DEFORESTATION INDUCES SHALLOW LANDSLIDING IN THE MONTANE AND SUBALPINE BELTS OF THE URBIÓN MOUNTAINS, IBERIAN RANGE, NORTHERN SPAIN

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ABSTRACT

In this study the spatial distribution of shallow landslides in the upper montane and subalpine belts of the Urbión Mountains (Iberian Range, northern Spain) was investigated, particularly in relation to the spatial organization of deforestation and land cover. The upper montane and subalpine belts have been deforested several times since the Neolithic Period, to enlarge the area of summer grasslands for feeding transhumant sheep flocks. Consequently, the timberline was lowered by 400–600 m, and increased the occurrence of severe erosion processes, particularly shallow landslides. This study shows that most of the landslide scars are in the summer grasslands area, and that a remarkable extent of the subalpine belt area has been subjected to mass movements. In addition to land use, the soil characteristics and topography help explain the development of conditions most favorable to landsliding. Shallow landslide susceptibility was highest in the upper parts of the slopes near the divides, in areas having slope gradients of 10–30° and deep soils with an increasing proportion of clay with depth. The landslides were clustered and not randomly distributed, and the causes

of this spatial distribution are discussed. The current trend of woody encroachment in the upper montane and subalpine belts, resulting from decreasing livestock pressure, will probably reduce the susceptibility of these areas to shallow landslides in the future.

KEYWORDS: shallow landslides; deforestation; subalpine belt; land cover changes; Urbión Mountains

1. Introduction

The subalpine and alpine belts of mountain ranges are extremely sensitive to environmental change (Ives and Barry, 1974; Price, 1981; Caviezel et al., 2014; García-Ruiz et al., 2015). Since Neolithic times, changes in land cover have occurred in many of the subalpine belts in European mountains (Galop and Jalut, 1994; Tasser and Tappeiner, 2002; Colombaroli et al., 2008; Vannière et al., 2011), and in the Iberian Peninsula ranges in particular (Lasanta and Vicente-Serrano, 2007; Améztegui et al., 2010; Bal et al., 2011; Cunill et al., 2012; Lasheras-Álvarez et al., 2013; Pérez-Sanz et al., 2013; García-Ruiz et al., 2016). Early changes in these mountain areas involved the substitution of forests by grasslands, done to promote summer pastures for feeding increasing numbers of livestock. This was the case in the Urbión Mountains (Iberian Range, Spain), where the presence of large transhumant sheep flocks explains the early deforestation of many of the hillslopes above 1500 m a.s.l. (García-Ruiz et al., 2016). Climatic fluctuations also directly affect the subalpine and alpine belts, where: (i) snow accumulation and melt change the extent and duration of the snowpack (López-Moreno, 2005; López-Moreno et al., 2009); and (ii) the treeline changes in altitudinal position, depending on temperature trends (Camarero et al., 2015). These changes condition geomorphic processes and the hydromorphological functioning of hillslopes and channels (Höllermann, 1985; Del Barrio and Puigdefábregas, 1987; García-Ruiz et al., 2010).

The occurrence of shallow landslides has been reported in mountain areas following deforestation, although few studies have been performed in the subalpine belt. This was the case in New Zealand during European colonization (Crozier and Preston, 1999; Glade, 2003), the Alps (Tasser et al., 2003; Caviezel et al., 2014), and the Pyrenees (García-Ruiz et al., 1990, 2010), where the result was a lowering of the solifluction belt (Höllermann, 1985), and changes in runoff generation (Alvera and Puigdefábregas,

1985; Lana-Renault et al., 2010, 2011), fluvial dynamics (Liébault et al., 2005; Gómez-Villar et al., 2014; Sanjuán et al., 2016), and the productivity of summer grasslands (García-Ruiz and Valero-Garcés, 1998).

Recent studies of Holocene and historical deforestation, and the recent woody encroachment in the Urbión Mountains, have demonstrated the complex history of land cover during the last few thousand years (García-Ruiz et al., 2016), and highlighted geomorphic processes (mainly shallow landslides) associated with deforested areas. This study investigated the spatial distribution of shallow landslides in the upper montane/subalpine belts of the Urbión Mountains, northern Spain, particularly in relation to deforestation and the spatial organization of land cover. The key finding of this study is the demonstration of a close relationship between early deforestation (since at least Neolithic times) and the occurrence of shallow landslides in an area where the presence of large transhumant sheep flocks has contributed to the maintenance of subalpine grasslands by impeding tree recolonization

2. The study area

The Urbión Mountains (maximum altitude 2228 m) are located in the northwestern part of the Iberian Range, Iberian Peninsula (Fig. 1). They are mainly composed of Upper Jurassic and Lower Cretaceous quartz conglomerate, red mudstone, and limestone. The alpine tectonics resulted in a monoclinal relief with an abrupt front facing north.

