calculate the displacements. The SPs of the first acquisition are aligned on their corresponding displaced SPs for each acquisition date. The displacement of the center point of each SP of the first acquisition is used to determine the displacement vector. The average misfit between the SP of the first acquisition and the corresponding SP is estimated at ca. 0.01 m.

Fig. 7. Example of seismic tomography data from the profile L2 (see location in Figure 5). a) Example of recordings of the P-waves and Surface-waves arrival time. The first arrivals of the P-waves are used for the inversion of the P-wave velocity tomography. A spectral analysis of the surface waves (SASW) is realized to measure the surface wave dispersion curve and invert the corresponding shear wave velocity b) Example of the analysis of the dispersive character of surface waves for the 1D inversion of a vertical profile of shear wave velocity. The phase velocity represents the velocity of individual wave propagation in the media according to its wavelength and frequency (Park et al. 1998). The relative good determination of the dispersion curve (dashed line) gives confidence on the inverted shear-wave velocity profile.

Fig. 8. Morpho-structural maps derived from the interpretation of an Airbone Laser Scanning (ALS) survey. a) Major discontinuities and sub-units identified in the scarp and in the crown areas. b) Differential DEM highlighting the retrogression direction of the landslide for the period 1960-2009 constrained by the discontinuity D2 and D3.

Fig. 9. 3D view of the seismic tomographies with their location in the crown area a) P-wave velocity tomography. b) S-wave velocity tomography. A velocity contrast at about 12 m deep highlights the presence of the discontinuity D4 interpreted as an internal sliding surface.

Fig. 10. TLS point cloud comparisons according to the reference date of 18 May 2008. The displacements are calculated with the shortest distance comparison in the vertical direction. A negative value means that the point elevation is lower than the point of the reference. Several blocks are clearly individualized through time. The blocks 1a, 1b and 1c belong to the sub-unit 1. The block 2 belongs to the sub-unit 2. The displacement vectors of the SPs are also indicated.

Fig. 11. Cumulated displacements of the SPs calculated with the roto-translation technique.

Fig. 12. Stereonet (equal angle, lower hemisphere) of the major discontinuities observed in the scarp and in the crown areas. The direction of displacement located between the great circles of D4 and D1 near the intersection of D2 and D3 strongly suggest that the upper part of the La Valette landslide is structurally-controlled by planar failure confined within a wedge geometry.

Fig. 13. Conceptual geological model of development of the upper part of the La Valette landslide. a) Proposed concept of the failure mechanism affecting the scarp and the crown areas. The morpho-structural and the kinematics analyses highlight a planar failure mechanism along D1 and D4 confined in a wedge geometry delimited by D2 and D3. The location of the cross-section is shown in Fig. 13b. b) 3D view of the discontinuity D4 extended with the SLBL computation. The ground topography above D4 has been removed.

Figure 1
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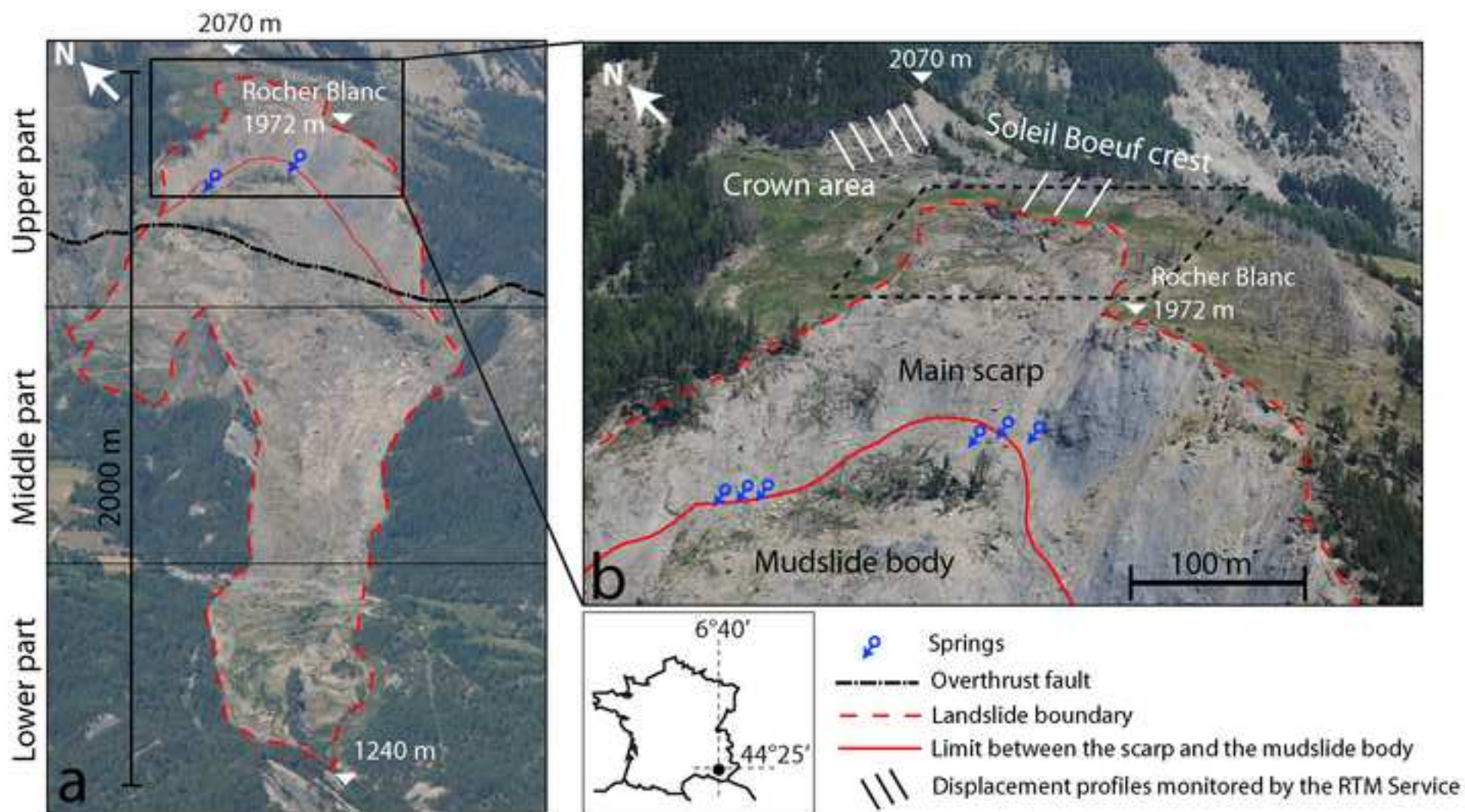


