Regional-scale high-plasticity clay-bearing formation as controlling factor on landslides in Southeast Spain

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ABSTRACT

Complex landslides in clay-bearing sediments are investigated in two moderate-relief regions of Southeast Spain. Both regions, more than 100 km apart show landslides affecting the same Flysch formation, which outcrops widely in the central and western Betic Cordillera along the contact between the External (South Iberian Domain) and Internal (Alborán Domain) zones. Intense rainfall episodes can be considered as the main triggering factor for slope failures in these two areas. We have chosen two landslides (Diezma and Riogordo landslides), one from each area of study, to investigate their morphological and geotechnical features in order to establish the relative importance of the different controlling factors. From a kinematic point of view, the two features studied in detail can be referred as to rotational failures, evolving downhill to slow earthflows. The movement was concentrated on several surfaces developed on a clay-rich layer mostly constituted by smectite. This clay mineral is of critical relevance to the mechanical behaviour of soils and Flysch-like formations, being very consistent at dry conditions, but rapidly losing its strength at wet conditions. Thus, softened smectite-rich clay layers with high water contents can approach the properties of a lubricant, which, in turn, can be critical for slope stability. In addition to their high plasticity, these clays have a high swelling potential, which can induce significant vertical overpressure, thus reducing even more the strength properties of the Flysch formation. In Southeast Spain, a region with a Mediterranean rainfall regime, slope stability can be seriously influenced by the presence of these smectite-rich clay layers in a formation of regional extent, as is the case of the Flysch formation. Therefore, this lithologically-controlled factor should be taken into account when evaluating landslide hazard in the Betic Cordillera.

Keywords: Complex landslides Flysch formation Smectite Swelling pressure Betic Cordillera

1. Introduction

Slope features related to long-term geomorphologic evolution can be considered as one of the main controlling factors of landmass movements. Topography controls the distribution of both elastic stresses in rock masses and pore-water pressures, both of which eventually can produce landsliding (Iverson and Reed, 1992). Apart from the influence of topographic and geologic features as controlling factors for slope instability, the most common landslide triggers are intense rainfall events (e.g. Crosta, 1998). Numerous researchers have shown the relationship between landslide occurrence and intense rainfall periods (e.g. Canuti et al., 1985; Azzoni et al., 1992; Ferrer and Ayala-Carcedo, 1997; Finlay et al., 1997; Polemio and Sdao, 1999;

Zeze et al., 1999). In the case of Mediterranean climatic environments, the relationships between landslides and extreme rainfall events have been extensively investigated (e.g. Guzzetti et al., 1992; Polloni et al., 1992; Crosta, 1998; Corominas and Moya, 1999). Although steep topography has classically been viewed as the main indicator of susceptibility to rockfall and deep-seated landslides, intense rainfalls produced during wet seasons can be the triggering factor in the initiation of slope failures. Rainfall events can also produce an increase in the rate of movement on landslides.

In the Iberian Peninsula, the areas with high susceptibility to slope instabilities concentrate along the Alpine orogenic belts, such as the Pyrenees and the Betic Cordillera. In the latter, slope morphology is well attested as a controlling factor in rockfall events and deep-seated failures (e.g. Thornes and Alcantara-Ayala, 1998). However, the most important landslides in terms of volume of mobilised material and economic losses have occurred in areas of low to moderate relief (Ferrer and Ayala-Carcedo, 1997). Most of the slope instabilities in

these low to moderate-relief regions are complex slow movements with an important component of flow (mainly earthflows). The controlling factors on these complex movements are related to the regional geology and the Mediterranean rainfall regime, their relationships with topography and seismicity being less straightforward (e.g. Ferrer and Ayala-Carcedo, 1997).

In the study of these complex movements, the behaviour of high-plasticity expansive soils and/or sedimentary formations submitted to high pore-water pressures can be considered especially interesting. Smectites, a clay mineral group with high plasticity and swelling potential, are the main constituents of these high-plasticity expansive soils and of some layers of sedimentary formations. These soils and clay-rich layers experience periodic swelling and shrinkage during alternate wet and dry periods (i.e. Day, 1994). Such cyclic swell-shrink movements can be considered to be critical for the stability of a natural slope. In some cases, the moisture content can be considered as a triggering mechanism of landslides in slopes made up of high-plasticity expansive materials (Yilmaz and Karacan, 2002).

