Hillslope scale soil moisture variability in a steep alpine terrain

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SUMMARY

In this study we analyse space–time variability of soil moisture data collected at 0–6, 0–12 and 0–20 cm depth over three hillslopes with contrasting steep relief and shallow soil depth in the Dolomites (central-eastern Italian Alps). The data have been collected during two summer seasons (2005 and 2006) with different precipitation distribution. Analysis of soil moisture data shows that different physical processes control the space–time distribution of soil moisture at the three soil depths, with a marked effect of dew on the 0–6 cm soil depth layer. The range of skewness values decreases markedly from the surface to deeper layers. More symmetric distributions, characterised by relatively low skewness, are found for mid-range soil moisture contents, while highly skewed distributions (generally with more log–normal shape) are found at dry and wet conditions. Scatter plots drawn for the whole data set and the analysis of the correlation coefficients suggest a good persistence of soil moisture with depth: the highest degree of correlation was observed between data collected at 0–12 and 0–20 cm.

Examination of correlation between soil moisture fields and topographical attributes shows that, notwithstanding the steep relief and the humid conditions, terrain indices are relatively poor predictors of soil moisture spatial variability. The slope and the topographic wetness index, which are found here the best univariate spatial predictors of soil moisture, explains up to 42% of the time-averaged moisture spatial variation.

A negative relationship between the soil moisture spatial mean and the corresponding spatial standard deviation is found for mean water contents exceeding 25–30%, while a transition to a positive relationship is observed with drier conditions. Overall, soil moisture variability shows the highest values at moderate moisture conditions (23–29%) and reduced values for wetter and drier conditions for all depths. A negative linear relationship between mean soil moisture content and the coefficient of variation was observed.

A soil moisture dynamics model proved to successfully capture the soil moisture variability at the hillslope scale. The simulated time series of hillslope-averaged soil moisture are in good agreement with the observed ones. Moreover, the model reproduces consistently the observed relationships between soil moisture spatial mean and corresponding variability.

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Introduction

Soil moisture plays a central role in the global water cycle by controlling the partitioning of water and energy fluxes at the earth's surface and constitutes the physical linkage between soil, climate and vegetation (Albertson and Montaldo, 2003; Pan et al., 2003; Rodríguez-Iturbe and Porporato, 2004). At the point scale soil moisture is crucial for the infiltration process (Bronstert and Bárdossy, 1999; Raats, 2001) and plant dynamics (Porporato et al., 2004). At the hillslope and catchment scale, the spatial and temporal distribution of soil moisture controls the flood formation process (Borga et al., 2007). At the regional and continental scale, soil moisture controls water distribution through land surface atmosphere feedback mechanisms (Koster et al., 2004).

Due to soil heterogeneity, atmospheric forcing, vegetation and topography, soil moisture is variable in space and time. Understanding and characterizing this variability is one of the major challenges within hydrological sciences. Information characterizing space–time variability of soil moisture is important to understand the contribution of soil moisture variability at smaller scales towards the effective soil moisture observed at larger scales or its role in the parametrization of, e.g., climate and watershed models (Famiglietti et al., 1999; Ryu and Famiglietti, 2005). As such, this information can provide guidelines for the design of field experiments and for the efficient use of remote sensing estimates.

Characterization of space–time soil moisture variability has been attempted by analysing the trends of spatial soil moisture
variability with spatial mean moisture content. Empirical analyses found that, generally, the standard deviation increases during drying from a very wet stage, reaches a maximum value at a specific or critical mean moisture content and then decreases during further drying (Famiglietti et al., 2008). Western et al. (2003) provided a qualitative interpretation of why variance peaks at intermediate moisture contents by making use of the combined results from several field campaigns. In their interpretation, differences in behaviour in humid and semi-arid regions are related to differences in the patterns of controlling processes. They found that the location and magnitude of the variance peak changed between catchments. Depth appeared to have only a small effect on the relationship. In the last 10 years, a number of studies have attempted to examine quantitatively how different processes act to either increase or decrease the spatial variability of soil moisture. By using the similar media concept, Salvucci (1998) showed how variability in soil texture leads to different soil moisture variability states in different limiting cases. Albertson and Montaldo (2003) showed how covariances between soil moisture and fluxes, originating from variability in soil moisture, forcing and/or land surface properties, can lead to either an increase or decrease in soil moisture variability. Teuling et al. (2005) developed a simple soil moisture dynamics model and showed how vegetation, soil and topography controls interact to either create or destroy spatial variance. By accounting for effects of spatial variability in soil and vegetation characteristics, in combination with atmospheric forcing (precipitation and potential evapotranspiration), different observed relations between spatial mean soil moisture and its variability can be explained in this way. Using stochastic analysis of the unsaturated Brooks–Corey flow in heterogeneous soils, Vereecken et al. (2007) showed that parameters of the moisture retention characteristic and their spatial variability determine to a large extent the shape of the soil moisture variance–mean water content relationship. They found that the standard deviation of soil moisture peaked between 0.17% and 0.23% for most textural classes and that the peak value was controlled by the parameters which describe the pore size distribution of soils. The theoretical results obtained by Vereecken et al. (2007) correspond well with Ryu and Famiglietti (2005) who experimentally found that the soil moisture variance–mean water content relationship tends to peak around a value of 0.2%.

These studies generally examined soil moisture variability for gentle topographies. Examples are provided by the Tarrawarra and Mahurangi experiments (Western and Grayson, 1998, 2000, 1999, 2004; Wilson et al., 2003, 2004) and SGP97, SGP99, SMEX02 and SMEX03 (Famiglietti et al., 1999; Mohanty and Skaggs, 2001; Choi and Jacobs, 2007). Few field studies have examined variability of soil moisture patterns in steep terrain and high altitude (above tree line) conditions (Grant et al., 2004). In general, the spatial variability of soil moisture in mountainous regions is expected to be high relative to other landscapes due to heterogeneous conditions of surface and bedrock topography, soil characteristics, wind patterns, interaction between evaporation, condensation and precipitation. Moreover, it is expected that the soil moisture spatial variability in humid and steep conditions with shallow soils (more favourable to lateral flow occurrence) exhibits a stronger relationship with topographic variables than in more gently sloping landscapes (Grayson and Western, 2001).

The main objective of this study is to characterise the variability of field scale soil moisture for three hillslopes characterised by contrasting steep relief and shallow soil depth located in the high mountainous Vauz river basin (Dolomites, central eastern Italian Alps). Soil moisture data were collected over three depths: 0–6, 0–12 and 0–20 cm. The three hillslopes show marked differences in topography, representing concave, planar and convex structure. Because of the relative small area, we do not expect significant differences in precipitation, relative air humidity, temperature and other climatic variables. This allows to isolate the effects of topography and soil depths on space–time variability of soil moisture fields. The hillslopes considered may be deemed representative of pedological, topographic, climatic conditions frequently met in pasture areas of the Dolomites.

