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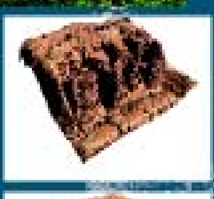

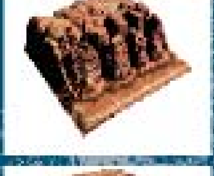

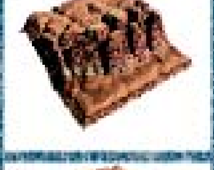

FIELD TRIP GUIDE

C-2

SIXTH INTERNATIONAL CONFERENCE ON GEOMORPHOLOGY

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FIELD TRIP GUIDE - C2

DIVERSE GEOMORPHIC PROCESSES IN THE SOUTH EASTERN PYRENEES.

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The excursion addresses the geomorphic and hydrological processes in the middle mountain area of the Eastern Pyrenees and its piedmont. The 'Salt Mountain' and the salt mines at Cardona, as well as the karst processes in saline deposits, will be visited during the first day. The second day will focus on a visit to the experimental area of Vallcebre, where hydrological and erosion processes in abandoned agricultural fields and badlands, as well as large mass movements, have been studied and monitored for more than 10 years. Finally, some of the mass movements triggered by an extreme rainfall event in 1982 will be visited. Structural landforms in limestones and conglomerates will afford a splendid scenery on the way.

1. Introduction to the geology and geomorphology of the upper Llobregat catchment and the Central Catalan Depression.

The headwaters of the Llobregat catchment is a part of the meridional Pyrenees, a middle-mountain area with summits up to about 2,500 m a.s.l., built up by a system of nappes of sedimentary rocks displaced several tens of kilometres to the South. The oldest rocks are Palaeozoic limestones and shales, followed by thick formations of marine limestones, marls and clays of Mesozoic age. During the Paleocene, the early tectonic deformations caused the occurrence of continental basins with the deposition of lignites, clays, gypsum and limestones. The youngest rocks are massive limestones and marls that belong to the Eocene. The nappes are divided in two groups. The upper ones are the upper and lower Pedraforca nappes that take a synform structure, whereas the lower ones form an antiformal stack that includes the Cadí nappe. The upper where displaced earlier (since the late Cretaceous) and the lower where displaced later (up to the early Oligocene) carrying the upper ones. The Cadí nappe experienced a displacement of about 15 km, whereas the Pedraforca nappe underwent a total displacement of more than 50 km.

The more meridional nappe thrusts over the materials of the Central Catalan Depression which is the eastern part of the Ebro Depression. The main sedimentation in this large unit started in the Eocene when the uplift of the Catalan Coastal Ranges closed a sea where the streams deposited large amounts of sediments in fan-deltas. In the early Oligocene, the desiccation of this sea caused the deposition of thick saline deposits. The sedimentation continued with torrential and alluvial facies that had endorrheic characters to the west, closer to the drier and central areas.

The more relevant features in the structure of the materials in this depression are the general tilt of the beds towards the centre of the depression and the occurrence of a series of long folds with E-W axes and diapiric characteristics. Indeed, the geophysical surveys demonstrated that the basement is not affected by the folding, whereas the arrangement of the folds points to the thrust of the Pyrenean mantles as the origin of the deformation. Finally, it is worth to state that most of the sedimentation in this depression was synchronous with the deformations, as evidenced by the scenic progressive unconformity at the Serra de Busa (first stop, Fig. 1).

The geological structure affords the main controls for the relief in this area. The main drainage net is arranged N-S, consequent to the mountain front, and the secondary drainage net is roughly W-E, subsequent to the main structural alignments. Yet, the limestone formations provide the main elevations, whereas marly and clayey units occur usually in the depressions. There are many structural landforms (cliffs, cuestas, structural plateaus), as well as direct and reverse Jurassic-type forms (monts, vals, combes...). In the Ebro Depression the lithological controls are much less important and the drainage net has a more dendritic structure, whilst there are still some Jurassic landforms produced by the diapiric folds.

The geomorphic evidences of Pleistocene climates are few in this area. In spite of the relatively high summits, there are only a small number of glacial landforms. Nevertheless, the study of some deposits suggest the role of some glacial features that would not produce conspicuous landforms because of the strong structural controls (Clotet et al., 1984). Furthermore, there are many quaternary deposits with large boulders that have not been investigated; some of them may be attributed to recent or pleistocene mass movements but some others might have a glacial origin. The more frequent periglacial features are *grèzes littées* (sloping bedded screes of angular rock fragments) in the lower parts of limestone cliffs.

The climate is sub-Mediterranean, with a characteristic water deficit period in summer, when potential evapotranspiration exceeds precipitation. Mean annual temperature at 1440 m a.s.l. is 7.3 ºC, with more than 100 freezing days per year. In winter, high atmospheric pressure situations are frequent, allowing strong thermal inversions. The mean annual precipitation is in the range 800-1000 mm with the main precipitation in autumn and spring. The highest rainfall intensities occur in summer. Snowfalls are occasional and snow blanket may last for some months only above 2000 m a.s.l..

First day: September 12th

Stop 1.1: Creu del Codó Belvedere.

The purpose of this stop is to provide a general view of this part of the Southern Pyrenees and the boundary with the Ebro geological depression. To the North, the Port del Compte calcareous plateau, with Pleistocene periglacial hillslope deposits, and the Cardener valley. To the North-East, the El Verd range, with some badlands, and the Llosa del Cavall reservoir. To the East, the Aigua de Valls valley and the ridges on vertical beds of Eocene conglomerates of the Ebro Depression in an unconformity with the upper (Oligocene) beds. To the South-East, the dipping of the lower beds decreases and the unconformity disappears (Fig. 1), evidencing that the northern beds were deformed at the same time the southern ones were deposited.

Stop 1.2 : Earthflow at la Coma (November 1982).

On the 6, 7 and 8th November 1982, an extreme rainfall event occurred in North-Eastern Spain. The rainfall depth measured in the nearest rainfall gauge (Guilanyà) was 250 mm for the whole event. By 1 a.m. on the 8th, a liquid debris flow about 3 m thick arrived to the lawns of the houses, that deposited a 1 m-thick layer but produced little damages. In the morning, the neighbours observed the slow advance of an earth tongue that stopped near the houses after destroying some water tanks (Fig. 2).

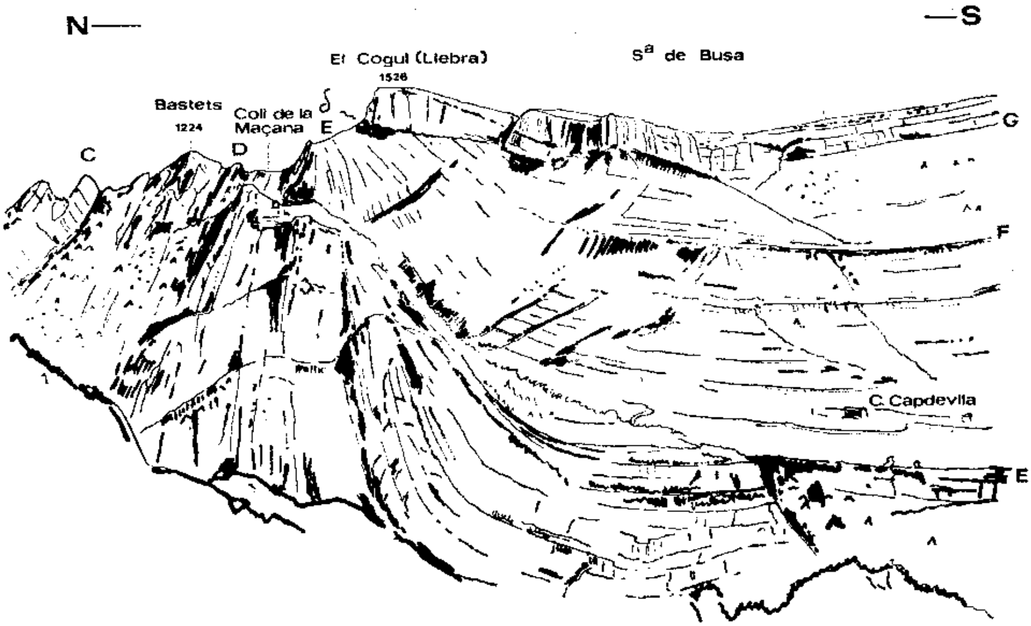

Figure 1: Progressive unconformity in the Eocene-Oligocene sediments of the Ebro Depression at the contact with the Southern Pyrenean nappes (Riba, 1973).

