Monitoring soil-water and displacement conditions leading to landslide occurrence in partially saturated clays

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1. Introduction

The soil physical properties in the unsaturated zone affect a large variety of processes determining the occurrence of a shallow landslide. Landslides depend on a combination of variables such as slope, rainfall amount, rainfall intensity, soil mantle thickness, soil state parameters, subsurface hydrology and vegetation (Dietrich et al., 2001; Rybar et al., 2002). Among these factors, unsaturated soil hydrology has been shown to be a key factor for shallow landslide formation and dynamics (Lu and Godt, 2008; Tsai and Chen, 2010; Tsai, 2011).

Several authors (e.g., Montgomery et al., 1997; Lu and Godt, 2008) maintain that the most important cause for shallow landslides is the decrease of matric suction after a rainstorm and the development of positive pressures above the water-table. In particular, the soil shear strength decreases non-linearly with increasing soil matric suction, therefore when suction becomes less negative and the soil approaches saturation, the soil becomes more susceptible to failure (Lim et al., 1996; Vanapalli et al., 1996).

The interactions between mechanical and hydrological processes are more complex in clay soils because of the different physico-chemical processes affecting shear strength and soil stability. Clay soils experience additional processes such as softening, weathering, slaking and soil weakening caused by cycles of wetting–drying or freezing–thawing. Hight et al. (2002) and Takahashi et al. (2005) proposed that even simple swelling can provoke weakening of bonded clays, because accumulation of plastic strains determines a loss of soil shear strength.

For practical purposes, such as mapping of shallow landslide susceptibility, it is advantageous to apply distributed models such as SINMAP (Pack et al., 1998), SHALSTAB (Dietrich et al., 2001), TRIGRS (Baum et al., 2008) or SLIP (Montrasio and Valentino, 2008), which are based on the infinite slope model. The infinite slope model often assumes that the failure plane is coincident with the soil–bedrock interface. However, in deep clay deposits the soil–bedrock interface may be tens
or hundreds of meters below the soil surface making the model assumptions questionable and requiring use of other models for slope stability, or otherwise requiring an experimental analysis for identification of the depth of the slip plane. Nevertheless, the occurrence of shallow landslides is very common on clay deposits, such as the ones of the Apennine mountain range in Italy (Simoni et al., 2004; Montrasio et al., 2009), therefore raising questions about landslide triggering factors and how slope stability models can be successfully applied to clay soils.

Other factors, such as a detailed description of partially saturated conditions in the infinite slope model, are only recently coming under examination (Baum et al., 2010). For example, Lu and Godt (2008) described the stability analysis of an infinite slope, taking into account the suction stress concept proposed by Lu and Likos (2006). Lu and Likos (2006) introduced the concept of suction stress to describe the state of stress and the mechanical behavior of unsaturated soils, in the context of a generalized effective stress principle. According to the authors, the macroscopic stress, called “suction stress,” combines physico-chemical forces, cementation forces, surface tension forces and the force arising from negative pore-water pressure, allowing for the capture of complex mechanisms involved in the mechanical behavior of unsaturated soils.

To identify the depth of the failure plane as well as other landslides features, and to test the validity of a specific modeling approach, there must be continuous monitoring of unsaturated soil properties and slope stability. Several techniques have been proposed to monitor landslides and a large variety of experiments have been performed to assess their reliability (Rahardjo et al., 2011). Among the available techniques, the use of time domain reflectometry (TDR) for mass deformation studies is increasingly applied for landslides studies. O’Connor and Dowding (1999) presented a description of the technique and several works have been published describing applications in different settings (Dowding and Huang, 1994; Arenson et al., 2002).

The TDR technique is more commonly used for measuring soil water content and electrical conductivity (Topp et al., 1980; Robinson et al., 2003). However the two applications can be combined within a single experimental station, using TDR for both mass deformation and soil-water content measurement.

The aim of this research was to test Lu and Godt’s (2008) slope stability model for clay soils. The model was tested by implementing a case study where soil hydrological properties and soil deformations were measured to investigate: (a) the effect of unsaturated soil hydrological processes on shallow landslide formation, (b) the processes involved in the formation of the failure plane, (c) the soil mantle thickness involved in shallow landslide formation on clayey soils, and (d) the validity of the suction stress concept for describing landslide occurrence on clayey soils.

Therefore, the new contribution of this paper is the application of the model to clay soils, by detailed monitoring of soil water content and soil suction during the occurrence of a landslide.

2. Theory: slope-stability analysis under unsaturated conditions

2.1. Computation of suction stress

Lu and Godt (2008) applied the concept of suction stress to a simplified slope stability analysis. They pointed out two mechanisms causing suction stress in soils: inter-particle physico-chemical forces and inter-particle capillary forces. When the soil approaches saturation, the suction stress is reduced, and this phenomenon is the triggering mechanism for landslides. For certain kinds of soil, suction stress depends on the degree of saturation, and matric suction, through the suction stress characteristic curve (SSCC). These factors can be directly linked with the well known concept of the soil water characteristic curve (SWCC) for unsaturated soils (Lu and Likos, 2006). In particular, Lu et al. (2010) state that suction stress can be determined based on the soil matric suction:

$$\sigma^s = -\frac{\theta - \theta_s}{\theta_s} (u_a - u_w) = -S_e (u_a - u_w)$$

(1)

where $\theta$ is the volumetric water content, $\theta_s$ is the residual volumetric water content, $\theta_i$ is the saturated volumetric water content, $S_e$ is the degree of saturation, $u_a$ is the pore air pressure and $u_w$ is the pore water pressure (or soil suction), when $u_a$ is zero.

