Influence of surficial crusts on the development of spreads and flows in Eastern Canadian sensitive clays

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ABSTRACT
Spreads and flows are the two main types of large retrogressive landslides occurring in Eastern Canadian sensitive clays. In spreads, the soil mass mobilized during failure is dislocated in a succession of horsts and grabens leading to a typical ribbed topography in the landslide scar. The failure mode for flows, on the contrary, is characterized by a succession of rotational slides propagating rearward, which requires that clays liquefy during the movement. These mechanisms are now relatively well understood. However, conditions leading to the development of either a flow or a spread are not yet clearly identified. Some numerical results published in the literature suggest that spreads form preferentially when a non-sensitive crust overlying a sensitive clay deposit is present. We examine in this paper whether this result is supported by observations made on several spreads and flows that occurred in southeastern Ontario and Quebec. It is shown that the presence of a crust is likely not a discriminating factor.

RÉSUMÉ
Les étalements et les coulées sont les deux principaux types de grands glissements de terrain rétrogressifs se produisant dans les argiles sensibles de l'est du Canada. Dans le cas des étalements, la masse de sol mobilisée lors de la rupture est disloquée en une succession de horsts et de grabens, laissant une topographie nervurée typique à l'intérieur des cicatrices. Le mode de rupture pour les coulées est au contraire caractérisé par une succession de glissements rotationnels se propageant vers l'arrière, ce qui nécessite que l'argile se liquéfie lors du mouvement. Ces mécanismes sont maintenant assez bien compris. Cependant, les conditions conduisant au développement d'un étalement ou d'une coulée restent mal identifiées. Certains résultats de modélisations numériques publiés dans la littérature suggèrent que les étalements se forment surtout lorsqu'une croûte recouvrant les dépôts d'argiles sensibles est présente. Nous regardons dans cet article si ces résultats sont validés par l'observation de cas réels d'étalements et de coulées dans le sud-est de l'Ontario et au Québec. On montre que la présence d'une croûte n'est probablement pas un facteur discriminant.

1 INTRODUCTION

Large retrogressive landslides occurring in eastern Canadian sensitive clays can be grouped into two main types depending on the failure mode: spreads and flows (Fig. 1). In spreads, the soil mass mobilized during failure is dislocated in a rearward succession of horsts and grabens, also called prisms and wedges, leading to a characteristic ribbed topography in the landslide scar. The formation of horsts and grabens is the result of an extensional, active state of failure, and the overall movement along the basal failure surface is translational, as exemplified by the horizontal layering of strata often observed in intact horsts, indicating that no rotation occurred. The amount of soil remaining in a spread scar is variable but can be important, with only a localized remolding of clays. This failure mode was first identified by Odenstad (1951) and further analyzed by Carson (1977), Locat et al. (2011), Quinn et al. (2012), and Dey et al. (2015). The failure mode for flows, on the contrary, is characterized by a succession of rotational slides propagating rearward (Bjerrum, 1955; Tavenas et al., 1971; Gregersen, 1981; Tavenas, 1984; Demers et al. 2014). In each rotational failure, the displaced soil mass must reach a sufficiently fluid state to be able to be evacuated from the slope toe and to allow retrogression. Otherwise, the backscarp slope is buttressed by debris, which can stop the retrogressive movement. This implies that clays must have a low remolded shear strength to flow away from the slope toe. Typically, only a veneer of strongly remolded clays with patches of the upper crust of the soil profile are left in a flow scar. The floor topography is relatively smooth with no significant relief variations. The retrogression process for flows can be initiated by a first rotational slide along a slope resulting from various causes, such as a riverbank erosion, an excavation at the slope toe, an overloading at or near the slope crest, an increase in porewater pressures following rainfall or snowmelt, or by seismic shaking. For spreads, the current understanding is that any perturbation conducing to a horizontal unloading along a slope can trigger retrogression if appropriate conditions are met. These mechanisms may include as for flows an initial slide or seismic shaking. In some cases, field observations suggest that these two types of failures can occur successively during the same landslide event (Geertsema et al., 2006; Tremblay-Auger et al., 2018). Both spreads and flows can propagate rapidly on flat grounds or on very gentle slopes, over large distances often exceeding several hundreds of meters.

