Gypsum scarps and asymmetric fluvial valleys in evaporitic terrains. The role of river migration, landslides, karstification and lithology (Ebro River, NE Spain)

J. Guerrero (1)* and F. Gutiérrez (1)

(1) University of Zaragoza, Earth Science Department, Spain

* Corresponding author. Tel.:0034 976762781; E-mail: jgiturbe@unizar.es

Abstract

Most of the Spanish fluvial systems excavated in Tertiary evaporitic gypsum formations show asymmetric valleys characterized by a stepped sequence of fluvial terraces on one valley flank and kilometric-long and more than 100-m high prominent river scarp on the opposite side of the valley. Scarp undermining by the continuous preferential lateral migration of the river channel toward the valley margin leads to vertical to overhanging unstable slopes affected by a large number of slope failures that become the main geological hazard for villages located at the toe of the scarps. Detailed mapping of the gypsum scarps along the Ebro and Huerva Rivers gypsum scarps demonstrates that landslides and lateral spreading processes are predominant when claystones crop out at the base of the scarp, while rockfalls and topples become the dominant movement in those reaches where the rock mass is mainly constituted by evaporites. The dissolution of gypsum nodules, seasonal swelling and shrinking, and dispersion processes contribute to a decrease in the mechanical strength of claystones. The existence of dissolution-enlarged joints, sinkholes, and severely damage buildings at the toe of the scarp from karstic subsidence demonstrates that the interstratal karstification of evaporites becomes a triggering factor in the instability of the rock mass. The genesis of asymmetric valleys and river gypsum scarps in the study area seem to be caused by the random migration of the river channel in the absence of lateral tilting related to tectonics or dissolution-induced subsidence. Once the scarp is developed, its preservation depends on the physicochemical properties of the substratum, the ratio between bedrock erosion and river incision rates, and climatic conditions that favour runoff erosion versus dissolution.

Keywords: fluvial scarp; interstratal dissolution; evaporites; slope movements; tilting; differential erosion

1. Introduction

Fluvial gypsum escarpments are prominent landforms in evaporite terrains (Gutiérrez et al., 1994). Their formation is often associated with the development of an asymmetric fluvial valley. Stepped fluvial terraces form narrow, laterally continuous benches on one valley flank, whereas the opposite margin of the valley is defined by tens of kilometres long prominent gypsum escarpment. Gypsum scarps may also develop in both sides of the valley leading to narrow valleys with paired terraces (Lucha et al., 2012). The scarps tend to show a rectilinear form controlled by the regional joint pattern and may reach more than 100 m high (Silva et al., 1988; Gutiérrez et al., 1994). They undergo a rapid retreat because of the persistent lateral migration of the fluvial channel leading to triangular facets and hanging valleys. The continuous undermining of the base of the escarpment due to river lateral erosion and downcutting causes vertical to overhanging unstable slopes affected by rockfalls, topples, and landslides that involve a risk for the villages located at the toe of the scarps. According to Ayala et al. (2003), slope movements are responsible for over €40 million damage per year in Spain. Rockfalls from gypsum escarpments cause important economic loss and occasionally result in casualties. For instance, several rockfalls that occurred in the locality of Azagra, located at the base of a Paleogene gypsum scarp, were responsible for 11, 100, 1, and 2 casualties in 1856, 1874, 1903, and 1946, respectively. In 1988 a gypsum block fell from the Jalón River gypsum scarp and collided with a house killing one of its inhabitants in the locality of Calatayud (Gutiérrez and Cooper, 2002).

The asymmetric configuration of the valley and the existence of a scarp incised into evaporitic gypsum rock are common features of fluvial systems that cross Tertiary evaporitic formations in Spain (Gutiérrez et al., 1994). This geomorphological phenomenon is well developed in the Ebro depression in the valleys of the Ebro River (Pellicer et al., 1984; Faci et al., 1986, 1988; Benito, 1989; Gutiérrez et al., 1994), Ginel River (Burillo et al., 1985), Jalón River, Aragón River (Carcedo et al., 1988; Lenároz, 1993), Huerva River (Guerrero et al., 2005, 2008), Ega River (Carcedo et al., 1988; Lenároz, 1993), Arga River (Lenároz, 1993), and Gállego Rivers (Benito, 1989). In the Barbastro anticline in the Spanish Pyrenees, Lucha et al. (2008) reported them in the Cinca, Noguera Ribagorzana, Farfaña, and Segre rivers (Lucha et al., 2008). In the Iberian range, an intraplate Alpine orogen in the northeast of Spain, Gutiérrez (1998) and Gutiérrez and Cooper (2002) described gypsum scarps in the Calatayud graben and Gutiérrez (1998) in the Teruel graben. In the Madrid Tertiary basin in central Spain, Silva (2003) and Silva et al. (1988) studied the Tajo, Tajuña, Jarama, and Manzanares gypsum escarpments and the fluvial sedimentology of thickened terrace deposits. They have been also described in fluvial systems that cross evaporitic formations around the world, such as the Ure River in Ripon, England (James et al., 1981), Jordan River in Jordania (Hassan and Klein, 2002), Sylva River in the central Urals in Russia

(Andrejchuk and Klimchouk, 2002), Fischells, Shamattawa, Shubenacadie, Cheticamp, Salt, and Lesser rivers in Canada (Tsui and Cruden, 1984), and the Eagle River in Colorado, USA.

The genesis of fluvial escarpments and valley asymmetry is often attributed to the influence of base level changes that cause a preferential migration of the river channel due to tectonic tilting, diapirism, or differential karstic subsidence (Bridge and Leeder, 1979; Leeder and Alexander, 1987; Osborn and du Toit, 1991; Mizutani, 1998). Conceptual models of river response to lateral tilting indicate that channel migration happens by avulsion (Heller and Paola, 1996) or by a steady process of preferential downslope cutoff and minor avulsion (Leeder and Alexander, 1987; Leeder and Gawthorpe, 1987). Examples of progressive down-tilt movement due to erosion on the down-dip river bank have been documented in the Madison and South Fork rivers in Montana where the majority of the meander loops were convex to the opposite tilting direction (Leeder and Alexander, 1987). The avulsive response to lateral ground tilting are well documented from the Holocene stratigraphy of grabens and half-grabens, such as the Carson River graben in Nevada (Peakall, 1998) and the stratigraphic record such as the Bighorn basin in Wyoming (Kraus, 1992), the Mugello half graben in central Italy (Benvenuti, 2003), and the Palomas and Mesillas half-grabens in New Mexico (Mack and Seager, 1990). Osborn and du Toit (1991) pointed out surprising examples about the capacity of river migration. For instance, in North America the Bow, Potomac, Missouri and Yellowstone rivers have undergone lateral migrations of 100 km in 6 Ma, 90 km during the Quaternary, 80 km in 10 Ma, and 56 km in 7 Ma, respectively. In Asia, the Brahmaputra and Kosi rivers have migrated laterally 10 km in just 150 years and 113 km in the last 228 years.

Independently of the shifting process (avulsion versus meander cutoff), the continuous existence of a subsidence-induced transverse slope on the floodplain causes the river channel to persistently reoccupy the axis of maximum subsidence leading to the stacking of channel or lacustrine facies in the depocenter of the basin (Leeder and Alexander, 1987; Leeder and Gawthorpe, 1987). Simulations in half grabens show that there is a marked clustering tendency of channel-belt or lacustrine deposits adjacent to the active margin that increases interconnectedness ratios and thickness of the alluvial fill (Heller and Paola, 1996). The spatial distribution of channel sheet sandbodies of the lower Eocene Willwood Formation in Wyoming was attributed to tilting associated with movement along basement-controlled faults that influenced the position of major channel systems (Kraus, 1992). The Walker River in Nevada crosses an active asymmetric graben and discharges into Walker Lake located in the highest subsidence area close to the active margin (Blair and McPherson, 1994). The asymmetric depositional wedge of the lower Cretaceous fluvial sandstone beds in Berry field in southeastern Alberta were caused by the continuous divertion of the fluvial channel into an area of synsedimentary salt dissolution subsidence (Hopkins, 1987). Other examples that explain the stacking of channel or lacustrine facies associated to tilting include: (i) in relation with tectonic tilting, the Rio Grande rift in New Mexico (Smith et al., 2001), the Megara basin in Greece (Bentham et al., 1991), the Nen river in Songnen plain in China (Bian et al., 2008), and the Namurian Kincardine graben in Scotland (Read and Dean, 1982); (ii) in response to halokinesis, the Permian and Triassic fluvial Cutler, Moenkopi, and Chinle Formations in salt-wall anticlines in Utah (Banham and Mountney, 2013) and the Cinca River valley in the Barbastro anticline in northeast

Spain (Lucha et al., 2008, 2012); and (iii) the Cambrian salt basin in south Oman (Hewards, 1990) associated with salt dissolution.

Regardless of facies distribution, one of the most cited example of surface lateral deformation is the existence of an asymmetric valley with stepped terraces at one side of the valley and a prominent scarp at the opposite side (Leeder and Alexander, 1987; Osborn and du Toit, 1991; Mizutani, 1998). Nevertheless, the existence of a stepped sequence of unpaired terraces does not need to be necessarily caused by surface tilting (Leeder and Alexander, 1987; Mizutani, 1998; Burbank and Anderson, 2001; Hancock and Anderson, 2002). Mizutani (1998) and Hancock and Anderson (2002) suggested that the number of terraces preserved in a valley depended on the time, the rate of lateral planation, and the shifting direction. The experimental studies conducted by Mizutani (1998) under conditions of constant base level, discharge, and sediment supply demonstrated that valley shape, terrace distribution, and preservation were significantly different in every running experiment. Asymmetric valleys with a stepped sequence of unpaired terraces in one side were formed when the channel shifted toward a preferred direction. In contrast, symmetric valleys with terraces in both sides of the valley were formed when the river shifted from one side to the other randomly eroding some previously created terraces. This experiment demonstrated that terrace preservation is fortuitous and asymmetric valleys may form in the absence of an imposed unidirectional channel migration. Therefore, the asymmetry of a fluvial valley should only be attributed to tilting when there is evidence of a preferred distribution of channel facies and an increase in their thickness toward the most subsiding margin (Bridge and Leeder, 1979; Leeder and Alexander, 1987). From the detailed geomorphological map of the Ebro and Huerva rivers, this manuscript demonstrates that gypsum scarps and asymmetric valleys in evaporitic terrains in the study area are not associated with tilting but to a combination of processes under semiarid climatic conditions. To our knowledge, despite the fact that fluvial gypsum escarpments are frequent landforms in evaporitic terrains, this is the first detailed study focused on their genesis and evolution.

2. Geological setting

The study area is located in the central sector of the Ebro Tertiary basin, which constitutes the southern foreland basin of the Pyrenees in the NE of the Iberian Peninsula (Fig. 1). In late Eocene times the Ebro basin became a land-locked depression surrounded by mountain ranges. Deposition during this endorheic stage was dominated by alluvial facies in the marginal areas of the basin grading into lacustrine evaporitic and limestone sediments in the most subsiding sectors, which migrated progressively toward the south (Riba et al., 1983; Ortí, 1997). In middle-upper Miocene times, once the basin was captured and opened toward the Mediterranean Sea (García-Castellanos et al., 2003; Pérez-Rivarés et al., 2004;), a new drainage network started to develop and dissect the endorheic basin fill by headward expansion, generating stepped sequences of mantled pediments and terraces.

