DISCUSSION ON "LARGE LANDSLIDES ASSOCIATED WITH A DIAPIRIC FOLD IN CANELLES RESERVOIR (SPANISH PYRINEES): DETAILED GEOLOGICAL-GEOMORPHOLOGICAL MAPPING, TRENCHING AND ELECTRICAL RESISTIVITY IMAGING" BY GUTIÉRREZ ET AL. (2015)

Núria M. Pinyol	Researcher. Centre Internacional de Mètodes Numèrics en Enginyeria. Department of Geotechnical Engineering and Geosciences. Universitat Politècnica de Catalunya, Barcelona, Spain.
Eduardo E. Alonso	Professor of Geotechnical Engineering. Department of Geotechnical Engineering and Geosciences. Universitat Politècnica de Catalunya, Barcelona, Spain.
Jordi Corominas	Professor of Engineering Geology. Department of Geotechnical Engineering and Geosciences. Universitat Politècnica de Catalunya, Barcelona, Spain
Jose Moya	Associate Professor of Engineering Geology. Department of Geotechnical Engineering and Geosciences. Universitat Politècnica de Catalunya, Barcelona, Spain

Corresponding author: E. E. Alonso
Department of Geotechnical Engineering and Geosciences.
Edificio D-2. Campus Nord. UPC. 08034 Barcelona
Phone: 34 93 401 6866; 34 93 401 7256
Fax: 34 93 401 7251
e-mail: **eduardo.alonso@upc.edu**

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N.M. Pinyol, E.E. Alonso, J. Corominas and J. Moya

1. Introduction

Gutiérrez et al. (2015) describe a geomorphological study of the left margin of the Canelles reservoir in the Spanish Pyrenees. The study reveals the presence of several landslides based on local geology, stratigraphy, trenching techniques, electrical resistivity imaging, geomorphological mapping and geophysical surveys. No borehole data were used in the study.

An important part of Gutiérrez et al. (2015) focuses on the evaluation of a previously published paper by Pinyol et al., 2012). The paper identified and analysed one of the landslides described in Gutiérrez et al. (2015) in detail, following the discovery of a continuous crack, subparallel to the reservoir water line, approximately 1 km long. The aims of Pinyol et al. (2012) paper were clearly noted in the paper:

- Identification and description of the unstable mass with the purpose of determining its geometry, volume, position of the sliding surface and materials involved.
- Evaluation of the risk of potential acceleration of the landslide
- Establishing the relationship between reservoir operation and landslide stability for the future management of the reservoir.

- Proposal of corrective measures to ensure that the slope remains stable during the management of the reservoir in the future.

The criticisms presented by Gutiérrez et al. (2015) to Pinyol et al. (2012) refer to three aspects: 1) the geological model described and the methodology used for its identification - Gutiérrez et al. (2015) state that the geological model shows a significant mismatch with respect to the actual sedimentary sequence -; 2) the conclusions from the analysis to evaluate the risk of rapid sliding; and 3) the effectiveness of the corrective measures proposed.

In the following sections these points will be discussed. A section of additional comments is also presented regarding some statements by Gutiérrez et al. (2015) that are, in our opinion, misleading.

2. Geological model of Landslide L4

After describing the geological setting and stratigraphy of the study area, Gutiérrez et al. (2015, p. 228) state that Pinyol et al. (2012) present a litho-stratigraphy different from the Garumn facies, because the borehole data used were from the Canelles landslide. This is not correct. . The litho-stratigraphy was based on the analyses of the sedimentary rock units outcropping in the area as well as the borehole logs (Pinyol et al. 2012, p. 33). The lithotypes shown by Pinyol et al. (2012) were simplified for publication purposes. No significant discrepancy exists between the stratigraphic logs presented by Gutiérrez et al. (2015) and the litho-stratigraphic units recognized by Pinyol et al. (2012) (Table 1).

Table 1.Comparison of stratigraphic logs of Canelles landslide.

