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**GIS-analyses of ice-sheet erosional impacts on the exposed
shield of Baffin Island, eastern Canadian Arctic**

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1 GIS-analyses of ice-sheet erosional impacts on the exposed shield of Baffin
2 Island, eastern Canadian Arctic

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6 **Abstract**

7 The erosional impacts of former ice sheets on the low-relief bedrock surfaces of Northern
8 Hemisphere shields are not well understood. This paper assesses the variable impacts of glacial
9 erosion on a part of Baffin Island, eastern Canadian Arctic, between 68 and 72°N and 66 and 80°W.
10 This tilted shield block has been covered repeatedly by the Laurentide Ice Sheet during the late
11 Cenozoic. The impact of ice-sheet erosion is examined by GIS-analyses using two geomorphic
12 parameters: lake density and terrain ruggedness. The resultant patterns generally conform to
13 published data from other remote sensing studies, geological observations, cosmogenic exposure
14 ages and the distribution of the Chemical Index of Alteration for tills. Lake density and terrain
15 ruggedness are thereby demonstrated to be useful quantitative indicators of variable ice-sheet
16 erosional impacts across Baffin Island. Ice-sheet erosion was most effective in the lower western
17 parts of the lowlands, in a west-east oriented band at around 350-400 m a.s.l. and in fjord onset
18 zones in the uplifted eastern region. Above the 350-400 m a.s.l. band and between the fjord onset
19 zones, ice-sheet erosion was not sufficient to create extensive ice-roughened or streamlined bedrock
20 surfaces. The exception where lake density and terrain ruggedness indicate that ice sheet erosion
21 had a scouring effect all across the study area was in an area from Foxe Basin to Home Bay that does
22 not exceed elevations of 400 m a.s.l.. These morphological contrasts link to former ice sheet basal
23 thermal regimes through the Pleistocene. The zone of low glacial erosion surrounding the cold-based
24 Barnes Ice Cap probably represents its greater extent during successive Pleistocene cold stages.

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Inter-fjord plateaus with few ice sheet bedforms remained cold-based throughout multiple Pleistocene glaciations. In contrast, zones of high lake density and high terrain ruggedness are a result of the repeated development of fast-flowing, erosive ice in warm-based zones beneath the Laurentide Ice Sheet. These zones are linked to greater ice thickness over western lowland Baffin Island. Adjacent lowland surfaces with similar elevations, however, of non-eroded, weakly eroded, and ice-scoured shield bedrock indicate that even in areas of high lake density and terrain ruggedness the total depths of ice-sheet erosion did not exceed 50 m.

Key words: GIS-analysis, ice-sheet erosion, glaciated shields, Baffin Island.

1. Introduction

The erosional impact of ice sheets on resistant Precambrian crystalline shield bedrock has been long debated. Two sharply divergent views exist: (i) Late Cenozoic glacial erosion amounted to depths of hundreds of meters (Laine, 1980; Bell and Laine, 1985) or (ii) that ice sheets have removed no more than a few tens of meters of bedrock from the Canadian and Fennoscandian Shields (Sugden 1976; Lidmar-Bergström 1997; Olvmo et al. 1999; Kleman et al. 2008; Ebert and Hättestrand 2010; Hall et al. 2013; Ebert et al. 2015). Establishing depths of glacial erosion on shields is important for its implications for understanding the origins of the low-relief bedrock geomorphology, for studies of sediment provenance in marginal sedimentary basins and for modeling crustal response to ice sheet loading.

The erosional impact of ice sheets on large-scale bedrock morphology in high-relief plateau and valley terrain has long been recognized as being selective (Sugden 1968). On glaciated passive margin edges, this differential erosion means that there is intense fjord glaciation and preservation of plateaus in between fjords under cold based ice (Kessler et al. 2008; Kleman 2008; Hall et al. 2013b). Ice sheet erosion patterns on low-elevation low-relief shield bedrock are less clearly demarcated (Ebert et al. 2015). The westward-tilted block of Baffin Island in the Eastern Canadian Arctic has been entirely covered by Laurentide ice sheets during the Late Cenozoic (Dyke et al. 2002, 2003). This

50 tilted block forms a low-relief surface without an obvious topography-induced erosion pattern and
 51 therefore provides an excellent area to investigate the impact of glacial erosion on low-relief shield
 52 terrain (cf. Sugden 1978; Andrews et al. 1985; Dowdeswell and Andrews 1985; Briner et al. 2006,
 53 2008; Staiger et al. 2006).

54 The area has previously been investigated by field studies, dating, geochemistry and remote sensing
 55 regarding its glacial geomorphology (Andrews et al. 1970; Sugden 1978; Andrews et al. 1985; Tippet
 56 1985; Staiger et al. 2006; De Angelis and Kleman 2010; Briner et al. 2008, 2014; Refsnider and Miller
 57 2013), providing a solid dataset for testing and refining the GIS-results of this study. The GIS-analysis
 58 brings a detailed and quantifiable overview of the erosion pattern in the study area that is
 59 transferable to other glaciated regions.

60 The aim of this paper is to analyze the cumulative impact of Late Cenozoic ice-sheet erosion on shield
 61 bedrock morphology of Baffin Island in the area between 68 and 72°N and 66 and 80°W. GIS analyses
 62 are used to identify the patterns of ice-sheet erosion, using the combination of two quantifiable
 63 geomorphic parameters: lake density and terrain ruggedness. The GIS results are compared to a
 64 range of other parameters linked to depths of glacial erosion, namely TCN-dates, CIA-values and
 65 locations of Paleozoic erratics. The zones of glacial erosion identified on Baffin Island constrain
 66 maximum depths of ice-sheet erosion and link to the dynamics of successive Laurentide ice sheets
 67 through the Late Cenozoic.