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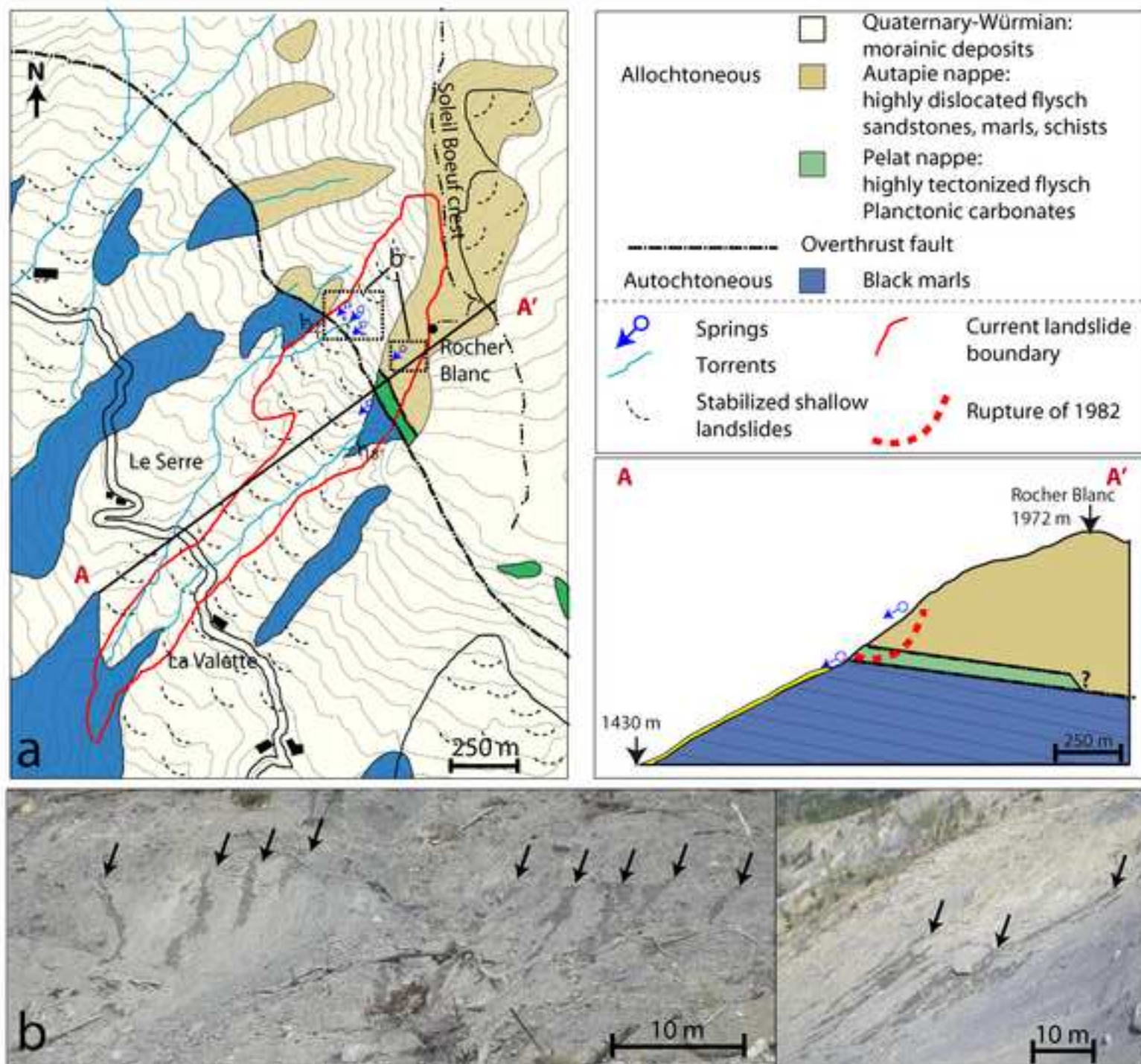


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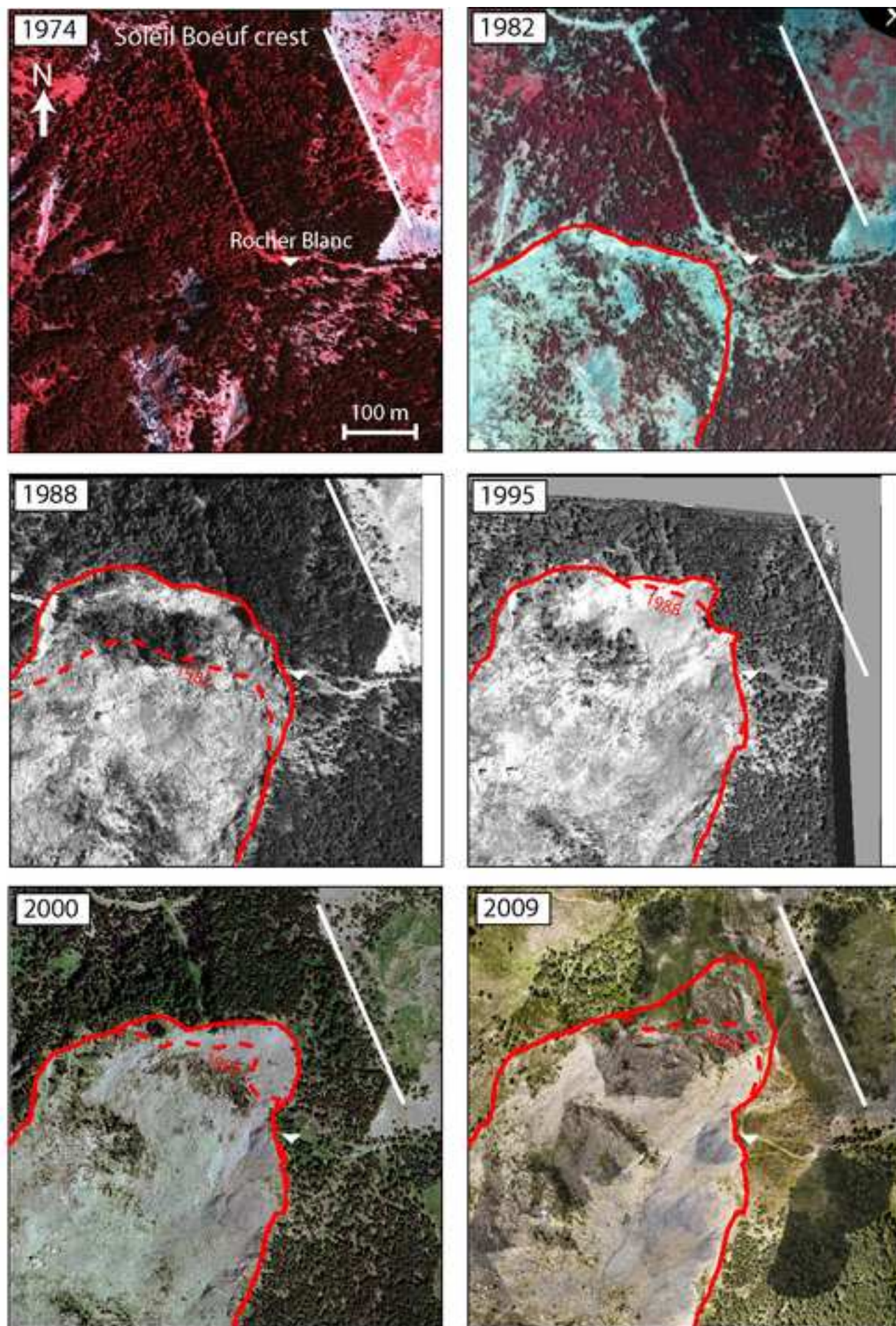


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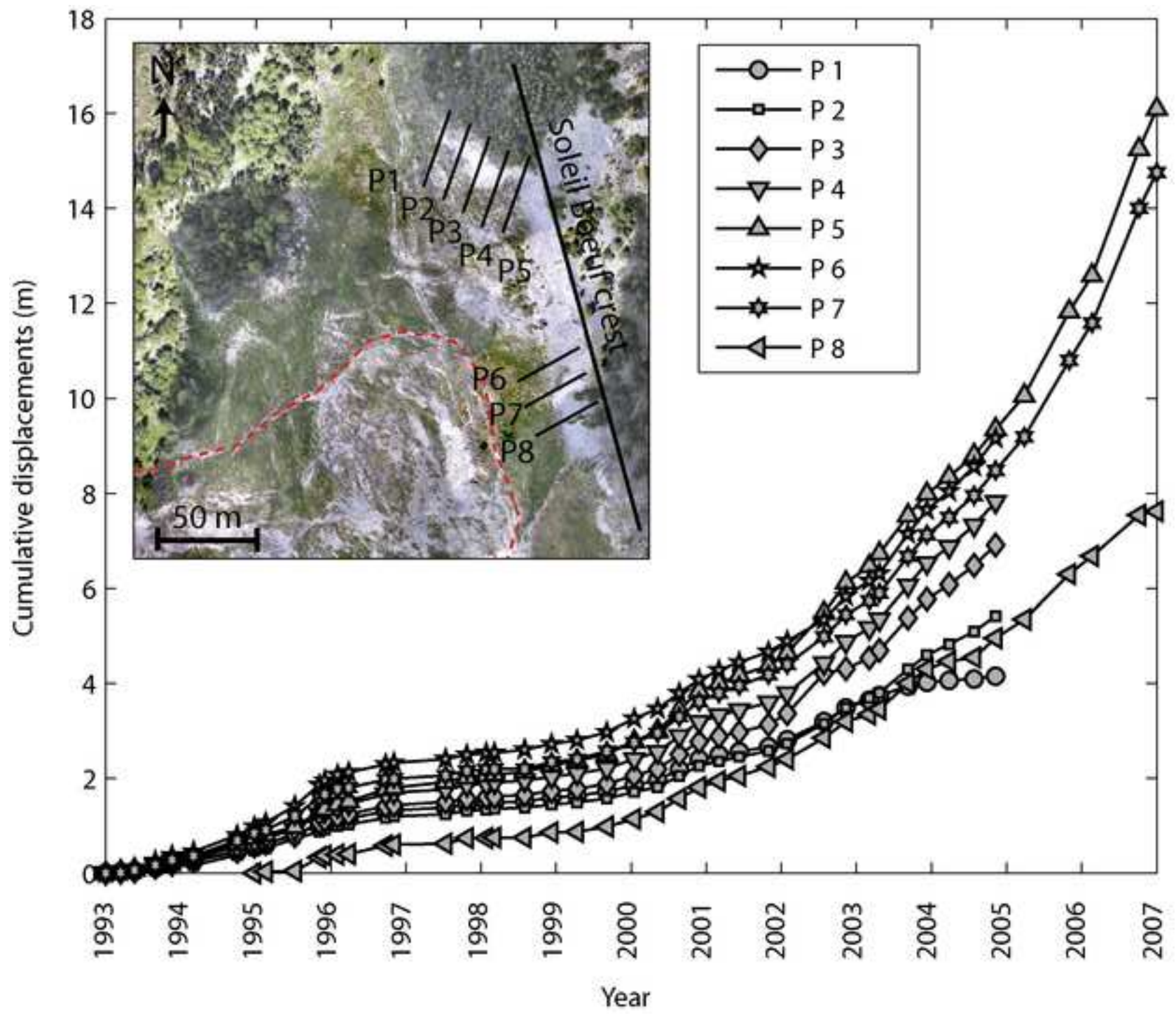