Weather stations, located on the southern face of the massif (between 1100 and 1150 m a.s.l.), have recorded a temperature increase in the last 60 years: the mean annual temperature at Vinuesa was 7.7°C for the period 1957–1965, and 9.6°C for the period 1990–2001 (García de Celis et al., 2008). On the northern face, the weather station at Valdezcaray (1620 m a.s.l.) recorded a mean annual temperature of 7.1°C for the period 1990–2008. Mean annual precipitation exceeds 900 mm above 1000 m a.s.l., and probably reaches 1500–1600 mm in the main divide, as estimated by Arnáez (1987) for the neighboring Sierra Demanda.

Plant cover is dominated by extensive *Pinus sylvestris* forests on the south-facing slopes, whereas the north-facing areas show a complex landscape with *P. sylvestris* and *Fagus sylvatica* forests, and grasslands and shrubs including *Calluna vulgaris*, *Erica cinerea*, *Vaccinium myrtillus*, and *Juniperus communis*. The area has a history of summer use for transhumant sheep herds involving two periods of intense activity, in

the 15th century and between the end of the 17th and the end of the 18th centuries. By the end of the 18th century the main two villages (Viniegra de Abajo and Viniegra de Arriba) accounted for almost 40,000 sheep in total (approximately 400 sheep per km²), whereas at present the total number of sheep is only 2,500. The human population has also declined in both villages, which together had 131 inhabitants in 2015. Viniegra de Abajo had 689 inhabitants in 1842, and 86 in 2015; Viniegra de Arriba had 418 inhabitants in 1842, and 45 in 2015.

The present study was carried out in the Ormazal Valley (2,513.8 ha) (Fig. 1), which ranges from 1,240 to 1,887 m a.s.l., and has a heterogeneous landscape from the lower montane belt (cultivated and abandoned fields) to the subalpine belt (grasslands and forests).

3. Methods

An inventory of shallow landslide was compiled by mapping all the landslide scars identifiable in the 2006 color orthophotos from the National Plan of Aerial Orthophotography, at a resolution of 0.5 × 0.5 m. This map was later validated in the field.

A map of land cover was also compiled by manual delineation using the same orthophotos. Various types of land use/land cover (LULC) were distinguished, including dense forest, open forest, shrublands, abandoned fields, reforestation, and summer grasslands. Eroded areas were also mapped because they are areas affected by intense erosion processes following deforestation, and may be related to the occurrence of shallow landslides. A set of topographic variables was derived from a digital elevation model at a spatial resolution of 18 × 18 m, and included: (i) altitude, (ii) slope gradient, (iii) slope aspect, (iv) distance to the hydrographic network, (v) solar irradiation, and (vi) topographic index (Beven and Kirkby, 1979).

Field studies involved measurement of the main features of the shallow landslide scars (width, length, depth), and the collection of soil samples to determine the pH, carbon and organic matter content, the distribution of grain sizes (percent sand, silt, and clay), and the liquid and plastic limits as well as the plasticity index according to the Casagrande test. The above analyses were undertaken for the A, B, and C soil horizons, and for soils affected and not affected by landslides. Analyses were performed at the Laboratory of the Pyrenean Institute of Ecology (CSIC).

Boxplots were used as an exploratory tool for assessing differences in soil characteristics among soil horizons, and between samples from landslide and no landslide areas, and the interactions between these two factors. ANOVA tests at the $\alpha = 0.05$ confidence level were performed to confirm differences in sand, silt and clay content and liquid and plastic limits between soil horizons, and between landslide and no landslide sampling sites. A general omnibus test was performed in a first stage to assess the main effects and interactions present in the data set, using the function `anova` in R (R Core Team, 2017). Pairwise comparisons were then performed for testing the significant effects found in the omnibus tests, using the function `pairwise.t.test` in R (R Core Team, 2017). When testing for the interaction between horizon and landslide/no landslide area, a correction of the confidence level was made by dividing by the number of factor levels (i.e., $\alpha = 0.05/3 = 0.017$).

To assess whether landslide areas had distinct topographic signatures, histograms of the topographic variables and the LULC types corresponding to landslide sites and the entire study area were compared. The Kolmogorov-Smirnov test was used for comparing the continuous (topographic) variables, and the Pearson's Chi-squared test was used for comparing the nominal (LULC) variables. The functions `ks.test` and `chisq.test` in R (R Core Team, 2017) were used for this purpose, and the significance level was set at $\alpha = 0.05$.