68 **2. Study area**

69 The study area of this paper comprises the area on Baffin Island between 68 and 72°N and 66 and
 70 80°W (**Figure 1**). Active valley glaciers on the eastern passive margin edge, draining into fjords, and
 71 the Barnes Ice Cap (**Figure 1**) are the only ice bodies existing today in the study area. The present ice-
 72 caps of Baffin Island are in large parts cold-based and clearly non-erosive, preserving underlying
 73 vegetation and patterned ground (Falconer 1966; Anderson et al. 2008; Miller et al. 2012; Margreth

et al. 2014). Baffin Island is barren and sparsely vegetated, facilitating remote sensing studies of the bedrock surface morphology (cf. Sugden 1978; Andrews et al. 1985).

2.1. Geology and large-scale morphology

Baffin Island is mainly comprised of Precambrian basement, primarily granites, granite gneisses, and a variety of metamorphic and igneous facies (Bridgewater et al. 1973; Behnia et al. 2013). Around the margin of Foxe Basin, outliers of Palaeozoic cover rocks occur (Sanford and Grant 2000) (**Figure 1b**).

The basement surface below the Palaeozoic cover is flat and was already planate when the sediments were deposited.

By the late Proterozoic large parts of the bedrock of the entire Canadian Shield had been eroded down to within a few hundred metres of its modern topography (e.g. Ambrose 1964; Bird 1972). The shield was probably completely covered by Ordovician sediments (Ambrose 1964; Bird 1972; Patchett et al. 2004), with further sedimentation in the Devonian. Baffin Island shows two planation surfaces that are partly in contact with Ordovician cover rocks and therefore might represent remnant morphology of the sub-Ordovician surface (cf. Ambrose 1964; Bird 1967; Utting et al. 2006).

During the late Cretaceous, rifting between West Greenland and Baffin Island led to the onset of uplift of the basement block in the west and tilting towards Foxe Basin (Dowdeswell and Andrews 1985). Westward drainage systems caused the stripping of the Palaeozoic cover rocks from wide areas. The flat forelands along the coast of eastern Baffin Island are covered by shallow-water and littoral sediments, deposited during periods of Quaternary marine transgression and regression (Andrews et al. 1988).

2.2. Glacial history

Baffin Island has been repeatedly covered by the Laurentide Ice Sheet during late Cenozoic glaciations (Refsnider et al. 2013). The Last Glacial Maximum was about 18 ky ago (Dyke et al. 2002, 2003) when Baffin Island was entirely covered by ice (Briner et al. 2006) (inset in **Figure 1**). Generally, the main ice center was on the Canadian Shield or in Foxe Basin (Andrews 1989; De Angelis and

Kleman 2010), with ice drainage from west to east up slope across Baffin Island. Under conditions of restricted glaciation, isolated ice caps existed on Baffin Island, similar to the polythermal Barnes Ice Cap of today (Dyke et al. 2002; 2003; De Angelis and Kleman 2008; Kleman et al. 2010). Along the elevated edge of the study area, ice drained eastwards through the fjords. To the north and south of the study area, across the entire Baffin Island, ice drainage was convergent northwards in the north of the island and southwards in the south, following the direction of the fjords and inlets (Kleman et al. 2010) (**Figure 1a**).

3. Data and methods

3.1. GIS input-data

DEM data was obtained from GeoBase, (moved to GeoGratis by end of Dec 2014, www.geogratis.gc.ca © Department of Natural Resources Canada. All rights reserved.). The Canadian Digital Elevation Data (CDED) consists of an ordered array of ground elevations at regularly spaced intervals. The source digital data for CDED at the scale of 1:50,000 is extracted from the hypsographic and hydrographic elements of the National Topographic Data Base (NTDB) or various scaled positional data acquired from the provinces and territories. Depending on the latitude of the CDED section, the grid spacing, based on geographic coordinates, vary in resolution from a minimum of 0.75 arc seconds to a maximum 3 arc seconds for the 1:50,000 National Topographic System (NTS) tiles. The elevation data of Baffin Island used in this paper has a resolution of ~18.5 m. All data presented in the figures is projected in the reference system NAD 1983 UTM Zone 14N.

The majority of the surface of Baffin Island displays a thin till cover that does not obscure the bedrock surface morphology (e.g. Andrews et al. 1970, 1988; Staiger et al. 2006; Briner et al. 2014; Refsnider et al. 2014). Snow cover might have existed during the production of the DEM; however, sources (air photos and satellite imagery) with no cover or the least amount of snow were selected (personal communication with Jean Pinard, Canada Centre for Mapping and Earth Observation).

With the exception of the present glaciers, the digital elevation model (DEM) therefore is regarded to represent the shield bedrock surface forms.

Geology and interpreted geological structures were downloaded from Behnia et al. (2013), Geological Survey of Canada open file (**Figure 1**). They consist of GIS-shapefiles of Behnia et al. (2013), who used the interpretation after the method of Harris et al. (2012) to identify geological structures such as dykes, faults, folds and structural form lines from landsat images and magnetic survey data. The hydrology, i.e. a water body database with digitized lakes and water courses, was downloaded from Geogratis (see link above) (**Figure 2**).

3.2. Lake density, lake surface area and lake elevation

Lake density is a recognized indicator for the areal scour of bedrock by glacial erosion, and has previously been used as such on Baffin Island (Sugden 1978; Andrews et al. 1985; Briner et al. 2008). Sugden (1978) and Andrews et al. (1985) calculated lake density from satellite images, in combination with topographic maps and aerial photographs for areas where lake density was concealed under snow cover on the satellite images. Lakes are now available as shapefile-layers for many areas on Earth, also for Baffin Island. In a recent study on NW Iceland (Principato and Johnson, 2009), lake density was analyzed in GIS, using a lake shapefile, to calculate both number and area of lakes in a 25x25 km² grid, using grid cells of 5x5 km.

