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# **Landslide Activation Behaviour Illuminated by Electrical Resistance Monitoring**

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## **Abstract**

A common factor in landslide activation (or reactivation) is subsurface moisture and associated pore pressure variations linked to rainfall. Monitoring of these subsurface hydrogeological processes is necessary to improve our understanding of water-induced landslide activation. Geophysical approaches, electrical methods in particular, are being increasingly applied to landslide monitoring because they provide non-invasive spatial information in heterogeneous subsurface environments that can be difficult to characterise using surface observations or intrusive sampling alone. Electrical techniques are sensitive to changing subsurface moisture conditions, and have proven to be a useful tool for investigating the hydrogeology of natural and engineered slopes.

The objectives of this investigation were to further develop electrical resistance monitoring for slope stability assessment, and to validate the approach at an intermittently-active UK landslide system to advance the understanding of complex landslide activation mechanisms. A long-term transfer resistance dataset was collected from a grid of

25 electrodes to allow spatial monitoring of the landslide. These data were interpreted using a  
26 synthesis of rainfall, temperature, GPS and piezometric records. The resistance data were  
27 corrected for seasonal temperature variations and electrode movements were monitored,  
28 as these processes were shown to mask moisture related changes. Results reveal that  
29 resistance monitoring is sensitive to soil moisture accumulation, including changes in  
30 piezometric levels, and can be used to study the principal activation mechanism of slow-  
31 moving shallow earthflows. Spatial monitoring using resistance maps was shown to be  
32 particularly valuable as it revealed the evolution of subsurface moisture distribution, in the  
33 lead up to landslide activation.

34 Key benefits of this approach are that it provides a simple, rapid and non-invasive means  
35 of spatially monitoring subsurface moisture dynamics linked to landslide activation at high-  
36 temporal resolution. Crucially, it provides a means of monitoring subsurface hydraulic  
37 changes in the build-up to slope failure, thereby contributing to early warning of landslide  
38 events.

### 39 **Introduction**

40 If the effects of landslides are to be mitigated and avoided then landslide activation and re-  
41 activation mechanisms must be better understood. One way of developing a better  
42 understanding of landslide activation events is by monitoring subsurface changes during  
43 the period leading to activation. The most common change in the subsurface leading to  
44 activation (or reactivation) is the movement of water and associated pore pressure  
45 variations, which in turn are closely linked to antecedent rainfall conditions (Moore et al.,  
46 2007; O'Brien 2007).

47 If changes in slope hydrogeology can be observed in advance of activation then an early  
48 warning of slope movement may be possible. The moisture content, and therefore  
49 propensity to fail, of natural soils is directly affected by climatic, seasonal and  
50 environmental factors such as rainfall amount and intensity, as well as evapotranspiration.

51 Intense rainfall and rapid infiltration is widely accepted as one of the principal landslide  
52 triggers as slope materials show a reduction in mobilized strength with changing water  
53 pressure and associated effective stresses (Friedel et al., 2006; Dijkstra and Dixon, 2010;  
54 Dijkstra et al., 2014). In addition, a major contributing factor in clay slope instability is the  
55 dissipation of pore suction associated with elevated moisture content (Toll et al., 2011;  
56 Lourenco et al., 2011; Merritt et al., 2016).

57 Many landslide warning systems rely on the use of rainfall thresholds (e.g. Tiranti and  
58 Rabuffetti, 2010; Reid et al., 2012; Papa et al., 2013; Segoni et al., 2015), but there is a  
59 growing appreciation that direct observation of water in the subsurface is also desirable  
60 (e.g. Intrieri et al., 2013; Stahli et al., 2015). This is because the link between rainfall  
61 events and failure can be complex, requiring an understanding of both long-term  
62 antecedent weather conditions and subsurface heterogeneity (Dijkstra and Dixon, 2010;  
63 Take and Bolton, 2011). In recent years monitoring of landslide processes by geoelectrical  
64 methods has become more common (Perrone et al., 2014). Examples include short term  
65 studies using electrical resistivity tomography (ERT) to intensively monitor simulated  
66 rainfall events on vulnerable slopes to determine subsurface moisture variation in  
67 controlled conditions (e.g. Travelletti et al., 2012; Lehmann et al., 2013). Studies using  
68 ERT to monitor landsliding under natural conditions over a period of months have revealed  
69 the link between subsurface moisture distribution and rainfall, and demonstrated the ability  
70 of this approach to observe dynamic and complex hydrogeological processes in landslide  
71 systems (e.g. Lebourg et al., 2005 and 2010; Jomard et al., 2007; Bievre et al., 2012;  
72 Supper et al., 2014; Gance et al., 2016). Longer term multi-year studies have also been  
73 described. Uhlemann et al. (2017) describe the use of four-dimensional ERT to monitor an  
74 active landslide over a three year period, showing the relationship between increasing  
75 subsurface moisture content and failure events. However, the high spatial and temporal  
76 resolution presented here focuses on shallow landslide (re)activation to place emphasis on

77 the movement patterns of the type of landslide commonly affecting infrastructure assets  
78 (Loveridge et al, 2010). Palis et al. (2017) used three-dimensional (two-dimensional (2D)  
79 image plus time) ERT monitoring over a two year period to distinguish between moisture  
80 driven processes above and below the base of the landslide. In addition, and of particular  
81 relevance to this study, they correlated raw apparent resistivity measurements (i.e.  
82 unprocessed measurements) from their 2D line of ERT electrodes with subsurface  
83 moisture changes associated with individual rainfall events as well as longer term  
84 seasonal changes. These previous investigations reveal that time-lapse electrical  
85 measurements are a useful tool to observe hydrogeological processes due to their  
86 sensitivity to moisture content variation, and therefore have the potential to provide  
87 information on moisture driven landslide activation mechanisms.

88 The aim of this study is to investigate the benefits of applying a multi-sensory system,  
89 incorporating novel-geophysical monitoring, which records in near-real-time *both*  
90 environmental inputs and the resulting subsurface response. Presented here are the  
91 results of nearly five years of high-temporal resolution geoelectrical and environmental  
92 monitoring of a periodically active inland landslide located within landslide-prone Liassic  
93 rocks of the UK –representing one of the longest-term geophysical monitoring studies of  
94 an active landslide. To the best of our knowledge this is the only study where spatially  
95 distributed (i.e. using a grid of electrodes rather than a linear array) raw electrical  
96 resistance data have been used to monitor an active landslide using fully automated data  
97 acquisition. The overarching objective of this investigation is to utilise electrical resistance  
98 monitoring measurements to advance the understanding of complex landslide activation  
99 mechanisms, and is achieved through integration and analysis of monitoring campaign  
100 results. The sensitivities and benefits of using rapidly-generated resistance  
101 measurements, that only require minimal manipulation and without time-consuming  
102 inversion modelling are highlighted.

103

104 **Hollin Hill Study Site**

105 The study site is a landslide located in the UK county of North Yorkshire 20km North of  
106 York and 11km West of Malton (Merritt et al., 2014; Dixon et al., 2014), Ordnance Survey  
107 grid reference SE672706. It is situated on a south-facing slope approximately 450m by  
108 200m, which is used as pasture. The slope is approximately 50m high from the base to the  
109 top of the slope (mean slope angle of 12°). Beyond the base of the hillslope is a wide  
110 topographic embayment. The slope is composed of four geological formations of Lower  
111 and Middle Jurassic Age (Figure 1). The base of the Hollin Hill slope is formed of Redcar  
112 Mudstone Formation (RMF) and marks the oldest formation at the field site. This is  
113 overlain by Staithes Sandstone Formation (SSF) which gives way to Whitby Mudstone  
114 Formation (WMF), with Dogger Formation (DGF) capping the hill slope.

115 Present at the study site is a complex landslide system that exhibits a variety of landslide  
116 types and activity, with WMF being the most susceptible to instability (Jones et al., 1994;  
117 Foster et al., 2007). The landslide system extends ~250m laterally along the hill slope  
118 beyond the limits of the study site. Several types of slope failure can be observed at the  
119 test site, with the landslide system described as '*a very slow to slow moving multiple earth  
120 slide – earth flow*' (Chambers et al., 2011). The whole system would correctly be referred  
121 to as a complex landslide (Cruden and Varnes, 1996). However, as the focus of this study  
122 requires the differentiation between earthflow and earthslide regions of the landslide  
123 system, the landslide will be referred to using this terminology.

124 The landslide system has been the focus of previous geotechnical and geophysical  
125 investigations (Chambers et al., 2011; Gunn et al., 2013), including assessment of  
126 landslide structure, activation timings (Smith et al., 2014; Uhlemann et al., 2015b),  
127 conceptual model development (Merritt et al., 2014) and ERT monitoring (Uhlemann et al.,  
128 2017). The earthflow region of this landslide system is the most frequently active (Figure 1

129 and 2), with movement rates of up to 3.5m per year observed since monitoring began in  
130 2008 (Uhlemann et al., 2017). The earth flows are composed of highly weathered WMF,  
131 characterised as a high plasticity clay, with a thickness of up to approximately 5m (Merritt  
132 et al., 2014). Failure surfaces are predominantly within the upper two metres of the earth  
133 flows, but there is evidence of deeper failures surfaces at the base of the earthflows  
134 (Uhlemann et al., 2015b).