Multivariate logistic regression analysis was then used to assess for landslide susceptibility, and its dependence on topographic and land use factors. A stratified random sample of grid cells was selected, consisting of all the grid cells that contained a landslide and an equal number of cells having no landslides. The latter were selected using a randomized strategy that maximized the spatial separation among samples, and ensured similar proportions of LULC types that occurred in the entire area. The t-test was used to assess the significance of the covariates (topographic variables and LULC types), and the McFadden pseudo- R^2 statistic (McFadden, 1974) was used to compare alternative models. The functions in the package `mlogit` for R (Croissant, 2013) were used to perform this analysis.

Universal indicator Kriging (UIK), based on the covariates selected as being significant in the logistic regression analysis, was used to produce a map of landslide susceptibility. Although the model resulting from the logistic regression analysis could be used for this purpose, UIK was preferred because of the spatially-aggregated nature

of the landslide scars. UIK enables the influence of covariates (topography and LULC) and the spatial pattern of the landslide scars (how far they are from other scars) to be incorporated into the susceptibility model. If the spatial distribution of the landslides exhibited a non-negligible degree of spatial auto-correlation, as is often the case with landslides, the UIK process would predict higher landslide susceptibility in the vicinity of previous scars, as well as lower uncertainty associated with the prediction. The functions of the package `gstat` in R (Pebesma, 2004) were used for performing the UIK analysis.

The Clark and Evans (1954) R aggregation index was used as a means of determining the spatial structure of the landslides. The R statistic, which is calculated as the ratio of the observed mean nearest neighbor distance with respect to that expected from a (random) Poisson process of equal intensity, is a measure of how ordered or clustered a given point pattern is. Values of $R < 1$ indicate clustering, whereas values > 1 suggest ordering. The function `clarkevans.test` in the `spatstat` package in R (Baddeley and Turner, 2015; Baddeley et al., 2015) was used for determining the significance of the departure of R from 1 (null hypothesis), at the $\alpha = 0.05$ confidence level.

4. Results

4.1. Inventory and characteristics of shallow landslides

A total of 307 shallow landslide scars were identified and mapped (Fig. 2), representing an average frequency of 12.2 shallow landslides km^{-2} . They had an average width of 5.9 m and a depth of 40–70 cm; in some cases the landslides exceeded 1 m in depth, particularly in concave areas and in the lower parts of the slopes (Fig. 3).

Each scar consists of a semicircular crown with a talus of 1–2 m in length, caused by erosion of the scar. The sliding surfaces are usually covered by stones or consist of the underlying bedrock (Fig. 4). Thin remnants of the original soil tend to evolve towards gelifluction terracettes having pioneer vegetation on the steps. A few meters downslope of each scar a corrugated lobe usually occurs composed of accumulated landslip material, which has tended to evolve into solifluction lobes (Fig. 5), or is the origin of a new landslide scar, probably as a result of water accumulation (overland flow from the bare bedrock upslope) and instability at the front of the accumulation lobe.

Soil samples were collected at 37 sites from shallow landslide scars, in small ditches carved in the upslope sector of the landslide, and 23 sites in areas unaffected by landslides. The latter samples were collected with a hand-operated corer. At each sampling site, samples were collected at different soil depths, coinciding with the main soil horizons that were visually identified. A total of 37, 37 and 15 samples of the A, B and C horizons were collected, respectively, from the shallow landslide scars; and 23, 22 and 4 samples, respectively, were collected from the area not affected by landslides. The small number of C horizon samples in the second group was due to difficulties in reaching the C horizon with the manual corer.

The soils, from the Kastanozem group, are dark colored and have a granular structure in the A horizon, whereas the B and C horizons are brown–yellow and have a subangular to angular structure (Fig. 6). In general the soil depth exceeds 40 cm. The organic matter content was typically 6–9% in the A horizon, and 4–6% in the B horizon. Both horizons also had a high carbon content and C/N ratio. Conversely, the C horizon had relatively little organic matter content (< 2%). The pH was neutral in the A horizon and progressively increased towards the C horizon, showing the effects of lixiviation. Based on the scar profiles we concluded that the landsliding plane was usually close to the zone of contact between the C horizon and the bedrock.

The distribution of the grain sizes in the soils affected by shallow landsliding differed markedly among the horizons (Fig. 7). The A horizon tended to have a high sand content (70–80%), a low silt content (approximately 20%), and a very low clay content (< 5%). Consequently, most A horizon samples (50% of the total) were classified as loamy sands and sandy loams (23%). The B horizons had a similar silt (40–45%) and sand (40–45%) content and a low clay content (10–15%), so they were mainly identified as loams (66% of the total). The C horizons had a high silt content (45–50%) and very similar sand (25–30%) and clay (20–30%) contents, and were mostly classified as loams (81% of the total). Thus, a progressive change occurred from the A to the C horizon, with increasing proportions of clay and silt, and a decreasing proportion of sand.

