High-resolution reflection seismic investigations of quick-clay and associated formations at a landslide scar in southwest Sweden

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1. Introduction

Landslides are one of the most commonly occurring natural disasters; the economic cost of the damage caused by them is estimated to be in billions of dollars, and they claim hundreds of human lives every year (Petley, 2011). Sweden is also affected by this natural hazard (Nadim et al., 2008; Fig. 1). This study is about one particular kind of rapid earth flow that is caused by quick-clays, which mainly exist in Sweden, Norway and Canada (Rankka et al., 2004; Rosenquist, 1953; Solberg, 2007). Undisturbed quick-clay resembles a water-saturated gel that accumulated as flocculated silty-clay to clayey-silt sediment in a marine to brackish environment; uplifted above sea-level, and has been leached to low salinity by fresh water in a marine to brackish environment; uplifted above sea-level, and has been leached to low salinity by fresh water.

We present high-resolution reflection seismic data from four lines (total 1.9 km) that cross a quick-clay landslide scar located close to the shore of the Göta River in southwest Sweden, and compare the results with geotechnical data from boreholes. The seismic data allow the imaging of bedrock topography and normally to weakly consolidated sediments to a subsurface depth of about 100 m. Different types of seismic sources, including sledgehammer, accelerated weight-drop and dynamite were utilized and compared with each other. Analysis of their power spectra suggests that weight-drop and dynamite have higher frequency content and energy than the sledgehammer, which makes these two sources suitable also for waveform tomography and surface-wave data analysis. The shallowest non-bedrock reflector is observed at about 10–20 m below the surface, it overlays the bedrock, and is interpreted to originate from the contact between clay formations above and a coarse-grained layer below. The coarse-grained layer appears to be spatially linked to the presence of quick-clays. It is a regional scale formation, laterally heterogeneous, which deepens to the west of the study area and correlates well with the available geotechnical data. Continuity of the coarse-grained layer becomes obscured by the landslide scar. There may be a link between the coarse-grained layer and landslides in the study area, although this possibility requires further hydrogeological and geotechnical investigations. Reflectors from the top of the bedrock suggest a depression zone with its deepest point below the landslide scar and a bowl-shaped structure in the northern portion of one of the seismic lines.

According to our knowledge, there have been no previous applications of reflection seismic methods for delineation of quick-clays and associated formations in Sweden. Refraction seismic methods have larger deposits are sensitive to greater stress changes, such as increased saturation by excess rainwater (e.g., Vanneste et al., 2012). Considering the nature of quick-clay formations, application of geoelectrical methods alone to delineate them is a challenge, although they are often used. The presence of thick and conductive marine clay does not allow deep current penetration (Bogoslovsky and Ogilvy, 1977). These thick layers of conductive clay, which can also be fully water-saturated at depth, sometimes already as shallow as 0.5 m, make ground penetrating radar (GPR) techniques inefficient (Annan, 2005). However, each geophysical method has its own advantage and disadvantage for a given site location or problem and for specific geologic target at different depths (Bichler et al., 2004; Bogoslovsky and Ogilvy, 1977; Carpentier et al., 2012; Jongmans and Garambois, 2007). Success is not always guaranteed for a single geophysical method. Therefore, a combination of various geophysical methods and implementing those that can also provide deeper information is needed (Bogoslovsky and Ogilvy, 1977; Jongmans and Garambois, 2007; Solberg et al., 2012). This is particularly important if quick-clays extend to greater depths (e.g., >20 m), such as in this case study.

According to our knowledge, there have been no previous applications of reflection seismic methods for delineation of quick-clays and associated formations in Sweden. Refraction seismic methods have
been extensively used for landslide studies (Bekler et al., 2011; Carroll et al., 1972; Jomard et al., 2007; Samyn et al., 2012), however, the use of reflection seismic methods has been limited to countries outside Sweden (e.g., Bachrach and Nur, 1998; Baker et al., 1999a,b; Bichler et al., 2004; Büker et al., 1998; Eichkitz et al., 2009; Kaiser et al., 2009; L’Heureux et al., 2010, 2012; Medioli et al., 2012; Polom et al., 2010; Roberts et al., 1992; Schmelzbach et al., 2005). Reflection seismic methods are more expensive than the traditional geophysical methods for landslide studies, but if properly designed and implemented they can provide high-resolution images of layer boundaries, provided that there is sufficient acoustic impedance contrast (product of seismic velocity and density) between the layers. For example, seismic methods can be used to map bedrock geometry and overlying clays that can sometimes be critical for landslide risk assessments. Integration with other geophysical and geotechnical methods can also be important in modelling and interpreting the data (Eichkitz et al., 2009). Reflection seismic methods experience problems when imaging the top few metres (smaller than one wavelength) of the near surface (<5 m), but this gap can often be filled by geoelectrical and GPR methods (Bichler et al., 2004; Bogoslovsky and Ogilvy, 1977).

Our geophysical investigations of areas prone to quick-clay landslides began in September 2011 over a known landslide scar near the Göta River (Figs. 1 and 2a) in southwest Sweden. Göta River (Fig. 1) is the source of drinking water for about 700,000 people and is used extensively for industrial transportation. Therefore, areas near the river are highly industrialized and populated. The geophysical investigations involved 2D and 3D P- and S-wave source and receiver seismic surveys, geoelectrics, controlled-source and radio-magnetotellurics, Slingram, ground gravity and magnetic surveys as well as passive seismic monitoring (Fig. 2a and b). Prior to our investigations, the Swedish Geotechnical Institute (SGI) studied the site using various geotechnical and hydrogeological methods (Löfroth et al., 2011). Therefore, a wealth of geotechnical borehole data, mainly CPT (cone penetration test with tip friction generated by the rod string; Robertson, 1990), CPTU (cone

Fig. 1. (right) Landslide risk map of Sweden (from the Geological Survey of Sweden), and (left) showing location of the study area and recent quick-clay landslides in southwest Sweden (e.g., Surte, 1950, Göta, 1957, Tuve, 1977, and Småröd, 2006). Our study area is about 7 km north of the city of Lilla Edet, where the Göta landslide occurred in 1957.
penetration test with friction sleeve measuring pore pressure data) and laboratory measurements are available from the site. Among geophysical methods, only surface electrical resistivity tomography (ERT) and induced polarization (IP) methods had been carried out by SGI (Löfroth et al., 2011).