For the present study, lakes were extracted from the water body database, provided by GeoBase. The water body database displays both lakes and water courses. Water courses are, unlike lakes, not created by glacial scouring, but instead may be formed by preglacial rivers and meltwater, with some water courses retaining their fluvial V-shapes and others being glacially overdeepened to U-shaped valleys. Water courses were therefore excluded from the analysis. The database shows 141 882 lakes in the study area. Lake density was analyzed using two methods:

147 Firstly, in the method described by Principato and Johnson (2009), the number of lakes and the
148 percentage of lake area for each grid cell were calculated. Principato and Johnson (2009) tested grid
149 cells of 25 km² and 100 km² and found that the resolution of the 100 km² grid was too low. To get an
150 even more detailed output map, ten times smaller grid cells were used in this study, namely
151 2.5x2.5 km grid cells (2500x2500 m or 6,25 km²). This grid cell size gives a clear visual of the lake
152 density (**Figure 3**). Also, the amount of data can be processed in ArcGIS for the size of the study area.
153 For lake density, i.e. the number of lakes per grid square, lakes were converted to points. The points
154 were positioned in the centroid of the lake, each point representing one lake. For lake surface area
155 per 6,25 km² grid cell, the shape-lake layer was converted to raster in the polygon to raster function
156 of ArcGIS. The quantitative method of Principato and Johnson (2009) gives a pixelated visual output,
157 but a quantitative measure of lake density and lake surface area in defined areas (**Figure 3a,b**).

158 Secondly, point density was calculated using the point density function in ArcGIS, using the lake point
159 layer. This methods results in a raster layer with soft transitions, in the calculation of the distribution
160 of areas with low and high lake densities (**Figure 3c**). The resulting values give the expected lake
161 density per pixel, divided by the pixel area. The raster resolution is 50m, i.e. each pixel covers
162 0.0025 km², and expected lake count per pixel would result in less than one lake per pixel (**Figure 3c**).
163 The values of the point density output are therefore not as quantitatively clear as calculations in a
164 larger grid (**Figures 3a and 3b**).

165 Elevation values were extracted for each lake, using the 'extract values to points' function of ArcGIS
166 spatial analyst, meaning that elevation data values were appended to points representing the
167 centroids of the lakes in order to provide consistent detail on lake elevation. Lake elevations are
168 presented in a histogram (**Figure 2**).

3.4. Surface roughness/terrain ruggedness

Surface roughness has been used to divide landscapes in landscape elements of different geologies or geomorphological processes, but only in areas of considerable relief to identify valley and mountain topography (Ascione et al. 2008; Jenness 2004; Li et al. 2010; references therein).

Areas of glacial scouring in lowland shield terrain are characterized by sharp local transitions in the relief, whereas areas of glacial preservation display no or few lakes, and no sharp edges of glacial erosion but a smooth texture (Briner et al. 2014; Ebert et al. 2015). In landscapes of areal scouring, the pre-glacial weathering front is stripped of saprolite to excavate the geological structure of the preglacial weathering front. In areas of most intense erosion, this structure is even further roughened by glacial erosion and plucking (Krabbendam and Bradwell 2014). Therefore, terrain roughness is a measure of ice sheet erosional impact in low-relief shield terrain. However, the relative relief of lowland terrain is limited and differences between areas of low and high glacial erosional impact are subtle.

Surface roughness for the study area was analyzed in three different ways: 1) the standard deviation of slope (cf. Ascione et al. 2008); 2) the Relative Topographic Position Index (**Figure 4**), a metric for terrain ruggedness (cf. Jenness 2004); 3) the Fast Fourier transformation (FFT) (cf. Lie et al. 2010) (**Figure 4**). All methods were tested with a number of parameters in order to make them function for low-relief terrain.

To calculate the Topographic Position Index, the system of Cooley (2013) was used:

$$(\text{"smoothed DEM"} - \text{"minimum DEM"}) / (\text{"maximum DEM"} - \text{"minimum DEM"})$$

The minimum, maximum and smoothed (mean elevation) DEMs were calculated in the focal statistics function of ArcGIS Spatial Analyst, with a resampled 50m elevation model. The output of this model gives the relative location of a cell in comparison of the mean elevation in a specified neighbourhood.

194 The higher the value, the more the cell elevation differs from the mean elevation, giving a measure
195 of “terrain ruggedness”.

196 The Fast Fourier Transformation was calculated in ENVI using the resampled 50m elevation model.
197 The Fast Fourier Transformation results in a raster file that re-projects the DEM into frequency space
198 displaying the spatial frequency of the data. Higher wavelength frequency indicates greater terrain
199 roughness. The data was then filtered to exclude low frequency elements after which the result was
200 re-projected into the image or spatial domain where a 5x5 pixel low pass filter was applied to reduce
201 noise.

202 *3.5. Overlay*

203 Input parameters for the GIS-overlay study were lake density, derived through point density (**Figure**
204 **3c**) and surface roughness derived through the Topographical Position Index (**Figure 4a**). Areas with
205 present day glaciers, both valley glaciers on plateaus in between fjords on Baffin Island’s eastern
206 margin, as well as Barnes ice cap in the centre of Baffin Island, were not included in the analyses. The
207 DEM shows the present land surface, including the glacier surface on top and consequently GIS-
208 analysis of these surfaces is not valid. Present day glaciers are therefore shown as shapefile-layer in
209 all figures to make clear which areas are excluded from all analyses. The result of the overlay displays
210 pixels in the DEM that indicate a deviation from the average elevations, with other words surface
211 roughness, in different colours that represent increasing degrees of lake density (**Figure 6**).

212 *3.6. Previous studies in the study area*

213 *3.6.1. Remote sensing*

214 Sugden (1978) showed landscapes of glacial erosion in northern Canada, including Baffin Island,
215 interpreted by calculating lake density from air photos, existing maps, and field work (**Figure 5A**). The
216 map of Baffin Island was refined by Andrews et al. (1985), who analyzed lake density and elevations
217 from satellite images, in combination with topographic maps and aerial photographs for areas were

lake density was concealed under snow cover on the satellite images (**Figure 5B**). Both serve as starting point for this paper; these studies provided the first ice-sheet erosional impact maps of Baffin Island.

