Characterizing large-scale glaciotectonic sediment deformation using electrical resistivity methods

R L Aylsworth Jr\textsuperscript{1,2}, R L Van Dam\textsuperscript{1,3}, G J Larson\textsuperscript{1} and M A Jessee\textsuperscript{1,4}

\textsuperscript{1} Department of Geological Sciences, Michigan State University, 206 Natural Science Building, MI 48824, USA
\textsuperscript{2} Ingrain, Inc., 3733 Westheimer Rd., 2nd Floor, Houston, TX 77027, USA
\textsuperscript{3} Queensland University of Technology, Institute for Future Environments, Gardens Point Campus, 2 George Street, Brisbane, QLD 4000, Australia
\textsuperscript{4} Department of Geological Sciences, Indiana University, 1001 East 10th Street, Bloomington, IN 47405, USA

E-mail: rvd@msu.edu

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Abstract

Large-scale sediment deformation structures formed by glaciotectonic processes have been identified south of Ludington, USA. Here, several apparent clay diapirs rise from below beach level to near the top of an approximately 60 m high bluff along the eastern shore of Lake Michigan. Throughout the area, the surface topography and locations of springs indicate a complicated subsurface structure and a preferred pattern of groundwater drainage. Since public borehole information is sparse, it is not known whether the structures exposed in the bluff are true diapirs or ridges, and if the latter, what is their orientation. In this paper we present the results of field, laboratory, and modeling studies to characterize the inland extent and orientation of these deformation structures using galvanic-source electrical geophysical methods. We exploit the large electrical contrast between a sandy sedimentary layer and an underlying clayey silt sedimentary layer in which the deformation occurred. Constant-spread traverses and multi-electrode tomographic data demonstrate that at least one of the narrow structures extends a significant distance inland.

Keywords: sediment deformation, glaciotectonism, electrical resistivity, subsurface characterization, diapirs, coastal bluff

(Some figures may appear in colour only in the online journal)

Introduction

Glaciotectonic deformation has long been studied as a means to reconstruct ice-flow dynamics (Evans and Rea 1999, Benediktsson et al 2008), ice-flow directions (Hieck and Dreimanis 1985, Aber and Ber 2007), glaciotectonic and glacio depositional processes (Hart and Boulton 1991, Hambrey and Huddart 1995), and stratigraphic and structural relations (Henriksen et al 2001, Morton et al 2004), among others. Common approaches for characterizing glaciotectonic deformation include outcrop studies, surface mapping, and coring methods (e.g. Boulton 1999, Arnaud 2008). Studies of outcrops and bluffs can provide excellent 2D information of grain size distribution, stratigraphy, and structural features (Ó Coi Gaigh and Evans 2001, Larson et al 2003), but are limited to available exposures. Mapping data from satellite remote sensing and aerial photos in terrestrial settings and from sonar in offshore settings allow for large spatial coverage in characterizing glacial deformation structures, but do not provide direct subsurface information (Smith et al 2006). Borehole methods, including geophysical borehole logging, core penetrometer tests, and direct-push methods, allow gathering of high-vertical-resolution data on textural, geotechnical, and hydrogeological properties (Bakker and van der Meer 2003,
Bakker 2004, Dietze and Dietrich 2012), but are naturally limited in their spatial distribution. Also, the invasive nature of borehole methods is a potential limitation.

Geophysical exploration methods are uniquely positioned to overcome some of the above limitations, as they allow for non-invasive characterization of a landform’s internal structures in multiple dimensions (Yoshikawa et al 2006, Van Dam 2012, Parkes et al 2013, Chwatal et al 2015). Two of the most frequently used geophysical methods for the study of glacial landforms and deformation structures include ground-penetrating radar (GPR) and seismic reflection. GPR has been successful for imaging glacial deformation structures (Lønne and Lauritsen 1996, Hansen et al 1997, Busby and Merritt 1999, Jakobsen and Overgaard 2002, Bakker and van der Meer 2003, Bennett et al 2004, Sadura et al 2006), but the method does have some limitations related to penetration depth and a tradeoff between image resolution and penetration depth related to signal frequency. Seismic reflection methods do not have the penetration depth limitation and have been successfully for studies of glacial landforms and glaciotectonic deformation in on- and offshore settings (e.g. Williams et al 2001, Larsen and Andersen 2005, Rüther et al 2011, Höyer et al 2013).