An overall assessment of different geophysical methods used in the study area was recently reviewed by Malehmir et al. (in press). Here, we present results and details of high-resolution shallow reflection seismic data acquisition, processing and interpretation along four seismic profiles (lines 2, 3, 4 and 5), some of which cross the landslide scar (see Fig. 2).

Fig. 3. A series of field photos taken (September 2011) from the study area, showing (a) the landslide scar (also see Fig. 2) and location of the seismic lines 4 and 5, its retrogressive nature and (b) tilting trees suggesting a slow mass movement (soil creep) toward the river.

Fig. 4. Two cross sections from (a) central and (b) western parts of the study area showing CPT and CPTU measurements including cone resistance (shown in MPa) and sleeve friction (shown in kN) in the boreholes (Fig. 2b). Section (b) partly coincides with the location of seismic line 2. CPTU data (sleeve friction) clearly suggest presence of a coarse-grained layer in the central part of the study area. This layer is not clearly observed in the western parts of the study area but is believed to be intersected where the boreholes ended (difficult to drill into it). A thick layer of quick-clay above the coarse-grained layer is estimated to be present. The coarse-grained layer is one of the main targets of our seismic investigation. Boreholes U07063 and U07065 were used for monitoring pore-pressure at four depth levels (P1–P4). These sections are modified from Löfroth et al. (2011).
lines 4 and 5 in Fig. 2). Only data along line 5 were presented by Malehmir et al. (in press). The main objectives of this study are (i) to evaluate the performance of high-resolution reflection seismic methods for quick-clay studies in Sweden, (ii) to image a coarse-grained layer that is overlaid by beds of quick-clay formations in the site, and (iii) to image the bedrock topography, and to obtain a better understanding of its relationships with the formations above it that might be important to pre-condition landslides in the site.

2. Quick-clays

Quick-clays in Sweden and Norway were (and still forming) formed as a result of processes that occurred since the last deglaciation (Bjerrum, 1954), which caused the study area to become ice-free c. 13 ka BP. As a result of isostatic depression, the deglaciation was accompanied by eustatic sea-level rise and a regional marine transgression. Melting water from the glaciers brought silt and clay materials from land into marine environment (Magnusson et al., 1963). Wasting of the Fennoscandian ice-sheet caused isostatic rebound and regional sea regression (e.g., Berglund, 1992 and references therein). The highest post-glacial coastline in the study area is approximately 150 m. Deeper clay formations that were deposited before the marine regression took place have had their salts leached out by artesian groundwater flow through fissures in the bedrock and permeable coarse-grained materials such as sandy to silty layers (Solberg, 2007; Solberg et al., 2012). When salt is leached from the clay formation, it becomes sensitive to excess stress (such as load from constructions or a significant increase in the pore water pressure due to excess water from rain or melting snow) and can result in a quick-clay landslide (Lundström et al., 2009; Vanneste et al., 2012). Quick-clay landslides are usually retrogressive (and often fan shape) or flake-type (Gregersen, 1981), starting near a river or lake, and progress gradually upslope. They have been known to penetrate several kilometres inland, and consume everything in their path. A well-known example of this regressive nature is the Rissa landslide in Norway, which lasted more than 45 min and progressed at a rate of 10–20 km per hour (Gregersen, 1981; L’Heureux et al., 2010, 2012; see also http://en.wikipedia.org/wiki/Rissa_Norway).

Quick-clay landslides are not particularly constrained to steep topo-slopes and have been known to slide even in low-to-moderate angle topographical slopes (e.g., 1:10 for the Rissa landslide). Some known examples of quick-clay landslides from Norway are Trøgstad (year 1967, 1 million-m$^3$, 4 human casualties), Rissa (year 1978, 33 ha, one casualty), and Verdal, (year 1893, 55 million-m$^3$, 116 casualties), and from Sweden are the Surte (year 1950, 22 ha, 1 casualty; Caldenius et al., 2012). Quick-clay landslides are usually retrogressive (and often fan shape) or flake-type (Gregersen, 1981), starting near a river or lake, and progress gradually upslope. They have been known to penetrate several kilometres inland, and consume everything in their path. A well-known example of this regressive nature is the Rissa landslide in Norway, which lasted more than 45 min and progressed at a rate of 10–20 km per hour (Gregersen, 1981; L’Heureux et al., 2010, 2012; see also http://en.wikipedia.org/wiki/Rissa_Norway).

3. Study area

The study area has been recently studied by SGI in a nation-wide project, which included investigations of areas prone to sliding along the Göta River (Löfroth et al., 2011). The focus of our study is a landslide scar about 300 m by 300 m in size located in the northeast of the selected area (Fig. 2). According to the Quaternary geological map of the study area (Fig. 2a), the southern margin of the river is dominantly composed of post-glacial tills and occasionally contains pockets of organic silt and clay. In locations where the river meanders, small pockets of river sediments (mainly silt) are also observed. Highland areas (Fig. 2b) consist primarily of glacial clays. Bedrock, of crystalline nature (granite to granodiorite), is occasionally exposed within the glacial clays; it is often surrounded by small unit of sandy to silty materials (Fig. 2a). This sudden exposure of the bedrock within the clay materials is an indication of its undulating nature.

Fig. 3a shows a series of field photos taken from the scar, which depict its regressive nature (fan shape). The landslide occurred at the southern bend of the Göta River where a small stream and river enter the main channel at about 100 m and 200 m east of the scar (Fig. 2), respectively. Tilting trees close to the riverbank suggest a slow mass movement towards the river (Fig. 3b). The scar is located at the southern bend of the river with possible large fluvial deposits, which may be leading to less erosion. However, the small stream and river may contribute to increased erosion at the southern bank of the river, and also locally disturbing the direction of the river flow. In following section, we describe available geological, geo-physical and hydrogeological data from the study area. Malehmir et al. (in press) summarizes these investigations. We present them again here for completeness.