135

## 136 **Methodology**

### 137 ***Time-lapse transfer resistance measurements***

138 A permanently-installed geoelectrical monitoring system called Automated time-Lapse  
139 ERT (ALERT) developed by the British Geological Survey (Wilkinson et al., 2010;  
140 Chambers et al., 2015) was deployed on site. The remotely configurable system can be  
141 interrogated by wireless telemetry from the office via GSM (GPRS or 3G) or wireless  
142 internet link. Via this link pre-programmed data acquisition schedules are uploaded and  
143 measurement results downloaded. The system is powered by high-capacity batteries  
144 which are recharged by a combination of wind-turbine, solar panels and a methanol fuel  
145 cell. The ALERT sensor arrays were arranged in five parallel lines each comprising 32  
146 stainless steel electrodes, creating a grid of 160 electrodes. The electrodes were located  
147 0.1 m below the ground surface. Electrode lines are orientated downslope, i.e. 165°S,  
148 having a 9.5 m line spacing and 4.75 m electrode spacing. Thus, the monitoring grid  
149 covered an area of 147.25 m by 38 m. The ALERT system is designed to measure  
150 electrical transfer resistances using four-point measurements, comprising a current dipole  
151 (i.e. pair of electrodes) used to inject current, and a potential dipole that is used to  
152 measure the resulting potential difference. The system automatically undertakes  
153 measurements using predefined combinations of electrodes within the monitoring grid.  
154 Resistance measurements were acquired, using a standard dipole-dipole array

155 configuration, on an alternating daily basis with occasional gaps due to system, battery,  
156 electrode array, or telemetry failure. The first resistance measurements were taken  
157 11/07/2008.

158

### 159 ***Electrode position interpolation***

160 Electrode arrays at the study site are buried just below the ground surface. This was to  
161 prevent damage to the arrays by the livestock that graze the site. Consequently, when  
162 ground movement occurs the exact positions of the electrodes are not known; therefore a  
163 method to derive the best-estimate of electrode positions is required (Wilkinson et al.,  
164 2010; 2015; Uhlemann et al., 2015a).

165 An estimate of the location of electrodes, and therefore dipole-dipole measurement array  
166 size, is important when interpreting resistance measurements, so that resistance changes  
167 associated with movement can be differentiated from those associated with changing  
168 moisture conditions. Significant electrode movements (i.e. tens of cm) cause significant  
169 measurement variation, i.e. moving electrode closer together will lead to a smaller  
170 resistance, while moving them apart will lead to a higher measured resistance (Wilkinson  
171 et al., 2010). Also, accurate electrode location information is required in the modelling  
172 steps used to temperature correct resistance measurements (see following section).

173 Electrode positions are estimated using a known set of reference points (i.e. a coarse grid  
174 of GPS benchmarks), following an approach by Uhlemann et al. (2015a). From the  
175 reference points the electrode locations can be estimated using a piecewise planar  
176 interpolation scheme, where movements are assumed to be represented by the location  
177 changes of three non-collinear reference points. Uhlemann et al. (2015a) show that by  
178 using this methodology, electrode movements can be estimated to about 10 % of the



179 electrode spacing, thereby removing significant movement related artefacts from the  
180 resistivity data.

181

### 182 ***Temperature correcting transfer resistances***

183 Where time-lapse electrical resistance data are being compared over several months, it is  
184 important to correct measurements for the seasonal variation in subsurface temperature  
185 distribution (Hayley et al., 2010). This is necessary because the electrical resistance of  
186 rock and soil is not only sensitive to moisture content, but also temperature (Brunet et al.,  
187 2010). Therefore, without removing the effects of temperature variations from the electrical  
188 measurements it is difficult to differentiate between moisture and temperature driven  
189 changes.

190 The method used here to correct the time-lapse transfer resistance data for temperature  
191 variations is a two-stage process. Firstly, the temperature variation within the subsurface is  
192 approximated by a simplified homogeneous model subject to a yearly sinusoidal  
193 temperature variation at the ground surface (Brunet et al., 2010; Chambers et al., 2014).

194 The solution to the heat equation (Cannon, 1984) for this model is given by

$$T_{\text{MOD}}(z, t) = T_{\text{MAT}} + \frac{A}{2} e^{-\left(\frac{z}{d}\right)} \sin\left(\omega t + \varphi - \frac{z}{d}\right)$$

#### **Equation 1.**

195 where  $T_{\text{MOD}}$  is the subsurface temperature at day  $t$  and depth  $z$ ,  $T_{\text{MAT}}$  is the mean annual  
196 air temperature,  $A$  is the peak-to-trough magnitude of the annual air temperature variation,  
197  $d$  is the characteristic depth of the temperature variation,  $\varphi$  is a constant phase offset, and  
198  $\omega$  is the angular frequency ( $2\pi/365 \text{ day}^{-1}$ ). The constant phase offset ensures that the  
199 surface temperature is in phase with the air temperature. Seasonal subsurface  
200 temperature changes were recorded over a two year period using vertical arrays of  
201 temperature sensors at three locations on the landslide site (see Figure 1), and were fitted

202 to Equation 1 to define the temperature model (Chambers et al., 2014). The fitted  
203 parameters are listed in Table 1 along with the RMS misfits between the modelled and  
204 measured temperatures. Four separate models were fitted, one each for data from the  
205 individual locations and one for data combined from all three locations (which was the  
206 model used to correct the resistance data). For simplicity, the parameters were assumed  
207 to be independent of position and time. The misfit values are quoted for the two years of  
208 available data.

209 The second step is to correct the transfer resistances for the seasonal temperature  
210 variations. This involves assuming a linearized model for the variation of resistivity with  
211 temperature which is given by

$$\rho(T) = \rho(T_{MAT}) \left( 1 + \frac{c}{100} (T - T_{MAT}) \right)$$

212 **Equation 2.**

213 where  $c$  is the percentage resistivity change per °C, which is typically  $c \approx -2.0 \text{ } ^\circ\text{C}^{-1}$  (Hayley  
214 et al., 2007). To calculate the temperature-corrected transfer resistance  $R_{tc}$ , it is assumed  
215 that the seasonal variations due to temperature changes are small compared to the overall  
216 range of the resistances due to the resistivity structure of the ground. This is a similar  
217 approach to that taken by Hayley et al. (2010). A further simplifying assumption is made  
218 that, for a given measurement configuration, the ratio of  $R_{tc}$  to the uncorrected (measured)  
219 resistance  $R$  can be approximated by

$$\frac{R_{tc}}{R} \approx \frac{R_h}{R_v}$$

220 **Equation 3.**

221 where  $R_h$  and  $R_v$  are modelled transfer resistances for the same configuration.  $R_h$  is the  
222 transfer resistance resulting from a homogeneous half-space of resistivity  $\rho_h$ , and  $R_v$  is that  
223 resulting from a 1-D layered model where the variation of resistivity with depth is given by

224 Eq. 2 with  $\rho(T_{\text{MAT}}) = \rho_h$  and  $T = T_{\text{MOD}}$  as given by Eq. 1. Ratio corrections such as these  
225 have previously been used to model the effects of other types of small perturbations (e.g.  
226 those due to topography, Tsourlos et al. (1999)). The use of a 1-D model allows the  
227 correction factors to be calculated rapidly (Ingeman-Nielsen and Baumgartner, 2006) as a  
228 function of time. Therefore, for each measurement configuration, the temperature-  
229 corrected transfer resistance is given by

$$R_{\text{tc}} = \frac{R_h}{R_v} R$$

#### 230 **Equation 4.**

231 The process of temperature correcting transfer resistance data for the analysis of  
232 subsurface physical processes adjusts raw data by  $\pm 0.04\Omega$ , depending on whether  
233 modelled subsurface temperature is higher or lower than the averaged modelled  
234 subsurface temperature for the depth of interest. When comparing raw transfer resistance  
235 data with temperature corrected resistance data (e.g. Figure 3), it is apparent that the raw  
236 data varies much more seasonally, and is systematically higher in winter and lower in  
237 summer than temperature corrected data. There is a lag of  $\sim 1.5$  months between weekly  
238 air temperature and resistance change ( $\Delta\Omega$ ), which is due to the time taken for air  
239 temperature changes to propagate to the median depth of investigation (see Eq. 1).

240

#### 241 ***Transfer Resistance Monitoring***

242 The geoelectrical monitoring campaign comprised 695 geophysical surveys of all five lines  
243 during the four years and nine months of monitoring equating to 1740 days of monitoring,  
244 with ERT surveys performed on average every 2.5 days. The results of the geoelectrical  
245 monitoring campaign are a series of 'raw' transfer resistance measurements, which were  
246 corrected for the effects of subsurface temperature variation. Each dipole-dipole transfer  
247 resistance measurement presented here was performed using a four-electrode

248 arrangement of adjacent electrodes, comprising two current (C) and two potential (P)  
249 electrodes arranged in the following order, C2-C1-P1-P2. Each four-electrode array had a  
250 length of 14.25m and a median depth of investigation of 1.9m. Measurements were  
251 performed using all available C2-C1-P1-P2 combinations of along each of the five lines.

