

1 Landslide characterization using P- and S-
2 wave seismic refraction tomography – the
3 importance of elastic moduli

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11
12 [Abstract](#)

13 In the broad spectrum of natural hazards, landslides in particular are capable of changing the
14 landscape and causing significant human and economic losses. Detailed site investigations form an
15 important component in the landslide risk mitigation and disaster risk reduction process. These
16 investigations usually rely on surface observations, discrete sampling of the subsurface, and
17 laboratory testing to examine properties that are deemed representative of entire slopes. Often this
18 requires extensive interpolations and results in large uncertainties. To compliment and extend these
19 approaches, we present a study from an active landslide in a Lias Group clay slope, North Yorkshire,

20 UK, examining combined P- and S-wave seismic refraction tomography (SRT) as a means of providing
21 subsurface volumetric imaging of geotechnical proxies.

22 The distributions of seismic wave velocities determined from SRT at the study site indicated zones
23 with higher porosity and fissure density that are interpreted to represent the extent and depth of
24 mass movements and weathered bedrock zones. Distinguishing the lithological units was facilitated
25 by deriving the Poisson's ratio from the SRT data as saturated clay and partially saturated sandy silts
26 showed distinctively different Poisson's ratios. Shear and Young's moduli derived from the SRT data
27 revealed the weak nature of the materials in active parts of the landslide (i.e. 25 kPa and 100 kPa
28 respectively). The SRT results are consistent with intrusive (i.e. cone penetration tests), laboratory,
29 and additional geoelectrical data from this site. This study shows that SRT forms a cost-effective
30 method that can significantly reduce uncertainties in the conceptual ground model of geotechnical
31 and hydrological conditions that govern landslide dynamics.

32 [Keywords](#)

33 Seismic Refraction Tomography; Landslide characterization; Elastic Moduli; Site Investigation

34 [1. Introduction](#)

35 Landslides form one of the major natural hazards and accounted for at least 4600 fatalities per year
36 between 2004 and 2010 (Petley, 2012). In addition there is significant economic impact, by affecting
37 transport and utility infrastructure (Bird and Bommer, 2004; Dijkstra et al., 2014; Glendinning et al.,
38 2014), and due to material loss which accounted for at least 1.7 billion US\$ in the last century
39 (Lacasse and Nadim, 2009; Nadim et al., 2013; Petley, 2013). Detailed investigations of slopes, which
40 pose a risk to communities and infrastructure, are needed to reduce the uncertainty of the ground
41 models (BSI, 2015). This involves characterisation (in space and in time) of the mechanical and
42 hydrological conditions that define the stability of a slope (Leroueil, 2001). Determining the spatial
43 distribution of parameters, such as soil thickness, weathering profile, and elastic material properties

44 are crucial for landslide hazard and risk zonation (van Westen et al., 2006). Being able to provide a
45 better defined ground model will lead to the design of more appropriate intervention, improved risk
46 mitigation, and landslide disaster risk reduction strategies (Crozier and Glade, 2005; Popescu and
47 Sasahara, 2009).

48 Geotechnical investigations, such as cone penetration tests and laboratory studies, are focussed on
49 discrete points of a landslide, sampling a small volume of the material only. Landslides, due to their
50 geomorphological characteristics, are complex structures, showing high variability in their physical
51 properties (Cascini et al., 2015). Thus, geotechnical investigations, delivering “true” mechanical and
52 hydrological properties, need to be supplemented by methods that allow for a definition of their
53 spatial variability (Jongmans and Garambois, 2007; Perrone et al., 2014). Therefore, landslide studies
54 frequently comprise geophysical measurements alongside geotechnical and laboratory
55 characterization (Sass et al., 2008; Schrott and Sass, 2008; Gunn et al., 2013; Springman et al., 2013;
56 Lissak et al., 2014; Salas-Romero et al., 2015). Out of the range of available geophysical techniques,
57 electrical resistivity tomography (ERT) and seismic imaging methods are perhaps the most frequently
58 applied to landslide studies (Jongmans and Garambois, 2007; Schrott and Sass, 2008; Van Dam,
59 2012; Perrone et al., 2014).

60 In a landslide characterization context, P-wave seismic refraction tomography is most commonly
61 applied, as seismic velocities usually show significant differences between the landslide mass and
62 the underlying bedrock (Heincke et al., 2006; Donohue et al., 2012; Yamakawa et al., 2012).
63 However, for slopes that consist of similar sediments, a delineation of the different units and the
64 effect of geomorphic processes is usually not possible as seismic velocities overlap (Schrott and Sass,
65 2008). This may be overcome by employing P- and S-wave SRT, as P- and S-waves are affected
66 differently by changes in saturation, porosity, or elastic moduli (Gregory, 1976; Macari and
67 Laureano, 1996; Mondol et al., 2007; Pasquet et al., 2015). Derivation of Poisson’s ratio from a
68 combined imaging of P- and S-wave velocities has recently been successful in detecting saturation

69 characteristics of shallow aquifers (Grelle and Guadagno, 2009; Pasquet et al., 2015). However, most
70 of these studies implement a sequential acquisition of P- and S-wave refraction data or a
71 combination of refraction and surface wave methods (Grandjean et al., 2009; Hibert et al., 2012),
72 which may introduce potential pitfalls due to different source and signal signatures, and offsets in
73 the acquisition layout.

74 This study employs simultaneous P- and S-wave SRT to study the elastic properties of a shallow
75 clayey landslide. From the SRT results, distributions of shear and Young's moduli, as well as Poisson's
76 ratio are derived. As these parameters define the elastic properties of the slope material, the likely
77 modes of deformation of the landslide can be defined (i.e. whether this is characterised by plastic,
78 brittle or flow-type failure; what the likely position/shape of the main slip surface is; and, potentially,
79 how strains are expected to develop in slopes). The outcome of this study highlights the benefit of
80 deriving elastic moduli and Poisson's ratio to cost-effectively conduct a thorough investigation of the
81 mechanical and hydrological conditions defining the landslide behaviour and provides insights into
82 how spatial distributions of elastic properties can be used to reduce the uncertainty in the landslide
83 ground model and improve characterisation of the landslide behaviour.

84 1.1 Study area

85 The studied landslide is located at Hollin Hill, a south-facing hillslope with a mean slope angle of
86 about 14°. It is close to the town of Terrington, about 10 km west of Malton, North Yorkshire, UK
87 (54°06'38" N, 0°57'30" W), set in the Howardian Hills, an escarpment running approximately NW-SE.
88 It is underlain by four formations (Fig. 1c) of Lower and Middle Jurassic age comprising, in ascending
89 order, Redcar Mudstone (RMF), Staithes Sandstone and Cleveland Ironstone Formation (SSF),
90 Whitby Mudstone Formation (WMF), and Dogger Formation (DF). The DF is a limestone- and
91 sandstone-dominated unit, which caps the hill and forms a potential perched aquifer above the
92 WMF (Gunn et al., 2013). The thickness of the DF varies considerably over the region as an effect of
93 the formation occupying hollows in the underlying WMF, and reaches a local maximum of 8 m to the

94 north of the site. The WMF is composed of grey to dark-grey mudstone and siltstone, including
95 scattered bands of calcareous and sideritic concretions. It has a thickness of about 25 m and shows a
96 sharp boundary with the overlying DF. The WMF is the failing formation at site and in the
97 surrounding area. The formations of the Upper Lias Group, and the WMF in particular, are known to
98 cause slope instabilities throughout the UK, accounting for as much as 7.5 % of all UK landslides, with
99 a density of 42 landslides per 100 km² outcrop (Hobbs et al., 2012). The SSF, which underlies the
100 WMF, comprises ferruginous, micaceous siltstone with fine-grained sandstone and thin mudstone
101 partings, and has a thickness of about 20 m. It is heavily bioturbated and shows locally occurring
102 masses of siderite and pyrite (Gaunt et al., 1980). In the lower and middle part of the slope it is
103 associated with relatively well-drained mixtures of clay, silt and fine sand. The lower boundary
104 shows a gradational change to poorly-drained RMF, which comprises grey, silty, calcareous, and
105 sideritic mudstone and thin shelly limestones (Chambers et al., 2011).

106

107 The bedrock succession shows a local dip of about 5° to the north (Merritt et al., 2013). It is overlain
108 by a thin layer of head deposits, ranging from 0.2 to 1.3 m, which are characterised by gravelly,
109 sandy, silty clay with occasional organic inclusions. It is formed of locally derived material (mainly
110 from the DF), reworked by a combination of geomorphological processes, such as hillwash, slope
111 failure, and soil creep (Chambers et al., 2011; Uhlemann et al., 2016).

