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Geospatial analysis of controls on subglacial bedform morphometry in the New York Drumlin Field – implications for Laurentide Ice Sheet dynamics

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ABSTRACT: The outline and trend of 6566 subglacial bedforms in the New York Drumlin Field have been digitized from digital elevation data. A spatial predictive model has been used to extend values of bedform elongation over an area measuring 200 km × 110 km. The resulting surface is used in conjunction with depth-to-bedrock data and an assumed duration of ice residence to test three proposed controls on bedform elongation. Upon comparison, the resulting display of morphometry is best explained by differences in ice velocity across the field of study. The existence of multiple zones of fast-moving ice located along the southern margin of the Laurentide Ice Sheet is implied by the observed patterns of bedform elongation and orientation. We present two interpretations that are consistent with the observations. First, enhanced basal sliding caused by decreasing effective pressure near a calving margin is suggested as a possible mechanism by which localized fast ice flow is initiated and maintained. Second, topographically controlled ice streams likely occupied the fjord-like troughs of the Appalachian Upland northern rim. Contrary to previous understanding of the Laurentide southern margin in New York State, the resulting palaeoglaciological reconstruction illustrates a dynamic mosaic of ice stream and/or outlet glacier activity. Copyright © 2009 John Wiley & Sons, Ltd.

KEYWORDS: geospatial analysis; subglacial bedform morphometry; New York Drumlin Field; Laurentide Ice Sheet

Introduction

The use of satellite imagery and digital elevation data to reconstruct former ice sheet dynamics from geomorphologic evidence has become more common over the past decade (Knight, 1997; Clark, 1997; Smith and Clark, 2005; Wellner *et al*., 2006; Briner, 2007; Napieralski, 2007). In particular, reconstructing the location of former ice streams has received increased attention due to their role in ice sheet response to climate change (Oppenheimer, 1998; Bennett, 2003; Rignot and Kanagaratnam, 2006). The prediction of future ice sheet behaviour will only result from improvements in numerical ice sheet modelling, remote sensing techniques and platforms, understanding of subglacial processes, and palaeoglaciological reconstructions (Vaughan *et al*., 2007). Studies that utilize a glaciological inversion scheme, i.e. deciphering ice sheet dynamics by investigating the landform record (De Angelis and Kleman, 2007), are a critical component of this synergistic undertaking.

The southern lobes of the Laurentide Ice Sheet (LIS) that flowed over New York State have been interpreted to be dynamic outlets that transported large quantities of ice and sediment from the ice sheet interior to periphery (Jennings, 2006). The Ontario Lobe was a sector of the LIS that crossed the Lake Ontario basin (Figure 1) and influenced the landscape of western and central New York State (Mickelson *et al*., 1983). Several studies have characterized ice flow within the Ontario Lobe as radial spreading from a dome located in the Lake Ontario basin (e.g. Ridky and Bindschadler, 1990; Hart 1999). This observation has been based on the orientation of bedforms in the New York Drumlin Field located south of Lake Ontario.

A growing body of literature supports the idea that subglacial bedforms serve as palaeoflow indicators and provide crucial information regarding ice sheet behaviour (e.g. Clark, 1994; Hart, 1999; Stokes and Clark, 2002, Sejrup *et al*., 2003). The drumlins and megaflutes that compose the New York Drumlin Field (generally referred to as subglacial bedforms in this study) display considerable variation in size, shape, and orientation. The processes responsible for the formation of these bedforms have remained elusive over decades of active research (Menzies, 1987; Boyce and Eyles, 1991; Shaw, 2002). Similar to the lack of consensus regarding formation processes of subglacial bedforms, the factors that control their morphometry remain enigmatic. Several studies have identified the velocity of overriding ice as the primary control on bedform elongation (Chorley, 1959; Miller, 1972; Stokes and Clark, 2002; Briner, 2007). Other studies have concluded that the time available for bedform creation and modification along with constancy