In the Betic Cordillera, with a typical Mediterranean rainfall regime, high-plasticity expansive soils and certain sedimentary formations are associated with numerous slope instabilities. In this paper, we investigate the influence of high-plasticity clays on two complex landslides which caused important damages in highway and national roads, developed on a Flysch formation of the Betic Cordillera. Detailed geotechnical, geological, and mineralogical data have been obtained to identify the mechanism of slope failure. Conclusions can be drawn concerning the controlling factors and mechanical features of these landslides. We also analyze the role of the rainfall as the main trigger of these landmass movements. Findings of this investigation can provide guidelines for future studies on landslide hazard in Southeast Spain.

2. Geographical and geological setting

The two landslides at issue here (Diezma and Riogordo landslides, Fig. 1) are located in moderate-relief areas of the central sector of the Betic Cordillera (SE Spain). Both areas are affected by landslides of variable sizes and types, most of them probably representing reactivated ancient landslides. The first landslide is located just to the north of Riogordo (Fig. 1), in an area characterized by a high frequency of diverse landmass movements (Fig. 2) such as landslides, earthflows and rockfalls (terminology after Cruden and Varnes, 1996). The most important landmass movement in this area, the Riogordo landslide (Chacón Montero et al., 1988; Irigaray Fernández and Chacón Montero, 1991; Irigaray Fernández et al., 1991), occurred in January 1970 causing damage to the old national road Granada-Málaga (Fig. 3). Due to this landslide, the new Granada-Málaga highway does not follow the old trace, being located north of the Sierra Gorda (Fig. 1). The area of this landslide has been calculated in 75 ha, with a mobilised volume of debris around 4.5 hm³.

The second landslide is located north of Sierra Nevada, near Diezma (Figs. 1 and 4). This area is around 100 km to the NE of Riogordo but, interestingly, it has the same geological substratum (see below). The Diezma area, with moderate slopes, is close to the drainage divide between the Granada and Guadix topographic depressions (Fig. 1). A geomorphic analysis of this zone shows the presence of abundant old inactive landslides, some of them having historical reactivations (Fig. 4). The last substantial landslide in this area occurred in March 2001, producing damage to the A-92 highway Sevilla-Almería (Fig. 5). The area of this landslide has been calculated in 6.2 ha, with a mobilised volume of debris around 1.2 hm³.

Climate in the areas of the two landslides is typically Mediterranean with average annual rainfall of 500–800 mm and monthly

Fig. 1. Simplified geological sketch of the central part of the Betic Cordillera. The two areas studied are located with rectangles.

Fig. 2. Landslide inventory in the Riogordo area, modified from Irigaray Fernández and Chacón Montero (1991). This image looking north has been obtained by superposing an aerial orthophotograph and a digital elevation model.

Fig. 3. Map of the Riogordo landslide, slightly modified from Irigaray Fernández and Chacón Montero (1991). This view looking north has been obtained by combining an aerial orthophotograph and a digital elevation model.

Fig. 4. Landslide inventory in the Diezma area. This image looking north has been obtained by superposing an aerial orthophotograph and a digital elevation model.

average rainfall of 60 mm, mainly concentrated in a wet period (October–April). Mean annual temperatures range from 12.6 to 16.2 °C, with extreme temperatures oscillating between −2.2 and 40 °C.