3.6.2. Geological evidence

A boulder train of Paleozoic erratics, deposited on shield bedrock of Baffin Island between Foxe Basin and Home Bay, was mapped by Tippet (1985) (**Figure 5C**). The boulder train has a source area for the boulders in the Paleozoic cover rocks west of Baffin Island (Andrews and Sim, 1964) and requires significant erosion of these rocks. The boulder train is elongate and falls within the area of highly scoured terrain that crosses the study area. The boulder train can be traced eastwards to the coast of Home Bay fjords where Palaeozoic erratics have been mapped (Andrews et al. 1970) (**Figure 5C**). The erosion and long distance transport of Paleozoic erratics requires warm-based ice flow from Foxe Basin to Home Bay.

3.6.3. Terrestrial cosmogenic nuclide (TCN) dating

Briner et al. (2006) identified differentially weathered bedrock in fjord zones, demonstrated by cosmogenic exposure ages of bedrock from ^{10}Be and ^{26}Al concentrations, and linked these to differential ice-sheet erosion (**Figure 5C**). Cosmogenic exposure ages derived from ^{10}Be and ^{26}Al indicate a lack of glacial erosion on smooth summits 500-800 m above ice-scoured valley floors in the eastern edge of Baffin Island (Miller et al. 2006) (**Figure 5C**). The pattern is consistent with selective linear glacial erosion beneath this part of the Laurentide Ice Sheet in the area during at least the Middle and Late Quaternary. In an investigation of a combination of Be^{10} and Al^{26} concentrations, ice-sheet model simulations, and field studies of till characteristics, Staiger et al. (2006) identified the thermal basal regimes of the ice-sheets which deposited till across northern central Baffin Island, namely cold-based till, intermediate till (with changing conditions) and warm-based till (**Figure 5C**). This indicates the ice-sheet to be warm based and erosive in valleys leading to fjords, and cold-based and protective on adjacent high plateaus (Staiger et al. 2006). ^{10}Be exposure ages were also used for

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calibrating the ice-sheet erosional impacts in a fjord onset zone (Briner et al. 2008) (**Figure 5C**). This study is a further confirmation of previous studies finding the selective character of ice sheet erosion in fjord- and upland terrain (Sugden 1968; Kleman 2008; Kessler et al. 2008).

An erosion line in north central Baffin Island divides glacially scoured, uneven terrain, with the presence of lakes, from non-scoured, low-relief, till and felsenmeer covered terrain (**Figure 5C**); verified by Briner et al. (2014) with ^{10}Be , ^{14}C , ^{26}Al exposure dating, confirming greater ice-sheet erosion in the northern side of the line. The line divides two areas with subtle differences in terrain, compared to areas of most intense erosion on Baffin Island; the line therefore provides a good calibration line for the present study.

3.6.4. The Chemical Index of Alteration (CIA)

Refsnider and Miller (2013) calculated CIA values for till across a large part of the study area using the equation of Nesbitt and Young (1982). The CIA sample sites cover a considerable part of the north of the present study area (**Figure 5C**). The higher the CIA, the more intense the weathering of a site, and the more time available for weathering processes. In areas of warm-based ice, where material was repeatedly removed by glacial erosion, the CIA is low; in areas of cold-based ice, the material stays *in situ* and the weathering process continues in ice-free times with whenever the climate is sufficiently warm and humid to promote weathering. The CIA helps to identify key areas of glacial preservation through the Pleistocene glaciations (Refsnider and Miller 2013) and allows reconstruction of pre-glacial landscape conditions (cf. Ebert et al. 2012b; Hall et al. 2015).

Datapoints of the previous results in the study area, such locations of TCN dating, Paleozoic erratics and CIA values (described in 2.3) were collected and digitized in a GIS-dataset (**Figure 5C**) to compare the results against the GIS analysis.

4. Results

4.1. Lake density

Lake density here, as in previous studies on Baffin Island (Sugden 1978; Andrews et al. 1985; Briner et al. 2008) and Iceland (Principato and Johnson 2009), provides a good measure of aerial scouring of bedrock and thereby ice sheet erosion. The three methods to display lake density or lake area display a similar pattern (**Figure 3**). Highest lake density, and thereby most intense impact is displayed in the lower southern parts of the Baffin Island block with a peak in a west-east band around 350-400 m a.s.l. (lake elevation histogram, **Figure 2a**); across entire Baffin Island in the lower area from Foxe Basin to Home Bay (names and profile across this area in **Figure 1a**); and in the fjord onset zones with particularly intense erosion in the onset zones east of the current Barnes ice cap. The forelands in front of the fjord zone of selective glacial erosion show low lake densities or no lakes (**Figure 3**), confirming the results of Briner et al. (2006) in the area of Clyde and Aston Forelands (see location of sample points in **Figure 2**). However, even though displaying seemingly correct results, it clearly cannot be used as a good proxy for areas built of Quaternary sediment (like the forelands), because lake density in this study represents the scouring effect of ice on shield bedrock. Lakes on sediments may have other origins such as dead-ice melt-out.

The two different analyses of lake density, namely the quantitative count of the number of lakes in grid cells of 6,25 km² (**Figure 3a**) and the expected lake density for each cell in a raster with 50 m resolution (**Figure 3c**) miss out the area of extremely strong erosion in the area of **Figure 2d** (indicated in **Figure 3**), where alignment of bedrock structure and ice flow direction lead to an intense streamlined carving of the bedrock structure. The reason for this is that lakes are converted to one point in the centroid of each lake in order to conduct lake density calculations. In areas where the streamlining has led to the formation of relatively large, elongate lakes, the lake count will correctly display a lower number of lakes and therefore a lower lake density, however this wrongly indicates lower ice sheet erosional impact (**Figure 2d**).

This problem does not occur in the calculation of **Figure 3b** where the actual lake area was calculated for each grid cell – correctly, maximum lake area is displayed in the south-western corner of **Figure 2d**. On the other hand, the calculation of lake area displays maximum values also in the area of **Figure 2e**, the locations of Conn and Bieler lakes. These lakes are ice-dammed lakes, dammed by the eastern edge of Barnes ice cap, and do not give a measure of aerial scouring. The methods presented to analyze lake count and density detect the area of weak aerial scouring north of the “erosion line” of Briner et al., 2014 (**Figure 3**), and confirm that subtle differences in glacial erosion intensity, as validated by TCN dating, are correctly detected by GIS-analysis of lake density.