The subsequent observations evidenced that the main movement was a planar-slump slide of Eocene marine green clays over a 27º dipping sandstone bed. Presumably the debris flow was triggered when the main movement occurred, whereas the toe of the slump subsequently flowed as a slowly moving tongue. The central part of the tongue was about 40 m wide and showed transversal ridges with the edges curved upslope that evidenced the existence of internal shear surfaces. The sides of the tongue in its middle part were in the form of levees made of materials much more clayey than the main tongue. (Clotet and Gallart, 1984).

The landforms in the nearby of this mass movement evidence that other mass movements of different kinds occurred in the past. Nevertheless, the forms are not fresh; suggesting that the return period of these movements is rather long (several centuries).

The La Coma landslide is one among many mass movements that occurred during this extreme rainfall event in the area. An inventory of these mass movements in the headwaters of the Llobregat river demonstrated that all the major (deeper) movements occurred in areas with evidences of older movements, whereas the occurrence of shallow mass movements was closely related to the spatial distribution of precipitation totals (Gallart and Clotet, 1988; Gallart, 1995).

Figure 2: Mass movement at La Coma on November 8th ,1982 (Photo F. Gallart).

Second day, September 13th

Morning: Hydrolological Behaviour of Mediterranean Mountain Catchments; Erosion and sediment transport in badlands.

1. Introduction

Sub-humid Mediterranean mountains share the hydrological processes from both wet and dry environments and are the source of water necessary for human life and activity in the drier downstream areas. The study of the hydrological functioning of these areas, besides its fundamental interest, may help to anticipate the hydrological consequences of both climate and land cover change, as well as to design land-use strategies that might counteract these changes.

The Vallcebre catchments were instrumented in 1989 to address these questions, as well as the erosion and sediment transport issues (see below). This is a scientifically driven research whose results are expected to be useful in the long term, following the methodology established for the International Hydrological Decade 1965-1975 (Toebes and Ouryvaev, 1970) renewed in some aspects for the IAHS Decade on Predictions in Ungauged Basins (PUB): 2003-2012 (Sivapalan et al., 2003).

The main results obtained after over fourteen years of study in the area are reviewed here, most of them addressed in Ph Dr. works and published in several scientific journals. The characteristics of climate and precipitation, the hydrological role of vegetation, the spatial and temporal patterns of soil moisture, and the runoff generation mechanisms are the main aspects addressed.

2. Characteristics of the studied area

The Vallcebre catchments are located in the headwaters of the Llobregat River, on the southern margin of the Pyrenees (Catalonia, northeastern Spain) at 42°12N, 1°49E (Fig. 3), with altitudes between 1100 and 1700 m.

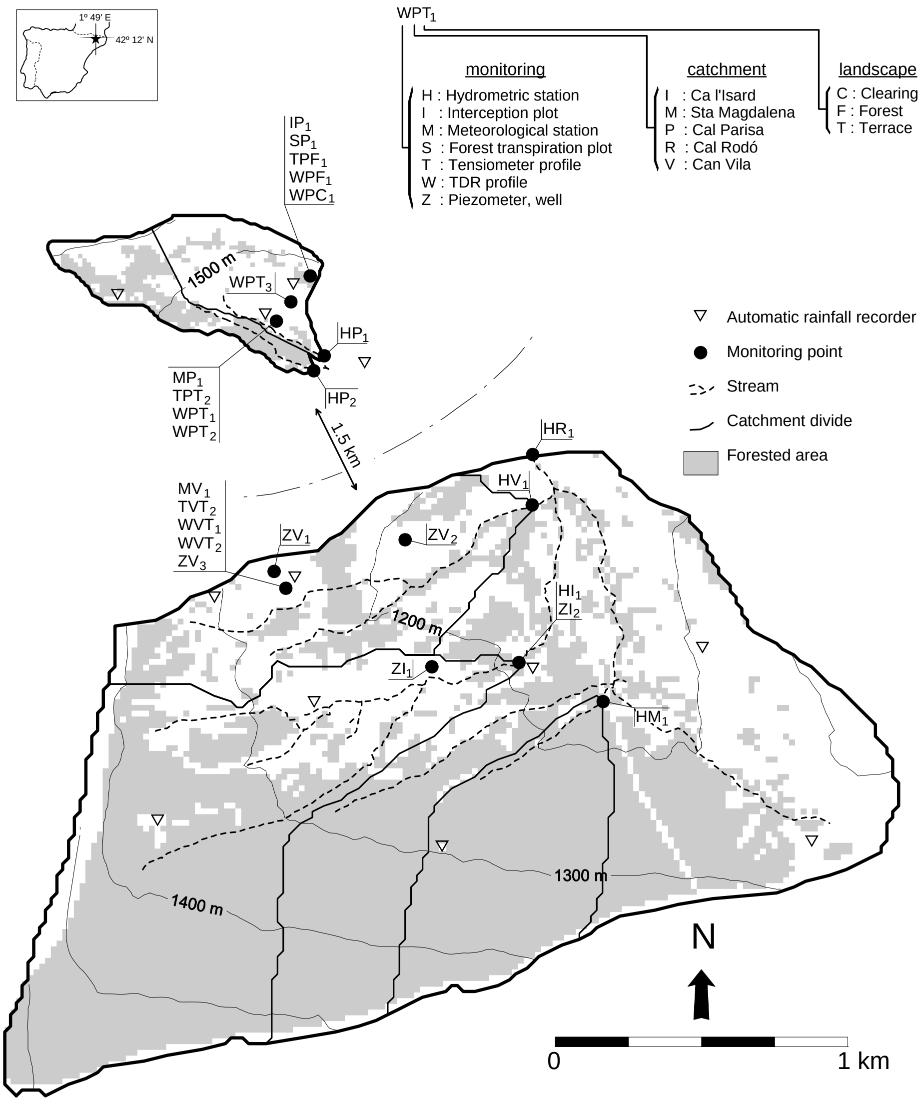

Figure 3: Map of the Vallcebre catchments, showing location of the instruments.

The bedrock is formed by red clayey smectite-rich mudrocks with massive limestone beds of continental facies attributed to the Palaeocene. The soft mudrocks are prone to landsliding and badland erosion. Soil thickness varies greatly, depending on lithology, geomorphology and the changes induced by terracing. Badland areas exhibit regoliths whose thickness varies throughout the year, but which rarely reach 15 cm; limestone areas are overlain by discontinuous soils up to about 40 cm thick; soils on hillslopes over clayey rocks are up to 80 cm thick and agricultural terraces can have soils thicker than 3 m. Topsoils are loamy and show high infiltration capacities due to their good structure. Nevertheless, hydraulic conductivity drops in the deeper horizons, inducing the formation of shallow semi-permanent aquifers.

The climate is sub-Mediterranean, with a characteristic water deficit period in summer (Fig. 4). Mean annual temperature at 1440 m a.s.l. is 7.3 ºC and the mean annual precipitation is 924 mm with a mean of 91 rainy days per year. Snowfalls are occasional and represent less than 5% by volume. Annual reference evapotranspiration is about 900 mm.