Pinyol et al. (2012)		Gutiérrez et al. (2015)	
Lithotypes	Age	Lithotypes	Age
(a) Lower grey limestones. Grayey limestones of lacustrine origin interbedded with grey marls	Campanian to Maastrichtian	Montclús Fm. Grey, micritic and fetid lacustrine limestones.	Late Cretaceous, Maastrichtian
(b) Grey sandstones: predominantly medium to coarse grained sandstones of gray and ochre colour, interbedded with thin layers of multicoloured (grey, red, ochre) clayey siltstones, sandy siltstones and conglomerates of age.	Maastrichtian	Lower detrital unit (G1). Red mudstones and abundant intercalations of ochre cross-bedded and massive sandstones (…). The sandstone packages are up to several meters thick.	Maastrichtian
(c) Clayey limestones: thin layer of grey and white limestones and marly limestones that appear only locally (boreholes)	Maastrichtian		
(d) Red claystones. Clayey siltstones and silty claystones reddish and ochre coloured of continental origin. Locally interbedded with thin limestone and marly layers 1-2 m thick. Lower facies).	Paleocene age (Garumnian)		
(e) White limestones: massive grey to white limestone layer having either micritic or brecciated facies, the	Paleocene age (Garumnian facies).	Intermediate calcareous unit (G2). White and light-grey micritic limestone with charophytes and marl	Paleocene, Danian

latter with silty or clayey multicoloured matrix (reddish, grey, ochre, brown)		intercalations in the lower part	
(f) Siltstones and limestones. Heterogeneous unit composed of clayey silts and siltstones, silty clays, and multicoloured calcareous marls, which are predominant and layers of calcarenites, micritic limestones and brecciated limestones.	Lower Paleocene age (Garumnian facies)	Upper detrital unit (G3). Red mudstones with tabular beds of micritic charophyte bearing limestone, more abundant in the upper part of the unit.	Paleocene, Selandian-Thanetian

The main difference shown in Table 1 is that Gutiérrez et al (2015) consider G1 as a single unit, while Pinyol et al. (2012) split it into three units (b,c,d). Unit b, which is composed predominantly of sandstones, and unit d, mostly claystones, match respectively to the lower and upper parts of the Unit G1 of the Blancafort section (see Fig. 2 of Gutiérrez et al. 2015). Unit c is a thin layer of limestones which, as mentioned by Pinyol et al. (2012), appears locally in some boreholes and, also, in an outcrop located 400m to the west of the landslide boundary. This unit was unnoticed by Gutiérrez et al. (2015) probably because they worked with the regional stratigraphy rather than the local stratigraphy of the Canelles landslide provided by the borehole logs. On the other hand, splitting G1 into three lithotypes was fundamental for preparing the geological model of the Canelles landslide (landslide L4 of Gutiérrez et al., 2015). The reason is that the working hypothesis of Pinyol et al. (2012, p. 37) was that the slip surface should develop parallel to the strata, along a weak layer. Pinyol et al. (2012) considered unit d as a potential layer where the slip surface developed.

Gutiérrez et al. (2015), in their Introduction and Discussion sections, implicitly suggest that the geological model of the Canelles landslide provided by Pinyol et al.

(2012) was wrong and that this could have affected the results of the paper. However, Gutiérrez et al. (2015) did not provide any evidence of the supposed mismatch and, more importantly, the consequences of such mismatch. Rather, Gutiérrez et al. (2015, p. 238) indicate the slip surface develops through G1 (more precisely unit c) as Pinyol et al. (2012) concluded and as it has been corroborated with recent inclinometric measurements obtained after Pinyol et al. (2012) publication.

3. Landslide analysis and the risk of rapid sliding

Gutiérrez et al. (2015, p.232) refer to the description of the L4 landslide by Pinyol et al. (2012). The slide is described as a planar landslide with an average dip of the slip surface of 9–10º. This is a simplification of the landslide geometry, which Pinyol et al. (2012) never mentioned. Despite a description of the landslide as a reactivation of a dormant translational slide, Pinyol et al. (2012, Fig. 29) described the geometry as a double interacting block. Gutiérrez et al. (2015) did not realize that the landslide is actually a compound landslide (Hutchinson 1988; Hungr et al. 2014).

The simplification of a planar landslide by Gutiérrez et al. (2015) may lead to a stability analysis different from that by Pinyol et al. (2012). The specific geometry of the landslide determined by the topography and the failure surface geometry is a relevant factor to understand landslide mechanics. Pinyol et al. (2012) first identified the geometry of the landslide and they selected a representative section for a hydro-mechanical coupled analysis, which allows the estimation of the risk of rapid sliding. Despite the necessary simplification of the actual geometry of the selected cross section, Pinyol et al. (2012), maintained a fundamental aspect, namely that the failure surface was defined by two interacting masses which describe a compound slide (Fell et al., 2007). The moving mass is described as an upper wedge dipping 18º and a lower wedge sliding on a horizontal plane. The upper part acts as an active wedge which is inherently unstable because the inclination of the failure surface is higher than the residual friction angle assigned to the failure plane (10–12º) on the basis of tests performed. No cohesion is expected in the failure surface of a reactivated slide. The lower wedge and the interaction between both wedges provide the necessary strength to maintain the slope stable.