As mentioned by Demers et al. (2014), the geotechnical properties of soils involved in large retrogressive landslides are quite similar, and conditions leading to the development of either a flow or a spread are not yet well identified and understood. Several geometrical and mechanical factors can interact in a complex way, making it difficult to identify discriminating conditions. In this respect, numerical simulations are of interest because they allow parametric analyzes, which is obviously impossible to do with real landslides in the field (Dey et al., 2015; Locat et al. 2015; Wang et al., 2016a, 2016b; Wang and Hawlader, 2017; Zhang et al., 2018; Tran and Sołowski, 2019). In some of these simulations, it appears that spreads only develop when a crust having a significant thickness relative to the thickness of the mobilized soil mass overlies the sensitive clay unit in which the failure propagates. For example, Wang and Hawlader (2017) show that a 5-m thick crust with a shear strength of 50 kPa overlying sensitive clays with shear strengths linearly increasing with depth, from 40 to 100 kPa at a depth of 35 m, was needed to generate a spread. The initial slope was 30 m high with an angle of 26.6°. A simulation with no crust led to the development of a flow. However, not all the parameters of the analyses were exactly the same in the simulations, in particular for the post-peak properties of sensitive clays. The extent to which this affects the results of the simulations is not discussed in the paper, but the authors observe that the presence of a crust is a potential cause of the formation of horsts and grabens. In Dey et al. (2015), it is mentioned that a sufficiently high undrained shear strength is needed for a spread to occur but that this failure mode is prevented for a very thick crust representing, in the case analysed by the authors, 47% of the height above the failure surface. With all other parameters being equal, a very thin crust (5% of the height above the failure surface) only leads to the development of a horizontal shear band that initiates at the base of the slope. The slope height for the cases analysed was 19 m and the slope angle 30°. Similar results on the influence of a crust were obtained by Q.A. Tran (personal communication, Nov. 2018; Tran and Sołowski, 2019). In all these studies, the crust was modeled as a non-sensitive material in undrained conditions.

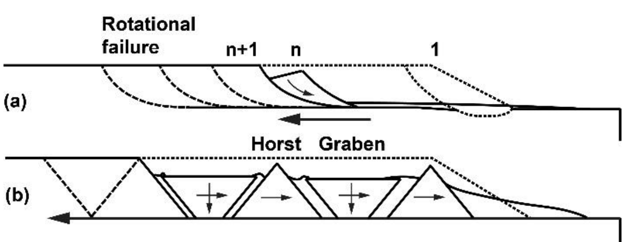

Figure 1. Conceptual sketch of the mode of failure (a) for a flow with successive rotational slides propagating rearward, and (b) for a spread involving the dislocation of the soil mass into horsts and grabens. Large arrows indicate the propagation direction of the failures. Small arrows indicate the main components of soil movement.

These simulation results are intriguing because crusts with variable thicknesses and strengths are almost always present on the top of sensitive clay deposits in Eastern Canada lowlands, including areas where large retrogressive landslides occurred. The goal of this paper is to examine whether these results are supported by observations made on several spreads and flows that occurred in southeastern Ontario and Quebec. First, the formation and the properties of crusts are reviewed. Then, criteria used to identify the failure mode are explained. Finally, results are presented and discussed.

2 CHARACTERSTICS OF CRUSTS OVERLYING SENSITIVE CLAY DEPOSITS

Surficial crusts overlying sensitive clay deposits are widespread in Eastern Canada lowlands due to the climatic and geological contexts of the region. In the literature, the terms ''crust'' most often implicitly refers to ''clay crust'' (e.g. Moum and Rosenqvist, 1957; Lefebvre et al., 1987). For this study, the definition of ''surficial crust'' is extended to any surficial unit having a shear strength or cone tip resistance well above values determined with in-situ field vane or piezocone tests immediately below in the intact clay deposit. This includes weathered clay crusts *sensu stricto*, and sandy units associated with regressive or fluvial-estuarine geological facies representing the final depositional stages of postglacial marine seas. According to this definition, a sandy cover does not need to be cemented to be considered as a crust. It was decided to include sandy crusts because their behavior during the failure propagation is probably not very different to the behavior of clay crusts as it will be seen.

2.1 Clay crusts

A surficial clay crust can develop from an unweathered clay deposit in response to a variety of chemical and physical processes. Seasonal cycles of freezing and thawing over millennia, along with fluctuations of the location of the groundwater table, are the major physical causes leading to a change in the mechanical properties of clays. Depending on the snow cover, the present-day average depth reached by freezing is about 0 to 2 m under the latitudes of the region of interest but could have been larger during colder periods in the last 10,000 years or so. Frost action causes water migration, which results in the formation of ice lenses and in a major change of the fabric of intact clays (Leroueil et al., 1991; Konrad et al., 1995).

The groundwater table fluctuations generate cycles of desiccation and wetting. These fluctuations can easily affect depths of two to four meters, especially close to slope crests where piezometric monitoring systematically indicates during dry periods deeper water tables than those on flat grounds. Infiltration of the surface waters, which usually have a chemistry different to the chemistry of the original pore water, is a secondary factor that can produce the oxidation and a partial cementation of the intact clays (Moum and Rosenqvist, 1957).

Trees also contribute to the weathering process by drawing groundwater from their roots. In addition, trees extract nutrients from the soil surrounding the roots, which can locally induce a strong chemical weathering. These effects are probably not negligible on regional scales if we consider the time period during which they were active. Most areas left free of water after the marine regression were rapidly colonized by trees (Dyke, 2005). By 6,000 years before present, a mixed forest with deciduous trees covered the entire St. Lawrence Valley and its tributaries up to the Saguenay region, while the lower north shore of the St. Lawrence River was occupied by a boreal forest. The maximum depth at which roots have a noticeable impact depends on the tree species, on the availability of water, and on the soil type. In studying the effect of trees on building settlement in the Ottawa and Montreal areas, Crawford (1968) and Silvestri et al. (1994) shown that changes in water content extended to depths exceeding three meters in clay deposits in the vicinity of some tree species common in the St. Lawrence Valley.