Figure 8 shows a boxplot comparison of the percent sand, silt, and clay among horizons from landslide and no landslide areas. This shows clear differences in the percent sand, silt, and clay contents among soil horizons, with a progressive increase in

the proportions of silt and clay particles with soil depth ($A < B < C$). This pattern appeared to be stronger for samples from landslide areas than for those from no landslide areas, although a pairwise comparison showed significant differences only for the C horizon, which had a higher clay content in samples from the landslide areas. Differences in soil texture, as determined by the ANOVA test, were significant among the horizons. Generally no significant differences were found among samples taken in areas affected by and not affected by shallow landslides; the exception was the C horizon, for which a significantly higher clay content was found in the landslide-affected samples ($p = 0.0095$). The low number of samples of the C horizon in the “no landslide” group ($n=4$), however, recommends considering this result with caution.

The plastic and liquid limits were determined in 37 and 15 samples of the B and C horizons, respectively, in the landslide sites, and in 22 and 3 cases, respectively, in the no landslide sites. In one sample of the C horizon in the no landslide area the plastic behavior could not be found. The median values of the plastic and liquid limits, expressed as the volumetric water content at which the behavior of the samples changed, were (respectively) 0.48 and 0.39 for the B horizon and 0.40 and 0.33 for the C horizon, and 0.46 and 0.36 for landslide samples, and 0.36 and 0.44 for no landslide samples. Therefore, these soils can be considered only slightly plastic, but having a relatively high tendency to become liquid, for instance during large or intense rainstorms or following snowmelt. This characteristic favors the triggering of shallow landslides, with the sliding plane located within the C horizon or in the contact zone between the C horizon and the bedrock. The boxplot (Fig. 9) suggested differences between landslide and no landslide samples, but also among horizons from within landslide areas. In particular, a lower liquid limit in the C horizon was found in landslide samples relative to no landslide samples, as confirmed by the ANOVA, which showed significant differences between the landslide and no landslide samples ($p = 0.00048$ and 0.00002 for the plastic and liquid limits, respectively), and among soil horizons for the landslide samples considered alone ($p = 0.00002$ and 0.0088 for the plastic and liquid limits, respectively). Thus, lower plastic and liquid limits were found in the samples from landslide-affected areas, especially those from the C horizon. Again, these results cannot be considered conclusive, since they concerned the C horizon for which only a small sample size existed.

4.3. The topographic signature

Observations of the spatial distribution of landslide scars in the Ormazal Valley (Fig. 2) revealed several distinct features. Shallow landslide scars are present throughout the valley, but are clearly concentrated in certain places, including the southwestern and eastern parts. In contrast, only isolated scars are present in the southern, northwestern and central sectors. The scars are predominantly at mid to high altitudes.

These observations are reinforced in the histogram comparison of the main topographic characteristics of the landslide scars and those of the entire area (Fig. 10). Thus, most landslide scars (95.8%) were found above 1500 m a.s.l. (the highest frequency, 32.9%, was found between 1700 and 1800 m) on moderate and steep slopes (80.8% were present on slopes of 10–30°, and the highest frequency, 48.9%, was on slopes of 10–20°) facing east, southeast, southwest, and west. This was confirmed using the two-sided Kolmogorov-Smirnov test analyzing slope and elevation, which enabled rejection of the null hypothesis of equal distribution of the variables in landslide and no landslide areas ($D = 0.241$, $p < 0.0001$; and $D = 0.184$, $p = 0.0003$, respectively; $\alpha = 0.05$ confidence level).

The landslides had an aggregated spatial pattern, i.e. they were spatially clustered and not randomly distributed; this was confirmed using the Clark-Evans aggregation test ($R = 0.397$, $p = 0.002$).

4.4. Relationship of landslides to land cover

The map in Figure 2 suggests the prevalence of landslide scars in areas dominated by grasslands and eroded areas, and to a much lesser extent in areas dominated by shrublands. The last histogram on Figure 10 confirms this strong relationship with LULC. Of 307 scars, 193 (62.9%) were located on summer grasslands, 27 (8.8%) were in shrubland areas, and only 5 (1.6%) occurred in forests. The Pearson's Chi-squared test confirmed that, with respect to LULC, a significant difference exists in the number of landslide scars between landslide areas and the entire area (Chi-squared = 114.96, $df = 5$, $p < 0.001$).

The number of landslides in the summer grasslands represented an average occurrence of 28.9 landslides km^{-2} . 76 landslide scars (24.8%), however, occurred in eroded areas within the grasslands. Based on the combination of intact and eroded grasslands areas, a total of 269 shallow landslide scars were found in these areas, representing 87.7% of the total.