The major criticism of Gutiérrez et al. (2015) to the analysis by Pinyol et al. (2012) concerns the dynamic analysis of the post-failure response of the Canelles landslide. It is important to highlight that Gutiérrez et al. (2015) do not discuss Pinyol et al.'s (2012) analysis of the causes leading to the landslide in the summer of 2006.

The criticism from Gutiérrez et al. (2015) includes two arguments:

1) The landslide was never catastrophically reactivated in the past, although it reactivated several times.
2) No catastrophic reactivations have been documented in other large translational rockslides if the sliding surface has an average dip as low as 10º, which corresponds to their simplistic interpretation of the geometry of the L4 landslide.

These two points are discussed in the following. Note that in the present discussion we will refer to catastrophic landslides with a dominant sliding mode of deformation. Flow-like motions require a different consideration of the mechanical and hydraulic process involved in the run out.

First, why the catastrophic failure has not yet occurred? Gutiérrez et al. (2015) conclude that the behavior of the landslide during its life time is completely different from the model predictions. They argue that the slide has been affected by historical earthquakes and drawdowns of the reservoir level without any catastrophic failure (Gutiérrez et al., 2015, p.240). The implicit assumptions of Gutiérrez et al. (2015) is that the slope conditions have remained constant over time. However, no evidence indicates that the slope had been subjected to similar hydrologic conditions (full saturation and rapid drawdown) in the past. Despite the authors' statement, the conditions following the drawdown event of 1991 cannot be used as an analog as it is discussed in more detail later. In addition, Gutiérrez et al. (2015) have not provided any rigorous analysis showing that the stability conditions of the slope under the 1373 earthquake. As Gutiérrez et al. (2015,p.228) mention, halokinesis may have caused the progressive steepening of the slopes, as well as the increase of dip of the bedding that control the development of translational slides. They also state that halokinesis is currently an active processes. Therefore, the present-day and past conditions have to be evaluated with care.

In addition, the fact that something has not happened in the past does not mean that it will not happen in the future. Several documented rapid landslides were described as reactivated ancient landslides. This is the case of the Grijalva landslide in Mexico (Alcántara-Ayala and Domínguez-Morales, 2008),the Qianjiangping landslides, China (Wang et al., 2004; Dai et al., 2004), Val Pola landslide (Govi et al., 2002), and Sale mountain landslide (Zhang et al. 2002). They slid along a pre-existing shearing surfaces associated with older geological events. Therefore, they had been mobilized prior to the documented catastrophic landslide. Another well-known case is the Vaiont landslide, Italy, which failed in 1963 (Hendron and Patton, 1985). In this case persistent displacements were registered for more than 3 years previous to the catastrophic event. More than 3 m of displacement were accumulated and velocities during this "creeping" period previous to failure ranged between a few mm.day^{-1} to around 3 cm.day^{-1}. This case indicates that a relatively slow motion may suddenly evolve to fast landsliding with a velocity of 30 ms^{-1}. This well documented case, as well as the cases mentioned before, highlight that no evidence of previous catastrophic motion does not imply that it cannot happen in the future.

Therefore, the main point under discussion is not whether the slide has been mobilized or not prior to a catastrophic event. In order to identify the relevant factors causing the acceleration of a landslide, the discussion and any relevant analysis should be rooted on well-established mechanical and physical knowledge, and it should be validated by experience. Unfortunately, Gutiérrez et al. (2015) do not present any argument against the hypothesis and methodology used in Pinyol et al. (2012).

The model presented in Pinyol et al. (2012) to determine the risk of acceleration of the landslide, taking into account the drop of effective shearing strength due to thermally induced pore water pressures in the sliding surface, is not able to explain the non-accelerated motion of the slide and its subsequent stabilisation observed in the field in 2006. This situation is plainly stated in the conclusion presented by Pinyol et al. (2012). The acceleration of landslides is a complex topic under active discussion within the scientific community. However, it seems to be generally accepted that the main reason for slide acceleration is a significant loss of rock strength along the main sliding surface and other internal shearing bands.