4.2. Terrain ruggedness/topographical roughness

Three methods were used to derive terrain roughness. The standard deviation of slope did not show satisfying results, with no meaningful pattern at all, and will not be further discussed.

The other two methods used to display terrain ruggedness, or topographic roughness, principally give a good visual impression of the patterns of glacial impact on Baffin Island, corresponding to the results of the lake density analysis and the basic patterns identified by Sugden (1987) and Andrews et al. (1985) (**Figure 5 A,B**). However, each method has their weaknesses:

In the topographical position index (TPI) (**Figure 4a**), rectangular structures are visible that represent the boundaries of satellite images or other images that the DEM was derived from, especially south and south-west of Barnes Ice Cap. Lake shores are steep and therefore comprise a major part areas representing terrain ruggedness – which displays the lake density, in an indirect way. However, many locations within rougher low-relief plateau terrain that are not lake shores are identified as well. Density of pixels with deviating topographic positions is a measure for terrain ruggedness; the denser the net of pixels with deviating topographic positions, the greater the terrain roughness, thereby indicating more intense ice-sheet erosional impact (**Figure 4a**).

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The Fast Fourier Transformation (FFT) also gives a good overview over the ice-sheet erosional impact pattern, especially valleys aligned along the direction of ice flow, fjord onset zones, and the ice stream zone that runs from Foxe Basin to Home Bay (**Figure 4b**). However, values are erroneous at the northern, eastern and southern edges of the elevation model, where an edge effect to the kernel-based analysis disrupts the values at the borders of the study area. Like in the TPI, the edges of satellite images or other images that the DEM was derived from was derived from get visible, especially west of Barnes Ice Cap.

4.3. Overlay

In the overlay analysis, the areas of deviating relative topographic position that indicate terrain ruggedness were coloured according to lake density, emphasizing the areas of most intense terrain ruggedness and greatest lake density (**Figure 6**). Threshold values for colouring areas of decreasing ice sheet erosional impact are the Standard Deviation of the lake density (see **figure 3c**). In the overlay analyses it becomes clear that lake density is the main indicator for ice sheet erosional impact in the study area, however this is further enhanced by terrain roughness, which shows a denser net of pixels with terrain ruggedness in areas of most intense ice-sheet erosional impact. The TPI would be a possible alternative option to show the overall pattern of ice sheet erosional impact if lakes are not available as a digital layer.

4.4. Patterns and boundaries of bedrock morphology, tested against previous studies

Lake density and terrain ruggedness, and thereby inferred ice-sheet erosional impacts, are highest in areas partly comprised of Paleozoic cover rocks along the west coast of Baffin Island; generally in the lower western area of Baffin Island, intensifying uphill, with a peak at 350-400 m a.s.l.; across the area between Foxe Basin and Home Bay; in fjord onset zones at the eastern edge of Baffin Island (**Figure 6**).

338 An absence of lakes, and low terrain ruggedness, and thereby inferred weak or no ice-sheet erosional
339 impacts, is present between the zone of intense scouring at 350-400 m a.s.l. and the fjord onset
340 zones, at elevations of about 400-600 m a.s.l. Barnes Ice cap is situated in the centre of this zone,
341 concealing the bedrock surface beneath, but melting out across areas of lowest lake density and
342 terrain ruggedness especially in the northern part of the ice cap. In the area between Foxe Basin and
343 Home Bay that does not reach elevations above 400 m a.s.l. (**profile C-D in figure 1**), ice-sheet
344 scouring extends across the entire Island.

345 Sometimes very clear boundaries exist between two adjacent classes of ice sheet erosional impact,
346 e.g. in fjord onset zones, where ice started to funnel in the drawdown-effect of the fjord and linear
347 erosion starts to emphasize (**Figure 6c**). Here the transition between zones of differential impact of
348 ice sheet erosion narrows to 10-20 km, to finally in the fjords become a clear boundary between
349 valley erosion and shoulders of preserved plateaus. On the plains, when no clear selectivity or ice
350 funneling in valleys takes place, a wide transition zone of 50 km or more occurs between the highest
351 and lowest impact intensities (**Figure 6 b and d**).

352 Local differences in relative relief in the study area do not exceed 50 m, with exception of valleys that
353 may be > 100 m in depth. Relative relief does not show great differences in range neither within nor
354 across the boundaries of different zones of glacial erosional impact. Therefore, no greater relief
355 differences were present before or after ice-sheet cover, and ice sheet erosion can not exceed
356 depths of 50 meters and was likely much lower, with only exception of glacially reshaped valleys,
357 positioned in ice sheet flow direction, that allowed a funneling of ice.

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5. Discussion

5.1. Lake density and terrain ruggedness as measures of ice-sheet erosional impact on low-relief shield terrain

The results of the lake density analyses and the terrain ruggedness fit with patterns of previous large-scale remote sensing studies (Sugden 1977; Andrews et al. 1985) (**Figure 5 A, B**), showing these patterns on a much more detailed level. Most intense erosion is indicated in a band across the western part of the study area, and along a transect of an eastwards linear zone of high ice-sheet erosional impact in the area from Foxe Basin to Home Bay. On a more detailed level, the results of the GIS-analysis conform to CIA-values (**Figure 6b**), TCN ages (**Figure 6c**) and the degree of excavation of geological structures (**Figure 6d**). Weak erosional impact is calculated in areas that give high CIA values, old TCN ages and a smooth land surface whereas strong erosional impact is calculated in areas that give low CIA values, young TCN ages and where the geological structure is excavated. Also in areas of subtle ice-sheet erosional impact, like along the erosion line of Briner et al. (2014), GIS-results confirm field results, on a detailed level. Even though the CIA might vary with bedrock types (Refsnider and Miller 2013), its overall pattern shows a statistically highly significant correlation with the lake density in areas of shield bedrock (**Figure 7**). A debris train of far-transported Paleozoic erratics falls in the centre of the linear zone of high ice-sheet erosional impact from Foxe Basin to Home Bay (Tippett 1985), and limestone erratics of the same rock type are located along the fjord walls of Home Bay (Andrews et al. 1970). The results of the GIS-analysis, in confirmation with the checks against previous data, confirm the suitability of lake density and terrain roughness to assess the impact of ice-sheet erosion on low-relief shield surfaces. Basic preconditions are a digitized lake layer for analysis in GIS, a DEM of resolution of 50m or higher, and knowledge about sediment thickness in the analyzed area. Areas with thick sediments that conceal the underlying bedrock topography in the DEM would not be suitable for this kind of study.