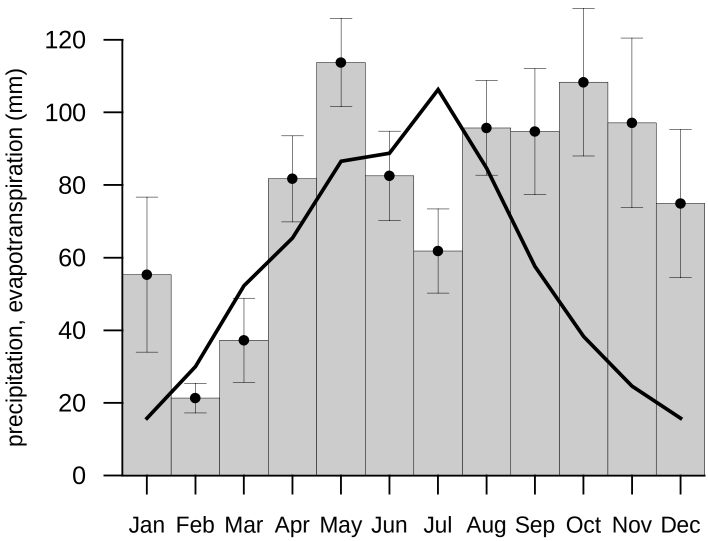

Figure 4: Monthly precipitation (bars) and reference evapotranspiration (thick line) at Vallcebre. Error bars represent standard deviations of precipitation (Gallart et al., in press).

Climax vegetation is woodland of *Quercus pubescens* with *Pinus sylvestris* on the cooler north-facing aspects. Nevertheless, most of the gentle hillslopes were deforested and terraced in the past for agricultural use and subsequently underwent a progressive abandonment during the second half of the 20th Century. Former agricultural terraces are now covered by mesophile grasses with hydrophile patches. Following abandonment, spontaneous afforestation by *Pinus sylvestris* has occurred on 30% of the catchment. Marginal areas, which were the first to be abandoned, nowadays have a rather dense forest cover, whereas those areas abandoned later exhibit forest patches and scattered trees (Poyatos et al., 2003).

An analysis of vegetation changes in the last 40 years (Poyatos et al. 2003) showed that in 1957, forests of diverse density covered 39 % of the catchment and grassland and crops 28%, whereas in 1996 the relative areas of these covers changed into 64 and 18% respectively.

3. Instrumental design.
The research area (Fig. 3) consists of two catchment clusters whose centres are 2.5 km apart. The main cluster is the Cal Rodó catchment (4.17 km^2), which has been subdivided into three sub-catchments (Ca l'Isard, Can Vila and Sta. Magdalena), whereas the smaller cluster is called Cal Parisa and consists of a pair of catchments of similar size (0.15 and 0.17 km^2). Instrumentation of the research catchments started in 1989 (Llorens and Gallart, 1992; Balasch *et al.*, 1992). Along with the monitoring of weather, precipitation and stream flow, several state variables (soil moisture and tensiometry, water table) and processes (rainfall interception, tree transpiration, grass evapotranspiration) are measured permanently or during several-year long periods.

The pluviometric network consisted of 12 tipping-bucket rain gauges, connected to data-loggers that recorded 0.2 mm precipitation increments at a temporal resolution of 1 s. Two weather stations were installed in the respective catchment clusters. (see Figure 3 for instrumentation locations).

All the stream gauging stations were provided with control structures where water level and temperature measurements were logged at time intervals between 2 and 60 minutes. Gauging stations HR_1 HI_1 and HV_1 were equipped with infrared backscattering turbidity sensors as well as automatic water sampling devices controlled by data-loggers. These first two stations were also provided with suspended sediment sensors based on ultrasonic beam attenuation, which are appropriate for the high range of sediment concentrations observed during storms (Gallart *et al.*, 1998).

Soil water content has been measured in the Vallcebre catchments since 1993 using the Time-Domain Reflectometry (TDR) method at 9 profiles distributed in the main geo-ecological units of the (7 in the Cal Parisa catchment and 2 in Can Vila sub-catchment). These profiles consisted of sets of four vertical 20 cm-long probes permanently installed in the ground at 0-20, 20-40, 40-60 and 60-80 cm depth. These were read every week with a Tektronix 1502-C cable tester.

A network of soil tensiometers was installed in late 1995. Depths to water table were measured weekly at four old wells and four continuously recording piezometers were instrumented in the catchments in late 1995, in order to study the dynamics of the water table during rain events.

An experimental forest plot in the Cal Parisa basin (IP_1 in Fig. 3) has been monitored since 1993 to evaluate the water balance of a representative afforestation patch. This 198-m^2 plot has a monoespecific cover of *Pinus sylvestris* with no understorey. Stand density is about 2400 stems ha^{-1}. The plot was instrumented for the continuous monitoring of rainfall interception (Llorens *et al.*, 1997a), tree transpiration (Oliveras and Llorens, 2001) and soil water potential, as well as periodical measurement of soil moisture.

4. Main outcomes of the research.
4.1 Rainfall interception
Mean throughfall in the experimental forest plot during the period July 1993–August 2000 represents about 74% of bulk rainfall, while stemflow accounts for only 2%, leading to a rainfall interception rate of 24% of bulk rainfall (Llorens et al., 2003b). Observation of the role of meteorological conditions on rainfall interception at the event scale, analysed by Llorens et al. (1997a), lead to the identification of three types of events, considering their duration, the differences in atmospheric conditions and rainfall intensity. Long events with low rainfall intensities and wet atmospheric conditions are the most frequent events measured. Short events with high rainfall intensities and dry atmospheric conditions, and medium events with low rainfall

intensities and very dry atmospheric conditions are less frequent but more characteristic of the Mediterranean climate. The application of a common rainfall interception model provided acceptable results (Llorens, 1997).

No clear seasonal control on rainfall interception rates has been observed at the event scale, as a consequence of the compensation between the characteristics of the events, principally its magnitude and the atmospheric conditions (Gallart et al., 2002).

4.2 Forest and pasture transpiration

Tree transpiration during the studied vegetative periods (May–September from 1995 to 2000) showed important differences between wet and dry summers (Llorens et al., 2003b). During a wet summer (1995), mean soil water content was about 0.28 cm^3 cm^{-3}. Pine trees relative transpiration was about 45% (ratio between actual and reference evapotranspiration), with a marked temporal correlation with reference evapotranspiration, because there was enough water available for transpiration. During a dry summer, as in 1998, mean soil water content was only 0.18 cm^3 cm^{-3}. The great dependence of tree transpiration on soil water content clearly broke the relationship between transpiration and evaporative demand, with values of the relative evapotranspiration lower than 25%. In these conditions, trees showed a continuous decrease in transpiration rates with respect to atmospheric demand due to soil water depletion.

The study of actual evapotranspiration from a terrace with a mesophilous pasture (Poyatos and Llorens, 2003) showed that during the mild and warm season (mean air temperature higher than 7ºC), the relative evapotranspiration rate is close to the unity when volumetric soil moisture was higher than 0.30 cm^3 cm^{-3}, whereas it decreases down to 35% when soil moisture is about 0.20 cm^3 cm^{-3}. During the colder season, in spite of the high soil water availability, the relative evapotranspiration is low (about 30%) as a consequence of grass senescence, and increases progressively with decreasing albedo in springtime.

4.3 Soil water.

Gallart et al. (1994) showed that frequently saturated areas occur on downslope locations as a consequence of the outcrop of the shallow water table. The spatial pattern of these saturated areas is partly coherent with the topographic index values (Beven and Kirkby, 1979), derived from a 20m-mesh DEM, although some saturated areas do appear upslope, often in the internal parts of terraces.

The analysis of soil water content records (Gallart et al., 1997) showed that soils under forest cover are typically drier than under grass, due to rainfall interception, and that underground water transfer that feeds frequently saturated areas is interrupted during dry periods. Llorens et al. (2003a), using a longer time series, observed that spatial soil moisture variability is higher during intermediate wetness conditions and decreases during both wet and dry conditions.

Three representative soil moisture profiles were selected for a more detailed analysis of soil moisture regime along the year (Gallart et al., 2002). The first profile is located in a frequently saturated area downslope, covered by hydrophile grass. The two others are located in a mid-slope position, covered by pine trees and mesophile grass respectively.

The first profile shows a marked intra-annual variability (between 0.35 and 0.59 cm^3 cm^{-3}). In winter, the soil profile is always saturated and remains so until the end of spring. In June, the soil water content decreases rapidly due to the increasing evapotranspiration demand and the interruption of the subsurface water transfer. The lowest soil water content is reached by the end of July. Later, following the rainfall inputs of August and the successive months, soil water content tends to increase until the end of November. Saturation during the first part of the year and its breakdown in late June are predictable, as water content during this period presents low interannual variability, but the depth and duration of the dry period are much less predictable, as these depend on large rainfall events restoring the underground water transfer.

