Hillslope-to-valley transition morphology: New opportunities from high resolution DTMs

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A B S T R A C T

The search for the optimal spatial scale for observing landforms to understand physical processes is a fundamental issue in geomorphology. Topographic attributes derived from Digital Terrain Models (DTMs) such as slope, curvature and drainage area provide a basis for topographic analyses. The slope–area relationship has been used to distinguish diffusive (hillslope) from linear (valley) processes, and to infer dominant sediment transport processes. In addition, curvature is also useful in distinguishing the dominant landform process. Recent topographic survey techniques such as LIDAR have permitted detailed topographic analysis by providing high-quality DTMs. This study uses LIDAR-derived DTMs with a spatial scale between 1 and 30 m in order to find the optimal scale for observation of dominant landform processes in a headwater basin in the eastern Italian Alps where shallow landsliding and debris flows are dominant. The analysis considered the scaling regimes of local slope versus drainage area, the spatial distribution of curvature, and field observations of channel head locations. The results indicate that: i) hillslope-to-valley transitions in slope–area diagrams become clearer as the DTM grid size decreases due to the better representation of hillslope morphology, and the topographic signature of valley incision by debris flows and landslides is also best displayed with finer DTMs; ii) regarding the channel head distribution in the slope–area diagrams, the scaling regimes of local slope versus drainage area obtained with grid sizes of 1, 3, and 5 m are more consistent with field data; and iii) the use of thresholds of standard deviation of curvature, particularly at the finest grid size, were proven as a useful and objective methodology for recognizing hollows and related channel heads.

1. Introduction

The morphology of alpine headwater basins is strongly influenced by erosion processes. The relationship between landforms and erosion processes has been analyzed based on the relationship between slope and drainage area (Hack, 1957; Tarboton et al., 1989; Montgomery and Foufoula-Georgiou, 1993; Tucker and Bras, 1998; Montgomery, 2001), because among parameters derived from a DTM (Digital Terrain Model), slope and drainage area are deemed to be pertinent for studying overall erosion dynamics (Willgoose et al., 1991; Willgoose, 1994; Ijjasz-Vasquez and Bras, 1995).

Montgomery and Foufoula-Georgiou (1993) suggested a partitioning of the landscape into drainage and slope regimes that include hillslopes, unchanneled valleys, debris flow-dominated channels, and alluvial channels. Fig. 1 shows a schematic illustration of this partitioning with two inclusions in a log–log drainage area–slope relation at about 10⁻⁴ km² and 10⁻¹ km². The first inclusion is related to the hillslope-to-valley transition, while the second correlates well with the transition from debris flow-dominated colluvial channels to gentler channels flowing over alluvial deposits. These transitions most likely depend on climate, uplift rates, rock strength, and the history of the fluvial system (e.g., glaciation). Ijjasz-Vasquez and Bras (1995) analyzed a DTM and identified four scaling regimes in the slope–area diagram that depict the change from diffusive to fluvial sediment transport processes. Tucker and Bras (1998) analyzed slope and drainage area to identify topographic thresholds defining erosion/landsliding processes. Stock and Dietrich (2003) interpreted the curved shape of the relation between slope and area on log–log diagrams as the topographic signature of valley incision by debris flows. Threshold criteria have also been proposed by other researchers to identify the extent of a channel network using a slope-scaling diagram. Two general methods have been used to simulate network sources with DTMs: a constant critical support area (e.g., O’Callaghan and Mark, 1984; Band, 1986; Mark, 1988; 1989; Tarboton et al., 1989, 1991) and a slope-dependent critical support area (e.g., Montgomery and Dietrich, 1992; Dietrich et al., 1993). The first approach considers a DTM cell from a river basin to be part of the channel network if its contributing area is larger than the defined threshold value. For many soil-mantled landscapes, the second approach is more appropriate both theoretically and empirically in defining the extent of channel networks (Montgomery and Foufoula-Georgiou, 1993).
of a slope-dependent contributing area that supports a channel head has been the object of both theoretical and field studies (Montgomery and Dietrich, 1988, 1989, 1994; Dietrich et al., 1993; Montgomery, 1994, 1999). Although this threshold is similar to Horton’s critical distance, it accounts for overland saturation flow and pore pressure-induced landsliding.