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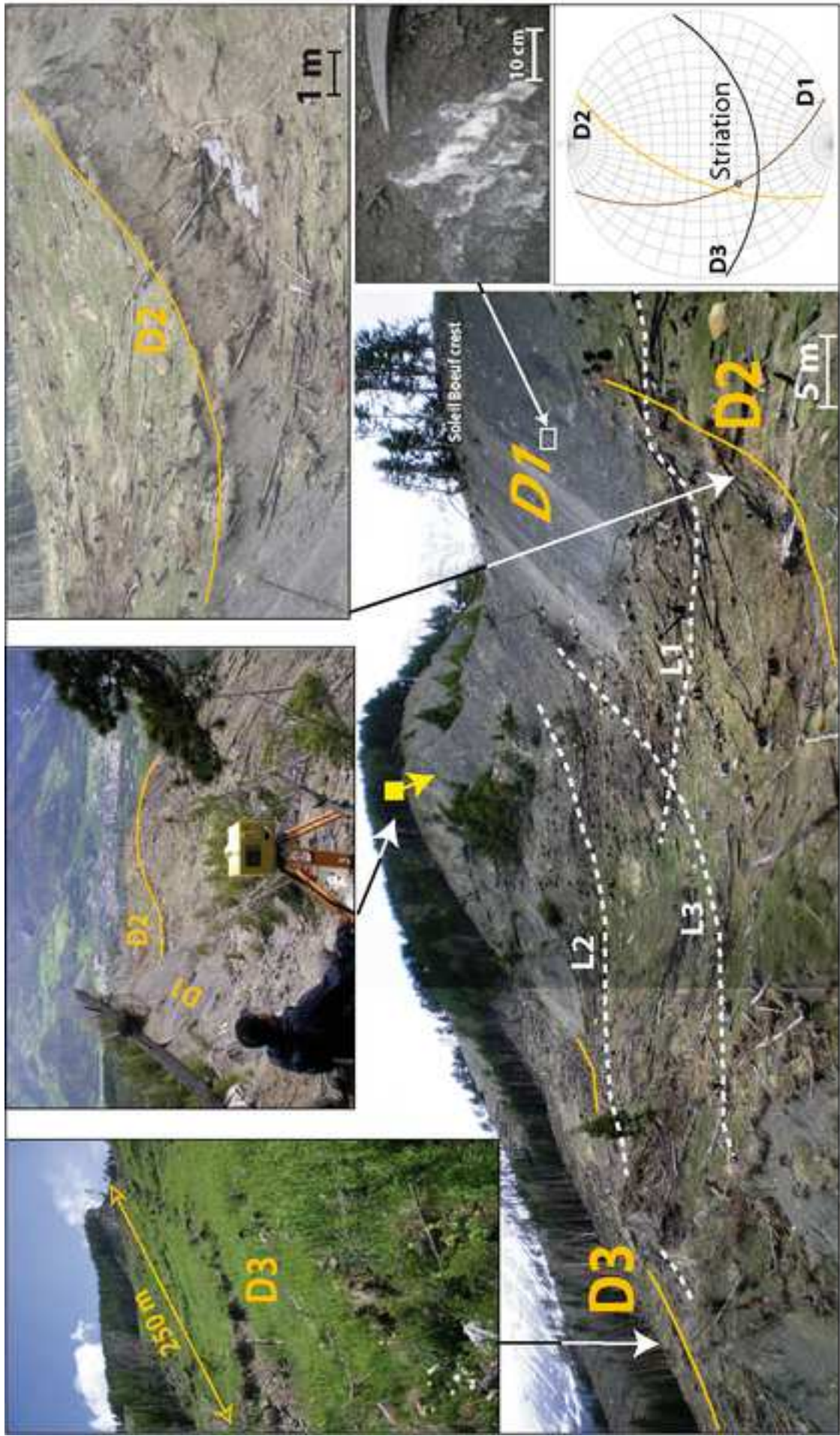


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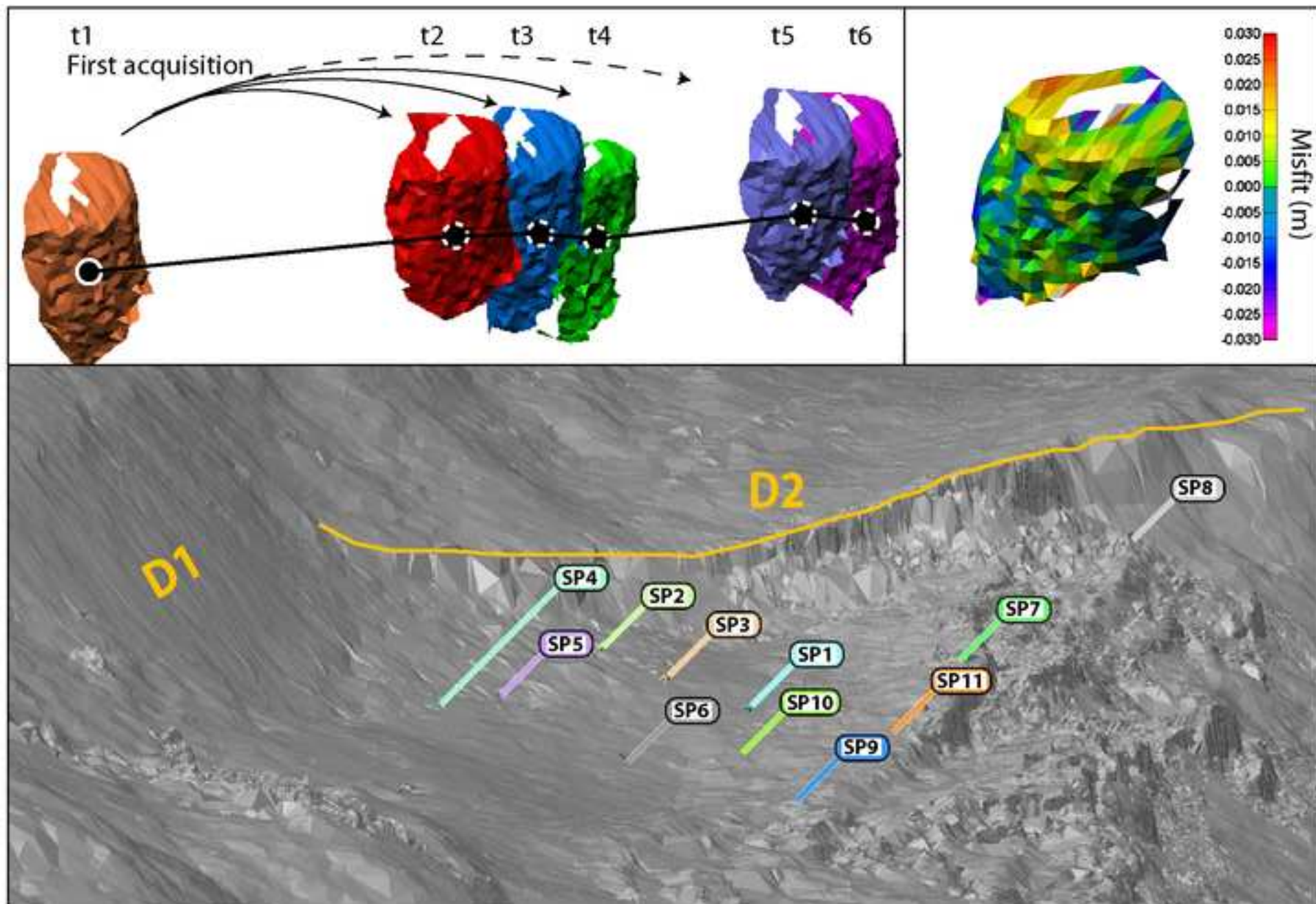


