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ORIGINAL PAPER

A. Demoulin · A. Pissart · C. Schroeder

On the origin of late Quaternary palaeolandslides in the Liège (E Belgium) area

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Abstract A series of palaeolandslides is described in eastern Belgium, with geophysical investigations, trench analysis and a number of ^{14}C dates complementing the field description. The proposed sliding mechanism stresses the importance of initial liquefaction of the upper Cretaceous Aachen sands underlying the ~30-m-thick Vaals Clay Formation, in which all landslides are developed. Slope-stability analyses support the hypothesis of sudden landsliding and yield further useful information to discuss their origin. A main difficulty in determining the landslide trigger arises from insufficient absolute dates. Assuming that all landslides occurred simultaneously, we weigh the probabilities, respectively, of a climatic and a seismic trigger. Although the 150 A.D. possible date of landslide initiation falls close to one of the wettest periods of the Holocene, the spatial distribution of the deep slides and their proposed mechanism strongly suggest a seismic origin in connection with a rupture of the nearby Ostend segment of the active Hockai fault zone. However, considering that the nearby 1692, Verviers earthquake apparently caused no ground failure, it is probable that both the climatic and seismic triggers were jointly needed to provoke such deep and extended landsliding on moderate, generally stable slopes.

Keywords Landslide · E Belgium · Palaeoseismicity · Slope stability · Geophysical investigation

A. Demoulin (✉) · A. Pissart
Dept of Physical Geography and Quaternary,
University of Liège,
Sart Tilman, B11, 4000 Liège, Belgium
e-mail: ademoulin@ulg.ac.be
Tel.: +32-4-3665660
Fax: +32-4-3665722

C. Schroeder
Department GeomaC/Lab of Applied Geology Hydrogeology and Geophysical Prospecting,
University of Liège,
Sart Tilman, B18, 4000 Liège, Belgium

Introduction

Beyond the influence of environmental factors favouring slope instability, the most common active triggers of landslides are climatic conditions, seismic shaking and coastal or fluvial undercutting (Crozier et al. 1995). In W Europe, human interference (through deforestation, slope alteration) has also to be taken into account from ~2000 B.P. While the evidence for slope undercutting is mostly direct (Hutchinson 1976), the distinction between climatically and seismically induced palaeolandslides is much more difficult. Both causes may indeed trigger every type of slope failure and even the temporal and spatial clustering of landslides may equally result from an earthquake or a rainstorm. In recent years, growing attention has been paid to the recognition of earthquake-triggered palaeolandslides in various settings (e.g. Perrin and Hancox 1992; Schuster et al. 1992; Jibson and Keefer 1993; Keefer 1994; Crozier et al. 1995; Jibson 1996; Ota et al. 1997), aided by many available descriptions of modern analogues (e.g. Hansen 1965; Pain 1975; Harp et al. 1981; Cotecchia 1987; Alkema et al. 1994; Tibaldi et al. 1995; Harp and Jibson 1996; Okunishi et al. 1999). Crozier (1992) proposed a list of criteria whose combined fulfilling should be indicative of the seismic origin of mass wasting. However, the evidence for earthquake-induced palaeolandslides often remains ambiguous, especially in regions of moderate or low seismicity where the seismotectonic framework itself is poorly documented. In this study, we present results strongly suggesting a seismic origin for clustered deep-seated palaeolandslides located to the east of Liège, in E Belgium.

Aim of the study

The main goal of our study was to produce a prediction map of the landslide hazard in the area affected by prehistoric slides [Demoulin and Chung: "Landslide hazard prediction in the Pays de Herve (E Belgium): a probabilistic susceptibility map." Geomorphology, submitted]. This obviously required not only determining the environmental conditions leading to failure, but also understanding the mechanism of landsliding. In this respect, we could not avoid addressing the central question of what triggered these landslides. The aim of this paper is therefore to identify the geological formations involved in landsliding and to get information about the internal structure, the depth and the age of the slides, in order to compare potential scenarios of landslide triggering and development. The possibility of seismic triggering in E Belgium, an area of moderate seismicity, is of considerable interest for regional seismic hazard assessment. However, as mentioned above, it is extremely difficult to unequivocally demonstrate the seismic origin of landslides (Jibson 1996), and this paper only lists a series of observations, none of which is definitely decisive, but which nevertheless allow the question of the seismic trigger to be raised.

Methods

The inventory of landslides has been carried out through the systematic analysis of black and white vertical aerial photographs (date: 1947; scale: 1/10,000). All forms interpreted as landslides on the photos have been verified in the field. Besides the general geomorphological analysis, three main trenches and two additional small excavations were dug in two major landslides (Fig. 1, Uh and Th). The trenches were ~20 m long across the toe scarps. Although they could not provide a comprehensive view of the landslide structure, they yielded several worthwhile indications of the toe architecture. In addition, they provided abundant charcoal and small pieces of wood, present in the soil buried under the toe and in humic layers developed higher in the landslide masses during stable phases, thus offering several opportunities for ^{14}C dating. A geophysical reconnaissance was conducted in the Th landslide (Fig. 1), located on the southern slope of the main E-W divide. A 2D electrical resistivity tomography complemented by electrical and seismic refraction profiles provided an image of the formations involved, of the internal structure of the slide and of the depth of its basal shear surface. The interpretation of these geophysical data was confronted with the available borehole data and the results of cone penetration tests carried out in another nearby landslide. Finally, in order to test various scenarios of slide development, stability analyses were performed on original slope profiles where the safety factor for several possible slip lines with characteristics similar to those of the actual basal shear surface of the landslides were calculated. In computing the safety factor associated with a number of complex slip lines, we used the Janbu method (Janbu et al. 1956).