In recent years, electrical resistivity (ER) methods, which are relatively low-cost compared to seismic methods and typically have a larger imaging depth than GPR, have become increasingly popular, and have numerous applications in landform studies and geomorphology, including for the characterization of landslides, sedimentary environments, and permafrost (Van Dam 2012). Most work on glacial landforms has focused on moraines (e.g. Parkes et al 2009, Kristensen et al 2009a). The number of examples in the literature of the use of ER for the imaging of glaciotectonic deformation is still very limited (e.g. Kristensen et al 2009b). However, previous work has demonstrated the ability of electrical resistivity tomography (ERT) to effectively characterize subsurface structures with a deformed or near-vertical nature (Smith and Sjøgren 2006, Beauvais et al 2007, Robineau et al 2007, Van Dam et al 2009).

In this paper, we discuss the use of traditional (soundings, traverses) and current tomographic methods for the characterization of large-scale glaciotectonic deformation. Few examples of such applications currently exist in published literature. We present data from the Ludington Ridge in Michigan, USA, where large-scale deformation has been identified in an approximately 60 m high exposed bluff face along the shore of Lake Michigan (Larson et al 2003). The first objective of this work was to assess the use of different electrical methods for such studies. To this end, we used vertical electrical sounding (VES) to study the general stratigraphy on the Ludington Ridge. These soundings were located away from the main deformation structures to limit the effect of lateral inhomogeneities and equivalences. Assisted by laboratory-derived petrophysical data of the texture-dependent relationship between resistivity and water content, the VES results were compared with available outcrop and drilling information. The second objective of this work was to assess whether the deformation structures observed in the bluff face (Larson et al 2003) are isolated structures (diapirs) or continuous transverse ridges, and if the latter, to identify their orientation. To identify the inland extent and orientation of the structures, we collected a series of constant-spread traverses (CST) and multi-electrode ERT transects mostly parallel to the bluff and coast line. Finally, we present some general considerations for the application of galvanic electrical geophysical methods to study large-scale sediment deformation.

Methods

Electrical resistivity

In the galvanic source ER method, a low-frequency alternating current of known strength is applied to the earth using a pair of electrodes that have been inserted into the ground, while two separate electrodes measure the potential difference over the circuit. Using Ohm’s Law and the electrode geometry it is then possible to calculate the resistivity. The electrical resistivity of a material depends on textural properties, surface conductance effects of clays, water saturation, fluid ion concentration, and temperature (Revil and Glover 1998, Rey and Jongmans 2007, Jayawickreme et al 2010). Because subsurface material is not homogeneous (e.g. layers, inclusions), and the exact current density through each of these is unknown, the measured quantity is expressed as the apparent resistivity ($\rho_a$). Electrodes can be deployed in an infinite number of ways, but a limited set of standard configurations with electrodes along a straight line (e.g. wenner, schlumberger, dipole–dipole) are typically used for data collection. The depth of investigation of the various array types is related not only to electrode separation but also to the subsurface resistivity distribution. Resolution decreases with increasing depth of exploration (Szalai et al 2009).