4. Previous geotechnical and geophysical investigations

4.1. Site characteristics and geotechnical boreholes

For quick-clay identification, CPT, CPTU, CPTU-R (cone penetration test with resistivity measurements) and laboratory measurements had been done. Unfortunately, most of these measurements provide information at points only and thus no geological profile (cross section) is available from the site. These studies, especially CPTU and some laboratory measurements (at some selected depths) suggested the presence of coarse-grained materials at varying depth (from the surface) but in average about 20–30 m below the current surface topography. Interestingly, in most places quick-clays were found to overly coarse-grained materials. Due to the limitations of CPT methods, only a very few boreholes could penetrate through the coarse-grained layer. Geotechnical boreholes (CPT and CPTU) indirectly suggest that the coarse-grained layer varies in thickness from 0.5 m to more than 10 m. Fig. 4 shows two cross sections (no geological observation is available from most of these boreholes) from previous geotechnical investigations (Fig. 2) carried out by SGI in the study area. The results of CPT and CPTU measurements clearly show the presence of coarse-grained materials intersected by the boreholes. Detailed description of the cross sections can be found in Löfroth et al. (2011) and we present only a short summary relevant to our study.

In the section just west of the seismic line 5 (Fig. 4a), the uppermost layer consists of 2 m dry surficial crust. The results from CPT and CPTU measurements in point U07060 (for simplicity shown as 7060 in Fig. 2a) show that a 24 m thick sequence of clay may exist underneath the dry crust. A thick layer of coarse-grained material replaces a sandy layer found in the boreholes east of the scar. Results from static pressure soundings indicate that the thickness of this coarse-grained layer exceeds 15 m immediately west of the landslide scar. It could not be verified if there is clay underneath this layer. The
Fig. 5. CPTU-R and laboratory measurements carried out in boreholes (a) 7202, (b) 7203, (c) 7206 and (d) 7208 (Fig. 2b), showing a sudden change in the resistivity in boreholes 7202, 7203 and 7206 due to the presence of a coarse-grained layer. Borehole 7208 does not intersect this layer at least down to 35 m below the surface. Note that quick-clays are estimated to be present above the coarse-grained layer in both boreholes 7202 and 7203. Borehole 7202 is located in the landslide scar (Fig. 2a). A large thickness of quick-clay is estimated (based on CPT measurements; not shown here but reported in Löfroth et al. (2011)) to be present in borehole 7206 (see also Fig. 4a) but it is not clear if this occurs immediately above the coarse-grained layer. These data are kindly provided by SGI (from Löfroth et al., 2011).
quick-clay in this section also is about 15 m thick and overlies the friction materials (i.e., coarse-grained materials).

In the section shown in Fig. 4b, the thickness of the fluvial deposits is about 3 m, which increases to more than 6 m towards the west. The fluvial deposits consist of sand and silt with small amounts of organic materials and gyttja. Underneath the fluvial deposits, there is clay that continues to a depth greater than 45 m. The friction materials observed in the eastern section (Fig. 4a) are not clearly seen in any of the boreholes, which extend to a maximum depth of 45 m. However, we believe that most of the boreholes (e.g., 7063 to 7067) ended in hard materials (or presumably coarse-grained materials) and this is the reason why some of them are much shallower than the others (Fig. 4b). This is just a speculation and further investigations are required to prove this interpretation. The northwestern part of seismic line 2 overlaps with the section shown in Fig. 4b. Interestingly, no quick-clay is found neither in this section, nor in the boreholes further towards the west (Fig. 2b). Our seismic lines, which are shown later, suggest the presence of coarse-grained layer or a continuation of the coarse-grained materials that deepen westwards, implying that the boreholes in the western part of the study area ended in these materials. The coarse-grain materials come again closer to the surface and are intersected in line 3, which may also be an indication of its lateral heterogeneity and discontinuity (e.g., size and shape). Geotechnical data along line 3 (Fig. 2) are reported in Löfroth et al. (2011), thus, not shown here as their original format but discussed later in the paper.

Fig. 5 shows a selection of CPTU-R and laboratory measurements (e.g., fall-cone tests) on samples taken at a few locations from four boreholes adjacent to the seismic lines (Fig. 2b). We present the CPTU-R data here since they show a clear increase in the frictional resistance and electrical resistivity (Fig. 4) where the coarse-grained layer is intersected (Löfroth et al., 2011). As evident from the boreholes and the cross sections (Figs. 4 and 5) from the east to the west of the study area, the coarse-grained materials deepen with no indication of the coarse-grained layer in borehole 7208. The resistivity measured in borehole 7203 shows a clear change at about 20 m below the surface, close to the landslide scar and near the seismic line 5 (Fig. 2). Borehole 7202 shows a similar feature at about 10 m below the surface in the landslide scar, which is likely the same layer as observed in borehole 7203. The sensitivity of clays was also examined in both laboratory (e.g., fall-cone tests) and using CPT methods suggesting the presence of quick-clays above the coarse-grained materials (e.g., boreholes 7202 and 7203). However, a thick layer of quick-clay was estimated (based on CPT rod friction results) to be present in borehole 7206, but it is not clear if it is situated immediately above the coarse-grained layer.

4.2. Previous geophysical investigations

Previous geophysical investigations only consisted of ERT and IP methods and were carried out in an irregular grid consisting of eight 2D profiles (Löfroth et al., 2011). The 2D resistivity model along a line close and nearly parallel to the seismic line 5 showed a sharp vertical change in the resistivity from more resistive (30–200 Ω m) top layer (20–25 m thick) and a considerably more conductive (1–16 Ω m) underlying layer. The conductive layer dips gently towards the river. Note that these observations contrast with the CPTU-R results underlying layer. The conductive layer dips gently towards the landslide scar or to the east. In their report, Löfroth et al. (2011) concluded that the electrical resistivity data could not be used to identify both the coarse-grained layer and quick-clays. However, they mentioned the importance of using electrical resistivity data, which enabled them to distinguish a top resistive crust and a transition to an underlying more conductive clay that may be quick at the transition zone. Sudden variations of the electrical resistivity at depth detected by CPTU-R and CPT data, especially those from the coarse-grained layer (e.g., variations observed in boreholes 7202 and 7203; Fig. 2) could not be resolved by the ERT models (Löfroth et al., 2011).