The other two profiles show a less pronounced intra-annual variability (between 0.23 and 0.38 cm^3 cm^{-3}). Soil water content decreases after the second fortnight of January and there is no increase in soil water content until early May. The summer drought includes currently June and July with the lowest soil water content during the second fortnight of July. Finally, there is a progressive wetting-up of the soils from August to the end of the year. The inter-annual variability for these two profiles was small and regular throughout the year.

The seasonal trend of the profiles described illustrates a catchment hydrodynamics pattern characterized by saturated areas surrounded by wet soils in early winter, that suffer a progressive drying until middle spring, although some saturated areas are still present. After an increase in wetness during late spring, the summer drought dries out the saturated areas. Finally, autumn is characterised by the wetting of the whole catchment, this being more pronounced in the saturated areas and more irregular and delayed in the forest-covered profiles.

4.4. Runoff generation.

From the very first results (Llorens, 1991, Llorens and Gallart, 1992) it was apparent that the response of these catchments is much more driven by antecedent conditions than by rainfall intensities. More detailed results (Latron et al., 1997, 2000, 2003; Gallart et al., 2002) showed that during the year, the dominant runoff generation mechanisms change gradually, as a result of both varying catchment antecedent wetness conditions and changing rainfall events characteristics (intensity and duration).

Using information on rainfall, stream flow, soil tensiometry and depths to water table, three main kinds of runoff events have been identified in the Vallcebre catchments (Latron, 2003):

a) Summer runoff events, which tend to occur as a result of short-duration, high-intensity convective storms over dry soils and deep water table. Runoff coefficients for this type of event are very low (typically less than 1%), as runoff is restricted largely to the poorly permeable rocky and badland areas of the catchment, resulting in a flashy hydrograph and low peak flow rates. Soil moisture responds to these events, but usually the water table does not.

b) Wetting-up transition events, which usually occur in autumn and eventually in spring, as a result of prolonged, lower intensity frontal rainfall over dry catchments with a deep water table, which rises up days after the rainfall event. Runoff coefficients are intermediate (3-15%) and recession limbs are relatively short. Overland flow is produced over areas saturated ‘from above’ where temporary perched aquifers occur due to limited permeability of the deep soil horizons.

c) Wet events, which occur in late autumn or early winter when large rainfall events occur on wet soils with shallow water table, which quickly responds to precipitation. Runoff coefficients are high (10–60%) and recession limbs contribute significantly to the flow volume. Overland flow occurs mainly on areas saturated ‘from below’.

An early analysis of the role of the terraces on runoff generation (Gallart et al. 1994) suggested that the saturated areas created by the terraces, along with the role of drainage ditches, would enhance the runoff by contribution from saturated areas of the catchment for a given water reserve, compared with the response for a non-terraced catchment, but this hypothesis has never been tested. On the other hand, the delays between rainfall and runoff peaks in the more terraced sub-basins Cal Parisa (0.18 km^2) and Can Vila (0.56 km^2), are respectively of 1.5 and 2.5 hours (Llorens, 1991; Latron, 2003), delays very long when compared with the common response of catchments where saturation overland flow is the main runoff-generating mechanism (Anderson and Burt, 1990)

The role of the scattered situation of the saturated areas (Latron and Gallart, 2002) linked to the terraces, and the pathways of water from these terraces to the main drainage net may explain this delayed response and deserve more attention.

5. Conclusion

During most of the year, the Vallcebre catchments behave as temperate wet ones; soil water content does not restrict plants transpiration, the evaporation of precipitation intercepted by canopies is driven by the duration of rainfall events, the subsurface flow along hillslopes feeds both saturated areas in downslope locations and stream baseflow, and the major part of stormflow corresponds to saturation excess overland flow over saturated areas during rainfall events.

Nevertheless, the increase of evapotranspirative demand during summer produces the depletion of the soil water that is not sufficiently replenished by precipitation; the limited soil water restricts plants transpiration and ceases the recharge of subsurface flow along hillslopes, producing the drying out of saturated areas and baseflow decline. Also during summer, the intensity of precipitations increase, rainfall interception events become more complex, and badlands are the only contributing areas to storm flow.

A large degree of spatial complexity must be added to this temporal arrangement: patches of forest and pasture influence local water balance and soil moisture, the agricultural terraces disturb the shallow aquifers, increase the spatial variability of soil properties and condition the formation of saturated areas, yet the artificial drainage ditches modify overland flow.

The above picture has been composed using field observations and different kinds of measurements, and several models were able to reproduce, at different extents, the observed behaviour. Nevertheless, none of the tested models attempted to cope with the complexity of some characteristics such as the terraced topography and the patchiness of vegetation. In fact, the used models, through the utilisation of ‘effective’ parameters, overrode these sources of complexity. Although most likely any model will be able to cope with all the complexity of any catchment (Beven, 2001), the progress on the knowledge of the actual hydrological functioning of catchments is necessary for the progress of both applied and scientific hydrology (Dunne, 1998; Beven, 2001; Sivapalan et al. 2003).

Stop 2.1 : “El Carot”. Badland weathering processes.

The study of weathering and erosion processes on “El Carot” badland areas was performed through the monitoring of several regolith properties using different field methods:

a) Regolith development and depletion, using bulk density as indicator on undisturbed samples (0-5 cm deep) with a periodicity of one or three times by month and along three years (Regüés, 1995; Regüés et al., 1995).

b) Regolith physical conditions: Surface regolith moisture (crust, 0-5 and 5-10 cm deep) using the weighing method on manual samples. Regolith hourly temperature profiles in three exposition conditions (north, south and flat), using sensors inserted in the crust and at 10 cm deep, during the same sampling period (Regüés, 1995, Unpubl. PhD Thesis; Regüés et al., 1995).

c) Regolith micro-geomorphological dynamics, was analysed with the support of surface stereoscopical photograps in four plots, one-two times by month and along one and half year (Regüés, 1995, Regüés et al., 1995).

d) Regolith response to simulated rainfall, analysed using repeated rainfall simulation experiments in two plots, two times per season during three years (Regüés and Gallart 2004).

The main weathering agent is freezing-thawing cycles during winter, which produce the breakdown of the soft bedrock, decrease bulk density below the unity (Fig. 5) and develop ‘popcorn’ features at the regolith surface, especially on the north-facing slopes. During spring, popcorn morphology is progressively eliminated, affected by the raindrop splash, inducing the ephemeral formation of sediment accumulation fans on the slope foots. During summer, the effect of the driest annual condition in combination with some convective rainstorms produces some wetting-drying cycles, induces crust development and surface sealing, which favours runoff and rainfall erosion rising and increasing the volume of footslope accumulations. This annual cycle is closed in autumn, when the crust development, the regolith depletion and the wetter conditions are favourable to runoff and erosion.

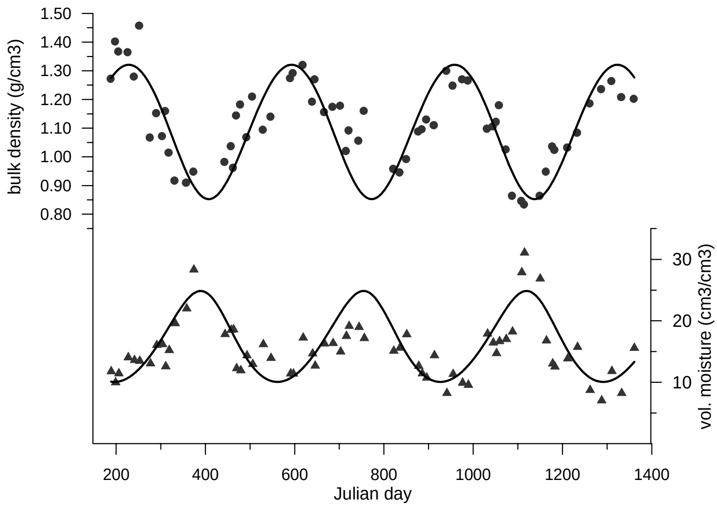

Figure 5: Regolith bulk density and moisture temporal evolution, symbols are measured values from samples and continuous lines represent modelled values (Regüés, 1995)

To verify these findings, the energy involved on weathering processes, associated to the water state changes, was modelled for northern and southern exposures using the temperature records. The results showed that the weathering energy was two orders of magnitude greater than the rainfall kinetic energy, whereas the differences between north and south expositions were about 3-fold (Fig. 6). Vegetation density and variety was greater in southern aspects, demonstrating the main role of low temperatures in the formation and dynamics of these badland areas (Regüés et al., 2000).