Based on these data, this study (1) characterises the statistical properties of the soil moisture measurements with varying soil depth and hillslope topography, and investigates the correlation between soil moisture measurements collected at various soil depths; (2) examines the influence of topographic variables on soil moisture spatial variability; (3) investigates the relationship between mean moisture content and spatial variability and (4) analyses the suitability of applying a soil moisture variability dynamics model to reproduce the observed spatial mean moisture content – spatial variability relationship. The use of a water balance model to characterise the spatial and temporal organisation of soil moisture fields is appealing, since this could reduce the need for ground based measurements significantly in numerous applications. The model applied here (Teuling and Troch, 2005) accounts for variations in soil and vegetation properties but not for redistribution due to lateral flow. It is therefore interesting to evaluate the quality

![Figure 1. Rio Vauz catchment and location of the three hillslopes.](image)
of the model simulations in the steep topography landscape considered in this study. Models of similar complexity have been shown to correctly simulate the root zone soil moisture dynamics under different climatic conditions (Albertson and Kiely, 2001; Teuling et al., 2005), but not in complex topographic settings.

Study area and data collection

Soil moisture data were collected in the lower portion of a small headwater alpine basin (Rio Vauz catchment, 1.9 km²), located in the central-eastern Italian Alps (Fig. 1), with altitude ranging from 1835 m ASL to 3152 m ASL. The basin has a typical alpine climate with a mean annual rainfall of about 1220 mm, 49% of which falls as snow. The precipitation monthly distribution shows a peak in early summer and a second one during fall (Fig. 2). In the lower parts of the catchment the snow cover period typically lasts from November to April. Runoff is usually dominated by snowmelt in May and June but summer and early autumn floods represent an important contribution to the flow regime. The average monthly temperature in the lower Vauz varies from −5.7 °C in January to 14.1 °C in July. Potential evapotranspiration, computed by means of the Penman–Monteith equation (Monteith, 1965), is in phase with precipitation. Moreover, monthly mean precipitation always exceeds monthly mean potential evapotranspiration. The climate can be considered humid, according to the Budyko classification (Norbiato et al., 2008 and references therein).

The basin is almost undisturbed by human activity: neither roads nor urban areas are present. The soil is primarily vegetated with grass, with root zone depth around 20 cm. Sparse woody vegetation (mainly larch) is distributed on a few hillslopes in the lower part of the catchment (Borga et al., 2002). A refined DEM with a 1 m resolution, delineated based on point data obtained from a total station survey, was developed for the basin.

Three hillslopes were selected in the lower part of the basin (Fig. 1) to provide detailed soil moisture data. The experimental sites have been named “Piramide”, “Emme” and “Vallecola”, with an area of 0.46, 0.47 and 0.57 ha, respectively. The hillslopes are characterised by markedly different relief shape. Accordingly with the hillslope relief characterisation introduced by Norbiato and Borga (2008), Piramide is a divergent-convex hillslope, Emme is a relatively planar hillslope, and Vallecola is a convergent-concave hillslope. An eroded landslide scar exists in the central portion of Vallecola. On several occasions, pipe flow was observed on the walls of the eroded gully.

The soils in the lower part of the Vauz catchment are relatively homogenous and consist of a Cambisol with a clay loam A horizon (up to 20 cm deep) with 32% clay and 42% sand, followed by a silty clay loam B horizon with presence of gravel and cobbles. The B horizon is located atop a weathered C horizon rich of gravel and coarse rock fragments. Soil depths, measured directly by using an iron pole, vary from 10 cm at some points on the ridgetops to 130 cm at the hillslope grounds. In these soils, structural soil pores constitute important pathways for water and may carry water before the finer pores of the soil matrix are fully saturated. Macropores are represented mainly by animal burrows, such as worm tunnels or structures built by small mammals.

The rainfall accumulation during the experiments amounts to 130 mm and 99.6 mm, for 2005 and 2006, respectively. The water table depth are also available from a network of 39 recording wells. The water level was measured every 5 min using capacitance rods (Trutrack, New Zealand). Soil samples were collected at the three different soil depths for porosity and particle size analysis. Porosity over 0–6 cm soil depth ranges from 48% to 75% with a mean of 59%; samples collected over 0–12 and at 0–20 cm depth exhibit similar distributions, ranging between 45% and 58%, with mean values around 53%.

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Field-saturated hydraulic conductivity was measured by means of a Guelph ‘constant head’ permeameter at several points over the hillslopes. Values ranged from 1.1 × 10⁻⁶ to 2.0 × 10⁻⁷ m/s with a mean of 1.1 × 10⁻⁶ m/s. The higher saturated conductivity values may reflect rapid pipe flow through worm holes and other preferential flow conduits. Topographic and soil characteristics are reported in Table 1 for the three hillslopes.

### Soil moisture data collection and hydrologic monitoring

Soil moisture data were collected at 0–6, 0–12 and 0–20 cm depth during two field campaigns: from 28 June to 21 July, 2005 and from 21 June to 16 July, 2006. Difficulties with deeper sampling, due to the presence of cobbles, prevented measurement at 0–20 cm over Vallecola. In the 2006 field campaign, soil moisture was investigated only for Piramide and Emme and data over Vallecola were not collected. Number of sampling sites and times for the three sites and the three depths are reported in Tables 2a and b for 2005 and 2006, respectively. Locations of the sampling sites are reported in Fig. 3.

Precipitation data were collected by means of a tipping bucket rain gauge located at 1923 m ASL, close to the sampled hillslopes. The rainfall accumulation during the experiments amounts to 130 mm and 99.6 mm, for 2005 and 2006, respectively, while the climatological average for the period is 120 mm. Data of water table depth are also available from a network of 39 recording wells. The water level was measured every 5 min using capacitance rods (Trutrack, New Zealand). Soil samples were collected at the three different soil depths for porosity and particle size analysis. Porosity over 0–6 cm soil depth ranges from 48% to 75% with a mean of 59%; samples collected over 0–12 and at 0–20 cm depth exhibit similar distributions, ranging between 45% and 58%, with mean values around 53%.

Soil moisture values at 0–6 cm depth were sampled by means of a Guelph ‘constant head’ permeameter at several points over the hillslopes. Values ranged from 1.1 × 10⁻⁶ to 2.0 × 10⁻⁷ m/s with a mean of 1.1 × 10⁻⁶ m/s. The higher saturated conductivity values may reflect rapid pipe flow through worm holes and other preferential flow conduits. Topographic and soil characteristics are reported in Table 1 for the three hillslopes.