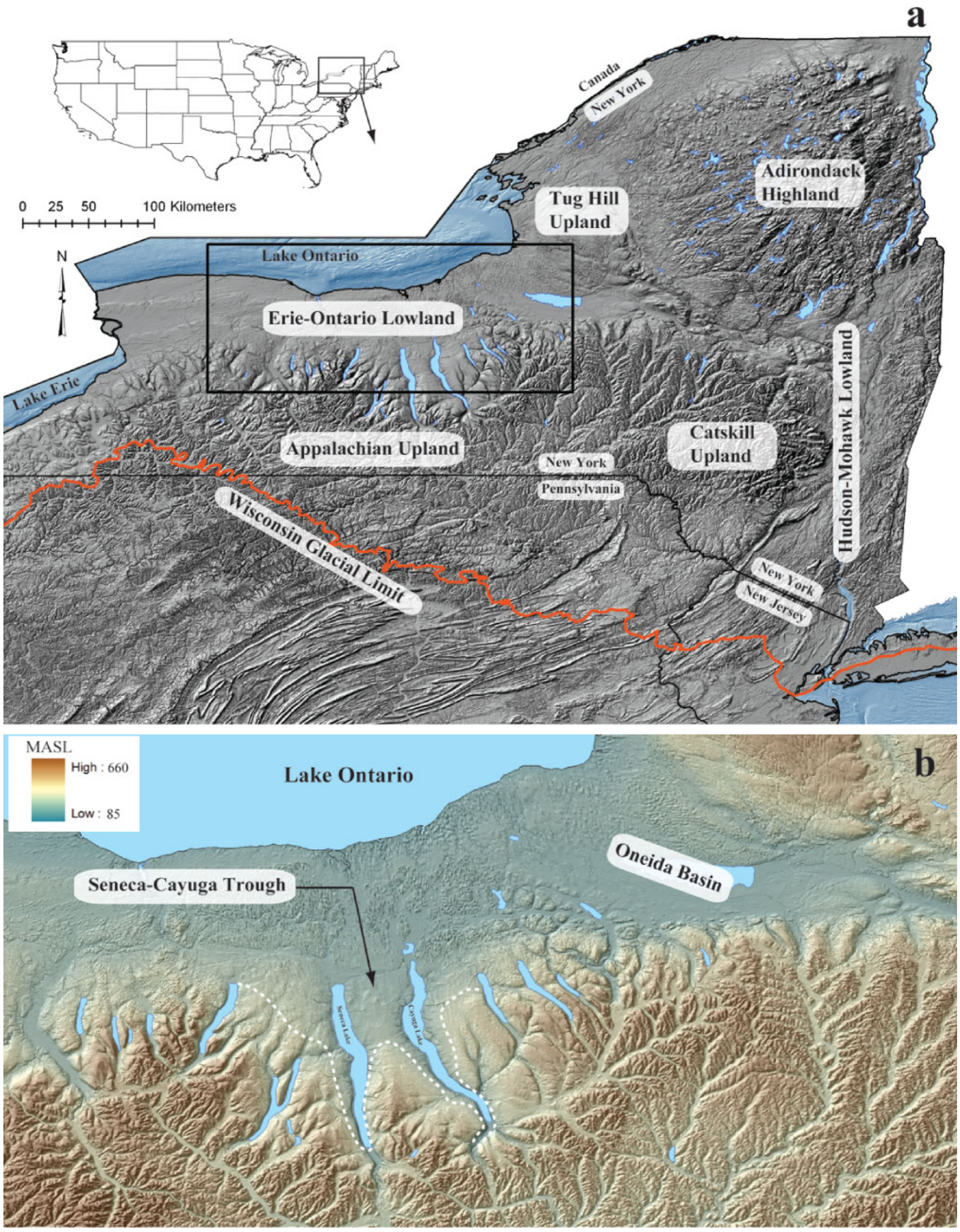

Figure 1. a) Hillside map showing physiographic provinces of New York State (Isachsen *et al.*, 2000). Outline of New York Drumlin Field indicated by bold rectangle. b) Elevation map showing the topography of the area. Approximate palaeoshoreline of Lake Newberry indicated by white dashed line. This figure is available in colour online at www.interscience.wiley.com/journal/espl

in ice flow direction are critical to determining the resulting shape (Gravenor, 1953; Mills, 1980; Boyce and Eyles, 1991). In addition, recent work hypothesized that the thickness of unconsolidated sediments upon which bedforms reside is the primary control on their morphometry (Kerr and Eyles, 2007). The fundamental contribution of this paper is a geospatial comparison of these controls in the New York Drumlin Field. Unlike several previous studies in the region that focused on subsets of the field (e.g. Stahman, 1992; Briner, 2007), more than 6500 features have been investigated in this paper providing a comprehensive view of the late-Wisconsin Laurentide southern margin in central New York State.

Setting

New York State is composed of multiple provinces of dissimilar physiography that illustrate the varying influence of the LIS on the region (Figure 1). This study is focused in an area where the Erie-Ontario Lowland forms a transition to the Appalachian Upland in central and western New York State. The topography in this region is developed on primarily undeformed strata of Ordovician and Devonian age (Isachsen *et al.*, 2000). The Erie-Ontario Lowland resides on sedimentary bedrock that is susceptible to erosion and, as such, is relatively flat and low in elevation. As an exception, the resistant Niagara and Onondaga Escarpments, residing in the Erie-Ontario Lowland, are exposures of their respective southward-dipping formations. The northern rim of the Appalachian Upland is characterized by a mature fluvial architecture and several valleys that have been incised over multiple glaciations.

The LIS covered New York State during the last glacial maximum except for a small area near the southwest corner of the state (called the Salamanca Re-entrant) and the southern coast of Long Island (e.g. Cadwell and Muller, 2004). A number of end moraines (Figure 2) and ice-marginal lakes formed as the LIS retreated from central New York State (Fairchild, 1909;