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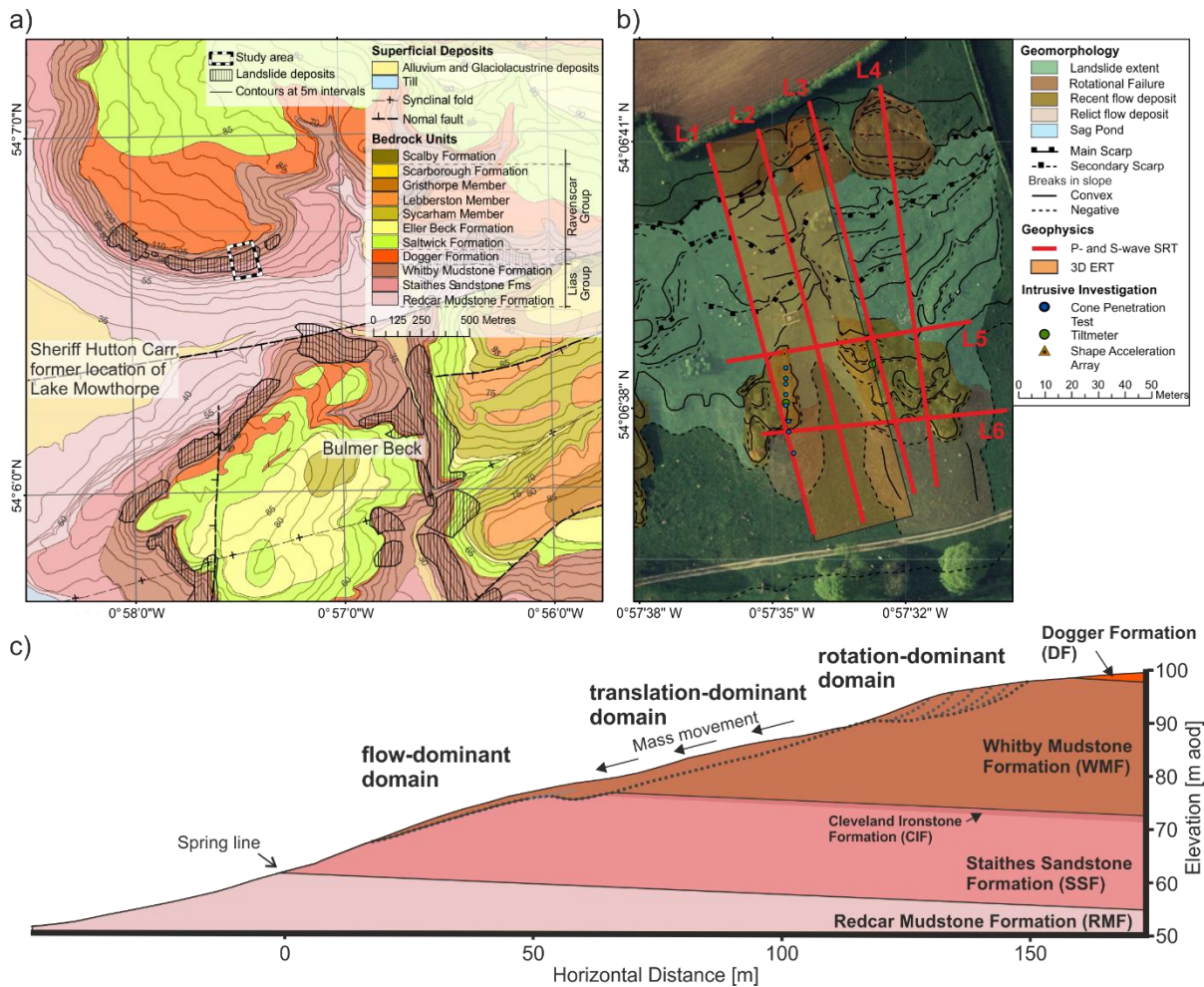
113 Using the nomenclature of Cruden and Varnes (1996), the landslide can be defined as a very slow to
114 slow moving composite multiple earth slide-earth flow, with maximum movement rates of up to 3.5
115 m/y observed in recent years (Uhlemann et al., 2016). Based on previously published data, different
116 authors have developed and continuously improved the geomorphological understanding of this
117 landslide (Chambers et al., 2011; Merritt et al., 2013; Uhlemann et al., 2016). The latest
118 understanding is that the translation-dominant domain (WMF) is the main driver for mass
119 movement processes on this slope. Substantial rainfall leads to additional loads, a rise in pore water

120 pressures and a loss of effective stress in the near-surface leading to the (re-)activation of shear
121 strains along (pre-existing) shear surfaces at critical depths of around 2 to 3 m. As material slides
122 towards the boundary between WMF and SSF it encounters a thin drape of aeolian sands overlying
123 the SSF that act as a toe drain and causes the slides to slow down and build up ridges along the
124 slope. Further phases of deformation can lead to local breakthrough and rapid acceleration of
125 flow/slide-like movement forming lobes towards the base of the slope. Thin sand lenses
126 incorporated within the slide mass can act as preferential flow-paths potentially leading to local
127 substantially elevated pore pressures (Uhlemann et al., 2016). The upper parts of the slope are
128 retrograding as shallow rotational slides, triggered by the progressive loss of support along the local
129 toe of the slopes through ongoing deformation in the translation-dominant domain. Thus, the
130 landslide complex shows translational movements towards the WMF-SSF boundary, which evolves
131 to slide/flow-like behaviour forming lobes towards to toe of the slope and drives rotational failure
132 retrograding into the upper slopes (Fig. 1c). For more general explanations on the different landslide
133 mechanisms the reader is referred to, e.g., Hungr et al. (2014).

134

135 The paleo-landscape in this area was affected by the water level dynamics of an ice-marginal lake
136 (Lake Mowthorpe) during the last glaciation in the Pleistocene. This lake was formed due to
137 landslides damming the gorge through which meltwater and surface-water runoff took place. As
138 water level in the lake rose and a spill point at the eastern edge of the lake was reached (at Bulmer
139 Beck, Fig. 1a), rapid incision occurred and this drained the lake (Chambers et al., 2011). This likely
140 caused changes in effective stresses in the slopes and potential over-steepening, causing landslides
141 that again blocked the drainage pathways and reinitiated the process. Thus, this area is
142 characterized by repeated slope movements and therefore by highly heterogeneous and poorly
143 compacted sediments, which are prone to landsliding.

144



145

146 Fig. 1 a) Geological map of the study area. Note the high landslide occurrences that are constrained to the Whitby
 147 Mudstone Formation. b) SRT line locations superimposed on geomorphological map after Merritt et al. (2013) and aerial
 148 photograph © UKP/Getmapping License No. UKP2006/01. c) Ground-model of the study site, delineating the
 149 different landslide domains (modified after Gunn et al., 2013, and Uhlemann et al., 2016).
 150

151 Hollin Hill is a well-studied landslide acting as a field laboratory to support UK landslide research. It is
 152 mainly focussed on technological developments in acoustic emission and electrical resistivity
 153 tomography, underpinning landslide monitoring and early warning (Wilkinson et al., 2010; Dixon et
 154 al., 2014; Smith et al., 2014; Smith and Dixon, 2015; Uhlemann et al., 2015; Wilkinson et al., 2016).
 155 Chambers et al. (2011) and Merritt et al. (2013) provide a thorough description of the landslide
 156 geology and geomorphology, which is mainly based on geoelectrical and borehole data, while
 157 Uhlemann et al. (2016) use long-term geotechnical monitoring data to derive an understanding of

158 the geomorphological processes and triggering mechanisms controlling the landslide movements.
159 This paper describes the result of a seismic characterization of the landslide, which can potentially
160 aid in determining the elastic properties of the landslide material and thus may provide crucial input
161 parameters for a physical modelling. It employs P-wave and S-wave seismic refraction tomography
162 (SRT) with a specific focus to determine the spatial distribution of the elastic moduli of the landslide.
163 To our knowledge, this is the first application of deriving elastic moduli from P- and S-wave SRT in a
164 landslide context, and this paper will highlight its benefits to landslide research and characterization.

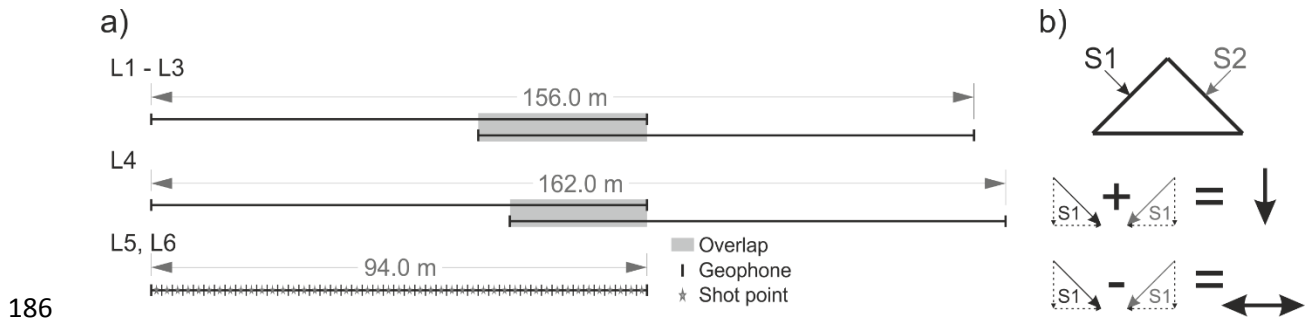
165 2. Methodology

166 2.1 Data acquisition

167 2.1.1 Survey parameters

168 The seismic survey consisted of six profiles, four of which (L1 to L4) extended from the toe to the top
169 of the slope, and two (L5 and L6) were perpendicular to these. The perpendicular profiles covered
170 the upper and lower part of two lobes (Fig. 1). While line L2 was located in a gully between two
171 lobes, lines L1, L3, and L4 covered actively moving lobes, where L3 and L4 are located along the
172 recently most active part of the landslide, showing movement rates of up to 3.5 m/year. Lines L1 to
173 L3 were located adjacent to a permanently installed 3D electrical resistivity tomography (ERT) array,
174 which also provided geoelectrical data during the SRT acquisition. Seismic data were acquired with a
175 2 m geophone and shot spacing, where shots were located between geophone locations (Fig. 2a).
176 Each spread consisted of 48 three-component geophones with a natural frequency of 4.5 Hz,
177 measuring vertical and two horizontal particle velocities. These were connected to six Geometrics®
178 Geodes with 24 channels each. As each spread spanned over 94 m L1 to L4 were measured in two
179 parts with an overlap of 32 and 26 m for L1 to L3, and L4, respectively. Each shot was recorded with
180 a 0.5 ms sampling interval and a recording length of 1.5 s. These parameters were chosen based on
181 test shots at site, which revealed very slow velocities that required long recording lengths. A 4.5 kg
182 sledgehammer hitting a steel prism was used as seismic source. The prism was oriented

183 perpendicular to the spread; for each side of the prism three recordings were acquired. The data
 184 acquisition of all six lines took 5 days and comprised a total number of 3156 shots. Each shot and
 185 geophone location was surveyed using RTK-GPS equipment.