All these topographically-based analyses are strongly affected by the accuracy and resolution of the DTM. Various researchers have investigated the effect of DTM resolution on landscape representation (Montgomery and Foufoula-Georgiou, 1993; Zhang and Montgomery, 1994; Dietrich and Montgomery, 1998; Wilson and Gallant, 2000; Schoorl et al., 2000; Claessens et al., 2005; Tarolli and Tarboton, 2006). However, less attention has been paid to systematic effects of different DTM resolutions and accuracies on geomorphological models, and to the most appropriate scale for observing geomorphic processes.

Montgomery and Foufoula-Georgiou (1993) found that in moderately steep topography, a DTM finer than 30 m is required to accurately identify the hillslope-to-valley transition. Their results suggested that the transition from divergent to convergent elements occurs at a contributing area/unit contour length of 30–50 m, being equal to the hillslope length inferred from high-resolution 2 m DTMs generated from low-altitude stereo aerial photographs. Zhang and Montgomery (1994) examined the effect of DTM grid size on land surface representations and hydrologic simulations, observing that grid size influenced physically-based models of runoff generation and surface processes. Their analysis suggested that a 10 m grid size offered a rational compromise between data resolution and volume. Tarolli and Tarboton (2006) noticed that a very high-resolution DTM may lower the performance of an infinite-slope stability model, because slopes derived from such a DTM may no longer represent the reference critical slope in the model.

Airborne Laser Swath Mapping technology (ALSM), also known as LiDAR (Light Detection And Ranging) provides high-resolution DTMs to permit better land surface representation (Ackerman, 1999; Kraus and Pfeifer, 2001; Briese, 2004). The DTMs may cover large areas with grid sizes smaller (0.5–1 m) than previously possible (Slatton et al., 2007). A LiDAR-derived DTM can dramatically improve model performance and the detail of the results, benefiting land management.

LiDAR data have already been used for studying landslide morphology and distribution (McKean and Roering, 2004; Glenn et al., 2006), depositional features on alluvial fans (Frankel and Dolan, 2007), and channel bed morphology (Cavalli et al., 2008), as well as for the numerical modelling of shallow landslides (Tarolli and Tarboton, 2006), headwater channel network analysis (Vianello et al., 2009), and objective extraction of valley and channel networks (Lashermes et al., 2007). Despite such development of LiDAR applications, some essential issues concerning the spatial scale of observing geomorphic processes...
remains to be explored: 1) To what degree do different grid resolutions affect the partition of the landscape into channel and slope domains? 2) How does a high-resolution LiDAR-derived DTM affect the analysis of sediment transport processes in a headwater alpine basin? and 3) Does the spatial distribution of morphometric parameters obtained from LiDAR data offer a tool for recognizing hollows and channel heads?

The main purpose of this paper is to answer these questions using the relation between slope, drainage area and curvature values obtained from LiDAR-derived DTMs with different resolutions (1–30 m). We also surveyed 30 channel head locations in the field to validate our findings.

2. Study area

The study area is a small (4.4 km²) wilderness basin in Carnia, a tectonically active alpine region in northeast Italy (Fig. 2). The elevation ranges from 834 to 2075 m a.s.l with an average of 1530 m. The slope angle is 34.7° in average and 74.1° at maximum. The area has a typical rainy climate of the Eastern Italian Alps with short dry periods. Recorded annual precipitation ranges from 1300 to 2500 mm with an average of 2200 mm. Precipitation occurs mainly as snowfall from November to April; runoff is usually dominated by snowmelt in May and June. In summer, flash floods with large sediment discharge and debris flows are common. Vegetation covers 91% of the area and consists of forest stands (64%), shrubs (19%), and mountain grassland (17%); whereas bare ground (9%) consists of landslide scars (8%) and bedrock outcrops (1%).

The head of the basin has a uniform bedrock lithology of acidic sandstone and some vegetation such as green alder (Alnus viridis (Chais) DC), the dominant species between 1600 and 2000 m a.s.l. The area is also characterized by moraines with vegetated talus deposits, and calcareous, calcareous-marly, and arenaceous rock formations. The basin has lithological and physiographical conditions frequently observed in Carnia, and detailed information on the topography (including LiDAR), channel network, and land use data are available.

Within the study area, shallow landslide scars amount to 0.35 km², about 8% of the total area. All the mapped landslides are classified as shallow translational ones and have generated several debris flows in the last few years. The study area is a typical alpine debris-flow dominated catchment (Tarolli and Tarboton, 2006).