Alternatively, they also proposed computing the suction stress by applying the relationship between the degree of saturation and matric suction, after the normalized degree of saturation has been defined through van Genuchten’s (1980) SWCC model. In this case, the expression of suction stress is the following:

$$\sigma^s = -\frac{S_e}{\alpha} \left( \frac{S_e}{n} \right)^{n-1} u_e$$

(2)

where $n$ and $\alpha$ are fitting parameters of unsaturated soil properties (van Genuchten, 1980). The value of $S_e$ in Eq. (2) can be obtained on the basis of field measurements of volumetric water content, while $n$ and $\alpha$ are determined as SWCC fitting parameters. The advantage of the second approach is that the suction stress can be obtained without direct measurement of matric suction, but rather from measurement of the degree of saturation and SWCC parameterization.

The starting point of our analysis is the determination of the SWCC for the different soil layers involved, based on field measurements of matric suction and water content to compute the suction stress. In particular, for each investigated sample site depth, we evaluated suction stresses through Eq. (2), on the basis of field daily measurements of degree of saturation and from van Genuchten parameters, taking into account the effects of the unsaturated conditions on the slope stability analysis.

2.2. Model for the slope-stability analysis

With regard to the limit equilibrium analysis of the slope, we employed a model that takes into account the trend of suction stress in time (Lu and Godt, 2008; Godt et al., 2009). The model is based on the computation of the factor of safety ($F_s$), defined as the ratio between the shear strength and the shear stress. It was applied on a one-dimensional infinite slope as function of vertical depth, $z$, below the ground level. The model employs the infinite-slope scheme, which can be successfully applied in situations where the failure plane is approximately parallel to the slope surface, with a maximum depth of approximately 1.5–2 m, as in the case analyzed in this paper.

The equation of $F_s$ is given by:

$$F_s(z) = \frac{\tan \phi' + \frac{2c'}{\gamma' \sin \phi'}}{\frac{\phi'}{\gamma' \sin \phi'} + \frac{\sigma^s}{\gamma'} (\tan \beta + \cot \beta) \tan \phi'}$$

(3)

where $\phi'$ is the friction angle of the soil, $c'$ is the effective cohesion, $\beta$ is the slope angle, $\gamma$ is the soil unit weight, $\sigma^s$ is the computed suction stress. As stated by Lu and Godt (2008), the third term of Eq. (3) represents the contribution given to shear strength by suction stress. In our stability analysis, daily value of $F_s$ has been computed by taking into account the computed value of suction stress at different depths. In particular, suction stress has been evaluated on the basis of field measurements of both the matric suction and the volumetric water content.

The instability condition corresponds to a value of $F_s$ equal to one. $F_s$ approaches one when the suction stress is greatly reduced, i.e. when the soil is near saturation. In this condition, $F_s$ is reduced drastically, creating unstable soil conditions.
2.3. Extension of soil water content information by numerical modeling

The experimental analysis was supported by numerical water flow modeling utilized solely to obtain estimates of water content at further depths. Indeed, the experiments provided soil water content and suction measurements at three depths: 0.2, 0.4 and 0.8 m. However, the total soil depth considered for the slope stability model was 1.4 m, a value selected on the basis of the observed failure surface. However, detailed estimated values of soil water content also at a depth of 1.4 m was needed, so as to retrieve a representative value across the whole profile. To compute soil water content at a 1.4 m depth, a numerical model was employed, based on the solution of Richard’s equation. The model is called CRITERIA and it is described in details by Bittelli et al. (2010). The CRITERIA model is a physically-based, 1, 2 or 3D model (the domains of simulation can be selected by the user) that simulates soil water dynamics by computing the soil water budget terms (infiltration, redistribution, runoff and evapotranspiration). The model was first calibrated against experimental data for the three measured depths, and then soil water content was simulated at a depth of 1.4 m. Input parameters required by the model were: a soil map, meteorological measurements (hourly shortwave radiation, relative humidity, air temperature, wind speed, and precipitation), soil physical properties (bulk density, texture, soil water retention curves, and vertical hydraulic conductivity) and crop parameters. All the input data were collected at the site during the experiment on an hourly basis.

Model’s performance was evaluated with the Nash and Sutcliffe (1970) statistical index (NS):

\[
NS = 1 - \frac{\sum (\theta_m - \theta_o)^2}{\sum (\theta_o - \bar{\theta}_o)^2}
\]

where \(\theta_m\) is the observed water content, \(\theta_o\) is the modeled water content, \(\bar{\theta}_o\) is the average of the observed water content, the superscript \(t\) is time and \(T\) is the total time for the time series. NS efficiencies can range from \(-\infty\) to 1, when \(NS = 1\) there is a perfect match between modeled and observed data. When \(NS = 0\) the model predictions are as accurate as the mean of the observed data, whereas for \(NS < 0\) the observed mean is a better predictor than the model since the residual variance is greater than the data variance.

3. Site description

The study site is located in the University of Bologna experimental farm (44°28′N, 11°28′E), which is situated in the Centonara watershed, southeast of Bologna, Italy (Fig. 1). The elevation at the study site is 200 m a.s.l. The geology is a fresh alluvial deposition of the upper Pleistocene, with undifferentiated clay moraine and yellow sand (Farabegoli et al., 1994).

Shallow soils developed on this parent material are inceptisols and vertisols. The area is geologically complex since various formations are present including alluvial deposits and yellow sands from the Pleistocene, the Bismantova formation and a chaotic clay complex (made by a block in matrix sequence, and scaly varicolored clay formation). The clay materials have been defined argile scaglione, or “scaly clays” (Pini, 1999). Shallow landslides are very frequent in this geological setting, and they usually occur on predominantly clayey soils with slopes ranging between 10° and 30°.