### Table 1

Topographic properties of the three study hillslopes.

<table>
<thead>
<tr>
<th></th>
<th>Piramide</th>
<th>Emme</th>
<th>Vallecola</th>
</tr>
</thead>
<tbody>
<tr>
<td>Area (ha)</td>
<td>0.46</td>
<td>0.47</td>
<td>0.57</td>
</tr>
<tr>
<td>Slope range (%)</td>
<td>21–84</td>
<td>21–90</td>
<td>25–90</td>
</tr>
<tr>
<td>Soil depth (cm)</td>
<td>20–130</td>
<td>10–130</td>
<td>10–130</td>
</tr>
</tbody>
</table>

Figure 2. Climatic conditions in the lower Vauz catchment.
and Miller, 1996; Miller et al., 1997). Soil moisture at 0–12 and 0–20 cm depth was evaluated by means of a TDR 300, a portable probe manufactured by Spectrum Technologies Inc. (the mention of trade and company names is for the benefit of the reader and does not imply an endorsement of the product) and operating on the basis of time domain reflectometry technology. The TDR probe is provided with two pairs of interchangeable rods of 12 cm and 20 cm length which allow to sample soil moisture over these different depths.

Both Theta Probe and TDR 300 were gravimetrically calibrated for the specific local soil conditions (Stenger et al., 2005; Walker et al., 2004). A split tube soil sampler was used in the field to obtain undisturbed samples and 55, 45 and 40 soil cores were collected at the three investigated depths (0–6, 0–12 and 0–20 cm), respectively. Sampling was carried out so that the collected cores were characterised by different water contents, necessary to calibrate the instruments with soil conditions varying from a dry to a wet status (Kaleita et al., 2005). The soil samples were weighted, oven-dried at 105 °C for 24 h and weighted again and a calibration curve was obtained for each instrument plotting measured volumetric versus probe-derived soil moisture values (Fig. 4). For TDR 300, samples of both depths (12 and 20 cm) were used together since they appeared to fit the same relationship. Calibration curves exhibit a root mean squared error (computed as in Cosh et al., 2005) of 2.35% and 2.09% for Theta Probe and TDR 300, respectively, with standard deviations of errors equal to 2.37% for Theta Probe and 2.1% for TDR 300. Bias is negligible. Note that throughout the paper soil water content is always reported as volumetric soil moisture.

Soil moisture was measured at 26 sites over Piramide and Emme, and at 16 sites over Vallecola (Tables 2 and 3; Fig. 3). At each measurement point, five measures were collected in order to ascertain the repeatability of the results and instrument errors;

**Table 2a**
Number of soil moisture measurements for 2005.

<table>
<thead>
<tr>
<th></th>
<th>Piramide</th>
<th></th>
<th></th>
<th>Emme</th>
<th></th>
<th></th>
<th>Vallecola</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>0–6 cm</td>
<td>0–12 cm</td>
<td>0–20 cm</td>
<td>0–6 cm</td>
<td>0–12 cm</td>
<td>0–20 cm</td>
<td>0–6 cm</td>
</tr>
<tr>
<td>No. of sampling points</td>
<td>26</td>
<td>26</td>
<td>26</td>
<td>26</td>
<td>26</td>
<td>26</td>
<td>16</td>
</tr>
<tr>
<td>No. of sampling times</td>
<td>24</td>
<td>24</td>
<td>8</td>
<td>25</td>
<td>25</td>
<td>16</td>
<td>24</td>
</tr>
<tr>
<td>Total no. of measures</td>
<td>624</td>
<td>624</td>
<td>208</td>
<td>650</td>
<td>650</td>
<td>416</td>
<td>384</td>
</tr>
</tbody>
</table>

**Table 2b**
Number of soil moisture measurements for 2006.

<table>
<thead>
<tr>
<th></th>
<th>Piramide</th>
<th></th>
<th></th>
<th>Emme</th>
<th></th>
<th></th>
<th>Vallecola</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>0–6 cm</td>
<td>0–12 cm</td>
<td>0–20 cm</td>
<td>0–6 cm</td>
<td>0–12 cm</td>
<td>0–20 cm</td>
<td>0–6 cm</td>
</tr>
<tr>
<td>No. of sampling points</td>
<td>26</td>
<td>26</td>
<td>26</td>
<td>26</td>
<td>26</td>
<td>26</td>
<td>0</td>
</tr>
<tr>
<td>No. of sampling times</td>
<td>23</td>
<td>23</td>
<td>23</td>
<td>23</td>
<td>23</td>
<td>23</td>
<td>0</td>
</tr>
<tr>
<td>Total no. of measures</td>
<td>598</td>
<td>598</td>
<td>598</td>
<td>598</td>
<td>598</td>
<td>598</td>
<td>0</td>
</tr>
</tbody>
</table>

**Figure 3.** Position of sampling points over the three experimental hillslopes.
the highest and lowest values were rejected and the mean was calculated over the three remaining values. The measurements at the three depths were taken almost concurrently at each site to reduce the effect of temporal variability on the comparison of results. The temporal gap between the acquisition of soil moisture values at the first and the last point was around 45 min: during such a short time and for no-rain periods, no measurable variations of water content among the sampling points occurred, as checked in the field by repeating at the end of the sampling program the measurement at the first sampled point. Measurements can be therefore considered as instantaneous. Points were sampled in the same order on each occasion. Measurements were generally carried out between 0900 and 1100 local time, to avoid showers which were likely to occur on the afternoon.

Fig. 5a and b shows the time series of the hillslope-averaged water content at different soil depths over the two years, respectively. Precipitation has clearly a strong influence on soil water content: dry periods are characterised by relatively low values which rapidly increase after rain events; the time series generally display comparable dynamics over the three hillslopes. Examination of the figures shows that the hillslope-averaged soil moisture content is highly dependent on sampling depth.

**Figure 4.** Regression between soil moisture values and probe outputs for the two instruments. (a) Theta Probe and (b) TDR 300.

**Table 3a** Summary of soil moisture statistics over the three experimental sites for the various depths for 2005 (only common sampling times are considered).

