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PROCESSES, SPATIO-TEMPORAL FACTORS AND MEASUREMENTS OF CURRENT EROSION IN THE FRENCH SOUTHERN ALPS: A REVIEW

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ABSTRACT

Present erosion in mountainous areas of Western Europe causes land management problems, particularly for areas located downstream of erosion zones. Except for transalpine roads and ski resorts, economic activities no longer require as much space as they did in the past. Therefore, natural reforestation has provided significant protection for alpine hillslopes during the 20th century. However, extreme floods continue to cause severe damage in intra-alpine valleys, as well as in piedmont and surrounding plains, making the study of present water erosion phenomena very important.

Many studies have investigated the processes and factors of water erosion on slopes at both the catchment and plot scales. They have focused on rock fragmentation and transportation in different fields, the spatial and temporal explanatory variables, the consequences downstream (flooding, sedimentation, river bed evolution) and the impact of floods.

In the French Alps, present erosion has been studied in a variety of outcrops, with several recent studies conducted in fields such as marls, clayey deposits, molasses and moraines. These kinds of outcrops are found throughout the alpine massif, including an area of special interest on the great Jurassic black marl outcrop where badlands are frequently observable. Geomorphologists and hydrologists have been particularly interested in the strong erosion processes in marls, seeking to determine the main patterns and the impact of spatial and temporal factors on soil loss quantities.

The main climatic factors of rock disaggregation were found to be the freeze–thaw and wet–dry cycles, which destroy rock cohesion, and the splash effect of rain. The principal site variables are vegetation cover, exposure and dip–slope angle. Erosion rates are two or three orders of magnitude higher on bare soils than on pastures; northern aspect slopes suffer two to four times as much soil loss as southern aspect slopes. Finally, the angle formed by the slope and the dip also determines different behaviours: erosion rates are higher when slope and dip are perpendicular than when they are parallel. The transportation agents are mostly debris flows and runoff caused by intense precipitation. Annual erosion depth in the marls is generally assumed to be substantial, up to 10 mm. The high value can be explained by the severity of the climatic conditions and the brittleness of the lithology, which results in numerous fractures.

KEY WORDS: solifluction; measurements; erosion processes; Southern Alps; erosion factors

INTRODUCTION

From the end of the Middle Ages, inhabitants of alpine valleys have observed that erosion and heavy rainfall effects were linked to deforestation. Similarly, the archives of the Little Ice Age also show that intense erosion of some areas caused significant changes in stream dynamics and in plains sedimentation downstream. Floods and low flows were accentuated during these periods, sediment deposition was accelerated in the alluvial cones and alluvial plains. River bed aggradation increased inundation hazard in downstream plains (Blanchard, 1945; Descroix, 1994).

In the mountains of the Mediterranean, eroded areas have expanded due to the conjunction of weathering of the lithology, the occurrence of intense rainfalls and demographic pressure on the environment. The impact of these factors on badlands and eroded landscape formations was analysed recently in Greece (Kosmas *et al*., 2000), as well as in southern Italy (Moretti and Rodolfi, 2000) and in southern Spain (Nogueras *et al*., 2000). In all these areas, soils have suffered substantial degradation in the last centuries.

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In the North Mediterranean basin, demographic pressure is no longer a contributing factor. However, eroded areas remain. Badlands in the Southern Alps, most often in marls, are an example of the morphology of these areas. The significant extension of badlands in relatively homogeneous outcrops provides a natural laboratory in which to study erosion in mountainous areas. The famous Oxfordo-Callovian and Upper Bathonian black marls, more than 2000 m thick, as well as those of the Toarcian, are known for their particular landscape. Many measurements, field surveys and modelling attempts have been carried out on this kind of generalized gullying to determine water erosion processes and their spatio-temporal variables (Descroix and Olivry, 2002).

Nowadays, substantial water erosion continues in already eroded hillslopes such as badlands in the Jurassic marls, which remain the main sediment supply zones because degraded zones have not revegetated and therefore continue to be an important source of deposits. Hydraulic constructions can partially control floods but they are still subject to intense sedimentation. Vegetation progression from reforestation and overgrowth in the southern Alps currently limits the development of badlands. In addition, excessive extraction of materials from river beds and embankments has caused what is called a sedimentary deficit (Descroix and Gautier, 2002). Therefore, an inverse problem has arisen in the North Mediterranean basin.

The influence of the lithology enhanced by rural desertification has led to an increase in solifluction processes, which poses severe problems in terms of mountain management. As a direct result of rural desertification, the canals, ditches and drains, retaining walls, and other constructions built in past centuries are no longer maintained and cause water logging, thus inducing solifluction processes. These processes are difficult to foresee.

The aim of this paper is to present a literature review of the recent advances in erosion research in the French Alps for the English-speaking research community. All references can be obtained from the authors of this paper.

This review focuses on the main themes related to weathered-terrain erosion in the Southern Alps. Many other reviews describe other aspects of alpine erosion:

- karst erosion, on limestone (Delannoy, 1984) or on gypsum (Chardon, 1992; Chardon, 1996);
- periglacial processes, scree formation, gelifraction on rock walls (Francou, 1988; Pech, 1995; Rovera, 1990, 1995);
- overgrazing and its consequent landforms such as the formation of terracettes on alpine pastures (Serrate, 1978);
- evolution of stream beds, gravel banks and islands as a consequence of sedimentary balance and heritage of mountain erosion (Piégay and Salvador, 1997; Piégay *et al*., 1999; Vautier, 2000; Liébault and Piégay, 2001);
- evidence of palaeoclimates and their influences on hillslope erosion, sedimentation and river bed evolution (Jorda *et al*., 1991; Gautier, 1992; Rosique, 1996; Ballais, 1997; Miramont, 1998);
- the role played by heritage and anthropization in alpine erosion (Descroix and Gautier, 2002);
- the impact of extreme flood events such as the millennial inundation of the Guil River in 1957 (Tricart, 1975), which induced a runoff coefficient greater than 1 due to rain falling on snow cover; more recently, the flooding of the Ouvèze river at Vaison-la-Romaine was the subject of a number of papers on the damage caused to vineyards (Wainwright, 1996) as well as stream beds and banks (Piégay and Bravard, 1997);
- consequences of major collapses, for example, the breakdown of the so-called Infernet occurred in the 12th century in the Romanche Valley, which devastated the city of Grenoble (Muller, 2000); or the collapse of Mont Granier (Figure 1) in 1248 (Chartreuse, northern Alps) (Goguel and Pachoud, 1972); the Claps de Luc collapse (Diois, Southern Alps) obstructed the Drôme valley in 1442 and the lake formed by this natural dam was filled over three centuries (Froment, 1973);
- influence of mountain management: gully erosion and solifluction were sometimes provoked by the development of ski resorts in the Northern Alps (Eglise and Ravoire Torrents downstream of Les Arcs resort, in the upper Isère valley, Savoie; Périnet, 1982) as well as in the Southern Alps, for example in Vars (Martin and Weber, 1996).