The test site was selected on a slope where different types of mass movements had occurred in the past. For this reason, the probability of occurrence of a new shallow landslide was relatively high. Some of these movements, in fact, had involved shallow soil layers of quaternary detritus. This is confirmed by the observation of the morphology of the site. In particular, the study site map in Fig. 2 shows the delineation of the landslide that occurred during the field experiment. In the figure a 2-m digital elevation model (DEM) shows how the trend of the contour lines of the study area is typical of a hillside subject to a process of retrogressive landslides. It is clear that landform evolution has been dominated by landslides, as frequently happens in the widespread earth flow terrains of the Emilia Romagna Apennines (Bertolini et al., 2005; Crozier, 2010). In such a geomorphological system, a surface of shallow failure occurs between the ridge and the earth flow head (Crozier, 2010).

4. Monitoring observations

The experimental station was installed on a slope (14°) with a shallow soil characterized by high clay content (60%). The slope faces the Centonara stream, which collects the water of the Centonara catchment.

A pedological classification was performed at the site, and the soil was classified as Aquic Chromic Haploxerert, smectitic, calcareous, and masic (USDA Classification). Five main horizons were identified: A1 (0–12 cm), A2 (12–30 cm), B (30–70 cm), Bw (70–90 cm) and Bg (80–170 cm). The horizons A1 and A2 were characterized by a light brown plastic clay with a polyhedral structure and many fine roots, while horizons B, Bw and Bg were characterized by dark gray plastic clay with a polyhedral structure and high packing density. Within the Bg horizon (at a depth comprised between 1.45 and 1.65 m) there was a layer of compacted material, with strong gleying, revealing stagnant saturated and reducing conditions.

A geotechnical classification was also performed, and the soil down to 1.7 m depth was classified as “inorganic clay with high plasticity” according to the USCS system (Atterberg limits: liquid limit of 68–80%, plastic limit of 27–33%, Tosi, 2007).

Determination of lithology, soil physical and hydrological properties was performed through auger, core drilling and soil pit excavation. Drained shear strength parameters had been determined by a previous study (Tosi, 2007) through direct shear tests on reconstituted samples, under normal stress comparable with those characteristic of the shallow layer at the site. The investigated soil presented a peak shear strength angle equal to 18° and a residual angle of 12°, a peak effective cohesion equal to 4.8 kPa and a residual effective cohesion equal to 0 kPa. The presence of the compacted layer at a depth of more than 1.45 m was confirmed by measurements of bulk density (Table 1).

The clay fraction is characterized mostly by smectite and vermiculite as determined by mineralogical analysis performed by employing X-ray diffraction with CuKa radiation (Philips XRC 3100, Philips Analytical Inc., Mahwah, NJ). The high clay content and the mineralogical composition of the clay fraction, causes pronounced swelling–shrinkin, with large spatially-distributed cracks facilitating preferential flow and a decrease in the shear strength along desiccation cracks. Table 1 shows the soil properties at the site.

The water retention curve was obtained from field and laboratory data (as described in the next sections) so as to derive the van Genuchten parameters through non-linear fitting (Marquardt, 1963). The non-linear fitting algorithm was written by the authors.

As shown in Table 1, the soil has a uniform profile in terms of texture and hydraulic properties (at least until the compacted layer at 1.45 m depth, where the bulk density increases). Therefore we derived one SWCC for the soil profile down to 1.4 m, and we used the same value of \(\alpha\) and \(n\), to compute the suction stress and for simulation of vertical water flow, as will be explained in the following sections.

For the installation of sensors, two soil pits were excavated, 4 m apart, perpendicularly to the contour lines, therefore along the line of maximum slope. Fig. 3 shows a schematic of the station. The site was equipped with an integrated experimental station, which automatically recorded soil water content and soil suction with a data logger and transmitted the data via satellite to a remote computer. The data regarding the formation of soil failure planes were periodically
measured. In this study the collected data are reported from October, 2004 to October, 2006.

4.1. Measurement and localization of soil failure planes

Slope stability measurements to detect potential failure planes were performed using TDR-based coaxial cables. The system is based on the insertion of a coaxial cable into a soil borehole; the cable is then grouted by compliant cement characterized by shear strength similar to that of the surrounding soil. Fig. 4 shows the different steps for the installation of a coaxial cable into a borehole and grouting. The grout for stiff clay soils was a mixture of Portland cement, bentonite and water with a ratio by weight of 1.0, 3.0 and 2.5, respectively (Mikkelsen, 2002).

When soil movement is sufficient to fracture the grout, cable deformation occurs and the change in cable impedance is detected by a TDR cable-tester. Since the cable tester detects the distance of the fracture along the cable, this information is then used to obtain the failure plane depth. In this study, 12 grouted cables were positioned in different locations in the catchment and grouted into 4-m deep boreholes. Monitoring was manually performed by periodically testing the cable with a cable tester (Cable tester 1502C, Tektronix Inc.).

4.2. Monitoring soil-water response

Soil water content was measured with a TDR system (TDR 100, Campbell Sci., Inc.), equipped with a multiplexer (SDMX50, Campbell Sci., Inc.). Six TDR probes (three-rod probes, model CS610, Campbell Sci., Inc.) were installed at 0.2, 0.4 and 0.8 m depths, for each profile (Fig. 3). We could not install sensors at greater depths because of the slope, high soil clay content and high soil compaction, which made sensor installation difficult, especially because the installation was designed to minimize soil disturbance. Multiple soil excavations down to 1.5 m and large enough to allow probe installation, would have required a significant disturbance of the slope, with potential modification of soil mechanical properties along the slope.

The dielectric permittivity was converted to water content by using the Topp et al. (1980) equation. Correction for soil water content overestimation due to the conductive clay material was performed, as described in Bittelli et al. (2008). Measurements were made in one-hour intervals. Periodic measurements of soil bulk density and gravimetric water content were performed for calibration and test.

