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1 **Statistical EOF analysis of spatiotemporal glacier**  
2 **mass-balance variability: A case study of**  
3 **Mittivakkat Gletscher, SE Greenland**

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30 **Abstract**

31 An Empirical Orthogonal Function (EOF) variance analysis was performed to map and elucidate  
32 in detail the spatiotemporal variability in individual stake mass-balances ( $b_a$ ) on the Mittivakkat  
33 Gletscher (MG) – in a region where at present five out of ~20.000 glaciers have mass-balance  
34 observations. The EOF analysis suggested that observed  $b_a$  was summarized by two major  
35 modes: EOF1 and EOF2 represented 80 % (significant) and 6 % (insignificant) of the explained  
36 variance, respectively. EOF1 captured a decline in  $b_a$  that was uniformly distributed in space at  
37 all stakes. The decline was correlated with: 1) albedo observations; and 2) surface air  
38 temperature observations from nearby maritime and coastal stations. EOF2, however, described  
39 variations in  $b_a$  that were heterogeneously distributed among stakes and associated with local  
40 slope and aspect. Low-elevation stakes (~<400 m a.s.l.) showed relatively negative (out of  
41 phase) correlation and higher elevated stakes relatively positive (in phase) eigenvector  
42 correlation values with EOF2. Such relatively negative and positive eigenvector correlation  
43 values were present where the surface of MG constituted of exposed glacier ice or snow cover,  
44 respectively. The results from this study show how EOF analyses can provide robust information  
45 on spatiotemporal patterns of glacier mass-balance. Understanding such detailed variabilities in  
46 mass-balance on a Greenlandic glacier is of interest because a fifth of the Arctic contribution  
47 from glaciers and ice caps to sea-level rise originated from Greenland.

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51 **Keywords** Empirical Orthogonal Function analysis; Greenland; observations; mass-balance  
52 stakes; Mittivakkat Gletscher

53 **1. Introduction**

54 At present, glaciers and ice caps are important contributors to eustatic sea-level rise (e.g.,  
55 Marzeion et al. 2012; Gardner et al. 2013; Allison et al. 2015). Over the last several decades,  
56 glaciers and ice caps in the Arctic have been observed to decrease in area and volume in  
57 response to climate changes contributing to sea-level changes (Leclercq et al. 2011; Bjørk et al.  
58 2012; Cogley 2012; Kargel et al. 2012; Marzeion et al. 2012; Zemp et al. 2015). Direct  
59 glaciological surface mass-balance time series and glacier meteorology observations from local  
60 glaciers (i.e. glaciers and ice caps surrounding the continental ice sheets; Weidick and Morris  
61 1998) are scarce and sparsely distributed in Greenland. In total, less than 20 local glaciers in  
62 Greenland have recorded mass-balance ( $B_a$ ) observations covering different time periods since  
63 1892-93 and to present (Machguth et al. 2016). At present, only five glaciers out of ~20,000  
64 individual local glaciers in Greenland have ongoing annual glacier mass-balance observation  
65 programs. This is a minor fraction of local glaciers in Greenland (Pfeffer et al. 2014; Radić et al.  
66 2014), which cover a total area of  $\sim 89,300 \pm 2,800 \text{ km}^2$  (Rastner et al. 2012). This lack of glacier  
67 mass-balance observations leaves us with limited information about local glacier conditions in  
68 Greenland and its contribution to sea-level changes. Local glaciers in Greenland are, for  
69 example, less well-studied than the main ice sheet, so processes driving their change and their  
70 sensitivity to climate variables are more unclear. However, it was recently stated by AMAP  
71 (2017) that a fifth of the Arctic contribution from glaciers and ice caps to sea-level rise originates  
72 from Greenland. It is, therefore, important to extract and generalize all relevant information  
73 about these monitored glaciers.

74 Mittivakkat Gletscher (henceforth MG), located in Southeast Greenland on Ammassalik  
75 Island, is Greenland's only peripheral glacier for which there exist long-term ongoing  $B_a$  records

76 continuously since the mass-balance year 1995/96 (Mernild et al. 2011). Even before this time,  
77 B. Fristrup in 1933, 1958, and 1970 and B. Hasholt in 1986 and 1998 have demonstrated  
78 relationships between climate-induced surface ablation and  $b_a$  variations and freshwater runoff  
79 on MG (Fristrup 1970; Hasholt and Jakobsen 2008). The four other local glaciers in Greenland  
80 with ongoing mass-balance monitoring programs are A. P. Olsen Iskappe (74.6°N, Zackenberg,  
81 East Greenland; Larsen et al. 2012), Freya Gletscher (74.4°N, Clavering Island, East Greenland;  
82 Hynek et al. 2014), Qaanaaq Gletscher, which is a part of Qaanaaq Iskappe (74.4°N, Qaanaaq,  
83 Northwest Greenland; Sugiyama et al. 2014), and Qassinnguit Gletscher (64.1°N,  
84 Nuuk/Kobberfjord, West Greenland; Abermann et al. 2014). The longest continuous  
85 observational program beside the MG mass-balance program goes back to 2007 and is operated  
86 by the Austrian Polar Research Institute on Freya Gletscher (Hynek et al. 2014; Machguth et al.  
87 2016). A common characteristic for all ongoing local glacier mass-balance programs in  
88 Greenland is that they show mean negative  $B_a$  for their period of observations.

89         As the MG mass-balance record started in 1995/96, MG is a valuable study site for  
90 obtaining detailed understanding of trends in mass-balance (for both  $B_a$  and  $b_a$ ) in Southeast  
91 Greenland. On catchment-scale, MG also serves as an important site for studying the variability  
92 of morphological characteristics (area, mean thickness, volume, surface slope, etc.) and ice  
93 dynamics (Mernild et al. 2013a) and the climate sensitivity of local glaciers in Greenland  
94 (Mernild et al. 2011, 2013b). Local glaciers in Southeast Greenland are influenced by air  
95 temperature and precipitation changes following the oscillation in the Atlantic Multidecadal  
96 Oscillation (AMO; Kaplan et al. 1998), where a relatively high temperature anomaly is in anti-  
97 phase with a relatively low precipitation anomaly (during positive AMO), and vice versa (during  
98 negative AMO) (Chylek et al. 2009; Mernild et al. 2012a). Since the end of the Little Ice Age

99 (~AD 1900), MG has undergone almost continuous retreat (Knudsen et al. 2008). MG has  
100 reduced in area by 18 % (1986–2011), mean ice thickness by 22 % (1994–2012), volume by 30  
101 % (1986–2011), and mean ice surface velocity by 30 %, which can be fully explained by the  
102 dynamic effect of ice thinning (Mernild et al. 2013a; Yde et al. 2014). Further, MG has  
103 undergone a surface elevation change where the vertical strain rate was able to compensate for  
104 ~50 % of the surface elevation lowering due to the surface mass-balance (SMB) conditions  
105 (Mernild et al. 2013a).

106         In this study, our aim is to address the knowledge gap in understanding the  $B_a$  variability,  
107 with special focus on the variations between individual stake-observed  $b_a$  mass-balance  
108 measurements. An improved understanding of long-term mass-balance variability is important to  
109 emphasize the link between glacier changes, meteorology and albedo feedbacks. We do this by  
110 analyzing a 19-year time series (1995/96–2013/14) of  $b_a$  data from 34 individual MG long-term  
111 observation stakes. More specifically, we analyzed, mapped, and evaluated the patterns of both  
112 temporal and spatial MG  $b_a$  variations using Empirical Orthogonal Functions (EOF). An EOF  
113 analysis allows us to describe simultaneously how spatial patterns of  $b_a$  change over time among  
114 the 34 stakes. It is possible to combine the EOF output with a cross-correlation analysis of local  
115 and regional climate and geometric surface data to make hypotheses about the factors driving the  
116 spatiotemporal patterns of  $b_a$  on MG. A detailed spatiotemporal mapping of  $b_a$  variation is  
117 needed if we want to fully understand the factors influencing  $B_a$  conditions and the impact on the  
118 hydrosphere in a “typical” Arctic landscape with a glacierized area, a downstream proglacial  
119 valley and outwash plain, and a delta and coastal zone (Hasholt and Jakobsen 2008).

120         To our knowledge this is the first time in Greenland an EOF analysis has been conducted  
121 on a local glacier scale to evaluate the spatiotemporal pattern in glacier  $b_a$ . MG mass-balance

122 data up to 2010/11 have earlier been published in Mernild et al. (2013a), although only the  $B_a$   
123 was analyzed. EOF analysis has previously been applied in glacier studies by Mair et al. (2002)  
124 on Haut Glacier d'Arolla, Switzerland, who analyzed the spatiotemporal surface ice velocity  
125 field. Also, Walters and Meier (1989) and Mernild et al. (2015a) have analyzed the large-scale  
126 spatiotemporal variability of glacier  $B_a$  conditions in western North America and along the  
127 Andes Cordillera to the sub-Antarctic islands, respectively. These analyses identified, for  
128 example, correlations between  $B_a$  and large-scale atmospheric and oceanic indices.

129

## 130 **2. Study area**

131 Mittivakkat Gletscher formerly known as Midtluagkat Gletscher (26.2 km<sup>2</sup> in 2011);  
132 65°41' N, 37°48' W) is located in the Ammassalik region, southeast Greenland (Figure 1). MG  
133 extends from 180 to 880 m above sea-level (a.s.l.) (Knudsen et al. 2008), and from 1986–2011  
134 the mean surface slope changed from 5.4 degree to 5.9 degree. Approximately ~19 % of MG (4.9  
135 km<sup>2</sup>) had a slope between 6–10 degrees, ~20 % (5.2 km<sup>2</sup>) between 11–15 degrees, ~ and ~18 %  
136 (4.8 km<sup>2</sup>) between 16–20 degrees (for other slope intervals, see Table 1, Figure 2). With regards  
137 to aspect ~20 % (5.2 km<sup>2</sup>) of the surface are facing towards the west, whereas ~14 % (3.8 km<sup>2</sup>)  
138 are facing northwestwards and ~13 % (3.5 km<sup>2</sup>) are facing southwestwards (for other slope  
139 intervals, see Table 1, Figure 2). Since 1995, the ELA has risen from ~350 m a.s.l. to ~>880 m  
140 a.s.l., with a mean ELA of ~750 m a.s.l. The ELA is the spatially averaged elevation of the  
141 equilibrium line, defined as the set of points on the glacier surface where the net mass-balance is  
142 zero. This ELA change has resulted in an average Accumulation Area Ratio (AAR) of 0.15  
143 (Mernild et al. 2011). Hence, MG is significantly out of balance with the prevailing regional

144 climate and will likely lose at least 70 % of its current area and 80 % of its volume even in the  
145 absence of further climate changes (Mernild et al. 2011).