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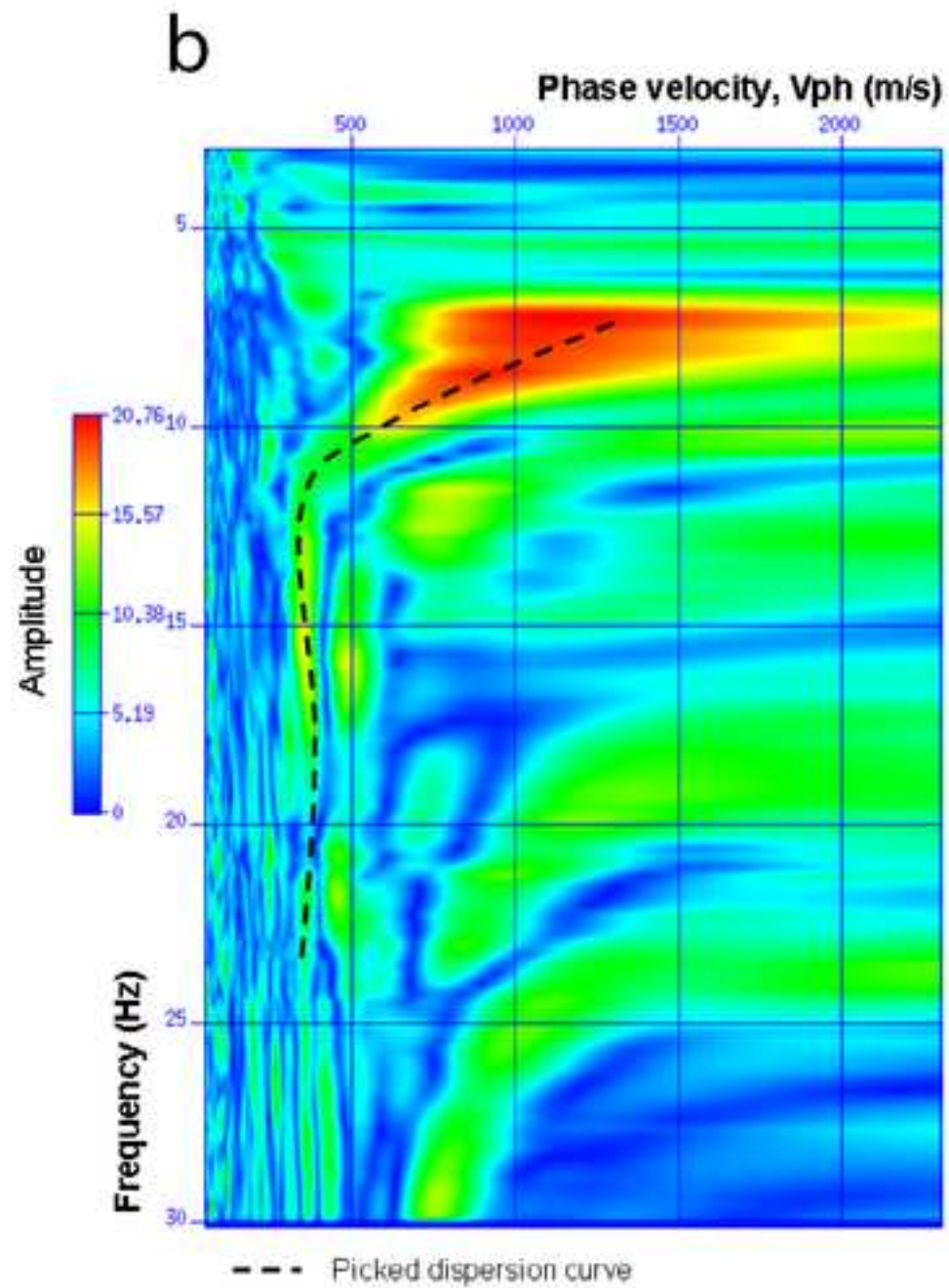
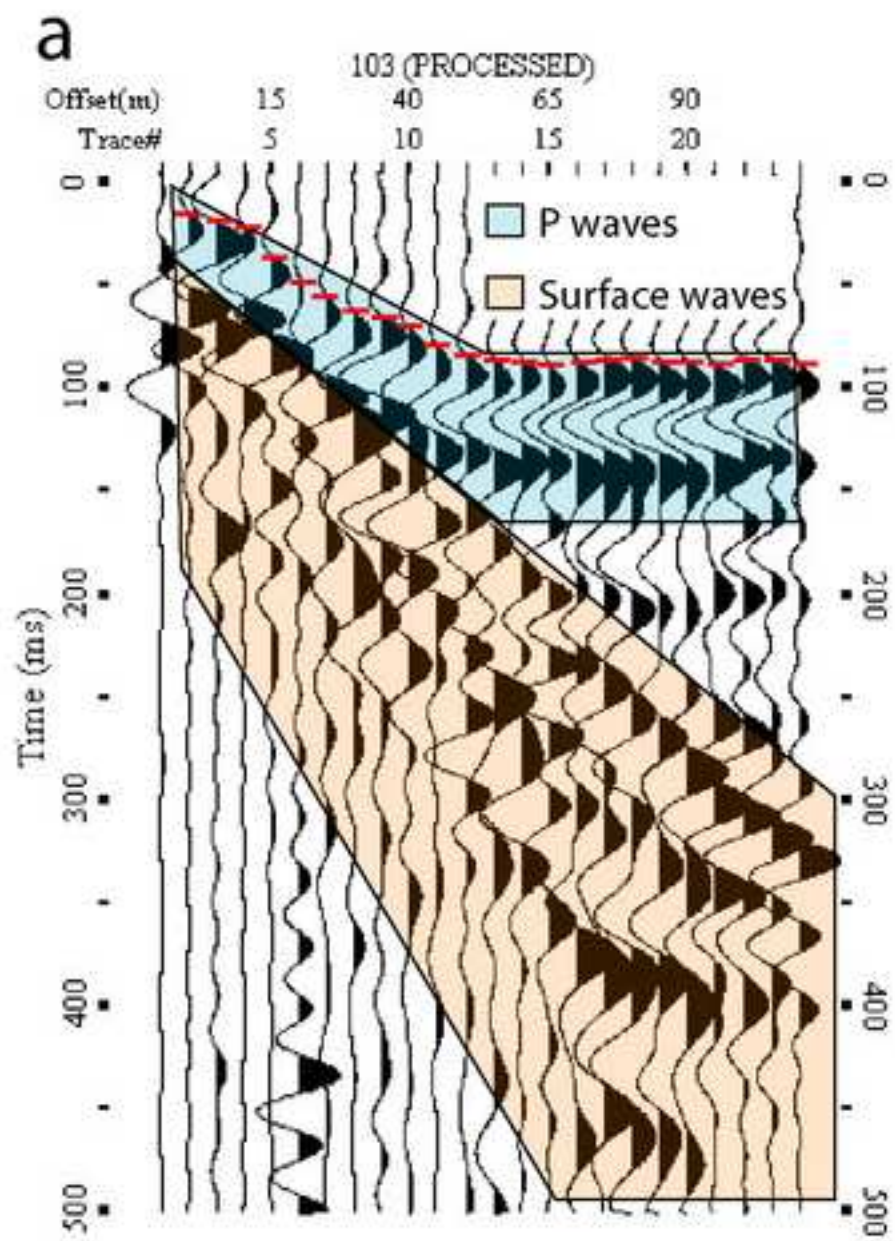