252 Long-term temperature corrected resistance monitoring results are given as both 1D time-  
253 series from four selected dipole-dipole measurement locations (ML1, ML2, ML3, ML4 -  
254 Figures 4 and 6) and as 2D maps (Figure 7) using all resistance measurements made on  
255 the five lines of electrodes (with resistances plotted at the midpoints of each individual  
256 four-electrode measurement array). The results are presented as resistance ratios in  
257 Figures 4 and 6, and as resistance in Figures 7 and 8. Resistance ratio is the resistance at  
258 time  $t$ , normalised to the initial (or baseline) resistance measurement, and is a useful way  
259 of displaying how the measured resistances changed of over time.

260

## 261 ***Environmental Monitoring and Modelling***

### 262 ***Rainfall and Evapotranspiration***

263 Rainfall was monitored at the research site (see Figure1b for rain gauge location) by 0.1  
264 mm tipping-bucket type rain gauge to complement the results of the geoelectrical  
265 monitoring regime with soil moisture input information. Rainfall data are presented as 2-  
266 week running mean (Figures 4 and 6), weekly total and weekly effective rainfall, with the  
267 latter requiring the estimation and removal of potential evapotranspiration effects, in  
268 mm/day, from total rainfall records using Hargreave's method (Hargreaves and Allen,  
269 2003). Note that effective rainfall can be either positive (i.e. moisture input from rainfall  
270 exceeds moisture loss due to evapotranspiration causing an increase in soil moisture) or  
271 negative (i.e. moisture input from rainfall is less than moisture loss from evapotranspiration  
272 resulting in drying of the near surface).

273

#### 274 ***Piezometric Levels***

275 Stand-pipe and water level loggers (Solinst Levellogger Junior Edge) were installed on  
276 each earthflow and recorded groundwater level from September 2009. A borehole was  
277 advanced to depths of 2.85 m and 2.8 m on the western and eastern lobes, respectively.  
278 These depths were chosen in order to place the active zone of the piezometers in the  
279 vicinity of the depth to predicted shear surfaces, which were determined using Cone  
280 Penetration Testing (CPTU) downhole tool measurements (Gunn et al., 2013). A 19 mm  
281 uPVC pipe, fitted with a 0.9 m slotted, porous piezometer tip was installed in each stand-  
282 pipe (see Figure 1 for location). Each hole was backfilled with clean sand to 1.95 m and  
283 1.75 m below ground for the western and eastern piezometer, respectively, forming an  
284 active zone that allows for monitoring of the pore water pressure in the vicinity of the slip  
285 surface (located at 1.6 m depth on the eastern lobe). The remainder of the borehole was  
286 backfilled using bentonite granules to ensure sealing. Since the active zones of the  
287 piezometers are in close proximity to the shear surface ( $< 0.35$  m), the measured pore  
288 water pressures are indicative for the conditions at the shear surface. The piezometer is  
289 located at a depth in the stand pipe corresponding to the depth of the deepest periodically  
290 active slip surface. Meanwhile, the depth of the most active slip surface ( $\sim 0.8$  m depth  
291 below ground level in the eastern earthflow) is annotated along with piezometry in Figures  
292 4 and 6.

293

#### 294 ***Ground Movement***

295 Ground movement and estimates of landslide activity were derived from GPS  
296 measurements of benchmarks, as described in the electrode position interpolation section  
297 (also see Uhlemann et al., 2015a and b), and from tilt meter records (Uhlemann et al.,

298 2015b and 2017). These records provide evidence of landslide activation and very slight  
299 slope displacements, which began early to mid-July 2012.

300

## 301 **Results**

### 302 ***Overview***

303 To our knowledge, the application of temperature-corrected electrical resistance  
304 measurements to observe the hydrogeological precursors to shallow landslide activation  
305 has not previously been reported in the literature. Therefore, the content of the results  
306 section aims to provide a complete analysis of the processes taking place within the  
307 shallow subsurface throughout the monitoring period. The results section is divided into  
308 four sections; the first, presents general geophysical observations of general hillslope  
309 processes, while the second and third sections focus on earthflow activation processes at  
310 both high-temporal and high-spatial resolutions, respectively. The landslide system is  
311 divided into several regions based on hydrogeological behaviours in Section 4. These  
312 regions were formulated through integration of monitoring datasets.

313

### 314 ***General Monitoring Results: 2008-2013***

315 Baseline data on landslide movement and environmental conditions were established in  
316 this period (Figure 4). Table 2 contains a summary of general hillslope and earthflow  
317 monitoring observations. Statistical analysis of piezometric levels and TC-res data reveal  
318 strong negative correlation coefficient of -0.65 (Schumann, 1998), while a  $p$ -value of  
319  $<0.001$  confirms that there is significant correlation between the two datasets. Note that  
320 the lag between the piezometric and resistivity data was taken to be zero since observed  
321 lags in similar studies have only been significant on timescales of hours to days

322 (Chambers et al., 2015). The results from the four measurement locations respond to 'wet'  
323 and 'dry' periods to different degrees.

324 Two wetter periods of substantially longer duration exist during the entirety of 2008 and  
325 only three short periods of negative effective rainfall occurred between July 2011 and  
326 March 2013 (see Figure 4 and Figure 6). The two years where landslide activations  
327 occurred during the monitoring period took place during years with higher than average  
328 annual rainfall (751 mm/year) (Figure 5). Both of these prolonged wet periods are  
329 associated with earthflow activation events with the latter being discussed in more detail  
330 later. Earthflow deposits activated during these prolonged wet periods, between July 2008  
331 - April 2009 and August 2012 – February 2013.

332

### 333 ***Pre-reactivation and Reactivation Monitoring: 2011-2013***

334 Temperature corrected resistance results in the fourteen months leading to earthflow  
335 activation are presented in Figure 6 and reveal several additional trends associated with  
336 progressive landslide activation processes. July 2011 was preceded by a three month  
337 period of negative effective rainfall and so is characterised by low piezometric levels and  
338 some of the highest TC-resistances recorded during the monitoring period (ML1 has  
339 resistance ratios close to 1.4). July and August 2011 see resistances fluctuate yet remain  
340 high, in response to two periods of rainfall. These rainfall events are not sufficient to raise  
341 the piezometric level and so the piezometer remains constant at 77.6 m AoD. Piezometric  
342 levels rise slightly at the end of August 2011 in response to rainfall at a time when the  
343 ALERT array was not fully functional.

344 Between August 2011 and February 2012 three periods of prolonged positive effective  
345 rainfall occurred, ranging between 21 mm and 17 mm of weekly rolling averaged rainfall.  
346 These three periods result in a 0.6 m rise in piezometric level and occurs in a stepped

347 manner. During this 7 month time-frame TC-resistances across all four earthflow  
348 measurement locations markedly decrease. ML1 exhibits the greatest decrease from 1.35  
349 to 1.15, with the other three measurement locations displaying less pronounced decreases  
350 of between 0.05 and 0.15.

351 March 2012 is a relatively dry month as it experienced only negative effective rainfall, and  
352 was accompanied by associated piezometric level falls and TC-resistances either slightly  
353 raise (ML1, 0.1 rise) or remain constant. The next six weeks (April to early May 2012) sees  
354 a rapid piezometric level rise from 77.8 m to 78.4 m AoD. TC-resistance for all four  
355 earthflow measurement locations either remain constant or decrease slightly during this  
356 time and could be indicative of the imaged slope material nearing saturation. The second  
357 half of May experiences negative effective rainfall and the piezometric level fall causes  
358 ML1 resistance ratios to increase by 0.15. The other three measurement locations again  
359 either remain relatively constant or reduce very slightly.

360 Between June and mid-August 2012, piezometric levels fall at a time when relatively high  
361 rainfall is recorded. TC resistances during this period initially decrease during June but  
362 then increase during the latter half of July 2012. Earthflows reactivate at a time when  
363 piezometric levels are falling and TC resistance values are at a 24 month minimum. Once  
364 movement is initiated, earthflows remain active until February 2013, a duration of just over  
365 six months. During this active period rainfall is at its most intense (the three highest peaks  
366 occur during this period).

367 One month after earthflow reactivation piezometric levels begin to rise once again and TC  
368 resistances reduce and levels off at between 0.9 (ML2 and ML3) and 1.15 (ML1). TC  
369 resistance values begin to jump (diverge), either more positive or negative, from October  
370 2013 until the end of the monitoring period. The earthflow-installed piezometer became  
371 trapped in the standpipe during this period, hinting at substantial earthflow displacement.



372 Piezometric levels reach their highest levels during the active earthflow period (November  
373 2012), and are coincident with substantial TC resistance divergence.