146 Mean annual air temperature (MAAT) for the study period is  $-2.1^{\circ}\text{C}$  (1993–2011) at  
147 Station Nunatak (Hanna et al. 2012; Mernild et al. 2014a), and mean annual corrected  
148 precipitation in the range of c. 1,400–1,800 mm water equivalent (w.e.) (1998–2006) (Mernild et  
149 al. 2008b) (precipitation was corrected after Allerup et al. 1998, 2000). For Station Tasiilaq a  
150 Danish Meteorological Institute (DMI) operated synoptic climate station located ~10 km  
151 southeast of MG (Station Tasiilaq is not shown on Figure 1), MAAT and mean annual  
152 precipitation sum changed (linear; both significant based on a linear regression t-test)  $\sim 0.8^{\circ}\text{C}$   
153  $\text{decade}^{-1}$  and  $\sim -200$  mm w.e.  $\text{decade}^{-1}$ , respectively (Hanna et al. 2012) . In this study, the term  
154 ‘significant’ is only used when the relationship is statistically significant at the 5 % level or  
155 better ( $p \leq 0.05$ ).

156

### 157 **3. Methods**

#### 158 3.1 Mass-balance program

159 The stake network used to measure the MG net annual balance (Figure 1) is based on the  
160 direct glaciological method (Østrem and Brugman 1991); summer balance was calculated as the  
161 difference between the measured net annual balance and the measured winter balance. As the  
162 MG observations were started in the spring 1996 by measuring the winter accumulation along  
163 transects and stakes were later drilled into the snow and ice during the early summer, there are no  
164 direct observations of stake changes to determine net balance during the balance year 1995/96.  
165 This made it possible to determine the summer balance in late-august 1996. The 1995/96 net



166 balance was determined as the winter balance subtracted the summer balance and not measured  
167 directly at stakes.

168         Since 1996/97, mass-balance measurements from individual stakes were obtained  
169 covering 16.3 km<sup>2</sup> of the main MG area, excluding the southeastern part of the glacier due to a  
170 high density crevassed area and the northern part of the glacier due to logistical reasons. This  
171 omission is not likely to bias the B<sub>a</sub> results as the surface of these areas follows the general  
172 hypsometric distribution of MG (Mernild et al. 2006). In total, 34 b<sub>a</sub> stakes were used for this  
173 study, where ~25 % have ≥15 annual observations during the period 1995/96–2013/14 and ~60  
174 % have ≥10 (Figures 3 and 4). The locations of the stakes are shown in Figure 1 and cover the  
175 elevation range of MG. Direct B<sub>a</sub> observations are subject to uncertainties (where specifically  
176 year 2002 seems to be underestimated, see further below). The methodological uncertainty of B<sub>a</sub>  
177 estimates on a single glacier is, according to Zemp et al. (2013), in the range of ±340 mm w.e.  
178 This uncertainty is due to a combination of measurement and analytic errors and has been added  
179 to the data set.

180

### 181 3.2 EOF analysis

182         Empirical Orthogonal Function (EOF) analysis is a standard method in earth and marine  
183 sciences for exploring spatio-temporal variation in a variable. It is a principle components  
184 analysis applied to a data matrix organized by location (space) and time. The method has been  
185 applied to glaciological analyses several times (e.g., Walters and Meier 1989; Mernild et al.  
186 2015a). In this study, the focus is on the spatiotemporal variability in b<sub>a</sub>.

187         Our approach to estimating the EOFs implements Data Interpolating Empirical  
188 Orthogonal Functions (DINEOF; Beckers and Rixon 2003) because our data are ‘gappy’ with

189 missing data in various years at various stakes. Our use of DINEOF via the R ‘sinkr’ package  
190 (<https://github.com/marchtaylor/sinkr>) fills gaps by iteratively decomposing the data field via  
191 singular value decomposition until a best solution is found as compared to a subset of reference  
192 values (Beckers and Rixon 2003; Taylor et al. 2013). We further extended the approach by  
193 interpolating and estimating the EOFs for our data 50 times and using the mean of these. This is  
194 because the interpolation process involves randomization and permutation.

195 As in any ordination technique, the major axes (e.g., EOF1 and EOF2) represent  
196 independent collections of information on the variable of interest, and in this case, variations in  
197  $b_a$  in space and time. The EOF analysis captures variations in  $b_a$  simultaneously in time and space  
198 (Figure 4). The significance of the variations captured by each EOF can be evaluated several  
199 ways. These tools are designed to reveal how many major axes of variations there are in the data.  
200 We relied on bootstrap randomization approach to estimate the significance (see  
201 <https://github.com/marchtaylor/sinkr>).

202 The eigenvectors associated with such an analysis are linked to locations and thus reveal  
203 the influence of different geographic locations on the summarized mass-balance patterns and  
204 further allow analyses of meteorological covariates linked to the EOFs. All data were centered  
205 around zero and scaled to unit variance (Mernild et al. 2015a).

206 We first examined the eigenvectors (loadings) of the first EOF, which showed how this  
207 major axis of spatiotemporal variation in  $b_a$  varies among stake locations. This provides a  
208 graphical visual summarization of the geographic patterns of temporal variation in  $b_a$ . The  
209 second assessment involved exploring how these correlations among sites and EOF varied  
210 regarding to site-specific characteristics such as elevation, slope, and aspect. The third  
211 assessment involved relating the temporal trends in EOFs to observed temporal trends in

212 meteorological surface conditions. These trends include: i) observed MAAT (September through  
213 August; following the mass-balance year) and mean summer surface air temperature (June  
214 through August) from Station Nunatak; ii) observed MAAT, mean summer temperature, and  
215 winter precipitation sum (September through May) from Station Tasiilaq, and iii) the observed  
216 mean MG glacier-wide albedo.

217         The surface albedo is defined as the reflected fraction of incoming solar shortwave  
218 radiation (e.g., Dumont et al. 2012) and is a parameter that governs energy availability for snow  
219 and ice surface ablation, and subsequently variabilities in  $b_a$  conditions. The mean MG glacier-  
220 wide surface albedo was estimated for a 16-day composite at the end of the mass-balance year  
221 period (27/28 July–12/13 August; 2000–2013), derived from the MODerate Imaging  
222 Spectroradiometer (MODIS MCD43A3) albedo product. For verification and technical details  
223 about the MODIS MCD43A3 MG albedo product, see Mernild et al. (2015b).

224         For estimation of MG surface slope and aspect a digital elevation model (DEM) was  
225 extracted from the Advanced Spaceborne Thermal Emission and Reflection Radiometer  
226 (ASTER) Global Digital Elevation Model Version 2 (GDEM v2) (based on the best observations  
227 between 2000 and 2010), with a vertical average precision of ~12 m over Greenland (Tachikawa  
228 et al. 2011). The vertical error was expected to be closer to the GDEM v2 standard  $\pm 8.7$  m  
229 precision due to the gentle slope (<10 degrees) of the MG surface (Table 1) from where the  
230 measurements were taken (Tachikawa et al. 2011). The lateral error associated with GDEM v2 is  
231 a little more than half a pixel (17 m). The ASTER GDEM v2 is a product of the US Ministry of  
232 Economy, Trade, and Industry and NASA.

233

#### 234 **4. Results and discussion**

#### 235 4.1 Variations in $B_a$ observations

236 In Figure 5, the MG  $B_a$  time series are shown. On average since 1995/96, annual  $B_a$  was -  
237  $1.00 \pm 0.70$  m water equivalent (w.e.) (where  $\pm$  equals one standard deviation), indicating a  
238 cumulative mass-loss of 19 m w.e. (Figure 5). The MG  $B_a$  loss has on average changed (linear)  
239 by  $-0.06$  m w.e.  $\text{yr}^{-1}$  (significant;  $r^2 = 0.26$ , where  $r^2$  is the square of the linear correlation  
240 coefficient). Overall, in comparison to the other local glacier mass-balance programs in  
241 Greenland (even though different time periods were compared), the mean annual  $B_a$  were  
242 negative for all observed glaciers:  $-0.55 \pm 0.56$  m w.e. Freya Gletscher (2007/08–2013/14)  
243 (Hynek et al. 2014),  $-0.40 \pm 0.56$  m w.e. Qaanaaq Gletscher (2012/13–2014/15) (Sugiyama et al.  
244 2014), and  $-0.17 \pm 0.34$  m w.e. Qassinnguit Gletscher (2012/13–2014/15) (Abermann et al.  
245 2014). Data were not available for A. P. Olsen Iskappe, as these data are too poorly constrained  
246 (pers. com., M. Citterio, September 2014). The available observed mean  $B_a$  conditions indicate a  
247 ‘snap-shot’ in time for various periods, where a mean negative  $B_a$  not only is a local phenomenon  
248 at MG, but present at other mass-balance observed glaciers in SE, NE, NW, and W Greenland.

249 In 2010/11, the  $B_a$  at MG was at a record setting  $-2.45$  m w.e. This was more than two  
250 standard deviations ( $-2\sigma$ ; Figure 5) below mean. In both 2004/05 and 2009/10,  $B_a$  was more than  
251 one standard deviation below mean ( $-1\sigma$ ), and in 1995/96 and 2002/03 more than one standard  
252 deviation above mean ( $+1\sigma$ ). These deviations were highly dominated both by relatively high  
253 mean summer temperature conditions and subsequently high ablation rates for 2004/05, 2009/10,  
254 and 2010/11 or enhanced winter precipitation conditions causing high accumulation rates for  
255 1995/96 and 2002/03. Annual and seasonal variabilities in surface air temperature have in  
256 general an impact on the snow and firm temperature conditions (on the cold content), as  
257 temperature changes propagate into the snow and firm (e.g., Cuffey and Patterson 2010).

258 Therefore, high MAAT likely indicates a relatively lower end of winter season (May 31) snow  
259 cold content and subsequently contributing to an early start of the melt season.

260 When  $B_a$  was either below (2004/05, 2009/10, and 2010/11) or above one standard  
261 deviation (1995/96 and 2002/03) between 66–100 % of the individual stake  $b_a$  observations had  
262 values that were below or above one standard deviation, respectively. For the remaining 14  
263 years,  $B_a$  was within one standard deviation. Overall for the observation period, changes in  
264 glacier winter balance ( $B_w$ ) and summer balance ( $B_s$ ) explain 59 % and 90 % of the variability in  
265  $B_a$  (both significant; not shown), respectively.  $B_w$  and  $B_s$  were observed in 13 out of 19 years.