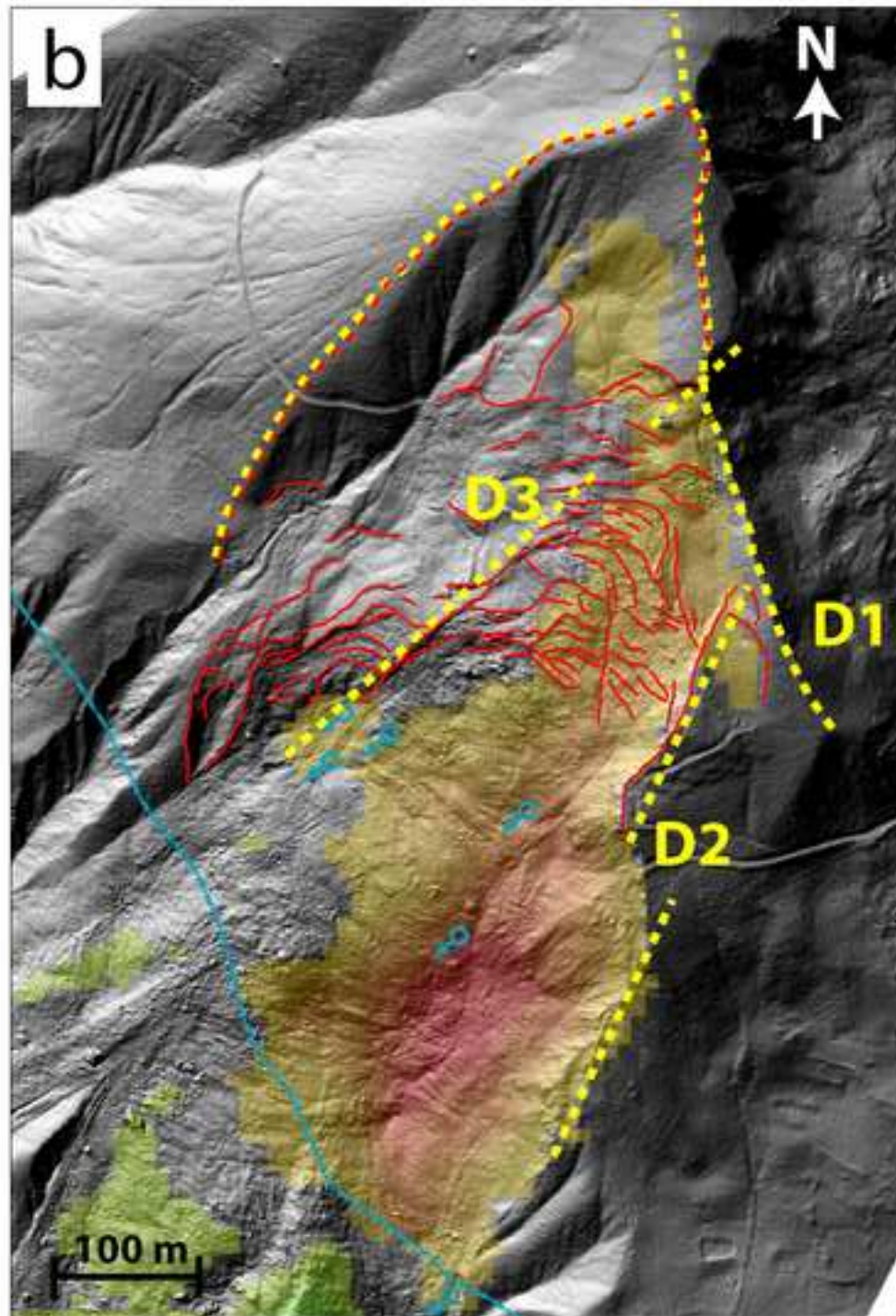
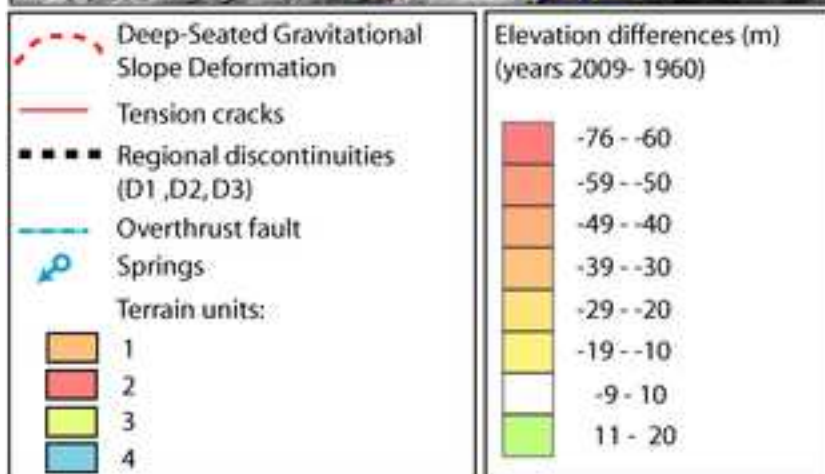
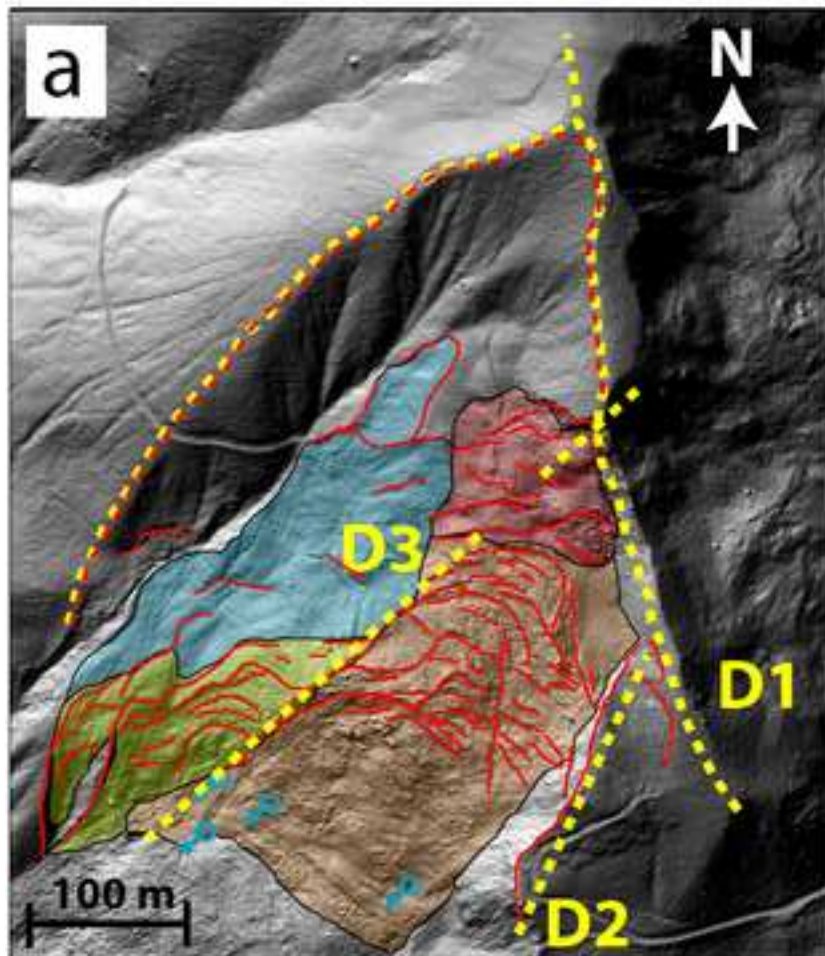