From a geological point of view, the Diezma and Riogordo areas are located just at the boundary between the South Iberian Domain and the Alborán Domain of the Betic Cordillera (Fig. 1). The landslides are made up of quartzite, sandstone, limestone and dolostone blocks, embedded in high- to moderate-plasticity clays, silts and marls. All of these lithologies were originally part of a Flysch-type formation outcropping in an intermediate position between the South Iberian and the Alborán Domains. This formation represents a turbiditic sequence of Cretaceous–Lower Miocene age (Bourgeois et al., 1974). It outcrops continuously in the western and central parts of the Betic Cordillera and discontinuously in the eastern part, but it is always linked to the contact between the South Iberian and Alborán Domains. At all places where the stratigraphic sequence is complete, this formation includes a continuous sandstone stratum over a mass of clay and marl with imbedded limestone, dolostone and sandstone blocks. This Flysch formation was intensively deformed during the Alpine orogeny, thus acquiring a chaotic appearance. As a whole, the Flysch formation is structurally superposed over rocks belonging to both the South Iberian and Alborán Domains. Nevertheless, Jurassic dolostones and limestones from the South Iberian Domain locally thrust onto the Flysch Formation. In the Riogordo area, a N–S cross-section shows exactly such structural configuration with the Upper Jurassic carbonate formations (South Iberian Domain) thrust onto the sandstone–argillaceous–calcareous alternation of the Flysch formation (Fig. 6). To the south of the area represented in the cross-section, the Flysch formation thrust onto the shale, phyllites, sandstones and conglomerates of the Malaguide Complex (Alborán Domain). Despite the intense tectonism, the Jurassic carbonate rocks from the South Iberian Domain and the Flysch formation gently dip to the north on a regional scale. These Jurassic limestones constitute a karstic aquifer

with ground-water flow to the north according to both the location of the main springs in the northern side of the ranges and the north-directed dip of the strata.

In the Diezma area, Upper Jurassic limestones and dolostones belonging to the South Iberian Domain thrust onto the Malaguide complex (Alboran Domain). The thrust surface dips to the north (Fig. 7), while the stratification in the hanging wall is very steep. The carbonate rocks of the South Iberian Domain outcropping just to the north of the Diezma landslide constitute an unconfined karstic aquifer which drains to the south. The main springs are located at the thrust trace which superposes the carbonate rocks over the low permeability Flysch-like rocks. Peak discharges in the springs are delayed 24–48 h with respect to rainfall events. Increased discharge is maintained 15–40 days after the rainfall event.

3. Description of landslides

Among the landslides in the Riogordo area (Fig. 2), we have selected the most prominent one (Fig. 3) in order to analyze its morphological and mechanical features. This landslide is a complex movement with an area of 75 ha, maximum length of 2860 m, and maximum width of 550 m (Table 1). The Riogordo landslide displaced a huge volume of debris (approximately 4,500,000 m³), which includes large carbonate blocks (up to 125 m³) floating on a clay-rich matrix (Fig. 8). Some of these large blocks were displaced floating on the clay-rich matrix several hundred metres (Fig. 5). The travel angle (Cruden and Varnes, 1996) of this landslide is 10° (Table 1). This landslide can be classified as an earthflow (Vertical interval/Horizontal extension = 0.22; classification follows Brunsden, 1973). This was a complex failure, occurring in several stages. The first stage was a rotational failure affecting the Flysch formation and located at the base of a carbonate escarpment. This rotational failure was probably the trigger of a huge rockfall produced in the carbonate rocks close to the initial escarpment (Figs. 5 and 6). Finally, the rotational

Fig. 5. Map of the Diezma landslide. The photograph used for depicting the different parts of the landslide is a vertical aerial view taken after the slope movement, when the A-92 highway restoration works were just started.

slide, together with the rockfall, evolved downhill into an earthflow affecting clayey and silty rocks of the Flysch formation; in this last stage, large carbonate rocks coming from the rockfall were passively displaced on the surface of the earthflow. The speed of the earthflow was around 9 m/day, until reaching stability equilibrium after 150 days (Oteo-Mazo, 2003). The Riogordo landslide occurred during

an intense rainfall episode after 2 years with annual rainfall higher than average values (1969 and 1970; Fig. 9).

The Diezma landslide can also be considered as a complex movement, affecting an area of 6.2 ha, with maximum length of 510 m and maximum width of 205 m (Table 1). Field observations clearly show two different parts in the landslide (Fig. 5). In the head zone located in the old Granada-Almería road, several scarps (1–5 m high) with a dip between 15 and 70° are observed. These scarps correspond to shallow rotational slides developed on clay-rich rocks from the Flysch formation. The impermeable character of these shear surfaces favoured the development of ponds at the foot of the main scarp (Fig. 5). Lateral spreading of the landslide produced lateral and oblique bulges with tension cracks at the crests (Figs. 5 and 10). These cracks are subperpendicular to the bulges. The landsliding generated slickenside features and striations on a relatively thin layer (2–5 cm) constituted by clay-rich rocks (Fig. 11). Numerous decimetre-scale lateral cracks (Fig. 10) almost parallel to the movement direction were opened. The rotational slide evolved downhill into an earthflow. Thus, the lower half of this landslide can be considered as the accumulation zone of an earthflow. The dip of slide surface is lower than 13°, being the stability angle around 7° (Table 1). Test drillings confirm that the maximum thickness of the landslide is 25 m at the toe. The Diezma landslide took place on 18 March 2001 and, as observed in the cumulative rainfall chart (Fig. 9), the antecedent precipitation was higher than average daily values extracted from the 10 precedent years.