266 During the period 2004/05–2013/14, for example, seven of the highest recorded  $B_a$  losses  
267 have occurred. During the past five years three of those losses were recorded in 2010/11,  
268 2009/10 (-2.16 m w.e.), and 2011/12 (-1.63 m w.e.) (Figure 5). In terms of the mean surface  
269 summer air temperature, the period 2004/05–2013/14 had six and seven of the highest values  
270 observed at Station Nunatak and Station Tasiilaq, respectively (not shown). For Station Tasiilaq,  
271 the period 2004/05–2013/14 also included seven of the driest winters recorded (not shown),  
272 indicating that the climate in the MG region both got warmer and drier at the same time  
273 (Cappelen 2015; Mernild et al 2012a).

274

#### 275 4.2 Variations in $b_a$ observations

276 The spatial distribution of mean MG observed  $b_a$  1996/97–2013/14 is illustrated on  
277 Figure 6a. Observed mean  $b_a$  values were lowest  $\sim$ -3.0 m w.e. at the lowest MG elevations close  
278 at the margin (180 m a.s.l.) and  $\sim$ -0.5 m w.e. at the highest elevations ( $\sim$ 650 m a.s.l.), indicating  
279 net ablation everywhere on the glacier. The mean annual  $b_a$  value of -1.0 m w.e. was observed at  
280  $\sim$ 500 m a.s.l., whereas mean  $b_a$  lower than  $-1\sigma$  (-1.7 m w.e.) and  $-2\sigma$  (-2.4 m w.e.) were observed

281 at elevations lower than ~400 m a.s.l. and ~300 m a.s.l., respectively (Figure 6a). The spatial and  
282 longitudinal distributions of observed  $b_a$  variability ( $1\sigma$ ) are illustrated in Figures 6b and 6e,  
283 respectively, indicating that the temporal variability in  $b_a$  increased in value with increasing  
284 elevation until 300–400 m a.s.l. Above this elevation the variability in  $b_a$  decreased slightly. In  
285 the frontal area below 200 m a.s.l., the mean  $b_a$  standard deviation was 0.48 m w.e. and above  
286 600 m a.s.l. it was 0.77 m w.e. The highest temporal variability of 0.85–0.88 m w.e. was present  
287 at elevations between 300–600 m a.s.l. (Figure 6e). This peak in  $b_a$  variability at elevations  
288 between 300–600 m a.s.l. along the central part of the glacier was likely identical to the average  
289 ascent of the ELA over the observation period (Mernild et al. 2015b), probably because the  
290 variability in  $b_a$  was influenced by the combined effects of changes in winter and summer  
291 meteorological conditions. This elevation range (300–600 m a.s.l.) was also the area, where the  
292 greatest change (-0.25) in the observed end of mass-balance year albedo occurred, identifying  
293 this zone as an important surface cover and albedo transitional zone – an ELA zone (Mernild et  
294 al. 2015b). The lowest variations in  $b_a$  together with the lowest changes in surface albedo both  
295 occurred where MG was either snow covered at the end of the mass-balance year (in the high-  
296 elevation accumulation zone) or constituted of exposed glacier ice (in the low-elevation ablation  
297 zone). Hence, in case the variability in  $b_a$  followed the  $-1\sigma$  or  $+1\sigma$  variability for a specific year,  
298  $b_a$  values in the frontal area would be as low as ~-3.5 m w.e. and at approximately 525 m a.s.l. ~-  
299 1.8 m w.e. for  $-1\sigma$  (Figure 6c), and ~-2.6 m w.e. and ~0 m w.e. (identical with the location of the  
300 ELA) for  $+1\sigma$  (Figure 6d), respectively. In other words,  $b_a$  followed the upper and lower  
301 boundaries of the longitudinal profile as illustrated in Figure 6e.

302

303 4.3  $b_a$  observations and EOF variance analysis

304 In Figure 3, the annual observed  $b_a$  time series are shown for all 34 individual stakes on  
305 MG. Overall, the variability in the  $b_a$  time series occurred both for each individual time series  
306 and between the annual stake values. For some years such as 2002/03 ( $\sigma = 0.45$  m w.e.), 2005/06  
307 ( $\sigma = 0.60$ ), and 2010/11 ( $\sigma = 0.37$ ) marked with blue colored squares in Figure 3, the  $b_a$  spatial  
308 variability was relatively low compared to other years 2000/01 ( $\sigma = 1.04$ ), 2001/02 ( $\sigma = 1.10$ ),  
309 and 2012/13 ( $\sigma = 1.09$ ) marked with red colored circles. This indicates that for the first three  
310 highlighted years relatively homogeneous  $b_a$  conditions and a low  $b_a$  gradient existed, and  
311 opposite for the last three highlighted years. The years with relatively negative  $B_a$  (Figure 5)  
312 were the years with relatively low spatial variability in  $b_a$  (i.e., relatively homogeneous  $b_a$   
313 conditions, Figure 3), and opposite for years with less negative or positive  $B_a$  ( $r^2 = 0.43$ ,  
314 significant).

315 In Figure 7, we report on two major axes (modes) estimated from our EOF analysis of  
316 spatiotemporal variation: EOF1 and EOF2 representing 80 % (significant) and 6 %  
317 (insignificant) of the explained variance, respectively. The temporal and spatial variability in  
318 EOF1 and EOF2 are illustrated in Figures 7 and 8, respectively. Regarding EOF1, we show a  
319 five-year running mean smoothing line that increased over the time-period, being negative until  
320 2003 and positive thereafter (Figure 7). For EOF2 the five-year running mean smoothing line  
321 oscillated with an approximately six-year frequency. EOF2 was negative during the periods  
322 1995–1999, 2003–2006, and 2009–2012 and positive during the periods 1999–2003, 2006–2009,  
323 and after 2012. Hence, the EOF2 five year running mean smoothing line was more complex than  
324 the EOF1 pattern.

325 Both the temporal EOF1 and EOF2 patterns were associated with individual eigenvectors  
326 for each individual stake (Figure 8). For EOF1, the bar-plot in Figure 8 illustrates negative

327 values for all stakes, indicating that EOF1 was capturing increasing mass loss at all stakes and  
328 therefore decreasing annual  $B_a$  loss. These negative correlations emphasize that variations in  
329 mass-loss were uniformly distributed in space and time at all stakes (Figures 8a and 9a).

330 We found that EOF1 showed a significant correlation with Station Nunatak mean  
331 summer air temperature and glacier-wide albedo (Figure 10 and Table 2). However, we also  
332 noted that values at year 2002 were a statistical outlier (due to an underestimation of measured  
333  $b_a$ ). As an example, if we ignore the year 2002 values EOF1 was correlated with MAAT and  
334 mean summer air temperature from both Station Nunatak and Station Tasiilaq and observed  
335 mean glacier-wide albedo, but not winter precipitation sum from Station Tasiilaq (Figure 11 and  
336 Table 2). We suggest that EOF1 strongly reflected variability in air temperature conditions and  
337 albedo, rather than variability in precipitation conditions and topographic conditions. None of the  
338 correlations to precipitation at both Station Nunatak and Station Tasiilaq were significant (Table  
339 2). We suggest that this is because precipitation in Greenland varies according to the topography  
340 and the patterns of weather systems and even within short distances precipitation is likely  
341 explained by prevailing wind circulation e.g., katabatic winds draining downslope from the ice-  
342 sheet interior, distance from the oceanic moisture source and the orographic effect of near-  
343 coastal mountains. The latter is especially important in Southeast Greenland (e.g., Hansen et al.  
344 2008) and in the area of MG, where the mean corrected annual precipitation varies from ~1,250  
345 mm w.e. (1999–2006) (uncorrected ~900 mm w.e.) at Station Tasiilaq to ~1,850 mm w.e. at  
346 Station Nunatak (1999–2006) (Mernild et al. 2008b, 2015c).

347 Station Tasiilaq is located around 10 km southeast of MG and highly influenced by  
348 maritime climate conditions, having a mean monthly air temperature range of 10–15°C  
349 (Cappelen et al. 2015). In comparison, Station Nunatak was influenced by coastal climate



350 conditions, having a mean monthly air temperature range of 15–25°C (Mernild et al. 2008b). At  
351 MG the  $B_a$  variability is significantly correlated with variabilities in both maritime and coastal  
352 climate conditions (for definitions of maritime and coastal climate conditions, see Przybylak  
353 2003). It therefore seems very likely that similar variabilities in  $B_a$  may occur for other glaciers  
354 located within maritime and coastal climate conditions in Southeast Greenland. Due to the  
355 significant correlation between MG  $B_a$  and observed air temperature both at Station Nunatak and  
356 Station Tasiilaq – two stations located in different climate conditions –, we propose that  
357 variability in  $B_a$  is not only a local phenomenon. Mernild et al. (2011) found that MAAT time  
358 series from Southeast Greenland’s coastal DMI stations were significantly correlated with  
359 MAAT time series from Station Tasiilaq. These data further suggest that  $B_a$  variability at MG,  
360 which has been driven largely by surface air temperature variabilities, is representative for mass-  
361 balance variations on a regional scale, which includes many hundreds of local glaciers. In  
362 support of this statement, the trends in glacier terminus recession at MG and mass-balance  
363 conditions are on average similar to glacier terminus recessions for land-terminating glaciers in  
364 the Ammassalik region (Mernild et al. 2012b) and overall for land-terminating glaciers in  
365 Southeast Greenland (Bjørk et al. 2012), and to simulated mass-balance conditions in the  
366 Ammassalik region (Mernild et al. 2014b), respectively.

367         In contrast to the EOF1 dataset, the EOF2 pattern was complex and showed a more  
368 spatial variable pattern with in total 11 (shown with light gray color) out of 34 stakes (~33 % of  
369 the stakes) being in phase with EOF2 and 23 stakes (~67 % of the stakes) being out of phase  
370 with EOF2 (dark gray; Figure 8b). While our bootstrap analysis indicates that the 6 % of  
371 variation captured by EOF2 is not significant given the current set of data, we explore it because

372 the EOF2 pattern was complex and showed a more spatial variable pattern than the EOF1  
373 dataset; we note these thus represent hypotheses about more localized patterns in  $b_a$ .

374 Surface elevation, slope, and aspect explained none of the  $b_a$  site-specific correlations  
375 with EOF1, but they explained variation in the spatial component of EOF2 (Figures 9a–9c). We  
376 suggest that the 11 stakes being in phase with EOF2 may have different mean geometrical  
377 surface terrain features such as different mean surface slope and aspect than the other 23 stakes  
378 in the mass-balance network. In general, we found that the 11 stakes being in phase with EOF2  
379 likely had lower mean surface slope than the other 23 stakes being out of phase with EOF2 (see  
380 also Figure 9b). For these 11 stakes, the mean surface slope was 8.5 degrees, whereas the mean  
381 slope at the other 23 stakes was 11.0 degrees (significant not equal). Further, our analysis  
382 indicates that the stakes being in phase with EOF2 mainly had south facing aspects; whereas  
383 stakes being out of phase with EOF2 mainly had west and north facing aspects (see also Figure  
384 9c). Even though the EOF2 pattern probably can be explained by the variability in surface  
385 conditions – by the mean surface slope and/or aspect, the more general EOF1 pattern seemed to  
386 be dominated by other controls than surface characteristics such as variations in meteorological  
387 conditions.