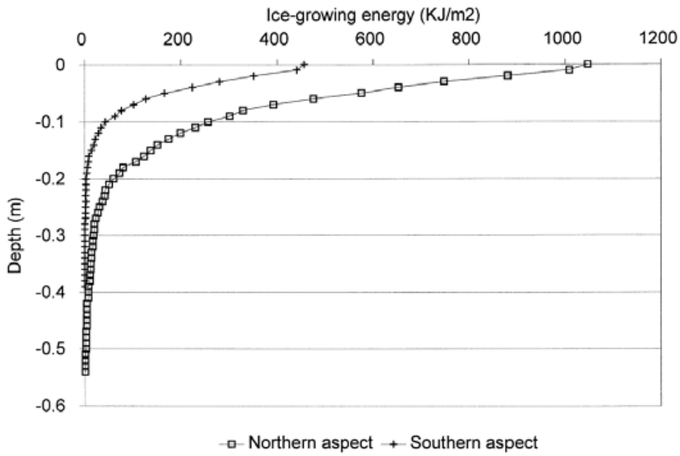

Figure 6: Potential weathering energy associated to ice growth on north and south expositions at El Carot (Regüés et al. 2000)

Stop 2.2 : Erosion and sediment yield at different scales.

The following results were obtained after several years of monitoring at different scales:

- Sediment yield rate from a small catchment without badlands (0.18 km^2) was only about 4 Mg $km^{-2}year^{-1}$, mainly coming from the channels, demonstrating the protective role of the old agricultural terraces and the vegetation cover (Llorens, 1991; Llorens et al., 1997b).

- Mean erosion rates at the badland surfaces, measured at the macroplot scale (1400 m2) during three years resulted in about 12 kg m^2 $year^{-1}$, mainly due to the high intensity summer rainstorms (Castelltort, 1995). The simulation of erosion rates at this scale using the KINEROS2 erosion model confirmed that relevant erosion is only active during high intensity rainfall (Martinez-Carreras, 2005). Nevertheless, erosion rates obtained with erosion pins (Clotet et al., 1988) were significantly higher (23 kg m^2 $year^{-1}$), suggesting that badland elements such as steep hillslopes, channels and banks suffer much higher erosion rates that the relatively gentle elementary catchments.

- The relationships between suspended sediment concentrations and water discharge during the events at the gauging stations showed a large scatter (three orders of magnitude) as well as diverse kinds of hysteresis loops (fig. 7), demonstrating the complexity of the runoff generation and sediment transport processes in the catchments. Clokwise hysteresis loops occur during the main events, when the catchments are wet, most of the runoff is produced by copious precipitation falling on saturated areas, and the high discharges erode and transport the sediments previously deposited in the channels. Counter-clockwise hysteresis loops occur during dry periods, when most of the runoff and sediment come from the badland surfaces because of the occurrence of high intensity rainstorms. Eight-shaped or complex loops occur in transition or composed events. (Balasch, 1998; Regüés *et al.*, 2000; Soler and Gallart, in press)

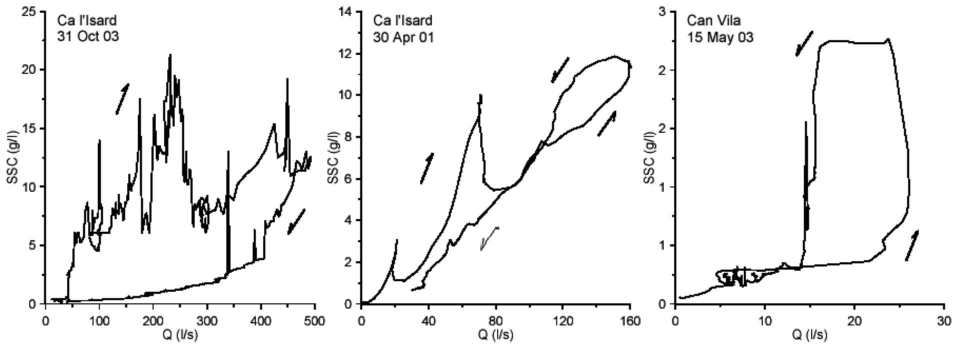

Figure 7: Examples of diverse kinds of hysteresis loops observed in the catchments. (Soler and Gallart., in press).

- Inter-annual scale sediment yield from catchments showed differences of up to two orders of magnitude (Gallart et al. in press). The drier years provided the lesser sediment transport, although the years with the highest sediment transport were not the wetter years, but the years with the larger runoff events. Figure 8 shows annual precipitation, annual sediment yield and averaged multi-annual sediment yield at the Ca l'Isard station. Inter-annual sediment yield from this figure was 1,400 Mg km^{-2} yr^{-1}, which is 2.8 times larger than sediment yield derived from plot measurements, but only 1.3 times larger than sediment yield derived from erosion pins measurement. Actually, if two years with extreme sediment yield (1997 and 1999) are excluded, the inter-annual average results in 535 Mg km^{-2} yr^{-1}, amount very close to the estimate obtained above from plot measurements. In average, sediment yield rates at Cal Rodó station are 4 times smaller than Ca l'Isard rates for the period 1995-1999.

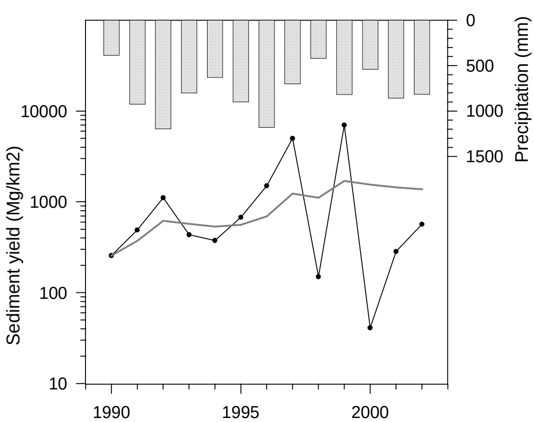

Figure 8: Annual precipitation, annual sediment yield and averaged multi-annual sediment yield at the Ca l'Isard station (Gallart et al., in press)

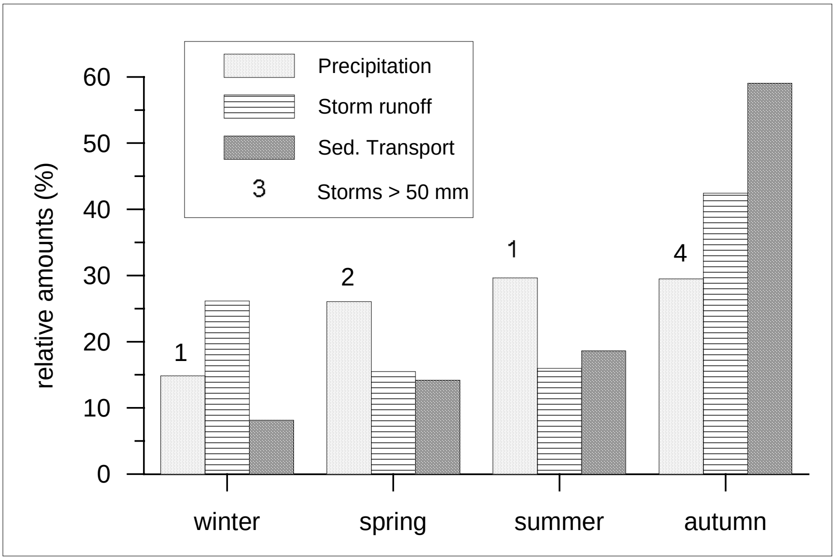

Figure 9: Relative seasonal precipitation, runoff and sediment transport at the Ca l'Isard station for the period 1995-99 (Gallart et al., 2002).