VES were traditionally used to obtain 1D information. Lateral profiling, also called constant-spread traversing (CST) was used to obtain information on lateral variability. Thanks to advances in equipment and data processing approaches, ERT is now often used to image the subsurface in two and three-dimensions. In a VES measurement only four electrodes are used. After each measurement, the electrodes are repositioned at an increasing distance from a central midpoint. The apparent resistivity of subsequent measurements will be representative of a larger volume of the subsurface and have a greater depth of investigation. A CST survey also uses only four electrodes but maintains an identical configuration. After each measurement, all four electrodes are moved along a line, thus creating a dataset that characterizes lateral variations in apparent resistivity. Most modern field equipment used for ERT measurements relies on multiplexers or electronically switching (active) electrodes to collect data for numerous electrode combinations along an array of electrodes. Multi-channel receivers can speed up data collection for array types where multiple potential measurements are made for a single current electrode pair. ERT data collected along 2D transects are typically plotted in pseudosections, where image points correspond to the lateral position of the electrodes and to the approximate depth of investigation (Dahlin and Zhou 2004).
Interpretation of both VES and ERT data requires inversion of the apparent resistivity values, where through an iterative process a model is calculated that best fits the measurements (Pelton et al. 1978, Loke and Barker 1996). Inversion of VES measurements is performed under the simplified assumption that the subsurface is layered, with each layer of constant thickness and resistivity and of infinite lateral extent. Ideally, VES measurements are conducted on level ground (Pous et al. 1996). CST apparent resistivity data are typically not interpreted in terms of true resistivity as no depth related information is available. Interpretation of ERT data typically occurs through inversion on a finite difference mesh, with a starting model that can be based on the average apparent resistivity in the data set or on \textit{a priori} information obtained from borehole or other geophysical data. To choose between multiple models that fit the data in VES and ERT inversions (equivalence), it is common to use smoothness constraints. These ensure that the resulting model explains the observed values to a level that is acceptable, while simultaneously avoiding large spikes in the modeled resistivity values. Constraints can also be placed on the ‘hardness’ of \textit{a priori} information. To interpret the final model in terms of the properties of interest (e.g. texture, water content), site-specific petrophysical relationships may need to be adopted (Jayawickreme et al. 2010, Hadzick et al. 2011).

\textbf{Site description}

The work presented here focuses on an area known as the Ludington Ridge, south of the town of Ludington, Michigan, which is an elongated NW–SE extending glacial landform of considerable height (figure 1). Stratigraphic units in the ridge include young eolian dune sand, glacial till and glaciolacustrine sediments. The sediment of interest is a so-called ‘lower stratified sand’ (LSS; Larson et al. 2003), which can be over 60 m thick, and is locally overlain by eolian sand. It consists primarily of horizontally and cross-bedded sands (LSS-sand) locally interbedded by beds of clayey silt. Most of these beds are less than a couple of meters thick. At the base of the LSS is a bed of clayey silt (LSS-fines) at least 4 m thick. Evidence for a subaqueous origin of the LSS includes drop stones and load structures (Larson et al. 2003).

The bluff face contains several large-scale deformation structures (Larson et al. 2003) comprised of several broad synclines and narrow anticlines (figure 2). Differential weathering has exposed the core of the anticlines as a series of diapir-like

\textbf{Figure 1.} Setting of the Ludington Ridge along Lake Michigan. Elevations are given in meters above 2000–2009 average lake level, which was 176 m above the International Great Lakes Datum 1985 at the Ludington NOAA gauge #9087023. During the study by Larson et al. (2003) from which this figure has been adapted, lake levels were ~1 m higher. The black outlined area refers to the map in figure 5.

\textbf{Figure 2.} Cross section illustrating large-scale glaciotectonic deformation observed in the ~60 m high bluff face. Refer to figure 5 for the location of the cross section. The vertical axis gives height in meters above lake level (h.a.l.). Eolian sand is not shown. Adapted from Larson et al. (2003).
structures (figure 3). According to Larson et al (2003) the structures formed as a result of glacial loading during an ice re-advance to the Port Huron moraine system. Primary sedimentary structures pre-dating the deformation are present in most of the folded layers but are largely absent in the core of the narrow of the anticlinal structures. Larson et al (2003) observed that the outcropping anticlinal structures have crests trending east–northeast, and suggest they form wall-like bodies that separate several broad structural depressions.

The structures observed in the bluff face are unique for the Great Lakes region. Glaciotectonic deformation has been documented in several studies around the Great Lakes region (e.g. Weaver and Arnaud 2011), but none at the scale observed in the Ludington Ridge. According to Larson et al (2003) the structures are locally reflected in the surface topography of Ludington Ridge and the location of springs along the bluff face. Despite the stratigraphy and deformation structures exposed in the bluff, the scarcity of wells inland and limited surface expression of possible below-surface deformation structures limits the ability to assess whether the structures are isolated (e.g. diapirs) or continuous features extending well into the Ludington Ridge.