The phenomenon invoked by Pinyol et al. (2012) has been widely acknowledged by the geotechnical and seismic scientific communities (Voight and Faust, 1982; Hendron and Patton; 1985; Vardoulakis, 2002; Rice, 2006; Veveakis et al., 2007; Goren and Aharonov, 2009; Pinyol and Alonso, 2010a,b; Cecinato et al., 2011; Cecinato and Zervos, 2012).This is, in particular, well exemplified by the continuing effort to explain the sudden acceleration of the Vaiont landslide. As far as we know, the question “why the Vaiont landslide did not accelerate when it was destabilized during the early filling of the reservoir” has not been answered yet. Recently, Alonso et al. (2015) tried to explain the interaction between a slow creeping motion and the possibility of a sudden acceleration to a very high velocity in a few seconds. They reviewed the mechanisms leading to strength reduction along the failure surface with special emphasis on the thermo-mechanical analysis. The transition from creep-like motion to a rapid event is analysed by combining strain-rate effects on friction and a thermo-poro-mechanical analysis of the shearing band and its vicinity. A sensitivity analysis, expressed in dimensionless fundamental parameters, provides considerable insight into the evolution of sliding velocity and its eventual blow-up. The blow-up takes place when thermal pressurization dominates the slide motion. It may occur for a combination of different factors mainly related to specific properties of the shear band, in particular friction rate effects and permeability. However, the blow-up depends also on the current straining rate of the landslide, affected by the unbalanced forces that are not constant during the creeping motion.

The second argument of Gutiérrez et al. (2015) to invalidate the dynamic analysis presented by Pinyol et al. (2012) is inappropriate. Gutiérrez et al. (2015) assert, based on cases collected from the literature, that rapid slides are typically associated with sliding planes dipping at least 20º. The fact that a sample of collected rapid slides presents a common feature is not a demonstration that a rapid failure may occur in other circumstances. Obviously the dip of the sliding surface affects landslide acceleration but equally significant is the available shear strength of the failure surface, directly controlled by pore water pressures. Consider the case of a planar landslide (which is not the case of the L4 landslide) of depth *D* and inclination β. The acceleration, *a*, can be easily calculated by solving the dynamic equilibrium equation:

$$a = \frac{1}{\rho D \cos\beta}\left(\rho g D \cos\beta \sin\beta - T\right) \qquad (1)$$

where ρ is the soil density, *g* the gravity acceleration and *T* the resisting force acting on the sliding surface. If the strength is reduced to values almost nil (as obtained in the thermo-hydro-mechanical coupled analysis by Pinyol et al., 2012) the block may reach an acceleration of 1.7 ms^{-2} for β=10º which leads to a velocity of 13 ms^{-1} in 100 m of displacement, which is an extremely rapid landslide (IUGS, 1995).This simple example indicates that a low inclination of the sliding plane does not limit the potential fast acceleration of a landslide.

Additionally, understanding the difference between a single block movement and a compound landslide is fundamental for the prediction of landslide kinematics. Otherwise, the potential for reactivation may be underestimated. Hutchinson (1988) already highlighted this mechanism of catastrophic failure. Fell et al. (2007) discussed the geometry of the failure surface of a compound landslide and, in particular the dips of the active and passive wedges. Glastonbury and Fell (2010) reviewed 51 cases of large rapid rockslides, 16 of which were compound landslides. They found that the inclination of the basal rupture surface (passive wedge) could be as low as 5º. They also found that the inclination of the basal rupture surface is typically smaller than the estimated basic friction angle by 5° to 10º and that the inclination of the rear rupture surface (active wedge) typically exceeds the estimated basic friction angle by 10° to 20º, suggesting that large out-of-balance forces are applied to the passive wedge. These conditions are fulfilled in the case of the Canelles landslide.

Gutiérrez et al. (2015) mention some limitations of the model proposed by Pinyol et al. (2012) that may justify the mismatch between the observed kinematics and the predicted response of the run-out. One limitation mentioned is that the reservoir water is not included in the model. It is worth noting that, although the water body is not modelled, the forces acting on the toe of the slope due to the weight of the water at the time of motion initiation are included. The effect of including the water of the reservoir in a more realistic way has not been evaluated.