4.3. Hydrogeological investigations

Pore-pressure monitoring carried out by SGI during 2010 in two boreholes (e.g., 7063 in Fig. 4) near the river for a period of about two years at 4 depth levels (3 m, 8 m, 15 m, and 30 m) did not show any strong temporal correlations between precipitation and pore-pressure changes (Swedish Geotechnical Institute, 2011). Therefore, pore-pressure effects for triggering landslides was suggested to be minimal, at least close to the river or where these investigations were carried out. Pore water pressure measurements in the area indicate that they are generally lower than the hydrostatic pressure implying a downward groundwater flow. At the time of measurements, pore water pressures were low, especially above the coarse-grained layer. Nevertheless, hydrogeological investigations suggest possible artesian groundwater flow especially near the river (Swedish Geotechnical Institute, 2011).

5. Seismic data acquisition

Seismic data acquisition (2D and 3D) was carried out in September 2011 for about two weeks. Four different types of seismic sources, namely weight-drop, sledgehammer, shear-wave source and explosive, were used. While majority of the seismic lines (including the 3D area) were acquired using a weight-drop source, data along lines 4 and 5 were acquired using 10–20 g of dynamite fired in holes about 0.5 m deep. The main reason to use explosives was to test their potential especially for imaging the bedrock at depth, reflections from and within the bedrock that may be important to understand the general structures of the subsurface, wide-angle reflections coming at large offsets and also

<table>
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<th>Source pattern 5 impacts/hole</th>
<th>Line 2</th>
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<th>Line 4</th>
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<td>SERCEL 428</td>
<td>SERCEL 428</td>
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Table 1 Main reflection seismic data acquisition parameters, September 2011.
to obtain wide frequency band seismic data that might be potentially useful for other studies, such as surface-wave analysis and waveform tomography. A SERCEL 428 acquisition system was used for the data recording and a differential global positioning system (DGPS) registered accurately the locations of shot and receiver positions. For lines 2 and 3 (Fig. 2) a 110-kg accelerated weight-drop (in most places) and a 15 kg iron sledgehammer (only in a few places where mobilizing the weight-drop was not possible) were used to generate the seismic signal. At every source location, five shots were generated and vertically stacked to increase the signal-to-noise ratio. We used 120 and 96 active/live geophones for lines 2 and 3, respectively. For lines 4 and 5 (Fig. 2), respectively 144 and 120 active/live geophones were used.

Fig. 6. (a) An example of a raw shot gather from line 2 (weight-drop source), (b) after vertical stacking of five repeated shots and (c) processed shot (including field statics, bandpass filtering and top mute). Bedrock reflection, B1, is clearly visible after some processing steps. Note the power spectra (0–200 ms window) in the lower right corner after each processing step.
Receiver spacing was 4 m while the shot spacing varied between 4 m to 12 m. These data were collected in about 8 days. We also collected a relatively higher resolution (2 m source and receiver intervals) reflection line (line 1) and only sledgehammer as the source to provide data for comparison with a shear-wave source and receiver set-up (lines S1a, S1b and 2S; see Malehmir et al., in press). These data are not discussed nor presented in this paper. Table 1 summarizes main acquisition parameters used to acquire data along lines 2, 3, 4 and 5.

Seismic data quality is generally good with clear direct and refracted arrivals observed on raw shot gathers but no clear shallow reflectors on the raw shot gathers. Data quality for the data along line 4 is partly lower than line 5 due to the presence of the river.

Fig. 7. (a) A raw shot gather acquired in the middle of line 5 (dynamite source), (b) after bandpass filtering and (c) field (refraction and elevation) static corrections. Note that at least three distinct reflectors (S1, S2 and B1) are already visible (see the red arrows) in the shot gather. Clear direct arrivals (1350 m/s) and refracted arrivals (apparent velocities are shown) are also observed on the raw shot gather. Also note that the critically refracted arrivals coincide with critically reflected arrivals (e.g., S1 and B1). Green curve shows the top mute function used to cancel the energy above and including the first arrivals.
The sledgehammer. All sources contain dominant signal frequencies in the frequency band of the weight-drop indicates better resolution than the sledgehammer (e.g., solid grey line in Fig. 8). The notably broader frequency content with broader bandwidth spectra as compared with power spectra are different for weight-drop and dynamite. Overall, the sledgehammer, weight-drop and dynamite sources. We used every fifth shot and then averaged all the power spectra for that specific source type. A comparison between Fig. 6a and b demonstrates improvement in the shot quality after the vertical stacking, especially at far-offsets. Fig. 7a shows an example of a raw shot gather using a dynamite source in the middle of line 5.

5.1. Comparison of various seismic sources

Selection of a suitable seismic source is important for acquiring good quality data during a shallow high-resolution seismic survey. Selection of source is also dependent upon the exploration targets and required resolution. The compromise between cost and effectiveness is always hard to balance; environmental restrictions and permission limitations are also other issues to keep in mind when choosing a seismic source. A sledgehammer is always cheaper and easier to operate and maintain in a near subsurface study. On the other hand, weight-drop and dynamite sources often consistently produce a higher frequency signal with higher amplitude, but cost more money and involve more safety and permission issues (Rashed, 2009). While there were various reasons (some mentioned above) to try different types of sources in the study area, the ultimate goal was to see which source could provide higher-resolution image for future studies in the site and in general for quick-clay landslide studies. Unfortunately, due to the limitation with time and budget, we were unable to test all these sources on one single line, but consider that near-surface conditions within such a small area is likely similar to allow such a comparison possible using different lines.

Fig. 8 shows the power spectra of reflection seismic data using sledgehammer, weight-drop and dynamite sources. We used every fifth shot and then averaged all the power spectra for that specific source type. Our analysis of the power spectra suggests that the weight-drop source (e.g., solid black line in Fig. 8) shows the highest frequency content with broader bandwidth spectra as compared with the sledgehammer (e.g., solid grey line in Fig. 8). The notably broader frequency band of the weight-drop indicates better resolution than the sledgehammer. All sources contain dominant signal frequencies around 30–150 Hz (>2 octave). The high frequency portions of the power spectra are different for weight-drop and dynamite. Overall, dynamite (e.g., solid grey and black arrowed lines) shows broader frequency range than the others. The loss of amplitude at higher frequencies represents the earth filtering effect at higher frequencies and at lower frequencies represents the effect of natural geophones frequency (i.e., 28 Hz). Power spectrum for the weight-drop source shows a different pattern for lines 2 and 3. Power spectrum along line 2 shows higher frequencies than line 3. Similar differences in spectral behavior are observed along lines 4 and 5 for the dynamite. Near-surface variations or topography may explain this behavior. Pullan and MacAulay (1987) explained a similar effect while comparing weight-drop data (75 kg) with shotgun data. When the surface was wet and contained fine-grained materials the shotgun produced a stronger signal than the weight-drop. However, in the case of dry and coarse-grain materials, the shotgun signal was weaker than the weight-drop. In our case, weather conditions were variable during the seismic data acquisition, including calm sunny days to rainy and stormy days. This variable weather can explain the difference between the power spectra from one line to another.