The Ebro River drains longitudinally across the Ebro depression central sector following the axis of a very open NW-SE trending synclinal structure (Quirantes, 1978). The strata are also affected by subvertical joints and small normal faults with NW-SE, E-W, and NE-SW azimuths (Arlegui and Simón, 2001). A number of studies document the strong influence of the NW-SE fracture set on the development of landforms and karst features in the central sector of the Ebro basin (Quirantes, 1978; Gutiérrez et al., 1994, 2008; Galve et al., 2009). The Huerva River is a right bank tributary of the Ebro

River that drains transversally the central Ebro depression (Fig. 1). The study area covers a 40-km-long reach of the Ebro River around Zaragoza city and the last 30-km-long reach of the Huerva River valley up to its confluence with the Ebro River in Zaragoza city. In this sector the Ebro and Huerva valleys are excavated in Miocene sediments of the Longares, Zaragoza, and Alcubierre Formations (Esnaola and Gil, 1995). Distal alluvial fan claystones and sandstones of the Longares Formation grade laterally into evaporitic facies of the Zaragoza Formation, extending 17 km upstream of the Huerva River confluence and 30 km upstream and 20 km downstream of Zaragoza city (Fig. 1). The Longares Formation is constituted by 150 thick, fining-upward metre-thick cycles of red claystones and horizontal laminated and cross-bedded orange sandstones with an increasing proportion of claystones to the top (Esnaola and Gil, 1995). The Zaragoza Formation reaches more than 850 m thick (Torrescusa and Klimowitz, 1990) and mainly consists of anhydrite, halite, and glauberite in the subsurface and of secondary gypsum in outcrop (Salvany et al., 2007). New data from the study of 19 mining exploration boreholes allowed Salvany (2009) to describe a detailed stratigraphic column of the uppermost part of the Zaragoza Formation. From base to top, he distinguished: (i) a lower 75-m-thick halite unit situated at 150-175 masl, ≈ 40-15 m below the Ebro valley floor. (ii) An intermediate, 150-175 m thick unit made up of glauberite and anhydrite and minor beds of halite and clay at 430 to 175 masl. (iii) An upper anhydrite unit, around 110 m thick, situated above the highest preserved terrace of the Ebro River. The thick gypsum sequences exposed in the area correspond to a secondary lithofacies derived from the replacement (gypsification) of anhydrite and glauberite related to weathering. The Longares and Zaragoza Formations are capped by the 70-m-thick limestone sequence of the Alcubierre Formation. This resistant limestone unit forms structural platforms 400 to 600 m above the valley bottom (Fig. 1A).

3. Geomorphological setting

The present-day Ebro River, with an average discharge of 250 m^3/s in Zaragoza city, is a gravelly meandering channel that flows along a broad floodplain more than 4 km wide dominated by the deposition of fine-grained sediments. A belt of abandoned channel reaches and meander lobes related to avulsion and cut-off processes is in the areas adjacent to the currently active channel (Fig. 1C). The main tributaries of the Ebro River in the studied reach include the Gállego River on the northern margin, and the Huerva and Ginel Rivers on the southern margin (Figs. 1A and C). The Huerva River, with a mean annual discharge of 3.5 m^3/s in Mezalocha (20 km upstream of the study area), shows a meandering pattern with a channel sinuosity of 1.6. Guerrero (2008, 2013) identified a stepped sequence of 11 (T1: 200-210, T2: 180-190, T3: 150-160, T4: 120-130, T5: 100-105, T6: 85-90, T7: 65-70, T8: 45-55, T9: 25-35, T10: 10-15, T11: 2-7) and 12 terrace levels (T1: 115-110, T2:105-90, T3: 93-75, T4: 75-62, T5: 61-55, T6: 52-50, T7: 42-39, T8: 56-34, T9: 35-30, T10: 25-18, T11: 17-7, T12: 8-2 m above local base level) in the Ebro and Huerva rivers, respectively, and seven mantled pediment levels that may be correlated with some of the terraces. A simplified version of the map is presented in Fig. 1. In the areas where the bedrock mainly consists of insoluble detrital facies of the Longares Formation, the terrace and pediment deposits of the Ebro and Huerva Rivers show a uniform thickness of ≈ 3-4 m and remain undeformed (Guerrero et al., 2008, 2013). In contrast, the deposits of the terraces underlain by glauberite- and halite-bearing evaporite bedrock show abrupt thickenings, locally more than 50 m, superposition of alluvial units bounded by angular unconformities, abundant gravitational synsedimentary and postsedimentary gravitational deformation structures, and an abrupt increase in the proportion of fine-grained sediments locally constituting more than 70% of the total thickness (Guerrero et al., 2008, 2013). Structurally controlled kilometer-scale flat-bottom karstic depressions up to 50 m deep are located at both margins of the Ebro and Huerva valleys (Guerrero et al., 2013). They are attributed to subsidence caused by interstratal karstification of glauberite and halite beds. The floor of these basins is typically underlain by a thin marly deposit resting on intensively karstified and deformed evaporite bedrock. Gravitational deformation includes collapse structures 10-50 m across and bending gravitational structures more than 100 m wide and 30 m deep with superimposed collapse structures in the hinge zone (Guerrero et al., 2013). Most of them are captured by a dense network of flat-bottom infilled valleys that dissect the secondary gypsum outcrops in the area and feed alluvial fans at the margins of the Ebro and Huerva valleys.

The entrenchment and preferent northeastward migration of the Ebro River throughout its evolutions has generated a markedly asymmetric valley (Fig. 1C). The southern margin displays a staircase sequence of terraces, whereas the northern side is bounded by a prominent linear gypsum escarpment 60 km long and up to 120 m high, whose development is controlled by the highly pervasive NW-SE joint set (Gutiérrez et al., 1994). It extends from the village of Remolinos (35 km upstream of Zaragoza city) to Osera Creek, abruptly merging upstream and downstream where the evaporitic Zaragoza Formation grades into the detrital sediments of the Longares Formation. The continuous, linear trace is interrupted in the Gállego River confluence where terrace deposits reach more than 50 m thick (Fig. 1C). From the height and spatial location of the oldest mapped terrace we estimate an entrenchment of 210 m and a lateral migration of 11 km. The presence of triangular facets and hanging valleys (Fig. 2A) together with the occurrence of old Muslim defensive castles of the eleventh century (partly in ruins at the scarp edge) evidence of the rapid retreat of the gypsum scarp (Fig. 3A). Regarding the scarp age, terrace remnants belonging to T6 in the confluence of the Gállego River on top of the scarp postdate the beginning of its development. The scarp is still forming and retreating today, as the river channel is located at the toe of the scarp along several kilometers upstream Zaragoza city, undermining it and rendering it unstable.

Despite the preferred northeastern lateral migration of the Ebro River, the existence of two shorter NW-SE trending gypsum escarpments on the opposite margin suggest that the river migrated laterally toward the southwest during particular time periods. The shorter one is located to the south of Fuentes de Ebro village (Fig. 1C). It is characterized by a length of 8 km and an average height of 20 m. Regarding its chronology, it developed in the time interval between the sedimentation of terraces T3 and T4, as the top of the scarp is capped by T3 terrace deposits while its base is covered by thickened T4 terrace deposits on the right bank of the Ginel River. The other scarp, which is better preserved and displays a maximum height of 60 m, extends for 11 km between El Burgo de Ebro and Fuentes de Ebro villages (Figs. 1C and 2B). It is capped by terraces T5 to T8, depending on the studied reach, suggesting it was formed after T8 sedimentation. The deposits belonging to T11 are the youngest ones mantling the base of the scarp predating it.

Two clearly distinguishable sectors can be differentiated in the studied lower reach of the Huerva valley (Fig. 1B). Between Botorrita and Cuarte villages, the entrenchment and lateral migration of the fluvial system throughout its evolution has generated a markedly asymmetric valley. In this sector, stepped fluvial terraces form narrow and continuous benches on the western valley flank. In contrast, the eastern margin of the valley is defined by a prominent gypsum escarpment more than 10 km long and up to 120 m high affected by numerous slope movements and oriented NE-SW (Guerrero et al., 2005). The presence of hanging valleys and triangular facets in the gypsum escarpment and the absence of well-developed alluvial fans and remnants of old terrace levels in the eastern margin of the asymmetric valley reach are indicative of persistent lateral migration and incision of the fluvial system and rapid scarp retreat (Fig. 2C). An entrenchment of 115 m and an eastward lateral migration in excess of 2 km have been estimated for the fluvial system since the formation of the oldest mapped terrace. Downstream of Cuarte village, the valley becomes flanked on both sides by paired terraces and acquires a roughly symmetric configuration (Fig. 1B). In this sector, the Huerva River is excavated in thickened and slightly cemented Quaternary terrace deposits of the Ebro and Huerva rivers that may reach more than 60 m thick in Zaragoza city (Guerrero et al., 2008).

The area has a continental, semiarid climate with an average annual precipitation and temperature of 315 mm and 14.6ºC, respectively.

4. Type and distribution of slope movements.

We have distinguished five types of movements using Varnes (1978) and Cruden and Varnes (1996) classifications: (i) rockfalls, (ii) topples, (iii) rotational and translational landslides, (iv) lateral spreading, and (v) complex movements that involve the intervention of two or more of the previously described movements. The lithostratigraphy plays an important role in the distribution of the different types of slope movements (Gutiérrez et al., 1994). Those areas where claystones outcrop at the base of the scarp are more affected by landslides, lateral spreading processes and complex movements. In contrast, rockfalls and topples are predominant when the rock mass is exclusively constituted by evaporites.

4.1. Slope movement distribution in the Ebro valley

Rockfalls and topples often happen upstream of the confluence of the Ebro and Gállego rivers and downstream of Zaragoza up to Alfajarín village. Here, the rock mass is composed of secondary gypsum interbedded with centimetre- to decimetre-thick grey marl beds (Pellicer et al., 1984). It shows vertical to overhanging walls, mainly in those stretches where the Ebro River was recently undermining the scarp (Fig. 2A). Blocks that may reach more than 100 m^3 in volume (Fig. 3A), often break up during the collision leading to talus slope deposits several meters thick at the toe of the scarp. Use of the gravel track that runs at the base of the escarpment to the northwest of Zaragoza city by private cars is forbidden and the road is periodically closed. A rockfall of 45 m^3 collided with a house in 8 July 2010 in Alfocea village (2 km northwest of the study area). Fortunately there were no casualties (Fig. 3B).