3. Data and methods

Field surveys were conducted in the study area during the summers of 2006 and 2007 in order to map the locations of channel heads. Fig. 2 shows the channel head locations numbered as progressively surveyed in the field. The 30 channel heads completely lack shrubs and medium–tall vegetation, permitting us to locate and measure channels correctly. Some channel heads located in inaccessible areas were not considered. The catchment was systematically walked along all drainage lines up to the catchment divide in order to locate and map the channel heads with an accuracy of a few centimetres using DGPS. The channel head or first-order stream head was defined as the point at which non-confined divergent flows on the hillside converge to a drainage line with a well-defined flow path, i.e., the upstream limit of concentrated flow (Dietrich and Dunne, 1993; see the example of channel head #7 in Fig. 3). The point of transition from hillslope to channel was quite distinct, and in most cases occurred over a distance of a few metres. The local slope of each channel head was measured as the slope of the surface 2 m along the flow direction. The drainage area of each channel head was determined by tracing catchment boundaries using 1–m contour lines derived from the LiDAR data for ground. Table 1 shows the statistics of the slope and drainage area of the channel heads. The LiDAR data were collected in snow-free conditions during November 2003, from a helicopter flying about 1000 m above ground, using ALTM 3033 OPTECH and Rollei H20 Digital cameras. The flying speed was 80 kn h⁻¹, the scan angle was 20°, and the scan rate was 33 KHz. The average density of the survey points was greater than 2 points/m². The first and last returns corresponding to vegetation and ground were filtered, resulting in an irregular density of ground returns with an average of 0.26 points/m² (4 m²/point). The ground point density for areas without continuous vegetation canopy, such as grassland, landslides, and bedrock outcrops, was in average 0.9 points/m² (~1 m²/point).

Table 1

<table>
<thead>
<tr>
<th>Channel</th>
<th>Slope (m m⁻¹)</th>
<th>Area (m²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.54</td>
<td>3467</td>
</tr>
<tr>
<td>2</td>
<td>0.53</td>
<td>1333.9</td>
</tr>
<tr>
<td>3</td>
<td>0.36</td>
<td>861</td>
</tr>
<tr>
<td>4</td>
<td>0.28</td>
<td>1521.64</td>
</tr>
<tr>
<td>5</td>
<td>0.17</td>
<td>1202.5</td>
</tr>
<tr>
<td>6</td>
<td>0.34</td>
<td>1364.58</td>
</tr>
<tr>
<td>7</td>
<td>0.20</td>
<td>504.29</td>
</tr>
<tr>
<td>8</td>
<td>0.23</td>
<td>538.47</td>
</tr>
<tr>
<td>9</td>
<td>0.36</td>
<td>601.7</td>
</tr>
<tr>
<td>10</td>
<td>0.33</td>
<td>3615</td>
</tr>
<tr>
<td>11</td>
<td>0.69</td>
<td>511.75</td>
</tr>
<tr>
<td>12</td>
<td>0.64</td>
<td>3197.7</td>
</tr>
<tr>
<td>13</td>
<td>0.74</td>
<td>540.28</td>
</tr>
<tr>
<td>14</td>
<td>0.58</td>
<td>1586.74</td>
</tr>
<tr>
<td>15</td>
<td>0.69</td>
<td>7775.2</td>
</tr>
<tr>
<td>16</td>
<td>0.60</td>
<td>10482.7</td>
</tr>
<tr>
<td>17</td>
<td>0.63</td>
<td>128.6</td>
</tr>
<tr>
<td>18</td>
<td>0.51</td>
<td>2571.6</td>
</tr>
<tr>
<td>19</td>
<td>0.80</td>
<td>1153.5</td>
</tr>
<tr>
<td>20</td>
<td>0.76</td>
<td>1257.47</td>
</tr>
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<td>21</td>
<td>0.78</td>
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<tr>
<td>22</td>
<td>0.62</td>
<td>1388.25</td>
</tr>
<tr>
<td>23</td>
<td>0.87</td>
<td>1441.7</td>
</tr>
<tr>
<td>24</td>
<td>0.72</td>
<td>7950.7</td>
</tr>
<tr>
<td>25</td>
<td>0.94</td>
<td>5916</td>
</tr>
<tr>
<td>26</td>
<td>0.77</td>
<td>9668.9</td>
</tr>
<tr>
<td>27</td>
<td>0.67</td>
<td>30551.8</td>
</tr>
<tr>
<td>28</td>
<td>0.86</td>
<td>4635</td>
</tr>
<tr>
<td>29</td>
<td>0.50</td>
<td>96680</td>
</tr>
<tr>
<td>30</td>
<td>0.76</td>
<td>5018.1</td>
</tr>
</tbody>
</table>