Study area

The study area belongs to the Pays de Herve, a moderately dissected tableland to the N of the Ardenne massif in E Belgium. Its mean altitude is about 250 m, with a relief of 80–100 m and slope angles rarely exceeding 12°. The topography is strongly marked by the presence of a main E-W trending divide, separating the Berwinne and Vesdre drainage basins. Several secondary watersheds have developed in a NNW direction within the former catchment, and only one significant secondary S-

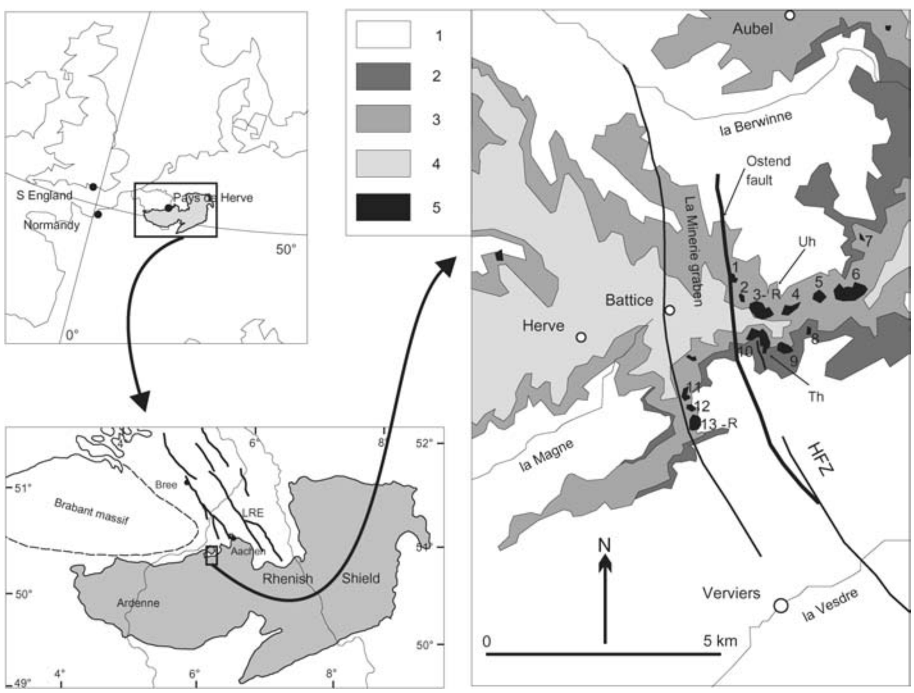


Fig. 1 Simplified geological map of the study area. *1* Shales and sandstones (Palaeozoic); *2* Aachen sands (Santonian); *3* Vaals clays and sands (Campanian); *4* Gulpen chalk (Campanian); *5* landslides. The landslide numbers refer to Table 1. *R* stands for landslides reactivated in 1998. *Uh* and *Th* locate the two landslides where trenches have been excavated and dating carried out. *HFZ* Hockai fault zone. In **bold**, the Ostend fault segment, on which the earthquake responsible for the palaeolandslides is suspected to have occurred. The Bree prehistoric earthquake(s) is located in the *lower left inset*

and then WSW-striking divide extends within the Vesdre basin to the south of Battice (Fig. 1).

The geology of the Pays de Herve is characterised by the extensive preservation of subhorizontal Cretaceous cover deposits resting on a folded Palaeozoic basement. Whilst the shales of the basement crop out in the valleys, the dissected Upper Cretaceous cover forms the bulk of the ridges (Fig. 1). At its base, the Aachen Formation is mainly comprised of fine sands interlayered with coarse sands and clays. Although the Battice area, where the landslides concentrate, corresponds to the westernmost extension of the formation, the Aachen sands may still be up to 10 m thick. The overlying Vaals Formation displays glauconiferous clays and marls (the so-called Herve smectites). Its maximum thickness is 25 m in the area affected by landslides. To the east, it gradually becomes sandier, with the occurrence of indurated layers. Resting on the Vaals clays, the chalks of the Gulpen Formation are almost absent from the Battice area. In contrast, they are well preserved on top of most other ridges of the Pays de Herve. In their upper metres, they generally are weathered to clay-with-flints and are then no longer geotechnically distinguished from the similarly weathered Maastricht chalks. They frequently display solution pockets up to several tens of metres in diameter and filled with remnants of Tertiary marine sands and Quaternary loess.

Geological, seismological and morphological data indicate that while the WSW-striking Variscan thrust faults had been inactive for a long time, several NNW-striking Variscan faults were reactivated as transtensional or pure normal faults during the Cenozoic opening of the Lower Rhine segment of the W European rift system. In the study area, the small graben of La Minerie, parallel to the main border faults of the Lower Rhine Embayment, is limited by active faults, namely the Ostend fault on its eastern side, which represents a segment of the seismogenic Hockai fault zone, with a suspected historical earthquake of M_w>6 close to the city of Verviers in 1692 (Camelbeeck et al. 1999) (Fig. 1). It should be noted that, from a detailed investigation of one slide of the area which had been reactivated due to highway construction at its toe, Graulich (1969) concluded that its initiation had to be related to the Quaternary seismic activity of the La Minerie graben.

With regard to hydrogeology, the main aquifer of the Pays de Herve lies within the Gulpen chalks, but numerous small aquifers also exist within the Aachen sands, trapped variously between the clays resulting from the Mesozoic weathering of the Palaeozoic basement, clay layers inside the Aachen Formation and the Herve smectites. Another aquifer is also present in the underlying fissured Palaeozoic shales. As for the climatic context, present-day yearly rainfall is 900 mm, without any significant spatial variation within the study area. The centennial 1 day rainfall is of 99 mm at the nearby station Thimister, with a 95% confidence interval between 76–123 mm (Dupriez and Demarée 1988).

Results

Landslide structure and age

Ten large (maximum size of a single slide is 22 ha) and a few smaller landslides, totalling 90 ha, are located in the Battice area, on both sides of the central part of the main E-W trending divide of the Pays de Herve and on a nearby secondary S-striking divide (Fig. 1). They have all developed within the Vaals clays, in places where the underlying Aachen sands are present. Indeed both materials, but especially the sands possess geotechnical characteristics highly favourable to instability. Several landslides also seem to have formed in places where the overlying Gulpen chalks have been more or less completely removed from the top of the ridge. Moreover, all the large landslides are located within the La Minerie graben or within a distance of 2.5 km of its eastern border fault.