This study is limited to the French Southern Alps. Research based on other alpine areas or in Provence (Martin *et al*., 1991, 1997), where problems are slightly different, will not be developed here. The *Revue de Géographie Alpine* (1996) has published a special issue on alpine erosion processes.

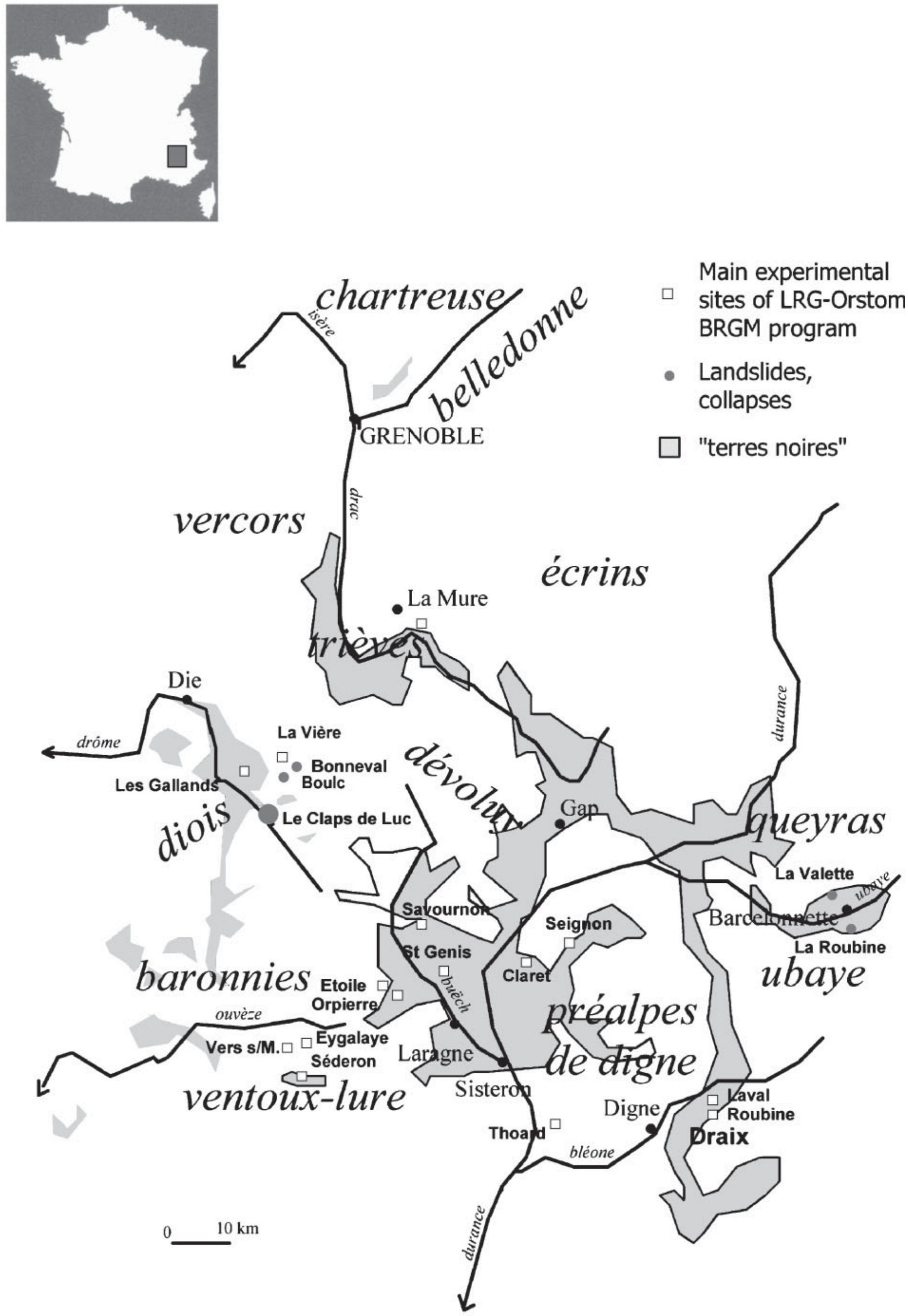

Figure 1. Location of the ‘terres noires’ areas in the French Alps

Firstly, this review presents the erosion factors and processes that vary according to spatial and temporal factors; then materials and methods will be described, followed by the analysis of measurements and their results, organized by catchment area, including dam sedimentation. Finally, it will be shown that solifluction processes are an increasing source of problems and must be taken into account in mountain management.

EROSION FACTORS AND PROCESSES

The black marls ('terres noires') are the most important erodible outcrop in the French Alps. They are 1500 to 2500 m thick and have a homogeneous facies and geotechnical behaviour throughout the entire thickness (Artru, 1972; Phan, 1993). The badlands in this region broadly correspond to the black marls. They are the subject of another paper (Descroix and Olivry, 2002); the present review is limited to alpine soft terrains.

Climatic factors

Most authors give great importance to climatic factors, particularly cycles of freezing–thawing and wetting–drying action, and their interaction (Descroix, 1985; Deshons, 1985; Peyronnet, 1988; Olivry and Hoorelbeck, 1990; Bufalo, 1989; Brochot and Meunier, 1995; Rovera *et al*., 1999b). The bedrock is weathered during freeze and thaw cycles, which are only efficient if rock water content is significant and if there are wet periods during the winter. Rainfall and snowfall generally correspond to periods of increased temperature during the winter. It is commonly agreed (e.g. Lecompte *et al*., 1998) that two opposite phases can be distinguished.

(1) During the winter season, materials are prepared and detached by cryoclastism, and often transported by solifluction processes. The rise in temperature during the spring results in a great accumulation of materials on slopes and at the bottom of hillslopes.
(2) During the summer season, characterized by intense and short rainfall events, the main morphogenic action is caused by water erosion: materials detached during the winter are removed by splash and transported by overland flow. Hydroclastism is active during the summer also because of a different marl element inflation coefficient (Birot, 1981).

The seasonal opposition results in the development of a thick layer of regolith in the badlands. This layer, locally up to 50 cm thick, strongly influences erosion patterns on marls and other soft terrains.

According to Brochot and Meunier (1995), the high porosity of the marls (6·8 per cent as opposed to 1·6 per cent for limestone), schistosity, low carbonate content and high density of joints explain the rapid weathering of marly terrains. Marl fracturing begins as soon as it outcrops, as can be observed when collapses or landslides open fresh cuts. Collapses are relatively frequent in the marls due to the steepness of the slopes on badlands (from 35° to 45°). Such a collapse occurred while the Laval station access path (Draix catchments) was being dug in 1987 (Figure 2) (Descroix, 1994). Near Barcelonnette, Malet *et al*. (2000) and Schmutz *et al*. (2000) observed that marl blocks that had collapsed 40 years before and had been overlapped by the landslide remained intact until they were exposed; subsequently, as noted by Malet *et al*. (2000), the blocks suffered rapid disaggregation due to wetting–drying and freezing–thawing cycles. Lecompte *et al*. (1998) evoked decompression as a likely cause of the strong penetration of disaggregation as and when superficial regoliths were being removed. However, rock weathering is probably a result of both the great porosity of the regoliths and the efficiency of freeze–thaw cycles.