Downstream of Alfajarin village, claystones belonging to the Longares Formation underlies the evaporitic strata at the base of the scarp favouring the development of landslides. Gutiérrez et al. (1994) indicated that: (i) landslides runout of more than 1 km in length and 3 hm^3 in volume, were predominantly rotational, multiple, and retrogressive (Fig. 4A); (ii) large, closed depressions developed at the head of the rotated slided mass are filled by gypsiferous silt and rockfalls coming from the main scar (Fig. 4C); (3) the lower mechanical strength of the claystones are responsible for the formation of multiple failure planes that exhibit abundant grooves and slickensides. The existence of claystone bulges at the toe of the landslides suggests that claystones are extruded outside in relation to intense plastic deformation; (iv) evaporitic strata overlying shear planes are highly brecciated and karstified; and (v) only two landslides located to the west of Alfajarín are active. They display a large number of aligned depressions, several metres wide sinkholes, and open grooves and cracks over 40 m in length and 2 m in width. The rest of the landslides are incised by the drainage network and covered by low-growing bushes indicating that they are old and inactive.

The two fluvial gypsum escarpments on the southern margin are very degraded by the drainage network displaying rounded slopes < 60º (Fig. 2B). Talus debris associated with rockfalls were removed by the river or covered by terrace deposits. Landslides are typically between 5 to 100 m long and show a drainage network with a very well ordered hierarchical arrangement. In the right margin of the Ginel River valley, the cuts of the Madrid-Barcelona high-speed railway has exposed terrace deposits more than 30 m thick belonging to the thickened T4 terrace covering a retrogressive rotational landslide (Fig. 4D) and rockfalls (Fig. 3C) developed from the southernmost scarp. Figure 3C shows up to 5-m-long gypsum blocks engulfed in a chaotic mass of 10 to 50 cm gypsum cobbles and boulders. The substratum rockhead is very irregular due to an intense karstification. The internal arrangement of the rockfalls allows four events to be differentiated. Every event is characterized by a mass of blocks that quickly wedges away from the scarp and is covered by fluvial deposits. Every rockfall event was probably related to a period of increasing dissolution and subsidence that caused the migration of the Ebro River toward the base of the scarp. Figure 4D shows a multiple landslide composed of two failure planes that define two, 10- and 15-m long slided blocks mantled by fluvial gravels. The strata of the lower block display a dip toward the scarp corroborating the rotational component of the movement. The beds of the upper block show a reverse dip due to a double rotation of the slided mass indicating the progressive behaviour of the landslide.

4.2. Slope movement distribution in the Huerva River

The existence of claystones of the Longares Formation seems to be a crucial factor in the type of slope movement as in the Ebro River valley. Rockfalls are predominant between María de Huerva and Cadrete villages where the rock mass is made up of secondary gypsum interbedded with metre thick beds of limestone, marls, glauberite, and halite belonging to the Zaragoza Formation overlying sandstones and siltstones of the Longares Formation. In this river reach, the scarp is undergoing a rapid retreat today, especially south of María village where the river flows along the base of the scarp (Fig. 1B). In Cadrete, the underlying sandstones and siltstones grade into claystones favouring the development of landslides and lateral spreading processes (Fig. 5C). The detrital Longares Formation merges toward the north downstream Cadrete and the scarp becomes mainly composed of evaporites. In this sector, rockfalls and topples are the main forms of slope instability.

Slope movements cause significant damage to numerous buildings of Cadrete and Cuarte village. The traffic along the road that connects both villages was interrupted several times in the last few years because of rockfalls and reactivated landslides. A few buildings constructed on landslides had to be demolished and those located at the toe of the scarp are severely damaged (Fig. 9F). From a detailed geomorphological map of the escarpment in Cadrete village at 1:1000 scale (Fig. 5A), we determined that rockfall and topples happen when the slope angle is over 70º. In addition they are more abundant in lateral valley creeks in the contact between the detrital Longares and evaporitic Zaragoza Formations where focused erosion on claystone beds by runoff led to the formation of overhung ledges in the contact of both units. As a result, the bottom of lateral creeks are often mantled by a mass of fallen blocks that may reach more than 10 m^3 in volume (Fig. 3D). Locally, rockfalls and topples result from complex slope movements evolving from landslides or lateral spreading processes.

Figure 5A shows that most of the landslides are rotational and that their failure planes are based in claystones of the lower unit. Translational landslides are small and show a single failure plane, whereas rotational landslides may reach more than 200 m long and 60 m wide and display multiple failure planes (Fig. 5A). Their major axis tends to orientate perpendicular to the slope gradient in lateral creek valleys and parallel in the scarp front (Fig. 5A). Failure planes often show a thin and parallel laminated claystone coating that reduces friction strength and favours sliding (Fig. 6A). As a result of the plastic behaviour of gypsum, the slided mass may occasionally bend to accommodate to the curvature of the failure plane leading to a synform with the axis parallel to the sense of movement (Fig. 6B). Those landslides located at the base of the scarp may be covered by or overlying T2 terrace deposits suggesting that they were formed previously or subsequently to T2 sedimentation, respectively (Fig. 5A). All of them were inactive displaying an important low-growing brush cover and an incised drainage network indicating they were developed under different climatic and geomorphological conditions to the present ones. However, five landslides have been recently reactivated due to slope changes caused by human activities (rotational landslides D1 to D5) (Fig.

5A). The excavation of the slope to lay the foundation for a building reactivated landslide D1 (Fig. 6B). The excavation of a 15-m-wide bench at the toe of landslides D2 and D3 caused their reactivation (Figs. 6C and 6D). Three collapse sinkholes between 0.5 and 2 m in diameter were formed on the bench surface demonstrating that karstification processes play an important role in the instability of the rock mass (Fig. 9C). Landslide D4 started moving after the construction of two water storage tanks at the bottom and middle parts of the slope. The one located in the middle sector was soon damaged and had to be demolished in 2008. The continuous water leakage of this tank triggered the sudden reactivation of a secondary landslide in April 2007 that collided with the tank located downslope (Fig. 9F). The reactivation of landslide D5 seems to be caused by the construction of several private houses at the head of the landslide (Fig. 5A). Those buildings display a large number of centimetre-wide cracks and the pavement is broken in pieces and tilted toward the scarp.

Lateral spreading movements are limited to areas close to Cadrete Castle and northeast of Cadrete (Fig. 5A) in the contact between the upper evaporitic unit and the underlying claystone unit. The plastic deformation of claystones causes the brecciation of the overlying gypsum unit into blocks of up to 45 m^3 that tend to flow and rotate toward the front of the scarp evolving into topples, rockfalls, and landslides (Fig. 7A) in a process called cambering (Cruden and Varnes, 1996). When claystones reach their maximum plasticity after rainfall events, gypsum blocks may subside into them becoming completely engulfed by a contorted claystone mass. The progressive flow of blocks on top of the claystone led to the formation of extensional cracks parallel to the scarp of more than 70 m long, 3 m wide, and 4 m deep partially filled by gypsiferous silts (Fig. 7B).

5. Factors involved in the development of slope movements

According to Crozier (1986) we may distinguish between conditioning and triggering factors. Conditioning factors refer to stratigraphic, structural, or topographical features of the rock mass that make the slope susceptible to instability processes. In contrast, triggering factors determine the temporal occurrence of slope movements. Lithostratigraphic distribution, joint pattern, and topography were classified into conditioning factors whereas rainfall, human activities, karstification, and river erosion into triggering factors.

5.1. Conditioning factors

The lithostratigraphic distribution seems to be critical in order to determine the spatial occurrence of slope movements. Most of the landslides and all of the lateral spreading processes happen where claystones crop out at the base of the scarp highlighting their critical role. Claystones are lithologies prone to slope movements because of their low mechanical shear strength that favours the development of failure planes (Taylor and Cripps, 1987; Bell and Pettinga, 1988; Sohby and Elleboudy, 1988; Bogaard et al., 2000). In addition, their constituent particles become reoriented along continuous bands parallel to the slip surface in the rock-shear plane interface reaching residual strength values and friction angles under small strain and little displacement (Kenney, 1984; Barton, 1988; Stark and Eid, 1994; Rouaiguia, 2010). Under undrained conditions, the low permeability of claystones causes high water pore pressures that decrease the effective normal stress in potential slip surfaces enhancing sliding (Kenney, 1984).

Weathering of claystones is known to contribute to slope instability. According to Taylor and Cripps (1987), weathering changes in mudrocks are accompanied by gradational decreases in shear strength and increases in water content and Atterberg limit value. Claystones of the Longares Formation have great potential to break down under minimum shear stress due to their considerable amount of gypsum nodules, swelling clay minerals, and high proportion of sodium. Quirantes (1978) and Esnaola and Gil (1995) quoted an increasing proportion of secondary gypsum nodules and veins to the top of the claystone unit. The dissolution of gypsiferous components causes the disintegration of the rock structure leading to a drop in cohesion and friction angle. Rick (1988) pointed out that landslides in a sector of the Swiss Alps were enhanced by the karst solution of gypsum crystals present in Keuper claystones. In addition, X-ray diffraction indicates that claystones in the Ebro and Huerva valley contain a high proportion of illite, kaolinite, and chlorite and a lower percentage of montmorillonite (Esnaola and Gil, 1995). Swelling phenomenon associated with clay minerals may be crucial in the triggering of landslides (Kenney, 1984; Sohby and Elleboudy, 1988; Bell and Pettinga, 1988; Selby, 1993; Gutiérrez et al., 1994). Changes in volume negatively alter the mechanical properties of the claystones especially along discontinuity planes (Selby, 1993). Crozier (1986), Selby (1993), and Gutiérrez et al. (1994) emphasized the importance of clay dispersion in the development of landslides. The high proportion of exchangeable sodium with respect to calcium and magnesium (SAR index) of the claystones is responsible for the development of piping and dispersion processes in the study area (Figs. 8A and 8B). The breakdown of clay aggregates by the leaching of sodium initiates the dispersion and structure instability of claystones with the subsequent decrease in shear strength (Gutiérrez et al., 1994; Sumner et al., 1998; Amezketa et al., 2003).

Rock masses usually contain numerous discontinuities such as bedding planes, faults, fissures, fractures, joints, and veins. They tend to possess low shear strength, negligible tensile strength and high hydraulic conductivity compared to the surrounding material, providing planes for shear failure and sliding (Priest, 1993). Rock mass permeability, strength, and proneness to failure are mainly governed by discontinuity orientation, spacing, size, frequency, and separation (Whalley, 1984; Selby, 1993; Winesa and Lillyb, 2002). In the study area, the stratigraphic sequence is affected by three structural subvertical regional joint sets oriented NW-SE, E-W, and NE-SW (Arlegui and Simón, 2001). If we consider that the limestone mesas of the Alcubierre Formation situated 400 to 600 m above the Ebro and Huerva valleys are the top of the Neogene sedimentation and assuming an average density of 2 g/cm^3 for the tertiary sequence, the strata cropping out at the evaporitic scarps have undergone an erosional unloading between 80 and 120 kg/cm^2. This important reduction in the compressive load resulted in widening of preexistent joints and the development of new release joints controlled by the topography (Figs. 8C and 8D). Most of the rockfall and landslide scars show a prevalent NE-SW or E-W orientation highlighting the structural control in the generation of slope movements.