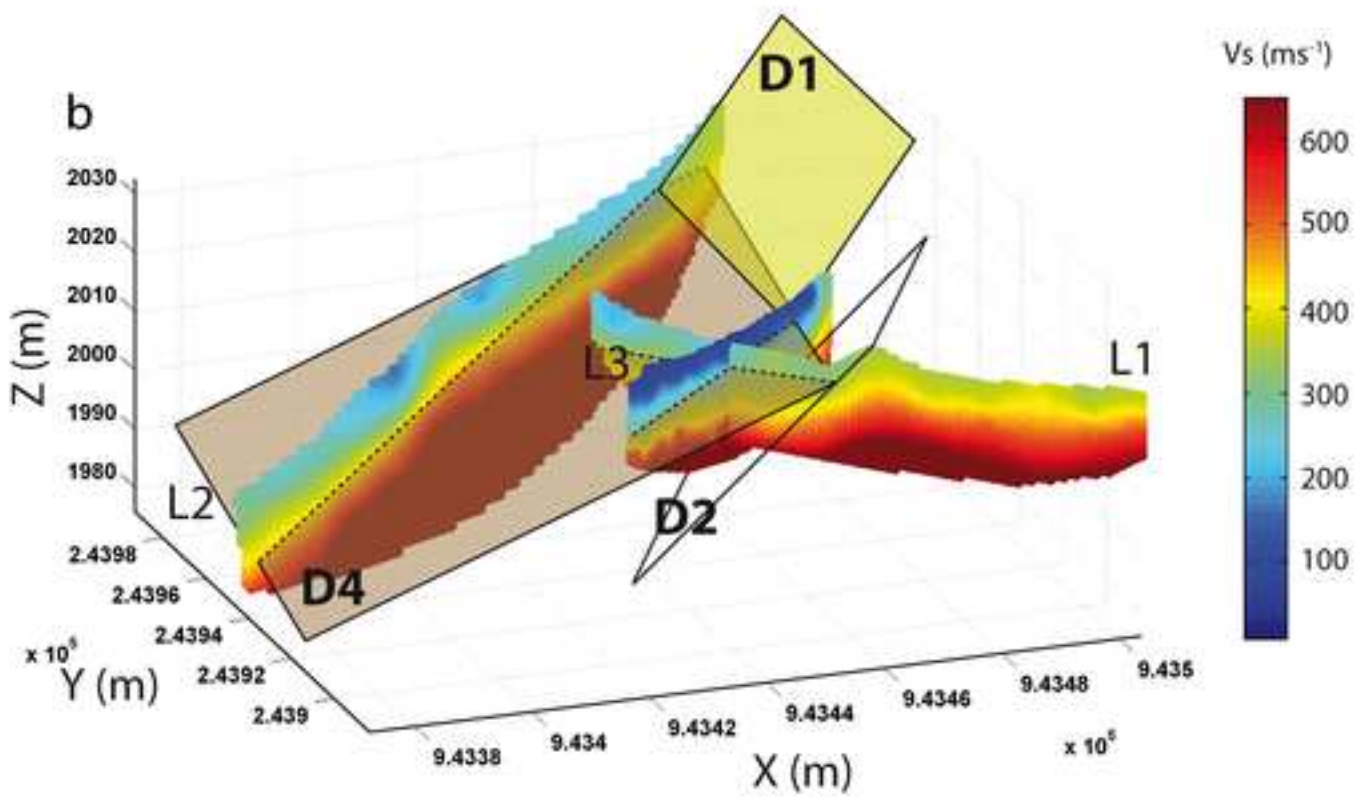
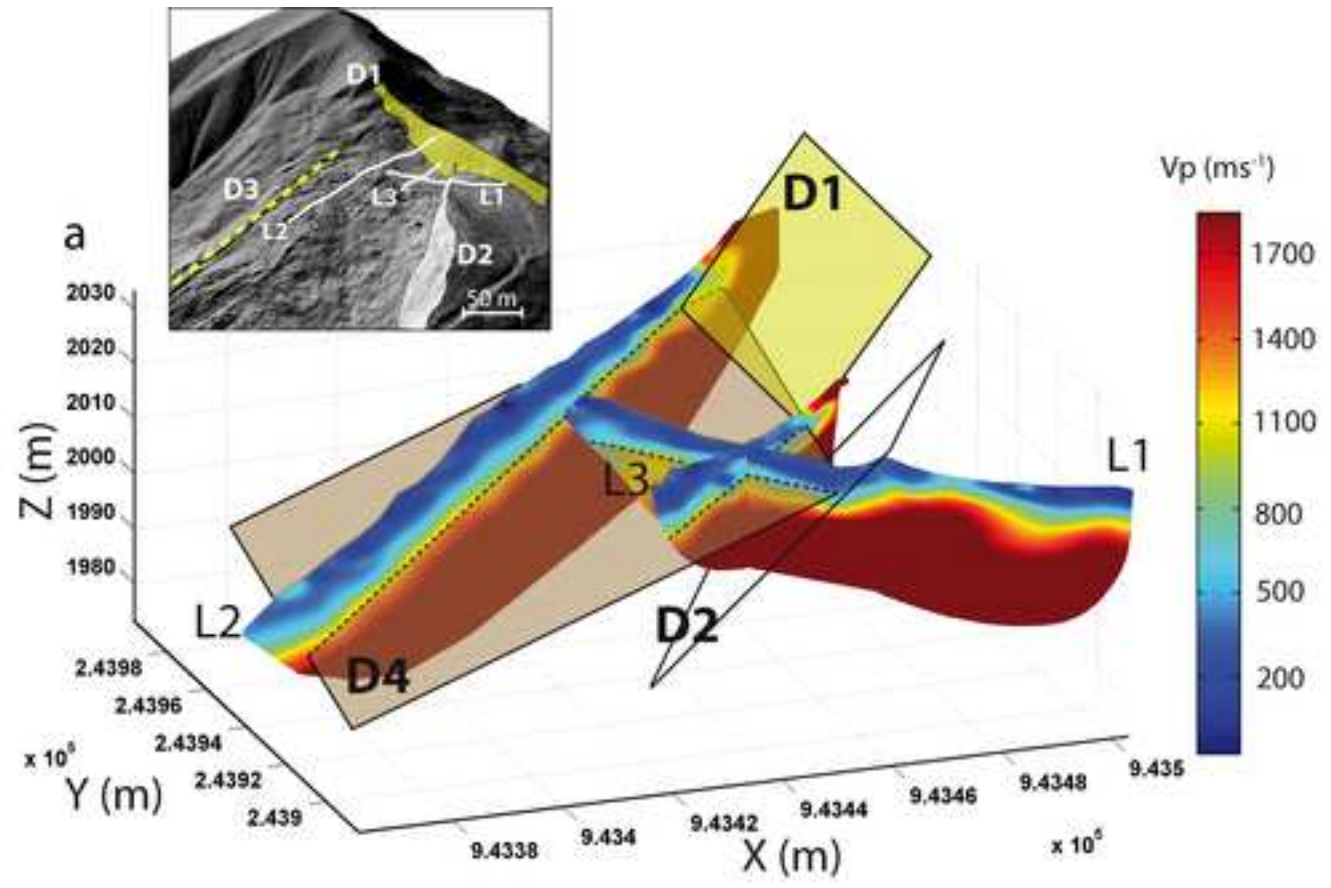
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