We investigated the interaction between topographical variables and land use, as the various land uses had very different topographic features, which affected the propensity for landsliding. Figure 11 shows a comparison of these features for a stratified sample of landslide and no landslide grid cells; in addition to elevation and slope gradient, Figure 11 also includes information on solar radiation and the topographic index.

Slope gradient is a major factor influencing the propensity of a soil to landsliding, as it determines the shear forces that act against the cohesional forces when the soil matrix is saturated. In the eroded and grassland areas the boxplots show slightly lower slope gradients in the landslide areas. Comparison of these land use areas (high prevalence of landslides) with dense forest areas (very low prevalence) is interesting because they were very similar in terms of slope gradient. Therefore, slope gradient was not the factor determining the slightly greater occurrence of landslides in grassland and eroded areas. The same is evident in the comparison of shrubland and open forest areas; although they have similar slope gradients they differed markedly in the occurrence of landslides.

With respect to solar radiation and the topographic index, clear differences occur among land uses. A gradient of decline in solar radiation was found from eroded areas to grasslands, shrublands, and forest areas. Relative to no landslide areas, landslide areas tended to receive higher levels of solar radiation, especially in the eroded areas and shrublands, but areas of similar solar radiation in the forest areas did not have a similar rate of landslide occurrence.

Very little difference was found among land uses with respect to the topographic index in areas having no landslides. The occurrence of landslides tended to correlate, however, with a higher topographic index for all land uses.

4.5. Regression analysis and maps of landslide susceptibility

The results of the multivariate logistic regression analysis confirmed the findings of the preliminary analysis. Thus, positive significant effects were found for elevation and solar radiation, and a strong negative effect was confirmed for the forest land cover (Table 1). Overall, the model was significant compared with a null model containing only the intercept (likelihood ratio test, $p < 0.001$), and achieved a McFadden pseudo- R^2 value of 0.178, which is considered good (although not exceptionally so).

Table 1. Regression coefficients of the multivariate logistic model. Provided are the estimated (mean) parameter and its standard error, and the value of the t statistic and its significance (Pr (> |t|)).

	Estimated	Std. Error	t value	Pr (> t)
(intercept)	-8.677	1.456	-5.960	< 0.001
forest	-2.978	0.548	-5.430	< 0.001
elevation	0.430	0.0920	4.669	< 0.001
radiation	0.177	0.06426	2.758	0.0058

The magnitude of the effects is summarized in Table 2. Compared with the 65.8% probability of landslide occurrence corresponding to average topographic conditions, the probability was reduced by 56.8% if the LULC was forest. An increase of 100 m in elevation represented an increase of 8.9% in the landslide probability, and an increase in solar radiation of 200 kW hr m⁻¹ yr⁻¹ represented an increase of 7.4% in the landslide probability. These probabilities were based on a stratified sample containing 50% each of landslide and no landslide grid cells. To obtain more accurate probabilities, these values need to be multiplied by the a priori landslide occurrence probability, which corresponds to the prevalence of landslide grid cells as a proportion of the total cells in the study area (0.0268). Comparison of the grid cells with and without a landslide scar showed that the logistic regression model made reasonable predictions, as most landslides had occurrence probabilities > 0.5, while most grid cells for no landslide areas had probabilities below that threshold (Fig. 12).

Table 2. Landslide probability corresponding to average topographic conditions, and change in probability as a function of land cover and topographic variables, according to the multivariate logistic regression model.

	Landslide probability
Mean topographic conditions: 1639 m, 1223 kW hr m ⁻² yr ⁻¹	0.658

Forest LULC	-0.568
Elevation: +100 m	+0.089
Solar radiation: +200 kW hr m ⁻² yr ⁻¹	+0.074

As a consequence of the marked spatial structure of the landslide scars, a geostatistical model (Universal Indicator Kriging, UIK) was used to produce a map of landslide susceptibility based on the variables that showed statistical significance in the logistic regression analysis. This spatial structure was confirmed by inspection of the semivariogram model fitted to the residuals of the logistic regression model (Fig. 13), which revealed strong spatial autocorrelation up to a distance of approximately 1000 m.

Only statistically significant variables from the logistic regression analysis were used for the UIK. Use of the UIK enabled computation of Best Linear Unbiased Predictions (BLUPs), which include the fixed effects of the topographic and land use variables, and the random effects of the observed landslide location were produced based on the UIK model (Fig. 14B). The BLUPs are different from Best Linear Unbiased Estimations (BLUEs), such as those obtained from the logistic regression analysis, which include only the fixed effects (Fig. 14A). The spatial configuration of landslide probabilities revealed the influence of the covariates elevation, solar radiation, and forest LULC, but also the marked influence of the location of previous landslide scars. Consequently, in the area surrounding one or several landslide scars, the landslide susceptibility was greater. Whereas BLUEs predict landslide probabilities up to 50%, and identify many ‘false positive’ areas, the BLUPs predicted landslide probabilities up to 90%, particularly in the areas adjacent to current landslides, whereas other topographically similar areas had much lower probabilities.