Soil matric suction was measured with three heat dissipation sensors (HD) (model HD229, Campbell Sci., Logan, UT) for each pit. The HD sensors measure soil matric suction in the range of $-10^1$ to $-10^5$ kPa (Flint et al., 2002), providing a wide range of matric suction measurements. As indicated by Flint et al. (2002), the sensors were first calibrated in the laboratory for measurement at both full saturation, as well as completely dry sensors. For full saturation the HD sensor was immersed into a beaker with de-ionized water and positioned into a vacuum chamber for complete removal of air bubbles. For the dry point, the sensors were positioned in an oven at 100 °C for 24 h, and measurements were performed after oven drying. Measurement of differential temperatures was then performed for full saturation and dry points (Eq. (2) in Flint et al., 2002).

The range of matric suction between 0 and $-10$ kPa is not provided by the HD sensors because the upper range is $-10$ kPa, which is the air entry pressure of the ceramic cups. However, information in this range, very close to saturation, is important for landslide studies.
Therefore, suction values between 0 and −10 kPa were derived by utilizing the soil water content data obtained from the TDR, and by inverting the van Genuchten (1980) equation.

### 4.3. Measuring the soil water characteristic curve

It is not known a priori if the SWCC is well constrained in the range 0 to −10 kPa, such that it is possible to derive matric suction from the degree of saturation. Therefore, to compare the data inferred from the TDR with measured suction, we also measured the SWCC in the laboratory.

Pressure plates are among the most commonly used techniques to measure SWCC. However, several authors (Gee et al., 2002; Bittelli and Flury, 2009) reported that low plate and soil conductance, lack of plate–soil contact, and soil dispersion, make this method often unreliable at low water potentials. In particular, Bittelli and Flury (2009) indicated that soil water retention curves determined from pressure plates may be in error at potentials less than −200 kPa.

Based on these results, two different sets of measurements were performed: (1) Stackman tables (Stackman et al., 1969) and pressure plate apparatus (Richards, 1948) measurements, and (2) dew point measurements. Stackman plates and pressure plates were used from 0 to −200 kPa, while dew point measurements (Scanlon et al., 2002), were employed below −200 kPa.

During the installation of the TDR and HD sensors, undisturbed soil samples were collected at 0.2, 0.4 and 0.8 m depths. The soil samples were placed on a pressure plate apparatus following the procedures described by Klute (1986). The soil samples were wetted from below with 0.1 M CaSO₄, and allowed to saturate overnight. The samples were equilibrated in the Stackman plates and pressure plate apparatus at −0.5, −1, −2, −3, −5, −10, −30, −50, −100, and −200 kPa. For each soil sample and water potential, three replicates were used. Total soil water potential at ≤200 kPa was measured independently using a commercial dew point potential meter (WP4T, Decagon Devices, Pullman, WA). The WP4T was calibrated with 0.1 M KCl salt solution. The soil samples were equilibrated at various potentials by wetting the soil with deionized water and letting the water evaporate. The soil samples were then placed into the WP4T, and the water potential was determined. The corresponding soil water contents were measured by oven drying immediately after removal of the samples from the WP4T. For each soil sample and water potential, three replicates were used.

### 4.4. Data acquisition

The field data were collected by a CR23X datalogger (Campbell Sci., Inc.) and transmitted to the Orbcocomm satellite transmitter (model IX G7100, Orbcocomm Data Communicator). For details about the Orbcocomm protocols for data transmission see [http://www.orbcocomm.com/](http://www.orbcocomm.com/). Hourly data were sent in real time by the satellite to an FTP address at the Department of AgroEnvironmental Science and Technology, University of Bologna.

### 5. Results

#### 5.1. Soil water characteristic curve

Fig. 5 shows the experimental and fitted SWCC for three depths (a to c), indicating the correspondence of the measured SWCC and the SWCC derived from inversion of the van Genuchten (1980) equation between 0 and −10 kPa. The good agreement between the field and laboratory experimental data between 0 and −10 kPa demonstrated that the inversion of the van Genuchten equation, used to obtain the missing soil water potential data close to saturation, was a valid approach.

<table>
<thead>
<tr>
<th>Depth range [m]</th>
<th>0–0.25</th>
<th>0.25–0.45</th>
<th>0.45–1.45</th>
<th>1.45–1.65</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Sand [% weight]</strong></td>
<td>16</td>
<td>17</td>
<td>15</td>
<td>16</td>
</tr>
<tr>
<td><strong>Silt [% weight]</strong></td>
<td>24</td>
<td>23</td>
<td>27</td>
<td>38</td>
</tr>
<tr>
<td><strong>Clay [% weight]</strong></td>
<td>60</td>
<td>60</td>
<td>58</td>
<td>46</td>
</tr>
<tr>
<td><strong>Bulk density [kg m⁻³]</strong></td>
<td>1120</td>
<td>1200</td>
<td>1230</td>
<td>1420</td>
</tr>
<tr>
<td><strong>Kᵣ [cm day⁻¹]</strong></td>
<td>4</td>
<td>4</td>
<td>3.8</td>
<td>1.5</td>
</tr>
<tr>
<td><strong>θᵣ [m³ m⁻³]</strong></td>
<td>0.055</td>
<td>0.54</td>
<td>0.54</td>
<td>0.46</td>
</tr>
<tr>
<td><strong>θₛ [m³ m⁻³]</strong></td>
<td>0.07</td>
<td>0.07</td>
<td>0.07</td>
<td>0.05</td>
</tr>
<tr>
<td><strong>α [kPa⁻¹]</strong></td>
<td>0.095</td>
<td>0.095</td>
<td>0.095</td>
<td>0.05</td>
</tr>
<tr>
<td><strong>n [−]</strong></td>
<td>1.3</td>
<td>1.3</td>
<td>1.3</td>
<td>1.1</td>
</tr>
</tbody>
</table>

Fig. 2. Digital elevation model (DEM) at 2 m resolution of the study area, with contour lines and delineation of a process of retrogressive landslides.
Fig. 3. Schematic illustration of the experimental station. Slope is not in scale.