187 *Fig. 2 a) Data acquisition layout. Note that the number of available channels and chosen geophone spacing of 2 m limited*
 188 *the maximum line length to 94 m. Thus, L1 to L4 comprise two spreads with 32 m and 26 m overlap for lines L1 to L3, and*
 189 *L4, respectively. The shot distribution applied on each line is shown on L5 and L6; shots were located with 2 m spacing*
 190 *between geophone locations. b) Source characteristics. A steel prism was hit from its two sides. Adding the two shots results*
 191 *in the vertical component of the wave field, while subtracting results in the horizontal component of the wave field.*

2.1.2 Wave component extraction

193 By using a steel prism as seismic source P- and S-waves were excited at the same time. This reduced
 194 the acquisition time as only one source type was required, and also ensured the same source
 195 location and signature for both P- and S-wave SRT. This is in contrast to many other studies that are
 196 using distinct P- and SH-wave data acquisitions (e.g., Jongmans et al., 2009; Turesson, 2007).
 197 However, it required an additional processing step, as P- and S-wave source signatures had to be
 198 extracted. This was achieved by adding or subtracting the shots of the two different sides of the
 199 prism. Adding the two shots results in a “pure” P-wave source signature, that is, a vertically oriented
 200 force, and subtraction provides a “pure” S-wave source signature (Xia et al., 2002), that is, extraction
 201 of the horizontally oriented force (Fig. 2b). As the prism was oriented perpendicular to the geophone
 202 spread, only the horizontally polarized S-waves S_H will be analysed in following, assuming an
 203 isotropic S-wave propagation; S_H waves are referred to as S-waves hereafter. The addition and
 204 subtraction of shots of the two prism sides not only resulted in an extraction of the required wave

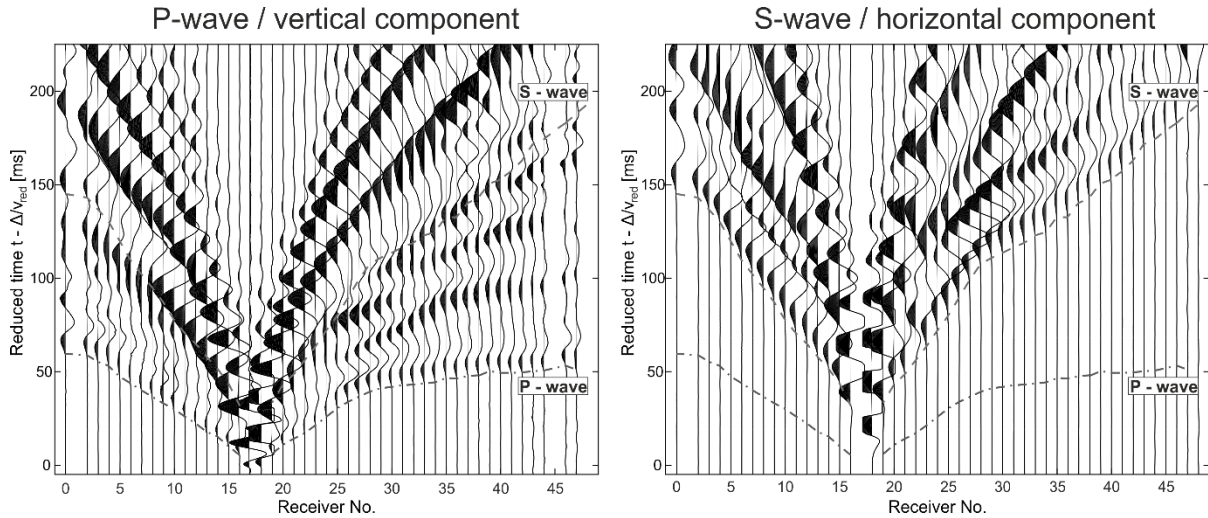
205 field, but also mostly increased the Signal-to-Noise (S/N) ratio by an additional stacking (i.e.
206 summation of two seismic traces). Note that inconsistencies between shots of different prism sides
207 may result in a deteriorated the signal.

208 2.1.3 Data quality

209 The raw data quality was generally good to very good, despite the comparably high attenuation
210 caused by the very soft material. The quality was further enhanced mainly by two procedures (1)
211 data stacking of the three shots of each prism side, and (2) the additional stack as part of the wave
212 field extraction. The initial stacking of shots from each prism side was guided by an analysis of the
213 correlation coefficient ρ_{XY} of the two seismic traces X and Y, which is defined as:

$$\rho_{XY} = \frac{\sum_{i=1}^N (X_i - \bar{X})(Y_{i+\tau} - \bar{Y})}{\sigma_X \sigma_Y} \quad (1)$$

214 with the variance σ , the number of samples N , and a lag τ . If two traces show a correlation
215 coefficient of $\rho_{XY} = 1$, the traces are identical. Correlation coefficients were calculated for each pair
216 of the three shots, and a stacking threshold of $\rho_{XY} > 0.85$ was applied; traces were only kept if at
217 least two of the three correlation coefficients were $\rho_{XY} > 0.85$. If, after this step, data acquired from
218 each of the prism side was available, horizontal and vertical wave components were extracted,
219 which implied a second stack. This requirement was fulfilled for more than 92 % of the data. These
220 steps significantly improved the S/N ratio from an average of 2.79 dB to 6.97 dB, aiding the correct
221 identification of the refracted waves (i.e. first arriving P- and S-waves, see Fig. 3).



222

223 *Fig. 3 Representative P- and S-wave shot gathers as generated after cross-correlation analysis, stacking, and wave-field*
 224 *extraction from the vertical and horizontal components, respectively. The two gathers show high S/N ratio, with first breaks*
 225 *clearly visible even at long offsets. Note that traces with low cross-correlation coefficients ($\rho_{xy} \leq 0.85$) were muted and*
 226 *gathers were reduced with a velocity of 3500 m/s.*

227

2.2 Data analysis

228

2.2.1 First break picking

229 The recorded wave field (Fig. 3) includes surface, reflected, and refracted waves. For the purpose of
 230 this study we concentrate on the refracted waves, as these contain information about the
 231 subsurface velocity structure and thus the elastic moduli. This structure can be determined from the
 232 first-arrivals (or first-breaks) of the transmitted waves (see Fig. 3; for receivers 30 to 45 first arrivals
 233 can be found between 40 and 60 ms). These were determined from the shot gathers by manual and
 234 semi-automatic picking of the P- and S-wave first arrival for each of the 526 shots. A picking error of
 235 ± 0.8 ms was determined from repeated picking of a subset of the data.

236

2.2.2 Inversion algorithm

237 The seismic P- and S-wave velocities of earth material can be defined in a simplified way as:

$$v_p = \sqrt{\frac{K + \frac{4}{3}G}{\rho}} \quad (2)$$

$$v_s = \sqrt{\frac{G}{\rho}} \quad (3)$$

238 where K is the bulk modulus, G the shear modulus, and ρ the density. K is defined as the ratio of
 239 hydrostatic stress to volumetric strain, and is a measure of a material's resistance to volume change
 240 under an applied stress. Similarly, the shear modulus is defined as the ratio of shear stress to shear
 241 strain (Mavko et al., 2009).

242 The methodology that was used to derive the subsurface velocity structure from the recorded travel
 243 times is described in detail in Lanz et al. (1998). In brief, tomographic images are derived from an
 244 algorithm that calculates the propagation of wave fronts through a 2-D heterogeneous medium and
 245 uses these results for an inversion to obtain the "true" subsurface velocity structures. The seismic
 246 problem can be simplified as a wave front traveling along the shortest ray-path in the time t from
 247 the source to the receiver i through a medium defined by its slowness (inverse of velocity) field u . If
 248 u is approximated by k cells with a constant slowness u , the forward problem can be formulated as
 249 (Lanz et al., 1998):

$$t = \sum_{k=1}^m G_{ik} u_k = \mathbf{G} \mathbf{u} \quad (4)$$

250 with G_{ik} representing the respective cell travel time derivatives. From a given slowness field u travel
 251 times t can be calculated by determining \mathbf{G} through minimization of the raypaths, using a finite-
 252 difference eikonal solver (Podvin and Lecomte, 1991). In the inverse problem, u is calculated from
 253 the determined first arrivals t . While in the case of the forward problem, the relationship between t
 254 and u is linear, in the inverse problem:

$$255 \quad \mathbf{u} = \mathbf{G}^{-1} \mathbf{t}, \quad (5)$$

256 due to the dependency of \mathbf{G} on \mathbf{u} , it is strongly non-linear and has to be solved iteratively. The
 257 inversion was performed separately for the P- and S-wave data.

258

2.2.3 Regularization

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Additional constraints are needed to determine \mathbf{u} from the seismic refraction data set, and are

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provided by the regularization parameters α and D_s . Including these parameters and error weights

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into the normal equation leads to the following notation of the inverse problem:

$$u_{est} = (\mathbf{G}^T \mathbf{W}_d^T \mathbf{W}_d \mathbf{G} + \alpha^2 \mathbf{I} + \mathbf{D}_s^T \mathbf{D}_s)^{-1} \mathbf{G}^T \mathbf{W}_d^T \mathbf{W}_d d_{obs} \alpha^2 \mathbf{I} u_{ref} \quad (6)$$

262

with the weighting matrix \mathbf{W}_d containing the data errors, the identity matrix \mathbf{I} , and the reference

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slowness field u_{ref} . The parameter α defines how much deviations from a starting model are

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penalized (i.e. damps the inversion), while D_s minimizes the roughness of the model (i.e. enforces

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model smoothness).

266

The starting model was chosen based on the guidelines given in Lanz et al. (1998). For the P-wave

267

inversion a starting model with a surface velocity of 500 m/s and a velocity gradient of 40 (m/s)/m

268

was chosen, with a maximum velocity of 2500 m/s, which represents a typical value for poorly

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consolidated sandstone (Telford et al., 1990). The S-wave starting model comprised a surface

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velocity of 100 m/s, a velocity gradient of 20 (m/s)/m, and a maximum velocity of 1500 m/s,

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representative of saturated clays (Mondol et al., 2007). Note that the defined maximum velocities

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are likely to overestimate the conditions of the study site, thereby ensuring sufficient ray coverage

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for the inversion (Lanz et al., 1998). For both P- and S-wave tomography the model was discretized

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in the same way, having initial cell sizes at the surface of 2.0 m and 0.5 m in horizontal and vertical

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direction, respectively. As ray coverage decreases with depth, cell sizes are slightly increased. A

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maximum model depth of 60 m was defined for profiles L1 to L4, and 35 m for profiles L5 and L6.