The mean annual number of freezing days ranges from 80 to 160 in the French Southern Alps: for example, 104 at the Laragne station (Figure 1) (Descroix, 1985) and 114 at Marcoux, near Draix (Oostwoud Wijdenes and Ergenzinger, 1997). Freezing–thawing cycles are more frequent on north-facing slopes ('ubac'). As a consequence, mud flows related to the thaw occur generally on north-facing slopes, where the soil water content decreases very slowly during the spring thaw (Olivry and Hoorelbeck, 1990; Descroix, 1994). In the Draix experimental catchments, Oostwoud Wijdenes and Ergenzinger (1998) have named these processes miniature debris flows (MDFs), which are widespread on bare marly fields. At the same site, Brochot and Meunier (1995) observed a laminar solifluction on south-facing slopes during the winter, as a result of the daily thaw. All these authors agree on the importance of these processes on marly slopes, related to cryocreeping, a combination of cryoturbation and creeping. Finally, Bufalo (1989) and Descroix (1994) observed a process previously analysed by Birot (1981): freezing and wetting of soil and/or regoliths induces a field surface inflation that is parallel to the slope while thawing or desiccation leads to vertical packing due to gravity: the overall result is a recurrent creeping of the regolith layer down the slope.

Figure 2. Blocks of fresh marl just after a collapse near Digne (Alpes de Haute-Provence, France), 1987. It is noticeable that rock fragments are not yet weathered and constitute solid blocks. Gravity remains the same; the excavation made for the road leads to collapse

Strong intensities and high kinetic energy values of rainfall from the end of summer to the beginning of autumn are considered the main variables explaining transportation by most of the authors who have studied erosion at the rainy event scale (Deshons, 1985; Descroix, 1985, 1994; Bufalo, 1989; Olivry and Hoorelbeck, 1990; Chodzko *et al*., 1991; Descroix and Olivry, 2002). These events frequently induce the formation of hyperconcentrated flows (with more than 500 g l^{-1} of solid particles): flows with more than 800 g l^{-1} were measured at Draix (Olivier, 1995), more than 600 g l^{-1} at the Savournon experimental site (Olivry and Hoorelbeck, 1990) and more than 400 g l^{-1} at the Orpierre site (Descroix, 1994). This type of flow reaches a competency that allows the transportation of very coarse elements.

Climatic conditions, particularly freezing–thawing cycles and precipitation, are frequently considered as determining factors of the occurrence of slope instability phenomena. The varved clay outcrop near La Mure (Isère) attracted great attention from a team of geographers from the Netherlands. Van Genuchten and Van Asch (1988) proved that the movements were linked to climatic, hydrological and morphodynamic factors. The cumulative displacements were correlated with the total annual rainfall. At the hour time scale, movements depend on the interstitial water pressure. Winter precipitation explains the sliding better than summer rainfall, which plays no role (Van Genuchten, 1989; Van Genuchten and De Rijke, 1989). The same team demonstrated that slow slides could be modelled, taking into account shearing stress and creeping principles (Van Asch and Van Genuchten, 1990).

However, in an extensive study on the climatic conditions of the Barcelonette and Vars basins (Alpes de Haute Provence), Flageollet *et al*. (1999) showed that these are important factors, but they alone cannot trigger slope destabilization. Seismic or anthropogenic factors are also necessary.

Site factors

Site variables include spatial factors of erosion, as opposed to event characteristics (mainly climatic variables). Black marl outcrops present a relative homogeneity and thus a great variety of site configurations exist. The gradient of completely bare hillslopes in badlands always ranges from 35° to 45° (Deshons, 1985; Descroix, 1994; Lecompte *et al*., 1998). Thus, the slope value does not appear as an explanatory variable of erosion. The following factors explain soil loss.

Vegetation. Since the end of the Middle Ages, inhabitants have noted the relation between deforestation and erosion. Significant evidence of this knowledge is available in historical archives (Blanchard, 1945).

The importance of the vegetation cover in black marls was emphasized by Mathys *et al*. (1996) at the Draix experimental catchments: the sediment yield of the Laval catchment (86 ha, with a vegetation cover of 32 per cent) is 40 times as great as in the Brusquet basin (108 ha, with a vegetation cover of 87 per cent). Crosaz (1995) and Crosaz and Dinger (1997) tried to vegetate badlands at the Draix catchments. In the same experimental site, Lukey *et al*. (2000) modelled the impact of reforestation at the basin scale using the SHETRAN model. In the Baronnies mountains, Cohen (1998) determined the main role played by vegetation density and disposition on badlands in reducing soil loss. Similarly, Rey *et al*. (1998) highlighted the influence of forests on erosion and the management of erosion in the Southern Alps. This is consistent with observations made on other parts of the Mediterranean basin (Sorriso-Valvo *et al*., 1995).

Exposure. At the Savournon and Saint Genis experimental catchments (Figure 1), Deshons (1985) observed that the regolith surface layer is thicker on south-facing slopes, because of the greater intensity of freeze–thaw cycles; on the other hand, using experimental spot devices at the same site, soil loss is always found to be two to four as great on north-facing slopes than on south-facing slopes (Descroix, 1994). This result gives rise to an apparent contradiction: the measured values of soil loss are higher on the hillslope which is subject to fewer freeze–thaw cycles. However, Rovera *et al*. (1999b) demonstrated that the number of freeze–thaw cycles at a depth of 6 cm is significantly higher on north-facing slopes (ubac) than on south-facing slopes (adret). In the Seignon catchment, 83 cycles were measured in the air; at a depth of 6 cm, only 12 cycles were observed on the adret, while 41 cycles were produced on the ubac. Therefore, it seems important to consider the temperature of the regolith instead of the air temperature. Nevertheless, it should be mentioned that these measurements were taken during a relatively mild winter; the mean number of freezing days in the year ranges between 100 and 105 in this area. In the coldest winters, it is assumed (Descroix, 1994) that the number of freeze–thaw cycles does not increase; in some cases, the persistence of very low temperatures can even induce a lower number of cycles on north-facing slopes than on south-facing slopes, a daily thaw on the latter the result of longer exposure to the sun.

In the Méouge valley, higher values for soil loss on adret slopes were originally recorded (Chodzko *et al*., 1991); later, it was concluded that exposure played no role in erosion yield (Lecompte *et al*., 1998). In the Seignon catchment, Robert (1997) emphasizes the difference between processes: the more disaggregation there is on the adret, the more solifluction processes there are on ubac. However, it has been demonstrated that the north aspect induces higher erosion rates (Descroix and Olivry, 2002).