Fig. 4. Installation of vertical TDR sensor cable in a borehole. (a) Anchoring of the cable to a wood cylinder. (b) Application of epoxy to prevent cable wetting. (c) Drilling. (d) Insertion of the cable into the borehole. (e) Cement grouting. (f) Grouted cable.
The standard deviations between the three repetitions for the laboratory data were always <0.03 m$^3$ m$^{-3}$, indicating a small variability among replicates for both measuring techniques; i.e., the precision of the measurements was good. However, a higher degree of scattering occurs at lower water content in the field data. This is mainly due to two factors: (a) SWCC hysteresis that determines numerous scanning curves between a main drying and wetting curve, and (b) a potential lack of good contact between the TDR probe and the soil, when the soil dries out and shrinks.

Although we had three SWCC (one for each profile), we fitted one SWCC only for the three measurement points and we integrated the data as shown in Fig. 5d. This choice was justified by the uniform soil properties (texture and bulk density) of the profile for these depths, as shown in Table 1. This allowed for an application of the soil stability model to the whole profile, without having to perform single analysis for each profile. From the practical standpoint and for the application of soil stability models, it is convenient to apply the model over an integrated soil profile, instead of separating the analysis for different profiles. This is particularly relevant when the models are applied over larger areas where information on profile horizons and layering is limited or unavailable.

Although the SWCC data displayed a degree of scattering, the data were purposely not filtered, so as to keep all the original experimental values. For instance, for the same value of degree of saturation there were often many different values of matric suction (“horizontal stripes” of matric suction data). We found that while the TDR was providing the same value of degree of saturation (i.e., for a few hours during the day), the HD sensors often provided varying values of matric suction. This was due to the different resolution of the two devices. The HD sensors detect changes in matric suction of 1 kPa (Flint et al., 2002), that may correspond to small changes in the degree of saturation (depending on the position on the SWCC, i.e. the soil water capacity) that the TDR cannot detect because of its lower resolution (2–3% volumetric water content) (Runkles, 2006).

Moreover, the SWCCs shown in Fig. 5 are data collected for different years and seasons. In some cases, during the experimental period different values of matric suction for the same values of degree of saturation were measured in different years (change of the SWCC over time). These differences are due to the shrinking and swelling of clay soils and to the different branches of the wetting and drying cycles. Indeed, TDR measures a volumetric value of soil water content, therefore if the soil bulk density changes over time, as a result of swelling–shrinking, so does the TDR reading.

It is indeed well known that the SWCC changes with time and it is not a constant property (especially in swelling soils) therefore laboratory determination of the SWCC may not be the best choice for this type of studies. Indeed, Nuth and Laloui (2008) pointed out that the term “soil water characteristic curve” is used as an intrinsic property, thus implying that there exists a unique water retention curve. On the contrary, they showed that it depends on the state of the material and is not invariant with time. The same results were found in this study and therefore a direct field determination of the SWCC, encompassing these time-dependent changes, may provide a better description of the field processes.

The lack of scatter close to saturation was due to the direct derivation of the matric suction from degree of saturation data (as described above) by inverting the van Genuchten equation, and not directly from experimental HD sensors data.

### 5.2. Localization of soil failure planes

Fig. 6 shows TDR traces for the coaxial cables for two different periods. The slope movement was only a few tens of centimeters, therefore the instrumentation was not destroyed. The landslide is approximately 25 m long and 7 m large and the center of the landslide is at N44°28 and E11°28 (Figs. 1 and 2). Since no deformation from TDR reflections were observed on March 2, 2006 while on May 1, 2006 two deformations were detected at 0.8 and 1.4 m depth, clearly the landslide occurred within this time interval. The important change in the reflections was due to a rupture in the cable and not to soil volumetric variations, which did not determine any fracture previously. Moreover, a visual occurrence of the
landslide was verified on May 1, 2006, when the TDR measurement was manually performed. The lower failure (of higher reflection) in Fig. 6 (right plot) was very likely the effective failure plane, while the most superficial fracture (of lower reflection) was probably due to a secondary fracture occurring as a result of the deeper fracture, which dragged the soil mantle downhill. Indeed, when the cement in the borehole is broken, it is not uncommon to find secondary breaks due to soil movement. Between 1.45 and 1.65 m depth (Table 1), the soil had a higher bulk density (1420 kg m\(^{-3}\)) than the overlying layers (1120 to 1230 kg m\(^{-3}\)), identifying a compacted layer characterized by a lower permeability. This high-density layer seems to have determined conditions very near to saturation of the upper layers for a prolonged period. This inference was corroborated by the slope stability analysis described below.

The transition zone between the shallow soil and the deeper compacted soil layer constituted the failure plane for the landslide, as shown by the depth of the failure planes and by the profile’s change in bulk density. Overall, shallow landslides occurred here because of slipping over compacted layers of clay materials of lower permeability.

5.3. Measured and simulated soil water content and matric suction dynamics

Fig. 7 depicts experimental soil water volumetric content and matric suction as a function of time for three different depths at the upper and lower profiles, as well as daily precipitation (upper graph). The inset (failure period, FP) details the dynamics of soil matric suction at the three depths for the period March–May, 2006, corresponding to the period when the landslide occurred.