277

The regularization parameters α and D_s were chosen based on inverting a wide variety of

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combinations of these parameters. Their magnitude controls the overall amount of regularization; if

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the parameters are too small the inversion becomes unstable and no solution can be found, while if

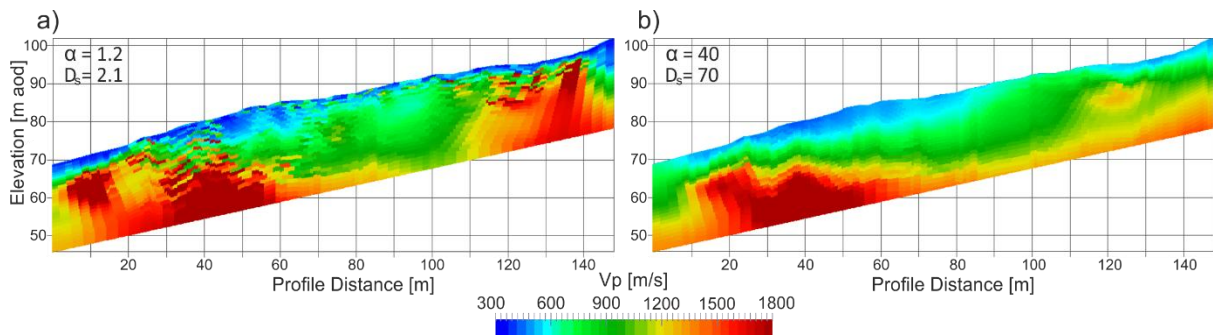
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they are too large the resulting tomogram will be overly smooth and/or show little deviation from

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the starting model (Fig. 4). After this test, the regularization parameters applied to all lines were

282 chosen as $\alpha = 8$ and $D_s = 14$; thus giving more weight to a smooth model than to a deviation from the
 283 starting model. The remaining root-mean-square (RMS) error between modelled and measured data
 284 ranged between 1.7 ms – 3.2 ms (for L2 and L6) for the P-wave travel time inversion, and between
 285 3.4 ms – 6.7 ms (L2 and L6) for the S-wave travel time inversion, and are slightly larger than the
 286 picking error.



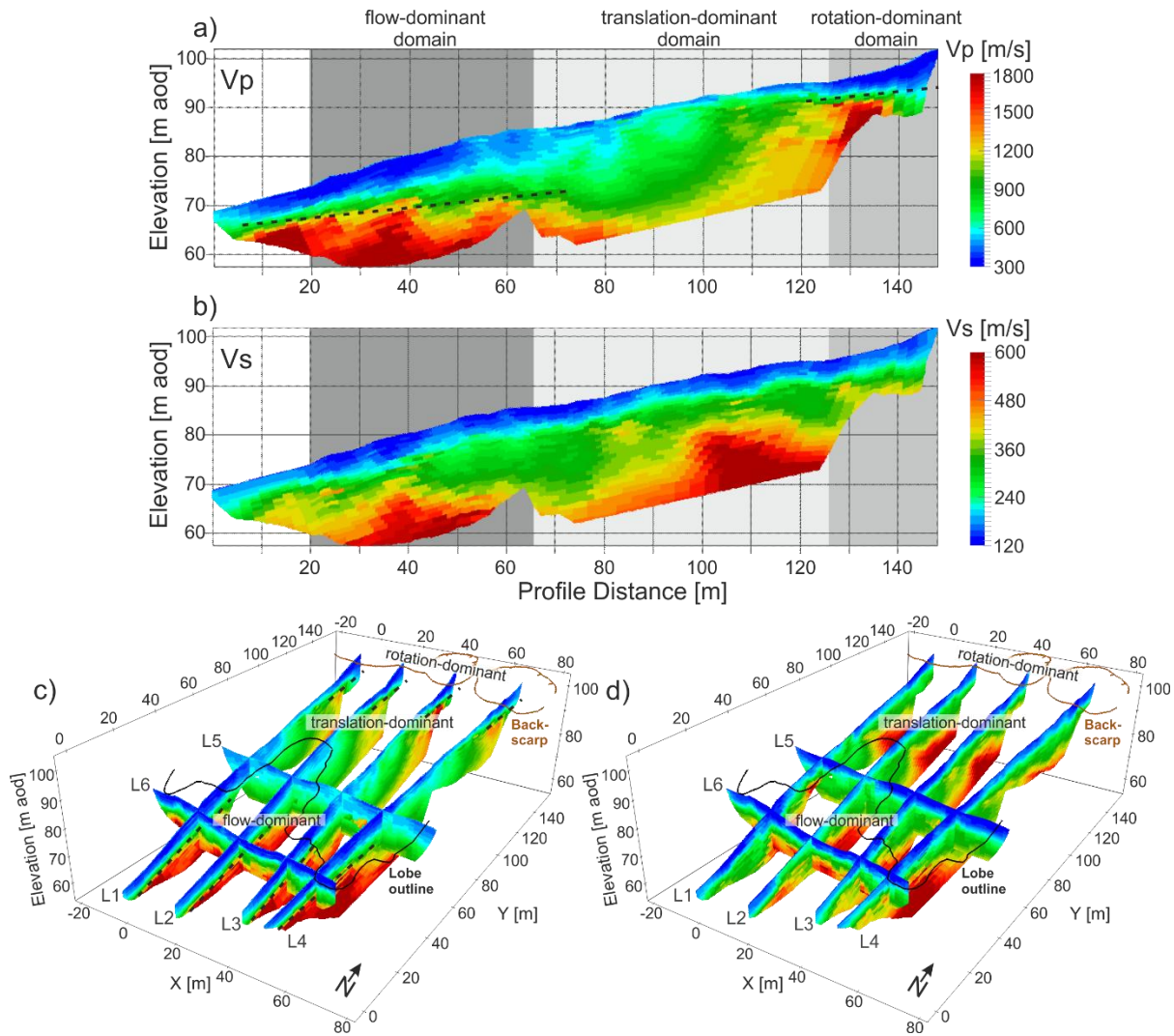
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 288 *Fig. 4 Data of Line 3 inverted using (a) small magnitude of regularization ($\alpha = 1.2$, $D_s = 2.1$) and (b) large magnitude of*
 289 *regularization ($\alpha = 40$, $D_s = 70$). The ratio between smoothing and damping has been kept constant. Note that a small*
 290 *amount of regularization results in larger small scale v_p variation, while a large amount of regularization leads to reduced*
 291 *resolution and an overly smooth image of the subsurface velocity distribution.*

292 3. Results

293 3.1 P-wave and S-wave tomography

294 The inverted P- and S-wave velocity models show generally very low to low velocities, with values
 295 ranging from 300 m/s to 1800 m/s, and 120 m/s to 600 m/s, respectively (Fig. 5). The smallest
 296 velocities in the P-wave tomograms ($v_p < 500$ m/s) are found less than 5 m below ground level (bgl)
 297 in the flow- and rotation-dominant domains of the landslide. In these domains, a sub-horizontal
 298 boundary can be found (dashed line in Fig. 5a), which in the flow-dominant domain increases in
 299 depth from about 5 m to 15 m bgl with increasing profile distance. At this boundary velocities
 300 increase rapidly from $v_p < 500$ m/s to $v_p > 1600$ m/s. This rapid increase is most pronounced at the
 301 flow-dominant domain, and is a consistent feature in all acquired profiles (Fig. 5c). Similar velocity
 302 gradients can be observed in the rotation-dominant domain of Line 3, but in the other profiles they

303 are smaller and the feature less pronounced. Common to all P-wave tomograms is a deep-reaching
 304 low velocity anomaly between $y = 65$ m to 110 m, thus characterising the translation-dominant
 305 domain. While shallow velocities (< 5 m bgl) are higher than in the neighbouring domains, the
 306 velocity gradients are much smaller, and thus a rapidly increasing velocity with depth is missing;
 307 velocities remain below 1000 m/s up to a depth of 25 m bgl.



308

309 *Fig. 5 a)-b) Images of P- and S-wave velocity distribution obtained from refraction data of Line 3. c)-d) 3D representation of*
 310 *all profiles (cross-sections of profiles L1, L2, and L4 to L6 can be found in the supplementary material). Highlighted are also*
 311 *domains of different movement characteristics (Gunn et al., 2013). Note that the lowest P- and S-wave velocities are within*
 312 *the lobes of the flow-dominated area of the landslides. Shown are only the parts of the tomograms with ray coverage of*
 313 *both P- and S-waves, and investigation depths < 25 m.*

314 The S-wave tomograms show no differences in the velocities of the shallow parts (< 5 m bgl) of the
315 translation-dominant domain compared to neighbouring flow- and rotation-dominant domains.
316 However, the lowest velocities ($v_s < 150$ m/s) are observed above 5 m bgl in the flow-dominant
317 domain. The lines of the eastern part of the landslide (Line 3 and 4) show a continuous shallow low-
318 velocity layer, while this thins out over the translation-dominated domain of the western part. A
319 significant increase in shallow velocities can be found just below the lobe (profile distance < 20 m,
320 Fig. 5b). This is a consistent feature of all profiles covering the lobes (Lines 1, 3, and 4). These lines
321 show significantly lower velocities in the flow-dominated domain than can be observed in Line 2,
322 which is located between two lobes. This can also be observed in the crosslines, Line 5 and 6, which
323 show higher velocities in this region ($15 \text{ m} < x < 25 \text{ m}$). The sharp boundary observed in the P-wave
324 velocity tomograms is less well-developed in the S-wave velocity sections, appearing slightly deeper
325 and with smaller velocity gradients. Similarly, a deep low S-wave velocity anomaly can be found in all
326 profiles, which is less distinctive than in the P-wave velocity profiles. There is also good spatial
327 consistency of the observed features in both the P- and S-wave velocity tomograms (Fig. 5c and d).