5. Discussion and conclusions

The upper montane and subalpine belts were deforested in many mountain areas to enlarge the area occupied by summer grasslands, as a way of feeding transhumant sheep flocks (Tinner et al., 2003; Miras et al., 2007; Gil-Romera et al., 2010; Guiguet-Covex et al., 2011; Rull et al., 2011; Roepke and Krause, 2013; Orengo et al., 2014; García-Ruiz et al., 2015). Many studies have confirmed the occurrence of deforestation processes since Neolithic times, based on the presence of charcoal in soil and lake

sediments. Thus, Montserrat (1992) found a thin 4,000 year-old charcoal layer in the sediments of Tramacastilla Lake (approximately 1,700 m a.s.l.) in the Gállego Valley (Central Pyrenees), which was attributed to a relatively early human-induced forest fire. A new and more significant forest fire occurred 1,000 years ago, coinciding with a general expansion of transhumance and the need to enlarge the subalpine summer grassland area. In the Urbión Mountains evidence also exists of early deforestation of the upper montane and subalpine belts, from 1,500 to 2,000 m (García-Ruiz et al., 2016). Below 1500 m, deforestation also affected most of the area, although the purpose was cultivation of cereals on south-facing slopes of the lower montane belt. In the middle of the 20th century, when the traditional management system collapsed, dense forests represented only 1.7% of the Ormazal Valley, shrublands occupied 33.9%, summer grasslands covered 36.2%, and eroded areas accounted for 16.1% (García-Ruiz et al., 2016). As eroded areas are located in the same topographic (altitude, slope gradient) and lithological context as summer grasslands, they have been considered to be the consequence of erosion (mainly shallow landsliding) in the summer grassland areas. Evidence of this is that the eroded areas retain small patches of soil intensively affected by shallow landsliding.

The occurrence of shallow landsliding in the subalpine belt is clearly related to the grassland domain (including intact and eroded grassland areas), confirming the interest of focusing geomorphological studies with a global perspective (García-Ruiz, 2015). Mapping of the spatial distribution of 307 shallow landslide scars in the Ormazal Valley showed that their occurrence is very much related to relatively intact summer grasslands, and the eroded areas within them (87.7% of the total), whereas only 1.6% occurred in the forest areas. Lorente et al. (2002), Beguería (2006), and Bathurst et al. (2007) reported the same relationship to land use in the Pyrenees. Landslide-prone land uses (grasslands) and those not prone to landsliding (dense forest) had similar topographic signatures (slope gradient, solar radiation, and topographic index), indicating that differences in topographic factors do not account for the occurrence of landsliding. We also showed that shrubland areas were more prone to landslides compared with open forest (not prone). This is a consistent demonstration of the effects of deforestation of the upper montane and subalpine belts on soil degradation in the long and short terms. The effect of land use on landsliding is related to the characteristics of the vegetation, which in turn is related to land use. Thus, whereas deep

rooted trees enhance soil stability in most cases, vegetation that does not have roots extending beyond the C soil horizon, including grasses and (to a lesser extent) shrubs, has decreased soil stability, because the slip surface is usually located between the C horizon and the underlying bedrock.

Relationships between land use and landsliding have been reported in previous studies. For instance, intense soil erosion occurred in the Tramacastilla Lake basin following deforestation 1,000 years ago (Montserrat, 1992), and shallow landslide activity has continued to the present, confirming long-term instability. For the Urbión Mountains no direct information exists on the erosion effects of deforestation during the Neolithic and Chalcolithic periods, the Bronze and Iron ages, or the Middle Ages (García-Ruiz et al., 2016). Its consequences are still evident, however, in the landscape, including thousands of shallow landslide scars in the summer grassland area (28.9 km²). These can reactivate during intense rainstorm or snowmelt events, as deduced from geomorphic activity in the subalpine belt of the Pyrenees (García-Ruiz et al., 2010). Thus, the extreme rainfall event of November 1982 in the Central Pyrenees triggered many shallow landslides in the subalpine belt; this event corresponded to a 25-year return period in the western valleys and more than a 200-year return period in the eastern valleys (García-Ruiz et al., 1983; Martí-Bono and Puigdefábregas, 1983). More recently, the extreme rainfall event of October 2012 in the Central–Western Pyrenees caused shallow landsliding in the area of summer grasslands above 1600 m (Serrano-Muela et al., 2015). Near the Urbión Mountains, a short and extremely intense rainstorm occurred in August 1988 following a forest fire in the neighboring Sierra Demanda. This triggered a number of confined shallow landslides and the formation of alluvial fans that threatened to block the Najerilla River (García-Ruiz et al., 2013). The rainfall threshold for initiating shallow landslides decreases following deforestation, and consequently ordinary rainstorms may be sufficient to trigger severe erosion events (Moody and Martin, 2001; García-Ruiz et al., 2013). Snowmelt can also trigger shallow landsliding, because of soil saturation and the low plasticity index of the soil. Soil saturation is a common phenomenon in the Sierra Demanda, where mudflows and debris flows often occur in May–June (Arnáez, 1987), and in the Pyrenees, as deduced from suspended sediment concentration peaks during the second half of the snowmelt season (Lana-Renault et al., 2011). The snowmelt season (May–June) in the subalpine