374

375

### 376 ***Pre-reactivation and Reactivation Monitoring Maps: 2011-2013***

377 The spatial variation of temperature corrected resistance is presented as a series of twelve  
378 time-lapse resistance maps (which show TC resistance change relative to a baseline)  
379 extending over the 14 month period preceding earthflow reactivation (Figure 7). The  
380 baseline is an average of resistance measurements made during 2010, a period when the  
381 landslide was inactive. Each map represents a snapshot in time, and it is therefore more  
382 difficult to identify trends in the data compared to the high frequency resistance ratio time-  
383 series given in Figures 4 and 6; nevertheless, temporal and spatial patterns can be  
384 observed. Firstly, the earthflow region and the SSF towards the base of the slope show  
385 significantly more variability in response to rainfall than the upper regions of the slope –  
386 perhaps indicating higher infiltration rates towards the base of the slope due to fissuring  
387 and, in the case of the SSF, coarser material and hence higher permeability. Secondly,  
388 around the time of earthflow reactivation the lower earthflow regions of the landslide show  
389 a very marked decrease in resistance (i.e. increase in moisture) to levels lower than at any  
390 time in the preceding 12 months. High levels of spatial variability are observed in the  
391 earthflow regions, which reflects ground movement resulting in fissuring and localised  
392 accumulation and drainage of moisture.

393

### 394 ***Pre-reactivation and reactivation Monitoring by Zones: 2011-2013***

395 The landslide system is divided into zones based on their electrical responses to  
396 environmental inputs. The landslide zones are shown on Figure 8 and, Table 2 is a

397 summary of interpreted hillslope hydrogeological behaviours. In brief, Zones 1, 2 and 3 are  
398 defined as the Backscarp, Head and Sag Pond, and Upper Body respectively. These three  
399 zones display relatively small changes in resistance over the period, which is perhaps due  
400 the low permeability clay soil and a relative lack of fissuring resulting in more consistent  
401 moisture retention. No significant decrease in resistance is observed in the months  
402 preceding landslide activation, although a small drop in resistance and greater variability  
403 accompanies the period of landslide movement towards the end of 2012. Zone 4 is defined  
404 as the Lower Body and Flow Lobes, and displays a steady drop in resistance in the  
405 months preceding landslide activation, which is in accordance with the resistance  
406 measurements shown in Figure 6. This is the most active region of the landslide system,  
407 with the greatest degree of fissuring. Zone 5 is defined as Between Flow Lobes and is a  
408 stable region of well drained SSF.

409

## 410 **Discussion**

### 411 ***Processing of Raw Resistance Measurements***

412 Plotting of raw transfer resistance monitoring data – without any form of processing –  
413 revealed that subtle resistance changes are masked by seasonal air temperature  
414 variations which propagate into the subsurface. This is a significant limitation for the  
415 monitoring and investigation of shallow landslides (<5 m). It will have less impact on  
416 deeper landslide systems in temperate climates. The sinusoidal nature of transfer  
417 resistance variation in response to air temperature variation acts to reduce resistance in  
418 the summer months when air temperature is higher than the annual average temperature  
419 and increases resistance in the winter months when air temperature is lower than the  
420 annual average.

421 Temperature correcting raw resistance monitoring results using the method proposed by  
422 Hayley et al. (2010) makes interpretation of resistance results for shallow landslides much  
423 simpler as one major external process which affects resistance seasonally has been  
424 modelled and removed. By altering the procedure outlined by Hayley et al. (2010) to model  
425 a correction ratio for every transfer resistance measurement, as oppose to modelling a  
426 single correction factor and applying it to all transfer resistances, the method was adapted  
427 to be more applicable to monitoring landslides, because measurement electrode  
428 geometries change when landslides activate.

429

### 430 ***Subsurface Environmental Conditions***

431 The main trends observed in temperature corrected resistance data include: TC resistance  
432 ratio highs during periods of low piezometric levels, and conversely, resistance ratio lows  
433 when piezometer levels are high. The eastern earthflow resistances reveal that some  
434 small rainfall events are not identified by piezometry but are responsible for small changes  
435 in resistance and, are attributed to the transfer resistances being sensitive to shallow  
436 moisture content and that piezometer observations only provide point data in the landslide  
437 system.

438 Resistances measured on the upper body of the landslide change more slowly in response  
439 to negative effective rainfall (i.e. drying). This is likely to be the result of fewer tension  
440 cracks within the slump section of the landslide, and them rapidly annealing after rainfall.  
441 Given that the main scarp of the slump transmits surface runoff into the earthflow systems  
442 the presence or absence of cracking here influences reactivation strongly. The upper body  
443 of the slumped region (Figure 1, Figure 8 - Zone 3, Figure 9) dips less steeply than  
444 earthflow regions, precipitation therefore has the time to penetrate the subsurface, and run  
445 off only occurs after crack annealing has taken place. This small resistance variation  
446 observed at this region is attributed to the soil moisture varying very little throughout the

447 year. Its small resistance variation and lack of rainfall infiltration flow pathways potentially  
448 indicates that the slump region of the landslide holds on to its moisture and doesn't freely  
449 release it like the heavily cracked earthflow region.

450 Resistance monitoring results were compared with piezometer measurements of the  
451 eastern earthflow region. Correlation coefficients between the two datasets suggest a  
452 negative correlation (-0.65) between piezometry and electrical resistance response.  
453 Therefore, as piezometric level rises due to rainfall infiltration the electrical resistances  
454 generally decrease, which is similar to the behaviour reported by Lebourg et al. (2010) in a  
455 short term landslide monitoring study in sandy clay materials. However, it should be noted  
456 there are deviations from this pattern in the data that have not be previously observed; for  
457 example in the month preceding activation piezometric levels show a consistent drop  
458 (albeit from a three year peak), whereas over the same period a drop in resistance is  
459 observed followed by a period of increase. This is discussed further in the following  
460 section.

461 Pervasive and deep cracking presents an impediment to electrical current flow, and  
462 therefore results in resistance increases in the vicinity of cracking. This trend of more  
463 elaborate resistance responses to environmental factors is attributed to thin earthflow  
464 regions being more susceptible to pervasive desiccation during dry summer months. They  
465 are more susceptible because the shear surfaces between individual flows act both as a  
466 conduit for water drainage, assisting flow through and out to underlying formation, and as  
467 an aid to joining up desiccation cracks, further encouraging their development. The  
468 opening of cracks within the silty clay dominated earthflows is thought to become more  
469 effective as the dry spell progresses, as cracks open up the ground to further drying and  
470 causes resistances to continue increasing. Resistance values reach a summer peak at  
471 around August/September associated with desiccation.

472 Desiccated earthflow toes of Zone 4 (Figure 8), composed of a series of overlapping and  
473 overriding thin landslide deposits, may permit more fluid to enter the subsurface when  
474 compared to less desiccated thick successions of WMF (Figure 8 - Zone 3). This is due to  
475 desiccation cracks being conduits for fluid to enter the subsurface. Furthermore, these  
476 cracked earthflow regions retain very little rain water and as a result resistances in these  
477 regions rise and fall sharply.

478 These observations accord with those of Bièvre et al. (2012) in that the geophysical  
479 signatures of fissure dynamics are similar and indicate that preferential flow is occurring;  
480 however, the longer term spatial monitoring presented here has shown a greater range of  
481 fissure behaviour (crack formation and annealing) and more variable drainage associated  
482 with longer term spatial monitoring over many seasonal cycles in a more complex  
483 landslide system.

484

#### 485 ***Earthflow Reactivation Mechanisms***

486 The sensitivity of the resistance monitoring system to soil moisture accumulation and  
487 piezometric level variation highlights an interesting process taking place in the months  
488 preceding earthflow reactivation (Figures 6 and 9). The system successfully identified the  
489 fall in piezometric levels up to June 2012 (Figure 6), which manifested as an increase in  
490 resistance up to this point. However, as described in the previous section the two  
491 measures deviated from one another shortly in advance of landslide reactivation; despite  
492 significant rainfall piezometric levels dropped, whilst resistance decreased. The drop in  
493 piezometric levels leading up to landslide reactivation was unexpected, as increased pore  
494 pressures are widely recognised as a key driver of landsliding (e.g. Iverson and Major,  
495 1987; Malet et al., 2005; Handwerger et al., 2013). However, the observed decreases in  
496 measured resistances (or increases in moisture content) to a 24 month low are consistent  
497 with landslide reactivation. The deviations between the two measures are likely to be

498 related to subsurface heterogeneity and sampling volume; the resistance measurements  
499 are sampling a significant volume of ground below a 14.25m array, whereas the  
500 piezometer is sampling a much smaller volume of ground, and is therefore likely to be  
501 more effected by very local heterogeneities and fissuring. Given that the piezometer is  
502 recording perched water levels within highly disturbed (and potentially mobile) earthflow  
503 material above permeable bedrock (Figures 2 and 9), it is probable that local fissure flow  
504 and drainage along failure planes caused the anomalous declining water levels observed  
505 from June to August 2012. This is supported by Gunn et al. (2013), who reported water  
506 and washed out fine material emerging from the slip surfaces at the earthflow toe's during  
507 periods of activation. Likewise this is also consistent with observations on other slow  
508 moving landslide where the role of fissuring, including the dynamic opening a closing of  
509 flow pathways, is seen to significantly influence ground water movement (e.g. Van Asch et  
510 al., 1999; Krzeminska et al., 2013; Bievre et al., 2012).