Most of the landslides start in the upper part of moderate (4–10°) slopes, about 5–10 m below the top of the ridges (Table 1). They are generally extended compound (multiple rotational+translational) slides (Hutchinson 1988; Dikau et al. 1996), up to 700 m in width (parallel to contour lines) and 400 m in length (perpendicular to contour lines), and show steep arcuate

Table 1 Morphometric parameters of the palaeolandslides of the Pays de Herve

Landslide	Slope value	Headscarp height (m)	Width (m)	Length (m)	Area (ha)	Estimated volume (m^3)
1	9°	~5	175	190	3.4	0.1×10^6
2	3.5–5.5°	~2	125	150	1.9	$<0.1\times10^6$
3 (Uh)	9.5°	>10	500	300	13.3	2×10^6
4	5.5°	~3	450	125	5.1	0.15×10^6
5	5°	~5	250	300	5.6	0.2×10^6
6	4.5–7°	17	700	400	21.8	3.3×10^6
7	9.5°	<2	125	75	1	<<
8	6.5°	<2	200	100	1.7	<<
9	man-reshaped	?	300	250	5.6	?
10 (Th)	9°	>10	550	250	12	1.8×10^6
11	4.5°	~5	200	150	2.4	0.1×10^6
12	9°	~5	175	75	2.1	$<0.1\times10^6$
13	4°	11	375	250	6.8	0.4×10^6

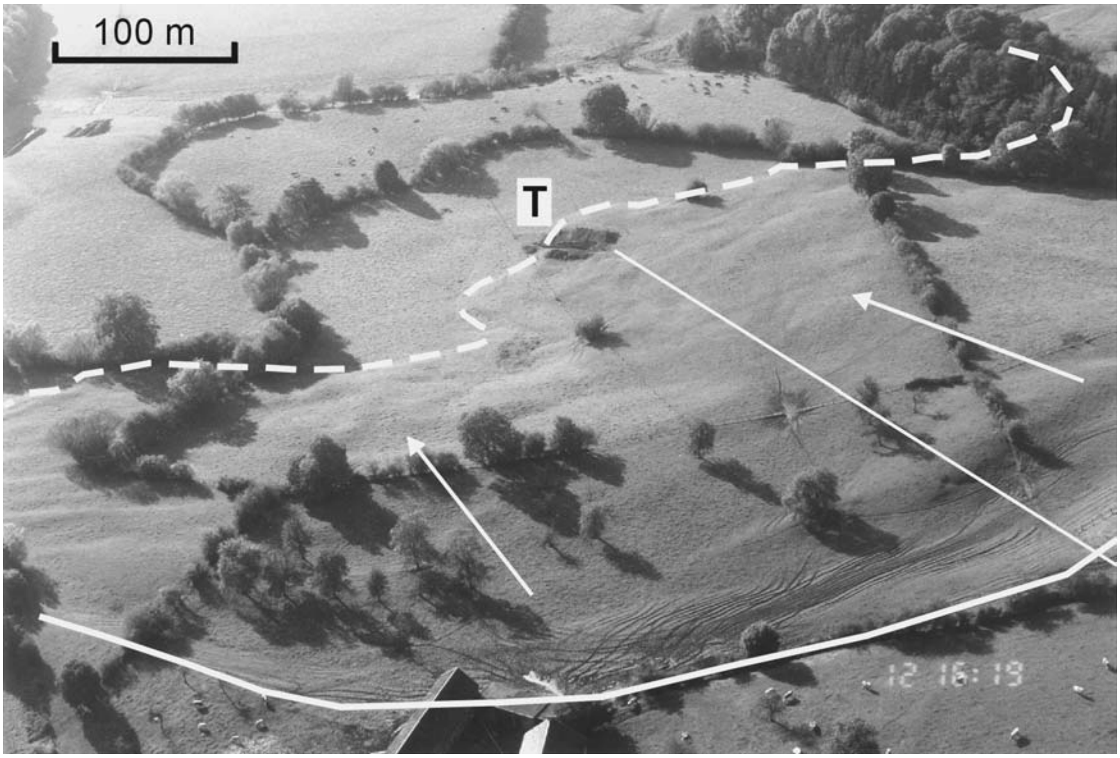


Fig. 2 Oblique aerial view of the "Th" landslide (see location on Fig. 1) showing the trench (*T*) cut into the landslide toe and the location of the electrical tomography profile (*thin white line*). The *bold line* marks the headscarp location and the *dashed line* the approximate extent of the landslide, while one rotated block is visible in the *middle right of the photo*

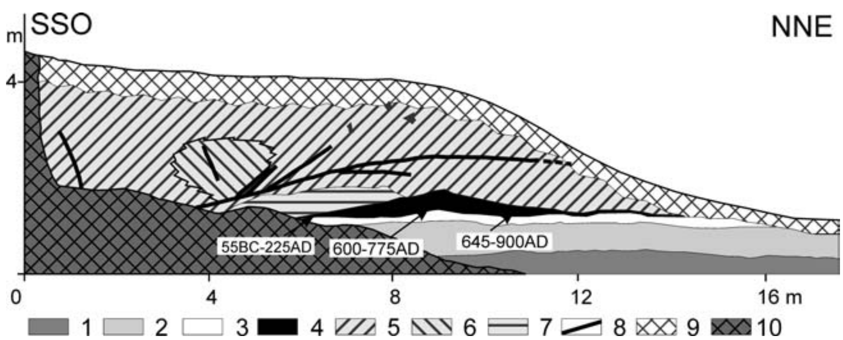


Fig. 3 Trench in the toe of the Uh landslide (see location on Fig. 1). *1* Paleozoic shales weathered to clay; *2* clay-with flint pre-landslide colluvium; *3*, *4* horizons of a podzolic soil developed in a silt of mainly eolian origin—note bulging of the *4* humic horizon in the centre of the section, with location of the dated samples; *5*, *6*, *7* slipped masses, chiefly made of reworked Vaals clays with sparse sandstone blocks, *8* major shear surfaces; *9* present-day soil; and *10* waste, hiding the base of the section

head scarps, up to 17 m high (Fig. 2). The upslope part of the slipped masses generally displays parallel, ~2-m-high ridges and intervening counterslopes determined by the upper edge of rotated blocks whose average length is of ~40 m. A downslope zone of hummocky topography often terminates in a sharp 2–3 m high frontal scarp. In some cases more than 200 m long, this zone is clearly longer than it might be expected of the accumulation zone at the foot of a purely rotational slide, and is evidence of mainly translational sliding, thus illustrating the compound nature of these features.