belt accounts for > 50% of the annual runoff, and 30–60% of the annual suspended sediment transport (Lana-Renault et al., 2011).

Shallow landslides in the Urbión Mountains occur from 1,200 to 1,700 m a.s.l. on slopes having gradients of 20–50° and relatively deep soils (> 40 cm) having an increasing proportion of clay from the surface to the C horizon. These are conditions similar to those in the subalpine belt of the Alps (Tasser et al., 2003) and the Pyrenees (Badía-Villas et al., 2002). In these areas the soils are unstable on slopes > 30°, even under dry conditions, but are stable on slopes < 15°, even under saturated conditions (García-Ruiz et al., 2010).

The shallow landslides showed clear clustering, with areas with multiple landslides alternating with areas having few or no landslide scars despite having similar characteristics in terms of land use and topography. Such clustering is often observed in landslide susceptibility assessments (e.g. Beguería, 2006). When generating a landslide susceptibility map, the modeler can decide whether to take this clustering into account (BLUP) or not (BLUE). In the latter case, the landslide probability in the susceptibility map will be overly spread across the territory, with large areas having a high susceptibility but a low density of landslides. Although these areas may be considered ‘false positives’, it has been usual to interpret them as areas that are highly susceptible to landsliding but which have not yet experienced the conditions necessary to trigger the process. Favoring this interpretation, BLUE susceptibility maps have been widely used in the landslide hazard literature. BLUP susceptibility maps, by contrast, show the landslide susceptibility concentrated in the vicinity of previous landslide scars, then rapidly fading with moving away from those areas. The degree of concentration of the susceptibility around the previous landslides is determined by the level of spatial aggregation of the scars. The idea behind BLUP maps is that new landslides tend to occur in the vicinity of existing ones.

In the absence of additional data it is difficult to favor one interpretation over the other, although in our opinion the BLUP map is to be preferred as it provides a better fit to the evidence (the observed landslides plus the spatial covariates). If new data were acquired, the map could be updated considering the new evidence. For instance, it could be useful to perform a new landslide inventory at a future time so as to determine whether new landslides occurred, and if so, where they occurred.

The clustered nature of the observed landslides could also be a consequence of differences in critical soil properties that affect the mechanical behavior of the soils but which were not included in the spatial analysis due to a lack of data. In our case, the susceptibility maps were based on land use / land cover and topographic information; however, analysis of a carefully designed soil sampling allowed the elucidation of possible differences in the soil characteristics between the areas with a high density of landslides and those with no landslides. Among these properties, the clay content and the liquid limit of the C horizon showed significant differences between landslide and non-landslide areas. This is an important result, since such differences may help explain the clustered pattern of shallow landslides in the area. Indeed, a higher clay content and a lower liquid limit (implying that less water is required to achieve liquid behavior) in the C horizon coincided with the landsliding plane in most cases, suggesting that these soil properties may help explain the spatial locations of the landslides. However, a fundamental question remains about the genetic relationship between landslides and soil properties: Can fundamental differences in soil properties explain why shallow landslides occur in certain areas and not others? Or does the occurrence of a landslide locally modify the soil properties, for instance by altering the hydraulic head gradient with respect to the topography, and hence increase the probability of occurrence of further landslides further up the slope? Our data are not suitable to resolve this issue. Answering this question would probably require repeated surveys or continuous monitoring of the hydraulic behavior of the soils, which is out of the scope of the current study. Also, the robustness of our results relevant to this issue was compromised by the small sample sizes available for assessing some combinations of factors, a problem that could only be solved by conducting additional surveys

These and other important uncertainties remain, including the possible evolutionary relationship between landslide clusters and the eroded grassland areas. They remain key issues that merit further research to achieve a more complete understanding of the morphodynamics of upper montane soils following land use changes. For instance, no data are available on the rate of extinction of landslide scars in the Urbión Mountains, because of the low quality of the 1956 aerial photographs compared with those from 2006. This is a critical problem in assessing hillslope stabilization and soil conservation (Beguería, 2006).