511 By September 2012 continuing rainfall resulted in increased piezometric levels once again  
512 as resistances briefly fall during September and October 2012. From November 2012  
513 piezometric levels are high (approaching ground level) and result in larger, deeper, more  
514 rapid earthflows activating. November onwards marks a time when electrodes are  
515 mobilising and is represented on temperature corrected plots as divergence and jumps in  
516 resistance values.

517 It should be reiterated at this point that the principal mechanism controlling movement of  
518 slow moving earthflows involves fluctuating pore pressures associated with changing  
519 groundwater levels. Increasing deformation rates are generally following a rise in these  
520 water levels that results in an increase in pore pressures, a concomitant loss of effective  
521 stress and thus a lower shearing resistance in these earthflows, particularly along  
522 bounding failure surfaces (Terzaghi, 1950). When associated with transitions from partial

523 to full saturation, instability is compounded by increased loading of the landslide mass  
524 (Varnes, 1978).

### 525 ***Summary of Findings***

526 Analysis of the results of this investigation reveals several new contributions to the  
527 understanding of landslide hydrogeological processes and resistance monitoring. These  
528 are stated in Table 4:

529

### 530 **Conclusions**

531 When applied to observe landslide processes, time-lapse electrical resistance makes use  
532 of its sensitivity to variation in moisture content in the subsurface. Resistance monitoring  
533 informs about the manner in which the slope responds to rainfall infiltration and soil  
534 moisture accumulation. Landslides respond to changing ground conditions, i.e. rising  
535 piezometric level or soil moisture content reaching plastic limits, which can bring about a  
536 change in internal physical properties, such as soil strength.

537 This investigation provides the longest term analysis of electrical resistance data for  
538 landslide monitoring that we are aware of in the literature, and extends the pioneering  
539 resistance monitoring work of Lebourg et al. (2010) and Palis et al. (2017) by providing  
540 spatial data from a grid of electrodes (rather than linear arrays) and detailed consideration  
541 of both the influence of temperature and electrode motion.

542 Compensating for temperature effects and accounting for electrode movements was  
543 shown to be essential in interpreting the geophysical events, as both of these processes  
544 can mask the moisture driven processes that the resistance monitoring system is designed  
545 to observe.

546 The spatial element of the monitoring described here was also shown to be highly  
547 significant. Landslides invariably display heterogeneous ground conditions with complex

548 hydraulic processes, which can be difficult to characterise using point sampling or linear  
549 monitoring arrays. The spatial geophysical monitoring presented here provides information  
550 on how the landslide system as whole responded to fissuring events, rainfall infiltration and  
551 changes in piezometric levels. This greatly assisted in identifying and characterising the  
552 different zones within the system (e.g. Figure 8 and Table 3).

553 Crucially, this study confirms the suitability of spatially distributed, temperature corrected  
554 resistance monitoring for landslide early warning by analysing multi-year variations in  
555 geophysical properties, which have permitted us to identify precursors to failure events.  
556 The sensitivity of this approach to changes in subsurface water distribution, and  
557 piezometric levels in particular, is key to its success because it can therefore observe the  
558 principal activation mechanism of slow-moving shallow earthflows i.e. the reduction in  
559 effective normal stress due to increasing pore water pressures.

560 The capability of observing increased moisture content with time provides a powerful tool  
561 to reveal hill slope hydrogeology, infiltration and landslide activation mechanisms. The  
562 technique highlights great potential to provide early warning of imminent slope failure when  
563 combined with additional a priori geotechnical data. Specifically, it provides a simple, fast,  
564 and non-invasive means of using resistance time-series data in order to monitor the  
565 moisture dynamics of landslide prone slopes thereby providing early warning of failure  
566 events.

567

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572



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801 and testing of displacement inversion to track electrode movements on 3-D electrical  
802 resistivity tomography monitoring grids. *Geophysical Journal International* 200: 1566-1581.

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<b>Sensor Location</b>	$T_{MAT}$ (°C)	$A$ (°C)	$d$ (m)	$\varphi$ (rad)	RMS (°C)
<b>T1</b>	9.81	14.62	2.073	-1.907	0.88
<b>T2</b>	9.99	15.62	1.968	-1.908	0.84
<b>T3</b>	10.25	16.49	2.697	-1.896	1.02
<b>All</b>	<b>10.03</b>	<b>15.54</b>	<b>2.264</b>	<b>-1.907</b>	<b>1.01</b>

Table 1. Fitted parameters for the temperature models using data from the individual (T1,T2,T3) and combined locations (All), Equation 1.

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<b>Environmental Inputs</b>	<b>Electrical Resistance Response (TC-Res)</b>	<b>Interpreted Ground Response</b>
Periods of higher than average rainfall (positive effective rainfall), e.g. October 2009 to April 2010. Results in rising piezometric levels.	TC-resistance values fall. e.g. Resistance ratio at Measurement Location 1 (ML1) reduces from 1.35 to 1.15.	Increase moisture content in the subsurface. Soil moisture accumulation manifests as both water level rise and very shallow soil moisture accumulation.
Periods of lower than average rainfall (negative effective rainfall), e.g. May 2010 to September 2010. Piezometric levels during these periods lowered by ~0.6 m (from 78.0 m ~77.4 m AoD).	TC-resistance values rise. e.g. resistance ratio at ML1 rises from 1.2 to 1.5.	Decrease moisture content in the subsurface. Soil moisture depletion manifest as both water table fall and drying of soil in the very shallow subsurface.
An intense period of rainfall (positive effective rainfall) during and after a period of low rainfall (negative effective rainfall). e.g. September 2010. Rapid piezometric level rise concurrent with rainfall event.	TC-res ratios fall rapidly. e.g. resistance ratio at ML1 falls from 1.5 to 1.25.	Annealing of desiccation cracking in response to rainfall, as soil moisture content increases in very shallow subsurface. Some phreatic water reaches the water table.
Minor, low intensity rainfall event, as occurred June to July 2011 and do not create a piezometric level rise.	Small rises and falls in resistance ratio between 1.4 to 1.3.	Soil moisture accumulation in the very shallow subsurface occurring. Insufficient quantity of rain water to reach the water table.
Full range of seasonal weather.	Measurement location (ML) 1 results reveal a greater temperature-corrected resistance ratio range (0.6) than the other three measurement locations, with ML4 showing the most subtle variation (0.25).	The thinnest region of the earthflows (nearest the toe) respond more extremely to environmental conditions than regions where the earthflow is thicker (closer to the main landslide body) and composed of several flows.
Earthflow activation caused by high and sustained piezometric levels.	November 2012 until the end of the monitoring period, the resistances recorded using the four measurement locations diverge markedly, with ML1 and ML2 displaying extreme increases and decreases in resistance respectively.	This occurs during a period of earthflow activity and is attributed to the resulting displacement of individual electrodes, fissuring in the near surface, and localised accumulations of moisture – all of which would influence electrical resistance measurements.

	It should be noted that this trend is not observed in the 2008 activation (within Figure 4) because a separate earthflow lobe was active during this activation event.	
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807 Table 2. Summary of General Monitoring Results and displayed in Figure 4.

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	<b>Landslide Zone</b>	<b>Electrical Resistance Response</b>	<b>Interpreted Ground Response Leading to Earthflow Activation</b>
<b>1</b>	<b>Backscarp</b>	Resistances display only small variations. However, resistances are slightly lower than baseline August to January and slightly higher than baseline between February and July.	Small moisture content variations due to soil moisture retention and lack of considerable fissuring. Potential supply of moisture from the Dogger Formation, a minor aquifer in the area.
<b>2</b>	<b>Head &amp; Sag Pond</b>	Resistances are equal to or slightly higher than baseline.	Region retains soil moisture throughout period due to slight back-tilt of beds, shading by backscarp and reed-beds.
<b>3</b>	<b>Upper Body</b>	Small resistance changes throughout the year. July to February is equal to baseline. March to June is higher than baseline.	Water table level not greatly variable in lead up to earthflow activation. Little evidence for soil surficial cracking, any cracks present are quickly annealed following rainfall.
<b>4</b>	<b>Lower Body &amp; Flow Lobes</b>	Response is more sudden and extreme. Rapid Resistance Change; uniformly higher 06/2011 and lower 05/2012.	Flow lobes are composed of a series of 0.5-1.0 m earthflow deposits. These permit easy fluid through-flow during wet periods and cracking when dry. Ground responds rapidly to negative effective rainfall events by dessication and lowering of perched water table within WMF.
		Winter 2011 (09/2011-01/2012) variable response, small regions of higher and lower resistance. When Upper Body of landslide system is near equal to resistance baseline the lower body and flow lobes show lower resistances.	Run-off occurring from Upper Body to penetrate the Lower Body and Flow Lobes.
		Lowest resistances measured occur during the month preceding earthflow activation. Very low resistances at earthflow toes.	Rising piezometric level within earthflow zone during the lead up to earthflow activation. Pore pressures high enough to permit earthflow activation. Groundwater flow occurring along slip surfaces and out at earthflow toe.
<b>5</b>	<b>Between Flow Lobes</b>	Markedly lower resistances than baseline from 05/2012 onwards and during the three month period preceding activation. Resistances remain low while earthflows are active.	Regional water table within Staithes Sandstone Formation rising in response to positive effective rainfall.