In the Uh site (Fig. 1), the dug distal part of the landslide toe is located a little bit downslope of the base of the subhorizontal Aachen sands, and in the Th site, the sands are very close beneath the bottom of the trench. In all trenches, the base surface of the slipped mass clearly displays a terminal irregularity resulting from limited scraping and subsequent downslope bulging of the underlying soil (Fig. 3). This podzolic soil is generally developed within pre-landslide silty-clayey colluvium resting on Palaeozoic shales weathered to clay. The soil bulge slightly stands back with respect to the frontal scarp of the landslide (by 6–7 m). Most striking within the slide

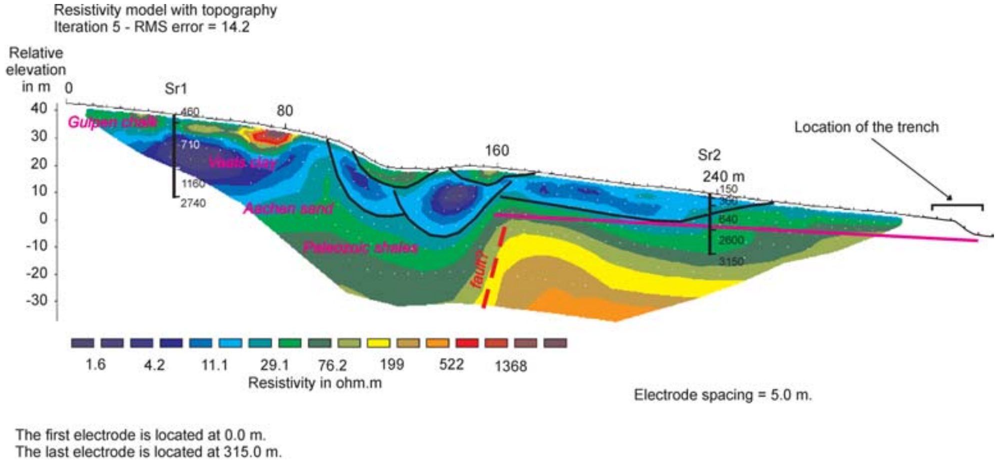


Fig. 4 Electrical tomography carried out in the Sérezé (*Th*) landslide, and interpretation of the subsurface structures. The line in magenta corresponds approximately to the Aachen–Vaals contact as it can be extrapolated upslope from its position in the trench. *SR1* and *SR2* are seismic refraction soundings with seismic velocity given in m/s

material is the systematic presence of major undulating shear planes greased by a striated clayey gouge and running across the whole length of the sections. In the rare cases where the displaced masses delimited by shear planes are not made of colluvium, they still preserve their original structure, one trench exposing, for instance, tilted blocks of intact cross-bedded Aachen sands interlayered with clays.

The geophysical reconnaissance carried out in the Th landslide identified three successive layers of varying resistivity and seismic velocity (Fig. 4). Their interpretation is based on the lithostratigraphy known from numerous boreholes drilled in the ridge area affected by the landslides (Graulich 1969). One recognizes first a 5–10-m-thick upper layer of intermediate resistivity corresponding to the more or less weathered chalk of the Gulpen Formation, then, the ~20-m-thick clays of the intermediate Vaals Formation characterised by a very low resistivity and, finally, a deeper layer of increasing resistivity including the Aachen Formation and the shales and sandstones of the top of the Palaeozoic bedrock. The tomography clearly evidences the listric shape of the basal shear and the presence of rotated blocks, mainly made of Vaals clays, in the upper part of the landslide. These blocks are about 25 m thick for an average downslope length of 40 m. This confirms the depth of at least 20 m, which could also be inferred for the basal slip surface of the slide from its size and morphological characteristics. Moreover, additional smaller blocks rest on top of the main tilted blocks, possibly bearing witness to reactivation episodes. In the tomography, the downslope half of the slipped mass appears as a 10-m-thick, more than 100-m-long slope-parallel body, slightly thinning at its far end.

The calibrated ^{14}C ages obtained for the Th and Uh landslides provide an insight into the chronology of the movements (Fig. 5). They point to a first episode of motion common to both landslides at 150±80 A.D. Later phases of displacement at ~400, 700–750 and 1250 A.D. are characteristic of only one or the other slide. Unfortunately, there was no opportunity to date material buried

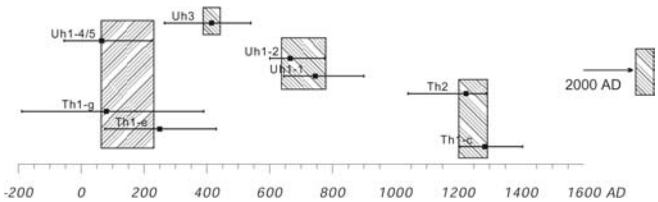


Fig. 5 Temporal distribution of landslide movement determined by the calibrated ^{14}C ages obtained in the Sérezé (*Th*) and Croix Polinard (*Uh*) landslides and recent movement. The 95% confidence intervals are indicated by *horizontal bars*. *Right hatching* denotes assumed time of landslide initiation. *Left hatching* indicates reactivation phases. The 2000 A.D. date does not come from ^{14}C dating, but is simply the time of present-day reactivation of one landslide

under the upslope part of the slipped mass, and despite the location of the samples dated back to 150 A.D. (Fig. 5), it cannot be proven that this is actually the time of the mass movement initiation.

Determining factors and landslide mechanism

Besides the nature of the stress/strength modifications leading to slope failure, the main question arising from the compound character of the landslide structure is that of the spatial (progressive or retrogressive) and temporal (slow or rapid) ways of its development.

The compound character is evidenced by the overall length of the mass displaced downslope of the rotated blocks and by the structures revealed by electrical tomography. Moreover, whereas neither the slide morphology nor its toe structure evokes flow, the presence of numerous shear planes and the preservation of original stratification within the toe's material and, to a lesser extent, the scraping and downslope bulging of the buried soil strongly suggest that the motion in the toe area primarily occurred as translational gliding involving overthrusting of coherent blocks.