Slope–dip angle. The angle formed by slope and dip is also an important discriminating factor. Surfaces parallel to the dip are usually more resistant to weathering, while cross-cut outcrop is more subject to both freezing and wetting front penetration (Deshons, 1985; Descroix, 1994). On the contrary, a perpendicular dip–slope configuration favours solifluction processes, which are less significant on cross-cut outcrops; in the latter case, soil particles or rock fragments are retained on the slope and do not creep. It has been observed, on both black and blue marls, that the dip–slope angle had an influence on soil loss values (Descroix, 1994). In the Draix experimental catchment, Oostwoud Wijdenes and Ergenzinger (1998) demonstrated that there is a great difference between miniature debris flows (MDFs) on hillslopes, depending on whether dip and slope are perpendicular or parallel: in the first case, the particles are smaller; in the latter, particles are bigger, and the transportation of detached material is easier due to creeping along joint planes. These authors observed that for a similar bulk density (ranging from 1·5 to 1·8), the grain size distribution was different.

The dip–slope angle has a great influence on mass movement classification, as has been clearly demonstrated by Peyronnet (1988), who defined three types of microlandslide that affect the badlands hillslopes, depending on the dip–slope angle and the depth of the regolith layer. Nevertheless, in some cases, irrespective of other factors such as water content and freeze–thaw cycles, gravity and human action alone can account for triggering collapse and other mass movements (Figure 2).

Other site factors. The position of the measuring site on the slope is of great importance. In the Eastern Baronnies (Lecompte *et al*., 1998) and in the Préalpes de Digne (Rovera *et al*., 1999a), the following opposition has been observed:

- interfluve areas continuously experience erosion processes;
- down-slopes and the bottom of the gullies are exposed to an alternating pattern of winter accumulation phases and summer transportation phases.

However, the slope value remains constant throughout the badlands. This can be explained by an annual balance, with the bottom of the gullies being excavated at the same rate that interfluves and slopes are eroded. When the excavation of the gully reaches a resistant calcareous horizon, sediment accumulation fills the bottoms of the gullies, and the slope value decreases. This configuration is widespread in blue marls, which are less thick and less homogeneous than the Oxfordo-Callovian black marls, and include stronger calcareous layers. Moreover, because of their different geologic age, the typical configuration exposes black marl outcrops to weathering in anticlinal positions; blue marls filling the synclines (Figure 3) are better protected. In this latter case, the elevation difference with the base level, and consequently the excavation capacity, are lower (Descroix, 1994).

In the Barcelonnette basin, Mulder and Van Asch (1989) used multiple statistical analyses to highlight the role of hydrological and topographical factors in slope instability. They proposed a deterministic model based on the hillslopes geotechnical equilibrium, determined by the ratio of slide resistance to triggering forces of these phenomena.

METHODOLOGIES FOR MEASURING SOIL LOSS

Methodological devices can be divided into two groups:

- those that measure soil loss *in situ*;
- those that measure erosion and material transportation at the outlet of a plot or catchment.

In situ *measurement*

The following list is not exhaustive, but presents some of the most important studies.

Site and rainfall variables were investigated in the Baronnies, Diois and Préalpes de Digne massifs (Descroix, 1985, 1994). The devices were simple but used at several scales: plots and sediment traps, on plots ranging from 1 to 10 m^2, microcatchments from 20 to 200 m^2, and catchments from 5 ha to 90 km^2.

In 1983 CEMAGREF started equipping the Draix experimental catchment (Préalpes de Digne), which remains the main mountain erosion field laboratory in France (CEMAGREF, 1987, 1995, 1997). The difficult measuring conditions compelled this team to develop experimental metrology devices, adapted to the measurement of flows with a very high suspended load (up to 800 g l^{-1}), as well as the following devices:

- experimental sediment samplers, which make it possible to measure sediment concentration; this device has frequent clogging problems;
- a turbidity assessment device based on a differential pressure sensor has been developed to measure the suspended load for higher rates (100 to 500 g l^{-1}) (CEMAREF, 1995);
- an optic sensor for suspended load measurement was developed by the same team; a gamma radiation absorber gauge was tested in the experimental catchment (CEMAGREF, 1995);
- water level recorders based on ultrasound sensors (CEMAGREF, 1995);
- experimental large-volume sediment traps (CEMAGREF, 1987, 1995, 1997).

In the Buëch valley, Olivry and Hoorelbeck (1990) attempted to measure erosion in black marls using graduated needles driven into the soil, through the regolith layer, perpendicular to the surface. But these fragile devices were quickly exposed by both excavation and solifluction.

Difficulties related to cryo-ejection were resolved in the Savournon catchment (Olivry and Hoorelbeck, 1990) using a portable microprofile meter (Figure 4), devised to measure erosion depth and placed on fixed rods. Although rods were driven 50 cm into the soil, they were still subject to cryo-ejection (devices are ejected from the soil or regoliths due to freeze–thaw cycles); some of them were exposed and ejected by the very first cycles. Consequently, at a second stage, rods were concreted and the problem avoided for at least six years (Descroix, 1994). Moreover, the microprofile meter avoids inaccurate measurements caused by scientists' activities on the site. This principle has recently been adapted by Rovera *et al*. (1999a) and Robert (2000), who designed a micrometric measurement device to study erosion depth in the Draix experimental catchment.