<table>
<thead>
<tr>
<th>Summary statistics 2005</th>
<th>Piramide</th>
<th></th>
<th>Emme</th>
<th></th>
<th>Vallecola</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>0–6 cm</td>
<td>0–12 cm</td>
<td>0–20 cm</td>
<td>0–6 cm</td>
<td>0–12 cm</td>
</tr>
<tr>
<td>No. of sampling points</td>
<td>26</td>
<td>26</td>
<td>26</td>
<td>26</td>
<td>26</td>
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<tr>
<td>No. of sampling times</td>
<td>8</td>
<td>8</td>
<td>8</td>
<td>16</td>
<td>16</td>
</tr>
<tr>
<td>Total no. of measures</td>
<td>208</td>
<td>208</td>
<td>208</td>
<td>416</td>
<td>416</td>
</tr>
<tr>
<td>Mean (%)</td>
<td>43.8</td>
<td>39.2</td>
<td>36.5</td>
<td>45.2</td>
<td>40.0</td>
</tr>
<tr>
<td>Median (%)</td>
<td>43.8</td>
<td>39.1</td>
<td>35.9</td>
<td>45.5</td>
<td>40.4</td>
</tr>
<tr>
<td>Standard deviation (%)</td>
<td>3.3</td>
<td>5.2</td>
<td>5.5</td>
<td>3.2</td>
<td>5.8</td>
</tr>
<tr>
<td>Coefficient of variation</td>
<td>0.07</td>
<td>0.13</td>
<td>0.15</td>
<td>0.07</td>
<td>0.15</td>
</tr>
<tr>
<td>Skewness</td>
<td>−0.8</td>
<td>0.0</td>
<td>0.3</td>
<td>−0.9</td>
<td>−0.3</td>
</tr>
<tr>
<td>Kurtosis</td>
<td>3.6</td>
<td>0.5</td>
<td>−0.5</td>
<td>1.5</td>
<td>−0.2</td>
</tr>
<tr>
<td>Inter-quart. range (%)</td>
<td>3.5</td>
<td>7.0</td>
<td>7.7</td>
<td>3.9</td>
<td>8.0</td>
</tr>
<tr>
<td>Minimum value (%)</td>
<td>26.6</td>
<td>20.4</td>
<td>26.1</td>
<td>31.7</td>
<td>23.1</td>
</tr>
</tbody>
</table>

**Table 3b** Summary of soil moisture statistics over the two experimental sites for the various depths for 2006.

<table>
<thead>
<tr>
<th>Summary statistics 2006</th>
<th>Piramide</th>
<th></th>
<th>Emme</th>
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</thead>
<tbody>
<tr>
<td></td>
<td>0–6 cm</td>
<td>0–12 cm</td>
<td>0–20 cm</td>
<td>0–6 cm</td>
</tr>
<tr>
<td>No. of sampling points</td>
<td>26</td>
<td>26</td>
<td>26</td>
<td>26</td>
</tr>
<tr>
<td>No. of sampling times</td>
<td>23</td>
<td>23</td>
<td>23</td>
<td>23</td>
</tr>
<tr>
<td>Total no. of measures</td>
<td>598</td>
<td>598</td>
<td>598</td>
<td>598</td>
</tr>
<tr>
<td>Mean (%)</td>
<td>39.9</td>
<td>36.6</td>
<td>33.3</td>
<td>41.7</td>
</tr>
<tr>
<td>Median (%)</td>
<td>43.4</td>
<td>40.1</td>
<td>35.4</td>
<td>45.2</td>
</tr>
<tr>
<td>Standard deviation (%)</td>
<td>8.3</td>
<td>11.4</td>
<td>9.1</td>
<td>8.3</td>
</tr>
<tr>
<td>Coefficient of variation</td>
<td>0.21</td>
<td>0.31</td>
<td>0.27</td>
<td>0.20</td>
</tr>
<tr>
<td>Skewness</td>
<td>−1.3</td>
<td>−0.9</td>
<td>−0.8</td>
<td>−1.4</td>
</tr>
<tr>
<td>Kurtosis</td>
<td>0.5</td>
<td>−0.3</td>
<td>−0.1</td>
<td>1.0</td>
</tr>
<tr>
<td>Inter-quart. range (%)</td>
<td>50.6</td>
<td>57.8</td>
<td>48.7</td>
<td>50.5</td>
</tr>
<tr>
<td>Maximum value (%)</td>
<td>15.7</td>
<td>8.0</td>
<td>8.9</td>
<td>14.1</td>
</tr>
</tbody>
</table>
The soil moisture content over the 0–6 cm soil layer is typically higher than over deeper layers. Furthermore, the shallow soil layer has less variability than the deeper layers.

**Distribution of data and correlation between moisture contents at different depths**

The comparison between soil moisture patterns at different sites and depths was evaluated by examining: (i) summary statistical properties of the data set; (ii) distributional properties and (iii) correlation between soil moisture contents collected over different soil depths.

**Summary statistical analysis**

Tables 3a and b show summary statistics for those periods when soil moisture data were available over the three depths, for 2005 and 2006, respectively. It is noteworthy in both tables that 0–6 cm soil moisture values are wetter and less variable than 0–12 and 0–20 cm values. This is contrary to what was reported by some researchers (Choi and Jacobs, 2007), who observed less variability for deeper layers than for shallower layers. Typically, several environmental factors (i.e., evaporation and rainfall) may cause higher variability at the surface than the subsurface. However, inspection of soil moisture data time series shows that hillslope-averaged data over different depths are relatively close to each other after rainfall events, and their difference increases with time during the dry-down. While influence of instrumental differences may not be excluded (even though the two soil moisture measurement devices were calibrated for local conditions), this calls for processes which are active during the dry periods and influence only the dynamics of the shallow layer. The partitioning of soil moisture in the first 6 cm with respect to deeper depths is consistent with the expected effects of observed dew formed through condensation, which reduces soil moisture depletion in the upper soil layers. The effect of condensation on soil moisture during the morning after clear-sky nights has been observed in the field and pretty high values of condensation (up to 1.8 mm per day) have been measured in the experimental site. These values are consistent with values reported by de Jong (2005). Similar effects were observed by Engstrom et al. (2005) in a site located in the Arctic coastal plain. Also, it is noteworthy that the formation of dew is essentially a nocturnal occurrence. The rate and duration of this process depends on the humidity, temperature and movement of the surface air layers, the sorption characteristics of the exposed surface, its radiation cooling and its heat supply from warmer soil layers. Therefore, the timing of the soil moisture surveys, conducted in the morning, may have influenced the measurement of soil moisture in the surface layer by enhancing the effect of dew.

In all cases, the standard deviation, the coefficient of variation and skewness are highly dependent on depth. In all cases but Piramid...
mide 2005, standard deviation and coefficient of variation increase from 0–6 cm depth to 0–12 cm and then decrease. In general, the distribution of moisture values over 0–6 cm depth at each of the three sites, while the other depths show either slightly negative kurtosis or values closer to zero. The increase of skewness with depth is consistent with the drier mean soil moisture over 0–12 and 0–20 cm depths, considering the evolution of skewness from being negative to symmetric then to positive with drying.