Minimum: 0.17, 128.6
Maximum: 0.94, 96680
Arithmetic mean: 0.58, 6956.95
Geometric mean: 0.53, 2221.33
Median: 0.63, 1481.67
Standard deviation: 0.21, 17907.51
The ground elevation points were interpolated into DTMs with different grid sizes from 1 to 30 m, using an algorithm with a spline function in the ESRI TOPOGRID tool. This approach can limit the occurrence of pits, producing hydrologically correct DTMs (Hutchinson, 1989). For curvature analysis (Section 4.3), however, we used the DTMs before pit filling, in order to obtain curvature not biased by artificial adjustments.

The accuracy of LiDAR data has been thoroughly investigated in the literature (Huising and Gomes Pereira, 1998; Baltsavias, 1999; Shrestha et al., 1999; Reutebuch et al., 2003; Alharthy et al., 2004; Hodgson and Bresnahan, 2004; Hodgson et al., 2005; Wechsler, 2007). The LiDAR dataset used for this study had already been verified in the field. Barilotti et al. (2006) compared the LiDAR-derived DTM for the study area with a corresponding DTM with an accuracy of 0.02 m, from a topographic ground survey using a pole mounted reflector and a total station with automatic target tracking. Both DTMs were obtained using the same interpolation method (spline) and parameter settings. The LiDAR DTM showed an average absolute vertical error of 0.324 m.

Fig. 4. Logarithmic diagrams of local slope versus contributing area for the a) 1 m, b) 3 m, c) 5 m, d) 10 m, e) 20 m, and f) 30 m DTMs based on binned and averaged data. The vertical black line shows the slope–area reversal at the hillslope-to-valley transition. The vertical dashed gray lines show the other scaling regimes with respect to the changes in negative gradient.
a mean relative error of 0.02 m, and an RMSE of 0.434 m. The largest error was found in areas under dense forest canopies, while smaller error (<0.1 m) was found in open areas.

4. Results and discussion

4.1. Scaling regimes of slope vs. area

Fig. 4 shows the relationship of slope versus log-bin averaged area for different DTM grid sizes (1–30 m). Extraction of slope and drainage area was performed using the TauDEM ArcGIS extension (http://www.engineering.usu.edu/dtarb/taudem). We used the D∞ flow direction algorithm (Tarboton, 1997) for the calculation of the drainage area, because of its advantages over the methods that restrict flow to eight possible directions (D8, introducing grid bias) or proportioned flow according to slope (introducing unrealistic dispersion). Slope was also evaluated using the D∞ method as the steepest outwards slope on one of eight triangular facets centered at each grid cell, measured as drop/distance, i.e. tan of the slope angle.

The slope–area data are based on the catchments rather than to single channel profiles. Slope values were averaged for each 0.25 log interval of drainage area. The amalgamation of many channel profiles

![Fig. 4](image)

![Fig. 5](image)