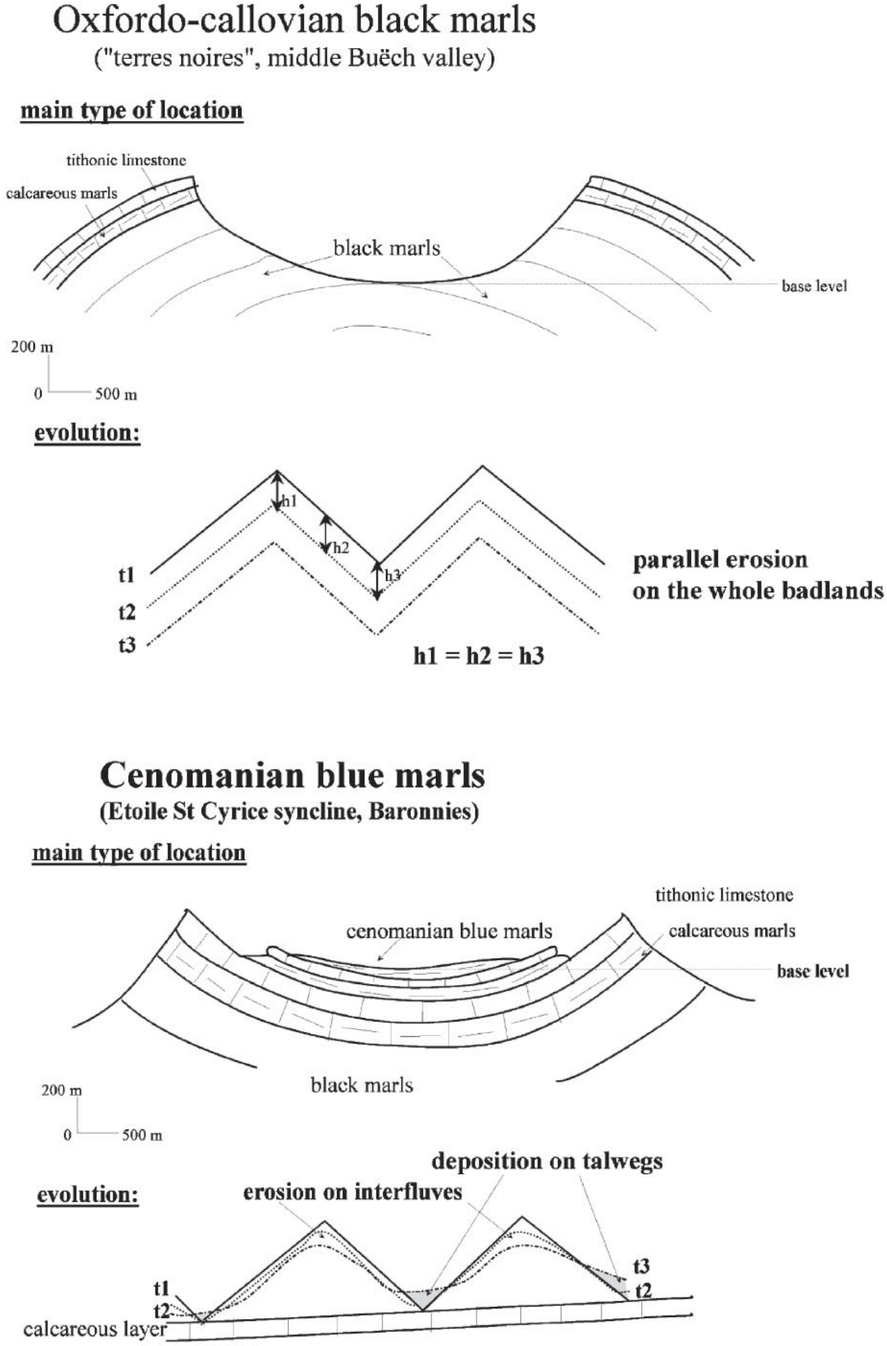

Figure 3. Schematic evolution of gullying in black marls (top) and blue marls (bottom)

A new type of light rugosimeter has been developed to measure profiles on steep slopes and in terrains with difficult access; its precision is about 1 mm, so it is well adapted to the substantial erosion of the marls and glacial moraines (Descroix, 1994).

A bottle-siphon was used to collect samples of suspended load in small creeks; it was adapted to collect samples with a sediment content up to 800 g l^{-1} (Olivry and Hoorelbeck, 1990).

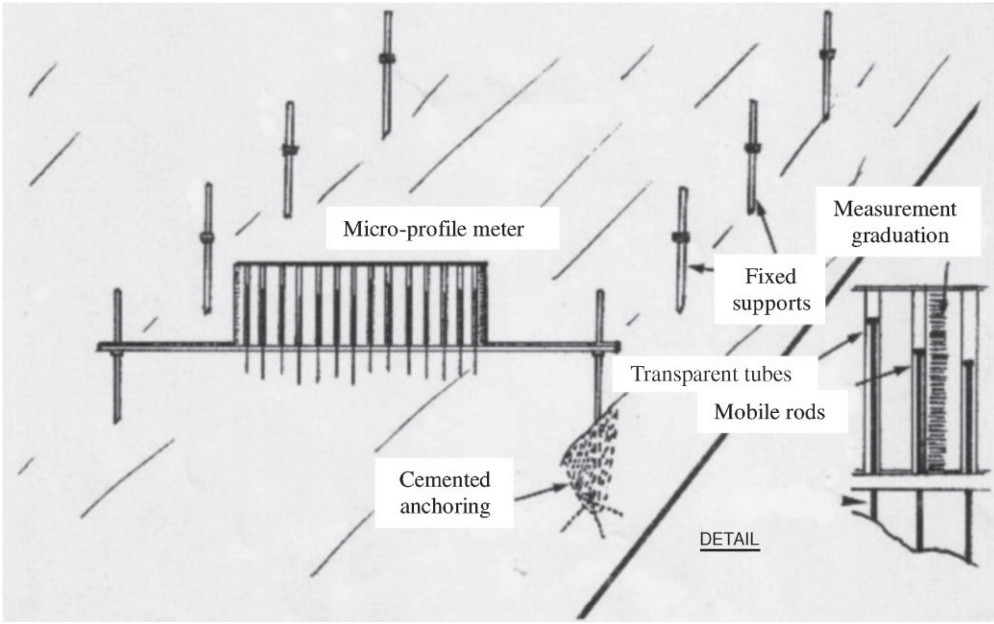

Figure 4. The complete microprofile meter *in situ* (Orstom-BRGM/LRG)

Soil losses were measured on black marls and their variables studied in the Méouge upper valley (Baronnies) (Lhénaff *et al*., 1993; Lecompte *et al*., 1996; Cohen, 1998). Needles fixed into the soil have also been used in the Méouge upper valley (Lecompte *et al*., 1996, 1998) and in the black marl in the Seignon catchment (Robert, 1997). The cryo-ejection problem was resolved by removing exposed devices.

Oostwoud Wijdenes and Ergenzinger (1998) used a large rainfall simulator to observe erosion processes on steep slopes. They determined and studied the important role of miniature debris flows (MDFs) in the Draix experimental catchments.

In 1997, studies on soil losses in black marls in the Seignon catchment were made (Robert, 1997; Rovera *et al*., 1999a). They developed new types of microprofilometer, which provide great precision in estimating soil loss.

Chemical erosion was analysed by Bufalo (1989) who emphasized the role of salts (haloclastism) in marl disaggregation. More recently, Simonnet *et al*. (1995) showed that the dissolution of calcite, pyrite and organic matter are the main factors of weathering in these outcrops.

A complete synthesis of research on erosion and solifluction in the Provence–Alpes–Côte-d'Azur region has been published (Collectif, 2000), summarizing the history of natural hazards and the numerous studies and experiments carried out in the Mediterranean part of the French Alps.

Plot-scale and basin-scale measurements

Erosion and transportation rates can be measured over a given area, from the plot scale to the catchment scale. Sedimentation in dams is also used to estimate erosion at the catchment scale.