328 3.2 Elastic moduli

329 The most commonly used moduli to characterize soils are the small-strain shear modulus G_0 and
330 Young's modulus E (either expressed in terms of undrained/total stress or drained/effective stress
331 conditions). Both moduli provide a measure of the materials stiffness and are defined as ratio of
332 stress to resulting strain along an axis resulting from shear (G) or loading (Young's E ; Mavko et al.,
333 2009; Clayton, 2011). G_0 is commonly defined as the ratio of shear stress to shear strain (γ_s) for very
334 small strains ($\gamma_s < 1 \times 10^{-3}$; e.g. Atkinson, 2000; Benz, 2007). Guadalupe et al. (2013) describe that G_0
335 of soils shows a linear relationship with the effective stresses at failure for dilatant soils,
336 independent of density, degree of cementation and confining stress. Both, G_0 and E , are frequently
337 used in the estimation of soil consolidation (Biot, 1941; Das, 2008) and deformation analysis (Paice

338 et al., 1996; Giannakopoulos and Suresh, 1997; Clayton, 2011), as well as physical landslide
339 modelling (e.g., Lacroix and Amitrano, 2013). They are related through the Poisson's ratio ν :

$$G = \frac{E}{2(1 + \nu)} \quad (7)$$

340 For the purpose of seismic wave analyses both moduli are considered in terms of total stress
341 conditions.

342 Equations 2 and 3 show that v_p and v_s are defined by the density and elastic moduli of the material
343 that the waves are travelling through. Hence, if the distributions of seismic wave velocities and
344 density are known, elastic moduli can be calculated, with the shear and Young's modulus being
345 defined as:

$$G_0 = \rho v_s^2 \quad (8)$$

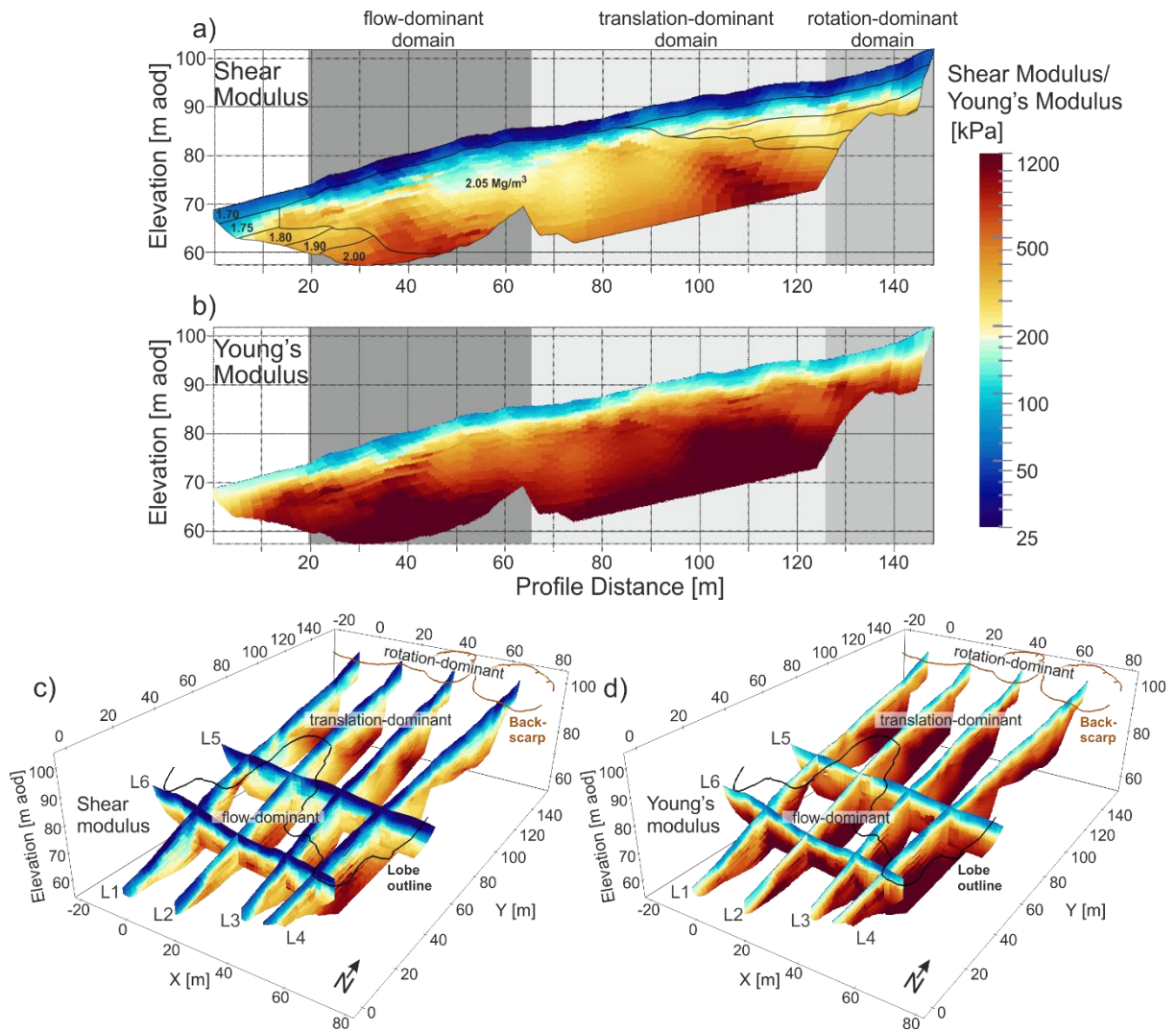
$$E = \frac{\rho v_s^2 (3v_p^2 - 4v_s^2)}{(v_p^2 - v_s^2)} \quad (9)$$

346 A density model (Figure 6a) was estimated based on laboratory analysis of samples taken from site
347 and by considering observed trends. The SSF was assigned a density of 2.05 Mg/m³, while for the
348 WMF and RMF a depth-varying density was assigned, increasing from 1.7 Mg/m³ at the surface to
349 2.0 Mg/m³ at about 15 m depth. These values were determined from site samples and informed by
350 characteristic values (Hobbs et al., 2012). This simplification is justified as shear and Young's
351 modulus show a linear dependence on the density, but quadratic to v_s . Thus the high sensitivity of
352 the elastic moduli to variations of v_s outweighs potential inaccuracies of the density model, which is
353 considered to be accurate to about 15 % of the true values.

354 As for v_p and v_s , the elastic moduli show low to very low values across the imaged landslide domains.
355 Fig. 6a shows the distribution of the shear modulus along Line 3, which spans across the recently
356 most active part of the landslide. The imaged features are comparable to the ones of the S-wave
357 velocity distribution. Very low shear moduli ($G_0 < 100$ kPa) are generally found at depths of less than

358 5 m bgl, with the lowest values located in the shallow, actively moving parts of the landslide (profile
359 distance > 20 m). The layer reaches its greatest thickness of up to 8 m in the upper part of the flow-
360 dominant domain. This is a feature that is observed in all lines covering actively moving parts (Fig.
361 6c). Values increase to more than 200 kPa below 5 m, with G_0 reaching maxima of about 1 MPa. At a
362 depth of about 20 m bgl, anomalies of higher shear moduli (>1 MPa) can be found below the flow-
363 dominant domain (20 m < y < 60 m) and the upper part of the translation-dominant domain (100 m
364 < y < 120 m). Young's modulus (Fig. 6b) shows a much thinner, shallow layer of E < 150 kPa, which
365 reaches down to about 2 m bgl only. This layer is thinnest in the most stable areas of the landslide
366 (line L2, Fig. 6d). Below this depth, Young's modulus rapidly increases to values of more than 1 MPa
367 in about 10 m depth. An anomaly with slightly lower E can be found below the boundary between
368 flow- and translation-dominant domains, with values of less than 500 kPa down to a depth of more
369 than 20 m.

370



371

372 *Fig. 6 Shear and Young's modulus. a)-b) Profiles of line L3, c)-d) 3D representation of all survey lines (cross-sections of*
 373 *profiles L1, L2, and L4 to L6 can be found in the supplementary material). The density model (in Mg/m^3) used in the*
 374 *calculation of the moduli is shown in a). Note that both shear and Young's modulus are plotted on the same colour scale.*
 375 *The location of the penetrometer test profile (P-P') is indicated in d). Shown are only the parts of the tomograms with ray*
 376 *coverage of both P- and S-waves, and investigation depths < 25 m.*

377

378 3.3 Poisson's ratio

379 Another commonly used parameter in slope stability analysis is the Poisson's ratio ν (e.g., Griffiths
 380 and Lane, 1999; Martel and Muller, 2000), which is strongly linked to the stress field in slopes and