Fig. 8 shows the simulated water content obtained from the CRITERIA model described in Section 2.3. The three graphs (Fig. 8a–c) show a comparison between the simulated and the measured data for the three experimental depths (0.2, 0.4 and 0.8 m). The model was able to simulate the soil water content fairly well, with Nash and Sutcliffe (1970) (NS) values of 0.49, 0.56, and 0.26 for the three depths, respectively. Fig. 8d depicts the numerically simulated water content at 1.4 m depth, confirming the soil water content at or close to saturation in the period March 2 to May 1, 2006 (values of saturation for the profiles are listed in Table 1). We can then assume that the simulated dynamics at 1.4 m depth are representative of the field condition and can be used for computation of the matric suction and degree of saturation needed by the stability model.

The soil water volumetric content and matric suction data depict the typical inverse relationship, with soil water content decreasing when the matric suction increases, representing a dynamical, and field measured SWCC. The 0.8 m depth of the lower profile shows scattered soil water content data with respect to the other measurement points, which may be due to an incomplete contact between the soil and the sensor. At the end of the winter and spring (also before the landslide of spring 2006), the soil was close to saturation. Indeed, soil water content oscillated between 0.4 and 0.5 m\(^{3}\) m\(^{-3}\) for most of the winter at all the three depths. Similarly, the matric suction displayed small variations ranging from −1 to −10 kPa, although because of the instrument range these measured data are less reliable. However, more reliable data can be obtained from the soil water content measurements and from the soil water retention curve as described above.

The unsaturated zone dynamics depicts a poorly drained soil, where the soil water content does not change significantly for most of the winter. The decrease in soil water content during the summers represents the climatic, pedological and vegetational conditions. The area was covered by a dense natural vegetation (various species of shrubs and herbaceous plants), typical of the natural settings of the Apennine Mountains, resulting in high plant water uptake, with roots that can reach down to 1.5 m depth (Tosi, 2007). Therefore the decrease in soil water content in the summer was mostly due to plant transpiration. The soil profile is then recharged by precipitation in the fall season. The soil is characterized by low saturated hydraulic conductivity (from 1.5 to 4 cm day\(^{-1}\)), therefore the low vertical conductivity keeps a high SWC value for long periods, as shown by the experimental data.

5.4. Slope-stability analysis

In agreement with the geomorphological observations reported in Section 3, a conceptual model of the observed landslide mechanism can be derived. The hillside of the study site was clearly affected by antecedent shallow earthflows, moving towards the stream in a north-east direction, which can be delineated by the high resolution DEM (Fig. 2). The small swales that appear in Fig. 2 are apparently maintained through rapid evacuation of colluvium by landsliding or gullying, followed by slow refilling by soil creep, biogenetic transport, scarp backwearing and sliding from areas further upslope (Lehre, 1987). The main mechanism may be a seasonal creep of the shallow soil in correspondence of the foothill. Given the geometry of the slope profile, the type of the shallow soil and its shear strength characteristics, the slope stability (within a soil thickness of about 2 m) had been maintained for a long time by stabilizing effects of roots and soil matric suction. The latter drastically decreases in consequence of prolonged rainfalls, as demonstrated by field measurements. Notwithstanding the lack of deformation measurements near the foothill, it can be reasonably supposed that the downslope transport has been driven by disturbances mainly related to wet–dry and freeze–thaw cycles and, probably, also by mineral weathering (Roering, 2004). Such a mechanism could be modeled through a finite element analysis by using appropriate constitutive models for soil in partially saturated conditions, and this will be done for this research after a detailed geo-mechanical characterization.
of the soil. However, on the basis of the available geotechnical parameters, a simplified but suitable slope-stability analysis can be carried out. For a simplified slope-stability analysis the soil could be considered substantially the same at the four depths. This assumption is very useful when slope stability models are applied to extended areas as predicting tools, and where detailed information on the soil stratigraphy are difficult to acquire. For this reason, the limit equilibrium has been applied at different depths by considering a single homogeneous soil layer with a very similar set of physical and shear strength parameters (Table 2). Moreover, based on soil properties analysis (Table 1) the soil is indeed fairly uniform with small differences in texture and bulk density for the first 1.4 m in depth. Therefore, this assumption is both convenient for practical purposes and representative of the experimental site. It is worth underlining that, since the same set of van Genuchten parameters has been selected for the entire soil thickness, one suction stress curve only represents the relationship between suction stress and effective saturation for the investigated soil at different depths. In particular, the suction stress reached the maximum value of 6.7 and 6.4 MPA at 0.2 and 0.4 m, respectively, while it exceeded 10 MPA for the soil at 0.8 m.

Regarding the soil layer at 1.4 m, the lack of soil water content and matric suction field measurements led us to use simulated daily values of volumetric water content by using a numerical model, as explained in Fig. 7. Soil water content and soil matric suction as function of time at the upper and lower soil profiles. The solid dots are soil water content data, while the open triangles are soil water suction data. The inset rectangle FP (failure period) shows the period where soil failure occurred (from March to May 2006).
such an approximation was reasonable due to the similarity of soil characteristic of the shallow soil layers and the good representation of soil hydrology by the numerical model in the shallow part of the soil, i.e. at 0.2, 0.4 and 0.8 m depths, the effective cohesion has been assumed to be 6.0, 7.7 and 3.6 kPa, respectively, including both the residual effective cohesion of the soil ($c' = 0$ kPa) and the contribution to the shear resistance by the roots measured by Tosi (2007).

With regard to the effective cohesion, different values have been used at different depths, taking into account root contribution. This area is densely vegetated and the roots reach depths of 1.2–1.5 m. For the shallow part of the soil, i.e. at 0.2, 0.4 and 0.8 m depths, the effective cohesion has been assumed to be 6.0, 7.7 and 3.6 kPa, respectively, including both the residual effective cohesion of the soil ($c' = 0$ kPa) and the contribution to the shear resistance by the roots measured by Tosi (2007). It is worth noting, in fact, that the sample site was characterized by the presence of some common autochthonous shrub species, such as I. Viscosa (L.). The root tensile strength of these shrubs has been analyzed by Tosi (2007) for the same site: he affirmed that the shear strength of the soil increased, thanks to the roots’ contribution, for a thickness of 0.6 m from the ground level much more than in the deeper layers. In Table 2, selected values of effective cohesion include the shear improvement given by the roots for the first three layers. On the contrary, the root contribution has been disregarded for the soil at 1.4 m from ground level, where only the residual effective cohesion ($c' = 0$ kPa) has been assumed (Table 2).