3.1. High-plasticity clay levels at the sliding surface

The two landslides studied here are characterized by movement concentrated on planar surfaces found by field and borehole investigation at the contact between a high-plasticity clay layer and the bedrock. The geotechnical properties of these high-plasticity clays are basically dependent on their mineralogical composition.

Mineral composition of the clay-rich layer has been determined by X-ray Diffraction (XRD) and confirmed by analytical Transmission Electron Microscopy (TEM). XRD analyses were carried out on both whole unoriented samples (Fig. 12) and oriented aggregates of the <2 µm fraction separated by centrifugation. Smectite identification was corroborated through ethylen-glycol treatment (Fig. 12, inset). TEM analyses were performed on dispersed samples deposited onto a Cu-grid. Individual mineral particles were in-situ chemically analysed by Energy Dispersive X-ray fluorescence (EDX). The two methods (XRD and EDX) showed smectite to be the dominant mineral, representing more than 95% of the clay fraction. EDX analyses on TEM allowed smectite mineral formula determination, yielding a typical composition of sedimentary smectite with beidellite and montmorillonite components. In the Diezma landslide, the smectite-rich levels are above a powdery white level, basically consisting of

Fig. 6. Simplified longitudinal cross-section of the Riogordo landslide.

Fig. 7. Longitudinal cross-section of the Diezma landslide. Vertical and horizontal scales are in metres.

pure calcite of very small grain size (Fig. 11). Scanning Electron Microscopy (SEM) images and TEM analyses of this white level show the existence, in addition to small-sized calcite, of beidellite-montmorillonite smectites at very low contents. From this observation, we postulate that these smectite-rich levels could have a secondary origin. In this regard, weathering processes, probably under saturated conditions, can contribute to leaching of carbonate minerals, thus producing a concentration of clay minerals.

Smectite content in clay soils can control their plasticity, compressibility and swelling pressure (Gillot, 1986). The strong interaction between clay minerals and water results from: a) the high specific surface area of the clay mineral; b) the structure of the clay minerals, and c) the polarity of the water molecule. The high specific surface area is a consequence of the small size and platy shape of the clay mineral grains. The weak ionic bond between filosilicate layers in the structure of the clay, due to the small interlayer charge, allows the bipolar water molecule to dissolve the weakly bonded cations in the interlayer space in proportions highly dependent on water availability. In the case of smectites, the specific surface area is around 760 m²/g. Thus, smectite-rich clays softened by increased water content can exhibit the properties of a lubricant and can be critical for slope stability. This assertion has also been made by Yilmaz and Karacan (2002), who showed that smectite-rich soils drastically change their geotechnical characteristics when saturated with water. These authors show that during the dry season, samples with a mean clay content around 25% (100% smectite in the <2 µm-fraction) have an average cohesion of

160 kPa, average internal friction angle of 20° and average density of 1.75 g/cm³. The same samples during the rainy season have very low shear strength, as revealed by the following average values: $c = 37$ kPa, $\phi' = 9^\circ$ and $\gamma = 1.92$ g/cm³. This change in the mechanical properties is of critical importance in understanding the controlling role that smectite-rich layers can represent in the generation of slope failure in areas with a mean slope of 10–12°, as in the case of the Diezma and Riogordo landslides.