Shear strength of discontinuities and consequently their susceptibility to sliding mainly depends on friction between their walls (Priest, 1993; Selby, 1993). In the study area, joint surface roughness is mainly reduced by karstification and infilling processes. Karstification is responsible for the widening of joints in evaporitic terrains (Durán and Val, 1984; Tsui and Cruden, 1984; Whalley, 1984; Williams and Davies, 1984; Faci et al., 1986, 1988; Carcedo et al., 1988; Gutiérrez et al., 1994). The absence of runoff after rainfall and the existence of a large number of springs at the contact between the upper evaporitic strata and underlying claystone unit (Fig. 5A) are evidence that the gypsum massif is highly karstified and that water quickly infiltrates along widened and enlarged joints by dissolution. Spring water samples in the Huerva gypsum scarp showed conductivity values over 30 mS/cm, were oversaturated in gypsum (10 g/l), and contained a high concentration in halite (8 g/l) demonstrating the important contribution of evaporite dissolution in slope instability. On the other hand, the marly insoluble residue that is left during karstification coats the discontinuity wall surfaces abruptly decreasing roughness. The geotechnical studies for the construction of Las Fuentes Bridge in Zaragoza city demonstrated that this karstic residue was mainly made up of high plasticity marls that display very low shear strength values (Serrano et al., 1990).

Finally, the high slope gradient of the scarp, usually over 70º, favours the development of slope movements. Rockfalls often happen in overhanging ledges undermined by lateral creeks or the Huerva and Ebro Rivers. Lateral spreading processes tend to occur in convex-shaped slopes with gradients between 30º and 70º. Although, slope gradient must have a positive effect in the generation of landslides, they seem to develop at any angle wherever claystones outcrop at the base of scarp, suggesting that the spatial distribution of claystones is the most relevant conditioning factor in their development.

5.2. Triggering factors

The downcutting and lateral migration of the Ebro and Huerva rivers control the topography of the scarp. The undermining of the base of the rock mass leads to oversteepened and overhung slopes that favour the formation of slope movements. Rockfalls and landslides that are mainly made up of evaporitic rocks are subsequently removed by the Ebro and Huerva rivers by mechanical erosion and dissolution contributing to keeping the slope profile vertical. The chemical erosion will be inversely dependent on the volume of the slided mass and proportionally on their solubility and brecciation level. The existence of halite and Na-sulphates in the bedrock enhances the dissolution phenomena due to their high solubility. Whereas the solubility of gypsum at 25ºC is 2.4 g/l, halite and glauberite solubilities reach 360 and 118, respectively (Ford and Williams, 1989). A rockfall of around 10 m^3 that fell into the Huerva River channel in 2003 between Cuarte and Santa Fe was completely removed within a year. This evolutional sequence of fluvial scarp base undercutting and slope movement formation and removal is the main cause for the rapid retreat of evaporitic scarps.

Evaporite karstification is probably one of the most significant factors contributing to the reduced mechanical strength of the evaporitic rock mass. The interstratal karstification of the bedrock at the base of the scarp causes the formation of voids that decrease the basal support of the cliff. Once a void is formed, the overlying evaporitic beds deform plastically thanks to crystal reorganization and varying amounts of interstitial muddy sediment (Bell, 1994; Karacan and Yilmaz, 2000; Gutiérrez et al., 2008) leading to synformal structures. During the flexure, the evaporitic roof often reaches the failure point giving way to subvertical failure planes that become potential shearing planes. Geotechnical tests of gypsum exhibit that it undergoes plastic-elastic-plastic deformation.

Building damage of Cadrete village was assessed in order to quantify the influence of karstic subsidence in the instability of evaporitic scarps. The effects of subsidence include severe building damage including tilting and cracking, sloping floors, sheared doors and window openings, bulging walls, collapsed and sagging roofs, broken pipes, and pavement collapse. The zonation of subsidence damage was evaluated by examination of the building façades. A damage category was assigned to each building on a scale of 1-4 based on the Subsidence Engineers' Handbook ranking system established by the British National Coal Board (N.C.B., 1975). Level 1 represents no damage and level 2 includes those buildings with appreciable damage such as millimetric cracks and/or doors and windows sticking. Open fractures (up to 1 cm), windows or doors distorted, and noticeable floor sloping were classified into level 3 (severe damage). Level 4 refers to very severe damage and includes centimetric wide cracks, roof and beam bearing lose, windows and doors broken, and severe slopes on floors. This ranking scheme has been usefully applied to evaporite dissolution subsidence studies in Ripon, England (Griffin, 1986; Cooper, 1998) and Calatayud (Gutiérrez and Cooper, 2002). This detailed survey is represented in a coloured map that shows the damage level of every construction in the village, borehole locations, and sinkhole distribution (Fig. 5A). Despite the damage level of a building depends not only on subsidence but on other factors such as the age, foundation type, and depth and characteristics of the supporting materials, this methodology may be a useful tool to determine the spatial distribution of subsidence. Borehole data indicate that Cadrete is located on a 15-m-thick terrace deposit that is overlying a substratum made up of 5-m-thick beds of gypsum, halite, and glauberite (Guerrero et al., 2005) (Fig. 5B). This anomalous terrace thickness is related to the synsedimentary dissolution of glauberite and halite layers (Guerrero et al., 2008). The buildings with the highest level of damage are located at the toe of the scarp suggesting that karstic subsidence is an important active process at this site (Figs. 9A and 9B). The often development of sinkholes in landslides (Fig. 9C) and at the base of the scarp (Fig. 9D) demonstrates the existence of interstratal cavities within the evaporitic bedrock that decrease the basal support and destabilize the gypsum massif. Recent excavations in Cuarte village expose a 75-m-long, 3-m-wide and 1- to 3-m-high, partially inundated, subcircular phreatic conduit at 10 m above the Huerva River channel and connected with T2 terrace level at the base of the gypsum scarp (Fig. 9E). This altitudinal correlation suggests that it was formed during T2 sedimentation. The bottom was covered by a variable thickness of insoluble residue and fallen gypsum blocks coming from the roof and walls. A breakdown pile of around 10 m long and 3.5 m high that was probably generated once the conduit became vadose were partially blocking it about 10 m away from the entrance. The existence of sinkholes and phreatic conduits and buildings severely damaged at the base of the scarp are evidence that karstification of bedrock is probably caused by regional flows discharging in the contact between terraces and the evaporitic scarp under phreatic conditions. Karstification was demonstrated to be a triggering factor in the genesis of landslides in the Iberian Range (Durán and Val, 1984) and Calatayud graben in Spain (Gutiérrez, 1998), French Alps (Rovera, 1993), Italian Alps (Alberto et al., 2008), eastern Germany (Reuter et al., 1977), and Alberta in Canada (Tsui and Cruden, 1984).

The close relationship between rainfall and landslide has been profusely studied. Rainfall is known to start the movement and reactivate or speed up preexisting landslides (Sowers and Royster, 1978; Bell and Pettinga, 1988; Battista and Surian, 1996; González et al., 1996; Julian and Anthony, 1996; Wieczoreck, 1996; Jiménez et al., 1999; Van Asch et al., 1999; Pair and Kappel, 2002; Schmidt and Beyer, 2002). In the study area, many landslides and rockfalls happen or accelerate after intense storms. Water is responsible for a drop in the mechanical resistance of the rock mass due to a decrease in the effective stress in shear planes, a rise in cleft water pressure, an increase in the weight of the slope materials, karstification of the evaporitic mass, and swelling and piping phenomena in claystones (Williams and Davies, 1984; Selby, 1993). Rainfall and karstification are closely related and often act together. Once sliding begins, the evaporitic mass starts breaking. Water infiltrates along the new fractures up to the failure plane increasing water pressure. As fractures widen by dissolution, water flows down faster getting to the shear plane before reaching their saturation point. Karstification concentrates on the sliding plane, roughness decreases, and aperture increases resulting in a drastic reduction of friction that favours sliding in a positive feedback mechanism.

Finally, the kinematics of the landslides, largely controlled by hydrological and karstification processes, was accelerated in some cases due to anthropogenic slope alterations, mainly excavation at the toe, overloading, and enhanced water infiltration. In Cadrete, construction works are responsible for the reactivation of landslides D1 to D5.

6. Genesis of gypsum scarps

The formation of gypsum scarps developed in fluvial valleys excavated across Spanish Tertiary evaporitic basins was often attributed to extensional tectonic tilting (Van Zuidam, 1976; Mensua and Ibáñez, 1977; Silva et al., 1988; Silva, 2003). Several lines of evidence counteract their tectonic origin: (i) the lack of relevant neotectonic structures and the low seismicity recorded in the central sector of the Ebro basin. The Spanish National Earthquake Hazard Map includes the central sector of the Ebro depression in the very low seismicity zone. This area is occasionally affected by earthquakes up to 3.0 in magnitude with epicentres located in the northern and southern ranges situated more than 70 km away from the study area (Martínez and Mezcua, 2002); (ii) the existence of gypsum scarps on both margins of the valleys suggests that there is no imposed shifting direction due to tectonic ground tilting; (iii) if the scarps were fault-related, their height should be proportional to the fault vertical throw. However, the strata located at both sides of the scarp (supposed footwall and hanging wall) are not displaced and remain undeformed; and (iv) the main river and its tributaries reveal gypsum escarpments on opposite margins with different alignments. For instance, the main scarp in the Ebro valley is on the left margin oriented NW-SE, while the Huerva and Gállego rivers tributaries that confluence in Zaragoza city display their escarpments in the right margins and oriented NE-SW. If every scarp had to be attributed to a fault plane, the city of Zaragoza would be cross-cut by three active normal faults which does not match borehole data, field observations, and the earthquake record.