- Seasonal and monthly water and sediment volumes exported from catchments show a lack of correlation between them and precipitation characteristics, attributable to the main role of antecedent conditions and sediment conveyance delay respectively (Regüés et al., 2000; Gallart et al. 2002; Gallart et al. in press). High intensity summer rainstorms are of minor importance for flood generation and secondary for sediment transport at the catchment scale. Runoff is more important in autumn and winter because of the wetter conditions, whereas sediment transport is effected mainly in autumn (Fig. 9).

Afternoon: The Landslide of Vallcebre

1. *Introduction*

Last decades, considerable research effort has been devoted to the development of physically-based landslide models. These models, however, require a large amount of good quality data. Heavily equipped landslides -like that of Vallcebre- can be regarded as a natural laboratory. The existing monitoring records, which are stored in large data-base give the opportunity to calibrate and test the models.

The landslide of Vallcebre is a translational movement located in the Eastern Pyrenees, Spain, 140 km North of Barcelona (figure 10). The slide mass is 1300 meter long and 600 m wide with an estimated volume of more than 20 million of cubic meters (figure 11). Activity features such as tilted trees, and the presence of fresh scarps and cracks indicate that the slide is active. Moreover, some houses built on the landslide show also damages (cracking and tilting of the walls). First research activities in the landslide consisted of a detailed geological and geomorphological mapping, and preliminary monitoring using conventional surveying and photogrammetry was carried out since 1987 (Gili & Corominas 1992), and it showed that the landslide is continuously moving.

In 1996 this landslide was included within the frame of NEWTECH project, funded by the European Union, as a test site carry out both hydrological and mechanical modelling. Later on, the landslide has received additional funding from the Spanish Research Council (CICYT-MCYT). These projects allowed us the set up of a monitoring network. Eighteen boreholes were drilled in the landslide, equipped with inclinometer casings, piezometers and wire extensometers (Corominas et al. 1999). At the same time, several campaigns with differential GPS have been carried out to check their feasibility as landslide monitoring technique (Gili et al. 1999).

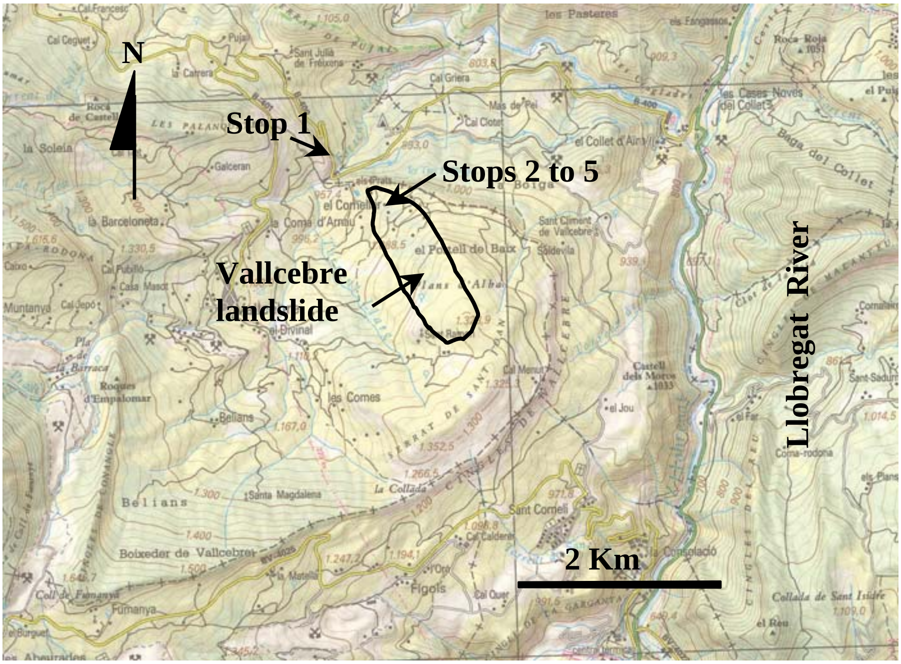

Figure 10. Location map of the Vallcebre landslide.

The objectives of the visit to the Vallcebre landslide are the following: (i) to discuss the geological context and observe the geomorphological features of the slide; (ii) to visit the monitoring network set up in the landslide to analyze its dynamics; and (iii) to present the main results on the mechanical and hydrological modeling.

Figure 11. General view of the Vallcebre translational slide. The main scarp can be observed in the background. The Vallcebre torrent runs from right to left and bounds the landslide foot.

2. *Description of the Vallcebre landslide*

The Vallcebre landslide has a stair-shape profile formed by three main morphological units of decreasing thickness towards the landslide foot. Each unit is formed by a gentle slope surface bounded by a scarp of a few tens of meters high. At the foot of each scarp, there exists an extension area that originates a graben. Northern and southern boundaries of the landslide are distinct lateral shear surfaces. The foot of the landslide reaches the torrent of Vallcebre and is being intensively eroded.

The unstable mass consists of a set of clayey siltstone layers sliding over a thick limestone bed (Upper Cretaceous - Lower Palaeocene age). The landslide takes place in the core of a gentle syncline, which axis is inclined more or less parallel to the ground surface. The borehole logs have provided direct information of the material involved in the landslide, from the bottom to the top: a) densely fissured shales, 1 to 6 m thick, showing slikensides; b) clayey siltstones, locally rich in veins and nodules of gypsum, and that also show intercalated gypsum lenses up to 5 meters thick. The logs have confirmed the presence of grabens between the landslide units, filled with colluvium.

The main direction of the movement is towards the northwest while a minor component of movement, towards the Torrent Llarg, is also observed in the upper slide unit. Figure 12 shows a geomorphologic sketch of the landslide and the location of the boreholes and targets.

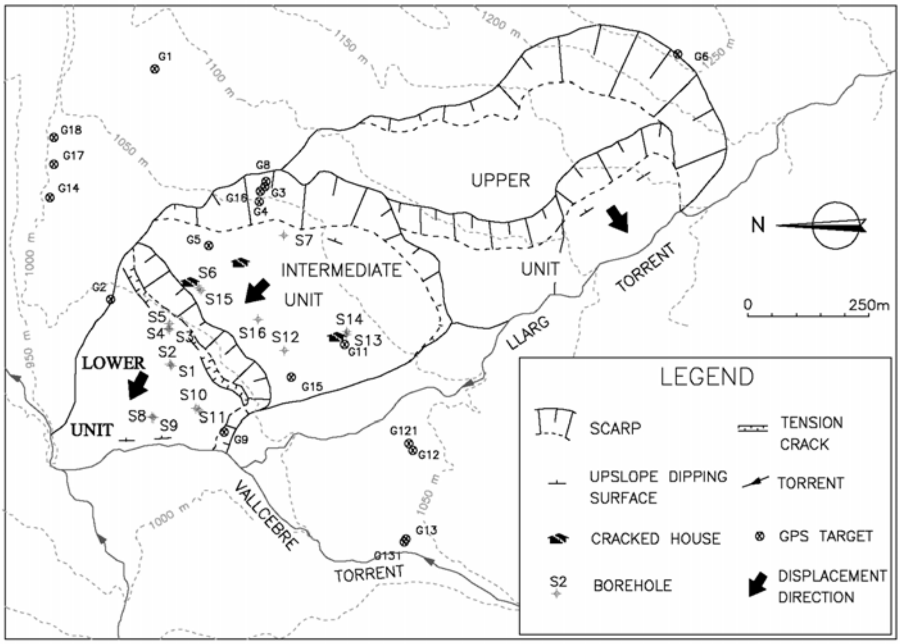

Figure 12.Geomorphological sketch of the Vallcebre landslide. The location of the boreholes and targets of GPS surveying is shown as well (Corominas et al. 1999).

The inclinometric profiles indicate that the failure occurs in a thin basal shear zone, with negligible deformation above it (figure 13). The shear zone runs along the fissured shale, close to the contact with the limestone and shows a gentle slope with an average inclination of 10º, similar to that of the ground surface. Inclinometric measurements also showed that the thickness of the slide mass is not constant. The lower slide unit (inclinometers S1, S3 and S8) has a thickness of between 10 and 15 m, whereas the intermediate unit (inclinometer S15) reach a thickness of at 45 m in the northern side and between 14 and 19 m in the southern one.