Another factor contributing to the development of a crust in clay deposits is the formation, by vertical and lateral erosion, of valleys, channels, and marine terraces. The associated stress relief results in the fracturing of clays in the eroded areas (Lefebvre and Morissette, 1984). It is often observed in fresh exposed lateral and back scarps of rotational slides that clays are traversed by a network of fissures reaching depths of two to three meters or more. These fissures are privileged paths for surface waters and tree roots which in turn facilitates weathering.

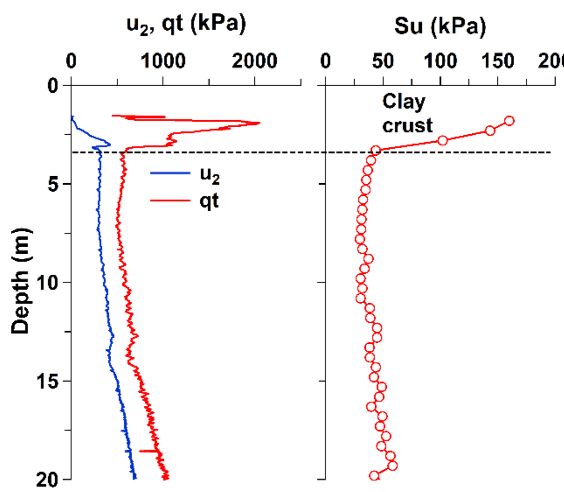

Figure 2. Piezocone and field vane tests profiles in a sensitive clay deposit overlain by a stiffer clay crust (case #32, Table 1). A hole was drilled to a depth of 1.5 m before beginning the piezocone test to avoid the desaturation of the porous element; u_2 is the pore pressure generated by cone penetration and qt is the cone tip resistance. Su is the soil shear strength determined from the field vane test.

As a result of a combination of these different processes, the thickness of the weathered clay crust generally varies from 1 to 6 m and is often of the order of 3 m on flat grounds (Lefebvre et al., 1987). It has long been recognized (e.g. Eden and Crawford, 1957) that the decrease of water content, and, possibly to a lesser extent, chemical weathering, lead to a significant increase of the undrained shear strength in the clay crust. Compared to unweathered sensitive clays, the plasticity indices and the remolded shears strengths in the crust are higher. Consequently, weathered clays are not sensitive to remolding. Typical resistance profiles are shown in Figure 2 for a site representative of the geotechnical conditions in Eastern Canada lowlands. This site is located just outside of a landslide scar that occurred along a small tributary of the Richelieu River about 30 km east of Montreal (case #32, Table 1). The undrained shear strength determined with the field vane test decreases from about 150 kPa close to the ground surface to about 40 kPa just below the crust. The cone tip resistance for a piezocone test done nearby follows the same trend.

2.2 Sandy crusts

As relative sea levels fell due to isostatic rebound following the last deglaciation, deltaic sands were deposited at the mouth of fluvial streams flowing into postglacial seas, like the Champlain Sea (e.g. Gadd, 1987; Parent and Occhietti, 1988). These deltaic units often cover marine clays and can reach thicknesses of a few tens of meters, particularly on the north shores of the Ottawa and St. Lawrence Rivers. They are generally organized in a complex succession of strata and channels of well sorted sands but can also locally contain finer grained sediments. Figure 3 shows a piezocone test result in a thick deltaic sand unit overlying a sensitive clay deposit at the rear of a spread (case #16, Table 1; see also Fig. 5). The cone tip resistance is much higher in the upper sandy unit than in the sensitive clays (note the difference in scales of Figs. 2 and 3) and shows strong variations over short vertical distances reflecting the heterogeneity of this sandy unit.

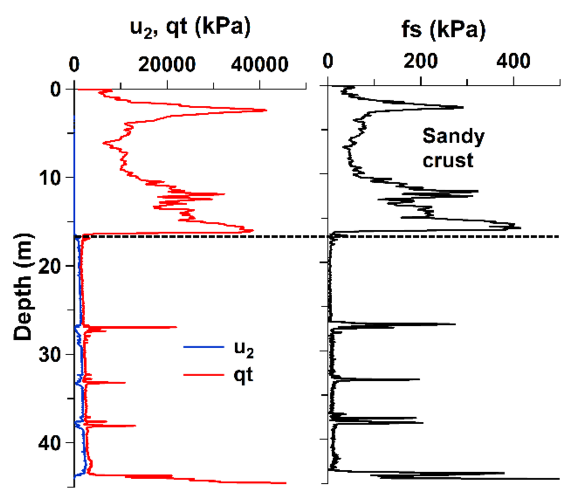

Figure 3. Piezocone test profiles in a sensitive clay deposit overlain by a stiffer sandy crust (case #16, Table 1); fs is the friction developed along the cone shaft during penetration.

During the last stage of the recession of marine waters, a fluvial system started to form in areas now approximately occupied by the Ottawa and the St. Lawrence Rivers (Gadd, 1987), while more open estuarine conditions existed farther downstream. Surficial sediments were reworked and redeposited as a thin blanket of sands or silts depending on the source material and on the distance over which they were transported. The thickness of these sandy fluvial or estuarine sediments is generally less than about 4 to 5 m and is on average 0.5 to 1.0 m. When this cover is relatively thin, a clay crust may have developed just below the sand cover once the lands emerged, as described previously.