Fig. 5. Channel heads in relation to scaling regimes of local slope versus drainage area for the a) 1 m, b) 3 m, c) 5 m, d) 10 m, e) 20 m, and f) 30 m DTM. The plots show the slope–area scaling regimes identified in Fig. 4.
across a catchment may lead to a smearing of trends and transitions, but removes uncertainties associated with the selection of individual channel profiles. The slope–area diagram for the highest grid resolution (1 m) (Fig. 4a) shows four regions with different scaling responses, i.e., trends with different gradients. In Region I, at small contributing area, the slope–area scaling has a positive gradient but it becomes negative in Region II. The negative trend is relatively indistinct in Region III, but becomes distinct in Region IV. These four regions are similar to those observed by Ijaz-Vasquez and Bras (1995). The change between Regions I and II is related to the change between hillslopes with diffusive sediment transport and concave unchanneled valleys (Fig. 1). Region I extends up to a contributing area of 17.5 m², which corresponds to 17.5 m in total hillslope length, because total hillslope length measured in raster is equal to the drainage area at the hillslope-to-valley transition divided by the pixel width (1 m in this case). The changes between Regions II, III, and IV are related to dominant sediment transport processes. The less distinct negative gradient in Region III is related to the dominance of debris flows and landslides (Montgomery and Foufoula-Georgiou, 1993) that led to valley incision (Tucker and Bras, 1998; Stock and Dietrich, 2003). Region IV is related to the dominance of alluvial channels. Although the four regions can also be observed in the 3, 5, and 10 m DTMs (Fig. 4b–d), only three regions are found in the 20 and 30 m DTMs, because Regions II and III are merged (Fig. 4e, f). The hillslope–to-valley transition (Regions I to II) occurred at drainage areas of 70, 160, 560, 1885 and 3770 m², or about 23.3, 31.9, 56.0, 94.3 and 125.7 m in total hillslope length, for DTM resolutions of 3, 5, 10, 20 and 30 m, respectively. Only hillslope lengths obtained from the 5 and 10 m DTMs are in the range reported in literature (Montgomery and Foufoula-Georgiou, 1993). The slope–area scaling of Region III demonstrates that the signature of debris flow/landslide dominated processes becomes more evident as grid size decreases, because the lower negative gradient observed in the 1, 3, 5, and 10 m DTMs seems to disappear in the coarser DTMs. Particular Region III in Fig. 4a shows a more pronounced debris flow/landsliding signature and demonstrates a clear shift toward higher slope values for a given area. This result is consistent with the finding of Stock and Dietrich (2003) that debris flow valley slopes become steeper, resulting in a more pronounced curvature in the slope–area diagram. The alluvial dominated area (Region IV) is also influenced by the change in DTM resolution because the smallest area of this region changes from 90,000 (30 m DTM) to 250,000 m² (1 m DTM). Although this change is significant, the decrease in resolution does not obscure the existence of the region.

4.2. Channel head location

The analysis of the slope–area relationship showed the capability of high-resolution LiDAR DTMs to depict the detailed change in the dominance geomorphic process. The hillslope–to-valley transition between divergent to convergent surfaces shown in the slope–area relationship does not necessarily correspond to channel head location (McNamara et al., 2006). Since the process of channel initiation is normally described as a threshold phenomenon on the slope–area relationship (Montgomery and Dietrich, 1992), we analyzed the distribution of 30 channel heads with respect to the scaling regimes of the relationship using DTMs with different resolutions. Table 1 provides a statistical summary of the channel heads studied. The channel head support area ranges from 128.6 to 96,680 m², with an arithmetic mean of 6957 m², a geometric mean of 2221 m², and a median of 1481 m². The local slope ranges from 0.17 to 0.94, with a median of 0.63.

Fig. 5 shows the channel head locations in the slope–area diagrams and different scaling regimes (Fig. 4) for different grid resolutions. Fig. 5a shows the result from the finest DTM resolution (1 m). The channel heads are mostly confined to Region II, while only seven (about 23% of the total channel heads analyzed) are found in Region III. No channel heads are located in Regions I and IV, where hillslope processes and alluvial dominated channels prevail. These results suggest that most channel heads occur in Region II where channelized processes prevail. Although a few channel heads occur in Region III where debris flow and landslide processes are dominant, these findings mostly agree with our field observation that all the channel heads are found where no shallow landslides and debris flows are documented. They are located in the transition zones between unchanneled areas near the catchment boundary and downslope areas with shallow landslides (Fig. 2). Fig. 5b (3 m DTM) appears similar to Fig. 5a, where eight channel heads are found in Region III and the others are in Region II, because the loss in detailed local morphology causes only minor shifts in the boundary lines between the regions. In Fig. 5c (5 m DTM), the distribution of channel heads in Regions II and III is similar to those observed in Fig. 5a, b, but one channel head (#17) is located in Region I. Fig. 5d (10 m DTM) shows channel head distribution across Regions I, II, and III. Five channel heads (#7, 8, 11, 13, and 17) are located in Region I, creating a higher possibility of misinterpretation about channel head distribution.

Fig. 5e, f (20 and 30 m DTMs) shows more channel heads in Region I. In Fig. 5f, 21 (70%) channel heads are found in Region I, pointing to a serious misinterpretation of channel head location. These results indicate that only DTMs finer than 5 m allow the correct identification of dominant geomorphic processes in relation to channel head location.