388 Further, EOF2 indicated a general eigenvector correlation pattern related to elevations on  
389 MG, where low-elevated stakes ( $\sim < 400$  m a.s.l.) showed relatively negative (out of phase)  
390 eigenvector correlation values and higher elevated stakes relatively positive (in phase)  
391 eigenvector correlation values (Figure 9a). Such relatively negative and positive eigenvector  
392 correlation values were present where MG either constituted of exposed glacier ice in the low-  
393 elevated ablation zone or snow cover in the high-elevated accumulation zone during  $b_a$   
394 observations. Also, these patterns were likely related to changes in both MG snow cover duration

395 and the occurrence of frequent air temperature inversion on the lower part of MG (~<300 m  
396 a.s.l.): Inversion and sea breezes associated with the adjacent relatively low temperature and  
397 frequently ice-choked fjords and oceans (Mernild and Liston 2010). Observations indicate that  
398 air temperature inversion is to be present 84 % of the time, and essential for accumulation and  
399 surface ice melt conditions (Mernild and Liston 2010). Therefore, in general these EOF2 stake  
400 eigenvector correlations likely indicated a change from eigenvectors being out of phase to  
401 eigenvector being in phase with increasing elevation, even though minor fluctuations in  
402 eigenvector correlations occurred due to variations in both surface slope and aspect (even within  
403 a distance less than a few hundred of meters). The EOF2 correlation values emphasize that  
404 variations in mass-loss were heterogeneously distributed (insignificant) in space and time at the  
405 MG stakes.

406 An understanding of such MG EOF correlations – its variability to changes in climate and  
407 surface conditions – is of interest for different purposes because local glaciers in Greenland are  
408 less well-studied than the Greenland Ice Sheet, but also for mass-balance upscaling proposes to  
409 regions from where no glacier mass-balance observations are available.

410

## 411 **5. Conclusions**

412 Our findings show that in 14 out of 19 years,  $B_a$  was within one standard deviation with a  
413 mean annual loss of  $-1.00 \pm 0.70$  m w.e., a change of  $-0.06$  m w.e.  $\text{yr}^{-1}$  during a period of climate  
414 warming and drying, following the variability in AMO (Mernild et al. 2012a). Relatively  
415 negative MG  $B_a$  equal low spatial variability in  $b_a$ . A change in  $B_a$  indicates that the greatest  
416 variability in  $b_a$  occurred for stakes located between 300–600 m a.s.l. and is likely related to the  
417 ascent of the ELA.

418 The EOF analysis is a robust way of obtaining spatiotemporal information on glacier  
419 mass-balance patterns. The use of a statistical EOF variance analysis suggests that observed  $b_a$   
420 on MG in time and space can be summarized by a single major axis of variation EOF1  
421 representing 80 % (significant) of the explained variance. Further, EOF1 emphasizes variations  
422 in  $b_a$  related to variations in albedo and surface air temperature observations from nearby  
423 maritime and coastal stations. However, we also explore the second major axis, which explained  
424 6 %, because it suggests possible localized geographic effects on  $b_a$  linked to surface conditions  
425 associated with slope and aspect. Our EOF calculations are crucial for our understanding of  
426 spatiotemporal mass-balance variabilities and for mass-balance upscaling possibilities in  
427 Greenland because variabilities in  $b_a$  are influenced strongly by variabilities in summer air  
428 temperature and MAAT and mean glacier-wide observed albedo, and insignificant against  
429 surface characteristics such as slope and aspect. Additionally, it is crucial for our understanding –  
430 in case of upscaling to regions with no observations – because we have few mass-balance  
431 observations from local glaciers in Greenland.

432

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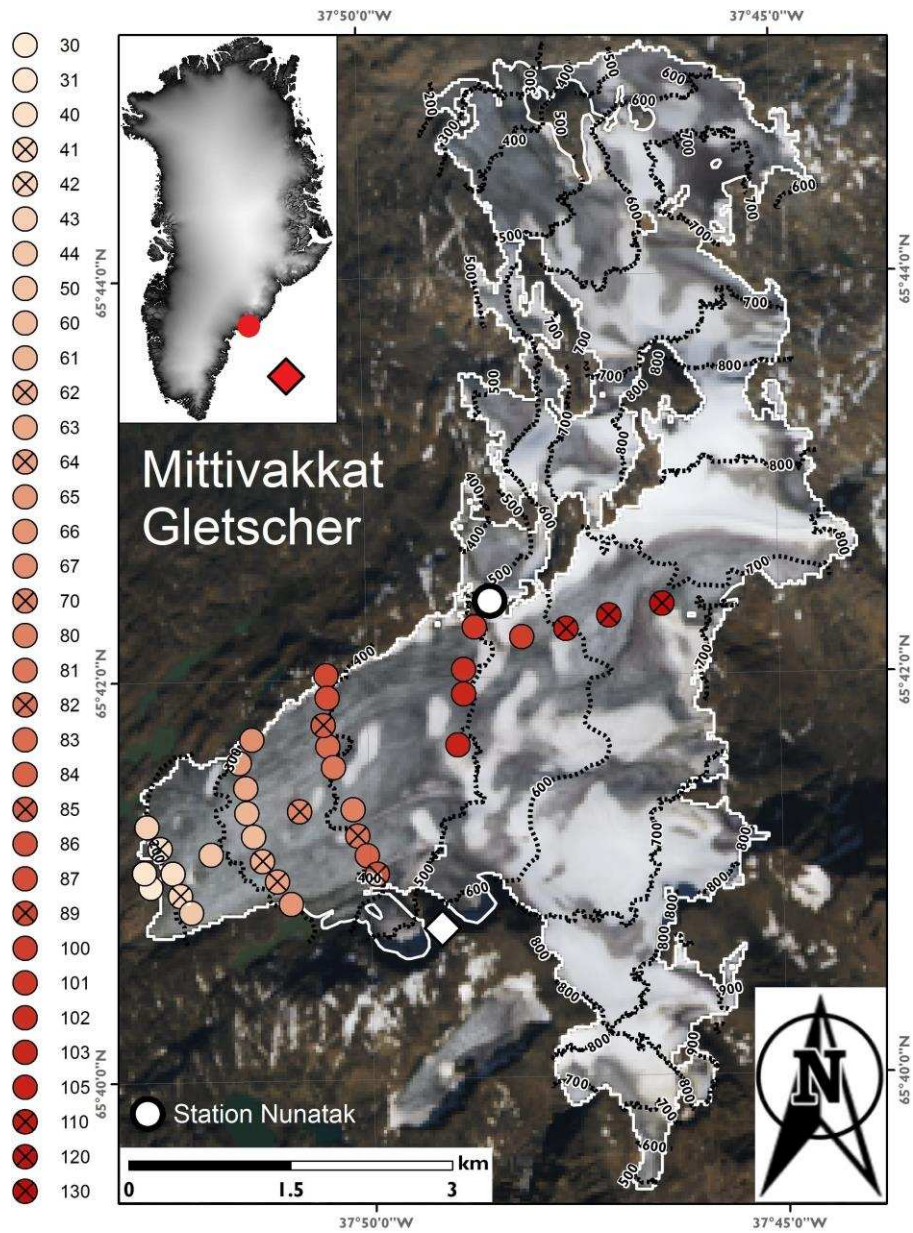
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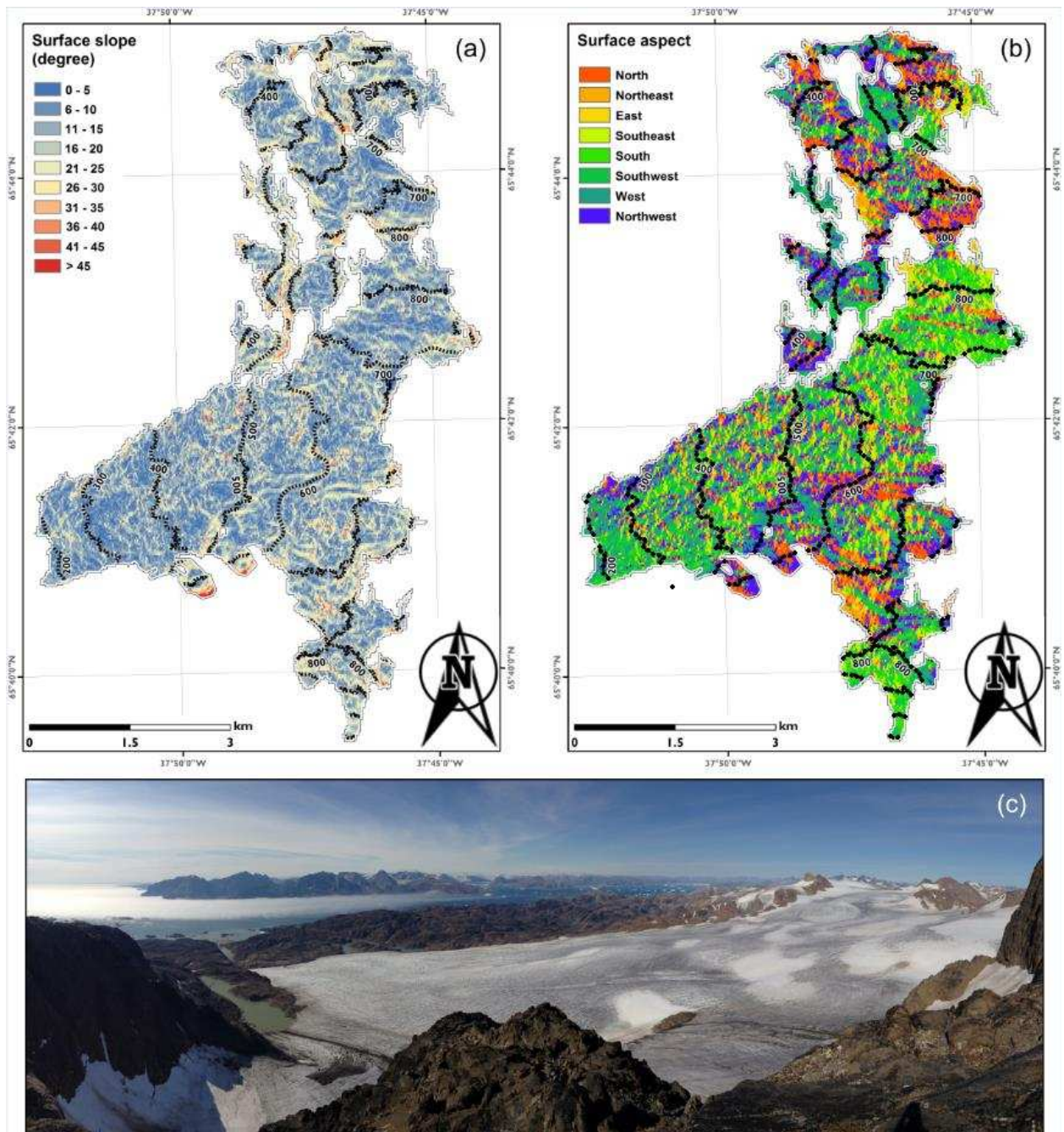
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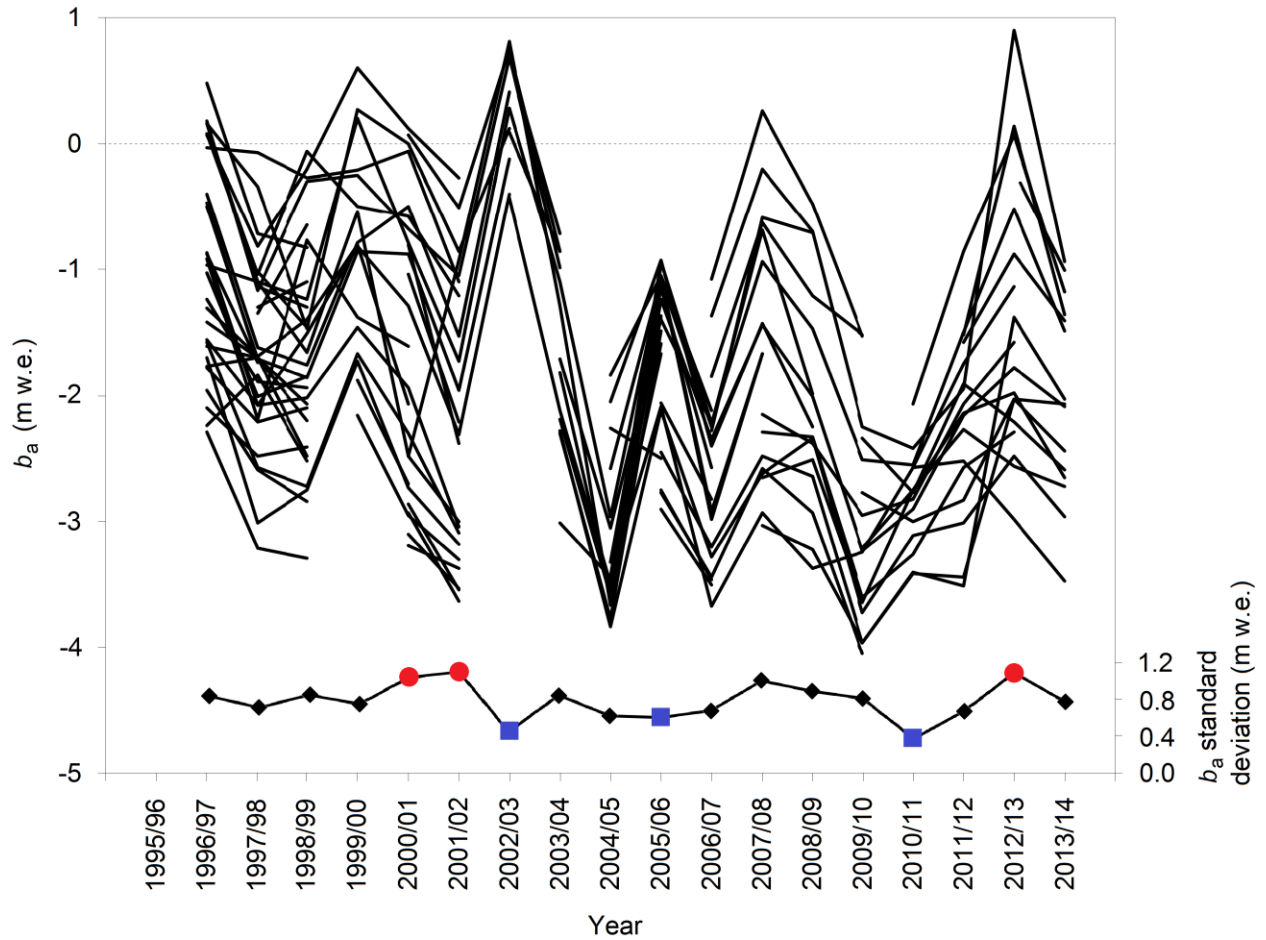