Survey design

Field data were collected in VES, CST, and ERT modes (figure 4), all in Wenner configuration. In this configuration, the distance \( a \) between each of the 4 electrodes is equal, for a total array length of \( 3a \), and the potential electrodes are always placed in between the two current electrodes. The Wenner configuration offers the largest depth sensitivity of the common quadripole arrays. Arrays with infinity electrodes have larger depth sensitivity, but were impractical for this study. Alternative approaches to characterize lateral resistivity variations, such as azimuthal measurements, require a more complicated field setup (Morris et al 1997, Watson and Barker 1999), and were not used in this study. To check data consistency, repeat measurements were collected for all survey modes. Despite relatively high contact resistances in the dry sandy surface material, the collected resistivity data were of high overall quality. Data were collected along paths parallel and perpendicular to the bluff face (figure 5) using an AGI Supersting (R8/IP) earth resistivity system, manufactured by Advanced Geosciences, Inc. For VES and CST measurements, the instrument was connected via 18-gauge wire to stainless steel electrodes that were planted an average of 20 cm into the ground. The 18-gauge heavily shielded wire ensured sufficient durability in the sometimes rugged environment.

VES data were collected at four locations along the flanks of the bluff face where deformation structures were expected to be minimal and surface topography variations were small. All four VES arrays were oriented north-south, parallel to the bluff face. VES1-3 were located on the top of the ridge; VES4 was located on the beach directly below the large diapirs in the bluff. The mid-point of each VES is shown in figure 5. As resolution decreases with depth (Szalai et al 2009), the electrode separation was increased logarithmically. The \( a \)-spacings for each VES ranged from 3 m to 95 m.

At each location, two CST data were collected using a set of 6 electrodes, where electrodes 2 and 5 were used as both current and potential electrodes. First, electrodes 1, 2, 5, and 6 were used for a large-scale measurement. Electrodes 2, 3, 4, and 5 where then used for a small-scale measurement with an \( a \)-spacing three times smaller than the large-scale measurement. After each measurement, the array was moved in increments equal to the half the \( a \)-spacing (45 m for CST2–6). This allowed for frequent re-use of already planted electrodes. The station spacing was reduced in areas of interest. Array lengths \( (3 \times a) \) for the large- and small-scale measurements on the beach (CST1, figure 5) were 225 m and 75 m, respectively. Array lengths \( (3 \times a) \) for the large- and small-scale measurements on the bluff top (CST2–6, figure 5) were 270 m and 90 m, respectively. Array lengths were chosen based on observations in VES and ERT data to give both a good depth...
penetration and adequate characterization of shallower features. Following measurement, the apparent resistivity values were plotted as a function of distance along the profile.

For the ERT measurements we used 6 multi-core cables (26-gauge wire) with 14 take-outs in each, allowing for roll-along surveys. The cables and electrodes were connected to the ER instrument via an 84-channel switch box. Electrodes were placed 6 m apart for a total line length (before roll-along) of 498 m. Data were collected along three lines, two parallel to the bluff and one perpendicular (figure 5). Total line lengths varied from 498 m to 1086 m, requiring 7 roll-along procedures for ERT2.

Figure 4. Principle of Wenner-array resistivity data collection. The distance, $a$, between each of the four electrodes is always equal. In ERT data collection, image points (represented by circles) are plotted as a function of both electrode separation (−image depth) and position. The two solid image points correspond to the electrode layouts at the surface, for electrode separations of $n = 1$ and $n = 3$. CST data collection focuses on data for a constant array length. VES collects data centered on a mid point; electrode separations are typically increased logarithmically due to decreased resolution with depth. ERT and VES apparent resistivity data are converted into models of subsurface resistivity through inversion.

Figure 5. Aerial photo of the field area with locations of the narrow anticlinal clay structures in the bluff face (figure 2), geophysical transects, and resistivity soundings indicated. For results of geophysical transects not included here (ERT1, CST6), see Aylsworth (2008).

Figure 6. Lab-derived petrophysical relationships between volumetric water content and electrical resistivity for different textural units identified in the bluff.

Results

Petrophysics

Grain-size analysis of the main stratigraphic units (figure 2) was performed by a combination of sieve analyses for the sand size fraction and hydrometer tests for the silt and clay size fractions, following standard procedures. Grain size distribution curves indicate a very high contrast between the sand and fine-grained beds. The eolian sand had 95.4% sand-size particles, with no clay. The horizontally and cross-bedded units in the LSS (LSS-sand) had 99.6% poorly sorted sand-size particles, with no clay. The lowest fine-grained layer (LSS-fines), comprising the core of the anticlinal structures, consists of 69.9% silt and 30.1% clay. The two upper fine-grained layers (Interbedded-fines) consist of about 75% silt and 25% clay.