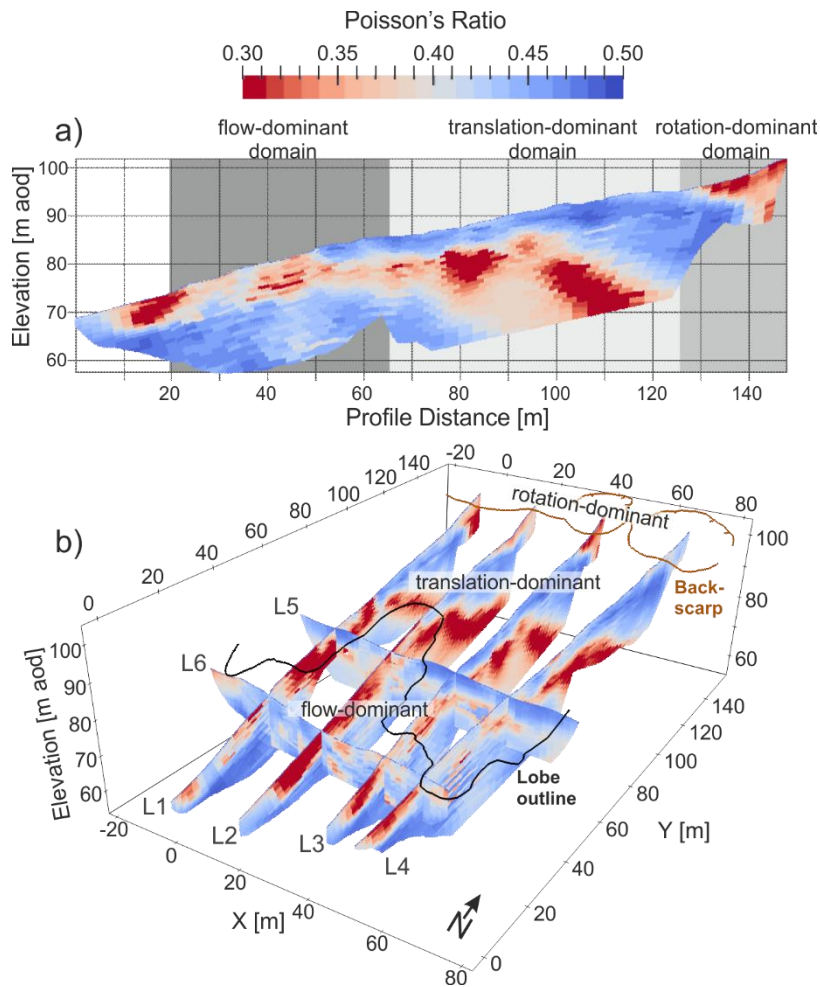
381 the degree of saturation of soil materials (Huang et al., 2012). It can be derived directly from the
382 inverted v_p and v_s distributions by (Mavko et al., 2009):

$$v = \frac{v_p^2 - 2v_s^2}{2(v_p^2 - v_s^2)}. \quad (10)$$

383 In contrast to the shear and Young's modulus, no density estimation is needed for the calculation of
384 v , highlighting the benefit of considering the Poisson's ratio by eliminating potential uncertainties
385 rising from an assumed density model.

386 The Poisson's ratio is usually positive and ranges between 0 and 0.5, where 0.5 is characteristic for
387 an incompressible fluid. For earth materials, v approaching 0.5 is characteristic for fully saturated
388 clays, while partially saturated silt or sandy clays show lower values between 0.2 and 0.4 (Davidovici,
389 1985; Bowles, 1988).

390 The Poisson's ratio profiles show spatially consistent features, delineating sub-horizontal, distinct
391 layers separated by v values of approximately 0.4 (Fig. 7). Throughout the survey area v shows a
392 minimum of about 0.08 and a maximum of 0.49. The shallow subsurface of the translation-dominant
393 domain is characterized by very high Poisson's ratios of $v > 0.40$, which reach deeper levels (down to
394 about 25 m bgl) towards the northern boundary of the study area. This is also evident in profile L3,
395 at profile distances between 100 m and 140 m; approaching the northern boundary, this layer of
396 high Poisson's ratio is overlain by a material with lower v . Note also that this layer, at its lower
397 boundary (at a profile distance between 50 m and 95 m) thins out and reaches the flow-dominant
398 domain. This is only evident on profiles covering lobes (L1, L3, and L4). Beneath, and extending to
399 the surface at the lower part of the flow-dominant domain (profile distance of 10 m to 50 m),
400 significantly lower values of v are found, ranging between 0.08 and 0.40. These values represent a
401 layer with a thickness ranging between 5 m and 20 m. In the shallow parts of the landslide complex,
402 this layer is most clearly distinguishable at L2, which is located between two lobes, without
403 accumulation of flow deposits. Below it, v increases again to values reaching 0.49.



404

405 *Fig. 7 Poisson's ratio of (a) profile L3, and (b) all profiles (cross-sections of profiles L1, L2, and L4 to L6 can be found in the*
 406 *supplementary material). Note the smaller values in the central part of the landslide. This area coincides with the previously*
 407 *known location of the SSF. Shown are only the parts of the tomograms with ray coverage of both P- and S-waves, and*
 408 *investigation depths < 25 m.*

409 4. Discussion

410 4.1 P- and S-wave tomography

411 P- and S-wave SRT was employed to delineate the thickness of the WMF deposits, as the WMF was
 412 expected to show lower seismic velocities than the SSF. This assumption was mainly based on
 413 expected differences in bulk density and elastic moduli; while the material of WMF can be classified
 414 as clay (Schaetzl and Anderson, 2005) with a bulk density expected to be about 1.7 Mg m^{-3} , the SSF is
 415 usually classified as a sandy clay to sandy clayey silt, with bulk densities exceeding 2.0 Mg m^{-3} .

416 However, neither P- nor S-wave velocity tomograms showed distinct velocities in areas known to
417 represent WMF and SSF (Fig. 1c). While shear wave velocities of less than 280 m/s are characteristic
418 for clay soils, soils of fine to coarse sand can show v_s values ranging between 70 and 800 m/s (Ohta
419 and Goto, 1978). Due to these overlapping ranges it was not possible to differentiate between WMF
420 and SSF solely from the S-wave SRT. Throughout the study area, weathering and destressing has
421 weakened these sedimentary lithologies to an extent that shear wave velocities are $v_s < 700$ m/s
422 above 20 m bgl (Yilmaz (2015) defines $v_s = 700$ m/s as a threshold to define 'geotechnical bedrock').
423 Hobbs et al. (2012) note that the bulk density of the WMF is likely to be reduced by periglacial frost
424 action, weathering and de-stressing in the near surface, affecting the material down to a depth of
425 about 10 m. In turn this will lead to a reduction in the shear modulus (e.g. Macari and Laureano,
426 1996) and, in conjunction with a high fissure density, causes the very low P- and S-wave velocities
427 observed in the upper 5 m bgl. Weathering usually decreases with increasing depth, and thus higher
428 P- and S-wave velocities are observed at deeper layers (Yamakawa et al., 2012).

429 P-wave velocities of about 800 m/s can be regarded as a critical stiffness threshold (CST) separating
430 'geotechnical bedrock' (in the sense of Yilmaz, 2015) from weathered/deconstructed materials
431 above. The depth at which this threshold is manifested at Hollin Hill is usually found between 5 m
432 and 12 m bgl. Above this depth, the lowest P-wave velocities are found, with minima being located
433 in the flow and translation dominated domains where materials are characterised by advanced de-
434 structuring and significantly increased porosities as a consequence of progressive straining and
435 reworking. The reduction of v_p with increasing porosity is higher for saturated material (Caris and
436 Van Asch, 1991; Mondol et al., 2007). This correlates with field observations where fully saturated
437 materials in the translations dominant domain are denser and thus have higher Poisson ratios in
438 comparison to the lower density flow deposits that deform more readily. The low values observed in
439 the backscarp area are generally due to partially saturated materials at the near surface during the

440 time of investigation. Comparing the laboratory results of Mondol et al. (2007) to the inverted P- and
441 S-wave velocities suggest that near-surface material may show porosities of up to 70 %.

442 While S-waves show only a limited response to changes in moisture content, P-wave velocities are
443 known to show a significant increase with increasing moisture content (Gregory, 1976). Thus the
444 high P-wave velocity anomaly ($v_p > 1500$ m/s), consistently found at depths between 5 m and 20 m
445 bgl in the lower part of the slope ($y < 60$ m) is likely to indicate the regional groundwater level
446 (Turesson, 2007). Extrapolating this boundary outside the study area coincides with the location of a
447 spring line below the toe of the slope. The DF at the northern-most part of the study area is known
448 to show a perched water table; increased v_p are likely to be caused by the elevated moisture content
449 in this area as well. Note that perched water tables are also found in the WMF and in the near-
450 surface materials of the flow lobes, particularly following prolonged or intense rainfall.

451 The study area is known to have been affected by paleo-landsliding (Chambers et al., 2011; Merritt
452 et al., 2013; Uhlemann et al., 2016). Thus the low-velocity anomaly in both v_p and v_s in the middle
453 part of the slope could reflect a potential paleo-landslide, leading to the formation of the relict
454 landslide deposits in the lower part of the slope (Merritt et al., 2013). However, the large lateral
455 extent and the abundance of this feature in the elastic moduli and Poisson's ratio are more likely to
456 suggest a lithological control (or increased weathering depth/extent). It is speculated that depth of
457 penetration of periglacial processes is greater where WMF is not covered by surface deposits in the
458 form of aeolian sands that have been found to cover the lower slopes (Uhlemann et al., 2016).

459 While the imaged P- and S-wave velocities do not provide much information about the extent of the
460 lithological units, it is possible to gain a clear indication of the depth to which weathering affects the
461 material, and, especially from v_p observations, provide an indication of the regional groundwater
462 table.