In summary, the explosive source produces the highest frequency band and best signal-to-noise ratio; because it contains higher energy related to higher burn/blast velocity and source containment than the others (Praeg, 2003; Yordkayhun et al., 2009). If the power spectra are described in terms of bandwidth, the weight-drop source is strongest in the intermediate frequency range compared with dynamite and the sledgehammer. Therefore, the weight-drop source might be better suited if for example a combination of shallow and deeper targets were considered. Moreover, it is cheaper and involves less permitting issues and is safer than dynamite.

6. Seismic data processing

For the data processing, we used a standard post-stack migration method but focused on important processing steps required for high-resolution imaging, such as field static corrections, bandpass filtering and velocity analysis (e.g., Bachrach and Nur, 1998; Baker et al., 1999a,b; Büker et al., 1998; Hunter et al., 1984; Kaiser et al., 2009; Malehmir and Bellefleur, 2010; Malehmir et al., 2011; Schmelzbach et al., 2005; Steeples et al., 1997). Low frequencies below 80 Hz for the dynamite data and 60 Hz for the weight-drop and sledgehammer data (Fig. 8) were rejected to reveal shallow hidden high-frequency

<table>
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<td>2.</td>
<td>Extract and apply geometry (CDP bin size 2 m)</td>
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<tr>
<td>3.</td>
<td>Trace editing</td>
</tr>
<tr>
<td>4.</td>
<td>Pick first breaks: full offset range, automatic neural network algorithm but manually inspected and corrected</td>
</tr>
<tr>
<td>5.</td>
<td>Refraction static corrections: datum 25 m, replacement velocity 1500 m/s, $v_0$ 600 m/s</td>
</tr>
<tr>
<td>6.</td>
<td>Geometric-spreading compensation: $v^2$t</td>
</tr>
<tr>
<td>9.</td>
<td>Direct shear-wave muting (near-offset) or attenuation (far-offset)</td>
</tr>
<tr>
<td>10.</td>
<td>Air-blast attenuation</td>
</tr>
<tr>
<td>11.</td>
<td>Trace balance using data window</td>
</tr>
<tr>
<td>12.</td>
<td>Velocity analysis (iterative)</td>
</tr>
<tr>
<td>13.</td>
<td>Residual static corrections (iterative)</td>
</tr>
<tr>
<td>14.</td>
<td>Normal moveout corrections (NMO): 60% stretch mute</td>
</tr>
<tr>
<td>15.</td>
<td>Stack</td>
</tr>
<tr>
<td>16.</td>
<td>$f_0$-deconvolution</td>
</tr>
<tr>
<td>17.</td>
<td>FK filtering</td>
</tr>
<tr>
<td>18.</td>
<td>$f_0$-deconvolution</td>
</tr>
<tr>
<td>19.</td>
<td>Trace balance: 0–200 ms</td>
</tr>
<tr>
<td>20.</td>
<td>Migration: finite-difference</td>
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<tr>
<td>21.</td>
<td>Time-to-depth conversion: using smoothed stacking velocities (1200 m/s to 1500 m/s)</td>
</tr>
</tbody>
</table>

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**Fig. 8.** Comparisons of power spectra (0–200 ms window) from sledgehammer, accelerated weight-drop and dynamite source data. The dynamite source produces the highest frequencies and best signal-to-noise ratio. On the other hand, the weight-drop shows higher amplitude energy at intermediate frequencies than the two other sources. Sledgehammer produces the weakest signal of all the three sources. Note that an average power spectrum containing more than ten shots is shown for every line or source type. See text for a detailed description about the figure.
Fig. 9. A series of unmigrated stacked sections along line 3 obtained using (a) 1350 m/s stacking velocity and (b) 800 m/s stacking velocity. (c) Merged data from (a) and (b). (d) FK-filtered and coherency enhanced version of (c). Reflection B1 likely represents bedrock topography (see Fig. 2); S1 and S2 are interpreted to be two sets of normally consolidated sedimentary layers above the bedrock. See text for further explanation about the figure and events marked in the sections.
reflections (Steeples et al., 1997). This rejection proved to be critical in revealing reflections as shallow as 20–30 ms (Figs. 6c and 7b). We used a time-variant bandpass filter to enhance deeper reflections from within the bedrock. Refraction static corrections were estimated using a single-layer model (e.g., Malehmir and Juhlin, 2010). Combined with elevation statics helped improving the continuity of the reflections in raw shot gathers (Figs. 6c and 7c). As demonstrated by several authors (e.g., Schmelzbach et al., 2005), top mute (cancelling amplitudes above and including the first arrivals) and the amount of stretch mute (wavelet stretching) allowed after NMO corrections are important to cancel first arrivals and avoid misinterpreting them as reflections. We added 5 ms to the picked first arrivals and used that as a top mute function (Figs. 6c and 7c). As evident from Fig. 7 (also Fig. 6), at far-offsets especially greater than 150 m, direct arrivals are received later than the refracted arrivals. Obviously, our top mute function is not able to remove these arrivals. Although a manual mute was possible, we left these arrivals in the data and hoped they would cancel each other during the stacking process, which did prove to be the case. Although most shear-wave and surface-wave energies were attenuated by the bandpass filtering applied to the data, some remained which were attenuated at near-offsets (using a median filter) and muted at far-offsets (Table 2). After various tests, we eventually decided to use 60% for the stretch mute. Higher percentage of stretch mute was still possible but generally resulted in lower frequency and partly distorted seismic images and, therefore, it was not applied. Bedrock reflections (e.g., B1 in Fig. 7c) are often observed at wide-angle (and large offsets). Significantly lower stretch mute than 60% (e.g., 30%) resulted in poor image of the bedrock and lower fold coverage at shallower travel times. Such low stretch mutes were not applied.