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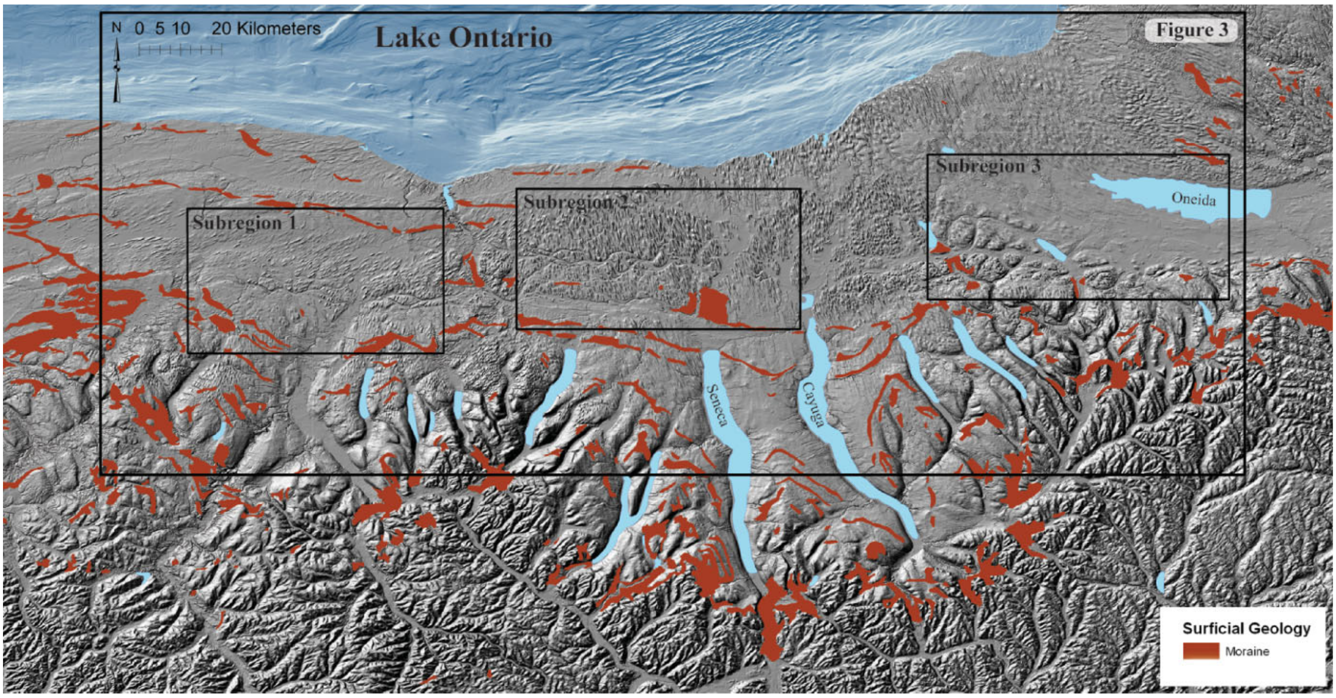

Figure 2. Hillside image showing New York Drumlin Field. Shaded polygons indicate moraines extracted from New York surficial geology data (Muller and Cadwell, 1986). This figure is available in colour online at www.interscience.wiley.com/journal/espl

Muller and Prest, 1985; Muller and Cadwell, 1986; Young and Burr, 2006); however, the chronology of LIS recession in central New York State is only broadly known (Muller and Calkin, 1993). Retreat from the Valley Heads Moraine, a complex of kame moraine features south of the Finger Lakes (Figure 2), began by 17·3 kyr BP (Muller and Calkin, 1993). Glacial Lake Iroquois, a high stand of present-day Lake Ontario, formed by 14·2 kyr BP when the LIS withdrew from central New York State but still blocked the St Lawrence River to the east (Muller and Prest, 1985).

Background

Previous studies of the New York Drumlin Field can be divided into two categories: studies of bedform shape, size, and distribution (morphometric), and investigations of internal sediment structure (stratigraphic). The first reference to New York drumlins in the literature dates to the mid-nineteenth century where the drumlin field is described as 'parallel ridges with a north–south orientation' (Hall, 1843). The early work of H. L. Fairchild (1907) laid the foundation for several descriptions of bedform pattern and shape (Slater, 1929; Miller, 1972; Francek, 1991). More recently, morphometric investigations within the New York Drumlin Field have been used as evidence for fast ice flow (Briner, 2007) and a systematic relationship between bedform shape and sediment thickness (Kerr and Eyles, 2007). Several studies have described the internal structure of drumlins where exposed at the southern shore of Lake Ontario with numerous observations of 'concentric banding' within the landform interior (Fairchild, 1929; Slater, 1929; Calkin and Muller, 1992; Hart, 1997). Till fabric analyses have been primarily limited to exposures along the Lake Ontario shoreline. The bulk of these studies imply drumlin formation via deformation of subglacial sediment (e.g. Dreger, 1994; Hart, 1997; Menzies *et al.*, 1997).

Methodology

Bedform elongation and orientation

Digital elevation data for the region were obtained from the United States Geological Survey National Elevation Dataset (USGS NED). The NED data for New York State were produced by the USGS via interpolation of digitized contours using a complex linear routine. Using 7·5 minute USGS topographic maps as a guide, the bounding break in slope for each of the 6566 bedforms was manually digitized from the digital elevation data in Environmental Systems Research Institute (ESRI) ArcGIS 9·2 using the editor utility. The length, width, and orientation of each bedform were measured and recorded in a geodatabase. Bedform elongation was determined by dividing length by width. An ordinary kriging technique was applied to the elongation data using a spherical semivariance model over a search area of 30 neighbours via the Geostatistical Analyst utility in ArcGIS. The resulting prediction surface was draped over the digital elevation data to illustrate the spatial distribution of bedform elongation (Figure 3).

Depth to bedrock

Three subregions (each characterized by a dense population of bedforms) were defined within the study area (Figure 2). The depth to bedrock at 434 wells was extracted from river basin planning reports covering each of the subregions. Subregion 1 resides within the Genesee River basin (Kammerer and Hobba, 1967), whereas Subregions 2 and 3 lie within the western (Kantrowitz, 1970) and eastern (Crain, 1974) Oswego River basins, respectively. For Subregions 1, 2, and 3 an ordinary kriging technique was applied to 61, 248, and 120 wells, respectively, using a spherical semivariance model to establish an isopach map of sediment thickness (Figure 4).