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 712 **Figure 1:** Mittivakkat Gletscher (26.2 km<sup>2</sup> (2011); 65°41'N, 37°48'W) topographic map (100-m  
 713 contour interval). The colored circles illustrate the 34 stake locations for the ongoing glacier  
 714 mass-balance observation program, 1995/96–2013/14. The stake colors on the glacier surface  
 715 correspond to the stake numbers illustrated to the left, where the low numbers correspond to the  
 716 stakes at the low-elevation part of the glacier. Stakes in anti-phase for the EOF2 analysis are  
 717 shown with a cross. Station Nunatak (515 m a.s.l.) is illustrated with a white circle. Station  
 718 Tasiilaq is not included as it is located ~10 km southeast of the glacier. The white diamond  
 719 indicates the location where the photo in Figure 2c was taken. Source: Landsat 8, OLI  
 720 (Operational Land Imager), 7 August 2014.





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 722 **Figure 2:** Mittivakkat Gletscher: (a) surface slope; (b) surface aspect; and (c) photo of the  
 723 southwestern part of the glacier (the photo was taken looking north northwest, and in the  
 724 background is Sermilik Fjord and the Greenland Ice Sheet, photo M. Lidström, August 2014).  
 725 Area values for each slope and aspect interval are shown in Table 1. Surface slope and aspect is  
 726 constructed based on the Advanced Spaceborne Thermal Emission and Reflection Radiometer  
 727 (ASTER) Global Digital Elevation Model Version 2 (GDEM v2). The glacier margin outline is  
 728 from 7 August 2014.



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730 **Figure 3:** Time series of individual observed  $b_a$  time series from Mittivakkat Gletscher for the  
 731 period 1996/97–2013/14, including calculated annual variability of  $b_a$  (at the bottom). The three  
 732 maximum and minimum annual variabilities are marked with red circles and blue squares,  
 733 respectively.

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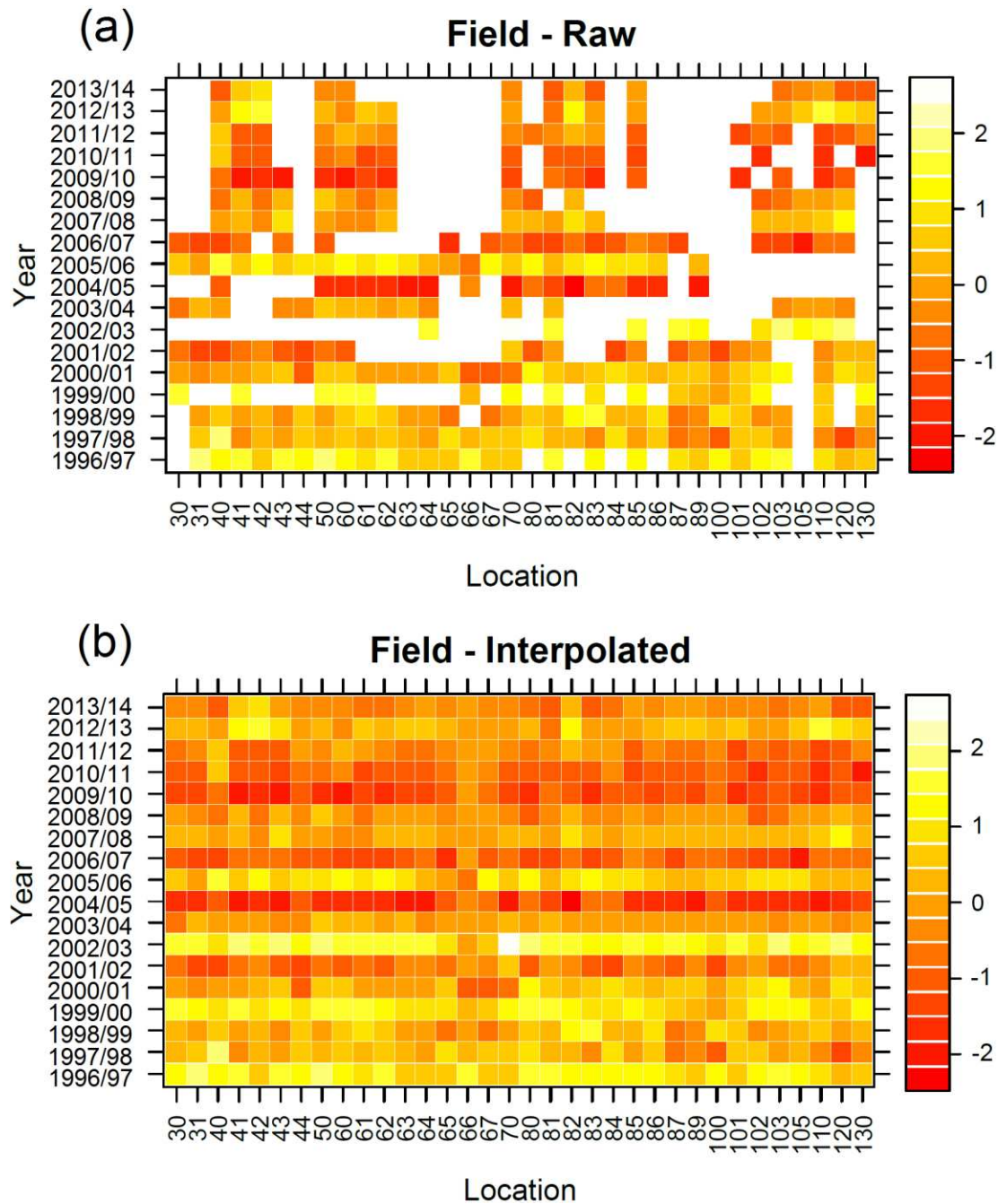
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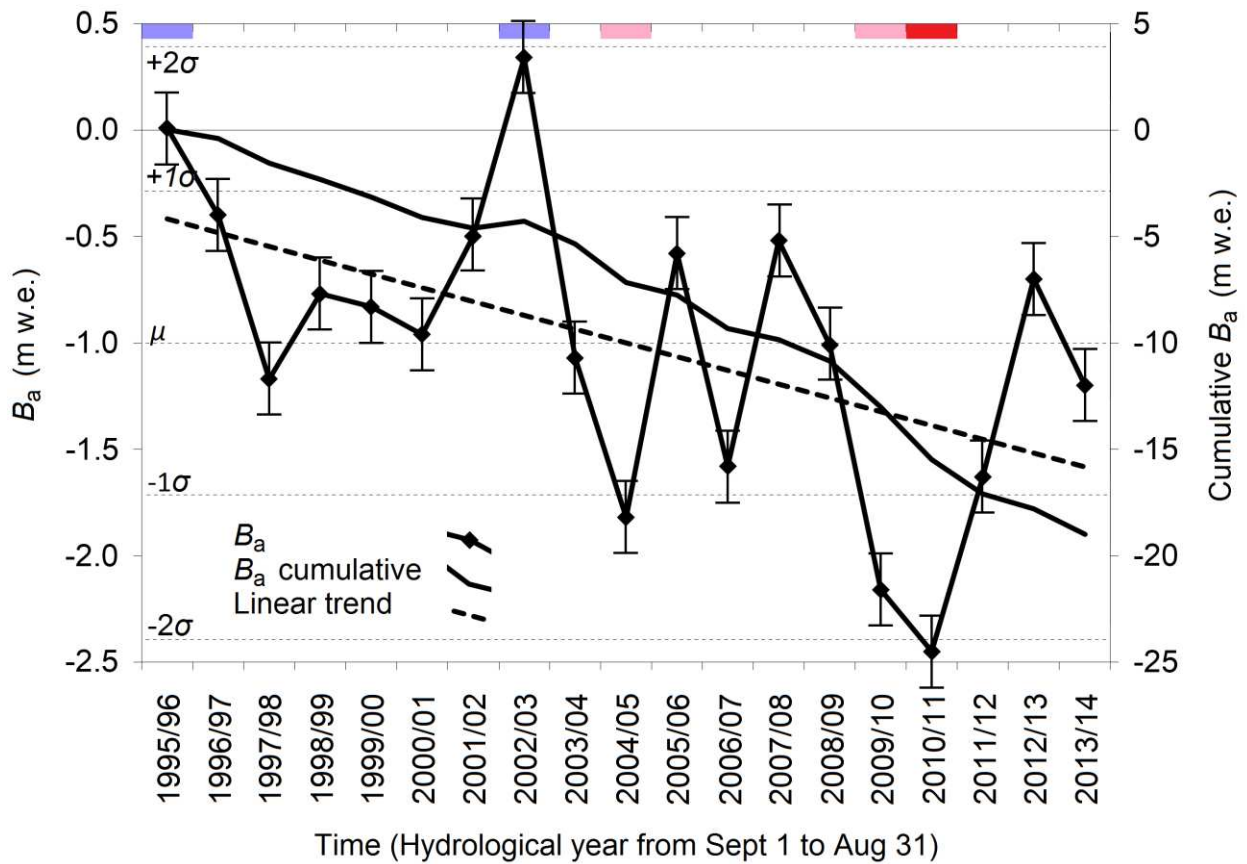
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743 **Figure 4:** Space-time field (1996/97–2013/14) for Mittivakkat Gletscher, where the stake  
 744 locations from left to right go from lowest stake numbers to highest stake number: (a) raw data  
 745 field (the white squares equal no data): and (b) reconstructed data field. The reconstructed field is  
 746 the mean of replicate reconstructions using the DINEOF method.