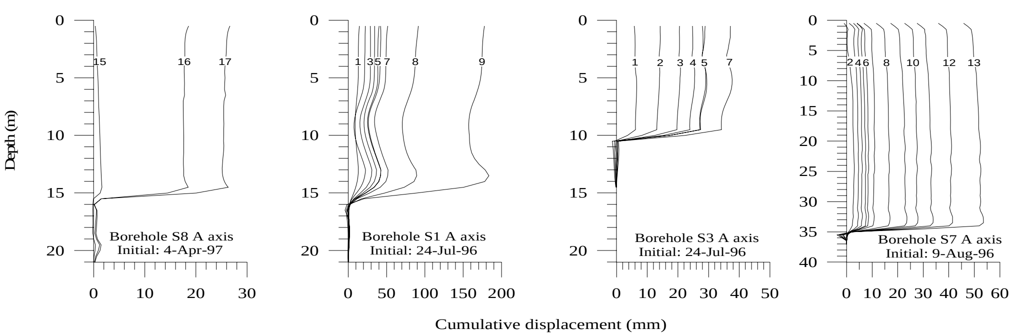

Figure 13. Inclinometric profiles. Numbers in profiles indicate campaign (i.e. number 7 was in 29-Oct-96) (Corominas et al. 1999b).

Even though the inclinometers have a short life when the landslide is very active, they have produced high quality information on soil displacement profiles, velocities and position of the shear surface. Superficial landslide displacements were also monitored by means of GPS techniques. 30 points were positioned on the landslide surface for periodic control. These points included reference points, fixed points adjacent to the landslide and targets within the landslide mass (i.e. buildings and upper ends of the boreholes). The GPS method used is based on a radiation with Real Time Kinematics GPS.

Automatic recording of the piezometers (each 20 minutes) provided critical information on the rapid water level changes due to rainfall. The piezometers showed an immediate response to the rainfall (figure 14). This fact suggests that water infiltration is controlled by fissures or macropores rather than by soil porosity. It is also observable that there is a practically simultaneous response of the piezometers (figure 15). Two basic types of responses to rainfall are observed depending on the location of the piezometers. The piezometers located in tension zones, as the S5, show smaller variations of the groundwater level (ranging between 0.5 to 2 m) and quicker drainage compared to the piezometers placed out of this zone (for example S2, S4 and S11). The latter ones experienced changes of 2 to 5 m and a slower rate of lowering of the groundwater level. We understand that the borehole S5 is located in one of this tension zones. The behaviour of the piezometer S5 is consistent with the presence of a very pervious zone. Consequently, it is assumed that cracks act as a preferential flow path within the landslide body.

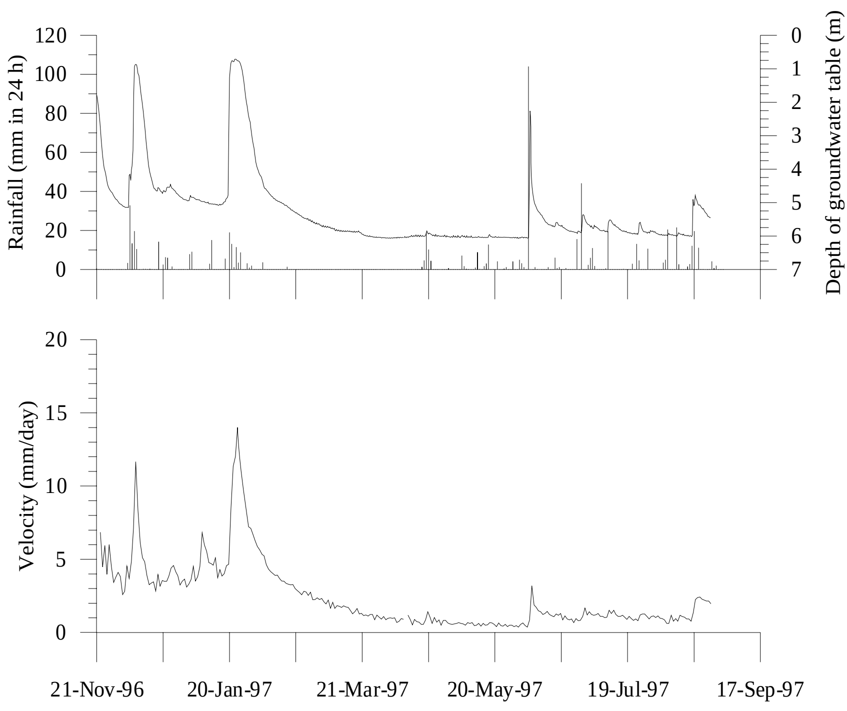

Figure 14. Above: rainfall (bars) and groundwater table changes at borehole S2. Below: rate of landslide displacement from wire extensometer readings at the same borehole (Corominas et al. 1999).

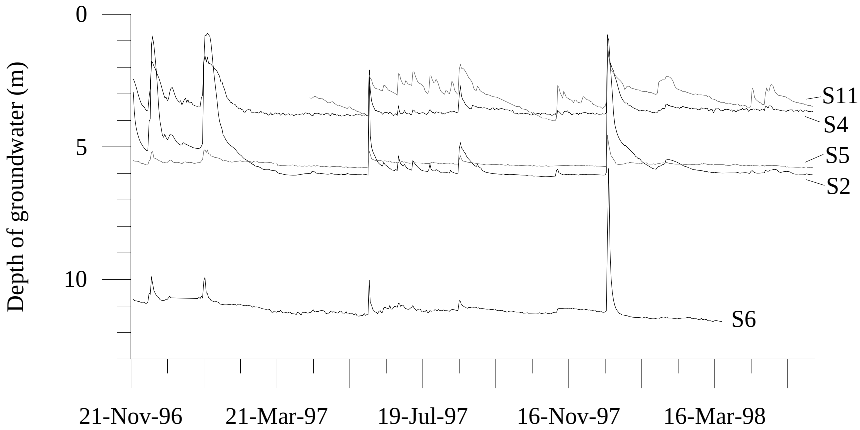

Figure 15. Piezometric records at the Vallcebre landslide.

Wire extensometers were specially designed by our group following an idea of Angeli et al (1988). It consists of a protected steel wire anchored to the limestone (below the slip surface) inside a piezometric pipe. The wire is kept in tension by means of a pulley and a counterweight of which rotation is continuously recorded using a potentiometer. The extensometric wire device has proved to be very useful in recording sudden changes in rates of displacements that can be directly related to the variations of the groundwater table and indirectly, to the rainfall. It is especially convenient when borehole inclinometers have been lost after large displacements.

Stop 2.3: General view

General view, from the west, of the Vallcebre landslide. Explanation of the landslide context, landslide boundaries and units, geological and geomorphological context. Description of the works carried out in the landslide since 1996. Modelling results and discussion on the viscous behaviour of the movement.

Stop 2.4: Landslide toe

Walk from the bottom of underlying limestone layer up to the foot of the Vallcebre landslide. We will follow a narrow pathway excavated in the rock through the gorge of the Vallcebre torrent up to a bridge located just North of the landslide. There we will observe: (i) the outcrop of fissured shales which are the materials where the rupture surface has developed; (ii) the landslide foot overriding the Vallcebre torrent bed; and (iii) the associated instability consequences on the opposite slope.