Table 3. Landslide system and hillslope zones and associated electrical resistance and interpreted ground response. The locations of the landslide zones are shown in Figure 8.

<b>Hydrogeological Landslide Processes</b>	
1	Significant seasonal and spatial variations in subsurface response to rainfall input can be seen across the different zones of the landslide, including in the lead-up to earthflow activation.
2	The intensity and distribution of desiccation cracking (and conversely, annealing) exerts a significant influence on slope hydrogeological dynamics.
3	The process of shallow slow-moving earthflow activation may be more complex than initially thought. Resistance monitoring identified the presence of fissure flow along the landslide slip surfaces.
4	Temperature-corrected resistance monitoring is sensitive to a key shallow earthflow activation mechanism, i.e. the rise in pore water pressure as a result of high and sustained piezometric levels.

<b>Temperature-corrected Resistance Monitoring</b>	
1	Temperature corrected resistance measurements are sensitive to both shallow rainfall driven moisture dynamics and piezometric level changes.
2	Upon landslide activation, electrode displacement causes divergence or convergence of resistance values.
3	Shallow resistance measurements are highly sensitive to seasonal temperature variations. These temperature variations act to mask subtle moisture content related resistance changes.

831 Table 4. Summary of the results of temperature-corrected resistance monitoring of an  
832 intermittently active complex landslide system.

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845 **List of Figures**

846 Figure 1. (a) Geological map of the study area and inset large scale map, (b) Engineering  
847 geomorphological map of study site. The depths of the sensors within the temperature  
848 sensors arrays are as follows: northern array - 0.1, 1.0, 2.5, 5.35 m; middle array - 0.1, 0.7,  
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850 National Grid. Figure is amended from Merritt et al (2014). BGS © NERC. Contains  
851 Ordnance Survey data © Crown Copyright and database rights 2016.

852 Figure 2. Cross sections of the complex landslide system at the Hollin Hill Test Site, (a)  
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854 (2015b)); (b) Detailed ground model of earthflow regions (western flow region  
855 represented). Ground model represented as Figure 2a extends from beyond the landslide  
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857 Figure 3. Quantifying temperature correction of transfer resistance datasets and seasonal  
858 air temperature variation. The example measurement shown here is located 19m (x-axis),  
859 95m (y-axis), within sliding region.

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866 2003), which accounts for the effects of evapotranspiration.

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868 Rainfall data is from rain gauge installed at Test Site. The dashed line shows the long-term  
869 (30 year) average annual rainfall in the area of the test site.

870 Figure 6. Geoelectrical monitoring of landslide deposit results for the 14 months leading to  
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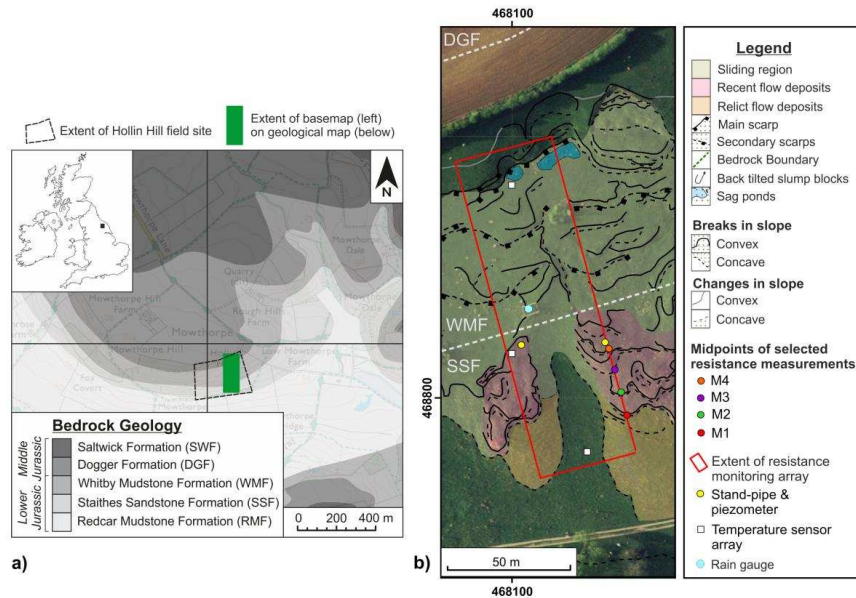


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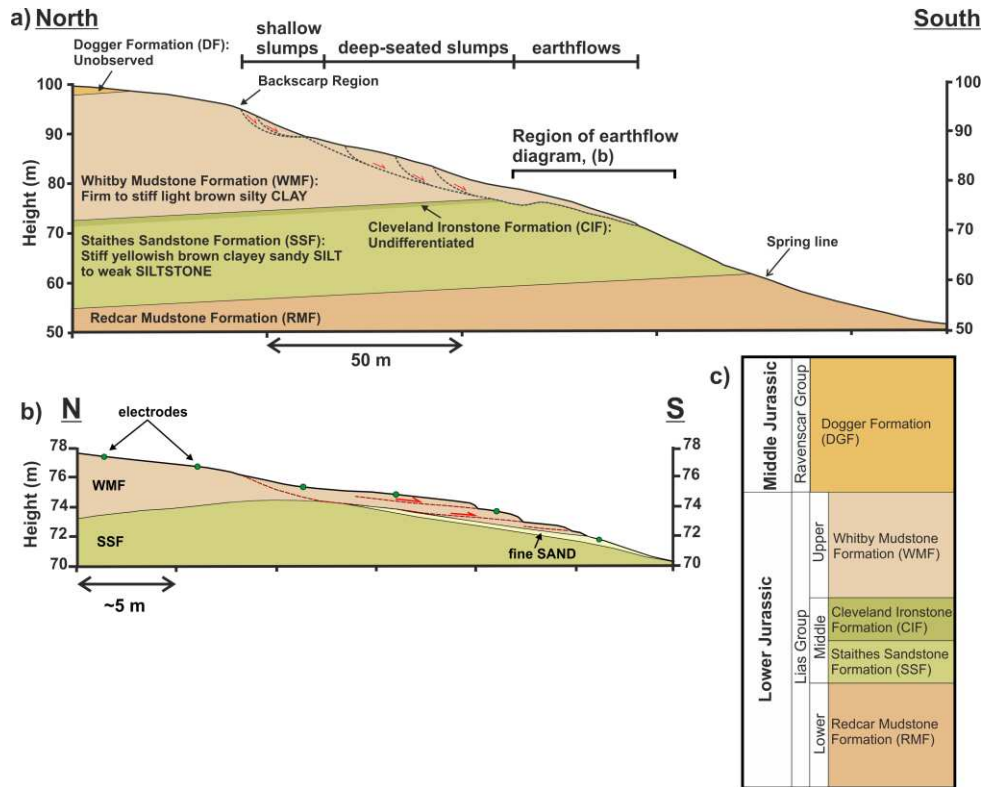


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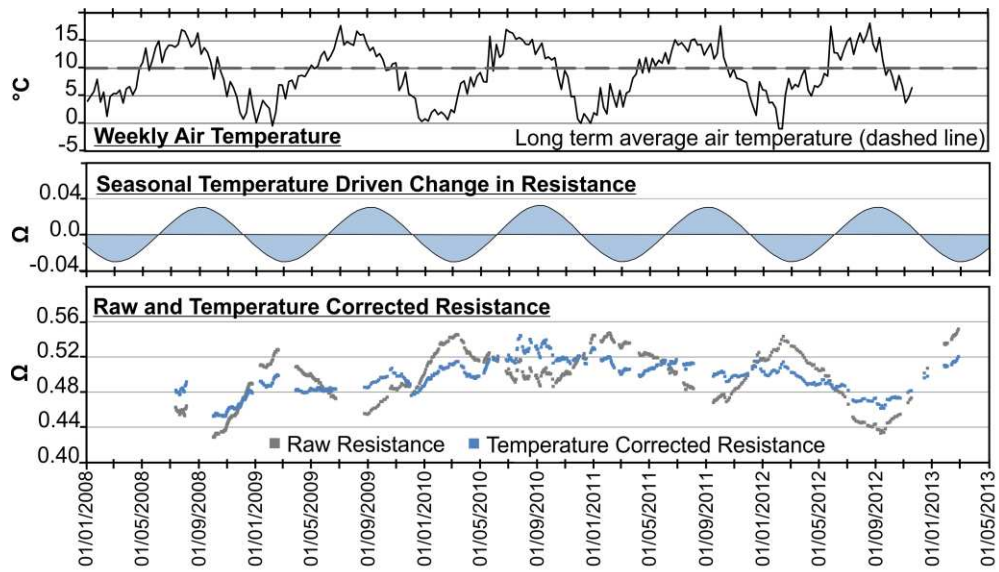