4.3. Landform curvature

In addition to the slope–area scaling regimes, landform curvature (i.e. Laplacian of elevation, $C = \nabla^2 z$) is also a useful measure for the interpretation of dominant landform processes (Lashermes et al., 2007; Istanbulluoglu et al., 2008). The curvature is given by:

$$C = \nabla^2 z = \left(\frac{\partial^2 z}{\partial x^2}\right) + \left(\frac{\partial^2 z}{\partial y^2}\right),$$

where $z$ is the elevation, $\frac{\partial^2 z}{\partial x^2}$ is the planar curvature representing the degree of divergence or convergence perpendicular to the flow direction, and $\frac{\partial^2 z}{\partial y^2}$ is the profile curvature showing convexity or concavity along the flow direction. In general terms, divergent-convex landforms ($C > 0$) are associated with the dominance of hillslope/diffusion processes, while convergent-concave ($C < 0$) landforms are associated with fluvial-dominated erosion. Table 2 shows the minimum, maximum, mean, and standard deviation ($\sigma$) of $C$ for the study area. The largest range is obtained from the finest 1 m DTM, and it decreases with decreasing DTM resolution due to smoothing effects. The mean curvature is close to zero regardless of DTM resolution.

Fig. 6 provides curvature values for each channel head at different DTM resolutions. Black bars in Fig. 6a show mean curvature within a 5-cell moving window. It shows a similar trend to the curvature for landform using DTMs with different resolutions. In general, divergent convex landforms have a positive curvature, which is consistent with our field observations. The curvature values range from $-4.542$ to $0.546$ for the finest 1 m DTM, and the mean curvature is $-0.225$ with a standard deviation of $0.073$. As the grid size increases, the curvature values decrease, indicating a smoothing effect.

### Table 2

<table>
<thead>
<tr>
<th>DTM resolution (m)</th>
<th>$C_{\min}$</th>
<th>$C_{\max}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>-4.542</td>
<td>0.546</td>
</tr>
<tr>
<td>3</td>
<td>-0.399</td>
<td>0.225</td>
</tr>
<tr>
<td>5</td>
<td>-0.225</td>
<td>0.073</td>
</tr>
<tr>
<td>10</td>
<td>-0.102</td>
<td>0.038</td>
</tr>
<tr>
<td>20</td>
<td>-0.056</td>
<td>0.022</td>
</tr>
<tr>
<td>30</td>
<td>-0.037</td>
<td>1.5 × 10^{-12}</td>
</tr>
</tbody>
</table>

$C$ values for each channel head at different DTM resolutions.
decreasing grid resolution, a progressive shift toward negative values (divergence/hillslopes) and a progressive decrease of curvature values occur (Fig. 6). Almost all channel heads are located in convergent/channelled zones for the 1, 3, and 5 m DTMs. For the coarser (10 to 30 m) DTMs, however, an increasing number of channel heads are located in divergent/hillslopes zones, and for the 30 m DTM, 17 channel heads (about 57% of all channel heads) are found in divergent zones (Fig. 6f). These analyses indicate that a high resolution DTM is required to obtain correct curvature values depicting convergent zones. This result is consistent with the field observations that the width of the valley floor near the channel heads surveyed was between 1 and 3 m.

4.4. Channel head density ratio

Table 3 provides a statistical summary of curvature C and channel head density within a range of curvature threshold values. For each DTM resolution, we considered eight threshold values of curvature calculated as multiples (1–3 times) of curvature standard deviation ($\sigma_C$). This methodology provides objective curvature classes for each DTM resolution, which was useful in comparing the effectiveness of different curvature maps in recognising hollow morphology. The convergent and divergent portions of the study area are almost equally distributed for all the DTMs, with only a slight (2%) shift toward negative values at the finest resolution. For each curvature