2.3 Remarks on the behavior of surficial crusts

In the numerical simulations cited in introduction, the analyses are performed for undrained conditions because the few available eyewitnesses accounts indicate that large retrogressive landslides may occur very quickly, sometimes in less than a few minutes (e.g. Tavenas et al., 1971; Demers et al., 2014). In such rapid mass movements, the soils may not have time to drain and dissipate excess pore pressures during the propagation of failure. Although justified for the modeling of low permeability sensitive clay deposits, postulating that the whole soil profile behaves in an undrained manner is questionable when a clay or a sandy crust is present.

Field and laboratory observations show that clay crusts are highly fissured, both at the micro and macroscopic scales (Konrad et al., 1995). It has been reported by Lafleur and Lefebvre (1980) and Lafleur et al. (1987) that hydraulic conductivities are not controlled by the clay matrix but by these discontinuities. According to these authors, hydraulic conductivities in the clay crust can be higher by at least two to three orders of magnitude than in intact clays (10^{-8} - 10^{-7} m/s, compared to 10^{-10} - 10^{-9} m/s), which is likely enough to change its behavior from undrained to drained.

A related consequence of the presence of fissures is that the mass strength of the crust is similarly not controlled by the clay matrix. In a study on the back-analyses of several natural slope failures in Canadian soft clay deposits, Lefebvre (1981) underscored that ''it is not reasonable to assume that the soil can resist tensile stresses under long term conditions, especially in the shallow superficial zone which is known to be fissured''. In the limit-equilibrium slope stability analyses presented in that paper, a vertical tension crack full of water was considered in the crust. The crust was therefore modeled with a null strength.

Even if the crust behavior is not fully drained during the development of retrogressive landslides, the undrained shear strength as determined with the field vane test has been shown to be not compatible with the mobilized shear strength in back-analyses of failed embankments and excavations where a clay crust was involved (Silvestri, 1980; Lefebvre et al., 1987; Lafleur et al., 1988). Lefebvre et al. (1987) recommended to use, for practical purpose, the field vane strength measured in the intact clay immediately below the crust, which typically has a significantly lower value than the observed average shear strength in the crust (Fig. 2).

From the above, it appears that the behavior of clay crusts and sandy crusts may not be as different from what one might think at first glance. To support that view, it is interesting to recall 1) that Leroueil et al. (1991) observed that a sensitive clay subjected to freeze-thaw cycles had a sand-like behavior due to its micro-fissured nodular fabric; and 2) that hydraulic conductivities of sands, of the order of 10^{-7} -10^{-4} m/s, partly overlap those of clay crusts.

In accordance with this line of reasoning, it would thus be surprising that a clear relation exists between the presence of a crust and the landslide type.

3 SELECTED SPREADS AND FLOWS

A total of 37 large retrogressive landslide cases were selected for this study from case histories documented by the Ministère des transports du Québec (Fig. 4; Table 1). Most of these landslides are located between the Ottawa region and Quebec City, in areas that were covered about 10,500 to 12,000 years ago by the Champlain Sea (e.g. Parent and Occhietti, 1988). One landslide is in the Saguenay-Lake St. John region (case #1, Table 1), and three others are located east of Quebec City on the north shore of the St. Lawrence River (cases #2, 3, and 7, Table 1).

For a landslide to be selected, at least two piezocone tests had to be available, one outside the scar to determine the crust thickness, the other inside the scar to locate the failure surface and thus determine the thickness of the mobilized soil mass. For case #3, however, the thickness of the mobilized soil mass was estimated from visual observations.

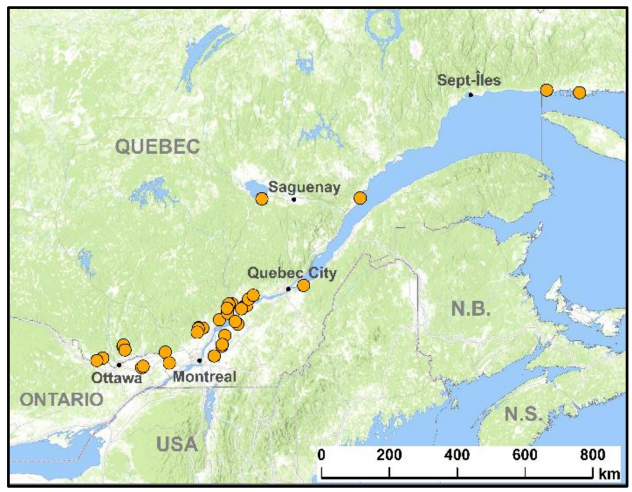

Figure 4. Location of large retrogressive landslides (orange symbols) considered in the present study.

3.1 Identification of landslide type

High resolution Lidar digital elevation models (DEMs) were systematically used to identify the type of large retrogressive landslides (see Demers et al., 2017, for an overview on the use of Lidar DEMs). In addition, geotechnical data and all other available information, including technical reports, historical documents, articles in newspapers, drawings and photographs, were considered to help determine or confirm the failure mode if necessary.