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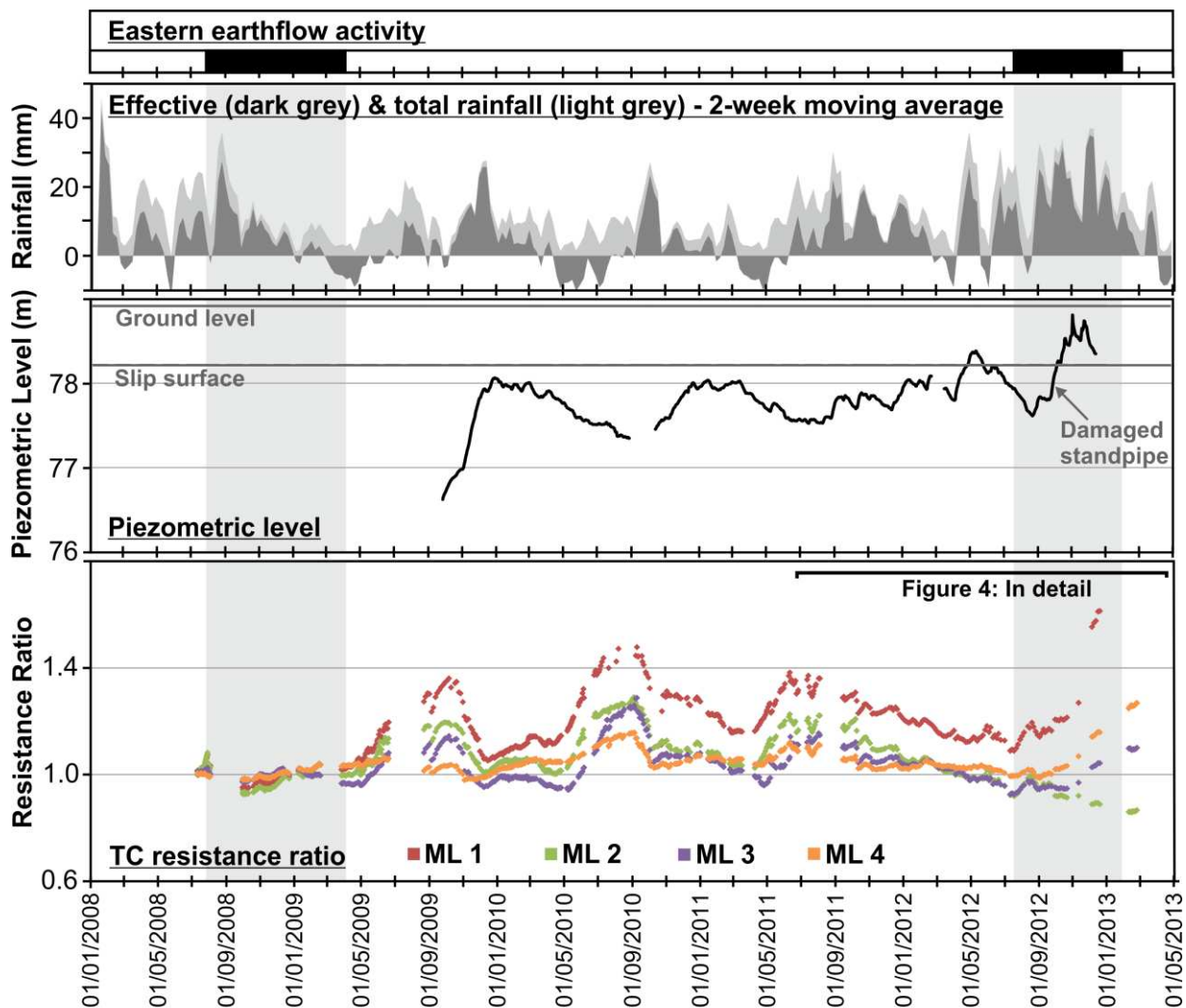


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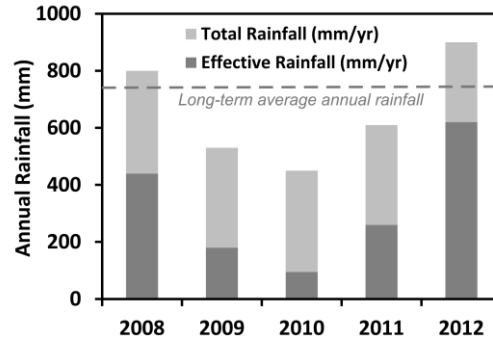