Gutiérrez (1998) and Benito et al. (2000) postulated that synsedimentary subsidence due to the karstification of the soluble substratum may play an important role in river migration. The sedimentological studies of the Huerva and Ebro river terraces by Guerrero et al. (2008, 2013) pointed out that the channel of both rivers have locally migrated to lower topographic areas in the floodplain. Nevertheless, the impact of synsedimentary subsidence in the development of valley asymmetry has not yet been demonstrated. According to Gutiérrez (1998) and Benito et al. (2000), focused karstification along one margin would cause a lateral slope gradient in the floodplain confining river channel on the subsiding side of the valley and preventing river migration to the opposite site. When a reach of a river valley is affected by differential subsidence, the fluvial system tends to adjust its profile by aggrading in the subsidence area (Ouchi, 1985). This synsedimentary subsidence is commonly recorded by thickened fluvial sequences made up of stacked, fining-upward cycles (Read and Dean, 1982; Johnson, 1984) that show a high proportion of floodplain facies (Bridge and Leeder, 1979; Heller and Paola, 1996). Therefore, if focused karstification at the base of the evaporitic scarp was responsible for the asymmetry of valley, terrace deposits at the toe of the scarp should display an anomalous thickness and numerous gravitational deformation structures. Figures 1B and 1C show the spatial distribution of thickened terraces in the Ebro and Huerva rivers. In the Huerva River valley, upstream of Cuarte village the valley is asymmetric and river terraces display a relatively constant thickness lower than 4 m at the base of the scarp with the exception of T2 terrace in Cadrete and Cuarte villages where it may locally reach up to 15 m thick. In contrast, downstream of Cuarte where river terraces thicken to over 60 m, the valley becomes symmetric and the gypsum scarp disappears (Guerrero et al., 2008). In the Ebro River valley, asymmetry is related to the existence of a prominent gypsum scarp more than 60 km long on the left valley margin indicating a dominant NE migration. However, there are two shorter more degraded and older gypsum scarps in the opposite margin suggesting that the channel preferentially moved to the SW during some periods of the valley evolution. If the genesis of the three scarps was attributed to a synsedimentary subsidence phenomenon, then we have to assume that karstic subsidence underwent a complex spatial and temporal distribution changing from side to side to explain the Ebro River migration pattern. A seesaw karstic subsidence hypothesis would imply important changes in thickness and sedimentology of fluvial facies across and along the valley because of depocenter migration. The spatial distribution of the Ebro terraces (Fig. 1) shows that their deposits are often undeformed and display a constant thickness of around 5 m across the floodplain and along the base of the three scarps, except for local anomalous thickenings in the eastern side of the Ginel river and in the left valley margin downstream of Zaragoza city. Regarding facies changes, Guerrero et al. (2013) indicated that the Ebro River terraces that were mainly comprised of channel gravel deposits did not undergo a significant increase in floodplain facies in the subsiding stretches. The lack of sedimentological evidence of changes together with parallel river terrace profiles allow Guerrero et al. (2013) to suggest that the Ebro River was able to keep pace with subsidence thanks to an aggradation/subsidence rate > 1.

The disappearance of gypsum scarps in those sectors more affected by karstic subsidence counteracts evaporite dissolution as the main mechanism in the development of gypsum scarps. Figure 1 shows that the gypsum scarps of the Ebro, Huerva, and Gállego rivers are not developed in their confluence in Zaragoza city coinciding with an abrupt increase in terrace thickness. In this stretch, the Ebro, Gállego, and Huerva rivers terrace deposits fill synsedimentary subsidence troughs of more than 50 m deep (Guerrero et al., 2013), 100 m deep (Benito et al., 2000), and 60 m deep (Guerrero et al., 2008), respectively, related with interstratal dissolution of salt and glauberite beds. The prominent scarp located in the left margin of the Ebro River valley is interrupted upstream of Zaragoza city when the channel starts incising into their own thickened fluvial sediments and reappears again downstream of Zaragoza city once terrace thickness decreases and becomes constant (Fig. 1B). The shorter gypsum scarps on the right margin of the Ebro valley were formed downstream of the main subsiding area located in Zaragoza city. In the Huerva valley, the gypsum scarp disappears downstream of Cuarte village where the Huerva River terraces reach more than 60 m in thickness.

This geomorphological and sedimentological information shows that: (i) the genesis of gypsum scarps are not caused by a synsedimentary subsidence that only affects local river migration and (ii) when magnitude, aerial extension, and rate of karstic subsidence are high enough to cause a steady and significant aggradation of the axial river, tributaries, and lateral creeks, the evaporitic bedrock becomes overlain by tens of meters of alluvial deposits through which the river flows, migrates, and downcuts. In the absence of evaporite outcrops, gypsum scarp formation is inhibited and the valley may become symmetric with paired terraces.

In the absence of an imposed preferent direction of migration related to tectonic or karstic subsidence, the development of gypsum fluvial scarpments and valley asymmetry in the Ebro depression seems to be related to a fortuitous river migration pattern. A random channel shifting mechanism would explain the possible existence of scarps at both margins in the same valley, scarps oriented in any direction and margin in every river valley, the chaotic distribution of channel and floodplain facies within the floodplain, and the lack of thickened deposits along the base of the scarp. Nevertheless, the genesis of fluvial gypsum scarps requires a combination of factors: (i) a particular distribution of facies characterized by evaporites grading into fine-grained detrital sediments, (ii) contrasting physicochemical properties of bedrock, (iii) semiarid climatic conditions that favour runoff erosion versus dissolution, and (iv) a river incision rate higher than basin erosion rates.

The existence of evaporites grading laterally into fine-grained sediments seems to be critical in the development and preservation potential of gypsum scarps. Facies distribution in semiarid continental land-locked basins is often characterized by the proximal sedimentation of alluvial fan detrital deposits in the margins and the precipitation of evaporites in saline playa-lakes in the distal depocenter. When the basin becomes exorheic, the drainage network incises into the endorheic infill leading to evaporitic outcrops surrounded by mudstones. Detailed mapping demonstrate that fluvial scarps form independently of the substratum lithology whenever the river randomly migrates preferentially to a particular valley margin. Thus, river scarps forms in evaporites and claystones in the study area but they are not preserved in claystones due to their lower mechanical strength while they become prominent in river reaches excavated in gypsum bedrock. According to mean rock strength values (Selby, 1993; Waltham, 1994; Salinas, 2004), compressive strength of gypsum ranges from 5 to 35 MPa depending on impurity content while claystones vary from 1 to 20 MPa depending on the proportion of silt-sized particles and grade of stiffness. Tensile strength of gypsum and claystone oscillates between 1 and 2 MPa and between 0.1 and 2 MPa, respectively. The geotechnical surveys for the Zaragoza-Barcelona highway construction indicate that compressive strength of gypsum and claystone in the study area oscillates between 0.8 and 8 MPa and between 0.4 and 1.5 MPa, respectively. This means that compressive strength of gypsum may be more than 20 times higher than claystone. In addition, Young´s modulus of gypsum was up to 12 times greater than claystone. Despite the much lower mechanical strength of claystones than gypsum, there are other factors that favour the erodibility of claystones in the study area. According to Desir and Marin (2006), the erosion of claystones in the Ebro depression is enhanced by their high sodium absorption ratios and amount of swelling clay minerals that lead to soil dispersion and piping processes. From the data collected in four erosion plots monitored in a 10-year period between 1991 and 2001 in the Ginel River valley and the left margin of the Ebro River, Desir (2001) obtained average annual erosion rates for gypsum between 2.70 and 0.5 Mg/ha·y with a maximum of 19.54 Mg/ha·y and a minimum of 0.005 Mg/ ha·y depending on slope gradient, vegetation cover, sun radiation, and intensity and volume of rainfall. Considering a density of 2.5 g/dm^3 for gypsum, we can estimate a surface lowering between 0.11 and 0.02 mm/y. These values are in agreement with other denudation rates obtained in gypsum in the Spanish territory. Marques et al.'s (2008) runoff and sediment loss studies in gypsiferous soils in the Tertiary Madrid basin in central Spain determined an average erosion rate of 0.34 Mg/ha·y and a soil thickness lowering of 0.02 mm/y from 36 erosion plots monitored during 5 years. In the Sorbas Karst in southeast Spain, gypsum denudation rates range between 0.28 and 0.42 mm/y in surface outcrops and between 0.004 and 0.22 mm/y at vadose and phreatic caves, respectively (Calaforra, 1996). Worldwide erosion rates range from 0.003 to a maximum of 1.15 mm/y under optimum conditions along karstic conduits (Klimchouk et al., 1996). Surface denudation rates of the gypsiferous Castile Formation in New Mexico reached 0.5 mm/y in some areas with an average of 0.3 mm/y (Shaw et al., 2011). Cucchi et al. (1996) estimated an average denudation rate of 0.9 mm/y during an observation time of eight years in the karst of Trieste in Italy.

Low erosion rates in gypsum contrast with those obtained in mudrocks. Sirvent et al. (1997) and Desir and Marin (2006) determined the erosion rates of the Ebro depression Tertiary claystones by means of erosion pins and erosion plots during a 12-year period from 1993 to 2004 in the Lanaja and Bardenas Reales sectors located 20 km to the northeast and 50 km to the northwest of the study area, respectively. The data collected yielded an annual average value between 32 and 156 Mg/ha·y with a maximum of 752 Mg/ha·y and a minimum of 13 Mg/ha·y depending on rainfall volume and intensity. Recently, from 14 experimental erosional plots in claystones in the Huerva River valley, González-Hidalgo et al. (2005) obtained an annual average erosion rate between 75 and 1650 Mg/ha·y depending on slope gradient and vegetation cover. Considering an average bedrock density of 2.5 g/dm^3, surface lowering of claystone ranges between 1.28 and 66 mm/y. This value is up to 3200 times higher than in gypsum in the study area. In Spain, average erosion rates in mudrocks are similar to the ones calculated in the Ebro depression and range between 0.7 and 8 mm/y in Almeria (Cantón et al., 2001), 0.6 and 4.2 mm/y in Murcia (Bergkamp et al., 1996; López Bermúdez et al., 2000, Romero Díaz and Belmonte Serrato, 2002), 1.4 and 12 mm/y in Cataluña (Clotet et al., 1989), and 4 and 22 mm/y in the Pyrenees (Regüés et al., 2000; García-Ruiz et al., 2008). These data reveal that any type of landform would be preserved longer in evaporites than in claystones due to their lower erodibility in the study area. For instance, a fluvial scarp of 10 m would be completely degraded in 7800 years in claystones and 500,000 years in gypsum using the lower denudation rates calculated for both lithologies.

Climate seems to be critical in the development of gypsum scarps since the climatic conditions of a region controls erosion rates. The latter erosion data reveals that under semiarid conditions vegetation cover is scarce and runoff and piping processes in mudstones overcome evaporite dissolution. As a result, evaporites become the lithology with the higher preservation potential. On the contrary, in wetter regions dissolution processes may become dominant due to an increase in the vegetation cover that reduces drop impact, runoff, and clay dispersion resulting in lower mudrock erosion rates. This is the reason why in high latitude areas with high rainfall rates, fluvial gypsum scarps are absent or poorly developed. Finally, it is essential that the relation between river incision rate and basin erosion must be positive. The higher the rate of river incision the higher would be the scarp, and so more time would be needed to erode it increasing their preservation potential.

7. Conclusions

Spanish rivers crossing Tertiary basins often lead to asymmetric fluvial valleys in those river reaches excavated in evaporitic substratum. They are characterized by a sequence of stepped terraces on one side and a prominent, several kilometre-long and up to 120-m-high fluvial scarp mainly made up of gypsum in the opposite side of the valley. The scarp quickly merges upstream and downstream once the river becomes incised into claystones.

The entrenchment and migration of the Ebro River throughout its evolution has generated a prominent, NW-SE trending, linear gypsum escarpment 60 km long and 120 m high in the left margin and two shorter escarpments of 8 and 11 km long and 20 and 60 m high in the opposite margin suggesting a changeable shifting direction during its evolution. The preferent migration of the Huerva River toward the east has developed a markedly asymmetric valley between Botorrita and Cuarte villages with a NE-SW trending fluvial gypsum escarpment more than 10 km long and up to 120 m high affected by numerous slope movements (Guerrero et al., 2005).

Rockfalls and topples are the main movements in those stretches where the scarp is mainly composed of evaporites, while multiple retrogressive rotational landslides and lateral spreading processes become dominant where claystones crop out at the base of the scarp. Claystone shows a great potential to break down under minimum shear stress due to their considerable amount of swelling clay minerals, high proportion of exchangeable sodium that favours piping processes, and the dissolution of gypsum nodules that causes a significant drop in cohesion and friction angle.