Stacked sections were extremely sensitive to the velocities used, especially for the data collected along lines 2 and 3. To obtain best stacking velocities, we ran a series of constant velocity stacks using velocities ranging from 500 m/s to 2500 m/s. These allowed us to successfully define velocities required to image individual reflections. Final stacking velocities used to image the reflections above the interpreted bedrock range from 800 m/s to 1500 m/s. Higher-velocities up to 5000 m/s, were used below the bedrock. This method of finding stacking velocities worked well for lines 2, 4 and 5 but failed for the data along line 3. For the data along line 3, we could not use a common stacking velocity function, which can be used to image both bedrock reflection and reflections...
They (i.e., bedrock and reflections above it) required significantly different velocities and choosing one velocity for one target resulted in losing the other ones. The time section along line 2 was stacked with stacking velocities ranging from 800 to 1350 m/s. A DMO (dip-moveout) approach was also tested but it did not improve the results, probably because these reflections are very close to each other (about one wavelength) our processing approach would not allow simultaneous imaging of all these reflections at the same time.

Neither a smooth nor a detailed and significantly varying velocity function was helpful. Fig. 9a and b shows two time sections for the stack with different stacking velocities around 1350 m/s and 800 m/s, respectively. To be able to provide an image containing all these reflections, we decided to merge the two individual stacked sections (i.e., Fig. 9a and b) by selecting odd traces from one section and even traces from the other one (Fig. 9c) and by applying an FK-filter to the merged sections (Fig. 9d). Juhlin et al. (2010) recently reported a similar approach for data acquired to image the upper crust.

A coherency enhancement filter was used after the stacking. This filter helped to improve the quality of the reflections, but it also artificially enhanced steeply dipping structures, which we believe were the artefacts of the surface-wave and direct arrival data. We used a dip filter to remove these steeply dipping artefacts in the stacked sections. A final pass of coherency enhancement filter successfully helped to improve the quality of the unmigrated stacked sections. Seismic data were migrated and depth converted using smoothed version of the stacking velocities. Table 2 summarizes main data processing steps used in this study.

7. Results and interpretations

7.1. Seismic lines 2 and 3

Fig. 10a and b shows migrated and time-to-depth converted stacked sections along lines 2 and 3, respectively. The shallower imaged reflector (B1), a few metres below the surface, is observed in line 2 and is interpreted to be from the top of the bedrock. Combined with the bedrock outcrops observed in both western and eastern sides of the line (Fig. 2), this interpretation suggests that the bedrock has a bowl-shaped geometry in the southeastern portion of the line. This geometry implies that clay materials in the most southeastern portion of the line, even if are quick-clays, are unlikely prone to slide because movement of these materials is restricted by the sides of the bedrock depression. Two sets of reflections are observed in the northwestern portion of the line, but it is not clear which one is from the bedrock. It is possible that the shallower reflector (i.e., dashed green line in Fig. 10a) represents coarse-grained materials onlapping the bedrock (i.e., dashed orange line in Fig. 10a) but it is
also possible that the shallowest reflector is from the top of the bedrock too. In either case, our interpretation of the geotechnical boreholes (Fig. 4b), if correct, suggests hard materials as shallow as 15 m below the surface (e.g., 7066 and 7067), which is in agreement with the shallowest reflector observed in this portion of the line. A good agreement between the depth of this reflector and data from geotechnical boreholes is also observed in the remaining portion of the line.

However, reflectors along line 3 show different characters than those observed in line 2 (Fig. 10b). A bedrock reflector (B1) clearly approaches the surface and correlates well with the exposed bedrock in the southeastern portion of the line. Two sets of reflectors, S1 and S2, are imaged above the bedrock gently dipping towards the river but onlapping the bedrock. We were not surprised to see some disagreement between the reflectors and the geotechnical data because the data along this line were processed somewhat differently (Fig. 9), however, a careful comparison between the reflectors and geotechnical data (Fig. 10b) suggests an excellent agreement between the two data sets, which further supports our processing approach for this line. In particular, reflection S1 matches well (except at 7070) with a thin (≤2 m) layer of hard materials (possibly coarse-grained) imaged as shallow as 10 m (also see Löfroth et al., 2011). Most geotechnical boreholes ended at reflection S2 suggesting a second coarse-grained layer at deeper levels.

Geotechnical boreholes did not suggest presence of any quick-clay in this part of the study area (Löfroth et al., 2011).

7.2. Seismic lines 4 and 5

Fig. 11a and b shows migrated and time-to-depth converted stacked sections along lines 4 and 5, respectively. Reflectors as shallow as 10–20 m (e.g., S1 below the landslide scar) are observed along line 4. They generally dip gently towards the west and are discontinuous in the middle of the profile. In contrast, reflectors along line 5 show both flat-lying and moderately north dipping characteristics, especially in the southern portion of the line. In general, three distinct reflectors are observed along seismic line 5: one at about 10–20 m depth (or about sea-level; S1 in Fig. 11b), one at about 40–45 m depth (i.e., about 20 m below sea-level; S2 in Fig. 11b) and another one that dips to the north (the bedrock reflection). The latter starts at about 10 m depth in the southern portion of the line and extends to a depth of about 90 m in the middle (B1 in Fig. 11b). We interpret the deepest reflector (B1) to represent the bedrock topography, with its deepest point located at a depth of about 100–120 m and potentially located below the landslide scar (Fig. 11b). No borehole intersects the bedrock along the seismic lines, thus, we can only speculate about the nature of the reflector and...
what it may imply concerning the landslide evolution. Bedrock is exposed at about hundred metres west of the southern portion of the line 5 and this maybe an indication that the reflection B1 is also from the shallow bedrock in the southern part of the line (Malehmir et al., in press). In contrast with the other three lines, the bedrock reflector is not evident at the landslide scar nor along line 4 and is as deep as 80–100 m in the northern part of the study area.