463

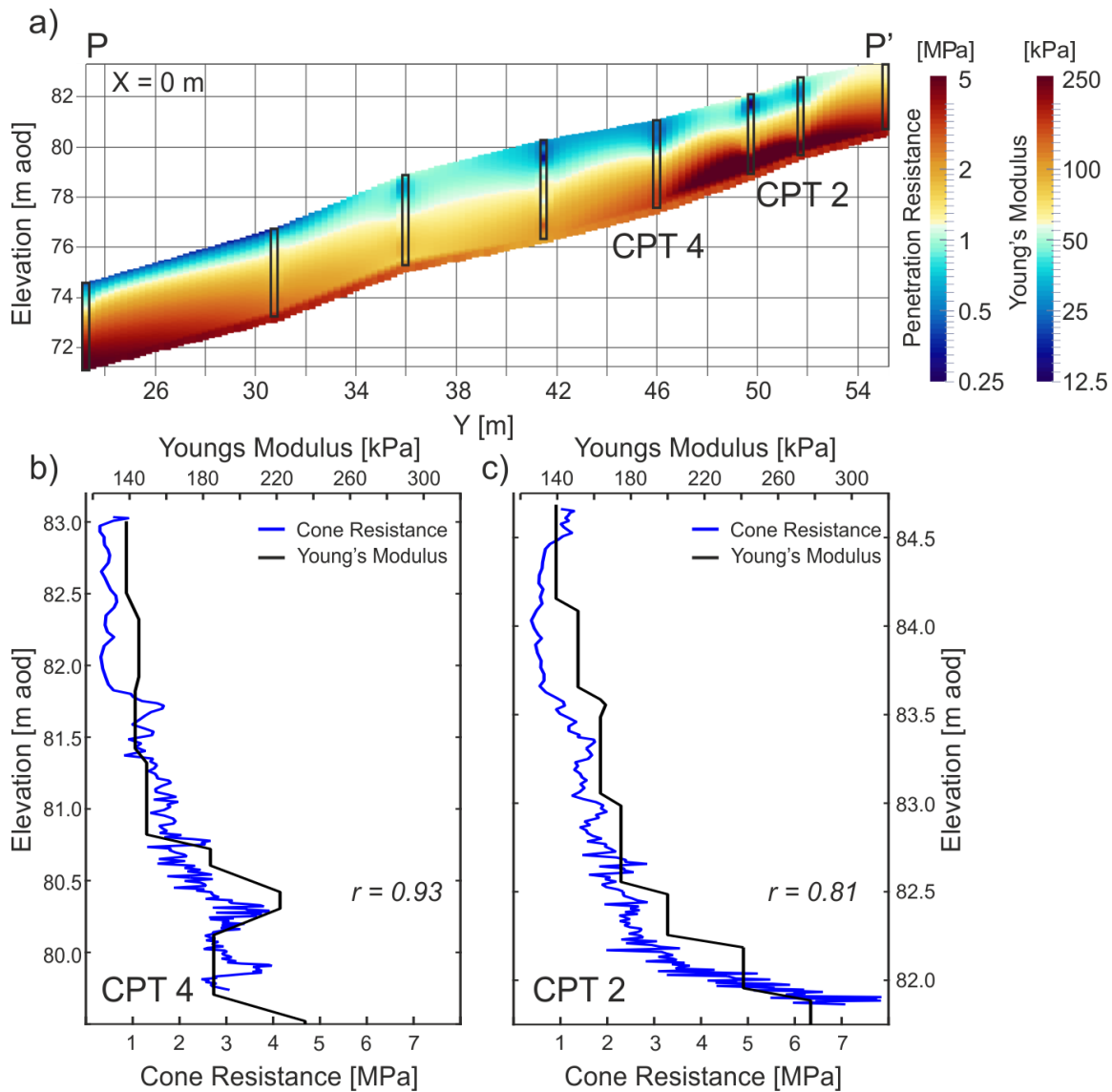
4.2 Elastic moduli

464 Similarly to the P- and S-wave velocities, shear and Young's modulus are reduced by weathering
465 processes (Macari and Laureano, 1996). The low values of the moduli in depths < 5 m bgl can
466 therefore be attributed to soil weathering and reworking through mass movements. In these shallow
467 depths G_0 remains mostly below 50 kPa, which is a typical value for clays and sands of low density
468 (Anderson and Stokoe, 1978). The small shear modulus indicates a low shear strength/internal
469 friction angle (residual friction angles are approximately 17 to 18 degrees at 0.5 m bgl; Merritt et al.
470 2013). Thus small elevations in pore pressures can decrease the effective stress at critical slip
471 surfaces to such an extent that landslide reactivation occurs despite a shallow slope angle of only
472 14° (Uhlemann et al., 2016). The rapidly increasing values of G_0 at depths > 5 m bgl indicate that the
473 majority of slope failures will occur above this depth, and hence deep-seated failures are unlikely.

474 Comparing the two moduli suggests that the weathering effect is not registered very clearly by the
475 Young's modulus; it shows values below 350 kPa only to about 2 m bgl. These are characteristic
476 values for very soft to soft clays with high plasticity (Kézdi and Rétháti, 1974). The same soil
477 classification was drawn from laboratory testing of samples of the WMF (Hobbs et al., 2012). With
478 values of up to 5 MPa, the Young's modulus of deeper layers takes values characteristic of soft to
479 firm clay and silt, and loose sands (Look, 2007). Examination of borehole logs obtained from shallow
480 boreholes (< 6 m) revealed a similar lithology and soil strength (Gunn et al., 2013) of material
481 representative for both WMF and SSF. A previous study employing cone penetration tests (CPT)
482 investigated the soil properties of the shallow material (< 4 m) of the lobes (Gunn et al., 2013). While
483 this formed a smaller and shallower investigation than was performed in the seismic study, it forms
484 an intrusive data set for comparison with the shear and Young's modulus derived from the seismic
485 data, between which commonly a linear relationship exist (Robertson, 2009). The agreement
486 between the CPT results (Fig. 8a) and the Young's modulus derived from the P- and S-wave SRT is
487 very good (Fig. 6), both in the magnitude and spatial correlation. Generally, the upper 0.5 m show
488 considerably smaller values than observed from the seismic data. This is most likely be caused by the

489 limited sensitivity of the seismic techniques within this layer. Both, CPT and SRT derived Young's
490 moduli show smaller values below the front of the lobe, between $y = 33$ m and 44 m. This is an
491 indication of a lower moisture content in this area (Gregory, 1976), but could also suggest a lower
492 local stress field and increased weathering/fabric dilation (Macari and Laureano, 1996). Direct
493 comparison of SRT derived Young's moduli with cone resistance at the CPT locations (Fig. 8b and c)
494 highlights this linear relationship between the two properties. The very good correlation between
495 SRT and intrusive investigation (Pearson's $r = 0.93$ and 0.81 for locations CPT2 and CPT4,
496 respectively) underlines their complementary nature.

497



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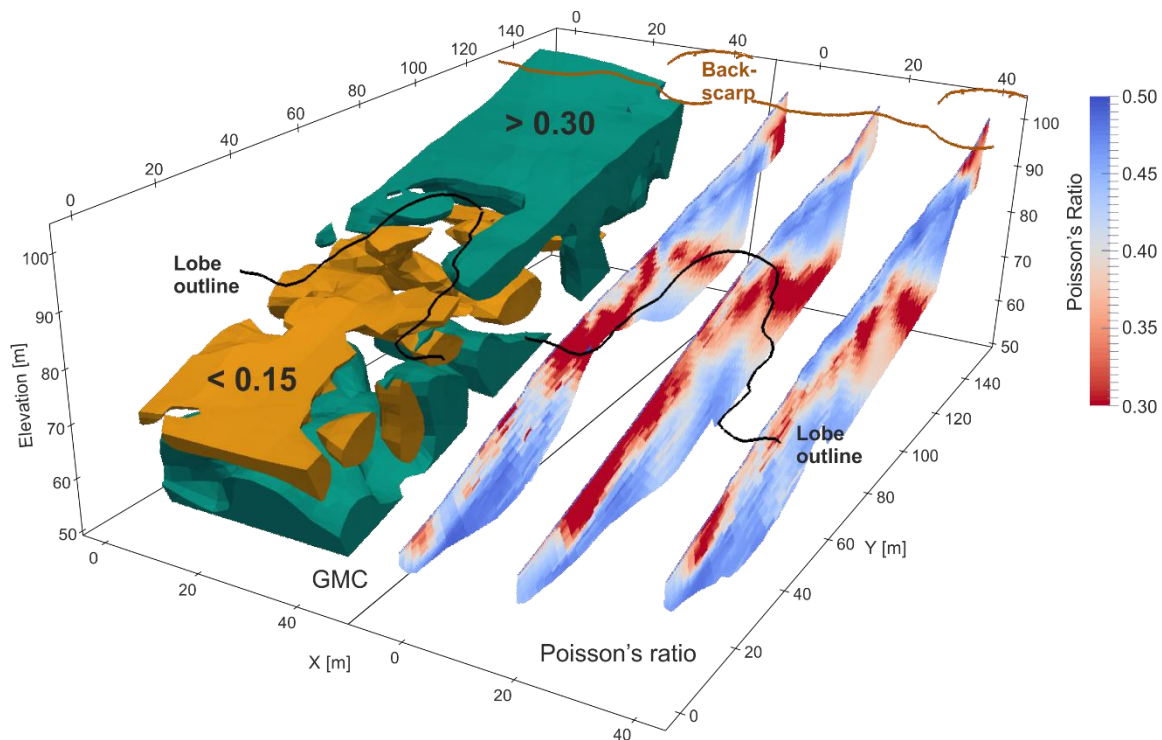
499 Fig. 8 a) Cross-section of penetration resistance, acquired along a 36 m long stretch next to L1 (see Fig. 1 and Fig. 5;
 500 modified from Gunn et al. (2013)). The area between test locations (black rectangles) was interpolated using an inverse
 501 distance weighting approach. The scale for the Young's modulus was derived from the cone resistance using a simplified
 502 linear relationship (Robertson, 2009). b) and c) intrusive cone resistance and SRT derived Young's modulus at CPT locations
 503 CPT 2 and 4. Note the very good correlation between the two methods.

504 **4.3 Poisson's ratio**

505 For the interpretation of the Poisson's ratio we define a threshold of $\nu = 0.4$, above which material
 506 can be classified as saturated clay or sand, while below this threshold the material is more likely to
 507 comprise partially saturated sand or silt (Bowles, 1988; Gercek, 2007). Applying this and comparing
 508 the imaged Poisson's ratio with the geological understanding of the site (Fig. 1c), a strong correlation

509 can be observed. While $\nu > 0.4$ in the translation-dominant domain coincides with the assumed
510 location of the WMF, $\nu < 0.4$ is found in the location of the SSF and DF. The layer of $\nu > 0.4$
511 underlying the central part of the slope indicates an increase in moisture content and perhaps
512 porosity (Gregory, 1976; Pasquet et al., 2015). It is likely to represent the saturated state of the SSF,
513 with its upper boundary representing the regional groundwater table. This is consistent with the
514 observations from the P-wave velocity profiles (Fig. 5c) and field observations. The decreasing values
515 at the southern-most part of the survey area may indicate the distribution of the RMF.