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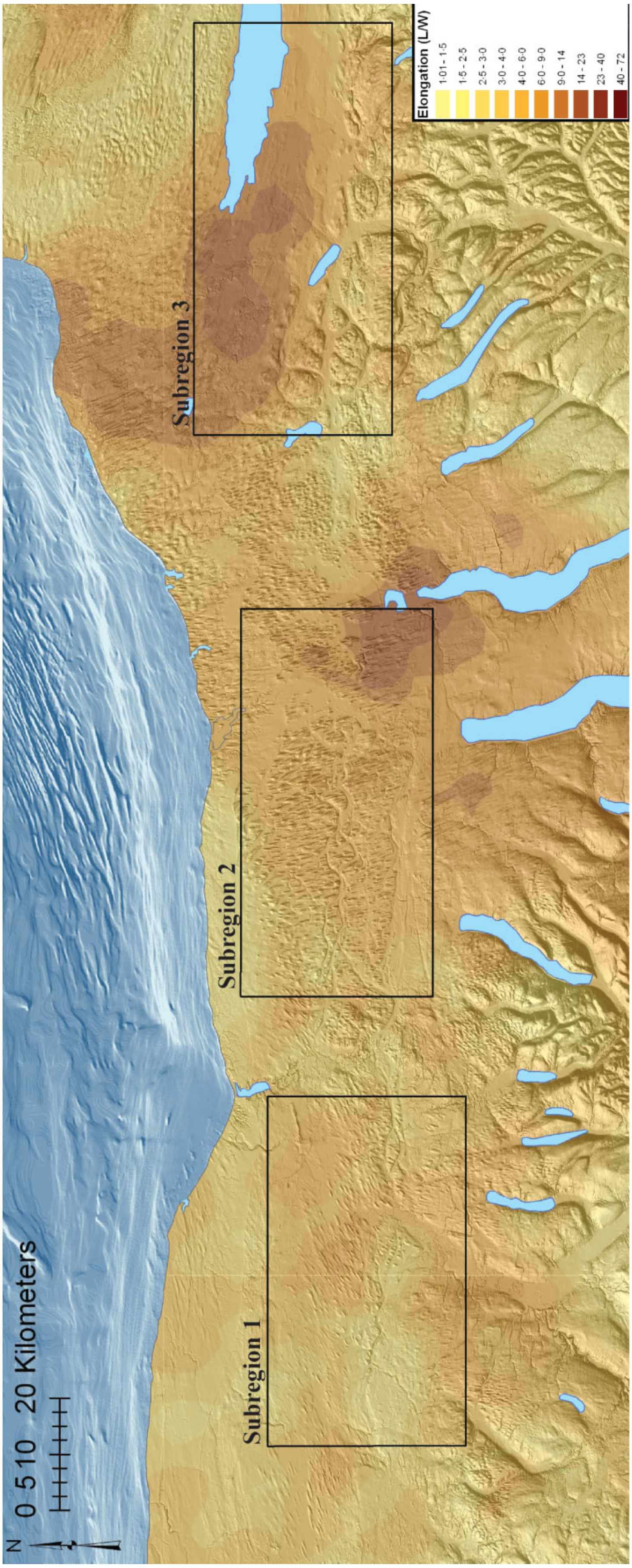

Figure 3. Prediction surface of bedform elongation. This figure is available in colour online at www.interscience.wiley.com/journal/espl