On the other hand, the location of the slip surface could have been controlled by the vertical reach of the roots. The failure occurred, in fact, at a depth where the root strength effect became negligible. We did not take into account the stabilizing effect of roots at 1.4 m, because the failure plane in the actual soil profile lies well below the major root zone.

It is worth underlying the importance of the effect of natural vegetation root systems on soil mechanical properties, as performed by Tosi (2007) for this study site and incorporated into this research. Without considering the contribution of root resistance, instability is reached at shallower depths. The slope angle ($\beta$) has been determined on the basis of a detailed topography of the site and $F_s$ has been calculated at each investigated depth ($z$). These results stress the importance of incorporating the variations in soil shear strength due to vegetation, which is often not accounted for in distributed models because of lack of reliable experimental data in many cases.

On the basis of the field measurements, we had at our disposal detailed soil water content and matric suction (Figs. 7 and 8) at three different depths (i.e. at 0.2, 0.4, and 0.8 m) over a period of 2 years. It was possible to calculate the daily value of suction stress and $F_s$ at each of the three depths by using Eqs. (2) and (3).

Table 2: Parameters used for slope-stability analysis.

<table>
<thead>
<tr>
<th>$z$ (m)</th>
<th>$\phi$ (deg)</th>
<th>$c'$ (kPa)</th>
<th>$\gamma_{sat}$ (kN m$^{-2}$)</th>
<th>$\beta$ (deg.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.2</td>
<td>12</td>
<td>6.0</td>
<td>18</td>
<td>14</td>
</tr>
<tr>
<td>0.4</td>
<td>12</td>
<td>7.7</td>
<td>18</td>
<td>14</td>
</tr>
<tr>
<td>0.8</td>
<td>12</td>
<td>3.6</td>
<td>18</td>
<td>14</td>
</tr>
<tr>
<td>1.4</td>
<td>12</td>
<td>0</td>
<td>18</td>
<td>14</td>
</tr>
</tbody>
</table>

Fig. 8 shows $F_s$ as function of time for the span period of 27 months. There is a seasonal trend with $F_s$ reaching very high values during dry seasons, when the soil is characterized by high values of matric suction. For the first three depths, $F_s$ was > 1, thus suggesting a condition of stability, for the whole monitored period. On the other hand, at a depth of 1.4 m, $F_s$ displays a very different trend compared to the shallower layers. $F_s$ was very close to unity for two winter periods, from mid-December to April. In detail, during the first winter, $F_s$ was > 1.2, while during the second period $F_s$ was well below 1.2 and a bit higher than 1.1. When $F_s$ gets closer to 1,
the slope is less stable, and when $F_s$ is equal to 1, the slope is unstable. Therefore $F_s$ approaching 1 is a condition of alarm.

$F_s$ was equal to 1 on December 1, 2005 and on March 16, 2006. The obtained values of $F_s$ on March 16, 2006 and in the following days are consistent with the TDR traces acquired at the experimental site between March 2 and May 1, 2006, showing the occurrence of a triggering mechanism, as caught by the model.

The results, based on Eq. (2), lead us to believe that the model could be used with a certain degree of confidence even in the absence of accurate field measurements of matric suction, but only after a correct determination of the SWCC and measurement of soil water content as discussed below. On December 1, 2005, the instability condition revealed by the model did not correspond to any landslide movement, notwithstanding that the amount of cumulative precipitation in the fall of 2005 was about three times that of winter 2006. Indeed, the cumulated rainfall from October 1 to November 29, 2005 was 330 mm, while from January 16 to March 15, 2006, it was 109 mm. This suggests that the abundant fall rainfall could have determined an acceleration of the soil creep in the foothill zone. Moreover, it suggests that the instability condition could be affected by the infiltration mechanism and by the variation of matric suction along the soil profile, more than by the cumulative precipitation. For instance, on September 18, 2005 there was the most intense rainfall event of the whole period, with 143 mm over 24 h. However, since the soil profile was fairly dry (it was one of the first significant rainfalls after the summer), $F_s$ did not fall to unity, and the strength contribution given by the suction stress was relatively high. Regarding the effectiveness of the stability model of Lu and Godt (2008), it is worth noting that the model was not calibrated, since the soil shear strength parameters were computed from previous field experiments, as detailed by Tosi (2007) and they were not changed or adjusted during model applications. Similarly, the hydrological parameters (i.e. the van Genuchten parameters) were derived from fitting of independent field measured data. We can therefore consider the present test and application of the model as a verification of the model capabilities (hind-cast study).

5.5. Suction stress analysis

We performed further analysis on suction stress and volumetric water content (from ground level down to 1.4 m), over 5 days prior and 2 days after the above mentioned dates, to explain the soil conditions leading to the triggering mechanism, and to understand why the first time $F_s$ was equal to one, a landslide did not occur.

The suction stress profiles referring to the period between November 29 and December 3, 2005 were obtained by using Eq. (2) on the basis of field measurements of volumetric water content for points at 0.2, 0.4 and 0.8 m, and, from the numerically simulated water content for the point at 1.4 m. The obtained trends are shown in Fig. 10a.