The minimum expected amount of pore water for these samples, defined as the residual volumetric water content ($\theta_r$) or water content after drainage by gravity, was estimated using the Rosetta model (Schaap et al 2001) as 0.055 cm$^3$ cm$^{-3}$ for both sand units (eolian sand and the horizontally and cross-bedded units in the LSS-sand layer) and 0.088 cm$^3$ cm$^{-3}$ for...
the LSS-fines layer. Although $\theta_r$ represents a minimum field value, it is a likely reasonable approximation for the sandy units as their larger pore diameters allow quick drainage. In and just above the LSS-fines layer, which may act as a perching unit resisting percolation, the water content may be higher than the $\theta_r$ calculation indicates.

To obtain starting values for resistivity inversions, we conducted laboratory experiments with sediment samples collected from the bluff face. Sample material was first oven dried to remove interstitial moisture and placed in a rectangular box with terminals for current transmission at the short ends. Resistance data were collected using two metal pins through the sample material near the center of the box. Repeat resistance measurements were made after thoroughly mixing the sample with deionized water at incremental steps equal to around $2.5-5\%$ of the box volume. Temperature was recorded before each measurement to correct for its influence on soil conductivity and the box was weighed to obtain exact water content for each measurement step (Jayawickreme et al. 2010). The obtained relationships between soil water content and electrical resistivity are linear in the log–log domain (figure 6). Although this approach does not explicitly quantify the effects of possible surface conductance effects (e.g. Robineau et al. 2007), these data and the minimum likely water contents obtained previously improve interpretation of inversion results.

**Synthetic modeling**

To understand the effects of a low-resistivity clay-rich structure protruding from below into a high-resistivity sandy formation, the apparent resistivity distribution for a series of simple models was calculated (figure 7). The dimensions were chosen so that the model and results would be easily scalable. Depth to the interface between the sand and the clayey silt (LSS-sand / LSS-fines) was held constant at 60 m (approximate bluff height). Specific attention was given to the effects of anomaly height and width and to the array $a$-spacing (30 and 90 m) that were used during CST measurements in the field. The resistivity of the sand and clay was set at 2400 and 40 $\Omega\text{m}$, respectively, which match the expected values based on calculations in Rosetta and petrophysical data (figure 6). As the models are not intended to be a direct replica of the field conditions, the effect of surface topography was excluded. Apparent resistivity values were calculated using RES2DMOD software (Loke 2002) for a Wenner array, similar to what was used in the field measurements.
The modeling results are presented using a series of graphs of apparent resistivity versus position. Near the edges of the model domain, without the effects of the disturbance, apparent resistivity has an inverse correlation with a-spacing. As a-spacing increases, the cumulative effect of the deeper low-resistivity layer becomes larger and the apparent resistivity drops (figures 7(b) and (c)). The results of the synthetic modeling also show that, for any given dimension (height or width) of the structure, the lateral range (distance along x-axis) over which the disturbance affects the apparent resistivity, increases with a-spacing (figures 7(b) and (c)). Simultaneously, the total range of apparent resistivity values ($\rho_{\text{max}} - \rho_{\text{min}}$) decreases with an increase in a-spacing (figures 7(b) and (c)). This has implications for the detectability of anomalies, especially when considering natural variability (anomaly shape, surface topography, and material properties) and noise not included in these simplified models. The effect of anomaly height on resistivity results is shown in figures 7(d) and (e). An increased height of the structure (structure approaching or at the surface; figure 7(e)), leads to sharper lateral changes in apparent resistivity. Also, the total range of apparent resistivity values ($\rho_{\text{max}} - \rho_{\text{min}}$) is lower for an increased depth of the anomaly below the surface (figure 7(d)). All results in figure 7 show the strong effect of anomaly width on resistivity variations, including the presence of a double resistivity low when the width of the anomaly is less than or approximately equal to the a-spacing.