Fig. 6. Curvature values for the channel heads computed from the a) 1 m, b) 3 m, c) 5 m, d) 10 m, e) 20 m, and f) 30 m DTMs. In Fig. 6a, gray numbers on the x-axis show the channel head numbers in Fig. 2. The black histograms are average values of curvature evaluated within a 5-cell moving window. In Fig. 6c–f, the values of curvature are $10^{-1}$ times the original value.
map corresponding to each DTM grid resolution and a range of curvature thresholds, the channel head density was determined as the ratio of the number of channel heads within the range to the area of the range. Then the ratio of the channel head density was derived using the following equation:

\[
\text{Channel head density ratio} = \frac{\text{CH}_{\text{thr}}}{\text{area}_{\text{thr}}} \quad \text{(2)}
\]

where \(\text{CH}_{\text{thr}}\) is the number of channel heads in areas within a threshold range of curvature (\(\text{area}_{\text{thr}}\)), and \(\text{CH}\) is the total number of channel heads within the basin area (\(\text{area}\)). This ratio was used to quantify the performance of each curvature map to discriminate the convergence of hollows. High values of the ratio correspond to better performance. A similar concept was used by Tarolli and Tarboton (2006), who employed the density ratio of the most likely initiation points to measure the effectiveness of stability index maps in discriminating landslide scars. Table 3 indicates that for the highest positive curvature threshold \(>3 \sigma_C\) obtained from the 1 m DTM, the density ratio is 53.28, which is the highest among values from all DTMs. For curvature \(>3 \sigma_C\), 60% of the channel heads were found in a small portion (1.13%) of total area, while all channel heads fell within curvature \(>\sigma_C\). The second highest density ratio (14.8) was found for the 3 m DTM and curvature \(>3 \sigma_C\). In this case, about 97% of the channel heads fell within curvature \(>\sigma_C\). The third highest density ratio (6.58) was found for the 3 and 5 m DTMs and curvature \(>2 \sigma_C\). For the 5 m DTM, about 93% of the channel heads fell within curvature \(>\sigma_C\), and for the 10 m DTM, 73%. Despite the decrease in the ratio, curvature is still effective in representing hollow morphology at these resolutions. By contrast, the 20 and 30 m DTMs were too coarse to resolve the channel hollow morphology in detail because of lower density ratios. For the 30 m DTM, the highest density ratio was only 2.47 (\(>3 \sigma_C\), \(\leq 2 \sigma_C\)), and 60% of channel heads were located in divergent (hillslope) areas. These results suggest that the 1 m DTM is the most suitable to represent hollow morphology. In order to confirm this, five curvature maps (Fig. 7a–e) obtained from the 1, 3, 5, 10, and 20 m DTMs for the 23, 24, 25, and 26 channel heads were interpreted because they present the largest difference in curvature values. Fig. 7a shows the effectiveness of the 1 m DTM in representing local morphology. The convergent areas of hollows and channels are clearly represented with the highest positive values of curvature. It is also interesting to note that curvature \(>3 \sigma_C\) \((>0.348)\) corresponds to a significant part of the channel network below each channel head. The maps for the 3, 5, and 10 m DTMs still well represent channel hollows. Most cells with higher positive curvature \((>0.126, 0.056, \text{and} 0.015\) for the 3, 5, and 10 m DTMs, respectively) are found along the main channels and correctly depicts the convergence of these areas. In contrast, the map from the 20 m DTM shows the ineffectiveness of coarser
DTMs in representing local morphology, because hollow morphology is poorly represented due to smoothing.

5. Conclusion

This paper discusses the potential of high resolution LiDAR data in the analysis of hillslope and valley morphology. The study area is a typical steep alpine catchment with frequent debris flows and landslides. The LiDAR technology provided unprecedented high-quality digital terrain data that allowed realistic topographic representations. Scaling regimes of local slope versus drainage area, spatial distribution of curvature, and channel head locations served as the basis for the analysis. The results demonstrated that high resolution DTMs (1–5 m cell size) are required to better recognize local morphology and provide interpretation of the hillslope-to-valley transition based on the slope-area relationship. DTMs finer than 10 m
permitted the recognition of the topographic signature of valley incision by debris flows and landslides from the slope–area diagram, and this signature is most evident in the 1 m DTM but disappears in the 20 and 30 m DTMs. Curvature computed from a high resolution DTM facilitated the discrimination of convergent hollow morphology around the surveyed channel heads. This discrimination capability was measured using the ratio of channel head density within threshold range identified as n-times the standard deviation of curvature. We found that the ratio was significantly higher for the 1 m DTM, enabling the most realistic channel head identification. In contrast, analyses carried out with coarser DTMs led to misinterpretation inconsistent with field data.

The channel head identification using the standard deviation of curvature provides an objective methodology for extracting channel networks. We can conclude that topographic signatures from high resolution LiDAR data better discriminate hillslope elements from river networks reducing uncertainties associated with the use of global criteria such as area or the area–slope threshold relationship.

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