5.2. Ice-sheet erosional impact patterns and possible links to ice dynamics

The ice-sheet erosional impact pattern links to ice sheet basal thermal regime, which may have a topographical control. Baffin Island does not exceed elevations of 400 m.a.s.l. in the corridor from Foxe Basin across to Home Bay (see profile **Figure 1b**), suggesting that the location of the Home Bay Ice Stream was influenced by this corridor (cf. Home Bay Ice Stream, Andrews et al. 1970; Tippet 1985; Briner et al. 2006). This indicated that rather subtle changes in elevation can produce linear zones of fast flow. In higher elevated areas to a maximum of 600 m.a.s.l. to the north and south of this area, ice was slowed moving upslope to the passive margin edge and remained largely cold-based. The fjord onset zones acted to funnel ice down into the fjords, to give fast, warm-based ice flow and to increase erosion capacity (cf. Andrews et al. 1985). The erosion patterns for areas below 400 m.a.s.l. in the study area must relate to times when the Laurentide Ice Sheet extended all across Baffin. Only then the area from Foxe Basin to Home Bay can have had a complete ice cover, thick enough to develop warm-based, fast flowing ice that could erode cover rocks in Foxe Basin and transport and deposit Palaeozoic erratics in a debris train (Tippet 1985) and in erratics at the east coast of Baffin (Andrews et al. 1970).

A radial pattern of ice-sheet erosional impact around Barnes Ice Cap, with a distance of between 50 and 70 km of its present margin, represents periods when the Barnes Ice Cap had a larger extent. If so, the patterns suggest a cold-based low-erosiveness of the ice cap in its northern and southern parts and warm-based erosive ice of the ice cap flowing into the fjords towards the east and also warm-based erosive ice on the plains in the west. No lakes and no terrain roughness indicated around the northern margin and around the south-western edge of Barnes Ice Cap indicate that the ice cap was cold based in these areas.

Ice sheet erosion has a streamlining effect on the bedrock surface in areas in areas of form lines that are in line with ice flow direction, like in the area SW of Barnes Ice Cap (**Figure 2d**), an area of Palaeoproterozoic sedimentary rocks. This area is exceptional for its streamlined appearance with

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elongated bedrock ridges in the study area – stemming from structural form lines that were further emphasized by glacial overprint. Also, this area is an exceptionally low lying area at the Baffin Island west coast and ice thickness under the Foxe Dome would be at a maximum. Generally, the more diversified the pre-glacial relief, with clear elevation differences, the narrower the transition between glacially scoured and glacially preserved areas. In flat terrain, the transition zones are wider and less clear.

5.3. Depths of glacial erosion

Geomorphological studies with the attempt to quantify glacial bedrock erosion of the Canadian shield (Sugden 1976, 1978) including Baffin Island (Staiger et al. 2006), and the Fennoscandian shield (cf. Lidmar-Bergström et al. 1997; Hall et al. 2013) point to limited erosion of bedrock by glacial erosion, in a range of tens of metres even in areas of strongest glacial impact, that does not overprint the preglacial large-scale morphologic expression of the landscape, with exception for fluvial V-shaped valleys being widened to glacially shaped U-shaped valleys. Other studies have pointed to much higher erosion rates (Bell and Laine 1985). More dating evidence of weathering remnants is needed, however, the geomorphic indicators presented here and in previous studies, along with existing dates, are in agreement with previous results that strongly point to limited glacial erosion of the exposed shield of Baffin Island.

Baffin Island fjords as fjords elsewhere were overdeepened by hundreds of meters through glacial erosion. Glacial erosional depth on the shield plains was much lower, in a range of tens of metres, as ice erosion was not linear but in maximum capacity achieved areal glacial scouring, excavating, but not destroying, the structure of the underlying shield bedrock. Ice sheet erosional impact might have been maximal on the shield where the grain of the geological structures, or water courses, are in line with ice sheet flow direction (area shown in **Figure 2d**) (cf. Ebert et al. 2015); one of the most intensely eroded areas according to the overlay (**Figures 6, 8**).

The results of this paper show that even subtle topography has an influence on ice erosional dynamics, like in the low area Foxe Basin to Home Bay, or generating erosion in areas with a parallel alignment of structure and ice flow direction. This influence of even subtle topography implies that, assuming that ice sheet erosion could not change the preglacial large-scale topographical forms, every ice sheet had a similar impact pattern with areas of cold-based and warm-based ice developing in the same areas.

Baffin Island fits well in the pattern of formerly glaciated areas in crystalline rock terrain that display weathering remnants adjacent to areas stripped of weathering. Differentially weathered landscapes can be the result of selective erosion by overriding ice sheets (Sugden 1978; Hall 1986; Hall and Sugden 1987). First regoliths had probably an effect on ice sheet geomorphology (Balco and Rovey 2010). Preglacial regolith was likely stripped with first glaciations (Balco and Rovey 2010; Refsnider and Miller 2013). However, preglacial regolith might be preserved in continuously cold-based areas, witnessing of preservation under cold-based ice during at least the latest, or all the Late-Cenozoic ice sheets (Ives 1975; Sugden and Watts 1977; Staiger et al. 2006; Leblanc-Dumas et al. 2015), like on the Canadian shield (Bouchard and Jolicoeur 2000; Jansson and Lidmar-Bergström 2004); in Fennoscandia (Peulvast 1985; Hirvas 1991; Migon and Lidmar-Bergström 2001; Hättestrand and Stroeve 2002; André 2004; Ebert et al. 2012a), and Scotland (Hall 1985). Dating of weathered remnants on these timescales are still few, but existing dates of saprolites (Ebert et al. 2012b) and tors with exposure ages of tors of >0,5-1.0 Ma on the Fennoscandian Shield (Stroeve et al. 2002; Darmody et al. 2008) point to preservation under cold-based ice of these weathering remnants likely during the last 1 Myr or possibly the entire Quaternary.