It is important to note that these results have been obtained for simplified subsurface resistivity models that do not include the effects of topography, a more natural shape of the anomaly without right angles, and internal resistivity variations. The modeling also does not take into account the 3D nature of the subsurface structures and the effect of survey lines that cross subsurface structures at non-normal angles. Nevertheless, these results highlight the capabilities and potential limitations of the resistivity method for characterizing subsurface structures with resistivity contrasts as expected at this site.

**Electrical soundings**

The VES data were inverted using DCINV software (Pirttijarvi 2005) to generate best-fit resistivity models. The optimization algorithm is based on a linearized inversion method. Adaptive damping is used to minimize the difference between the field data and the results computed for the model. VES data from all four locations could be reasonably fit with a 3-layer Earth model. Inversions using four- and more layer models further reduce the root mean square (RMS) error, but require unrealistic resistivity values.

For VES1 and VES2, located on South Lakeshore Drive near the bluff (figure 5), the best-fit solution gives a low-high-low sequence for the resistivity values of the three model layers (figure 8). The first layer, with a thickness of just over 4 m, has a resistivity around 100 $\Omega$m, and can be interpreted as the eolian sand, which has a lower resistivity than the LSS-sand (figure 6). The second layer, interpreted as the LSS-sand unit, varies in thickness between 25 and 40 m and has a resistivity of around 2500 $\Omega$m, similar to the expected value based on petrophysical measurements (figure 6). The deepest layer has a resistivity of approximately 18 $\Omega$m, and represents the LSS-fines near the bottom of the bluff (figure 2). The RMS error is near 6% for inversions of both VES1 and VES2 data. The VES3 data were collected a significant distance inland from the bluff. The best fit model for the VES3 data has a similar character to VES1 and 2, but the thickness and resistivity of the first and second layer, respectively, is notably smaller (figure 8). Differences between individual inversion results can be due to absence of eolian sands (VES3) and locally different properties (e.g. texture, thickness, and resistivity) of the other layers.

The VES data were collected in areas without significant amounts of known deformation. The inversion results reveal a high resistivity layer (dry, clay-free sand) several 10’s of meters thick, underlain by low-resistivity material, confirming the general stratigraphic model for the ridge (figure 2). The agreement between VES interpretations and bluff stratigraphy highlights the potential usefulness of ER methods at this location. The best-fit inversion for VES4 shows a shallow top layer with high resistivity, which represents the unsaturated beach sand, underlain by two layers with decreasing resistivity.

**Figure 8.** Resistivity sounding results for (a) VES1, (b) VES2, and (c) VES3. See figure 5 for locations on the Ludington Ridge. The plots give the measurement results (circles) and the computed best-fit curve for a three-layer model. Model properties and the RMS error between model and data are given in table insets.
The collected CST data show that one of the anticlinal structures has a clear inland extent. The lateral position of the drop in apparent resistivity associated with the second deformation structure at 600 m is gradually shifting north further away from the bluff (figure 10). The orientation of this structure below the ridge can thus be interpreted as WSW to ENE. The northernmost deformation structure is clearly visible in CST2 (at 1150 m), but less so in CST5 (at 1200 m). The shift northward is also less significant, suggesting a more W to E orientation. The absence of strong dips in resistivity for the third and fourth (northernmost) deformation structures in CST5, suggests that with increasing distance from the bluff these structures experience changes in their textural (and thus electrical) properties and/or dimensions. Similar observations were made in the multi-electrode surveys, discussed in the next section. Comparison of the CST data with the apparent resistivity values for comparable array lengths show very similar results.

**Multi-electrode surveys**

ERT data were inverted in EarthImager2D software to obtain best-fit models of the subsurface resistivity distribution. Data points with repeat errors larger than 3% were omitted from the inversions. This criterion applied to an average of 0.3% of points per data set. The initial inversion step involved calculating a forward model based on the pseudosection resistivity distribution, using a finite-difference mesh increasing by 10% for each deeper layer. The width of each cell was set to the electrode spacing. Eight padding cells were added on either side of the line and to the bottom of the mesh. An iterative Occam l2-norm robust inversion (Constable et al 1987) was used with a topographic correction to account for elevation changes along the profile. The investigation depth was limited by the roll-along pattern. By moving 14 electrodes in the roll-along process, data coverage for the largest a-spacings (greatest imaging depths) was incomplete. The effective maximum array length was 414 m (69 electrodes × 6 m). Topography was included in the inversions via a damped transform of the modeling mesh.