Field evidence and the geophysical survey also show that whether the head scarp is developed mostly within the Vaals clays or sometimes in the Gulpen chalks, close to their limit of outcrop, the lower part of the landslide's surface of rupture is always located approximately at the level of the Aachen sands. Furthermore, the clays clearly remain unaffected by landsliding in areas where the underlying sands are absent. We therefore suggest that the liquefaction of these highly liquefiable fine sands interlayered with silts and clays could have been the decisive factor in initiating the large slides. Although there was no direct evidence of liquefied sands in the exposed trenches, there exist a number of observations showing the quicksand behaviour of the Aachen sands (Graulich 1969). The absence of any homogenised mass of flowed sand in the trenches would then suggest that the initially liquefied sand body was certainly of limited thickness and extent, and only acted as the slide starter. It possibly corresponded to a single sand layer sandwiched between silt/clay intercalations within the Aachen Formation, and spread over a restricted area much upslope of the trenches opened in the toe of the landslide. Whatever the cause of liquefaction of some sand layers (see below), it removed any support of the overlying mainly Vaals material, especially in the lower, more gentle part of the slope where the Aachen sands crop out, thus setting off the first translational motions of coherent blocks of Vaals clay (and occasionally even of unliquefied Aachen sands themselves) in this area. A further indication that this was indeed the first stage in landsliding is given by one slide that did not develop further, leaving a small graben at the head of the translational glide. In the other cases of larger slide development however, the latter motion in turn brought upslope blocks in disequilibrium, inducing successive rotational slides in the higher and generally steeper part of the slope and the final carving out of a major steep head scarp close to the top of the ridge. The whole process might therefore be seen as retrogressive and most probably catastrophic, the bulk of the morphological features of which the landslides are comprised appearing more or less simultaneously.

However, the alternative scenario of a slow, progressive translational movement along an unstable downward part of the slope leading, after decades or centuries, to sudden failure of the upward slope portion and to the development of rotational slides also has to be considered since it would have considerable influence on the meaning of our ^{14}C dates and on potential triggers.

The stability analyses shed some light on this question. Considering that the whole process is initiated by translation in the lower half of the slope where the Aachen sands crop out, we first calculated that the safety factor largely exceeds 2.0 for a realistic slope-parallel (i.e. with a maximum dip of 4°) slip line at a depth of 10 m, whether it is located within the clays or the sands, and this even in the worst circumstances of a water-table close to the ground surface (Table 2). Within the limits of our assumptions on unit weight ρ, effective cohesion c' and effective internal friction angle φ' (based on sampling in several nearby areas and laboratory testing of the geological units involved in the landslides), any slip line would need to dip more than 9° to become unstable, which is clearly unrealistic here. The slopes of the Battice ridge are therefore stable in static conditions with respect to this type of motion.

Table 2 Slope stability analysis with respect to translational gliding. Calculated relationships between safety factor F, slope angle β and friction angle φ' in case of a 10-m-deep translational gliding within either saturated Vaals clay or Aachen sand (water table at the ground-surface level)

	ρ (kg/m^3)	c$'$ (kPa)	φ' (°)	β (°)	F
Vaals clays	1,500	9.81	15	4	**2.24**
	1,500	9.81	15	**9**	1
Aachen sands	1,500	0	30	4	**2.75**
	1,500	0	26	4	**2.32**
	1,500	0	30	**10.9**	1
	1500	0	26	**9.2**	1
	1500	0	**11.8**	4	1

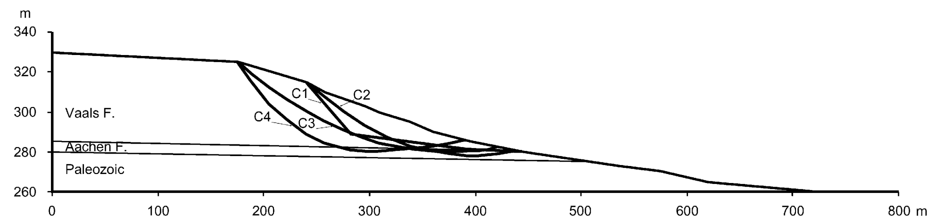
In the case that all slipped masses (including translated and rotated blocks) would be involved from the beginning in the movement, we also calculated the safety factor for more complex slip lines. Assuming that the geotechnical characteristics of the sands are crucial to the initiation of landsliding, we assigned ρ, c$'$ and φ' values to the clays and carried out a parametric analysis of the sands, which again showed that in static conditions the slopes in the Battice area are intrinsically stable (Table 3). However, especially for the steeper C4 slip line (Fig. 6), near-surface water-table conditions lower the safety factor to values hardly exceeding 1.0, and even <1.0 in the improbable eventuality of sands with a friction angle <26°. In any case, these results suggest that an additional external factor is necessary to destabilise the slopes, but that only a small action would suffice to get them sliding. Should this factor be related to exceptional rainfall, it must then be noted that our computations evidenced that heavy precipitations lasting for weeks or months and inducing the ascent of the water-table to the ground surface cannot alone explain an extended slope destabilisation. On the contrary, an extreme event of short-term intense rainfall can induce a head of water below the top of the ridge sufficient to cause a rapid increase in pore pressure within the underlying fissured Palaeozoic shales. Transmission of this additional pore pressure increase from beneath into the sands in the lower part of the slope may then provoke a sudden drop of their shear strength below the calculated value and initiate slope failure. Summarising, if it is climatically induced, the stability analysis makes a "catastrophic" (i.e. at the minute-to-day scale) process with an unique downslope displacement more probable than slow sliding, thus strongly supporting the first scenario proposed above.