5. Conclusions

GIS-analyses of lake density and the topographic position index, a measure of terrain roughness, give a good measure of ice-sheet erosional impact on the shield bedrock surface. Areas of high lake density, and high terrain roughness coincide and indicate areas of intense glacial scouring. This confirms to results of earlier studies of remote sensing, and cosmogenic dating and the chemical index of alteration in small parts of the study area. The glacial erosional signal can consequently be identified with this type of GIS-analysis and makes it transferable to other areas where elevation data in sufficient resolution, digitized lakes, or preferably both these datasets are available.

Ice sheet erosion on the low-relief shield plains on Baffin Island was most effective in the western part of Baffin Island, on the lower part of the lowlands below 400 m a.s.l., especially in areas where the grain of the geological structure and ice movement direction coincide, and in the fjord onset zones.

An approximate maximum intensity of glacial scouring occurs in a west-east band at around 350-400 m a.s.l., as indicated by highest lake density and terrain ruggedness. Above the 350-400m a.s.l. line and in between fjord onset zones, ice-sheet erosional impact on the bedrock surface was minimal, as indicated by low lake density, low terrain ruggedness, high CIA values and high TCN ages.

The exception where ice sheet erosion has a scouring effect all across Baffin Island was in the area from Foxe Basin to Home Bay, where high lake density, high terrain ruggedness, and transported Paleozoic erratics witness of an erosive ice sheet. Here, elevations do not exceed 400 m a.s.l., allowing the ice to maintain sufficient thickness to reach the pressure melting point and remain warm based.

The fjord onset zones to the east of the current Barnes ice cap show most intense traces of ice-sheet erosional impact, where ice thickened to become warm-based due to fjord drawdown. Fjord onset zones to the south and north show partly very weak ice-sheet erosional impact.

In areas of strongest erosion in the study area, the bedrock structure was excavated, not overprinted, by glacial erosion. No great differences in relief were shaped by ice-sheet erosion between preserved and eroded areas, merely the structural grain, the roughness of the landscape, was changed, pointing to depths of ice-sheet erosion of no more than tens of meters.

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Figure captions

Figure 1. a) Location of study area in larger context; main ice drainage directions at LGM (from Kleman et al. 2010) and maximum ice extent (extent boundary from Dyke et al. 2002, 2003). b) Study area. Profile lines and two profiles across the study area. c) Location of exposed shield and cover rocks and d) interpreted geological structures (c and d from Behnia et al. 2013).

Figure 2: a) Lakes in the study area. Data provided by © Department of Natural Resources Canada. All rights reserved. Lake elevations histogram, with clear peaks at sea level and at 350-400 m a.s.l. (cf. Andrews et al. 1985). b) area of high lake density and location of debris train of Ordovician cover rocks (Tippett 1985), indicating warm-based ice movement across this area (see also Andrews et al. 1970). c) erosion line, derived by TCN dating of Briner et al., 2014. Note no lakes south of erosion line (interpreted as low erosion by Briner et al. 2014) and presence of lakes north of erosion line (higher erosion). Note that this detail figure is more strongly zoomed in, and the difference of lake numbers in b) and c). d) Lake orientation adapted to interpreted geological structures by Behnia et al. 2013. e) large ice-dammed lakes at the northern margin of Barnes ice cap.

Figure 3: Lake density and area. a) lake count per 6,25 km² grid cell. b) lake area in % per 6,25 km² grid cell. c) lake density as point density. Detail areas of fig. 2 added to facilitate discussion.

Figure 4: Terrain roughness. a) terrain roughness by the Topographic Position Index (TPI). The higher the density of areas with high TPI, the more intense the ice-sheet erosional impact. b) Terrain

roughness by the forward Fourier Transformation (FFT), where low values indicate areas with low ice-sheet erosional impact. (Please not for b) that values are invalid at the outer limit of the area). Higher terrain roughness indicates strong ice sheet erosional impact, low terrain roughness weak ice sheet erosional impact. Both methods are slightly erroneous in areas where they reveal edges of background data used for DEM production, in straight and rectangular structures west of Barnes Ice Cap. Ignoring these errors, both methods confirm the patterns of generally stronger ice-sheet erosional impact in the west of the area, across the entire Peninsula in the lower terrain from Foxe Basin to Home Bay (see Figure 1) and in fjord onset zones, with strongest impact in the onset zones east of Barnes Ice Cap.

Figure 5. Calibration data used in this study. A) Map of glacial erosional impact assessed by remote sensing (redrawn from Sugden 1978). B) The percentage of lakes calculated from satellite images (redrawn from Andrews et al. 1985). C) Digitized locations of TCN dating, CIA values and Paleozoic erratics. This is a simplified presentation of the locations and outcomes of these studies, focusing on the values for ice-sheet erosional impact. Please consult the individual studies for more specific information (these are Andrews et al. 1970; Tippet 1985; Briner et al. 2006; Miller et al. 2006; Staiger et al. 2006; Briner et al. 2008; Refsnider and Miller 2013; Briner et al. 2014).

Figure 6. a) Ice sheet erosional impact in the study area – raster overlay of best results, calibrated with data of previous studies. b) area of weak ice sheet erosional impact. Soft landform structures, low terrain roughness, high CIA values > 70 (Refsnider and Miller 2013). c) fjord onset zone. High terrain roughness and young Be10 ages (Briner et al. 2008) in the area where ice was warm-based, erosive, and thickened to draw down in the fjord. Low terrain roughness, low lake density and old Be10 ages on higher elevations beside the fjord drawdown zone (Briner et al. 2008; Miller et al. 2006 – please consult these papers for deeper explanation of this simplified presentation). d) Area of areal scouring, with increasing gradient from left to right, indicated by increasing lake density and terrain

roughness. In the most intensely scoured area, the bedrock structure appears excavated (interpreted structures from Behnia et al. 2013).