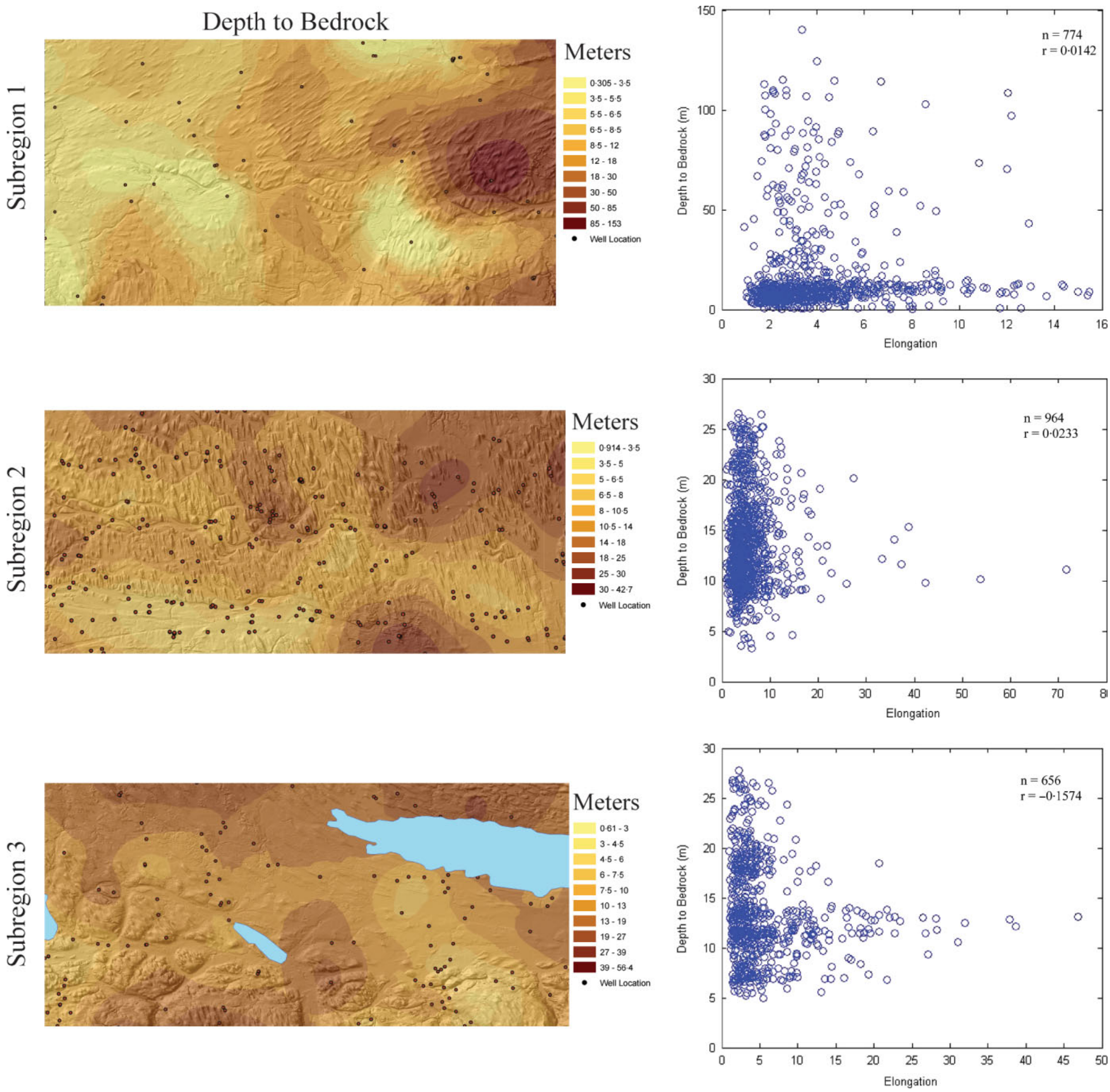

Figure 4. Prediction surfaces of depth to bedrock for three subregions. Bivariate scatter plots illustrate comparison of depth to bedrock beneath individual subglacial bedforms in each subregion (quantity = n) to associated elongation. Well locations are shown as points on each prediction surface. This figure is available in colour online at www.interscience.wiley.com/journal/espl

The number of neighbours used in each searching method was determined by using the value corresponding to the average standard error to root-mean-squared prediction error ratio closest to one.

Comparing elongation and depth to bedrock

Each isopach map was converted to raster format and a point was generated at the centroid of each polygon outlining the digitized bedforms. This reduced the polygon-shapefile to a point-shapefile that represented the collective attributes of the drumlin field. This point-shapefile was converted to raster format using the Point to Raster conversion utility included with the ESRI ArcToolbox with a value of one wherever a bedform was located and NODATA elsewhere. The ESRI Raster Calculator was used to multiply depth to bedrock by drumlin location thereby giving an output raster file with a cell value of depth to bedrock wherever a bedform was located, and NODATA everywhere else. This raster was converted to a point shapefile using the ESRI Raster to Point conversion utility. A spatial join was performed between the resulting point-shapefile and the original polygon-shapefile, thereby adding the corresponding depth to bedrock as an attribute to each digitized feature. Finally, the data were plotted on a bivariate scatterplot and a Pearson correlation coefficient was generated for each subregion.

Results

Bedform elongation

The minimum bedform elongation (1·01) was observed in the north-central portion of Subregion 2 and the maximum (71·58) near the southeast portion of Subregion 2 (Figure 3). The