However, such a mechanism can obviously explain only the first initiation of mass movement on undisturbed slopes. Later phases of landslide reactivation are evidenced by ^{14}C dates and by localised slide features superposed on the main slipped masses. The position of the soil bulge with respect to the present-day frontal scarp

Table 3 Parametric analysis of slope stability with respect to rotational sliding. Parametric analysis yielding safety factor F in function of Aachen sand cohesion c′ and friction angle φ' for different slip lines (see Fig. 6) and water-table conditions after having empirically fixed the c′ and φ' values of the overlying Vaals clay respectively at 9.81 kPa and 15°

Slip line	Water table at a depth of 2 m						Water table at a depth of 0.2 m						
	c′↓ φ' →	20	22	24	26	28	30	20	22	24	26	28	30
C1	0.0	1.39	1.4	1.41	1.43	1.44	1.46	1.18	1.19	1.2	1.21	1.22	1.22
	0.2	1.41	1.42	1.43	1.45	1.46	1.48	1.2	1.21	1.22	1.23	1.23	1.24
	0.4	1.43	1.44	1.45	1.47	1.48	1.5	1.22	1.23	1.24	1.24	1.25	1.26
	0.6	1.45	1.46	1.47	1.49	1.5	1.52	1.24	1.25	1.26	1.26	1.27	1.28
	0.8	1.46	1.48	1.49	1.51	1.52	1.54	1.26	1.27	1.28	1.28	1.29	1.3
C2	0.0	1.38	1.45	1.51	1.58	1.66	1.73	1.15	1.2	1.25	1.31	1.36	1.42
	0.2	1.42	1.49	1.55	1.62	1.7	1.77	1.19	1.24	1.29	1.34	1.4	1.46
	0.4	1.46	1.52	1.59	1.66	1.73	1.81	1.22	1.28	1.33	1.38	1.44	1.5
	0.6	1.5	1.56	1.63	1.7	1.77	1.85	1.26	1.32	1.37	1.42	1.48	1.54
	0.8	1.54	1.6	1.67	1.74	1.81	1.89	1.3	1.35	1.41	1.46	1.52	1.58
C3	0.0	1.15	1.18	1.22	1.25	1.29	1.33	*0.99*	1.02	1.05	1.08	1.11	1.14
	0.2	1.16	1.2	1.23	1.27	1.31	1.34	1.01	1.04	1.07	1.1	1.13	1.16
	0.4	1.18	1.21	1.25	1.28	1.32	1.36	1.03	1.05	1.08	1.11	1.14	1.18
	0.6	1.19	1.23	1.26	1.3	1.34	1.38	1.04	1.07	1.1	1.13	1.16	1.19
	0.8	1.21	1.24	1.28	1.32	1.35	1.39	1.06	1.09	1.12	1.15	1.18	1.21
C4	0.0	*0.93*	*0.98*	1.04	1.09	1.15	1.21	*0.85*	*0.9*	*0.95*	1	1.06	1.11
	0.2	*0.94*	1	1.05	1.11	1.16	1.22	*0.86*	*0.91*	*0.96*	1.01	1.07	1.12
	0.4	*0.95*	1.01	1.06	1.12	1.17	1.23	*0.88*	*0.92*	*0.97*	1.03	1.08	1.13
	0.6	*0.97*	1.02	1.07	1.13	1.18	1.24	*0.89*	*0.93*	*0.98*	1.04	1.09	1.14
	0.8	*0.98*	1.03	1.08	1.14	1.2	1.26	*0.9*	*0.95*	1	1.05	1.1	1.16

Fig. 6 Profile of original slope and different slip lines used in the slope-stability analysis



also indicates that after the initial main displacement the material still moved a few metres downslope, but these reactivations are either shallower or located at the distal end of the accumulation zone, where they are of much more limited extent. They are totally disconnected from the Aachen Formation and primarily result from the reduced shear strength along shear planes developed within the slipped mass.

The main determining factor of landsliding in the Battice area thus seems to be the combined presence of the Aachen sands and the overlying Vaals marls and clays, both with highly unfavourable geotechnical characteristics. But other possible factors also deserve attention. Namely, it must be explained why the landslides seem to concentrate in the areas more or less devoid of overlying Gulpen chalks. In this respect, it should first be stressed that, indeed, considering the respective percentages of displaced slopes within the ridge parts covered by the chalks and those uncovered parts, the latter are more affected. Conversely, the number of landslides in each of these two zones is however more or less similar, making the assumed relationship much less striking. This is also confirmed by a probabilistic analysis of the landslide hazard in the area (Demoulin and Chung: "Landslide hazard prediction in the Pays de Herve (E Belgium): a probabilistic susceptibility map." Geomorphology, submitted), which yields a slightly degraded prediction rate when the proximity of the Gulpen chalks is included in the set of determining factors. Moreover, the possible indirect role of the chalk cover remains unclear. From the mechanical point of view, this cover, especially with the perched water-table it retains above the clays, represents an additional weight on top of the slope, which theoretically should favour landsliding. However, it also prevents swelling and weathering of the buried clays whose shear strength is certainly higher than that of exposed and weathered Vaals clays. This is, however, of very limited concern to the clays cropping out even in the lower slopes of the chalk-covered ridge parts (Fig. 7). Hydrogeologically, on the one hand, one might argue that unweathered, "closed" clays below the weight of overlying chalks could partly reduce the amount of percolating water made available to saturate the Aachen sands. Nevertheless, on the other hand, one has to account for the presence of a perched water-table in the chalks which allows for continuous seepage through the clays, and should have a reverse effect. It is therefore very difficult to estimate how much, if ever, the presence of a chalk cover is

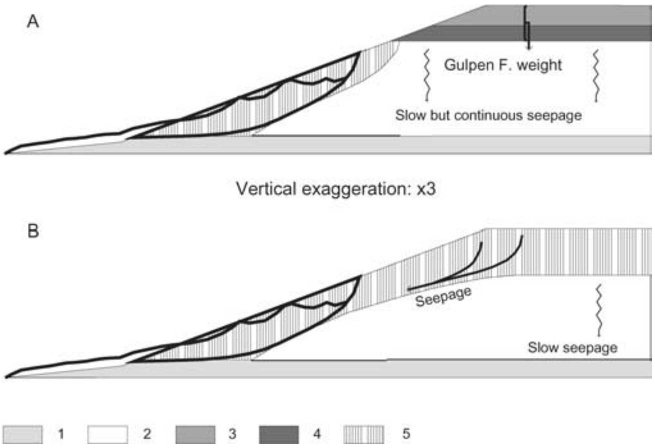


Fig. 7A, B Sketch comparison of the along-slope mechanical and hydrological conditions with or without a cap of Gulpen chalks. *1* Aachen sand; *2* Vaals clay; *3* Gulpen chalk; *4* perched water-table within the chalk; *5* superficial zone of swelling and weathering of the clay and subsequently increased permeability. The *thick lines* describe the slope disturbance caused by landsliding

determining for the landslide distribution in the Battice area.