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751 **Figure 5:** Observed  $B_a$ , cumulative  $B_a$ , and linear trend for  $B_a$  for Mittivakkat Gletscher  
 752 (1995/96–2013/14). The colors light blue and pink indicated  $B_a$  values one standard deviation  
 753 above or below mean, respectively, and red color two standard deviations below mean  $B_a$ . For  
 754 each  $B_a$  the errors from measurements and analytics are added.

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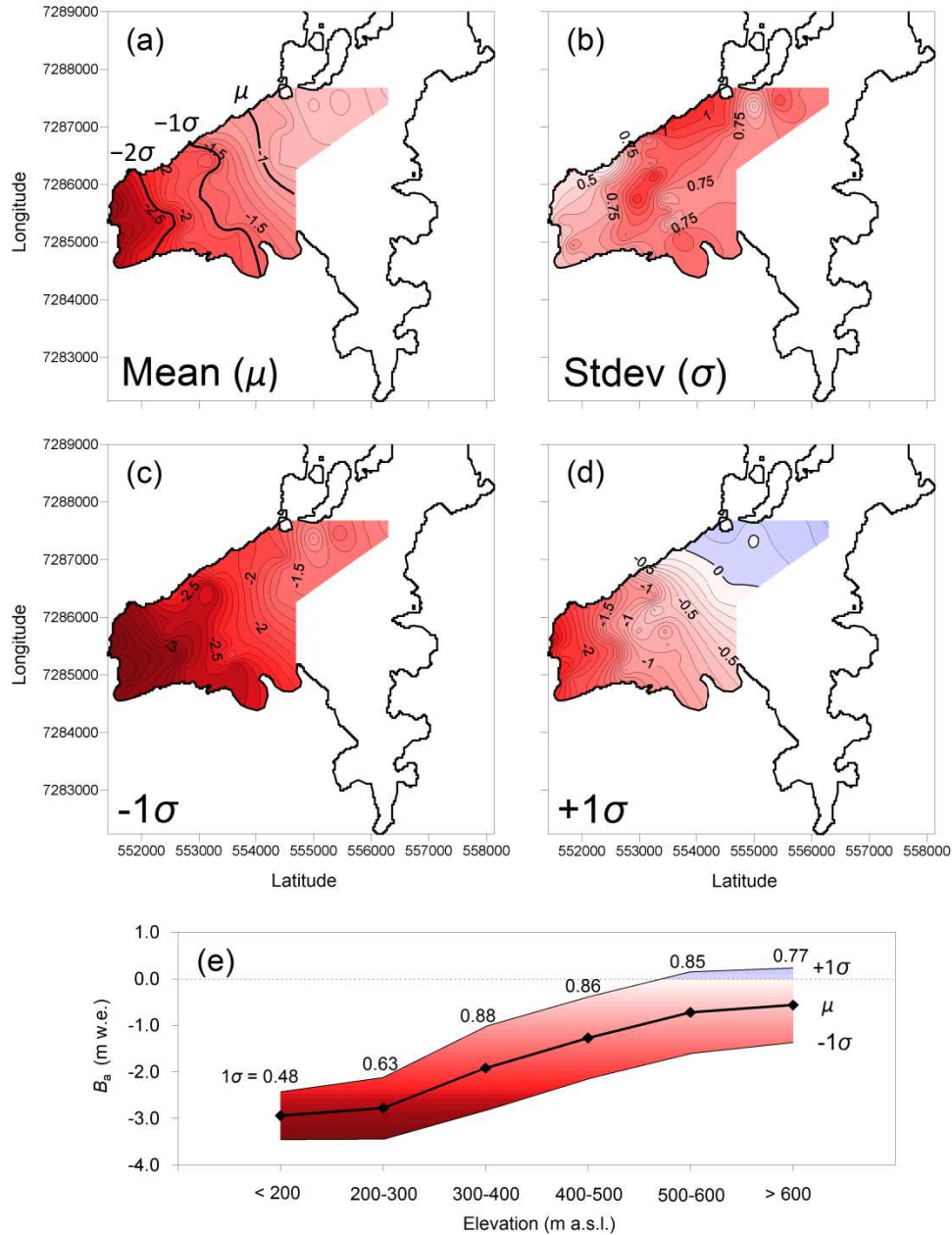
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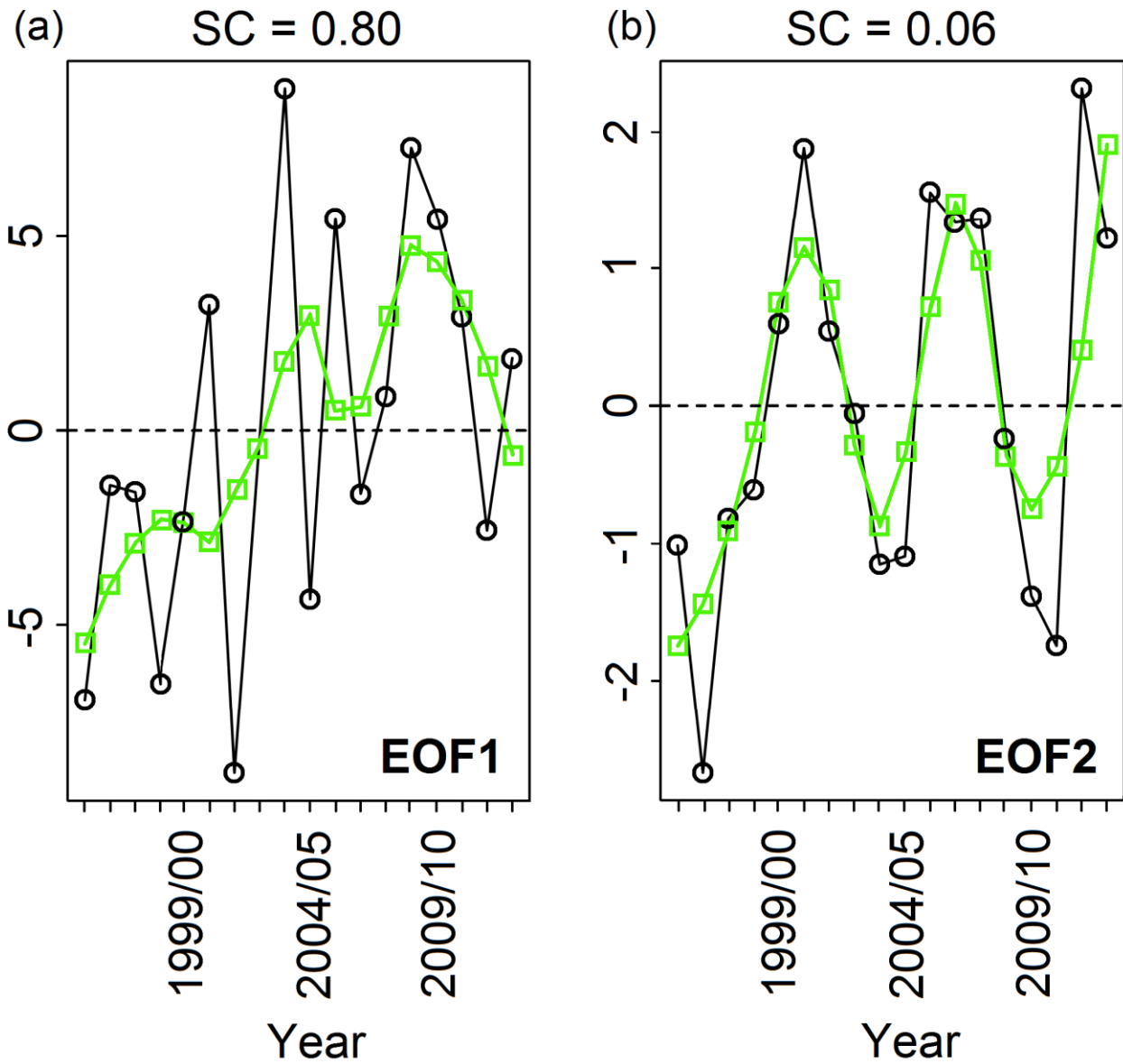


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 767 **Figure 6:** Observed Mittivakkat Gletscher (1996/97–2013/14): (a) spatial mean  $b_a$  ( $\mu$ ) where one  
 768 ( $-1\sigma$ ) and two ( $-2\sigma$ ) standard deviations are illustrated with bold lines; (b) spatial distribution of  
 769 one standard deviation ( $\sigma$ ); (c) spatial  $b_a$  minus one standard deviation ( $-1\sigma$ ); (d) spatial  $b_a$  plus  
 770 one standard deviation ( $+1\sigma$ ); and (e) longitudinal  $b_a$  profile with plus and minus one standard  
 771 deviation.