Intercomparison among the different hillslopes by soil depth can be carried out at best for 2006, because only 8 sampling times are available over the three depths for Piramide during 2005. Analysis of summary data (Table 3b) shows that statistics are generally comparable across the two sites, with Emme slightly wetter than Piramide and Vallecola. This is probably due its main westward aspect. Overall, this indicates that, in spite of the different topographic structures, differences in soil moisture distributions across the three hillslopes are negligible. Frequency plots for Emme and Piramide over the three depths are reported in Fig. 6 for 2006. It is interesting to note the bimodality in the distribution for 0–12 and 0–20 cm soil depths. This bimodality is consistent with the nonparametric test developed by Silverman (1986), which allows us to reject the hypothesis that the available soil moisture measurements are sampled from a unimodal distribution. Analysis of the data shows that bimodality in the distribution is originated from the temporal persistence around two distinct dry and wet moisture conditions.

**Distributional properties of spatial soil moisture fields**

The characteristics of the spatial soil moisture fields were also investigated. Appropriate probability distributions were identified for the spatial instantaneous soil moisture fields, differentiating by depth and by site. The normal distribution and log-normal distribution were analyzed as they are the most widely used PDFs for soil moisture analysis. Probability plot correlation coefficient (PPCC) tests (Vogel, 1986) were conducted to determine whether the data follow normal or log-normal distributions. Due to the relatively low number of sampling points at Vallecola site, the PPCC test was not performed for this hillslope. In the PPCC test, if the correlation coefficient ($r$) between the data and standardized quantile for the specified distribution is smaller than the critical $r_C$ (derived by Looney and Gulledge, 1985), the null hypothesis $H_0$ ($H_0$: the data are drawn from the considered distribution) is rejected. A large significance level (i.e. 0.1) was applied to increase the power to detect non-normality.

Results are reported in Fig. 7a–d for Piramide and Emme during 2005 and 2006, respectively. As expected, the distribution at the 0–12 and 0–20 cm soil depths differed from the one at the surface. The range of skewness values decreases markedly from the shallow layer to deeper layers. For Piramide 2006, skewness ranges between $-1.3$ to $1.3$ at 0–6 cm; this range is reduced by 50% and 40% at 0–12 and 0–20 cm, respectively. Similar reductions are found for Emme 2006 and for the data collected during 2005. In general, surface distributions have a larger percentage of negatively skewed cases than deeper layers. This corresponds to the
higher soil moisture content observed at this layer, due to the
effect of precipitation (negatively skewed distribution are found
after rainfall events) and dew (negatively skewed distributions
are found also during the dry-down). More symmetric patterns,
with low skewness, are found for mid-range soil moisture con-
tents, while highly skewed distribution (generally with more
log–normal shape) are found at dry and wet conditions. This is con-
sistent with repeated observations that soil moisture distributions
become skewed and less variable as the mean approaches each
end-member state, i.e., either the residual water content or the sat-
urated water content (Famiglietti et al., 1999; Western et al.,
2002).

Seventy one per cent of the distributions were well de-
scribed by both the normal and log–normal distributions. Nor-
mal and log–normal distributions were appropriate for 82%
and 74% of the datasets, respectively. Neither a normal nor a
log–normal distribution was appropriate for 16% of the
datasets.

Correlation between point measurements at different depths

Correlation between point measurements at different depths
are shown as scatter plots in Fig. 8a–g for the three hillslopes (in
Vallecola only 0–6 cm depth and 0–12 cm depth soil moisture data

Figure 7. Skewness and appropriate probability density functions (PDF) according to PPCC test for: (a) Piramide 2005; (b) Emma 2005; (c) Piramide 2006 and (d) Emma 2006. For each sampling time, the appropriate PDF is reported versus time.
for 2005 could be compared). Scatter plots suggest that a positive correlation exists between moisture contents over the three depths for the three sites, and that the relationships are generally relatively good, particularly over Piramide and Emme where $R^2$ ranges from 0.67 to 0.8.

Relatively large correlations are reported for relationships between 0–6 and 0–12 cm depth, and for 0–12 and 0–20 cm depth (with $R^2$ ranging from 0.76 to 0.84). The relationships between 0–6 cm depth and 0–20 cm depth are generally characterised by lower correlations, with $R^2$ equal to 0.67 and 0.73 for Piramide and Emme, respectively. All these linear relationships are significant with $a = 0.01$. Considering the persistent high soil moisture at 0–6 cm depth, the observed high correlations between 0–6 cm and 0–12 cm depth and between 0–12 and 0–20 cm depth are likely to be related to the effects of vertical percolation of soil water, with a reduced and/or lagged moisture response at depth to the soil moisture dynamics at the surface. In general, the correlations among the various soil depths found in this study are larger than those reported in other studies. For instance, Wilson et al. (2003) report $R^2$ values ranging between 0.0077 and 0.27 among moisture data collected over 0–6 and 0–30 cm for three sites in Australia. Better correlations in our study are likely due to the higher moisture content in our test sites, which facilitates the vertical redistribution of soil moisture resulting in higher correlations between different depth profiles.

**Relation of soil moisture to topographic variables**

In this section we analyse how well different topographic variables can predict the spatial pattern of soil moisture and how this explanatory power changes with soil depth and with time. The variables, which are shown in Table 4, were chosen because they represent flow processes through the hillslopes or spatial variation in the meteorological forcing. The relationship between surface soil moisture and topography was first explored by examining the correlation coefficients between time-averaged moisture content and the various topographic variables (Table 5). On going investigation shows that the moisture fields investigated here are characterised by considerable time stability. Therefore, the time-averaged soil moisture fields may be deemed representative of the spatial variability found in instantaneous soil moisture fields. Inspection of Table 5 shows that logarithm of specific area and wetness index are in general characterised by positive correlation; slope always exhibits negative correlation, whereas cos (aspect) and solar radiation are not characterised by a specific behaviour. In absolute values, the behaviour of slope is similar to that of upslope area and

---

Figure 8. Scatter plots for 2005 and 2006 soil moisture data as a whole at different depths. Piramide: (a) 0–6 cm vs. 0–12 cm; (b) 0–12 cm vs. 0–20 cm and (c) 0–6 cm vs. 0–20 cm. Emme: (d) 0–6 cm vs. 0–12 cm; (e) 0–12 cm vs. 0–20 cm and (f) 0–6 cm vs. 0–20 cm. Vallecola: (g) 0–6 cm vs. 0–12 cm.
tently with observations above, the correlations increase with behaviour noted for the other hillslopes (not shown here). Consistency at different soil depths and slope, wetness index and solar radiation is shown for Emme in Fig. 9a and b for years 2005 and 2006, respectively. Emme is reported because it exemplifies the content at different soil depths and slope found in more gentle terrain and less humid conditions (Western et al., 1999). This supports the view that, even under almost ideal conditions, there is limited topographic control on soil moisture patterns. The flow-related topographic variables (slope, contributing area and wetness index) are generally better related to the time-averaged soil moisture fields than radiation-related variables (cos (aspect) and solar radiation), and explain up to 42% of the time-averaged moisture spatial variation, defined as the correlation square. An exception is represented by the concave-convergent hillslope Vallecola, where we would have expected stronger relation with flow-related variables and instead radiation is the only variable able to capture a fraction of the spatial variability. The disruption of the potential flowpaths due to the landslide scar located on this hillslope may provide an explanation for the low explanatory power of upslope area and slope found here. Only for Emme, the sampling depth is characterised by a clear behaviour, with correlations increasing with soil depth. This is not unexpected, since correlation should increase with increasing spatial variability and therefore (in the conditions considered here) with decreasing moisture content and with sampling depth.