A detailed karstic subsidence damage map of Cadrete village based on the examination of facades shows that severe damaged buildings were located at the base of the scarp demonstrating that karstification causes a significant decrease in the basal support that favours the development of slope movements. The formation of sinkholes also contribute to destabilize the rock mass acting as preferent flow paths for infiltrated waters down to sliding planes and increasing water pressure. In addition, the enlargement of discontinuity planes by dissolution causes an important friction drop between discontinuity walls enhancing sliding.

The evolution and rapid retreat of fluvial gypsum scarps is related to the continuous and preferential lateral migration of the river channel toward a valley margin causing the undermining of the base of the scarp, destabilizing the rock mass and favouring the development of slope failures. Lateral migration was often attributed to normal faulting or dissolution-induced subsidence. The lack of relevant neotectonic structures, vertical displacements, and terrace thickness increase in the supposed hanging wall together with the low seismicity recorded in the central sector of the Ebro basin counteract the tectonic tilting origin hypothesis. The disappearance of river asymmetry and gypsum scarps in those sectors more affected by synsedimentary karstic subsidence does not support evaporite dissolution as the main mechanism in the development of gypsum scarps. In addition, the constant thickness of the Ebro and Huerva river terraces and lack of expected sedimentological changes related to a subsidence phenomenon along and across gypsiferous scarps suggest that the synsedimentary subsidence related to the interstratal dissolution of halite and glauberite layers was exclusively responsible for local channel shifting. In the absence of an imposed preferent direction of migration related to tectonic or karstic subsidence, valley asymmetry seems to be related with a fortuitous river migration pattern. A random channel shifting mechanism would explain the possible existence of scarps at both margins in the same valley, scarps oriented in any direction and margin in every river valley, the chaotic distribution of channel and floodplain facies within the floodplain, and the lack of thickened deposits along the base of the scarp.

The occurrence of prominent river scarps only in reaches excavated in evaporites seems to be exclusively related to the greater mechanical strength of evaporites than the surrounding lithologies. In the study area, despite the fact that river scarps develop in evaporitic and argillitic substratum, their potential of preservation is greater in gypsum than in claystones under semiarid climatic conditions as rainfall is scarce and evaporite dissolution processes are irrelevant in comparison to runoff erosion. Erosion rates of claystones may be over 3000 times higher than in gypsum in the study area.

Acknowledgements

This research work has been funded by the national project CGL2010-16775 (Ministerio de Ciencia e Innovación and FEDER). I want to thank Francisco Gutiérrez for his help, knowledge and support.

References

Alberto, W., Giardino, M., Martinotti, G., Tiranti, D., 2008. Geomorphological hazards related to deep dissolution phenomena in the western Italian Alps: distribution, assessment and interaction with human activities. Engineering Geology 99, 147-159.

Amezketa, E., Aragüés, R., Carranza, R., Urgel, B., 2003. Chemical, spontaneous and mechanical dispersión of clays in arid-zone soils. Spanish Journal of Agricultural Research 1, 95-107.

Andrejchuck, V., Klimchouk, A., 2002. Mechanisms of karst breakdown formation in the gypsum karst of the fore-Ural region, Russia (from observations in the Kungurskaja Cave). Implication of Speleological Studies for Karst Subsidence Hazard Assessment. International Journal of Speleology N31, 89-114.

Arlegui, L., Simón, J.L., 2001. Geometry and distribution of regional joint sets in a non-homogeneous stress field: case study in the Ebro basin (Spain). Journal of Structural Geology 23, 297-313.

Ayala, F.J., Olcina, J., Vilaplana, J.M., 2003. Impacto económico y estrategias de mitigación de los riesgos naturales en España en el periodo 1990-2000. Gerencia de Riesgos y Seguros, MAPFRE, 84, 19-27.

Banham, S.G., Mountney, N.P., 2013. Evolution of fluvial systems in salt-walled minibasins: a review and new insights. Sedimentary Geology 296, 142-166.

Barton, M.E., 1988. Controle sedimentologique des surfaces de cisamement des plans de stratification. In: Bonnard, C. (Ed.), Landslides-Glissements de Terrains. Proceedings of the 5th International Symposium on Landslides. Rotterdam, Netherlands, AA Balkema, 73-76.

Battista, G., Surian, N., 1996. Geomorphological study of the Faldato landslide, Venetian Prealps, Italy. Geomorphology 15, 337-350.

Bell, D.H., Pettinga, J.R., 1988. Bedding-controlled landslides in New Zealand soft rock terrain. In: Bonnard, C. (Ed.), Landslides-Glissements de Terrains. Proceedings of the 5th International Symposium on Landslides. Rotterdam, Netherlands, AA Balkema, pp. 77-83.

Bell, F.G., 1994. A survey of the engineering properties of some anhydrite and gypsum from the north and midlands of England. Engineering Geology 38, 1-23.

Benito, G., 1989. Geomorfología de la Cuenca Baja del Rió Gállego. Universidad de Zaragoza, Ph.D. Thesis, Zaragoza, Spain.

Benito, G., Gutiérrez, F., Pérez-González, A., Machado, M.J., 2000. Geomorphological and sedimentological features in Quaternary fluvial systems affected by solution-induced subsidence (Ebro basin, Spain). Geomorphology 33, 209-224.

Bentham, P., Collier, R.E.L., Gawthorpe, R.L., Leeder, M.R., Prossor, S., Stark, C., 1991. Tectono-sedimentary development of an extensional basin, the Neogene Megara basin, Greece. Journal of Geological Society of London 148, 923-934.

Benvenuti, M., 2003. Facies analysis and tectonic significance of lacustrine fan-deltaic successions in the Pliocene-Pleistocene Mugello basin, central Italy. Sedimentary Geology 157, 197-234.

Bergkamp, G., Cammeraat, E., Martínez Fernández, J., 1996. Water movement and vegetation pattern on shrub lands and abandoned fields in desertification threatened areas. Earth Surface Processes and Landforms 21, 1073-1090.

Bian, J., Tang, J., Lin, N., 2008. Relationship between saline–alkali soil formation and neotectonic movement in Songnen Plain, China. Environmental Geology 55, 1421-1429.

Blair, T.C., McPherson, J.C., 1994. Historical adjustments by Walker River to lake level fall over a tectonically tilted half-graben floor, Walker Lake basin, Nevada. Sedimentary Geology 92, 7-16.

Bogaard, T.A., Antoine, P., Desvareux, P., Giraud, A., van Asch, Th.W., 2000. The slope movements within the Mondorès graben (Drôme, France), the interaction between geology, hydrology and typology. Engineering Geology 55, 297-312.

Bridge, J.S., Leeder, M.R., 1979. A simulation model of alluvial stratigraphy. Sedimentology 26, 617-644.

Burbank, D.W., Anderson, R.S., 2001. Tectonic Geomorphology. Blackwell Science Ltd., Oxford, UK.

Burillo, F., Gutiérrez, M., Peña, J.L., 1985. Las acumulaciones holocenas y su datación arqueológica en Mediana de Aragón (Zaragoza). Cuadernos de Investigacion Geográfica 11, 193-207.

Calaforra, J.M., 1996. Karstología de yesos. Instituto de Estudios Almerienses. Servicio de Publicaciones de la Universidad de Almería. Alemería (España). Monografías. Ciencia y Tecnología 3.

Cantón, Y., Solé-Benet, A., Queralt, I., Pini, R., 2001. Weathering of a gypsum-calcareous mudstone under semi-arid environment at Tabernas, SE Spain: laboratory and field-based experimental approaches. Catena 44, 111-132.

Carcedo, A., Valls, A., Borel, C., 1988. Estudio de inestabilidades en los acantilados yesíferos de la Ribera de Navarra. In: Alonso, E., Corominas, J. (Eds.), II Simposio sobre taludes y laderas inestables. IGME, Andorra, pp. 657-668.

Clotet, N., García-Ruiz, J.M., Gallart, F., 1989. High magnitude geomorphic work in Pyrenees Range: November's 1982 unusual rainfall event. Studia Geomorphologica Carpatho-Balcanica 23, 69-91.

Cooper, A.H., 1998. Subsidence hazards caused by the dissolution of Permian gypsum in England: geology, investigation and remediation. In: Maund, J.G., Eddleston, M. (Eds.), Geohazards in Engineering Geology. Engineering Geology Special Publications, Geological Society, London, 15, pp. 265-275.

Crozier. M.J., 1986. Field assessment of slope instability. In: Brunsden, D., Prior, D.B. (Eds.), Slope Instability. John Wiley & Sons Ltd., Chichester, UK, pp. 103-142.

Cruden, D.M., Varnes, D.J., 1996. Landslide types and processes. In: Turner, A.K., Schuster, R.L. (Eds.), Landslides: Investigation and Mitigation. National Academy Press, Washington DC.,USA, pp. 36-75.

Cucchi, F., Forti, P., Marinetti, E., 1996. Surface degradation of carbonate rocks in the karst of Trieste. In: Fornos, J.J., Gines, A. (Eds.), Karren Landforms. Universitat de les Illes Balears, Palma, Spain, pp. 41-52.

Desir, G., 2001. Erosión hídrica en terrenos yesíferos en el sector central de la Depresión del Ebro. Publicaciones del Consejo de Protección de la Naturaleza de Aragón, Zaragoza, Spain.

Desir, G., Marín, C., 2006. Factors controlling the erosion rates in a semiarid zone (Bardenas Reales, NE Spain). Catena 71, 31-40.

Durán, J.J., Val, J., 1984. Incidencia de la disolución kárstica en taludes con materiales hipersolubles. El deslizamiento de Santa Cruz de Moya (Cuenca). In: Romana, M. (Ed.), Reconocimiento de macizos rocosos. VIII Simposio Nacional, Sociedad Española de Mecánica de las Rocas, Madrid, Spain, pp. 159-172.

Esnaola, J.M., Gil, C., 1995. Memoria y mapa geológico de España, 1: 50.000 Zaragoza (382). IGME, Madrid, Spain.

Faci, E., Rodriguez-Avial, J.I., Jugo, J., 1986. Estabilización y establecimiento de medidas preventivas en un talud rocoso en Azagra (Navarra). In: Alonso, E., Corominas, J. (Eds.), II Simposio sobre taludes y laderas inestables. IGME, Andorra, pp. 485-496.

Faci, E., Rodriguez-Avial, J.I., Jugo, J., 1988. Estabilización y mediadas correctoras en el talud rocoso de Las Tres Marías en Falces (Navarra). In: Alonso, E., Corominas, J. (Eds.), II Simposio sobre taludes y laderas inestables. IGME, Andorra, pp. 497-511.

Ford D., Williams, P., 1989. Karst Geomorphology and Hydrology. Unwin Hyman, Winchester, MA.