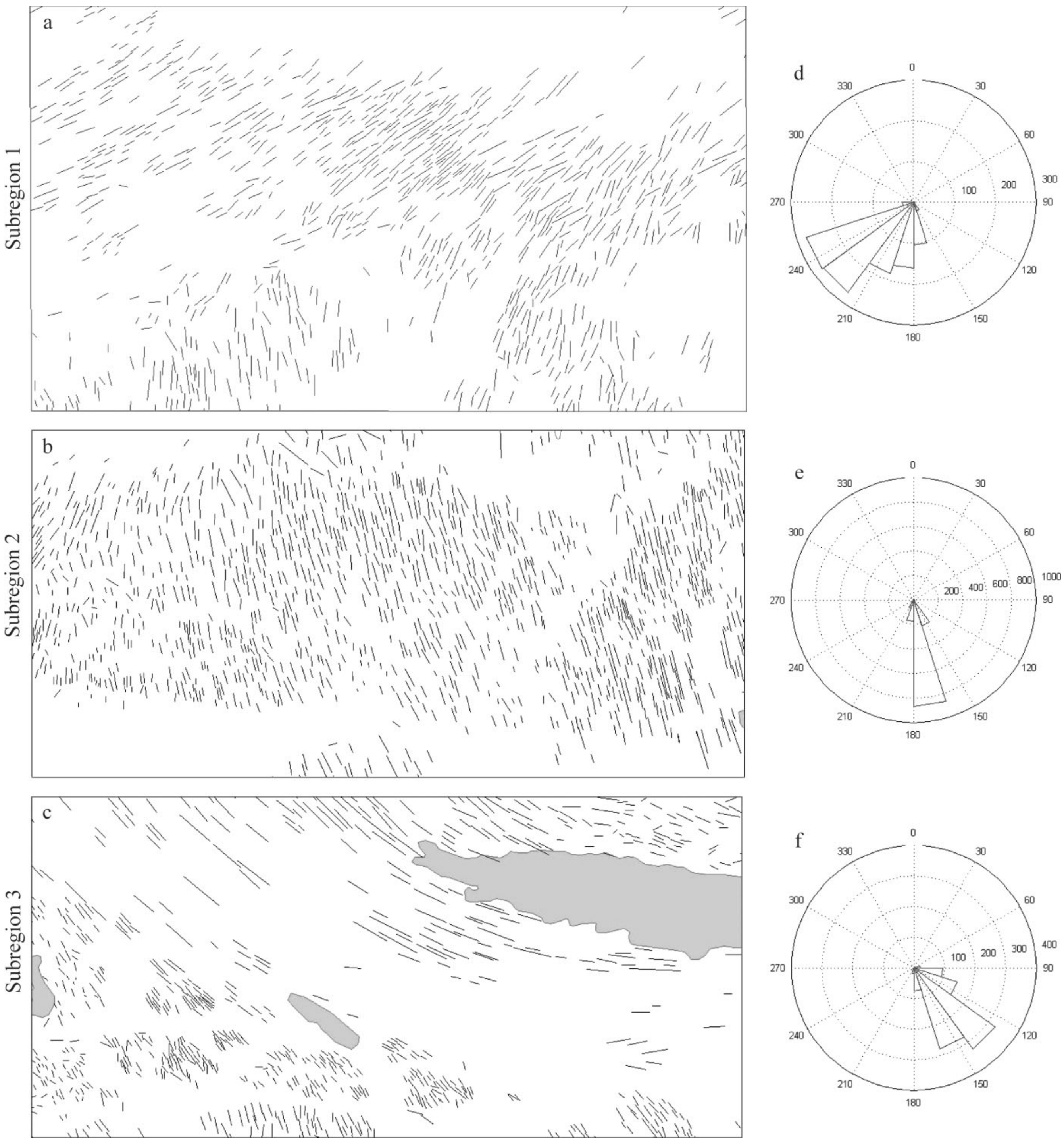

Figure 5. Bedform orientation (a, b, c) and orientation frequency (d, e, f) for 3 subregions.

majority of bedforms in the New York drumlin field classify as drumlins, although many can be defined as megaflutes (Benn and Evans, 1998). The most significant clusters of attenuated features are located north of Cayuga Lake in the lower-right portion of Subregion 2 and west of Oneida Lake in Subregion 3. Features with the lowest values of elongation are clustered near the shore of Lake Ontario, along the high ground north of Oneida Lake in Subregion 3, and to the west of Subregion 1.

The distribution of bedform elongation was spread among 10 classes about a mean elongation of 4·81. The standardized mean prediction error of 0·0004 indicates a prediction surface that is strongly developed upon measured values. The root-mean-square standardized prediction error of 1·07 indicates a slight underestimation in prediction surface variability; however, the results are sufficient for performing landscape-scale comparisons.

Bedform orientation

Each of the three subregions displays a unique pattern of bedform orientation (Figure 5). Those in Subregion 1 primarily display a southwest trend, although bedforms that are aligned north–south can be found in the southern portion of the area. The bedforms in Subregion 2 show the most consistent frequency of orientation with the vast majority of features

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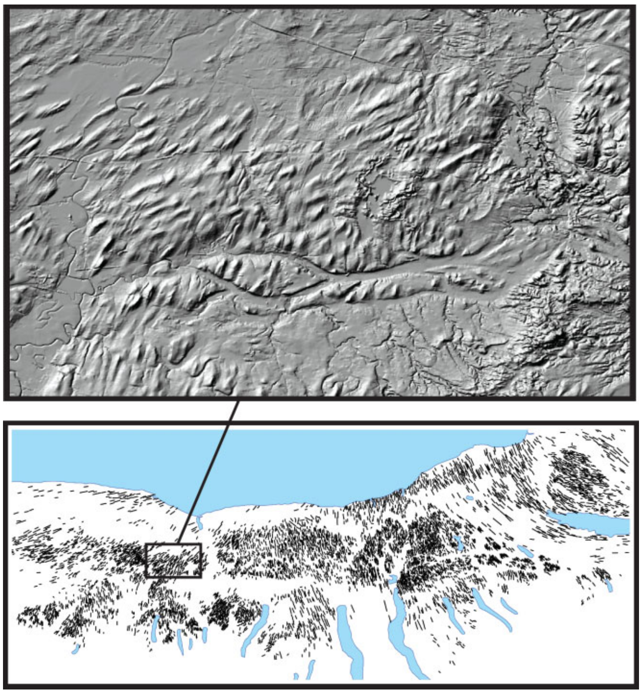

Figure 6. Hillshade image of suture zone between southwest trending bedforms in the west and north-south bedforms in central New York. This figure is available in colour online at www.interscience.wiley.com/journal/espl

aligned north–south. The distribution of orientation for those in Subregion 3 is bimodal with a strong southeast trend toward Oneida Lake and a scattered distribution of north–south oriented bedforms to the south. The zone of transition in bedform orientation from Subregion 1 to Subregion 2 displays drumlins oriented to the southeast with several features showing survival or significant modification from a north–south trend (Figure 6). The transition from north–south oriented bedforms in Subregion 2 to southeast in Subregion 3 is distinct.

Depth to bedrock

Depth to bedrock in Subregion 1 (Figure 4) varies from less than 1 m to more than 150 m with the thickest drift located near the eastern edge of the area. Development of the isopach surface for Subregion 1 utilized the fewest number of wells ($n = 61$). Correspondingly, the standardized mean prediction error (0·0836) and root-mean-square standardized error (1·56) are suggestive of a surface that is limited in its development from point values. This is further implied by concentric rings displayed in the eastern portion of the Subregion 1 isopach surface (Figure 4). In contrast, the distribution and density of point data for Subregions 2 and 3 are robust and yield standardized mean prediction errors of –0·0061 and –0·0018, respectively, with corresponding root-mean-square values of 0·94 and 1·19. A much smaller range in depth to bedrock is displayed in Subregion 2 with a maximum depth of 43 m occurring north of Seneca Lake. Depth to bedrock varies from less than 1 m to more than 56 m for Subregion 3. Each bivariate scatterplot illustrates a lack of correlation between depth to bedrock and bedform elongation (Figure 4). Corresponding Pearson correlation coefficients (r) for Subregions 1, 2, and 3 are 0·0142, 0·0233, and –0·1574, respectively.