Figure 7: The relationship between CIA (from Refsnider and Miller 2013) and lake density (values extracted from lake density shown in figure 4c). CIA values > 40 with only few exceptions occur in areas with low lake density, indicating weak ice-sheet erosional impact. CIA values are decreasing with increasing lake density, indicating intensifying ice-sheet erosional impact. A statistically significant correlation between CIA and lake density is only true for Precambrian shield rocks and not for Palaeozoic cover rocks (shown separately) (see location of dots in figure 2 and location of Palaeozoic cover rocks in figure 1b).

Figure 8. Main impact patterns of ice sheet erosional impact deducted from lake density and terrain roughness. Bathymetry (source: GEBCO <http://www.gebco.net/>) reveals troughs of varying dimensions in the shelf that may have been main ice drainage pathways for eroded shield bedrock into Baffin Bay.

Figure 1. a) Location of study area in larger context; main ice drainage directions at LGM (from Kleman et al., 2010) and maximum ice extent (extent boundary from Dyke et al., 2002, 2003). b) Study area. Profile lines and two profiles across the study area. c) Location of exposed shield and cover rocks and d) interpreted geological structures (c and d from Behnia et al., 2013).

148x147mm (300 x 300 DPI)

Figure 2: a) Lakes in the study area. Data provided by © Department of Natural Resources Canada. All rights reserved. Lake elevations histogram, with clear peaks at sea level and at 350-400 m a.s.l. (cf. Andrews et al., 1985). b) area of high lake density and location of debris train of Ordovician cover rocks (Tippett, 1985), indicating warm-based ice movement across this area (see also Andrews et al., 1970). c) erosion line, derived by TCN dating of Briner et al., 2014. Note no lakes south of erosion line (interpreted as low erosion by Briner et al., 2014) and presence of lakes north of erosion line (higher erosion). Note that this detail figure is more strongly zoomed in, and the difference of lake numbers in b) and c). d) Lake orientation adapted to interpreted geological structures by Behnia et al., 2013. e) large ice-dammed lakes at the northern margin of Barnes ice cap.

177x262mm (300 x 300 DPI)

Figure 3: Lake density and area. a) lake count per 6,25 km² grid cell. b) lake area in % per 6,25 km² grid cell. c) lake density as point density. Detail areas of fig. 2 added to facilitate discussion.
240x426mm (300 x 300 DPI)

Figure 4: Terrain roughness. a) terrain roughness by the Topographic Position Index (TPI). The higher the density of areas with high TPI, the more intense the ice-sheet erosional impact. b) Terrain roughness by the forward Fourier Transformation (FFT), where low values indicate areas with low ice-sheet erosional impact. (Please not for b) that values are invalid at the outer limit of the area). Higher terrain roughness indicates strong ice sheet erosional impact, low terrain roughness weak ice sheet erosional impact. Both methods are slightly erroneous in areas where they reveal edges of background data used for DEM production, in straight and rectangular structures west of Barnes Ice Cap. Ignoring these errors, both methods confirm the patterns of generally stronger ice-sheet erosional impact in the west of the area, across the entire Peninsula in the lower terrain from Foxe Basin to Home Bay (see Figure 1) and in fjord onset zones, with strongest impact in the onset zones east of Barnes Ice Cap.

141x168mm (300 x 300 DPI)

Figure 5. Calibration data used in this study. A) Map of glacial erosional impact assessed by remote sensing (redrawn from Sugden, 1978). B) The percentage of lakes calculated from satellite images (redrawn from Andrews et al., 1985). C) Digitized locations of TCN dating, CIA values and Paleozoic erratics. This is a simplified presentation of the locations and outcomes of these studies, focusing on the values for ice-sheet erosional impact. Please consult the individual studies for more specific information (these are Andrews et al., 1970; Tippet, 1985; Briner et al., 2006; Miller et al., 2006; Staiger et al., 2006; Briner et al., 2008; Refsnider and Miller, 2013; Briner et al., 2014).

184x201mm (300 x 300 DPI)

Figure 6. a) Ice sheet erosional impact in the study area – raster overlay of best results, calibrated with data of previous studies. b) area of weak ice sheet erosional impact. Soft landform structures, low terrain roughness, high CIA values > 70 (Refsnider and Miller, 2013). c) fjord onset zone. High terrain roughness and young Be10 ages (Briner et al., 2008) in the area where ice was warm-based, erosive, and thickened to draw down in the fjord. Low terrain roughness, low lake density and old Be10 ages on higher elevations beside the fjord drawdown zone (Briner et al., 2008; Miller et al., 2006 – please consult these papers for deeper explanation of this simplified presentation). d) Area of areal scouring, with increasing gradient from left to right, indicated by increasing lake density and terrain roughness. In the most intensely scoured area, the bedrock structure appears excavated (interpreted structures from Behnia et al., 2013).
221x327mm (300 x 300 DPI)

Figure 7: The relationship between CIA (from Refsnider and Miller, 2013) and lake density (values extracted from lake density shown in figure 4c). CIA values > 40 with only few exceptions occur in areas with low lake density, indicating weak ice-sheet erosional impact. CIA values are decreasing with increasing lake density, indicating intensifying ice-sheet erosional impact. A statistically significant correlation between CIA and lake density is only true for Precambrian shield rocks and not for Palaeozoic cover rocks (shown separately) (see location of dots in figure 2 and location of Palaeozoic cover rocks in figure 1b).

74x31mm (300 x 300 DPI)

Figure 8. Main impact patterns of ice sheet erosional impact deduced from lake density and terrain roughness. Bathymetry (source: GEBCO <http://www.gebco.net/>) reveals troughs of varying dimensions in the shelf that may have been main ice drainage pathways for eroded shield bedrock into Baffin Bay.
77x51mm (300 x 300 DPI)