The other limitation indicated by Gutiérrez et al. (2015) is that the permeability and stiffness of the shear zone are most probably high and low enough, respectively, to allow the dissipation of the potential excess pore fluid pressure related to water dilation by frictional heating. They note that Pinyol et al. (2012) document a fractured rock mass, several meters thick, associated with the slip surface. It is not true that we associate the slip surface with the fractured rock. The fractured rock refers to the mass above the sliding surface which has been mobilized during the previous sliding events. The assumption presented by Pinyol et al. (2012) is that the shear strains are localized into the Garumn clay layer. The value of permeability (k) selected for this high plasticity claystone was obtained from laboratory tests using an oedometer cell by means of a stationary flow method under a stress of 300 kPa. Two undisturbed specimens provided $k=4.2\times10^{-10}$ and 4.9×10^{-11} ms^{-1}. The value used in the calculation was 4.8×10^{-11}ms^{-1}. The stiffness was estimated introducing a Young's modulus equal to

500 MPa. However, the real support for the permeability adopted for the Garumnian clay layer is that, in the coupled hydro-mechanical analysis performed by Pinyol et al. (2012), the evolution of pore water pressure registered in the field piezometers installed at the end of 2007 were satisfactorily matched. In addition, the analysis presented, using these properties for the clayey layer, can explain the failure observed in 2006.

A higher value of permeability will allow the dissipation of pore water pressure in the shear band, which implies a reduction of sliding velocity. However, this higher value would also allow the dissipation of pore water pressure induced by reservoir level changes, and the failure observed in the 2006 summer could not be justified by the drawdown. It is therefore difficult to accept that the permeability value selected by Pinyol et al. (2012) was underestimated.

Gutierrez et al. (2015) also mentioned as a limitation of Pinyol et al.'s (2012) model that the opposite valley side and the submerged valley bottom are not included. Both effects will modify the results, particularly the velocity of the landslide when it reaches the opposite valley slope, but not the velocity when the reservoir water is hit by the moving mass, which is the key information for a tsunami analysis of the reservoir. Pinyol et al. (2012) selected 200 m as a maximum slide displacement because, according to the topography of the modelled representative section, the landslide does not reach the opposite valley side. In any case, the results given do not refer only to the velocity attained when the run-out is 200m. The calculated temporal evolution of the landslide motion shows that after 25 m of displacement, when the mentioned limitations of the model are probably less relevant, the sliding velocity is about 6 ms^{-1}, a very high and potentially destructive value.

Finally, the discussion on the triggering mechanism by Gutiérrez et al. (2015) is mostly speculative because no clear evidences are provided. They assume that the Canelles landslide was triggered by infiltration through karstic limestones (Congost de la Vall limestones), giving artesian conditions at the foot of the slope (p. 238). However, the piezometric measurements in the boreholes drilled in the landslide (Figs. 15 to 19 of Pinyol et al., 2012) apparently contradict this hypothesis.

4. Corrective measures

Gutiérrez et al. (2015, p.240, point 4) suggest that the sharp lateral facies change is the internal failure plane of a multiblock landslide, and that the stabilization fill identified by Pinyol et al. (2012) might lead to the reactivation of the lower block. As Gutiérrez et al. (2015) did not provide any cross-section or further details, we presume that they refer to the cross-section 2 (Fig. 29 of Pinyol et al., 2012) and that the multiblock landslide is the “Ls” located SE of Sant Salvador. We guess that Gutiérrez et al. (2015) have misplaced the location of “Ls” in the cross-section e. According to Fig. 5 of Gutiérrez et al. (2015), the “Ls” landslide is located just next to the location of borehole SI-2-1b, further North from the boreholes S-I2-1, S4-1 and S-1-3 (see Fig. 5 of Pinyol et al., 2012). Therefore, the proposed stabilization fill does not reach the “Ls” boundary. Gutiérrez et al. (2015) may have mistaken the “Ls” landslide with the one shown in our Figs. 5 and 11 of Pinyol et al. (2012), but this landslide does not extend to the west beyond the cross-section 2. In summary, the stabilization by removing weight from the upper part and loading the particular area selected in the lower part, as proposed by Pinyol et al. (2012) looks correct.

5. Additional comments

Gutiérrez et al. (2015, p. 233), in their conclusions on L4, neglected that the reactivation of the Canelles landslide and the presence of a secondary landslide were already mentioned by Pinyol et al. (2012) based on similar criteria. Furthermore, the conclusion that the lower unit of the Garumn facies is particularly prone to landslides (Gutiérrez et al., 2015, p. 238) is not new as shown by previous research (Corominas and Alonso, 1984; Corominas, 1989; Corominas and Baeza, 1992;Corominas et al., 2005).