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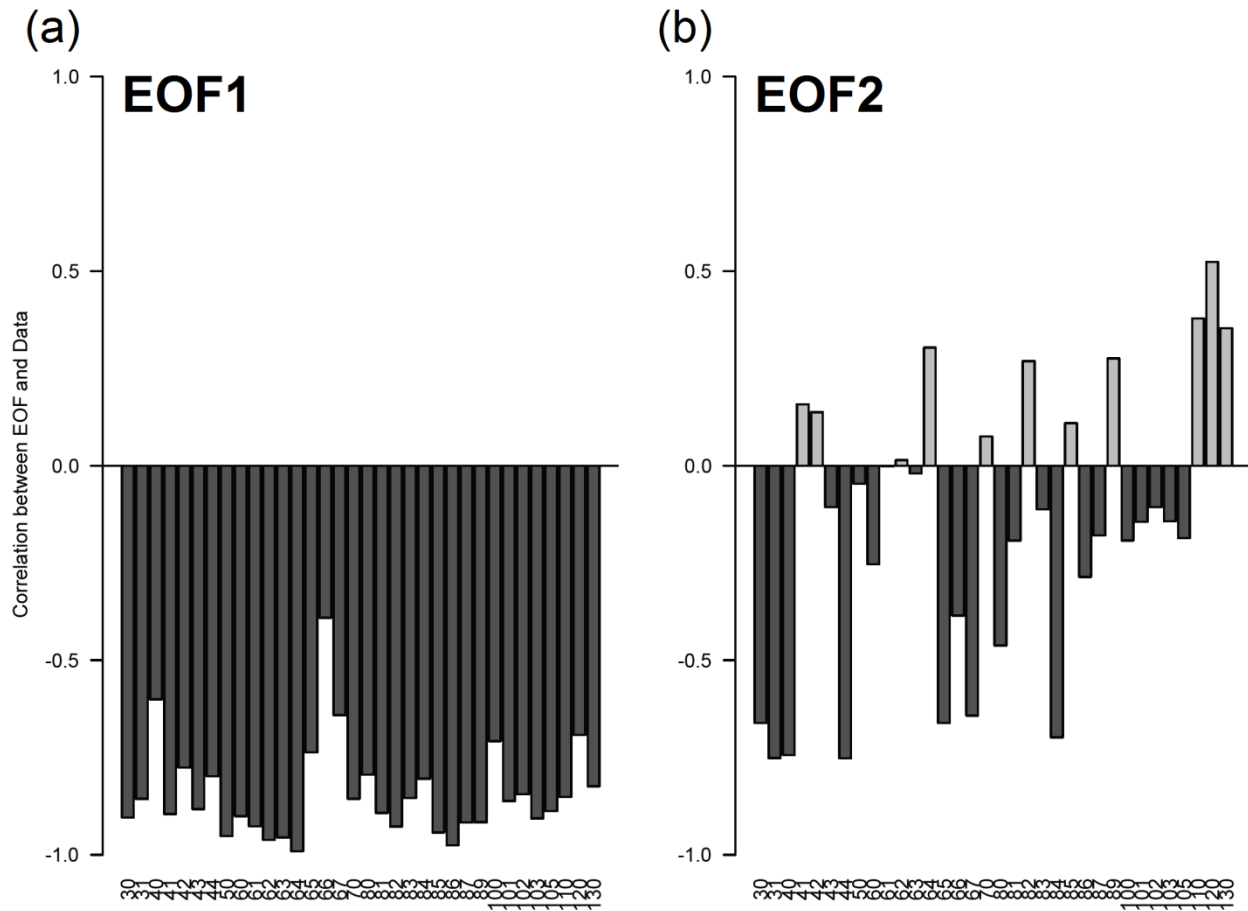
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776 **Figure 7:** Mass-balance time series (1996/97–2013/14) based on the empirical orthogonal  
 777 functions: (a) EOF1; and (b) EOF2. The explained square covariance (SC) is shown for each  
 778 EOF. The green line is a five running mean smoothing line.

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781 **Figure 8:** Eigenvector correlation values for each individual site for: (a) EOF1, and (b) EOF2.

782 Locations from left to right go from stake 30 to stake 130.

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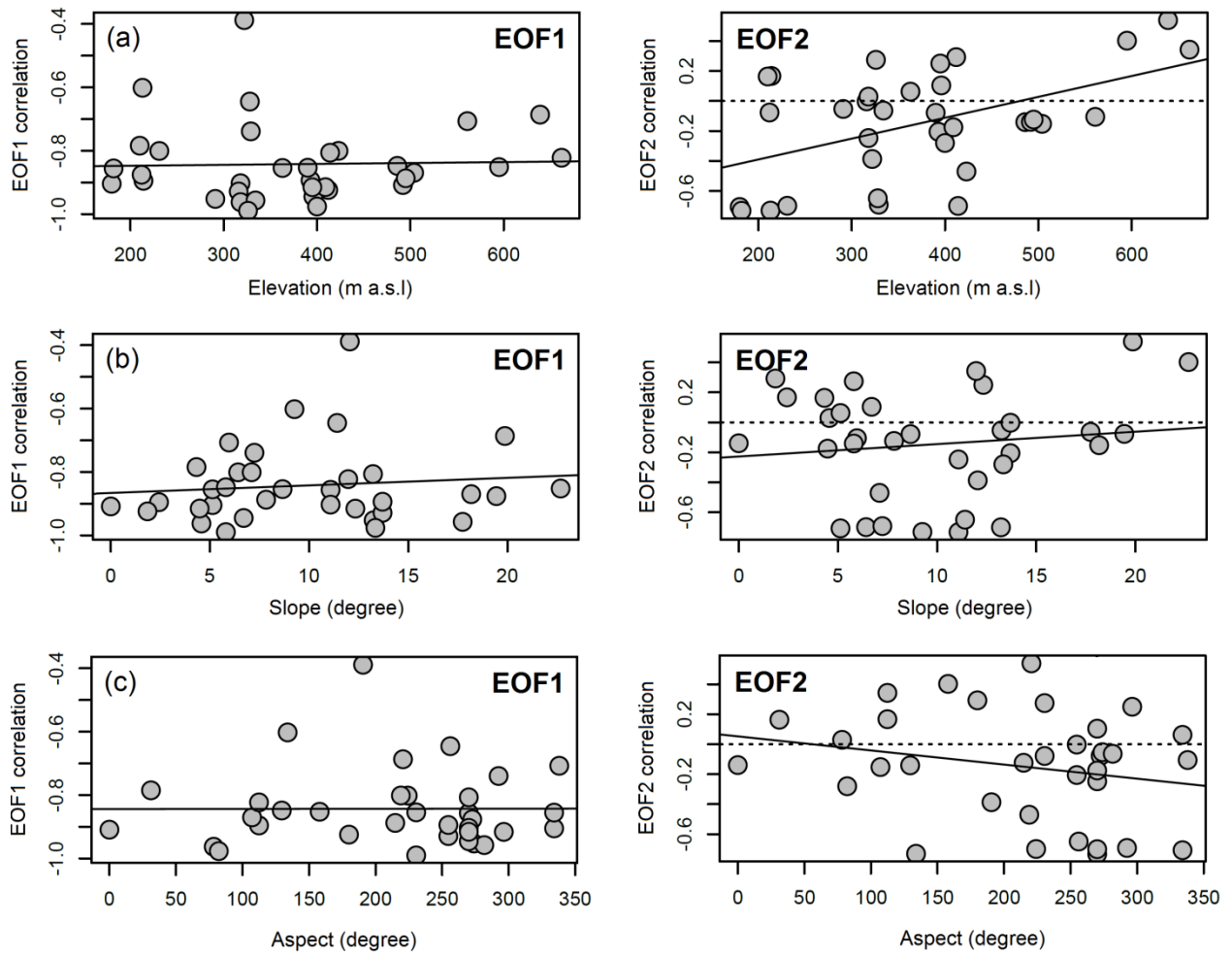
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794 **Figure 9:** EOF1 and EOF2 correlations between: (a) stake elevation; (b) stake surface slope; and

795 (c) stake surface aspect.