3.2. Geotechnical characteristics

From a mechanical point of view, three different lithological levels can be recognized in the landslide bodies (Fig. 13). In the Diezma landslide these are: a) substratum, which is constituted by shales, phyllites, conglomerates and greywackes, all of them belonging to the Malaguide Complex (Alborán Domain); b) centimetre-thick smectite-rich layer; c) landslide debris of chaotic nature, which are mainly composed of sandstone and dolostone blocks embedded in a marl-clay matrix. The Riogordo landslide also has these three lithological levels (Fig. 13). In this case, the substratum is constituted by quartzite sandstones, marls and clays belonging to the Flysch formation. Above the substratum, the landslide debris are composed by dolostone, limestone and carbonate breccia blocks, belonging to the South Iberian Domain, involved in a marl-clay matrix of flyschoid origin. Between the substratum and the landslide debris, a relative thick layer mostly composed of smectite-rich clay, also appears. All of the sliding surfaces in this landslide have been found within this intermediate layer.

From borehole data (only available for the Diezma landslide) and field observation, the main failure surface of both landslides has been found to coincide with the smectite-rich layer. We have performed different geotechnical tests (plasticity, permeability, granulometry, oedometer, direct shear and uniaxial compression) on unweathered samples recovered from drill cores in the Diezma landslide.

The dry unit weight of the landslide debris is relatively high, ranging from 1.7 to 1.8 kN/cm³. The landslide debris are fairly uniform with regard to granulometric distribution and plasticity. They are composed of a fine-grained fraction representing about 84–95% of the total weight (diameter < 0.075 mm). A representative grain-size distribution is shown in Fig. 14. The more relevant geotechnical properties are shown in Table 2.

The landslide debris can be described as very rigid and hard, based on several resistance values obtained with the pocket penetrometer on drill cores and trench samples. The resistance values obtained vary

Table 1
Geometric parameters of the Diezma and Riogordo landslides.

	Diezma	Riogordo
Maximum longitude	510 m	2.860 m
Maximum width	205 m	550 m
Maximum depth	26 m	30 m
Scarp's width	20 m	510 m
Scarp's height	6 m	25 m
Scarp's radius	39.3 m	324 m
Area	6.2 ha	75 ha
Volume	1.2 Hm ³	4.5 Hm ³
Mean slope before sliding	13°	12°
Mean slope after sliding	7°	10°
Vim/Hem	0.12	0.17
D/L (depth/length)	0.05	0.01
Travel angle	6.8°	9.9°

Vim/Hem: vertical interval (m)/horizontal extension (m). Travel angle according to Cruden and Varnes (1996).

Fig. 8. Block of limestone breccia on a smectite-rich layer in the Riogordo landslide. Inset shows a detailed view of the base of the breccia and the underlying smectite-rich layer.

between 500 and 600 kPa, with an average value of 550 kPa. In the oedometer tests, low compressibility values were obtained for the landslide debris samples. The pre-consolidation pressure values are difficult to interpret, since they are clearly below those inferred from the geological history of these rocks. In order to obtain proper values for this property, the sample should be tested in a high-pressure oedometer. The landslide debris are generally characterized by medium to high plasticity (Fig. 15; the liquid limit ranges from 30 to 57, while the plastic limit averages 20, ranging from 18 to 28;

Table 2), medium to high slaking potential, low to medium swelling pressure, and a tendency to soften under the environmental conditions found in the field.

The smectite-rich layer is highly plastic and extremely expansive (Fig. 15). The Liquid Limit ranges from 58 to 92, while the plastic limit averages 28, ranging from 24 to 32 (Table 2). The dry unit weight for this layer is lower than in the landslide debris, with a value of 1.57 kN/cm³. N content is higher in this layer than in the landslide debris (17–23%, below the plastic limit). We have performed most of the geotechnical

Fig. 9. Plots of daily and cumulative rainfall data for the Diezma and Riogordo areas.

Fig. 10. Panoramic view looking north of the central sector of the Diezma landslide.

Fig. 11. a) Secondary scarp in the Diezma landslide. Inset shows a detailed view with slickenside and striations on the landslide surface. b) closer view of the same secondary scarp to show that the landslide surface is constituted by a smectite-rich thin clay layer, which, in turn, overlays a centimetre-thick fine-grained powdery calcite-layer. The block diagram illustrates this layer distribution. Inset on the lower right corner is an image of an unweathered sample obtained from a drill core. The smectite-rich layer appears with green colour in the photograph.