Another key uncertainty is the likely evolution of soils in the study area. The trend to woody encroachment in the upper montane and subalpine belts of the Urbión Mountains (Sanjuán et al., submitted) will probably reduce the occurrence of shallow landslides in the future. In addition, the consistent trend of decrease in snow accumulation will shorten the snowmelt and soil saturation periods (López-Moreno, 2005; López-Moreno et al., 2009; Navarro-Serrano and López-Moreno, 2017), which will also reduce the probability of occurrence of shallow landslides. It is also likely that the increasing occurrence of drought periods, suggested in some studies (Gouveia et al., 2016; Vicente-Serrano, 2016), and the recent trend of increasing evapotranspiration (Vicente-Serrano et al., 2017), will moderate the conditions triggering shallow landslides, including changes in the dynamics of the main fluvial channels resulting from decreasing sediment yield and runoff (Beguería et al., 2003; Zhang et al., 2008; Gómez-Villar et al., 2014; Buendía et al., 2016; Sanjuán et al., 2016).

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FIGURE CAPTIONS

Figure 1. Map of the study area, showing the location of the Ormazal Valley in the Urbión Mountains (left), and the topography of the Ormazal Valley (right).

Figure 2. Land cover and spatial distribution of shallow landslide scars in the Ormazal Valley.

Figure 3. Example of a hillslope affected by shallow landslides, with large landslides evolving into long strips of soil (in the foreground) and small scars. (Photo: Paz Coba-Pérez, June 2014).

Figure 4. Example of a large landslide that occurred in the contact zone between the soil and the bedrock. (Photo: Paz Coba-Pérez, June 2014).

Figure 5. A solifluction lobe near a shallow landslide scar. (Photo: Paz Coba-Pérez, June 2014).

Figure 6. The scar of a large landslide showing the A horizon, containing densely packed grass roots; the B horizon, which is almost free of roots; and the C horizon, which includes a gravel layer. (Photo: Paz Coba-Pérez, June 2014).

Figure 7. Grain size distribution for grassland soil samples from the Ormazal Valley, distinguishing the A (1, 4), B (2, 5), and C (3, 6) horizons. The dots indicate landslide samples, and the triangles indicate no landslide samples.

Figure 8. Boxplots showing the percent sand, silt, and clay in Ormazal Valley grassland soil samples for the three soil horizons (A, B, C), and for landslide and no landslide samples.

Figure 9. Boxplots showing the liquid and plastic limits and the plasticity index for Ormazal Valley grassland soil samples for the B and C soil horizons, and for landslide and no landslide areas.

Figure 10. Distribution of shallow landslides as a function of slope gradient, elevation, slope aspect, and land use/land cover, compared with the distribution of these variables over the entire study area. G: grasslands; E: eroded areas; S: shrublands; R: reforestation; DF: dense forest; A: abandoned fields; CF: clear forest.

Figure 11. Boxplots showing the elevation, slope gradient, solar radiation, and topographic index for a stratified random sample from landslide and no landslide sites, classified by land use. For each soil horizon, the pair of values indicates the sample size for landslide and no landslide samples, respectively (n).

Figure 12. Boxplot showing the multivariate logistic model probabilities predicted for grid cells having landslides (TRUE) and no landslides (FALSE).

Figure 13. Semivariogram of landslide scars.

Figure 14. Maps showing Best Linear Unbiased Estimations (BLUEs; top left) and Best Linear Unbiased Predictions (BLUPs; top right) resulting from Universal Indicator Kriging, and difference among them (bottom left).

Figure 1rev (Color)
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Figure 2rev (Color)

Legend

- *Shallow landslide scars*
- Topography
 - *Contour lines (100 m)*
- Land Use / Land Cover
 - Abandoned fields*
 - Pine forest, clear*
 - Pine forest, dense*
 - Reforestation*
 - Shrubland*
 - Grasslands*
 - Eroded areas*
- Reference system: ETRS89 / ETRS-TM30 (EPSG:3042)

Figure3

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Figure4

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Figure5

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Figure6
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Figure 7rev (Color)

●(1) ●(2) ●(3) ▲(4) ▲(5) ▲(6)

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Figure10

■ % Surface
■ % Shallow landslides

Figure11

Figure12

Figure 13rev (Color)

Figure 14rev (Greyscale)

Legend

○ Shallow landslide scars

Topography

— Contour lines (100 m)

UIK landslide probability

BLUE, BLUP	BLUP-BLUE
0.0	-0.4
0.1	-0.2
0.2	0.0
0.3	0.2
0.4	0.4
0.5	0.6
0.6	
0.7	
0.8	
0.9	

Reference system: ETRS89 / ETRS-TM30 (EPSG:3042)

Figure 2rev (Greyscale)

Legend

- *Shallow landslide scars*
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