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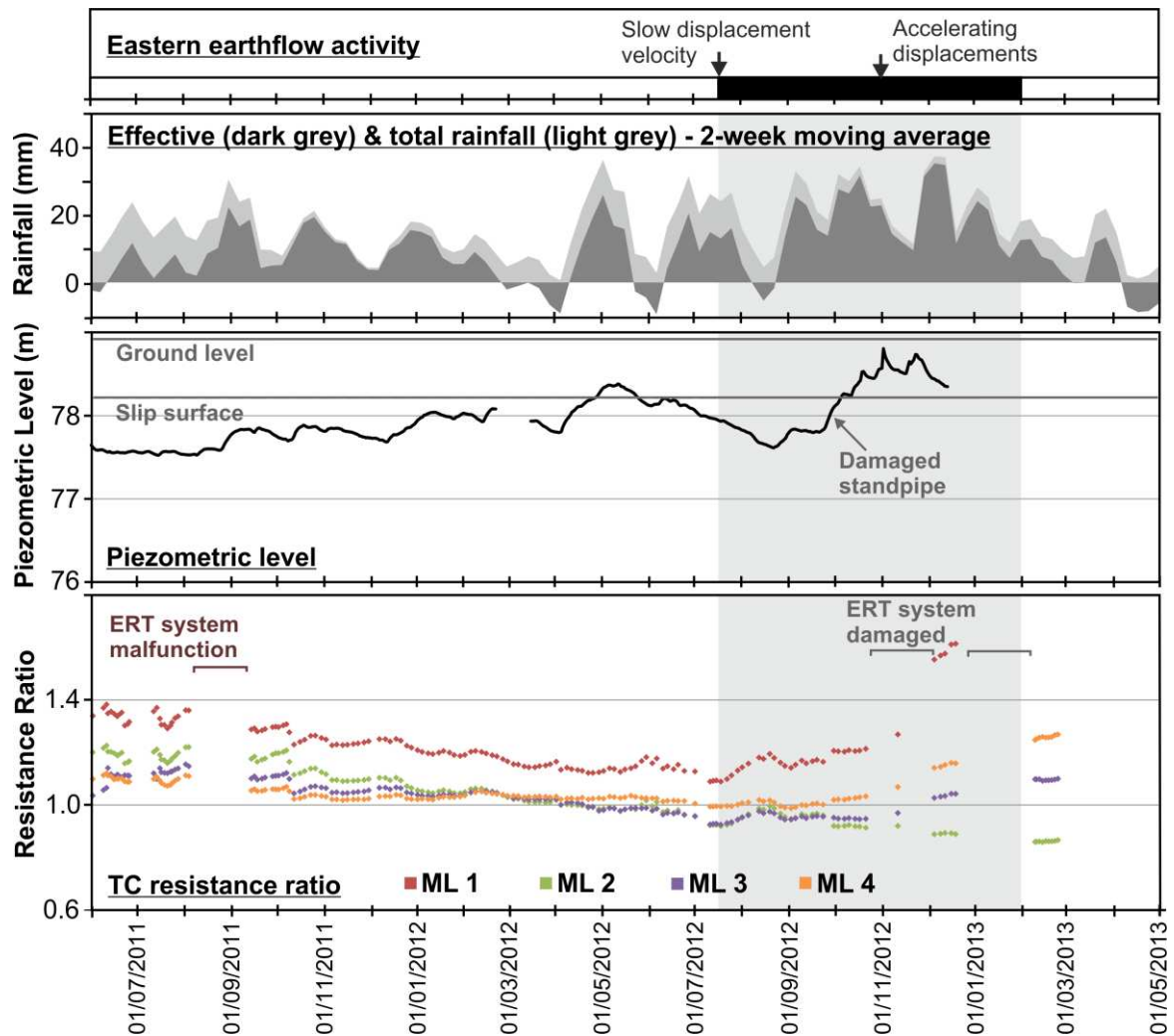
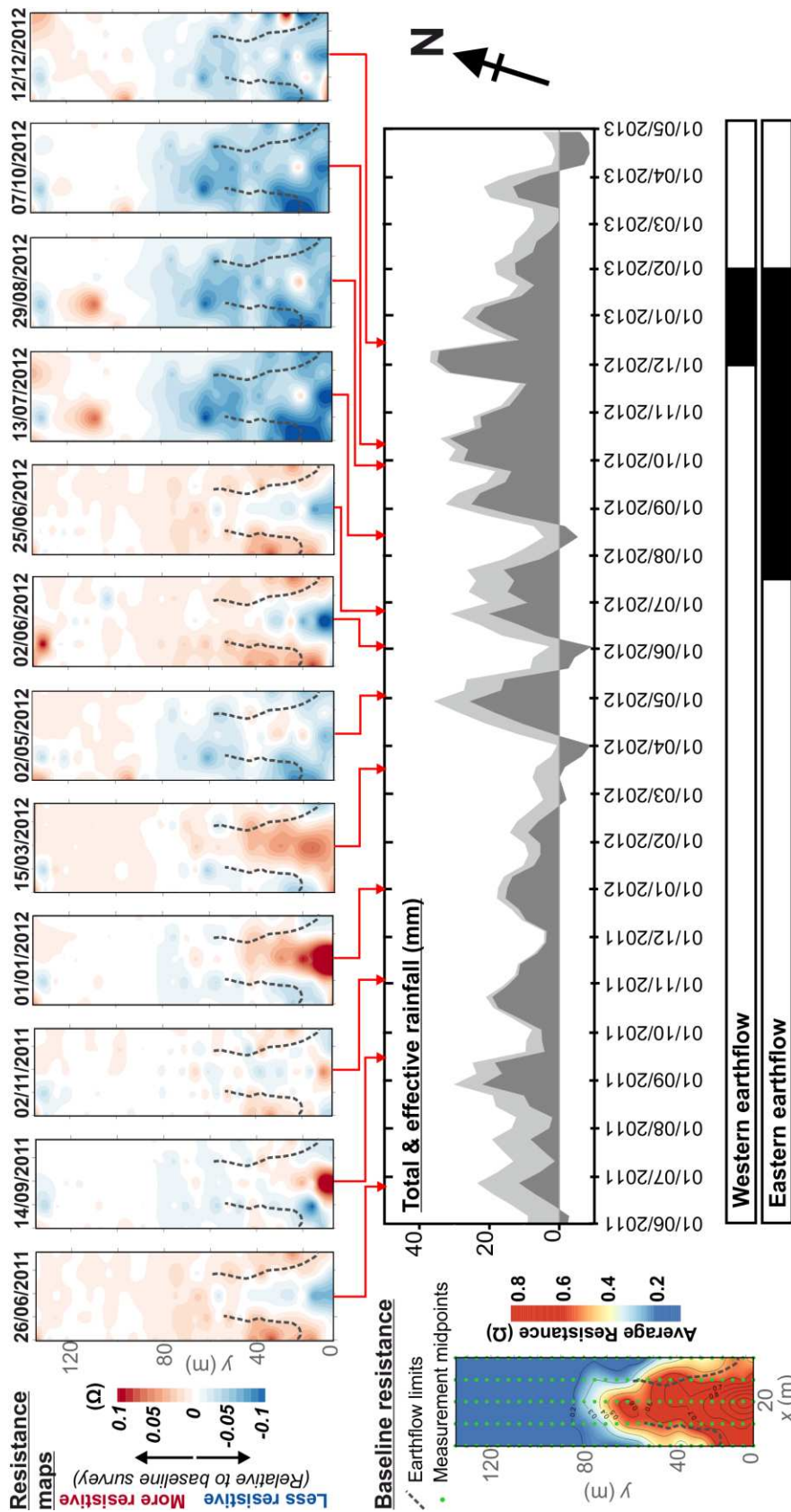
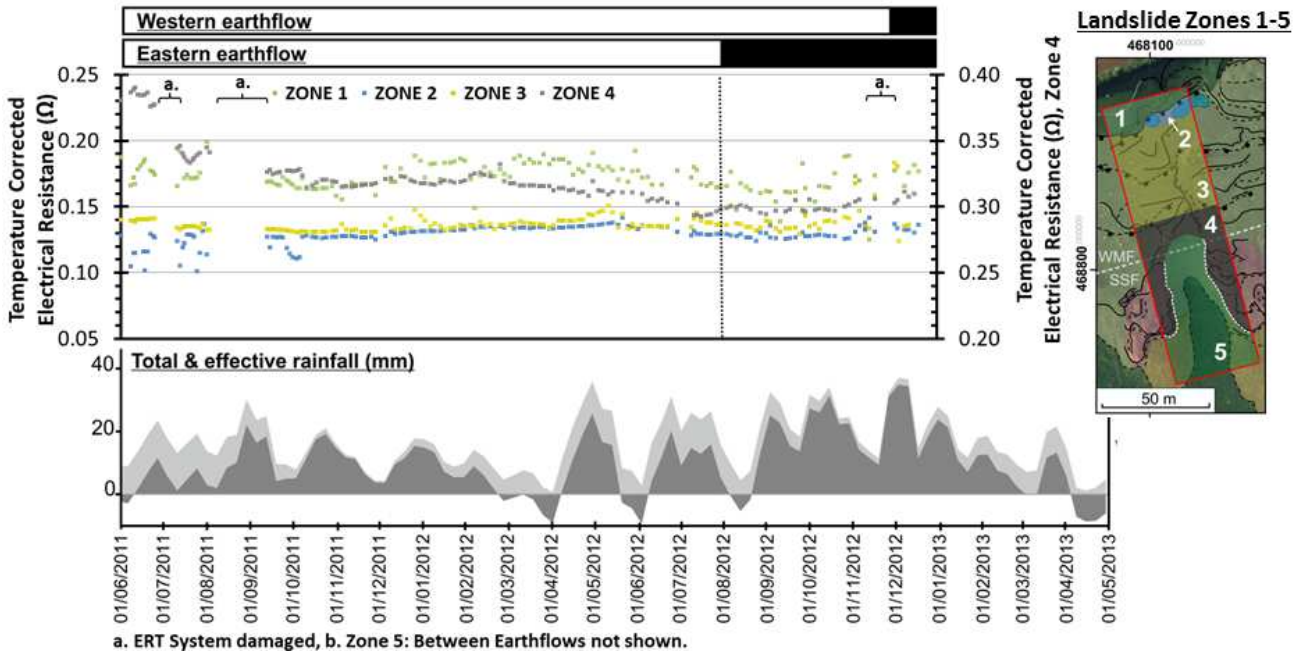


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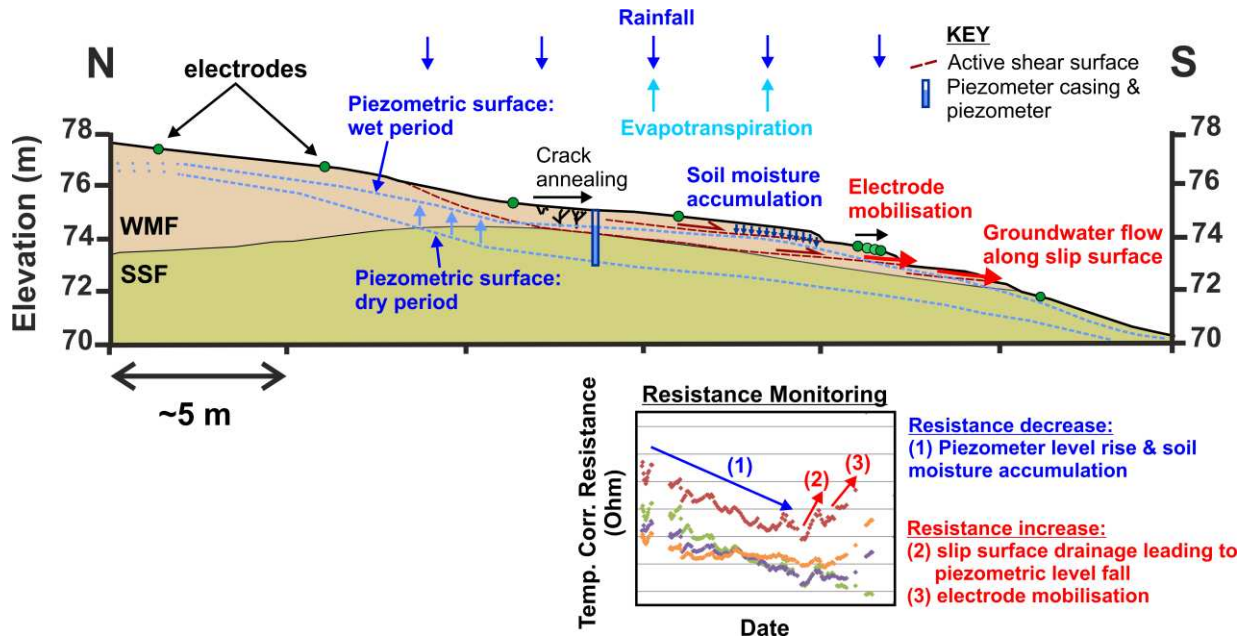


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