Discussion

Controls on bedform morphometry

Multiple explanations have been proposed for the diverse pattern of bedform elongation observed in the New York Drumlin Field. The controls explored by this study can be classified into three groups: the thickness of sediment cover upon which the bedforms reside, the time available for the creation of subglacial bedforms, and the velocity of ice responsible for their formation. Sediment rheology, another potentially important factor (Piotrowski, 1987; Rattas and Piotrowski, 2003), is not investigated here due to a lack of representative data over the large area of study. Moreover, exposures that provide a view of subglacial bedform internal composition and structure are rare in the New York Drumlin Field. Maps of surficial geology are available for the area; however, their representation is limited to the near surface and does not reveal the internal characteristics of individual bedforms. Within the region explored by this paper, a previous study focused on the drumlins and megaflutes located north of Cayuga Lake (Stahman, 1992). The results of the project show little correlation between internal composition (till texture) and elongation in a zone where drumlins and megaflutes reside within close proximity.

Depth to bedrock

The well log data reveal a complex underlying topography with several features that are supported by independently-derived observations. Although additional data for Subregion 1 were not available to improve model performance, thick glacial drift near the eastern edge (Figure 4) (maximum 153 m) is suggestive of a buried valley. Indeed, this area is aligned with the Genesee Valley, a sediment-infilled bedrock trough similar in geometry to the Finger Lake valleys to the east (Young and Sirkin, 1994). Secondly, a relatively linear west–east trending group of relatively shallow depth to bedrock measurements in Subregion 2 coincides with the Onondaga Escarpment. Thirdly, depth to bedrock for the area located north of Seneca Lake in Subregion 2 is consistent with values obtained by seismic reflection for the area (Mullins *et al.*, 1996). These similarities among the interpolated isopach surfaces and ground-truth observations provide confidence in the isopach maps. We find little correlation between depth to bedrock and bedform elongation. The low Pearson correlation coefficient (r) shown in each bivariate scatterplot (Figure 4) indicates little systematic dependence of elongation on depth to bedrock.

Time available for bedform attenuation

The long-axis trend of subglacial bedforms records ice flow direction during the time of their formation. The New York Drumlin Field displays southwest oriented bedforms in the west (Subregion 1), north–south in the centre (Subregion 2), and southeast in the east (Subregion 3) (Figure 5). The zone of transition from southwest trending features in Subregion 1 to north–south in Subregion 2 is highly suggestive of redirection by subsequent ice flow (Figure 6). We do not suggest contemporaneous bedform genesis among the subregions due to this imprint of discontinuous ice flow direction (Clark, 1993). The dominant north–south direction displayed in Subregion 2 suggests similar ice flow direction during bedform formation. Given the relatively small size of Subregion 2 in comparison to the overall Ontario Lobe, the duration of ice

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residence can be considered spatially homogenous along an east–west transect. Therefore, the time available for landscape modification was similar across the subregion yet significant differences in bedform elongation are present (maximum 71·58 to minimum 1·01). In addition, elongation increases down flowline for Subregions 2 and 3, not up-flowline (Figure 3). These observations cannot be explained by variation in the amount of time available for landscape modification.

Ice velocity

Of the three controls on bedform elongation tested, the observations noted in this paper are best explained by differences in ice velocity across the field of study. Using elongation as a proxy measure of velocity, the prediction surface of bedform elongation demarcates two zones of fast ice flow (Figure 3). In Subregion 3, megaflutes located west of Oneida Lake are characterized by bedform elongation ratios up to 47:1. The Adirondack Highland (Figure 1) was exposed as a nunatak during deglaciation promoting divergent flow southward out of Canada to the east of the Adirondack Highland and southeastward from the Erie-Ontario Lowland (Krall, 1977). Ice flowing from the lowland into the Oneida Lake basin (hereafter referred to as the Oneida Sublobe) created the bedforms located west of Oneida Lake. Very elongate megaflutes (maximum 72:1) are also located north of Cayuga Lake (southeast corner of Subregion 2; Figure 3). The highest concentration of elongate features is located immediately north of Cayuga Lake, but we cannot discount the possibility that bedforms of similar morphometry existed north of Seneca Lake and have since been removed by channelized meltwater or ice-marginal lake swash zone erosion.

We suggest two interpretations that are consistent with the observations. Recent geophysical work on the Rutford Ice Stream has shown that drumlins can evolve over sub-decadal time scales (Smith *et al.*, 2007). Therefore, the elongation displayed in the now-exposed drumlin field may be representative of the palaeoglaciological conditions present at the LIS bed immediately prior to deglaciation. In this case, ice-marginal lakes may have played a role in the initiation and maintenance of fast ice flow. In contrast, other studies of the New York Drumlin Field suggest that drumlin morphometry reflects LIS dynamics during the Valley Heads stage when the terminus was south of the Finger Lakes (Francek, 1991; Ridky and Bindschadler, 1990). Under these conditions, the observed drumlin morphometry may reflect flow within topographically-controlled ice streams that populated the fjord-like troughs along the northern rim of the Appalachian Upland.

Ice acceleration toward a calving margin

Fairchild (1909) outlined a series of palaeoshorelines that illustrate a time-transgressive sequence of pro-glacial lake stages in central New York State. A large lake covered the interfluve between Seneca and Cayuga Lakes during the Lake Newberry Stage (Figure 1). The modern depth of Cayuga Lake (128 m) is second only to that of Seneca Lake (188 m) (Von Engeln, 1961). Total depth to bedrock beneath Cayuga and Seneca Lake sediments is 366 and 439 m, respectively (Mullins and Hinchey, 1989). Seneca and Cayuga Lakes collectively reside in a relatively deep trough (hereafter referred to as the Seneca-Cayuga trough) surrounded by the high ground of the Appalachian Upland (Figure 1). A deep lake of significant volume was created within the trough as the Valley Heads Moraine prevented the escape of meltwater to the south. To the east of Oneida Lake, lacustrine beach deposits mark the outline of a significant ice-marginal lake residing in the Oneida basin (Fairchild, 1909; Muller and Cadwell, 1986). Again, Oneida Lake is located between the Tug Hill Upland to the north and the Appalachian Upland to the south (Figure 1). Thus, both the Oneida basin and the Seneca-Cayuga trough held significant pro-glacial lakes during deglaciation.