Fig. 12. Representative XRD plots of samples from the clay-rich layer of the Diezma and Riogordo landslides. Inset shows the results of the three tests (<2 μm clay fraction, glycolated sample and 100 °C-heated sample diffractograms) performed to prove that smectite is the dominant clay mineral in the clay-rich layer at both landslides.

tests on this thin smectite-rich layer, because its behaviour seems to be important for landsliding. The unconfined compressive strength, shear strength and compressibility parameters of this layer are tabulated in Table 2.

We have also carried out five consolidated and drained direct shear tests on samples coming from the smectite-rich layer and the overlying landslide debris (Fig. 14 and Table 2). Shear strength parameters are significantly different in the two materials involved in

the landslide. The landslide debris have relatively high shear strength values ($cp' = 0.3$ kPa and $\varphi' = 36^\circ$). In contrast, the clay-rich layer has very low shear strength values, especially in the case of the residual internal friction angle with values as low as 7° . These values for the clay-rich layer compare quite well with results of direct shear tests performed on pure clays and fully weathered marls (Cripps and Taylor, 1981; Moore, 1991). The residual friction angle for the landslide debris is also very low ($\varphi'r = 11\text{--}12^\circ$). These values coincide

Fig. 13. Simplified lithological sequence of the two landslides.

Fig. 14. a) Representative grain-size distribution of a sample from the landslide debris. b) Shear versus normal strength plot at peak and residual conditions obtained from several points of a direct shear test on landslide debris. Note that the plot at residual conditions is not a straight line, as it is usual in tests on overconsolidated materials. c) Representative plot resulting from a consolidation test on landslide debris.

Table 2
Geotechnical properties of the materials involved in the Diezma landslide.

Geotechnical properties	Max.	Min.	Number of tests	Remarks
% fines > 74	99.13	84.4	12	
Liquid limit	92.2	53	10	The highest value is for the smectite-rich clay
Plastic limit	32.5	16.6	14	The lowest value is for the smectite-rich clay
Plasticity index	65	30	10	The highest value is for the smectite-rich clay
% carbonates	15.5	6.34	4	
Natural moisture content	34.21	15.2	13	The highest value is for the smectite-rich clay
Specific gravity	2.34	2.34	1	–
Swelling pressure (kPa)	500	200	5	The highest value is for the smectite-rich clay
Dry unit weight (kN/cm ³)	1.53	1.91	12	The min. value is for the smectite-rich clay
Unconfined compression strength (kPa)	500	>600	50	–
Void ratio (e)	0.46	0.66		
Effective cohesion	0.3		4	–
Effective angle of internal friction	36°	28°	4	–
Residual angle of friction	12°	7°	5	Smectite-rich clay (7°)
Compression index, C _c	0.010	0.0119	2	–
Swelling index, C _s	0.0065	0.006	2	–

with the stable angle of the Diezma slope, which confirms that at drained conditions, after the failure, the strength of clay-rich materials on a shear surface falls to the residual values (cf. Skempton, 1985).

4. Discussion and conclusion

The Diezma and Riogordo landslides can be described as complex slope instabilities, occurred in moderate-relief areas after intense rainfall events. Both landslides have a rotational character in the head zone, evolving downwards to earthflows. In the case of the Riogordo landslide, large (up to 125 m³) carbonate blocks were displaced by the earthflow, reaching the lower parts of the landslide. These carbonate blocks come from a huge rockfall at the head of the landslide.

As for the controlling and triggering factors of these two landslides, a number of geologic, topographic and climatic parameters must be taken into account. For both landslides, the construction of the roads crossing the landslides is discarded as a triggering factor, since there is no temporal relationship between the road works and the occurrence of the landslides. Therefore, the roads in these two landslides acted as passive elements damaged by the slope instabilities. The real triggers for the landslides at issue here were intense rainfall episodes. In the case of the Riogordo landslide, the relationship between an intense rainfall event and the slope failure is clearly established (Fig. 9). Moreover, three intense fall–winter rainfall episodes predated the landslide, thus making reasonable the assumption that the Flysch formation was close to saturated conditions when the landslide finally occurred. On the contrary, the Diezma landslide occurred 20 days

after an intense rainfall peak (Fig. 9), being also preceded by several rainfall episodes, all of them occurring in a year rainier than the average one. This delay between rainfall peak and landsliding is probably related to the hydrogeologic behaviour of the Diezma area. In this area, a small carbonate aquifer located just behind the head of the landslide provides a continuous ground–water flow towards the Flysch formation outcropping in a lower topographic position. The increased ground–water discharge after the rainfall episode, together with the moisture increment directly attributable to the rainfall, are thought to be the triggers of the landslide. The delay in receipt the contribution of ground–water flow can explain why the Diezma landslide was delayed some days with respect to the rainfall peak. This ground–water influence is one remarkable difference between the Diezma and the Riogordo landslides. In the latter, ground–water flows to the north, without influencing the slope instabilities located in the southern side of the range.