516 Electrical resistivity tomography (ERT) data was acquired during the time of the SRT survey using a
517 permanently installed monitoring system (Wilkinson et al., 2010, 2016). Both SRT derived Poisson's
518 ratio and ERT data are sensitive to variations in moisture content. Assuming that the electrical
519 conductivity of the pore fluid is constant over the imaging volume, moisture content can be derived
520 from ERT data provided a property relationship between moisture content and resistivity is known
521 (for details on data acquisition and processing see Chambers et al., 2011, and Wilkinson et al., 2016;
522 for details on translation of resistivity to gravimetric moisture content see Chambers et al., 2014,
523 Gunn et al., 2014, and Merritt et al., 2016). Comparing the ERT derived moisture content (iso-
524 volumes in Fig. 9) with the Poisson's ratio shows a good correlation (Pearson's $r = 0.53$). Note the
525 excellent agreement showing high moisture content and Poisson's ratio of the WMF sliding over the
526 SSF at the top of the eastern lobe at $x = 40$ m and $y > 60$ m. Also the central part of the SSF ($x = 20$ m,
527 $y < 60$ m) is shown to be of low moisture content and Poisson's ratio. Thus, a P- and S-wave derived
528 Poisson's ratio can be used to assess the moisture content of these formations.



529

530 *Fig. 9 Iso-volumes of ERT derived gravimetric moisture content (GMC) and SRT derived Poisson's ratios of profiles L1-L3*
 531 *(cross-sections). Shown are values of GMC > 0.30 (turquoise) and GMC < 0.15 (orange).*

532 At L2, located between the two studied lobes, the low Poisson's ratios of the SSF are very
 533 pronounced and show a clear distinction to the WMF and the underlying higher values of ν . A higher
 534 degree of distortion, resulting in higher Poisson's ratios, can be observed along the lobes of L1, L3,
 535 and L4, indicating higher moisture content than observed in the central part of the slope. This may
 536 support the hypothesis that mass movements of the flow lobes are controlled by base drainage at
 537 the sliding surface (Uhlemann et al., 2016).

538 4.4 Landslide characteristics

539 The landslide characteristics can be derived from a joint interpretation of the P- and S-wave velocity,
 540 elastic moduli, and Poisson's ratio distributions. The landslide consists, in general, of three types of
 541 materials, (1) saturated clay of the WMF overlying (2) partially saturated sandy silts and clayey silts
 542 and (3) saturated sandy silts and clayey silts of the SSF (Fig. 10). Next to the lithology, the degree of
 543 saturation/material density is a crucial input parameter for landslide modelling, as it provides
 544 indications of which geotechnical properties may be most appropriate to underpin the

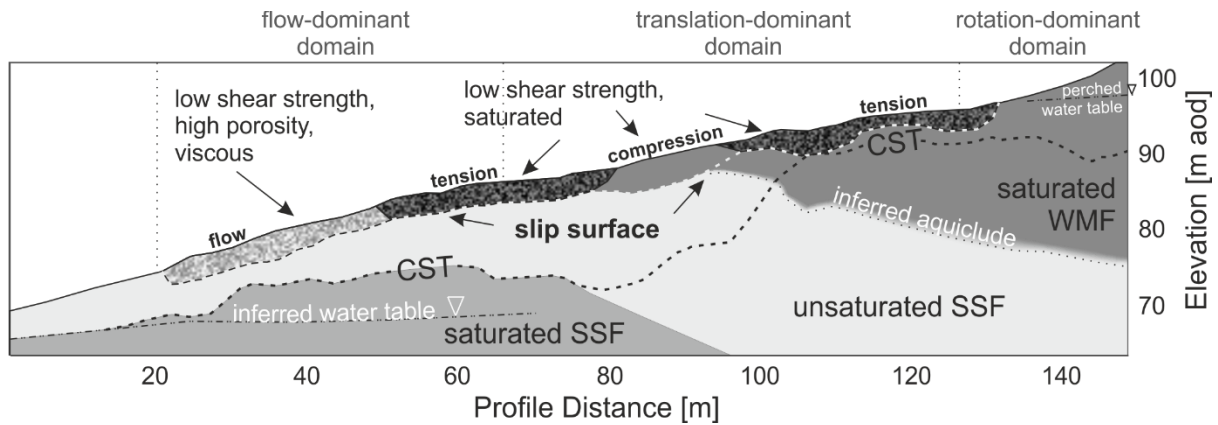
545 reconstruction of mass movement processes and support the numerical analysis of slope stability
546 (e.g., drained or undrained shear strength).

547 In the translation-dominant domain, a continuous surface zone of deforming deposits is observed
548 (Fig. 10) but with different relative densities. It is thought that these may relate to a wave of
549 deformation progressing through the slope at the time of observation. The darker shades represent
550 material that is likely under tension while the light-grey represents denser material that is likely
551 undergoing a phase of compression in the cascade of accumulating strains that progress downslope.
552 The lower part of the 'tension' zone extends into the top of the flow-dominant domain where
553 deformation is most likely controlled by undrained shear strength with transitions towards viscous
554 deformation. As deformation progresses, drainage of the reworked material is thought to take place
555 along the base of the flow lobes, leading to a gradual mobilisation of frictional resistance and
556 resulting in stabilisation of the mass movement even though the local slope becomes steeper.

557 Reactivation of landslide activity is a function of wetting up of the landslide body, predominantly
558 through direct infiltration following periods of prolonged/intense rainfall, further assisted by
559 groundwater inflow from the DF. During relatively dry periods comparably slow deformations (< 15
560 cm/year) within the lobes can be observed, and these are likely the local adjustments in strain to
561 even out imbalances in the tension-compression stress field in the translation-dominant domain.

562 The critical stiffness threshold (CST, $v_p = 800$ m/s) shows significant variation throughout the slope. It
563 is found between 2 m and 5 m bgl in the lower and upper parts, although increasing in depth
564 towards the northern boundary (profile distance > 135 m) of the study area. In the central part it
565 reaches depths of up to 17 m bgl. This could be a reflection of an enhanced susceptibility of a local
566 lithology to weathering and de-structuring. It is potentially possible that this boundary reflects
567 palaeo-mass movements, but this is not evident from existing borehole records (Uhlemann et al.,
568 2016). Note that P-wave velocities, on which the CST is based on, vary with saturation. Thus, its
569 shallower depth at $y > 100$ m could also be caused by a perched water table.

570 Under the current hydrological situation episodic deformation along a pre-existing slip surfaces
 571 predominantly caused by prolonged rainfall define the landslide behaviour. If drainage pathways
 572 close or reduce in the future, this may change and the risk of a comparably deep-seated failure
 573 should be reassessed.



574

575 *Fig. 10 Schematic ground model of the landslide, derived from the P- and S-wave SRT, elastic moduli, and Poisson's ratio*
 576 *distributions. The critical stiffness threshold (CST) indicates the perceived maximum depth of weathering and de-structuring.*

577 5. Conclusions

578 Site investigations are usually limited to surface observations, borehole or intrusive investigation,
 579 and laboratory measurements, providing surficial or information at depth profile or samples of
 580 discrete points only. In the case of landslide studies, where ground heterogeneities in both material
 581 and hydrological properties may define the failure mechanism and trigger, this is often not
 582 appropriate. The approach presented here overcomes this by employing P- and S-wave SRT, and
 583 deriving distributions of elastic moduli and the Poisson's ratio from this data. The main benefit of
 584 this study, and the information obtained from the Poisson's ratio in particular, is the spatial
 585 information relating to saturation state and potential strength of the ground. This information is
 586 crucial for an accurate definition of landslide models.

587 The P- and S-wave SRT indicated very low velocities of $v_p < 500$ m/s and $v_s < 150$ m/s in the depths
 588 above 5 m bgl. These could be related to a high degree of weathering, de-stressing and de-

589 structuring, with high porosity and low density. P-wave velocities of $v_p > 1500$ m/s close to the toe of
590 the slope were assigned to the regional groundwater table. Despite these features, v_p and v_s failed to
591 provide an indication of the different lithological units present at site. These were only imaged by
592 deriving the Poisson's ratio from the velocity distributions. The saturated clays of the WMF showed
593 Poisson's ratios $\nu > 0.4$, while the partially saturated sandy silts and clayey silts of the SSF showed $\nu <$
594 0.4 . Both shear and Young's modulus, also derived from the seismic velocity distributions, showed
595 small values ($G_0 < 1.0$ MPa, $E < 5$ MPa) throughout the slope, indicating the small strength of the
596 material constituting the slope. Minima of the elastic moduli were found at the actively moving parts
597 of the landslide, highlighting the reduced strength of the material leading to mass movements at
598 shallow slope angles. An interpretation of the mechanical properties derived from this study
599 concluded that deep-seated failures are unlikely, and occasional reactivation of landslide
600 movements in response to prolonged intense rainfall is the main failure mechanism.

601 It is difficult to directly compare material properties derived from field measurements and from
602 laboratory studies. Collecting truly undisturbed samples from the field is fraught with difficulty and
603 reconstructing the in situ stress field is very challenging. In addition, very small strain
604 characterisation of soft sediments and soils is very difficult using conventional laboratory
605 assessments (that are better at characterisation of intermediate to large strains). Further work is
606 needed to investigate the relationships in order to successfully combine the two approaches (e.g.
607 Mavko et al., 2009; Zhang et al., 2009).

608 This methodology has the potential to provide the spatial distribution of elastic moduli and Poisson's
609 ratio forming a major improvement upon the discrete sampling/testing programmes of standard site
610 investigations where large slopes are characterised by often very sparse data. The introduction of
611 spatially varying parameters in a 2/3D environment enables construction of detailed ground models
612 that form a step change in the analysis of landslide failure mechanisms and movement. In turn, a
613 better suite of tools to interpret landslide behaviour in greater detail will significantly contribute to

614 more appropriate management practices and disaster risk reduction strategies, particularly where
615 the landslide hazard affects vulnerable infrastructure and communities (Dijkstra and Dixon, 2010;
616 Dijkstra et al., 2014; Glendinning et al., 2015; Longoni et al., 2016).

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