Another determining factor of landsliding is generally the slope gradient itself. Here, the landslides occurred on slopes of varying but moderate steepness (between 4–10°), and the highest (10–12°) slopes of the area, cut in the chalks, were not affected, indicating that this parameter does not play any significant independent role. Slope aspect influence, with landslides preferentially occurring on N- to NW-facing slopes, also is only an indirect effect resulting from the gentle NNW-dip of the Cretaceous strata.

Triggering factors

The static stability analysis was performed, not only to evaluate the adequacy of the proposed slip mechanism, but also to test the necessity of invoking a seismic trigger for the existing set of landslides. Unfortunately, it left the second question unanswered, showing that, mechanically, the slides might have been produced by exceptionally intense short-term rainfall as well as by seismic shaking. The lack or insufficiency of other important information makes it still more difficult to definitely assign a cause to the landslides.

In this respect, dating is of critical importance. The few available ^{14}C dates, with a similar oldest age recorded in both studied sites, are nothing more than a faint suggestion that all large landslides in the Battice area could have occurred simultaneously. Further support to this assumption is nevertheless provided by the similar degree of scarp and general morphology degradation displayed by all slides. If we therefore hypothesise that all main forms appeared during the same event, the date of ~150 A.D. should be retained with great caution, as it cannot be demonstrated that it was the time of the mass movement initiation. However, some hints thereof could perhaps be found in the simultaneous occurrence of motion in both dated forms, and especially in the position of the 150-A.D.-aged charcoal on the stoss side of the underlying soil bulge, provided we accept that the latter might locate the end-point of the main initial displacement event. With all required reservations, the following discussion will thus be based on the hypothesis that the landslides of the Battice area record one single event dating back to ~150 AD.

Considering a first possible climatic trigger, an indication of the rainfall intensity required to set off extended landsliding in the study area is given by the amounts of precipitation which caused, or did not cause, some limited slide reactivation in recent years. During the autumn of 1998, on the night of September 13, more than centennial precipitation of 126 mm reactivated only two of the ancient landslides (R, Fig. 1). In one of them, it induced a downslope displacement of ~2 m along a pre-existing shear plane and the development of a 1–1.5-m-high, 70-m-long scarp. Continued rain in October 1998 (450 mm was recorded during the period of September and October 1998) provoked further decimetre-scale movements only in this same place. These quite isolated motions resulted from the combined influence of exceptional precipitation and local loading of the headscarp of the slide by waste. By contrast, such intense rainfall was not able to move any of the other landslides, where present-day human interference is limited, and still less to initiate new deep landslides in the Battice area. This suggests that the latter type of event is highly improbable under present-day climatic conditions, at least without the operation of an additional trigger.

It is indeed difficult to ascertain whether even the wetter conditions that prevailed over western Europe during several periods in the Holocene could alone have caused the development of extended deep landsliding in the Pays de Herve. Significantly moister climates characterise the periods 7000–6000 B.P., 2500–2000 B.P. and 1100–1300 A.D. (Starkel 1966; Lamb 1977; Delmonaco et al. 1999). However, in the Battice area, the 1100–1300 A.D. episode seems, for instance, not to have been able to induce more than limited reactivations of existing slides. On the contrary, it will be noted that the possible date of landslide initiation at 150 A.D. falls relatively close to the upper limit of the 2500–2000 B.P. episode, which has also been recognised as favourable to landslide development along the coast of and inland S England (Brunsden and Ibsen 1994a, 1994b). This comparison is still of limited value as the atmospheric circulation and the topographical conditions are clearly more landslide-prone along the cliffs of Normandy and S England than in the Pays de Herve. But the chief difficulty in assigning a climatic trigger to the Battice landslides lies in their spatial distribution. Why did they occur in certain zones and not in others within the area of combined outcropping of the Aachen sands and Vaals clays? In the absence of any slope influence, this distribution could be determined by facies variations within the Vaals Formation or by the presence or not of a chalk cover on top of the ridge. However, on one hand, the area where the clays predominate in the Vaals Formation extends amply beyond that affected by landsliding. On the other hand, we have already stated that the spatial correlation between landslide occurrences and the absence of overlying chalks is much looser than it seems at first glance, and does not lead anyway to a clear-cut cause-and-effect link.

In contrast, the landslide distribution is much more easily explained in the case of a seismic trigger. All landslides concentrate within an area of ~ 5 km diameter centred on the Ostend fault, which represents the eastern border of the La Minerie graben (Fig. 1). The Ostend fault is a NNW-striking, 7.5-km-long segment of the active Hockai fault zone. Not only is instrumental seismicity recorded on more southern segments of the fault zone (Camelbeeck 1993), but the epicentre of one of the most violent historical earthquakes having ever struck NW Europe (with an estimated M_w>6.0) has also been located in the Verviers area, to which the zone of landslide occurrence belongs (Camelbeeck et al. 1999). Moreover, fresh scarp and scarplet features which could betray Quaternary or even Holocene seismic activity along the Hockai fault zone between Verviers and Battice are currently under investigation. The invoked slide mechanism involving initial sand liquefaction is another hint of a seismic cause, as it appears to result more frequently from dynamic liquefaction (generally of seismic origin in subaerial environments) than from static liquefaction (Seed 1968). Considering the six criteria listed by Crozier (1992) to support the seismic origin of landslides in New Zealand, most of them are therefore satisfied here: (1) though weak, ongoing seismicity is recorded in the region, (2) the landslide distribution coincides with a segment of an active fault zone, (3) slope stability analyses suggest that either extreme climatic events or seismic shaking are required to induce slope failure, (4) the landslides are fairly large sized, (5) liquefaction is associated with landsliding, and (6) the landslide distribution cannot be definitely explained solely on the basis of geological or geomorphological conditions. The low angle of affected slopes and basal shear surfaces and the location of most landslide headscarps in the upper part of the slopes furthermore support a seismic origin of the Battice landslides (Keefer 1984; Densmore and Hovius 2000).