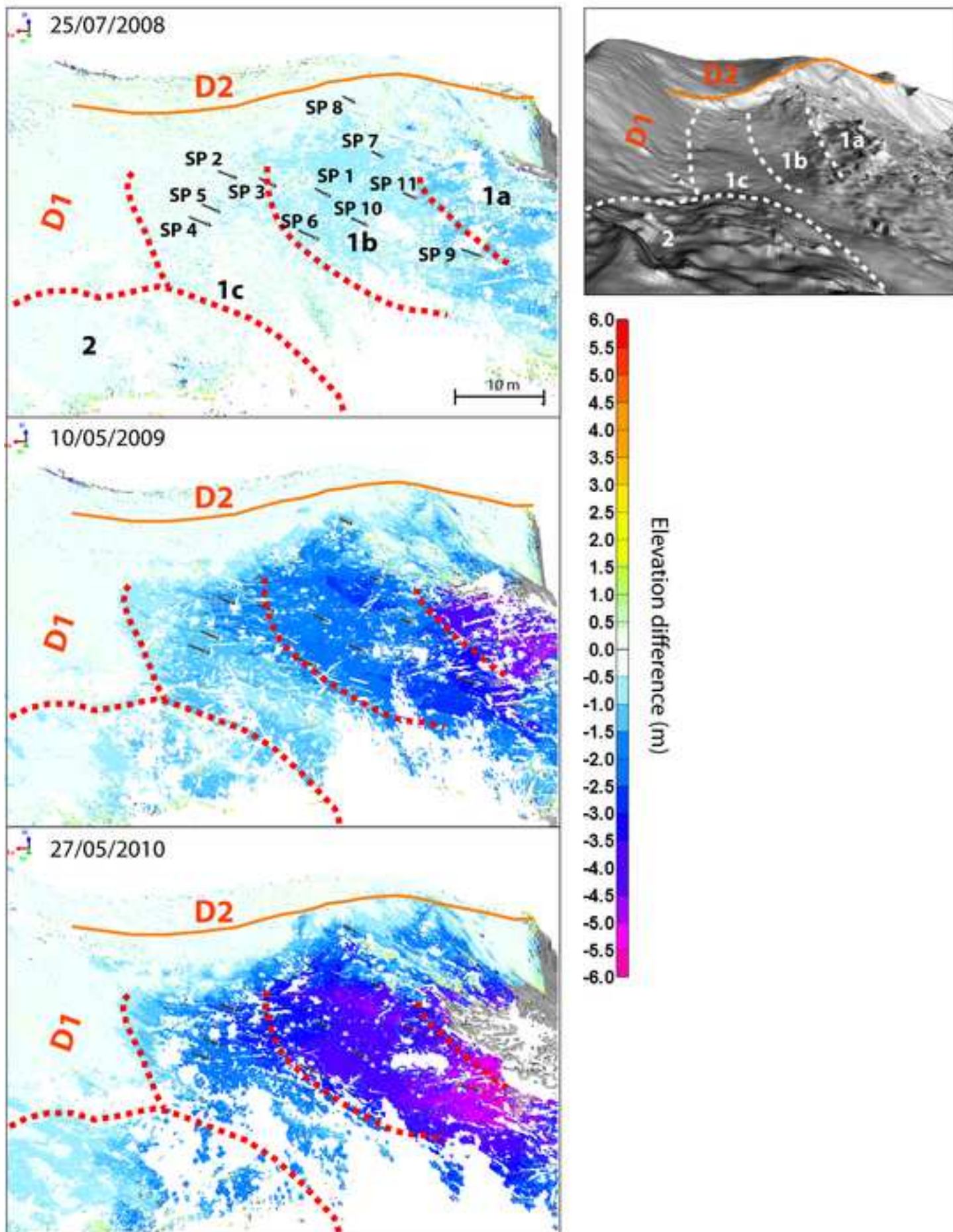
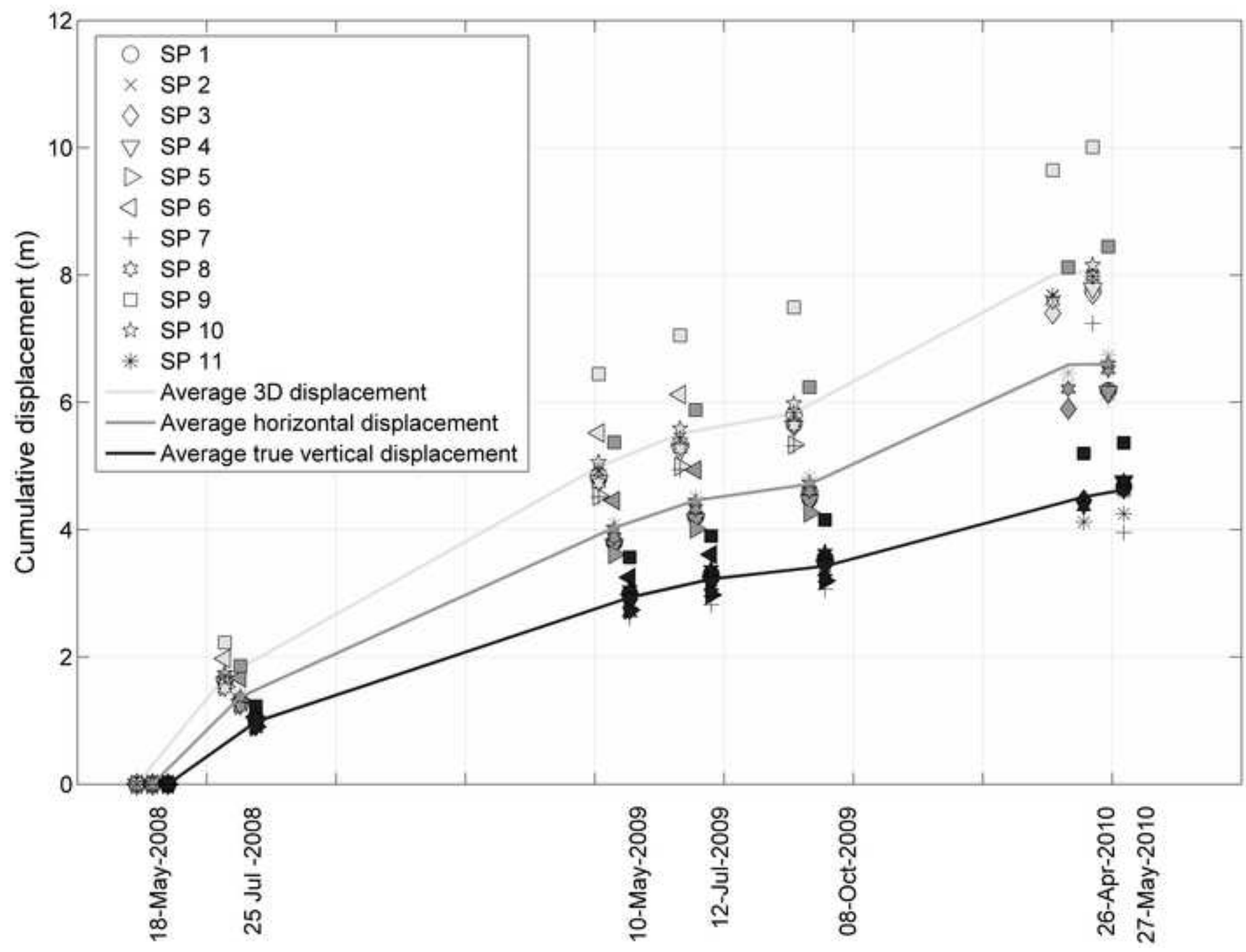
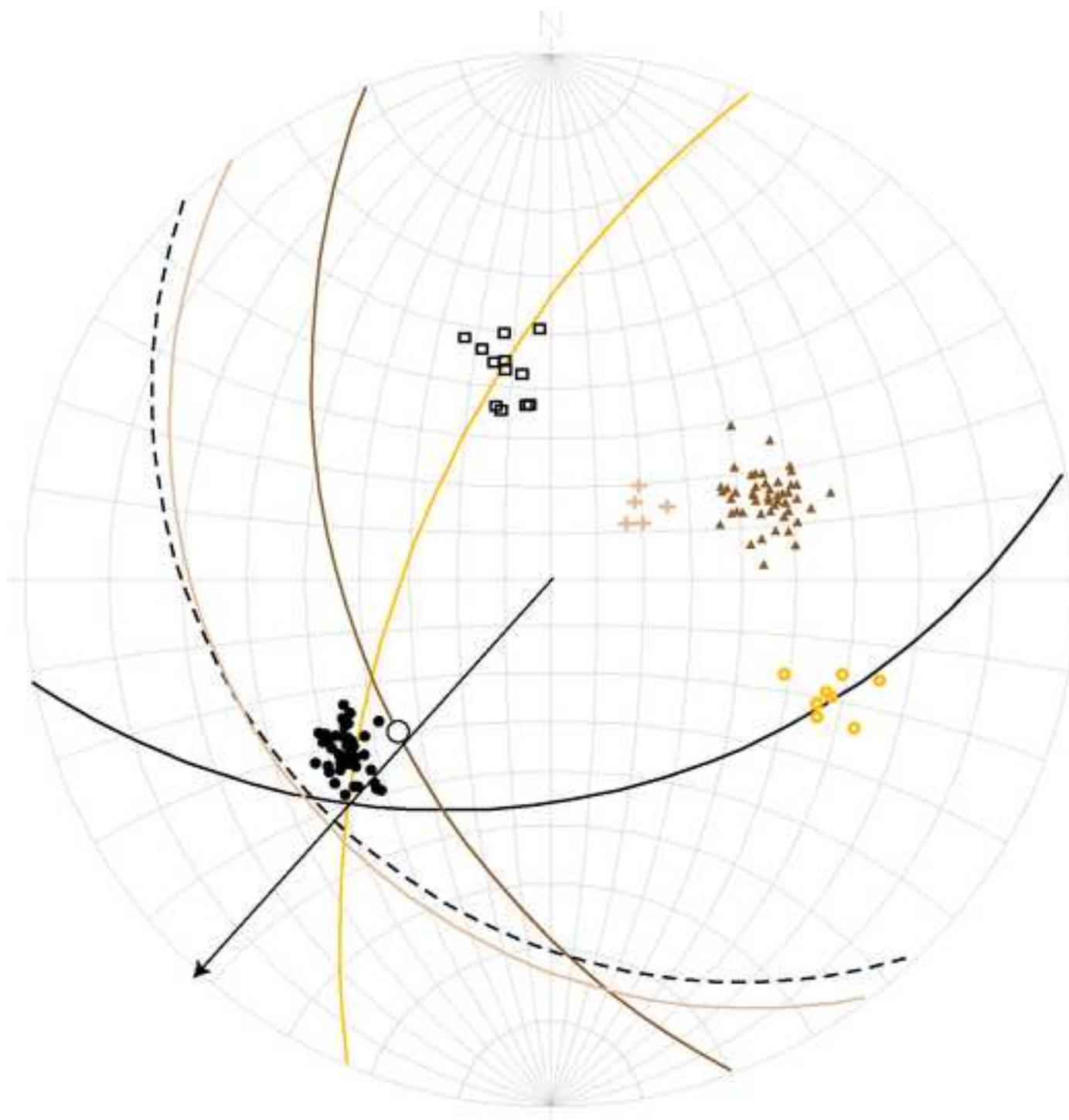