Modern observations of calving glaciers typically show an increase in surface velocity toward their terminus (Krimmel, 1992; Abdalati and Krabill, 1999; Lefauconnier *et al.*, 2000). Longitudinally decreasing effective pressure resulting from buoyant forces induced by ice-marginal water has been observed at Columbia Glacier (O'Neel *et al.*, 2005). Moreover, basal drag decreases toward the margin in response to decreasing effective pressure and reaches zero when the ice is fully supported by water pressure (Benn *et al.*, 2007). The increase in surface velocity observed as ice approaches the terminus of a calving glacier is partly an expression of enhanced basal sliding and, therefore, should be present in the landform record.

Increasing elongation down flowline is observed in Subregions 2 and 3. We propose that this pattern reflects a localized increase in basal sliding velocity during deglaciation. In Subregion 2, the most elongate bedforms are proximal to a deep region of Lake Newberry; now the basin within which Seneca and Cayuga Lakes reside. In response to different lake stages, one would expect different regions of elongate bedforms generated by a calving margin of varying width. Indeed, megaflutes are found to the west of Seneca Lake and may represent calving into a lake residing at higher modern elevation. In Subregion 3, the observed distribution in bedform orientation (Figure 5c) may represent the path of a topographically-controlled outlet glacier or ice stream flowing into the Oneida basin. Again, the pattern of increasing elongation down flowline is suggestive of calving into an ice-marginal lake. Meltwater freely drained from Subregion 1 and a significant ice-marginal lake was not created. In support of the calving margin interpretation, highly attenuated bedforms are not found in Subregion 1.

Topographically controlled ice streams

During the Last Glacial Maximum, ice moved southward in Central New York State onto the Appalachian Upland and eventually reached the Wisconsinan limit in Pennsylvania (Figure 1). Flow from north to south across the Erie-Ontario Lowland was restricted by the Appalachian Upland. The northern rim of the transition to higher elevation has been significantly dissected thereby generating preferential pathways for flowing ice. The surface gradient of the LIS during retreat was quite low along the southern margin due to the abundance of deformable bed material (Mickelson and Colgan, 2003). The interaction of a low-gradient ice surface and climbing topography of the Appalachian Upland surely resulted in strong glaciodynamic sensitivity to underlying topography (e.g. Kessler *et al.*, 2008). In previous work, a numerical flowline model yielded evidence of outlet glaciers residing in the Finger Lake troughs upon LIS occupation of the landscape (Ridky and Bindschadler, 1990). The study determined that the flux of ice transported by these glaciers generated zones of fast-moving ice projecting into the interior of the ice sheet. Our results show a significant clustering of elongate bedforms located proximal to the deeply incised Cayuga Lake trough. These results agree well with a previous study of the megaflutes located north of Cayuga Lake that found a strong correlation between bedform elongation and modern ice stream velocity (Hart, 1999). In Subregion 3, Oneida Lake resides within a trough between the Appalachian and Tug Hill Uplands. Bedforms located proximal to Oneida Lake are highly elongate

(Figure 3) and their orientation is descriptive of streaming ice directed into the modern Oneida Lake basin. A zone of fast ice flow within the Oneida Sublobe was identified by a recent study (Briner, 2007). Using morphometric data, the study outlined a discrete transition to highly elongate megaflutes from bounding, ovoid drumlins and illustrated a pattern of increasing elongation down flowline. The results shown here, derived independently from those of Briner (2007), express similar findings. Flow of relatively fast-moving ice is represented by the sharp discontinuity in bedform orientation shown in Subregion 3 (Figure 5c). The results are suggestive of fast ice flow within the Oneida Sublobe as it flowed into the Oneida basin.

Conclusions

In this study we generated continuous prediction surfaces via kriging analyses to compare spatially non-coincident point measurements of bedform elongation and drift thickness. Caution must be exercised, however, when performing such predictions. A combination of validation statistics and comparison to independent data confirmed our interpolations. Through implementation of this technique we have determined that ice velocity played a key role in determining the morphometry of subglacial bedforms in the New York Drumlin Field. We conclude that drift thickness did not significantly influence bedform elongation in the region.

Our results supplement the growing body of literature that utilizes a glaciological inversion scheme in an attempt to better understand ice sheet dynamics. Previous studies in the area describe a radial pattern of bedform orientation in the New York Drumlin Field reflecting flow from a central dome. In contrast, we present three distinct zones of ice flow with vastly differing reconstructed flow trends. Multiple zones of fast ice flow are developed, however the timing of bedform formation is unclear due to a lack of suitable chronological control on ice sheet withdrawal from the area. We propose two mechanisms that may have promoted the localized fast ice flow defined in this study. Topographic funnelling likely directed ice into the broad Seneca-Cayuga trough and Oneida basin during the Valley Heads stage thereby increasing ice flux through each zone. During deglaciation, ice-marginal water may have influenced the longitudinal velocity gradient north of the Seneca-Cayuga trough and west of the Oneida basin. These processes, each of which are supported by our observations, are limited in definition by the lack of chronological control on subglacial bedform genesis in the New York Drumlin Field.

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