Stop 2.5: Features near the stream

Observation of the (i) instability features at the landslide foot associated to the erosion of the Vallcebre torrent; (ii) groundwater features (seepage front, springs) associated to the landslide and weir for water flow measurement; (iii) discussion on the presence of a hidden torrent bed

Stop 2.6: Instrumentation

(i) Equipment for measuring landslide displacement (wire extensometer and inclinometer); piezometers; and data transmission system; (ii) development of the graben between landslide units;

(iii) the piping features; (iv) discussion on the role of both graben and piping features in the hydrological behaviour of the landslide

Stop 2.7: Lateral shear surface
Observation of the lateral shear surface of the landslide. Northern boundary

Thirst day: September 14th
The Cardona Salt Formation
During the Upper Eocene (Priabonian) the central part of the Catalan Central Depression was occupied by a marine transgression. Under this regime take place the deposit of the Igualada marls, which pass vertically into the Cardona Saline Fm. Overlying these marls, the reef limestones of La Tossa de Montbui Fm. are developed on the S, SE and N basin edges. The Cardona Saline m., widely extended, constitutes an evaporitic macrocycle which begins and ends with sulphate facies. The saline Fm. is composed of rock salt and potash beds (150-300 m thick). It crops out in the Cardona diapir, where is strongly disturbed by halokynetic movements. Towards the top, the Cardona Saline Fm. passes into the “top gray lutites” with Halite and Gypsum layers interbedded. This unit is only visible in boreholes. Finally, terrigenous influxes coming from the basin borders have contributed with major sedimentary masses to the continental infilling of the Catalan Central Depression, during the Upper Eocene and the Oligocene. These are the Artés Molasse, proceeding from the Catalan Ranges and the Solsona Molasse, proceeding from the Pyrenees.

An isobath map of the seismic reflector placed at the bottom of the Cardona Saline Fm. shows that the lower surface of the salt is flat or gently deformed. It contrast with the top which configures a disharmonic set of diapiric folds with direction SE-NW and ENE-WSW, developed at levels stratigraphically higher than the Cardona Saline Fm. A late Sannoisian tectonic phase reactivated the Frontal South Pyrenean Thrust gliding on the decollement level placed at the bottom of the salts. The halokynetic movements would initiate simultaneously. On the other hand, it seems likely that the lineations of diapiric folds would be related to the fault structures affecting the basement of the Cardona Saline Fm.

The stratigraphy of the salt formation may be summarised as follows:

- A basal sulphate-rich unit, 5 to 30 m thick, that outcrops from Vic to Igualada. Secondary gypsum in three units: a stromatolitic lower, a massive with Selenite pseudomorphs, and a laminated detritic upper.

- An upper chloride-rich unit, about 300 m thick in the central zone (without diapiric thickening) that disappears towards the marginal areas of the basin. The lower part is formed by banded Halite; the upper part is formed by potassium ores at the bottom (Silvinite and Carnalite), and laminated Halite at the top. Over the Halite, there is a 35m-thick cover of laminated Anhydrite.

There are four main locations for salt mining: Sallent (with the larger landfills), Balsareny, Súria and Cardona, where the salt formation outcrops.

Stop 3.1: Environmental problems related to the mining activities.
Surface water pollution is a major problem related to the salt mining. There are two main sources of pollution: the spills from the factories and the waste mounds. The first problem was partly fixed with the construction of a pipe that conveys the brine from the factories to near the sea.

Land subsidence because of the closing of mine galleries is another relevant issue. It caused serious problems in the buildings at Sallent.

Karst phenomena were not of much importance in the past, but in the last years, due to the pumping of water inside the mine galleries, the hydraulic gradient was reversed and the water from the Cardener river started recharging the salt aquifer. As a result of that, a large sinking hole opened in the river bed, threatening the mining activity and the long term quality of the river water downstream. The solution adopted was the diversion of the Cardener River through a tunnel that cuts off the meander were the sinkhole occurred.

Stop 3.2: The Salt Mountain and the Mina Nieves.
The Salt Mountain is the outcrop of the Cardona Salt Formation, where the dissolution of the salt by rain water is compensated by the upwards vertical movement produced by the diapirism. This geomorphic feature is known since the Roman times.

The stop includes visits to the Sal Mountain, to a photographic exposition about the history of the mining works, and to the Mina Nieves, along 500 m of galleries.

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ROAD LOG:

September 12th. Departure from the Conference Hall at 9:00

0-270 km	A-2 (E90) Motorway to LLeida, C-1313 road to Basella, LV-310 road to Solsona, LV-4241 road to Coll de Jou. Stop at the junction with the track that goes to the **belvedere** ‘Mirador de la Creu del Codó’. About 45 minutes round trip walk along a fair track at 1500 m a.s.l. Scenic view of the southern border of the Pre-Pyrenees and its contact with the Ebro Depression.
270-276 km	LV-4241 road to Sant Llorenç de Morunys. **Lunch** at La Brasa restaurant. Free visit to the medieval village (XIII Century).
276-281 km	LV-4012 road to La Coma. 15 minutes walk to examine the **earthflow** occurred in 1982. Free visit to the **karstic springs** of the Cardener stream.
281-310 km	LV-4012 road to Sant Llorenç de Morunys and LV-4241 road to Berga. Free visit to the town, founded by the Romans (*Castrum Bergum*) and head of the Berguedà District. **Dinner and night** at the Ciutat de Berga Hotel.

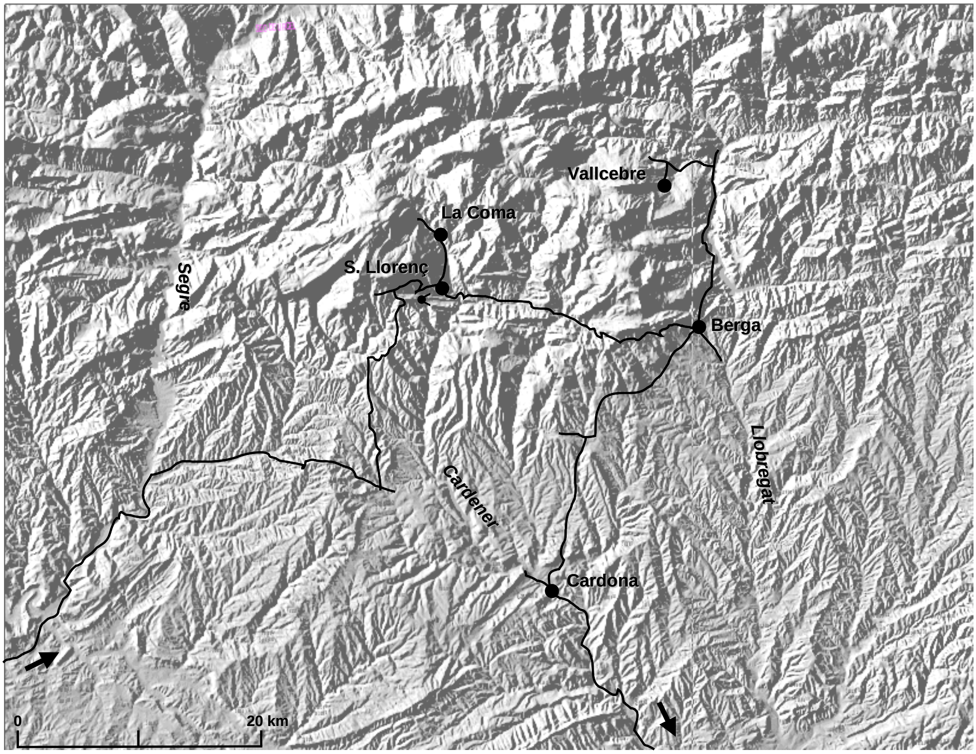
September 13th.

310-332 km	C-16 road to El Collet, B-400 and B-401 roads to Vallcebre. Morning: visit to the research sites for **hydrological and erosive processes**. Picnic **lunch**. Afternoon: visit to the **instrumented landslide**.
332-354 km	B-401 and B-400 roads to El Collet, C-16 road to Berga. **Dinner and night** at the Ciutat de Berga Hotel.

September 14th.

354-409 km	C-149 and B-420 roads to Cardona. This town has the more impressive medieval castle in Catalonia. Morning: visit to the **salt mines** and **Salt Mountain**. Free visit to the medieval village. **Lunch** at the Can Pujol restaurant. Afternoon: visit to the **salt dissolution** and collapse problems at the Cardener river.
409-705 km	C-1410, C-55 and C-25 roads, and A-2 motorway to Zaragoza.

Expected arrival time to the Conference Hall: 19:30.


Vallcebre
La Coma
S. Llorenç
Segre
Berga
Llobregat
Cardener
Cardona
0
20 km