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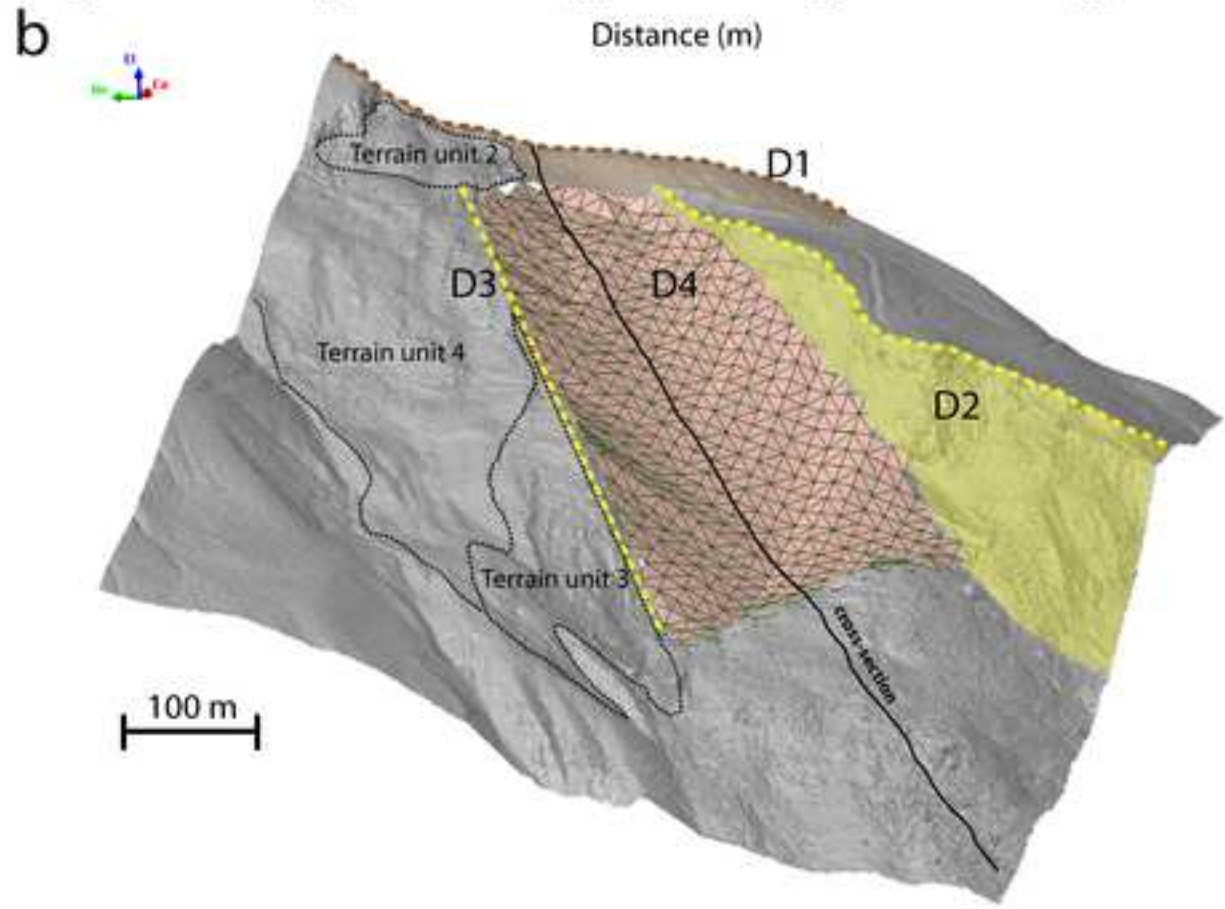
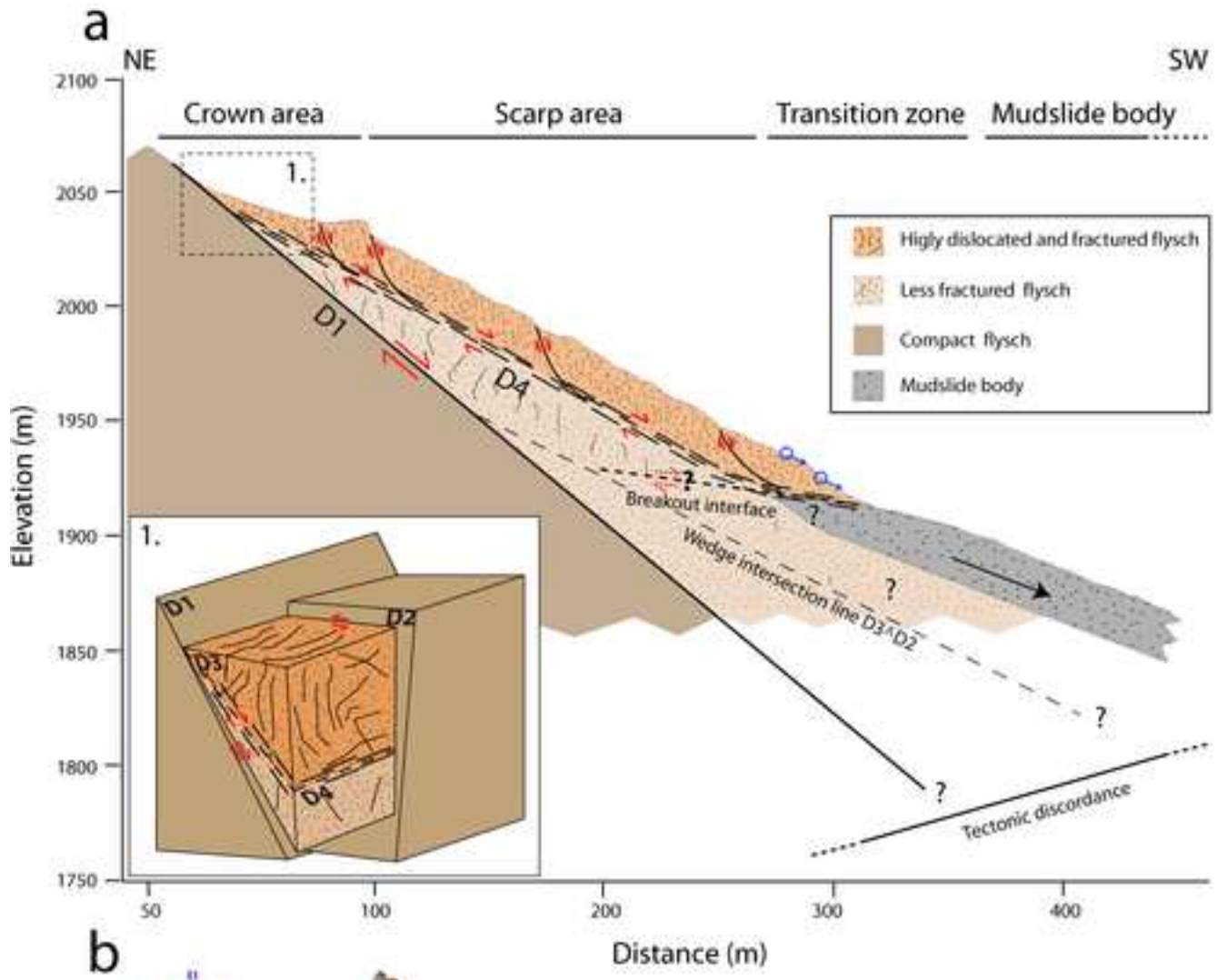
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Equal Angle
 Lower hemisphere

	Poles	Average great circles
D1	▲	⌒
D2	●	⌒
D3	◻	⌒
D4	+	⌒
Displacements	•	
Striation	○	
Mean terrain slope	- - -	
Wedge axis	↙	

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