When a landslide scar showed a ribbed pattern on a Lidar DEM, the identification was easy and unambiguous, as illustrated in Figure 5 for the Notre-Dame-de-Lourdes landslide that occurred in the north shore of the St. Lawrence between Montreal and Quebec City (case #16, Table 1). The absence of a ribbed pattern or of a topography with horsts, however, is not a proof that the landslide was a flow and not a spread. In the rich farmlands of eastern Canada lowlands, the hummocky topography in landslide scars is often smoothed by earthworks to reclaim the land for agricultural purposes. A striking example is shown in Figure 6 for a landslide that occurred in 1975 near the town of St-Ambroise-de-Kildare, between Quebec City and Montreal on the north shore of the St. Lawrence River (case #20, Table 1). On a recent Lidar DEM (Fig. 6a), the bottom of the scar shows an even surface, and, in a first analysis, this lack of relief could be associated with a flow. On a vertical air photo taken in the days following the landslide, however, a succession of very well-defined horsts and grabens is clearly seen, indicating that this landslide is in fact a spread (Fig. 6b).

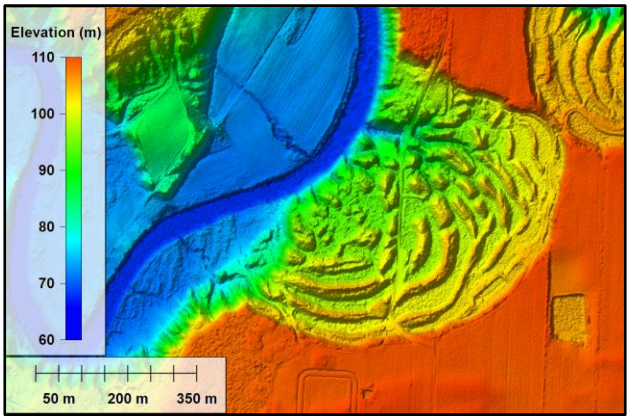

Figure 5. Lidar DEM of a landslide spread (case # 16, Table 1) showing a typical ribbed pattern with alternating horsts (protruding prisms) and grabens (sunken wedges). Part of another spread is visible in the upper right corner of the figure.

When only ambiguous visual information was available, identification was more challenging. In such cases, we considered the thickness of debris inside the scar, which is on average greater for spreads than for flows (Demers et al., 2014). In addition, the shape of the piezocone test profiles was taken into account. As the mobilized sensitive clays in flows can be almost entirely liquefied, the cone tip resistance and pore pressure profiles are typically shifted to much lower values in the debris, an attribute generally not observed in spreads. This is illustrated in Figure 7 for a flow that occurred in the Ottawa region in 2010. For old flows, a crust may have had time to develop in debris and the contrast in the profile shape may be less obvious.

In some landslides, the failure seems to have begun by a flow, which evolved into a spread (Tremblay-Auger et al., 2018). These cases are identified in Table 1 as compound landslides. We emphasize that this is a work in progress and that this classification is based on our current understanding. Some of these landslides could be reclassified in the future into either category as additional data become available or new conceptual models are proposed.

3.2 Determination of the thickness of the crust and of the mobilized soil mass

The determination of the thickness of crusts was most often straightforward. In the few cases where there was a gradual transition with alternating layers of sand and clay between the sensitive clays and a sandy crust, the determination was less easy, and some judgment was required to position the base of the crust.

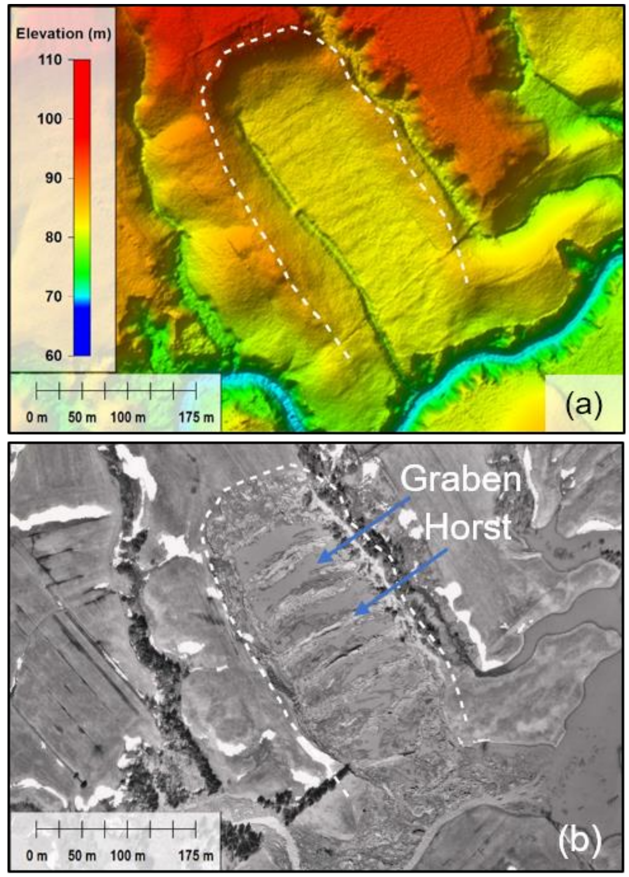

Figure 6. Effect of agricultural earthworks on the ground surface appearance at the site of the 1975 St-Ambroise-de-Kildare spread (case #20, Table 1). (a) Lidar DEM from a survey in 2008, (b) vertical air photo taken in 1975. The dashed white line circumscribes the landslide scar.