The time history of the correlation coefficient between moisture content at different soil depths and slope, wetness index and solar radiation is shown for Emme in Fig. 9a and b for years 2005 and 2006, respectively. Emme is reported because it exemplifies the behaviour noted for the other hillslopes (not shown here). Consistently with observations above, the correlations increase with sampling depth and a specular pattern is shown for slope and wetness index, with solar radiation characterised by an intermediate behaviour. In general, the correlation coefficient time pattern is not revealing any relation with precipitation events and dry-down periods, with the exception of the initial period of 2006, which is characterised by low soil moisture contents and by reduced correlations for 0–12 and 0–20 cm sampling depths.

### Relationship between spatial mean value and spatial variability

The relationship of instantaneous soil moisture spatial mean and spatial variability is explored over the three hillslopes and soil depths. Fig. 10a–h shows the relationship between the spatial mean soil moisture and the spatial standard deviation by soil depth and hillslope. While considerable scatter exists, a negative relationship between the mean soil moisture and the standard deviation can be recognised for mean water contents exceeding 25–30%, while a transition to a positive relationship is observed with drier conditions (for Emme and Piramide). This cannot be observed for Vallecola, because for this site only data from 2005 (characterised by high soil moisture) are available. Overall, soil moisture variability shows the highest values at moderate moisture conditions (25–30%) and reduced values for wetter and drier conditions for all depths and for the two sites. With further increasing drying, the spatial variability starts to decrease. Thus, a positive relationship between mean values and standard deviation can be identified for water content less than 25% over Piramide and Emme. These findings are consistent with previous results from Famiglietti et al. (1999), Western et al. (2003), Ryu and Famiglietti (2005) and Choi and Jacobs (2007), who reported observations of unimodal shape of the variability–mean relationship. These results also agree with theoretical considerations about the double-bounded character of the soil moisture distribution, which is physically bounded by porosity and wilting point, i.e. the point below which plants are no longer able to extract water from the soil matrix. Typically, double-bounded distributions exhibit minimums of variance at the boundary and a peak in variance between the boundaries (Western et al., 2003).

Empirical field observations revealed that when soil dried starting from saturation, a few sites remained wet resulting in an enhanced spatial variability. Examination of spatial patterns (not reported here for the sake of brevity) shows that the spatial organization of these sites was markedly homogeneous for the three depths, suggesting similar processes at work for the three soil depths. Sites remaining wet as soil dried were generally characterised by high clay content.

### Table 4

Summary of the topographic variables considered in the analysis.

<table>
<thead>
<tr>
<th>Topographic variable</th>
<th>Formula</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Slope</td>
<td>tan β = √(f² + f)</td>
<td>Mitasova and Hofierka (1993)</td>
</tr>
<tr>
<td>cos (aspect)</td>
<td>cos (aspect)</td>
<td>Mitasova and Hofierka (1993)</td>
</tr>
<tr>
<td>Contributing area</td>
<td>D-inf</td>
<td>Tarboton (1997)</td>
</tr>
<tr>
<td>Topographic wetness index</td>
<td>ln (e^tan p)</td>
<td>Beven and Kirkby (1979)</td>
</tr>
<tr>
<td>Solar radiation</td>
<td>Computed by summing direct and diffuse insolation originating from the unobstructed sky directions</td>
<td>Fu and Rich (2002)</td>
</tr>
</tbody>
</table>

### Table 5

Spatial Pearson’s correlations coefficients of hillslope-averaged soil moisture and surface characteristics (only common sampling times are considered). Significant correlations (with p < 0.05) are bold.

<table>
<thead>
<tr>
<th>Site</th>
<th>Sampling soil depth</th>
<th>Slope</th>
<th>Cos (aspect)</th>
<th>Log contributing area</th>
<th>Topographic wetness index</th>
<th>Solar radiation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Piramide</td>
<td>0–6 cm</td>
<td>-0.42</td>
<td>-0.03</td>
<td>0.39</td>
<td>0.58</td>
<td>0.14</td>
</tr>
<tr>
<td></td>
<td>0–12 cm</td>
<td>-0.26</td>
<td>0.10</td>
<td>0.31</td>
<td>0.45</td>
<td>-0.10</td>
</tr>
<tr>
<td></td>
<td>0–20 cm</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Emme</td>
<td>0–6 cm</td>
<td>-0.37</td>
<td>-0.13</td>
<td>0.17</td>
<td>0.25</td>
<td>0.12</td>
</tr>
<tr>
<td></td>
<td>0–12 cm</td>
<td>-0.52</td>
<td>-0.03</td>
<td>0.47</td>
<td>0.52</td>
<td>0.21</td>
</tr>
<tr>
<td></td>
<td>0–20 cm</td>
<td>-0.65</td>
<td>-0.18</td>
<td>0.47</td>
<td>0.56</td>
<td>0.44</td>
</tr>
<tr>
<td>Vallecola</td>
<td>0–6 cm</td>
<td>-0.03</td>
<td>0.03</td>
<td>-0.38</td>
<td>-0.32</td>
<td>0.15</td>
</tr>
<tr>
<td></td>
<td>0–12 cm</td>
<td>0.41</td>
<td>0.34</td>
<td>-0.12</td>
<td>0.03</td>
<td>0.60</td>
</tr>
<tr>
<td>Piramide</td>
<td>0–6 cm</td>
<td>-0.33</td>
<td>0.01</td>
<td>0.51</td>
<td>0.64</td>
<td>0.14</td>
</tr>
<tr>
<td></td>
<td>0–12 cm</td>
<td>-0.11</td>
<td>0.07</td>
<td>0.23</td>
<td>0.30</td>
<td>-0.13</td>
</tr>
<tr>
<td></td>
<td>0–20 cm</td>
<td>-0.32</td>
<td>0.11</td>
<td>0.38</td>
<td>0.53</td>
<td>0.03</td>
</tr>
<tr>
<td>Emme</td>
<td>0–6 cm</td>
<td>0.00</td>
<td>-0.09</td>
<td>-0.07</td>
<td>-0.05</td>
<td>-0.21</td>
</tr>
<tr>
<td></td>
<td>0–12 cm</td>
<td>-0.12</td>
<td>-0.04</td>
<td>0.23</td>
<td>0.22</td>
<td>-0.11</td>
</tr>
<tr>
<td></td>
<td>0–20 cm</td>
<td>-0.44</td>
<td>-0.21</td>
<td>0.39</td>
<td>0.44</td>
<td>0.20</td>
</tr>
</tbody>
</table>
Standard deviation ranges between 2.8% and 7.8% over Piramide and Emme for 0–12 and 0–20 soil depth, while it ranges between 1.1% and 6.5% for 0–6 cm soil depth. The range of values is much less over Vallecola, for which only 2005 data are available; due to this reason, any relationship between statistical moments is hardly identifiable over this hillslope. Values from Piramide and Emme are in the range of values reported by other researchers. For instance, Choi and Jacobs (2007) reported that maximum standard deviations of the soil moisture were 8.0% and 10.1% over the two study fields, whereas Martinez-Fernandez and Ceballos (2003) reported a value of 10% for the peak standard deviation of soil moisture data collected over a 1285 km$^2$-wide region in Spain.