As shown above, where $F_s$ is discussed, the minimum values of suction stress at a depth of 1.4 m were reached from November 29 to
December 1, 2005 and from March 14 to March 18, 2006. Field measurements of both matric suction and volumetric water content reveal how at a depth 0.8 m the soil was relatively drier than the shallow soil layer. In all the investigated dates, in fact, the maximum value of suction stress was reached at a depth of 0.8 m. Due to infiltration and redistribution processes, after a rainfall, the suction stress at 0.8 m increased, while it progressively decreased at 0.2 and 0.4 m. The suction stress profile seems to have gone in the opposite verse while the water content increased with depth. It is evident from the simulation (Fig. 8) how, on November 29, 2005, the soil was near saturation from ground level down to 0.8 m; then, the water content increased with depth after the occurrence of a rainfall of 20.8 mm on November 30, 2005 and another of 17.4 mm on December 1, until nearly complete saturation is reached at a depth of 1.4 m.

A similar analysis was carried out for the same profile over the period between March 14 and March 18, 2006 (Fig. 10b). In this case, the minimum value of suction stress is at 1.4 m depth, where the simulations reveal an increase of water content. The daily suction stress profile shows a trend very similar to that observed in the previous analysis, as a consequence of the unsaturated downward seepage. It is evident from the simulation how the soil is near saturation, from ground level down to 1.3 m, already 2 days before the main rainfall event on March 16, 2006. In this case, the thickness of the nearly saturated soil is greater than that observed previously.

The triggering mechanism of the observed soil movement, therefore, occurred on March 16, 2006. Two main characteristics distinguish the events on December 1, 2005 and March 16, 2006.

1. Field measurements of both matric suction and water content reveal how, on both the dates, the soil at 0.8 m depth was drier than at above but its water content was higher on March 2006 than on December 2005.
2. The loss of the shear strength contribution given by the suction stress, at 1.4 m depth, has to be analyzed in comparison with the reduction in suction stress in the entire soil profile involved in the sliding mechanism; in particular, if the dynamics of unsaturated seepage was similar in the soil profile for the two dates, a more important role is assumed by the absolute suction stress value in correspondence to the sliding surface, which was lower for the event that occurred in March.

6. Discussion and conclusions

The general picture emerging from this study reveals that the triggering factors of a shallow landslide in a clayey soil were changes in soil matric suction, associated with the presence of compacted layers of different hydrological properties. In particular:

(a) It was observed that slope failure occurred at a depth of 1.4 m from ground level, corresponding to the interface between the superficial soil (more permeable and less compacted) and a substrate of very low permeability and high compactness, in correspondence to a thin plastic soil layer, under partially saturated conditions.

(b) By comparing the timing of a landslide occurrence, rainfall amounts and results of the slope-stability analysis, the dominant factor triggering the landslide seems to be the water infiltration and redistribution mechanism within micro- and macro-porosity rather than cumulative rainfall. Field measurements of soil water content and matric suction revealed it to be fundamental to the slope stability analysis.

(c) The triggering mechanism has been modeled through the limit equilibrium method, with the hypothesis of infinite slope, by using the concept of suction stress in the stability model (Lu and Likos, 2006), and the results appear satisfactory. This method considers only the main aspects of the triggering mechanism. Swelling–shrinking phenomena and seasonal creep, for example, which have been disregarded in the present analysis, may induce secondary effects on soil deformation and instability.

(d) The model based on suction stress, accommodates both instability ($F_s = 1.0$) at the correct time and stability ($F_s > 1.0$) at all other times. The main difference, with respect to the results presented by Godt et al. (2009), is the application of the model to a clayey soil, instead of a sandy colluvium. The model takes into account soil conditions of partial saturation and provides $F_s$ of the slope as a function of the water content. Results obtained on the basis of field measurements of water content are reliable, after the determination of the SWCC. This result showed that the model could be used with a certain degree of confidence even in the absence of accurate field measurements of matric suction, but with measurements of water content and after the definition of SWCC.

(e) These results demonstrate that the correct experimental characterization of the SWCC is a key element for the successful estimation of $F_s$. Indeed, in field conditions the SWCC is affected by varieties of factors that are commonly not accounted for in laboratory conditions, such as time dependent bulk density, shrinking and swelling, and hysteresis, but that they are encompassing the time-dependent changes of hydraulic properties.

By combining the results of the slope-stability analysis and the observation of the site geomorphology, it is reasonable to suppose that during the rainfall event of early December 2005 the soil at a depth of 1.4 m underwent only a very small deformation, without reaching failure. The most probable date of failure is March 16, 2006, when the cumulative small plastic deformations that occurred near the foothill during previous months, became notable as a failure surface from TDR strain measurements near the hilltop. Even though the limit equilibrium method can be considered a rough means to model such a kind of soil deformation, we can state that the model identified the triggering mechanism of this shallow landslide.

The results obtained stress the importance of measuring soil hydrological properties (i.e. the SWCC). They confirm that using only rainfall data as a proxy for soil instability provides highly insufficient information for correct assessment of landslide processes, especially with regards to clayey shallow soils. As already well established by others, the presented study also strengthens the importance of the integration of different techniques for continuous monitoring of soil hydrological and stability properties, through both the installation of appropriate instruments in landslide susceptible areas and the remote acquisition of measurements by satellite transmission, in order to enhance the assessment of unstable slope conditions for early warning purposes. It has been shown, in fact, how the experimental setup of TDR devices used in this study could successfully detect the time-interval and the depth of the failure plane, as well as soil water content and soil matric suction.

However, hourly-based acquisitions of TDR traces for mass deformation studies, together with a larger number of instruments over a wider area, will provide better information on the relationship between landslides and single rainstorms since it would allow for the identification of the exact time of the landslide occurrence. Moreover, such an integrated system can be used as an alarm in susceptible areas or areas where an existing landslide danger has been already established. Finally, the soil mantle depths, necessary for modeling spatially distributed shallow landslide potential, were experimentally determined and can be used for assessment of landslide susceptibility on a catchment scale.

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References