As shown in Figure 7, identifying the failure surface was generally not a problem for flows. The identification of the basal failure surface was trickier for spreads because this basal surface can sometimes be mistaken with the inclined interface between a horst and a graben and be positioned at a higher elevation (Demers et al., 2000). When more than one piezocone test was performed inside a scar, the consistency between the tests was used to detect these possible misinterpretations. The thickness of the mobilized soil mass was then estimated for each landslide and at each piezocone test location by reconstructing the pre-failure topography.

4 INFLUENCE OF CRUSTS ON LANDSLIDE DEVELOPMENT

The crust thickness is plotted in Figure 8 as a function of the thickness of the mobilized soil mass for the 37 cases listed in Table 1. For comparison, a few results of numerical simulations taken from the literature are also plotted in Figure 8. Different symbols are used for spreads and flows, and for retrogressive landslides involving these two modes of failure (compound landslides). A symbol without bars or with a bar in one direction means that only one piezocone test was available, either for the crust thickness or for the thickness of the mobilized soil, or for both. The symbols represent the average value in the case where more than two piezocone tests have been considered. Bar lengths illustrate the natural variability in crust thickness at a same site as well as the multi-stepped topography of the failure surface when detected or suspected. However, it cannot be ruled out that some of the variability in the thickness of the mobilized soil mass is due to a misidentification of the basal failure surface for some spreads, as mentioned in the previous section.

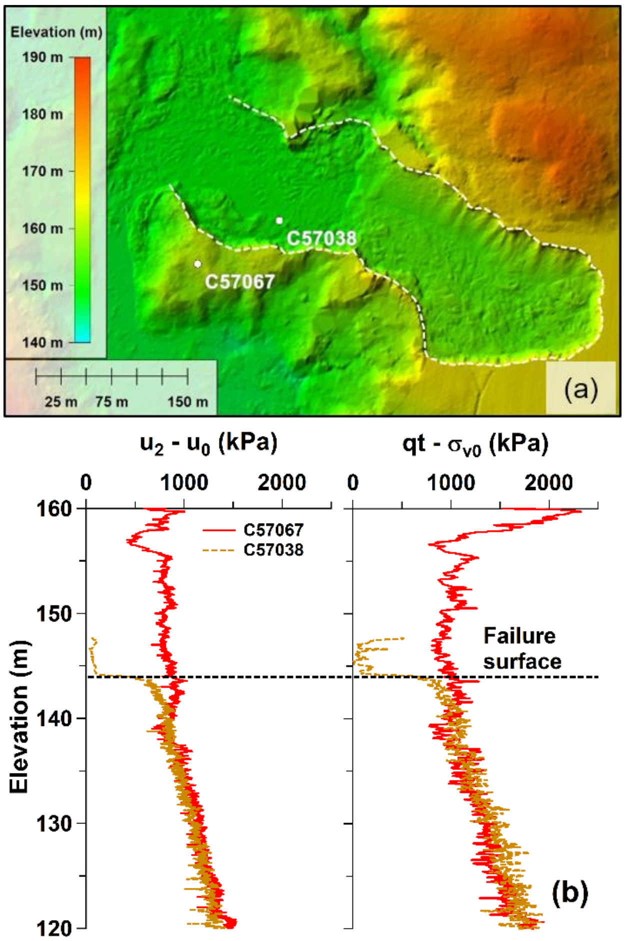

Figure 7. (a) Lidar DEM of the 2010 Notre-Dame-de-la-Salette flow (case #6, Table 1); (b) piezocone tests profiles showing the location of the failure surface. u_0 is the in-situ equilibrium pore pressure and σ_{v0} is the total vertical overburden pressure. The dashed white line in (a) delimits the boundary of the landslide scar.

Data are clustered into two groups, apparently irrespective of the thickness of the mobilized soil mass in each group: a first group with crust thicknesses lower than about 6 m, and a second group with crust thicknesses greater than about 10 m. Symbols for spreads, flows and compound landslides are intermixed with no discernible trend. Although not shown with a different symbol to avoid overloading the figure, all failures involving clay and sandy crusts in the first group overlap (Table 1). In the second group, which comprises spreads except for two cases, the crust consists mainly of sands. Interestingly, a spread in this second group occurred in a deposit with only a very small proportion of sensitive clay (case #16, Table 1). It is also worthwhile to highlight that the two flows in this group developed even if a thick sandy crust of the order of 40-50% of the thickness of the mobilized soil mass was present (cases #2 and 3, Table 1).