Discussion

The possible landslide-triggering earthquake would not in any case be an exceptional seismic event in the area. With respect to the empirical relationship established by Keefer (1984) between earthquake magnitude and maximum distance of landslide from fault-rupture zone, an earthquake along the Ostend fault segment should have had a minimum M_s of 5.0 to induce the Battice landslides at the observed maximum distance of 2.5 km (Fig. 8). According to the same author, while the observed landslide type, i.e. coherent soil block slides, has been described in relation with earthquakes as small as 4.5 in M_s magnitude, another relationship between magnitude and total area affected by landslides gives a minimum M_s of 4.7 with regard to the 20 km^2-large zone encompassing the Battice landslides. The latter zone is situated within the Hautes Fagnes (HF) seismic zone of Leynaud et al. (2000) for which these authors calculate a return period of 447 years for an M_s 5.0 earthquake. If we reasonably assume from morphological observations that the Ostend fault is one of the two to three most active among less than ten potential seismic sources within the HF seismic zone, the return period of such an earthquake on this particular fault

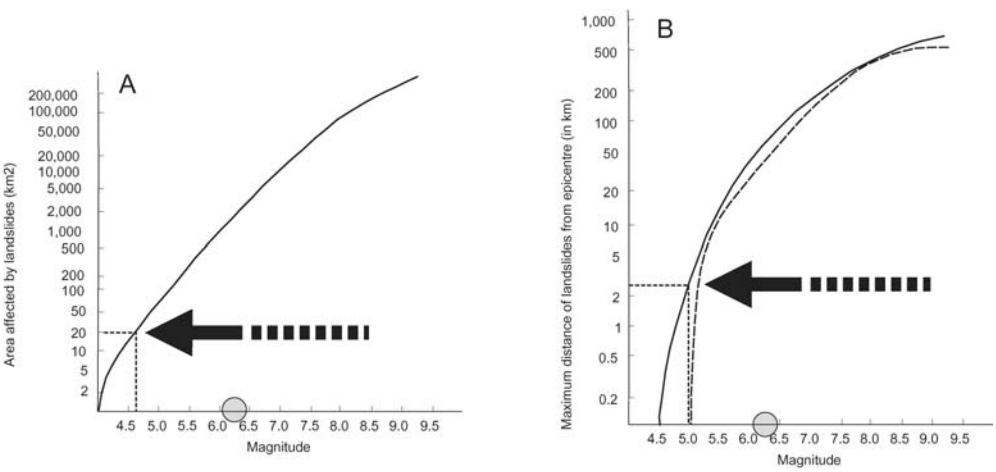


Fig. 8A, B Location of the 'Pays de Herve event' in the diagrams showing the relationships between magnitude of the triggering earthquake and the characteristics of the landslide distribution (after Keefer 1984) in case of a seismic origin of the landslides. In **B**, the *solid curve* refers to 'coherent slides' and the *dashed curve* to 'lateral spreads' (and liquefaction-induced mass movements). The *grey circles* locate the 1692, Verviers earthquake, which seems to have been unable to cause landsliding in the study area segment probably lies within the range of 2,000–3,000 years.

However, the possibility of an earthquake having ruptured the Ostend fault is not the only way of causing earthquake-triggered landslides in the Battice area. Camelbeeck and Meghraoui (1998) have identified a surface rupture event along the Bree segment of the Feldbiss fault, which bears witness to an M_s ~6.5 earthquake between 610–890 A.D. (Fig. 1, inset). They suggest that it could be related to the 782–836 A.D. period of seismic activity reported in the historical earthquake catalogue of Alexandre (1990), namely with a large earthquake mentioned in Aachen in 803 A.D. It is interesting to note that a phase of landslide reactivation also dates back to the same period, with a calibrated ^{14}C age between 600–900 A.D. at the 95% confidence level. Yet, Keefer's relationship indicates that an M_s ~6.5 magnitude corresponds to a maximum distance of landslide occurrence of 80 km, whereas the Bree fault segment is located less than 60 km away from Battice. This, added to the rough coincidence of dates, thus makes it quite possible that the remote Bree earthquake caused the reactivation of some of the Battice landslides. But, however tempting the extrapolation might be, a similar link is far less probable for the time of initiation of the landslides (at 150 A.D.?) since again, it would in no way be supported by their limited distribution.

Another peculiarity is that no dated landslide initiation or even reactivation coincides with the historical earthquake of 1692. Indeed, this tremor with an estimated $6.0<M_S<6.5$ and located within the region of Verviers should have induced important mass movements as secondary effects in the close landslide-prone area of Battice (Fig. 8). Of course, additional dates should first confirm that the earthquake caused no extended landsliding there. If this was the case, the alternative would then be either that the 1692 epicentre was located more than 40 km apart from—and to the south of—Battice, i.e. near Vielsalm or St Vith, which runs counter to the available historical data (Alexandre, 1997), or that maybe the soils were too dry and unsaturated when the 1692 earthquake occurred on September 18, at the end of the summer. The importance of the latter factor was again recently illustrated by the 1995 Kobe earthquake which, due to a preceding dry year, caused landsliding only at a maximum distance and within an area respectively ~10 and 1% of what could be expected from the Keefer relationships to magnitude (Okunishi et al., 1999). This also suggests that possibly the combined influence of a particularly wet season and an additional, for instance seismic, trigger was required to set off landsliding in the extent and depth as observed in the Pays de Herve.

Conclusion

In summary, although it cannot be proved without independent dating of other, preferentially primary evidence of the responsible earthquake, the seismic origin of the Battice landslides nevertheless appears to best fit the observational data set. However, we must recall that our interpretation rests on the assumption that all landslides occurred simultaneously. More dates will be needed in order to confirm this supposition, even though a similar degree of scarp degradation and ridge erosion is already another valuable indication (Jibson and Keefer 1993). But the main argument for a seismic trigger clearly lies in the landslide distribution, which cannot be understood solely on the basis of the topographical or lithological conditions. A good support to the seismic hypothesis is likewise provided by the reconstituted landslide mechanism involving the liquefaction of the Aachen sands. On the other hand, the rough coincidence of the 150 A.D. date with one well-documented period of wetter climate in western Europe is also certainly of significance, for such deep and extended failure of moderate, usually stable slopes most probably was set off by a combination of triggers (seismic event+period of heavy rainfall).

In conclusion, even if a seismic triggering may appear probable, no definite origin can as yet be definitively assigned to the Battice landslides. However, future palaeoseismological studies of the Hockai fault zone in the Verviers-Battice area should at least bear in mind the ~150 A.D. date.

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