Plots. Due to the general homogeneity of marl, plots could be used for measuring erosion. From plots used in black marls, Descroix (1985, 1994) defined climatic factors of erosion: efficient rainfall (rainfall necessary to trigger runoff or erosion), maximal intensities, kinetic energy, index R of Wischmeier, etc. However, collecting devices may be clogged and sometimes destroyed by extreme events and data could be lost.

Microcatchments. A global value of soil loss on a broader scale can be obtained at the microcatchment level, although this has the drawback of producing a mean value without separating the influence of each variable. However, plots and microcatchments could be used in a complementary fashion. Various research teams have made use of these field scales:

Figure 5. Draix: Laval (front) and Roubine (background) measurement stations (CEMAGREF)

- the LRG team (Descroix, 1985, 1994) on 20 to 20 000 m^2 areas;
- the Orstom-BRGM on 200 to 80 000 m^2 catchments (Deshons, 1985; Bufalo, 1989; Olivry and Hoorelbeck, 1990);
- Laboratoire de Géographie Physique of Paris VI University on 20 to 100 m^2 watersheds (Chodzko *et al*., 1991);
- CEMAGREF (Brochot and Meunier, 1995; Crosaz and Dinger, 1997) on microcatchments from 100 to 2000 m^2 (Figure 5);
- the Laboratoire de la Montagne Alpine (Grenoble University) in areas from 100 to 300 m^2 (Robert, 1997; Rovera *et al*., 1999a).

These experimental devices make it possible to model erosion processes, as shown by Bufalo (1989) in the Baronnies and Borges (1993) in Draix, both on similar areas.

Catchments. The same erosion measurement was used at the catchment scale. However, it is necessary to measure suspended load in the stream and bed load in the traps. This was made possible by using stacks of bottle-siphons (Olivry and Hoorelbeck, 1990) or electronic samplers (Olivier, 1995). Where larger areas were studied, the results were more significant. Sediment material was stocked at the bottom of the slope and in the thalwegs, but even if erosion depth measured on the event scale is not representative, the annual balance gave good results.

On the small Seignon watershed (Préalpes de Digne), Combes (1981) calculated soil loss in black marls as the volume of trapped sediments in a small dam divided by the eroded area of the catchment.

This methodology was also used at Draix (Mathys *et al*., 1996), at the Savournon experimental catchment (Bufalo, 1989; Olivry and Hoorelbeck, 1990), and in the Miocene molasses near Thoard (Alpes de Haute-Provence; Descroix, 1994). At the Draix experimental catchments, Richard and Mathys (1997) studied how to measure solid load and obtained interesting results. Finally, Brochot (1998) proposed an interesting synthesis on relations between gully erosion dynamics, transportation and sedimentation.

On black marls, Olivry and Hoorelbeck (1990) have shown that the distribution of particle size has a significant influence on deposition in traps or dams and that the mean size of particles decreased quickly downstream; therefore, this distribution changes depending on the basin area. According to Brochot and Meunier (1995), the percentage of deposit in the Roubine catchment (1500 m^2) is 85 per cent of the total transported load, and only 40 per cent in the Laval catchment (80 ha). This is due to the strong disaggregation of particles in the flows: the proportion of suspended load increases downstream, while the bed load rate is strongly reduced by transportation.

RESULTS ON WATER EROSION DATA

Table I shows that erosion values are similar for all kinds of marls: Oxfordo-Callovian black marls, Gargasian and Cenomanian blue marls. Almost all agree that the mean annual eroded depth ranges from 6 to 12 mm: these values have been obtained on different sites, on different configurations and with different devices. Most of the results presented below were obtained on marls. Only Descroix (1994) presented results for other lithologies.

- Bufalo (1989) measured soil losses on bare marls of 11·5 mm a^{-1} on small catchments over three years.
- Descroix (1994) and Descroix and Olivry (2002) presented values obtained on nine different sites and various lithologies. In marls (six different sites) and glacial fields, erosion rates ranged from 4 to 17 mm a^{-1} depending on exposure and slope–dip angle (see above); higher rates were measured on Oligocene clays (on only one site, the Bonneval mudflow): 30 mm a^{-1}. However, on Tertiary molasses, erosion depth was only 1·4 to 30 mm a^{-1}.
- On two microcatchments of 0·13 and 86 ha, Mathys *et al*. (1996) measured soil losses of 11 and 12·1 mm a^{-1}, respectively, over six years in the Draix experimental site. Measurements are still being taken, and mean values become more robust every year.
- Rovera *et al*. (1999b) measured an erosion rate of 3 to 10 mm a^{-1} on bare marls, depending on the year and the location. Using the same devices for another period, Robert (2000) measured a soil loss rate of 7 mm a^{-1} in the Draix catchments.
- Only in the Méouge valley were significantly higher values found; Lecompte *et al*. (1998) recorded annual means close to 3 cm with no extreme rainfall events. This should be due to a specific configuration or a local facies, if cryo-ejection has been taken into account. Otherwise, these results, confirmed by Cohen (1998), may be influenced by an artefact or the measurement devices.

Except for these latter data, all results in marls are similar, although they were obtained on different sites and for different areas, at different periods.

Before the previously cited authors, Combes (1981) calculated an erosion rate of 6·8 mm a^{-1} for the bare marls of the Seignon catchment. He considered only the degraded areas of catchments (40 per cent of the total area). This allowed him to estimate the mean annual volume of soil loss starting from the entire volume of stocked sediment (180 000 m^3). The filling of the Seignon small dam, in the Sasse Valley (Alpes de Haute Provence) is a major example of an involuntary sediment trap. This dam was built in 1962 at the outlet of a 3·6 km^2 basin, and filled in 1979. As the mean value from 17 years in the basin, it can be considered as a robust value, even though it was obtained at only one site.

The main local factors explaining the variability of erosion rates are exposure, vegetation, gradient and dip–slope angle.

Exposure (see Table I). Erosion is significantly higher on northern slopes (two to four times as high as on black marls). Over the year, aspect contrasts lead to a clear distinction between dominant transportation processes. Whereas rainfall impact is the same at the yearly scale on all aspects, processes caused by freezing (cryocreeping and debris flows, for example) cause higher yields and transportation values on north-facing slopes than rainfall events themselves (Figure 6). Aspect does not significantly influence the ability of vegetation to colonize slopes where the topsoil has been completely eroded on any aspect: vegetation regrowth is hindered on north-facing slopes by solifluction phenomena and on south-facing ones by summer drought.

Vegetation. The measured erosion rate in bare marly outcrops is more than 1000 times as high as in a close grassland (7 mm a^{-1} instead of 0·004 mm a^{-1}), all other conditions being equal.

The *gradient* seems to have no influence: measured on 1 m^2 plots, the erosion rate was twice as high on a 30° slope as on a 25° and a 48° slope. This is consistent with results obtained in the Western Sierra Madre by Descroix *et al*. (2001): soil losses are twice as high on a 20° slope as on a 30° one and three times as high as on an 8° slope.