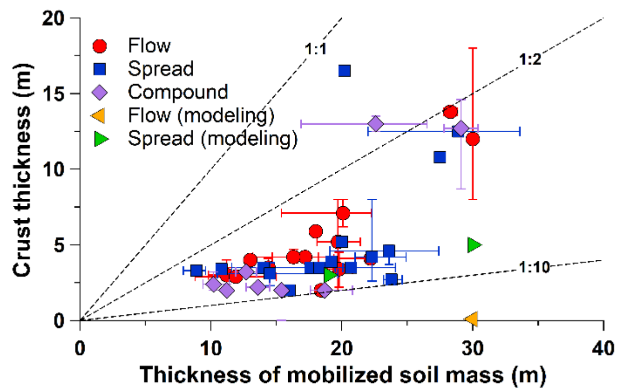

Figure 8. Crust thickness as a function of the thickness of the mobilized soil mass for the 37 landslide cases listed in Table 1. The bounds of horizontal and vertical bars, when displayed, correspond to the minimum and maximum observed values. The green and orange triangular symbols correspond to numerical simulation results (Saha, 2017; Wang and Hawlader, 2017). The different slope lines (1:1, 1:2 and 1:10) indicate the relative proportion of the crust thickness to the thickness of the mobilized soil mass.

5 DISCUSSIONS

As expected, no clear pattern emerges from the distribution of data points in Figure 8. The thickness of a crust, relative or not to the thickness of the mobilized soil mass, does not appear to be a discriminating factor controlling the development of either spreads or flows, or of retrogressive compound landslides in sensitive clay deposits. It was shown in Section 2 that crusts are almost ubiquitous in eastern Canada lowlands and that the nature of the crust, be it a clay crust or a sandy crust, is also probably irrelevant to explain why a landslide of a certain type can occur in a given area. Based on a brief review of field and laboratory observations, it was postulated that this absence of influence could be related to a similar behavior during the development of retrogressive landslides, both clay and sandy crusts sharing some common hydraulic and mechanical characteristics.

The clustering observed in Figure 8 into two apparently distinct groups may be an artefact due to under-sampling. The gap between the two groups would probably be reduced by investigating other landslide sites, particularly in geological environments where sands overlay sensitive clays. For reasons exposed in Section 2.1, it would be extremely unlikely to identify landslide sites with clay crusts much thicker than about 6 m. Incidentally, this is approximately the maximum clay crust thickness observed for the landslides listed in Table 1.

There is no evidence supporting that crusts can be modeled in numerical simulations as a unit having a higher undrained shear strength than the one in the underlying sensitive clays. On this subject, the point in Figure 8 corresponding to the simulation without a crust is unrealistic (but nevertheless useful in terms of parametric analyses), as no documented cases is characterized by an absence of crust. The fact that a flow can be followed by a spread during the same landslide event can be regarded as the best proof that the presence, the nature, and the thickness of a crust do not play a critical role in the development of a specific failure mode.

It is sometimes mentioned that the presence and the thickness of a crust tend to be correlated to the importance of salt leaching in the underlying clays, and indirectly to the mode of failure (e.g. Torrance, 2017). According to these views, leaching from below in response to ascending hydraulic gradients leads to a thin weathered clay crust while leaching from above by the downward infiltration of surface waters leads to a thick weathered clay crust. In the first scenario, the amount of highly sensitive clays with a low remolded shear strength is greater than in the second scenario, which enables the formation of flows. On the contrary, spreads would preferentially occur when a thicker crust is present because the amount of highly sensitive clays is proportionally less important. Although the amount of highly sensitive clays certainly does play a role, the crust thickness is not a good explanatory parameter as illustrated in Figure 8 by the distribution of data points in the first group of case histories.

An implicit assumption made in interpreting Figure 8 is that the investigations performed to characterize landslide sites are representative of prefailure conditions. The availability of high-resolution Lidar DEMs and, in several cases, of air photos taken before the landslides occurred allows for reconstructing the prefailure topography relatively easily. The thickness of the mobilized mass was therefore determined with reasonable accuracy. However, the thickness of the crust was measured outside landslide scars, in areas where the failure did not propagate further, meaning that conditions in these areas may not be representative of the prefailure situation. This is particularly possible in environments showing a high spatial variability of geological facies and of geotechnical properties. Although judgment has been used to exclude piezocone tests with unrepresentative profiles, this is an inherent limitation of such "after the fact" studies.

6 CONCLUSIONS

Among the many factors that may influence or control the occurrence of spreads and flows, results of numerical simulations presented in the literature suggest that the thickness of crusts overlying sensitive clay deposits is a possible discriminating factor. The analysis of the 37 landslide cases examined in this paper indicates that these numerical results are not consistent with field evidence, and that natural processes are probably more complex than currently recognized.

Crusts are present almost everywhere in regions that were inundated by postglacial seas. Based on field and laboratory observations, we argued that clay and sandy crusts behave similarly during large retrogressive landslides due to their hydraulic and mechanical properties, and that there is no obvious reason why they should play a critical role in the development of either flows or spreads, or compound landslides.

Other factors should better explain in which circumstances a flow or a spread can occur. To identify these factors and the way they may interact, numerical simulations with systematic parametric analyses should be pursued in conjunction with detailed field investigations of landslides that have occurred in as many different environments as possible. In particular, the role of the relative amount of potentially liquefiable clays with low remolded shear strength should be investigated in detail.