Interestingly, data from the literature show that the maximum standard deviation increases slightly with the area of the sampled region. This means that variability in the atmospheric forcing and topography have a relatively small impact on maximum spatial standard deviation of soil moisture.

A less scattered relationship between soil moisture variability and field mean becomes evident when standard deviation is scaled by the field mean (Fig. 11a–h). The coefficient of variation (called CV, hereinafter), calculated by the ratio of standard deviation of soil moisture to mean soil moisture, versus mean soil moisture (called $\Theta$, hereinafter) is well characterised by a negative linear fit: $CV = A\Theta + B$. It is interesting to note that the fitted negative linear relationships identified over Piramide and Emme for 0–12 and 0–20 cm soil depth are indistinguishable from a statistical point of view (with significance of 0.05). This shows that the topographic characteristics of the specific hillslopes have a negligible effect on the relationship between coefficient of variation and average values.

The use of a linear relationship between the coefficient of variation and the mean value leads to a parabolic relationship between mean soil moisture and spatial standard deviation. This allows to derive the position and the value of the soil moisture variability peak (Table 6). Inspection of results reported in Table 6 shows that the inferred soil moisture variability peaks correspond to mean soil moisture values ranging between 23% and 28%, with standard deviation ranging between 5–6% and 7–8%. In this case, influences of site and soil depth on inferred characteristics of the variability peaks are relatively minor. Interestingly, the soil moisture observations are organised around two quite different wet and dry preferred states. This agrees with the previous observation that the general soil moisture PDF for the study sites exhibit a bimodality character. Evidences for existence of preferred states and bimodality has been attributed to a number of causes, including seasonality in the meteorological conditions in combination with the non-linearity of the soil moisture response (Teuling et al., 2005), soil moisture–precipitation feedback

Figure 9. Time series of correlation coefficients between soil moisture at different depths and topographic variables for: (a) Emme 2005 and (b) Emme 2006.
mechanism (D’Odorico and Porporato, 2004), existence of local and non-local controls on soil moisture spatial patterns (Grayson et al., 1997). Our data indicate that the soil moisture variability peak lies in the region of rapid transition between the two preferred states, which are characterised by opposite trends in the relationship between the spatial mean and the spatial variability. These data also show that the hillslope topography and the soil depth have a negligible effect on these trends. We speculate that one major control on the observed trends is represented by the soil properties and their spatial variability.

Soil moisture dynamics model

In order to assess the influence of the soil moisture retention characteristics and their spatial variability on the soil moisture variability, we applied a version of the soil moisture dynamics model developed by Teuling and Troch (2005). The advantage of this model approach is that the number of parameters is small, while the parameters still reflect observable properties. The model developed by Teuling and Troch (2005) accounts for variations in soil and vegetation properties, but not for redistribution due to lateral flow. It is therefore interesting to evaluate the quality of the model simulations in the steep topography landscape considered in this study. Models of similar complexity have been shown to correctly simulate the root zone soil moisture dynamics under different climatic conditions (Albertson and Kiely, 2001; Teuling et al., 2005), but not in complex topographic settings. The equations of the model are given as follows (Teuling and Troch, 2005).

Soil moisture balance

The point scale soil moisture dynamics is spatially unconnected. Vertical redistribution of soil moisture is assumed to occur instantaneously (at the daily time step). The daily water balance for a number of independent soil columns is solved following:

$$\frac{d\theta}{dt} = \frac{1}{Z} (T - S - E - R - q)$$

(1)

where $\theta$ is the volumetric soil moisture, $Z$ is the depth of the root zone, $T$ the throughfall (i.e., the rainfall that is not intercepted by the vegetation), $S$ the root water uptake, $E$ the evaporation from the soil surface, $R$ the saturation excess runoff (i.e., the part of $T$ that causes oversaturation of the soil) and $q$ the deep drainage. Lateral flow is assumed to be negligible in the root zone.
Deep drainage is computed using the Campbell (1974) parameterization

\[ q = k_s \left( \frac{\theta}{\varphi} \right)^{2b+3} \]

where \( k_s \) is the saturated hydraulic conductivity, \( b \) is the pore size distribution parameter, \( \varphi \) is the porosity.
Root water uptake

The vertically integrated root water uptake is thought to be proportional to a maximum transpiration rate $E_p$, a soil moisture stress function $\delta(h)$ and a function accounting for spatially variable response of unstressed transpiration to atmospheric boundary layer conditions (Al-Kaisi et al., 1989). It is computed as follows:

$$S = f_r \delta(h) |1 - \exp(-c\xi)| E_p$$

where $f_r$ is the root fraction in the layer of depth $Z$, $\delta$ a soil moisture stress function, $c$ is a light use efficiency parameter, $\xi$ is the leaf area index. The factor $|1 - \exp(-c\xi)|$ allows to account for leaf area index following Al-Kaisi et al. (1989).

Soil moisture stress

Soil moisture stress is modelled as

$$\delta = \max \left[ 0; \min \left( 1; \frac{\theta - \theta_w}{\theta_c - \theta_w} \right) \right]$$

where $\theta_w$ is the wilting point and $\theta_c$ is the critical soil moisture content, which defines the transition between unstressed and stressed transpiration.

Leaf area index

Leaf area index is modelled with a spatial and temporal component.