Dip–slope angle. Summarizing observations on both spatial variables ‘exposure’ and ‘dip–slope angle’, a difference in processes makes it difficult to interpret the results. In marly terrains, the conjunction of a north-facing and a dip–slope angle close to 90° seems to be the configuration that yields the most material (Figure 6). On

Table I. Non-exhaustive synthesis of measurements made on erodible fields in the French Southern Alps (from Descroix and Olivry, 2002, reproduced by permission of IAHS Press)

Site	Reference	Years	Method‡	Bed rock	Soil loss (mm a^{-1})§			Measured points (*in situ*)	Area (measurement at the outlet)
					Average	Adret	Ubac		
Savournon (05)	Descroix and Olivry (2002)	1985–91	Rugosimeter	Black marls	9	6·5	11·6	140	
Savournon (05)	Descroix and Olivry (2002)	1987–90	Trap + BS	Black marls	*8·2**				7·84 ha and 75 ha
Saint Genis (05)	Descroix and Olivry (2002)	1985–91	Rugosimeter	Black marls	7	2·5	11·7	160	
Saint Genis (05)	Descroix and Olivry (2002)	1987–90	Trap + BS	Black marls	*6·7**				2·36 ha
Orpierre (05)	Descroix (1994)	1983–93	Trap	Black marls	*5·9**				100 m^2
Orpierre (05)	Descroix (1994)	1989–92	Rugosimeter	Black marls		3·8		80	
Gallands (26)	Descroix (1994)	1990–91	Rugosimeter	Black marls	16·2**			80	
Gallands (26)	Descroix (1994)	1990–91	Plots	Black marls	18·9**				2 m^2
Etoile (05)	Descroix (1994)	1989–92	Rugosimeter	Blue marls		8·5/1***		80	
Etoile (05)	Descroix (1994)	1988–93	Trap	Blue marls		*8·7*			70 m^2
La Vière (26)	Descroix (1994)	1990–91	Rugosimeter	Blue marls		10·3		80	
Thoard (04)	Descroix (1994)	1988–90	Trap + BS	Molasses	*1·4**				2 ha
Claret (04)	Descroix (1994)	1991–93	Trap	Glacial fields	*16·5**				12·5 m^2
Bonneval (26)	Descroix (1994)	1990–91	BS	Oligocene clays	*30**				28 ha
Laval (04)	Mathys *et al*. (1996)	1986–90	Trap + sam	Black marls	*11**				86 ha
Roubine (04)	Mathys *et al*. (1996)	1985–90	Trap + sam	Black marls	*12·1**				0·13 ha
Results of other authors									
St Genis (04)	Bufalo (1989)	1985–88	Trap	Black marls	11·5				200 and 2000 m^2
Seignon (04)	Combes (1981)	1962–79	Trap	Black marls	6·8				160 ha
Séderon (26)	Lecompte *et al*. (1998)	1990–95	Tods	Black marls	30			37	
Eygalaye (26)	Lecompte *et al*. (1998)	1990–95	Tods	Blue marls	7			52	
Vers s/M. (26)	Lecompte *et al*. (1998)	1990–95	Tods	Grey marls	8			71	
Izon la B.(26)	Cohen (1998)	1990–95	Tods	Blue marls	33			20	
La Motte C (04)	Robert (1997)	1995–97	Trap + Tods	Black marls	5	5	5		165

† (05) = Hautes Alpes department; (26) = Drôme department; (04) = Alpes de Haute-Provence department

‡ Trap = sediment trap; BS = bottles-siphons; Sam = electronic samplers; Tods = measurement sticks

§ Adret = south-facing slope; ubec = north-facing slope. * A ratio of 1/1·3 has been considered to calculate volume and eroded depths; density of regoliths has been fixed at 1·3. Values indicated in italic have been obtained in traps or plots and converted in runoff depth according to density = 1·3. ** Measurements affected by a decennial event: 100 mm rainfall in 2·5 hours, in July 1990. *** Influenced by a small landslide.

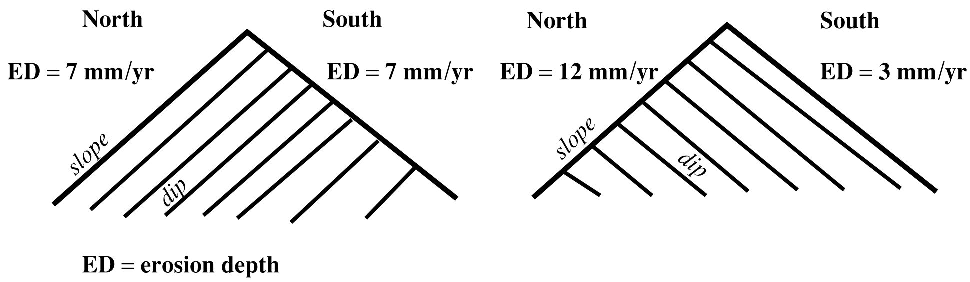

Figure 6. Impact of slope opposition on erosion depth (ED)

Table II. Some examples of soil losses values calculated according to sedimentation rate in dams (in Descroix and Olivry, 2002, reproduced by permission of IAHS Press)

Dam	River	Lithology	Watershed area (km^2)	Calculated soil losses ($mm\ a^{-1}$)	Measurement years (number)
Northern Alps					
Chambon*	Romanche	marly limestone, schist	220	0·12	50
Aussois*	Aussois	schists, marls	150	0·05	20
Sautet*	Drac	marls, moraines, granite	1000	0·37	30
Verney*	Eau d'Olle	metamorphic rocks, granite	120	0·27	(est.)
Southern Alps					
Serre Ponçon*	Durance	marls, moraines, metamorphic rocks	3000	0·5	30
Escale†	Durance	marls, marly limestone	3500	0·23	15
Cadarache†	Durance	marls, limestone, molasses	5500	0·05	15
Claps‡	Drôme	marls, marly limestone	182	1·4	350

* Sources: Vivian and Thomas (1982) and Descroix (1994)
† For dams situated downstream of some others, only the intermediary watershed has been taken into account
‡ Natural dam created by Luc en Diois Mountains collapse in 1442

blue marls and on south-facing slopes, a slightly higher soil loss value was obtained in the case where slope and dip are perpendicular, compared to the parallel configuration. Finally, it was always on north-facing slopes, and always in the perpendicular dip–slope configurations, that higher values of soil losses were obtained (Descroix and Olivry, 2002).

At a regional scale, global estimations of sediment supply were obtained from observations of sedimentation in dams at the outlet of larger catchments. Though dams are not built to become sediment traps, sedimentation data in this environment can be used to estimate soil loss in catchments (Vivian and Thomas, 1982). Table II summarizes erosion rates calculated from sedimentation data related to the entire watershed; therefore, the mean values are not specifically representative of the badlands.