Another reflector located nearly at sea-level is observed along line 4 (S1 in Fig. 11a) and is similar to the one observed in lines 3 and 5. The reflector is nearly flat lying along line 5 but shows clear westwards dip component along line 4 (Fig. 11a). This reflector onlaps the bedrock in line 5 and shows a good correspondence in both lines 4 and 5. However, its character becomes unclear at the location of the landslide scar (Fig. 11b). Borehole data suggest that the reflector (S1) is from the contact between the clay formations above and the coarser-grained materials that were identified by CPT and CPTU-R measurements, potentially sandy-silty or coarse-grained materials (Figs. 4 and 5). This reflector occurs at about 10–15 m below the surface at the landslide scar and reaches a depth of more than 50 m (below the surface) in the westernmost portion of line 4. This depth may correlate with the reflector (green dashed line in Fig. 10a) observed in line 2, although not clearly visible because it occurs, in both lines, at the edge of the seismic lines where the signal-to-noise ratio is usually low. It is also possible that there is no S1 reflector in line 2 which would then indicate geological changes in this location.

We also observe a short reflector (S2 in Fig. 11) below the coarse-grained layer and above the bedrock, which similar to our interpretation for line 3 (Fig. 10b) may represent another contact between a coarse-grained layer and clay formations. This layer is located below present day sea-level and may not directly be relevant for triggering landslides in the study area. However, it may be important for the presence of another artesian aquifer below the landslide scar. No borehole reaches the S2 reflector along lines 4 and 5. However, the presence of marine clays below the coarse-grained layer has been confirmed by a few boreholes (e.g., 7202). A noticeable difference in the reflection characteristics of the data between lines 4 and 5 maybe due to the lower signal-to-noise ratio along line 4 since it is close and parallel to the river. A more plausible scenario to explain this could also be that line 4 is parallel to the structural geometry of the bedrock (opposite to the dip direction) and this results in imaging out-of-plane reflections from undulated surfaces of the bedrock (Fig. 11a).

8. Correlation and 3D visualization with geotechnical boreholes

Fig. 12 shows 3D visualization of the seismic lines 4 and 5 with the geotechnical boreholes and it demonstrates a good correspondence between the reflectors and available borehole data. The borehole data are the CPT and CPTU-R measurements carried out by SGI (Löfroth et al., 2011), which attributes the sudden increases in the penetration resistance and resistivity to the presence of a coarse-grained layer. Seismic data clearly indicate the presence of a reflector at the location where this layer was detected and suggest that the coarse-grained layer can be a regional scale formation in the study area. Seismic data have been able to clearly image this reflector along line 4, showing a westward dipping component of about 10 degrees. This depth also correlates well with the borehole information further demonstrating why the layer was not intersected by for example borehole 7208 (Fig. 5) or any of those located in the western part of the study area (Fig. 12). Penetrations of these boreholes were simply too shallow (about 10 m) to intersect the layer or there is a sudden break in the continuity of the coarse-grained layer from the east to the west at the location of seismic line 2 (Fig. 10a). Geotechnical data from CPTU, CPT and laboratory measurements (Figs. 4 and 5) suggest the presence of a thick quick-clay formation above the coarse-grained layer especially in the eastern side of the study area except in the old landslide area. It should be noticed that borehole 7206 and those shown in Fig. 10a intersect a large thickness of quick-clay formation near where S1 reflector suggests a depression in the western part of line 4 (Figs. 11 and 12).

Fig. 13 shows 3D visualization of all the four seismic lines and our interpretation of the bedrock geometry and the S1 and S2 reflectors. Bedrock is well imaged in all the profiles except line 4 and this maybe a confirmation that recognizing the bedrock, especially if its structural geometry is parallel to the line is difficult. Seismic data along lines 2 and 4 at where they cross each other show a correlation in terms of reflectivity pattern but do not support the number of reflectors observed in each of them. The fact that this occurs at a place where the seismic fold is low and this inconsistency maybe resulted from that does not allow further speculation about the results in this part of the study area. It is clear that line 3 shows both S1 and
S2. Seismic data along line 1 (Fig. 2; see Malehmir et al., in press) also show two clear sets of reflectors down to the bedrock which further confirm the presence of two distinguishable units, most likely coarse-grained materials, above the bedrock in the study area.

9. Discussion and implications

The most significant feature in the seismic data is the reflection that originates from the coarse-grained layer (S1). If in fact this reflector is originated from a sandy-silty layer with higher permeability than the clays above and below, it could act as conduit for fresh groundwater. Water from the highland areas in the south into the landslide scar and eventually into the river. The bedrock is closer to the surface in the southern part of the study area (Figs. 10 and 11), and the reflector (the coarse-grained layer) onlaps the bedrock, suggesting that there might be stronger water infiltration into the clay from the southern part of the study area than the northern part. In this scenario, groundwater recharge (infiltration) can take place where there are outcrops of bedrock. It is not clear at the moment if a change in the water flow (pore-pressure) in the coarse-grained layer could alone or in combination with the water flowing from the river into the formation be the main pre-condition of the landslide in the site. Therefore, it is necessary to investigate this scenario using down-hole geophysical and hydrogeological methods in the southern part of the line 5. The fact that we are able to image the coarse-grained layer is encouraging and demonstrates the value of high-resolution reflection seismic methods in delineating such detailed (thin) structures that may contribute to the sliding potential of quick-clays and help to improve landslide risk assessment in Sweden. Shear-wave reflection seismic data (e.g., S1a) collected from the southwestern part of the study area also suggest two sets of reflections above the bedrock, consistent with the P-wave reflection seismic data (Malehmir et al., in press). Previous (Löfroth et al., 2011) and preliminary ERT models (Malehmir et al., in press) could not image the coarse-grained layer. The loose character of the coarse-grained layer at the location of the landslide scar (Fig. 11b) might be an indication that the layer was disturbed during the landslide or is an indication of large lateral heterogeneity across the layer. A depression zone of similar character, but of lesser extent, is also observed in the middle of line 4 and this zone can be an important factor for future landslides in the study area (Fig. 11a). The geotechnical data suggest the presence of a thick quick-clay formation at this location (Figs. 4 and 5) and we recommend monitoring and detailed studies of this part of the line.

Bedrock topography is particularly well defined along lines 2, 3 and 5 (Fig. 13). The surface elevation map (Fig. 2b) shows a gentle dip (about 5°) toward the river; however, bedrock topography imaged by the seismic data demonstrates a dip of about 25° towards the river (Figs. 10 and 11). This is certainly an important outcome of the seismic survey suggesting that, locally, surface topography is not exactly controlled by the bedrock topography and that the bedrock dips steeper than the surface topography. This is not surprise since this was an aquatic sedimentary nature prior to regression. Although geological features located below present day sea-level maybe irrelevant for quick-clay formation and generation of landslides, they are important structures that define the subsurface geometry of the normally consolidated sediments. Their shape and geometry may control the extent of the area that a landslide can affect. The fact that the bedrock is imaged as shallow as only a few metres (Figs. 10 and 11) again proves the potential of high-resolution reflection seismic data, provided that they are properly acquired and processed. Good ground conditions with highly saturated clays and a significant seismic contrast between the clay and coarse-grained layers have contributed to such an excellent data quality.