Once discussed the triggers of the Diezma and Riogordo landslides, the next issue to be discussed concerns the reasons for the common occurrence of mass movements in some moderate-relief sectors of the Betic Cordillera. From a climatic point of view, it must be emphasized that most regions in the Betic Cordillera have a Mediterranean climate, although the total average annual rainfall can vary between 400 and 500 mm (as in the case of Diezma) and 700–800 mm (as in the case of Riogordo). Despite these variations, there is not any particular relationship between landslide distribution and annual rainfall. Therefore, present-day geographic annual rainfall variations cannot be claimed to be responsible for the abundance of landslides in some particular areas of the Betic Cordillera.

Other contributing factor to be considered is the structure, referred to as the presence of possible mechanical discontinuities and the geometry of the rock formations. The great majority of the landslides in the External Zone of the Betic Cordillera (South Iberian Domain) are located either at stratigraphic contacts or along fault traces. In both cases, formations of contrasting mechanical properties appear at both sides of the stratigraphic contact or fault trace: typically soft clay- and/or marl-rich rocks on one side and hard limestones/dolostones on the other side. Thus, the head of the landslides coincide in many cases with old fault scarps or with steeply dipping stratigraphic contacts. In other cases, the heads of the landslides do not correspond to the mechanical discontinuity provided by the stratigraphic contact or the fault trace, but instead are located at the foot of a steep escarpment, which, in turn, coincides with a stratigraphic or fault contact dipping contrary to the slope. This is the situation of the Riogordo and Diezma landslides, whose heads are located at the foot of a steep limestone escarpment, the contact between the limestones and the underlying Flysch formation being a thrust fault dipping to the north (Figs. 6 and 7), i.e. opposite to the general inclination of the slope.

Fig. 15. Plot of samples from the Flysch formation in the Casagrande plasticity chart. Smectite-rich clays and landslide debris have been plotted separately.

The most important controlling factor of the landslides studied here is the lithology of the rocks affected by the slope instabilities. The Diezma and Riogordo landslides and many others in the South Iberian Domain involve a clay- and marl-bearing Flysch formation of regional extent. This formation outcrops continuously from Gibraltar in the western Betics to Sierra Arana in the central Betics, along the boundary between the Alborán and the South Iberian Domains. Despite important lateral variations, this Flysch formation can be described in many localities as a “chaotic” succession constituted by centimetre- to decametre-scale blocks of different lithologies (limestone, dolostone, sandstone) embedded in a marly-clayey matrix. Such “chaotic” structure seems to be a primary feature of this Flysch formation, which may, however, facilitate landsliding when the topographic, structural and moisture conditions are appropriate. Interestingly, the landslide surfaces coincide with centimetre- to decimetre-thick clay-rich levels of very particular mineralogical composition and mechanical properties. These clay-rich layers have a very high smectite content (more than 90%), which is thought to be of critical relevance for the mechanical behaviour of the whole Flysch formation. Clay minerals in general and smectite in particular are quite resistant in dry conditions, but rapidly lose their strength in wet conditions. Thus, softened smectite-rich clay layers with high water contents can have the properties of a lubricant, which, in turn, can be critical for slope stability. In addition to their high plasticity, these smectite-rich clays have a high swelling potential, which can induce significant vertical overpressure, reducing even more the strength properties. Therefore, the presence in the Flysch formation of smectite-rich levels seems to be the main factor controlling the tendency of this formation to slide during or after rainy periods.

At a regional scale, slope stability in southeast Spain can be seriously conditioned by the presence of this smectite-rich Flysch formation. Therefore, the presence of this high-plasticity formation should be taken into account when evaluating landslide hazard in this region.

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