The inversion models for the ERT data show resistivity values varying widely on account of the significant textural differences (figure 11). Away from the deformation structures the resistivity gradually decreases with depth. This may be due to an increase in water content with depth in the sand or the presence of interbedded fines (figure 2). The array length was not large enough to clearly capture the transition into the LSS-fines at depth, but the gradual resistivity decrease with depth may be related to this transition. Only one of the clay structures is clearly visible in the data (figure 11). This structure in figure 11(a) is equivalent to the low-resistivity anomaly identified in CST3-5 (figure 10), although the positions do not exactly match. The low-resistivity subsurface structure extends to transect ERT3 (figure 11(b)) where it is located up to 450 m from the bluff edge.
The 2D ERT inversion results are affected by the complex 3D nature of the system, including the subsurface texture and resistivity distribution and the surface topography (Holcombe and Jiracek 1984). In our inversions it has been assumed that both vary only in the transect-parallel direction, and are constant in the perpendicular direction. This simplification of the real conditions results in near-surface resistivity gradients that are not representative of the true distribution and can cause inversion errors. The data quality may also have been impacted by the higher contact resistances between the electrodes and the resistive sandy material. Inversion models with RMS errors as low as 4% had a significant number of inversion artifacts. The results shown in figure 11 used more smoothing constraints and have a RMS errors of around 15%. The effect of the surface topography and the 3D nature of the subsurface structures could potentially be addressed by using a full 3D data collection and inversion approach. However, the added time- and cost requirements would likely outweigh the potential benefits.

Conclusions
This paper presents one of the first investigations of large-scale glaciotectonic deformation using ER methods. It compares constant spread traverses (CST) and multi-electrode ERT surveys collected in an area adjacent to a bluff with known deformation structures. Two sandy units that are over 60 m thick in places are underlain by a clayey unit from which several apparent diapirs protrude.

Electrical resistivity was chosen as the method to investigate this system as it is able to exploit the large resistivity contrast between the two primary textures visible in the nearby
Ludington Ridge bluff exposure. Petrophysical properties of these sediments were determined from lab-scale measurements of the relationship between water content and current resistance. These petrophysical relationships, in combination with modeling of water retention, confirmed the large electrical contrast of the main sedimentary units. The values thus obtained were then used for forward modeling of the electrical resistivity response.

Forward modeling of the CST response for a simplified dyke model, with a low resistivity feature protruding into a high resistivity unit of height $h$, illustrates the applicability of the resistivity method at this site. The model results show the strong effects of array length and variable feature dimensions on the response. As expected, the signal response declines with lower height of the feature and increased depth below the surface. The modeling results also show that for large array lengths, anomalous features of around half the height of the unit in which it protrudes have low contrast and thus detectability. The synthetic modeling was effectively used to understand the basic ability of the method to characterize the structures of interest, despite the fact that the model was not a perfect replica of the site’s conditions.

VES measurements that were conducted away from the main deformation structures confirm the strength of the ER method for defining the main sedimentary units at this site. In the central area of the site with four large glaciogenic deformation structures, a series of parallel CST and ERT lines were measured. Of the four structures observed in the bluff face, one was clearly detected and imaged by both the CST and ERT resistivity measurements. The resistivity values in the parallel profiles CST3-5 identified a WSW to ENE trend. This observation is in general agreement with earlier observations of the antinclinal structures in the bluff face and confirms that this structure is a ridge rather than a diapir. The northernmost structure was clearly identified in profile CST2, closest to the bluff. However, it was only faintly visible in the data collected further from the bluff along South Lakeshore drive (CST5 and ERT2), suggesting that it had disappeared, or that its dimensions or properties kept it below the detection limit.

Overall, the CST and ERT measurements contributed to a much improved understanding of the extent of the subsurface structures beyond the bluff face. Nevertheless, the steep surface topography and complex subsurface structures, resulted in challenging conditions for a full characterization using 2D methods. A more complete 3D investigation is possible but requires a larger investment of time and equipment.

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