Galve, J.P., Gutiérrez, F., Lucha, P., Bonachea, J., Cendrero, A., Gimeno, M.J., Gutiérrez, M., Pardo, G., Remondo, J., Sánchez, J.A., 2009. Sinkholes in the salt-bearing evaporite karst of the Ebro River valley upstream of Zaragoza city (NE Spain). Geomorphological mapping and analysis as a basis for risk management. Geomorphology 108, 145-158.

García-Castellanos, D., Vergés, J., Gaspar-Escribano, J., Cloetingh, S., 2003. Interplay between tectonics, climate and fluvial transport during the Cenozoic evolution of the Ebro basin (NE Iberia). Journal of Geophysical Research 108, B7, 2347, ETG 8-1/8-18.

García-Ruiz, J.M., Regüés, D., Alvera, B., Lana-Renault, N., Serrano Muela, P. Nadal-Romero, E., Navas, A., Latron, J., Martí-Bono, C., Arnaez, J., 2008. Flood generation and sediment transport in experimental catchments affected by land use changes in the central Pyrenees. Journal of Hydrology 356, 245-260.

González, A., Salas, L., Díaz, J.R., Cendrero, A., 1996. Late Quaternary climate changes and mass movement frequency and magnitude in the Cantabrian region, Spain. Geomorphology 15, 291-309.

González-Hidalgo, J.C., Luís Arrillaga, M., Peña Monné, J.L., 2005. Los eventos extremos de precipitación, la variabilidad del clima y la erosión del suelo. Reflexiones ante el cambio del clima en los sistemas mediterráneos. Cuaternario & Geomorfología 19, 49-62.

Griffin, D.A., 1986. Geotechnical assesment of subsidence in and around Ripon, North Yorkshire, due to natural solution. MsC Thesis, University of Newcastle upon Tyne, UK.

Guerrero, J., Gutiérrez, F., Lucha, P., 2005. Peligrosidad, daños y mitigación de inundaciones, subsidencia por disolución y movimientos de ladera en la localidad de Cadrete (Depresión del Ebro, Zaragoza). Cuaternario & Geomorfología 19, 63-82.

Guerrero, J., Gutiérrez, F., Lucha, P., 2008. The impact of halite dissolution subsidence on fluvial terrace development. The case study of the Huerva River in the Ebro basin (NE Spain). Geomorphology 100 (1-2), 164-179.

Guerrero, J., Gutiérrez, F., Galve, J.P., 2013. Large depressions, thickened terraces and gravitational deformation in the Ebro River valley (Zaragoza area, NE Spain). Evidence of glauberite and halite interstratal karstification. Geomorphology 196, 162-176.

Gutiérrez, F., 1998. Fenómenos de subsidencia por disolución de formaciones evaporíticas en las Fosas Neógenas de Teruel y Calatayud (Cordillera Ibérica). Ph.D. Thesis, Zaragoza University, Spain.

Gutiérrez, F., Cooper, A., 2002. Evaporite dissolution subsidence in the historical city of Calatayud, Spain. Damage appraisal and prevention. Natural Hazards 25, 259-288.

Gutiérrez, F., Arauzo, T., Desir, G., 1994. Deslizamientos en el escarpe de Alfajarín (Zaragoza). Cuaternario & Geomorfología 8, 57-68.

Gutiérrez, F., Guerrero, J., Lucha, P., 2008. A genetic classification of sinkholes based on the analysis of evaporite paleokarst exposures in Spain. Environmental Geology 53, 993-1006.

Hancock, G.S., Anderson, R.S., 2002. Numerical modeling of fluvial strath-terrace formation in response to oscillating climate. Geological Society of America Bulletin 114, 1131-1142.

Hassan, M.A., Klein, M., 2002. Fluvial adjustment of the lower Jordan River to a drop in the Dead Sea level. Geomorphology 45, 21-33.

Heller, P.L., Paola, C., 1996. Downstream changes in alluvial architecture: an exploration of controls on channel-stacking patterns. Journal of Sedimentary Research 66, 2, 297-306.

Hewards, A.P., 1990. Salt removal and sedimentation in southern Oman. Geological Society of London Special Publications 49, 637-651.

Hopkins, J., 1987. Contemporaneous subsidence and fluvial channel sedimentation: upper Mannville C Pool, Berry Field, Lower Cretaceous of Alberta. AAPG Bulletin 71, 334-345.

James, A.N., Cooper, A.H., Holliday, D.W., 1981. Solution of the gypsum cliff (Permian, middle Marl) by the River Ure at Ripon Parks, North Yorkshire. Proceedings of the Yorkshire Geological Society 43, 433-450.

Jiménez, M., Farias, P., Rodríguez, A., Menéndez, R.A., 1999. Landslide development in a coastal valley in northern Spain: conditioning factors and temporal occurrence. Geomorphology 30, 115-123.

Johnson, S.Y., 1984. Cyclic fluvial sedimentation in a rapidly subsiding basin, northwest Washington. Sedimentary Geology 38, 361-391.

Julian, M., Anthony, E., 1996. Aspects of landslide activity in the Mercantour Massif and the French Riviera, southeastern France. Geomorphology 15, 275-289.

Karacan, E., Yilmaz, I., 2000. Geotechnical evaluation of Miocene gypsum from Sivas (Turkey). Geotechnical and Geological Engineering 18, 79-90.

Kenney, C., 1984. Properties and behaviours of soils relevant to slope instability. In: Brunsden, D., Prior, D.B. (Eds.), Slope instability. John Wiley & Sons Ltd, Chichester, UK, pp. 27-65.

Klimchouk, A., Cucchi, F., Calaforra, J.M., Aksem, S.D., Finocchiaro, F., Forti, P., 1996. Dissolution of gypsum from field observations. International Journal of Speleology (Italian edition) 25, 37-48.

Kraus, M.J., 1992. Alluvial response to differential subsidence: sedimentological analysis aided by remote sensing, Willwood Formation (Eocene). Bighorn basin, Wyoming, USA. Sedimentology 39, 455-470.

Leeder, M.R., Alexander, J., 1987. The origin and tectonic significance of asymmetrical meander belts. Sedimentology 34, 217-226.

Leeder, M.R., Gawthorpe, R.L., 1987. Sedimentary models for extensional tilt-block/half-graben basins. In: Coward, M.P., Dewey, J.F., Hancock, P.L. (Eds.), Continental Extensional Tectonics. Geological Society Special Publications 28, Blackwell Scientific Publications, Oxford, UK, pp. 139-152.

Lenaroz, B., 1993. Geomorfología y geología ambiental de la ribera de Navarra. Ph.D. Thesis, Universidad de Zaragoza, Spain.

López-Bermúdez, F., Conesa-García, C., Alonso-Sarría, F., Belmonte-Serrato, F., 2000. La cuenca experimental de la Rambla Salada (Murcia). Investigaciones hidrogeomorfológicas. Cuadernos de Investigación Geográfica 26, 95-112.

Lucha, P., Gutiérrez, F., Guerrero, J., 2008. Environmental problems derived from evaporites in the Barbastro-Balaguer anticlinal core, NE Spain. Environmental Geology 53, 1045-1055.

Lucha, P., Gutiérrez, F., Galve, J.P., Guerrero, J., 2012. Geomorphic and stratigraphic evidence of incision-induced halokinetic uplift and dissolution subsidence in transverse drainages crossing the evaporite-cored Barbastro-Balaguer anticline (Ebro basin, NE Spain). Geomorphology 171-172, 154-172.

Mack, G.H., Seager, W.R., 1990. Tectonic control on facies distribution of the Camp Rice and Palomas Formations (Pliocene-Pleistocene) in the southern Rio Grande rift. Geological Society of America 102, 45-53.

Marques, M.J., Bienes, R., Pérez-Rodríguez, R., Jiménez, L., 2008. Soil degradation in central Spain due to sheet water erosion by low-intensity rainfall events. Earth Surface Processes and Landforms 33, 414-423.

Martínez, J.M., Mezcua, J., 2002. Catálogo sísmico de la Península Ibérica (880 a.C.-1990). Monografía Núm. 18, IGN, Madrid, Spain.

Mensua, S., Ibáñez, M.J., 1977. Sector central de la Depresión del Ebro. Mapa de Terrazas Fluviales y Glacis. 3ª Reunión Nacional del Grupo de Trabajo del Cuaternario, Universidad de Zaragoza, Spain.

Mizutani, T., 1998. Laboratory experiment and digital simulation of multiple fill-cut terrace formation. Geomorphology 24, 353-361.

National Coal Board (N.C.B.), 1975. Subsidence Engineers' Handbook. National Coal Board Mining Department, London, Spain.

Ortí, F., 1997. Evaporitic sedimentation in the south Pyrenean foredeeps and the Ebro basin during the Tertiary: a general view. In: Busson, G., Schreiber, B.Ch. (Eds.), Sedimentary Deposition in Rift and Foreland Basins in France and Spain. Columbia University Press, New York, USA, pp. 319-334.

Osborn, G., du Toit, C., 1991. Lateral planation of rivers as a geomorphic agent. Geomorphology 4, 249-260.

Ouchi, S., 1985. Response of alluvial rivers to slow active tectonic movement. Geological Society of America Bulletin 96, 504-515.

Pair, D., Kapel, W., 2002. Geomorphic studies of landslides in the Tully valley, New York: implications for public policy and planning. Geomorphology 47, 125-135.

Peakall, J., 1998. Axial river evolution in response to half-graben faulting: Carson River, Nevada, U.S.A. Journal of Sedimentary Research 68, 788-799.

Pellicer, F.; Echevarría, M.T., Ibañez, M.J., 1984. Procesos actuales en el escarpe de yesos de Remolinos. Cuadernos de Investigación Geográfica 10, 159-168.

Pérez-Rivarés, F., Garcés, M., Arenas, C., Pardo, G., 2004. Magnetostratigraphy of the Miocene continental deposits of the Montes de Castejón (central Ebro basin, Spain): geochronological and paleoenvironmental implications. Geologica Acta 2, 221-234.

Priest, S., 1993. Discontinuity Analysis for Rock Engineering. Chapman and Hall, London, UK.

Quirantes, J., 1978. Estudio sedimentológico y estratigráfico del Terciario continental de los Monegros. Institución Fernando el Católico, CSIC, Zaragoza, Spain.

Read, W.A., Dean, J.M., 1982. Quantitative relationships between numbers of fluvial cycles, bulk lithological composition and net subsidence in a Scottish Namurian basin. Sedimentology 29, 181-200.

Regüès, D., Guàrdia, R., Gallart, F., 2000. Geomorphic agents versus vegetation spreading as causes of badland occurrence in a Mediterranean subhumid mountainous area. Catena 40, 173-187.

Reuter, F., Molek, H., Bochmann, G. (1977). Slope sliding as a secondary process in subsidence areas of chloride-karst. Bulletin of the International Association of Engineering Geology, 16, 62-64.

Riba, O., Reguant, S., Villena, J., 1983. Ensayo de sintesís estratigráfica y evolutiva de la Cuenca Terciaria del Ebro. Libro jubilar J. M. Ríos, Mapa Geológico de España, Zaragoza, Escala 1: 50.000, Segunda serie, IGME, Madrid, pp. 131-159.