Figure 12. Simulated and measured mean soil moisture at 0–20 cm depth for 2006: (a) Piramide and (b) Emme.

Figure 13. Relationship between mean soil moisture at 0–20 cm depth and standard deviation for simulated and observed data for 2006: (a) Piramide and (b) Emme.
\[ \xi = \xi_{\max} \left[ (1 - c_1) \sin \left( \frac{2 \pi \text{DOY} - c_2}{c_1} + \frac{\pi}{2} \right) \right] \]  

where \( \xi_{\max} \) is the local maximum of \( \xi \), and \( c_1 \) indicates the seasonal development of \( \xi \).

The size of the interception reservoir is taken proportional to the leaf area index, with a proportionality constant equal to 0.2 mm, and the reservoir is assumed to evaporate every day. Since \( 0 \leq \theta \leq \phi \), where \( \phi \) is the porosity, \( R \) equals \( T \) for \( \theta = \phi \), and is zero for \( \theta < \phi \). Bare soil evaporation is assumed to be small in comparison to the root water uptake over the entire soil profile. Drainage is calculated using Darcy’s law with the unit-gradient assumption.

\[
ks = \frac{\text{TransportPerDOY}}{\text{UnitGradient}}
\]

The simulated distribution agrees well with the observed one, particularly for higher values of soil moisture. Fig. 14a and b shows the relationship between spatial mean and spatial standard deviation, with overestimation of observed spatial mean leading to underestimation of observed spatial standard deviation (and vice versa).

The simulated and observed relationship between spatial mean and spatial standard deviation is reported in Fig. 13a and b for the two hillslopes. The simulated distribution agrees well with the observed one, particularly for higher values of soil moisture. Fig. 14a and b shows the relationship between spatial mean and spatial coefficient of variation for the two hillslopes. The simulated patterns are similar to those observed for both hillslopes. Both are well characterised by the negative linear fit, with very similar values for the slope of the line (Table 8). The regression lines computed for the simulations show a systematic underestimation of the observed variability (bias in the line interception), which again may be explained by the error-free character and the integration in time of the simulations. A quantitative evaluation of these specific effects is on going and will be reported in future papers.

The good fit between observed and simulated patterns obtained by this model, which does not consider topographic processes, and the similarity of the patterns obtained over the two hillslopes, characterised by markedly different topographic structures, support the hypothesis that the spatial mean–spatial variability relationship is essentially controlled by soil and vegetation properties. Moreover, these results indicate that a simple physically-based model can capture the soil moisture variability at the

\begin{table}[h]
\centering
\begin{tabular}{|l|c|c|c|c|c|}
\hline
          & \multicolumn{2}{c|}{Piramide} & \multicolumn{2}{c|}{Emme} & \\
          & A    & B    & A    & B    & \\
\hline
Observed  & -0.015 & 0.687 & -0.013 & 0.596 & 0.980 \\
Simulation& -0.014 & 0.610 & 0.897  & -0.013 & 0.557 & 0.911 \\
\hline
\end{tabular}
\caption{Regression relationship between the coefficient of variation and the mean soil moisture of observed and simulated values for the different soil depths in Piramide and Emme.}
\end{table}
hillslope scale even for the hillslopes considered here, characterised by steep slopes and shallow soils.

Conclusions

Analysis of soil moisture data at 0–6, 0–12 and 0–20 cm depth is presented for three experimental hillslopes with contrasting steep relief and shallow soil depth in the Dolomites (central-eastern Italian Alps). The hillslopes have mainly convex, planar and concave topographic structure. The data have been collected during two summer seasons (2005 and 2006) with different precipitation distribution.

Analysis of soil moisture data shows that different physical processes control the distribution of soil moisture at the three soil depths, with a marked effect of dew on the 0–6 cm soil depth layer. Due to this effect, the surface layer is usually wetter and shows lower space-time variability than deeper soil layers, particularly during dry-down.

As expected, the distribution at the 0–12 and 0–20 cm soil depths differed from the one at the surface. The range of skewness values decreases markedly from the shallow layer to deeper layers. In general, surface distributions have a larger percentage of negatively skewed cases than deeper layers. This is due to the combined effect of precipitation (negatively skewed distribution are found after rainfall events) and dew (negatively skewed distributions are found also during the dry-down) which increases the moisture content of the surface layer. The soil moisture distribution tends to evolve from positively skewed to negatively skewed shape as mean soil moisture changes from the dry-end to wet-end members. More Gaussian patterns are found for mid-range soil moisture contents, while data at dry and wet conditions tend to follow a log-normal distribution.

Scatter plots drawn for the whole data set and the analysis of the correlation coefficient suggest a good persistence of patterns along the soil profile: high correlations between 0–6 cm and 0–12 cm depth and between 0–12 and 0–20 cm depth were observed. Considering the persistent high soil moisture at 0–6 cm depth, the observed correlation patterns are likely to be related to the effects of vertical percolation of soil water, with a reduced and/or lagged moisture response at depth to the soil moisture dynamics at the surface.

Examination of correlation between soil moisture fields and topographical attributes shows that, notwithstanding the steep relief and the humid conditions, terrain indices are relatively poor predictors of soil moisture spatial variability. The slope and the topographic wetness index, which are found here the best univariate predictors of soil moisture spatial variability, the slope and the coefficient of variation of soil moisture. It is interesting to note that the fitted linear relationships identified for 0–12 and 0–20 cm soil depth are indistinguishable from a statistically point of view (with significance of 0.05). This shows that topographic structure has a negligible effect on the relationship between relative variability and spatial mean values.

The use of the linear relationship between mean and relative variability leads to infer soil moisture properties corresponding to the variability peaks. The soil moisture variability peaks correspond to mean soil moisture values ranging between 23% and 28%, with standard deviation ranging between 5–6% and 7–8%. The influence of site and soil depth on the characteristics of the peaks is relatively minor. The soil moisture observations are organised around two quite different wet and dry preferred states. This agrees with the observation that the general soil moisture PDF for the study sites exhibit a bimodality character.

A soil moisture dynamics model proved to successfully capture the soil moisture variability at the hillslope scale. The model reproduced relatively well the observed time series of the spatial mean and of the spatial standard deviation water content data at 0–20 cm depth. Moreover, the relationships between simulated soil moisture spatial mean and corresponding variability (measured by standard deviation and by the coefficient of variation) were remarkably similar to the observed one.

The good fit between observed and simulated patterns obtained by this model, which does not consider redistribution processes controlled by topography, and the similarity of the patterns obtained over the two hillslopes, characterised by markedly different topographic structures, support the hypothesis that the spatial mean–spatial variability relationship is essentially controlled by soil and vegetation properties, at least over the studied sites.

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