Although our ultimate goal in this study was to assess the potential of seismic methods for imaging structures associated with quick-clays, on the basis of the results, we are able to suggest a few different scenarios that may explain the formation of quick-clays and their landslide generation in the study area. Various people (e.g., Hjördis Löfroth, personal communication, 2012) have discussed these ideas but now we provide seismic evidence about the subsurface geometry, which can support some of these ideas. The first scenario (Fig. 14a) suggests that
the coarse-grained layer, a regionally important formation, but laterally heterogeneous, acts as a conduit for groundwater flowing from highland areas. As it plummets below the clay formation, an artesian aquifer is formed and salts can be leached from the nearby clays and transported to the river. Periods of high rainfall or snowmelt, and hence recharge of the aquifer would increase pore-pressure. A second scenario (Fig. 14b) suggests that during periods of higher river level (perhaps due to precipitation events or snowmelt further upstream) water is directed into the coarse-grained layer and a combination of erosion at the riverbank and increased pore-pressure in the layer close to the river is responsible for the generation of landslides in the study area. A high rate of erosion, due to the stream and the small river close to the landslide scar, perhaps expedited this process at this site. The third scenario (Fig. 14c), our preferred one, is a combination of the first and the second scenarios with some exceptions, suggesting that the quick-clays are formed mainly due to the infiltration of surface water into the clay formations and partly due to the circulation of fresh water into the clay formation from an artesian aquifer (i.e., the coarse-grained layer). It is possible that the coarse-grained layer, locally highly permeable, has acted as a conduit for water from either highland areas or the river and has contributed to the formation of quick-clays. It is also equally, if not highly, possible that the coarse-grain layer has acted as a drainage conduit directing the salts leached downwards from the overlying sediment toward the river. In their review article, Malehmir et al. (in press) suggested that a decreasing value of the resistivity observed on CPTU-R data (Fig. 5a) from the top of the quick-clay zone to its base might imply an increase in the pore water conductivity with depth and hence, an increasing salinity value (even if it is low). They concluded a possibility that the original marine pore water was displaced by water percolating downward from the surface. The coarse-grained layer would have then facilitated movement of water and its dissolved chemicals to the Göta River, as they exited from the base of the overlying clay layer (Malehmir et al. in press). What this implies is that the coarse-grained layer has a little significance in the development of the quick-clay zone, a point that needs to be confirmed by additional investigations (Malehmir et al. in press). A landslide triggering mechanism due to increased pore-pressure in the coarse-grained layer cannot be entirely ruled out, but this hypothesis is not supported by the results of pore-pressure monitoring carried out close to the river. Nevertheless, it should be noted that if the layer drains readily to the Göta River, it probably does not build up substantial excess pore pressures or artesian pressures. Erosion by the Göta River that steepens its bank, and saturation of the clay slopes associated with flooding stages of the Göta River, or periods of heavy or prolonged rainfall that saturates the slope, are more probable triggers; possibly after a flushing event and fall in river level after a high stand. The very small slides along the Göta River (Fig. 2) are most likely caused by erosion and slope saturation. An initial landslide in the slope materials, which almost always precedes a larger landslide that involves quick-clay, was probably also caused by the erosion/rainfall combination.

To check the validity of all these scenarios, further investigations and studies using ground-based and down-hole geophysical as well as hydrogeological site investigations should be carried out. A borehole located in the southern portion of the line 5 is best suited for this purpose. These scenarios are all important when climate change and water-soil interactions are considered. Long term geophysical (4D) and hydrogeological investigations using active and passive monitoring techniques (e.g., Pennington et al., 2009) would certainly be helpful to provide useful insights to explain mechanisms behind quick-clay formations and landslides. Future site investigations should also aim at drilling deeper boreholes at a few locations along lines 2, 4 and 5 to constrain our interpretations (e.g., Medioli et al., 2012) and if possible to include subaqueous geophysical measurements in the river.

Our results demonstrate the potential of high-resolution reflection seismic methods for not only delineating shallow structures that are associated with quick-clays but also filling the information gap between the geotechnical boreholes, which are often far apart. They should be attempted regularly in combination with other geophysical and geotechnical methods especially in areas like the one shown in this study where thick clay formations would limit depth penetration of traditional geoelectrical methods to only a few tens of metres.

10. Conclusions

Shallow high-resolution reflection seismic data, for the first time in Sweden, were acquired and processed along four lines to image subsurface formations at a quick-clay landslide site. All the seismic lines contain numerous reflectors with different characters. Reflectors that are probably of regional importance in the study area are (1) a flat-lying reflector at about 20–30 m below sea-level that originates from the contact between the clay formations and a coarse-grained or sandy-silty layer. The estimated depth of this reflector at the eastern parts of the study area correlates well with the results from the geotechnical investigations carried out in a few boreholes located close to the seismic lines. Borehole information suggests that the quick-clay usually occurs above this layer. However, exact relationships between the reflector and the occurrences of the landslides in the site require further investigations. The continuity of the reflector becomes loosely defined where the landslide scar exists, which maybe an indication that there is a link between the reflector and the occurrence of landslides in the site. The reflector dips about 15° towards the west also partly confirmed by geotechnical boreholes which show no evidence of its presence in boreholes drilled down to about 45 m depth in the western part of the study area, and (2) reflectors from the bedrock.

Various types of seismic sources were utilized in this study. Comparisons between the power spectra of the raw shot gathers suggest that the weight-drop and dynamite are better suited for seismic imaging in the study area than the sledgehammer. Dynamite source shows broader frequency bandwidth, which is useful if additional studies such as surface-wave and waveform tomography are also intended.

The level of the details revealed by the high-resolution reflection seismic data demonstrates the ability of the method to image fine structures associated with quick-clay landslides, which are important and valuable for any site risk assessment especially in areas where thick clay formations and deep bedrock topography are expected.

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