Furthermore, the values obtained in this way do not account for sediments cleared from the dam; they are lower than those encountered in other Mediterranean regions. In northern Algeria, for example, annual eroded depths are commonly several millimetres for large basins of hundreds or thousands of square kilometres. Several dams were completely filled in a few decades (Benchetrit, 1972).

RESULTS ON OBSERVATIONS AND MEASUREMENTS OF SOLIFLUCTION PROCESSES

It is considered here that solifluction processes can lead to phenomena such as landslides, mudflows and debris flows. A considerable amount of research carried out in the Alps deals with solifluction processes at different scales.

The small-plot scale (1 cm to 1 m)

Devices dedicated to measuring the creeping of regoliths have been installed in different kinds of marls (Descroix, 1994). Creeping represents a type of soil loss that is not taken into account by the microprofile meter, so it is necessary to complete depth erosion measurements with an estimation of creeping values. Calculated sliding speed values ranged between 2 and 6 cm a^{-1} (3 cm in average); the maximum value was obtained on a site with parallel slope and dip. This represents an additional erosion depth close to 0·3 mm a^{-1}.

In the Draix experimental catchments, Oostwoud Wijdenes and Ergenzinger (1998) studied the behaviour of miniature debris flows, which are an intermediary process between hyperconcentrated flows and landslides. They are an efficient erosion and transportation agent: these sampled flows had solid concentrations ranging from 550 to 1440 g l^{-1}, leading to an erosive power much higher than that of clear water, because of its high specific density. These authors clearly demonstrated that there is 'an annual pattern, with net erosion along concentrated flow lines in the spring and summer months, and production of weathered marls associated with spatial dispersion in the winters months'.

The medium-plot scale (10 m)

Observation of dense networks of fissures and the presence of inflated areas in the field led to the development of an experimental site in the eastern Baronnies to measure solifluction *in situ*. Twenty-five pairs of stakes were installed on a slope on both compressional and tensional areas and their spacing was measured every month for eight years. Definitive results have not yet been published, but measurements showed that speed values of field displacement are related to temperature and rainfall events (Descroix, 1994). It seems that most of the sliding, with values ranging from a few millimetres to 5 cm a^{-1}, is produced at the beginning of the spring, when overland flow can infiltrate frost cracks, and at the end of the summer, in the sun cracks. These cracks favour a strong increase in soil water content, accelerating the phenomenon.

The scale of a small hillslope (100 m)

Alpine geologists have studied all of the solifluction processes, in both the Northern and Southern Alps (Antoine and Fabre, 1980). All the major alpine landslides have been observed and most of them are well known. Sliding speed has been related to winter snowfall and rainfall. This low-intensity precipitation and snowmelt led to high levels of water infiltration (Van Genuchten and Van Asch, 1988; Van Asch and Van Genuchten, 1990): near la Mure, they measured a 0·4 to 2 m annual soil surface displacement on varved clays. However, the most important observation site on black marls was the Ubaye Valley near Barcelonnette (Alpes de Haute Provence).

With regards to debris flows, according to Van Steijn (1989), 'controlling factors of hillslope stability in respect to mud flows and debris flows are slope value, presence of fragile rocks and presence of overpressure in interstitial water'. Systematic research on erosion and solifluction landforms as well as on natural hazards has been carried out in the Ubaye Valley. The La Valette landslide, located on the right side of the Ubaye river (south facing), has been the centre of intensive research (Van Steijn and Van den Hof, 1983; Salomé and Beukenkamp, 1989). From the 12th century, deforestation and overgrazing have led to gully erosion and hillslope instability near Barcelonnette. Blijenberg *et al*. (1996) developed a mini-rainfall simulator to work on steeper slopes; this device makes it possible to determine the threshold for the occurrence of microscale mass movements. From their field results they concluded that the main factors explaining debris flows were the slope value and rainfall intensity. The threshold conditions defined for the occurrence of mass movement are a function of these two factors: a minimum slope angle of about 34–36° at very high rainfall intensities, up to 300 mm h^{-1}, and a minimum of about 60–70 mm h^{-1} at slope angles of 55° or more, are required.

In the same valley, a landslide occurred in 1960, fossilizing a gully network. This La Roubine complex landslide–debris flow, on a north-facing slope, has been observed and measured since 1995 (Figure 7). The aim of these studies is to determine the risks (debris flows, hyperconcentrated flows) for the downstream areas created by the presence of this landslide. It appeared that the highest soil water content and therefore the strongest displacement speed occurred during the snowmelt period: Malet *et al*. (2000) described and measured the evolution of the slide showing that the debris flow continues its progression rapidly downstream after the first collapse (more than 180 m from 1982 to 2000). The authors noted that a similar mass movement occurred

Figure 7. Barcelonnette: measurement station on the la Roubine landslide (CEREG, Strasbourg)

Figure 8. Collapse and mudflow in Bonneval en Diois viewed from downstream

in the Draix experimental catchment in the spring of 1999. Furthermore, the impact of surface features on runoff yield appeared clearly in the La Roubine landslide–mudflow event (Malet *et al*., 2000, 2003).

In the Diois mountains (Drôme), the Bonneval-en-Diois debris flow (Figure 8) was studied in 1990–1992 (Descroix, 1994). This debris flow and the Boulc landslide, very close to Bonneval (10 km downstream), were the subjects of a historical synthesis and a risk analysis (Leone, 1996) and of a recent study based on dendrochronology and geomorphological datings (Astrade *et al*., 1998). These methods led the authors to give precise dating for the last 50 years. The meteorological context analysis showed that all of the 17 active phases of these two mass movements except one occurred after a major precipitation event (including four snowfalls) during a thaw period in winter or spring.

CONCLUSION

Many authors have focused their research on the processes, factors and measurements of current erosion in the French Southern Alps. They have considered the detachment of particles and their transport at the plot scale as well as at the microcatchment and catchment scales. Almost all of the French alpine massifs were covered by these studies, with a special emphasis on the marly badlands. Two classes of processes are contrasted – gullying and solifluction – although they may at times be related. The numerous data sets made it possible to determine the main climatic and site factors of water erosion; the Draix field laboratory (CEMAGREF, 1995) plays an increasingly important role in these observations, which have led to improved hydrological and erosion modelling.

These research topics on erosion are also studied in other Mediterranean countries such as Italy and Spain (Torri and Rodolfi, 2000) and in the countries of North Africa. They also focused on factors and processes, with special regard to the respective role of climate, vegetation, lithology and the anthropogenic impact. Brochot (1998) showed an opposition between:

- upstream basins, where the tendency towards torrential behaviour is always strong but is not a major problem, unless mountains are overpopulated and overexploited;
- downstream areas, where sedimentation processes occur on densely exploited zones of plains or intra-alpine valleys.

All the measures and descriptions should be improved for accurate modelling of processes, catchment behaviour and sediment load. This modelling of erosion and runoff could lead to a better prediction of the natural hazards in the mountains as well as in the downstream plains.

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