Rick, B., 1988. Instability as a consequence of deep disintegration in gipseous marls. In: Bonnard, C. (Ed.), Landslides-Glissements de Terrains. Proceedings of the International Symposium on Landslides. A.A. Balkema, Rotterdam, Netherlands, pp. 73-76.

Romero Díaz, A., Belmonte Serrato, F., 2002. Erosión del suelo en ambiente semiárido extremo bajo diferentes tipos de litologías y suelos. In: Pérez González, A., Vagas, J., Machado, M.J. (Eds.), Aportaciones a la Geomorfología de España en el inicio del tercer milenio. Instituto Geológico y Minero de España, Madrid, Spain, pp. 315-322.

Rouaiguia, A., 2010. Residual shear strength of clay-structure interfaces. International Journal of Civil & Environmental Engineering 10, 5-14.

Rovera, G., 1993. Instabilite des versants et dissolution des evaporites dans les Alpes Internes; L´exemple de la montagne de Friolin. Revue de Geographie Alpine 81, 71-84.

Salinas, J.L., 2004. Diccionario guía de reconocimientos geológicos para ingeniería civil. Ministerio de Fomento CEDEX, Madrid, España.

Salvany, J.M., 2009. Geología del yacimiento glauberítico de Montes de Torrero. Universidad de Zaragoza, Zaragoza, Spain.

Salvany J.M., García-Veigas, J., Ortí, F., 2007. Glauberite-halite association of the Zaragoza Gypsum Formation (lower Miocene, Ebro Basin, NE Spain). Sedimentology 54, 443-467.

Schmidt, K., Beyer, I., 2002. High-magnitude landslide events on a limestone-scarp in central Germany: morphometric characteristics and climatic controls. Geomorphology 49, 323-342.

Selby, M.J., 1993. Hillslope Materials and Processes. 2nd Edition. Oxford Universty Press, Oxford, UK.

Serrano, A., Dapena, E., Villar, J.M., 1990. Cimentación de un puente de 331 m de longitud sobre pilotes en margas yesíferas. Simposio sobre el Agua y el Terreno en las infraestructuras Viarias. Sociedad Española de Mecánica del Suelo e Ingeniería Geotécnica. ISSMGE, Madrid, Spain, pp. 125-132.

Shaw, M.G., Stafford, K.W., Tate, B.P., 2011. Surface denudation of the Gypsum plain. Abstracts with Programs - Geological Society of America 43, 42.

Silva, P.G., 2003. El Cuaternario del valle inferior del Manzanares (Cuenca de Madrid, España). Estudios Geológicos 59, 107-131.

Silva, P., Goy, J.L., Zazo, C., 1988. Neotectónica del sector centro-meridional de la cuenca de Madrid. Estudios Geológicos 44, 415-427.

Sirvent, J., Desir, G., Gutierrez, M., Sancho, C., Benito, G., 1997. Erosion rates in badland areas recorded by collectors, erosion pins and profilometer techniques (Ebro basin, NE-Spain). Geomorphology 18, 61-75.

Smith, G.A., McIntosh, W., Kuhle, A.J., 2001. Sedimentologic and geomorphic evidence for seesaw subsidence of the Santo Domingo accommodation-zone basin, Rio Grande rift, New Mexico. Geological Society of America Bulletin 113, 561-574.

Sohby, M.A., Elleboudy, A.M., 1988. Instability of natural slope in interbedded limestone and shale. In: Bonnard, C. (Ed.), Landslides- Glissements de Terrains. Proceedings of the International Symposium on Landslides, A.A. Balkema, Rotterdam. Netherlands, pp. 121-123.

Sowers, G., Royster, D., 1978. Field investigation. In: Schuster, R., Krizek, R. (Eds.), Landslides: Analysis and Control. National Academy of Science, Washington, DC., pp. 81-110.

Stark, T.D., Eid., H.T., 1994. Drained residual strength of cohesive soils. Journal of Geotechnical Engineering 5, 856-871.

Sumner, M.E., Rengasamy, P., Naidu, R., 1998. Sodic soils: a reappraisal. In: Sumner, M.E., Naidu, R. (Eds.), Sodic Soils: Distributions, Properties, Management and Environmental Consequences. Oxford University Press, Oxford, NY, 3-17.

Taylor, R.K., Cripps, J.C., 1987. Weathering effects: slopes in mudrocks and over-consolidated clays. In: Anderson, M.G., Richards, K.S. (Eds.), Slope Instability. John Wiley and Sons Limited, Chichester, UK, pp. 405-445.

Torrescusa, S., Klimowitz, J., 1990. Contribución al conocimiento de las evaporitas Miocenas (Formación Zaragoza) de la Cuenca del Ebro. In: Ortí, F., Salvany, J.M. (Eds.), Formaciones evaporíticas de la Cuenca del Ebro y cadenas periféricas y de la zona de Levante. ENRESA-GPG, Barcelona, Spain, pp. 120-122.

Tsui, P.C., Cruden, D.M., 1984. Deformation associated with gypsum karst in the Salt River escarpment, northeastern Alberta. Canadian Journal of Earth Science 21, 949-959.

Van Asch, Th.W., Buma, J., Van Beek, L.P., 1999. A view on some hydrological triggering systems in landslides. Geomorphology 30, 25-32.

Van Zuidam, R., 1976. Geomorphological Development of the Zaragoza Region. Processes and Landforms Related to Climatic Changes in a Large Mediterranean River Basin. Enschede. Int. Inst. for Aerial Survey and Earth Science (I.T.C.), Utrecth, Germany.

Varnes, D.J., 1978. Slope movement types and processes. In: Schuster, R.L., Krized, R.J. (Eds.), Landslides: Analysis and Control. National Academy of Science, Washington, DC., pp. 11-33.

Waltham, T., 1994. Foundations of Engineering Geology. Chapman & Hall, Oxford, UK.

Whalley, W.B., 1984. Rockfalls. In: Brunsden, D., Prior, D.B. (Eds.), Slope Instability. John Wiley & Sons Ltd., Chichester, UK, pp. 217-256.

Wieczorek, G.F., 1996. Landslide triggering mechanisms. In: Turner, A.K., Schuster, R.L. (Eds.), Landslides: Investigation and Mitigation. National Academy Press, Washington, DC., pp-76-90.

Williams, A., Davies, P., 1984. Cliff failure along the Glamorgan Heritage Coast, Wales, U.K. In: Flageollet, J.C. (Ed.), Mouvements de terrains. Reserches en Geographic Physique de L´environment. Documents du BRGM 83, Association Française de Geographie Physique, Caen, France, pp. 109-119.

Winesa, D.R., Lillyb, P.A., 2002. Measurement and analysis of rock mass discontinuity spacing and frequency in part of the Fimiston open pit operation in Kalgoorlie, western Australia: a case study. International Journal of Rock Mechanics and Mining Sciences 39, 589–602.

Figure Captions

Fig. 1. (A) Geomorphological map of the central sector of the Ebro depression showing the location of gypsum scarps. (B) Detailed geomorphological map of the lowest reach of the Huerva River valley. (C) Detailed geomorphological map of the Ebro River valley downstream of Zaragoza city to Osera Creek.

Fig. 2. (A) Northwest view of the left margin gypsum scarp in the Ebro Valley at Alfajarin village characterized by triangular facets and hanging valleys (reproduced with permission~~(image taken by F. Gutiérrez~~). (B) Southeast view of the youngest scarp of the right margin in the Ebro valley from El Burgo village. (C) North view of the Huerva River prominent gypsum scarp and Cadrete village built on a T2 terrace at the toe of the gypsum scarp.

Fig. 3. (A) Ruins of the Castellar Castle at the Ebro River scarp edge located 18 km upstream of Zaragoza city. The photograph that was taken during the 2003 flood shows an unstable gypsum block detached from the scarp edge (reproduced with permission~~(image taken by F. Gutiérrez~~). (B) Rockfall colliding with a house in Alfocea village located 2 km northwest of the study area (reproduced with permission~~image taken by F. Gutiérrez~~). (C) Rockfalls covered by Ebro River T8 terrace deposits located at the base of the oldest right margin evaporitic scarp to the east of the Ginel River. (D) Private fence pushed over by rockfalls in Cadrete village in the Huerva River valley.

Fig. 4. (A) The 130-m-long, 230-m-wide, multiple, retrogressive, rotational active landslide downstream of Alfajarin village at the left margin gypsum scarp of the Ebro valley (reproduced with permission~~image taken by F. Gutiérrez~~). (B)

Rotational landslide located northwest of the confluence of the Ebro and Gállego rivers. The Juslibol oxbow lake is located at the base of the scarp suggesting that river undermining of the scarp is the main triggering factor in this sector. (C) Detail of the depression formed at the head of the landslide shown in (A). (D) Rotational landslide covered by T4-thickened terrace deposits of the Ebro River to the southeast of Fuentes village.

Fig. 5. (A) Geomorphological map of Cadrete village showing the distribution of slope movements and karstic subsidence damage in buildings based on the examination of façades. (B) Geologic cross section of the Huerva River valley in Cadrete village based on borehole data. (C) Correlation panel of the Longares and Zaragoza Formations in the gypsum scarp in Cadrete village.

Fig. 6. (A) Failure plane of landslide D1 at Cadrete village displaying reoriented shales due to shearing. (B) View of rotational landslide D1 with the location of a building at its head. Its additional weight seems to be responsible for D1 reactivation. (C) and (D) Reactivation of landslides D2 and D3 due to the excavation of a 15-m-wide bench at the base of the scarp.

Fig. 7. Spreading processes in the Huerva River gypsum scarp. (A) Displaced and rotated gypsum blocks due to cambering. (B) Gypsum blocks more than 45 m^3 individualised by extensional cracks up to 70 m long and 3 m wide because of the plastic deformation of the underlying clays.

Fig. 8. Conditioning factors in the development of slope movement in the study area. (A) and (B) Metric-size pipes related to dispersion of sodium-rich claystones of the Longares Formation at Cadrete and María villages in the Huerva River valley, respectively~~ (images taken by F. Gutiérrez)~~ (reproduced with permission). (C) and (D) More than 1 m wide, 2 m deep, and 100 m long opened by preexisting joints developed from erosional unloading and karstification parallel to the Ebro and Huerva river scarps, respectively.

Fig. 9. Triggering factors involved in slope movements in the study area. (A) and (B) Severely damaged buildings located at the toe of the Huerva River scarp with subsidence-induced decimetric cracks. (C) and (D) Collapsed sinkholes 1 m in diameter developed in a berm excavated at the toe of the D2 and D3 rotational landslides and at the base of the scarp in Cadrete village (see location in Fig. 2). (E) A 75-m-long, partly inundated old phreatic karstic conduit in the Huerva River valley located at 10 m above the Huerva river channel in Cuarte village~~ (image taken by F. Gutiérrez)~~. (F) View of the rotational landslide D4 showing the location of the middle and lower water storage tanks. Water leakage from the middle tank triggered the sudden reactivation of a secondary landslide in April 2007 that collided with the lower tank (picture at the upper right corner)~~ (image taken by F. Gutiérrez)~~.