In the description of the landslides L1 and L3, Gutiérrez et al. (2015) note that most landslide masses are located below the maximum normal water level of the reservoir, but no evidence of reactivation during the 2006 water level drawdown was detected unlike the L4 landslide. The size of landslide masses affected by the inundation of reservoir water is just one of the factors determining the slope instability due to a drawdown. Factors such as landslide geometry, particularly the shape of the impervious soil layer where the sliding surface is located, surface topography, and the drawdown velocity which is not constant and depends on the reservoir level, have a relevant effect (Pinyol et al., 2008). Without knowing them it is impossible to evaluate the potential reactivation of landslides such as L1, L3 and L4.

Regarding the L4 landslide and based on the examination of orthophotographs at several dates, Gutiérrez et al. (2015) state that there was already a rupture surface in January 2005, before the start of the drawdown, coinciding with the long crack detected in 2006. Gutiérrez et al. (2015). Without offering any stability analysis, they note that a potential trigger for the reactivation was the drawdown in 1991, which they describe as being similar to the 2006 drawdown. This is a major error. The reservoir water level since 1986 is plotted in Fig. 1. The drop of the reservoir level from 502 to 430 m a.s.l from 18/05/2004 to 21/08/2006 (27 months) is similar to the 69 m drawdownfrom17/02/1989 to 20/08/1991 (30 months). In terms of the total magnitude of the water level reduction and the average drawdown velocity, the 2006 drawdown was only slightly more critical. However, the main factor leading to a critical situation for the stability of the slope in 2006 was the long period of time (around 12 years) during which reservoir level was maintained at a relatively high level. This explains why the low permeability Garumn layer reached high pore water pressures, which is an important information to explain the instability described by Pinyol et al. (2012). On the contrary, the water level remained at a relative low level before the drawdown which started in 1989. At the beginning of the 1991 drawdown, pore water pressures in the clay layer were significantly lower than the pressures prevailing before the 2006 drawdown. As a result, the 2006 drawdown was more dangerous than the 1991 drawdown.

In conclusion, in order to discuss complicated landslide characteristics such as risk of failure, effects of water, run-out length, velocities and stabilization procedures, it is useful to combine geological inference, interpretation of geomorphological features with precise field data, and geomechanical analyses well rooted in physical phenomena. Lord Rutherford summarized it in a sentence: “All science is either physics or stamp collecting”.

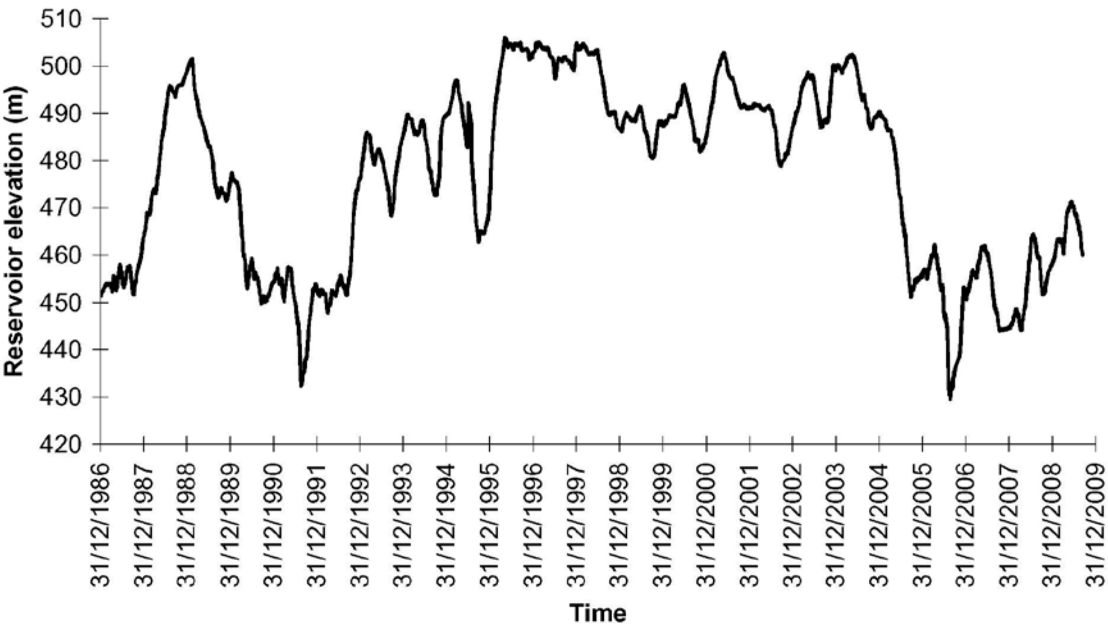

Fig.1. Water level evolution for the Canelles reservoir between 1986 and 2009.

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