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Table 1. Large retrogressive landslides considered for analysis

Site #	Name	Type	Date	Latitude	Longitude	Surface Area[1] (ha)	R^2 (m)	W^3 (m)	Th. Crust[4] (m)	Th. MSM[5] (m)	Crust Type
1	Desbiens	Flow	1989	48.4267	-71.9147	0.9	65	105	2.9	11.9	Clay
2	Havre-St-Pierre	Flow	Unknown	50.2552	-63.5000	13.5	350	540	13.8	28.3	Sand
3	Les Escoumins	Flow	1986	48.4394	-69.3101	1.4	70	140	12.0	30.0	Sand-Clay
4	Maskinongé	Flow	1840	46.2571	-73.0378	29.4	1000	320	3.0	11.2	Clay
5	Notre-Dame-de-la-Salette	Flow	1908	45.7687	-75.5925	6.5	125	460	4.1	22.0	Clay
6	Notre-Dame-de-la-Salette	Flow	2010	45.7941	-75.5837	5.6	425	150	4.2	16.3	Clay
7	Rivière-St-Jean	Flow	1970	50.2978	-64.3717	3.6	235	215	7.1	20.1	Sand
8	Shawinigan South	Flow	Unknown	46.5487	-72.7051	110.0	1400	1200	5.2	19.7	Sand
9	St-Boniface-de-Shawinigan	Flow	Unknown	46.5260	-72.7855	24.0	475	500	2.0	18.4	Clay
10	St-Boniface-de-Shawinigan	Flow	1924	46.5476	-72.7906	3.1	285	140	4.2	17.2	Clay
11	St-Jude	Flow	1954	45.7940	-72.9723	1.7	95	120	4.0	13.0	Sand
12	Ste-Geneviève-de-Batiscan	Flow	1870	46.5095	-72.3681	4.2	90.0	125.0	3.5	14.4	Clay
13	Ste-Marcelline	Flow	Unknown	46.1154	-73.5838	51.1	1160	390	3.4	19.8	Sand-Clay
14	Brownsburg	Spread	1988	45.6642	-74.4718	2.8	75	145	3.1	14.5	Clay
15	Eardley	Spread	Unknown	45.5543	-76.1281	11.8	210	490	3.9	19.2	Sand
16	Notre-Dame-de-Lourdes	Spread	Unknown	46.1074	-73.4783	23.5	410	630	16.5	20.2	Sand
17	Poupore	Spread	1903	45.7043	-75.5403	31.0	675	570	3.5	14.0	Clay
18	Quyon	Spread	~1000 BP[6]	45.5077	-76.2861	23.8	510	550	4.6	23.6	Sand-Clay
19	Rigaud	Spread	1978	45.4639	-74.3654	3.3	85	285	3.5	18.3	Clay
20	St-Ambroise	Spread	1975	46.099	-73.5954	5.7	350	170	4.2	22.3	Sand-Clay
21	St-Barnabé	Spread	2005	46.3802	-72.8239	3.6	110	175	12.5	28.9	Sand
22	St-Boniface-de-Shawinigan	Spread	1996	46.4691	-72.8369	27.7	200	955	10.8	27.5	Sand
23	St-David	Spread	2015	45.9716	-72.8932	0.7	130	80	3.3	8.9	Sand-Clay
24	St-Jude	Spread	1925	45.7790	-72.9757	7.9	105	525	3.5	17.6	Sand-Clay
25	St-Jude	Spread	2010	45.8046	-72.9641	3.8	75	265	2.7	23.8	Sand
26	St-Liguori	Spread	1989	46.0306	-73.6259	7.0	85	500	3.5	20.7	Clay
27	St-Luc-de-Vincennes	Spread	1986	46.4648	-72.4427	6.0	130	305	5.2	20.0	Sand
28	St-Vallier	Spread	1935	46.8841	-70.8096	4.3	230	150	3.4	10.8	Sand
29	Ste-Monique	Spread	1994	46.1785	-72.552	5.8	130	400	2.0	16.0	Sand-Clay
30	Casselman	Compound	1971	45.3767	-75.0992	27.7	400	770	13.0	22.6	Sand
31	Lemieux	Compound	1993	45.4010	-75.0584	16.9	560	275	12.7	29.1	Sand-Clay
32	Mont-St-Hilaire	Compound	1859	45.5920	-73.1772	4.4	240	160	3.2	12.7	Clay
33	Nicolet	Compound	1955	46.2268	-72.6198	2.2	170	110	2.0	11.2	Sand
34	Ste-Geneviève-de-Batiscan	Compound	1939	46.5190	-72.3091	5.3	120	395	2.2	13.6	Sand-Clay
35	St-Luc-de-Vincennes	Compound	2016	46.4612	-72.4529	1.8	145	160	2.4	10.2	Clay
36	St-Prosper	Compound	1953	46.6350	-72.2674	16.0	750	300	2.0	15.4	Sand
37	St-Thuribe	Compound	1898	46.7054	-72.1452	34.0	900	500	2.0	18.7	Clay

[1]Includes the surface area of the horizontally projected pre-landslide slope; [2]Retrogression distance, calculated from the slope crest to the farthest backscarp; [3]Width, calculated along a line normal to the retrogression axis; [4]Average thickness of crust; [5]Average thickness of mobilized soil mass; [6]Before Present.