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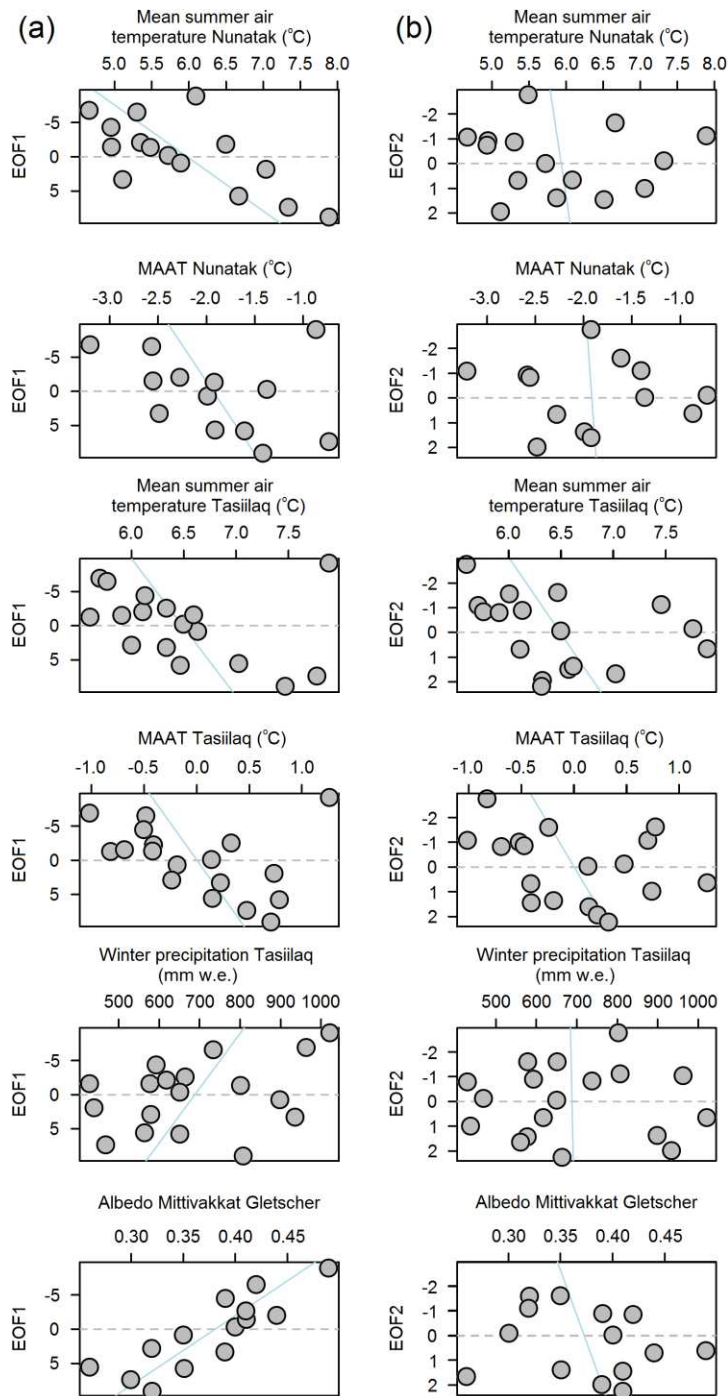
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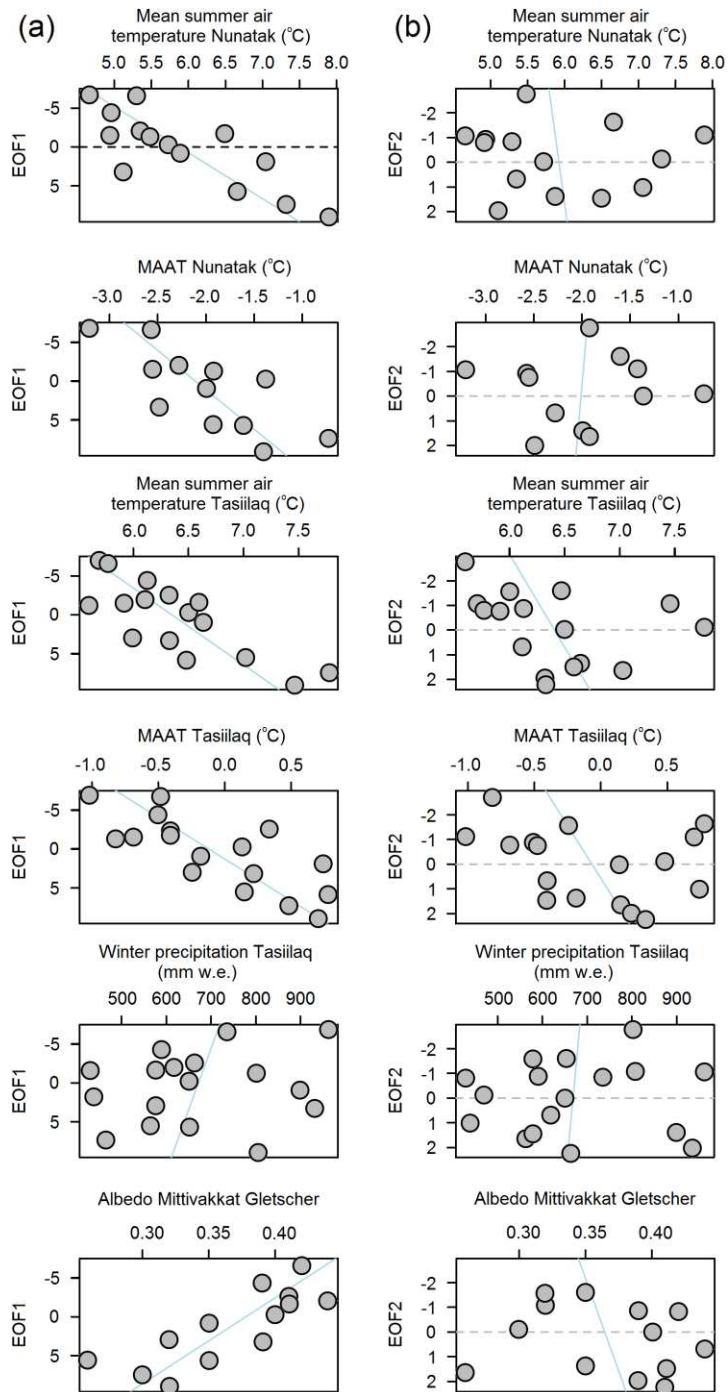
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804 **Figure 10:** (a) EOF1; and (b) EOF2 correlations between Station Nunatak mean summer air  
 805 temperature, Station Nunatak MAAT, Station Tasiilaq mean summer air temperature, Station  
 806 Tasiilaq MAAT, Station Tasiilaq winter precipitation, and Mittivakkat Gletscher mean glacier-  
 807 wide surface albedo by end of the mass-balance year for the 16-days mean period 27/28 July–  
 808 12/13 August (Mernild et al. 2015b).



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810 **Figure 11:** (a) EOF1; and EOF2 correlations without the statistical outlier from 2002 between  
 811 Station Nunatak mean summer air temperature, Station Nunatak MAAT, Station Tasiilaq mean  
 812 summer air temperature, Station Tasiilaq MAAT, Station Tasiilaq winter precipitation, and  
 813 Mittivakkat Gletscher mean glacier-wide surface albedo by the end of the mass-balance year for  
 814 the 16-days mean period 27/18 July–12/13 August (Mernild et al. 2015b).



815 **Table 1:** Mittivakkat Gletscher surface slope and aspect estimated from ASTER GDEM v2.

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| Slope (degrees) | Percentage of area and area   | Aspect    | Percentage of area and area   |
|-----------------|-------------------------------|-----------|-------------------------------|
| 0–5             | 8.3 % (2.2 km <sup>2</sup> )  | North     | 11.5 % (3.0 km <sup>2</sup> ) |
| 6–10            | 18.7 % (4.9 km <sup>2</sup> ) | Northeast | 9.8 % (2.6 km <sup>2</sup> )  |
| 11–15           | 20.0 % (5.2 km <sup>2</sup> ) | East      | 11.3 % (3.0 km <sup>2</sup> ) |
| 16–20           | 18.2 % (4.8 km <sup>2</sup> ) | Southeast | 8.8 % (2.2 km <sup>2</sup> )  |
| 21–25           | 12.0 % (3.1 km <sup>2</sup> ) | South     | 11.2 % (2.9 km <sup>2</sup> ) |
| 26–30           | 8.6 % (2.3 km <sup>2</sup> )  | Southwest | 13.3 % (3.5 km <sup>2</sup> ) |
| 31–35           | 5.6 % (1.4 km <sup>2</sup> )  | West      | 19.7 % (5.2 km <sup>2</sup> ) |
| 36–40           | 3.6 % (0.9 km <sup>2</sup> )  |           |                               |
| 41–45           | 2.2 % (0.6 km <sup>2</sup> )  | Northwest | 14.4 % (3.8 km <sup>2</sup> ) |
| >45             | 2.9 % (0.8 km <sup>2</sup> )  |           |                               |

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837 **Table 2:** Statistical relationships between EOF1 and EOF2 and meteorological variables  
838 observed near MG and MG mean glacier-wide surface albedo. The lower part of the table is  
839 without the statistical outlier from 2002. The letter S = significant, and InS = insignificant.

| All data  |  |        |         |         |               |
|---|--|--------|---------|---------|---------------|
| Relationship                                      |  | Slope  | F-value | P-value | $p \leq 0.05$ |
| EOF1  | Mean summer air temperature (Station Nunatak)  | 5.98   | 12.37   | 0.00    | S             |
|   | MAAT (Station Nunatak)                         | -1.93  | 1.81    | 0.21    | InS           |
|   | Mean summer air temperature (Station Tasiilaq) | 6.49   | 2.38    | 0.14    | InS           |
|   | MAAT (Station Tasiilaq)                        | 0.00   | 2.56    | 0.13    | InS           |
|   | Winter precipitation sum (Station Tasiilaq)    | 688.22 | 2.19    | 0.16    | InS           |
|   | Mean glacier-wide surface albedo               | 0.38   | 32.28   | 0.00    | S             |
| EOF2  | Mean summer air temperature (Station Nunatak)  | 5.93   | 0.06    | 0.81    | InS           |
|   | MAAT (Station Nunatak)                         | -1.92  | 1.94    | 0.18    | InS           |
|   | Mean summer air temperature (Station Tasiilaq) | 6.49   | 1.94    | 0.18    | InS           |
|   | MAAT (Station Tasiilaq)                        | 0.00   | 1.79    | 0.20    | InS           |
|   | Winter precipitation sum (Station Tasiilaq)    | 688.22 | 0.00    | 0.97    | InS           |
|   | Mean glacier-wide surface albedo               | 0.37   | 0.44    | 0.52    | InS           |
| All data except the statistical outlier from 2002 |  |        |         |         |               |
| EOF1  | Mean summer air temperature (Station Nunatak)  | 5.88   | 23.11   | 0.00    | S             |
|   | MAAT (Station Nunatak)                         | -2.11  | 13.74   | 0.00    | S             |
|   | Mean summer air temperature (Station Tasiilaq) | 6.35   | 23.59   | 0.00    | S             |
|   | MAAT (Station Tasiilaq)                        | -0.12  | 21.63   | 0.00    | S             |
|   | Winter precipitation sum (Station Tasiilaq)    | 672.09 | 0.49    | 0.50    | InS           |
|   | Mean glacier-wide surface albedo               | 0.38   | 19.21   | 0.00    | S             |
| EOF2  | Mean summer air temperature (Station Nunatak)  | 5.92   | 0.05    | 0.84    | InS           |
|   | MAAT (Station Nunatak)                         | -2.01  | 1.64    | 0.22    | InS           |
|   | Mean summer air temperature (Station Tasiilaq) | 6.41   | 1.64    | 0.22    | InS           |
|   | MAAT (Station Tasiilaq)                        | -0.07  | 1.53    | 0.24    | InS           |
|   | Winter precipitation sum (Station Tasiilaq)    | 668.51 | 0.03    | 0.86    | InS           |
|   | Mean glacier-wide surface albedo               | 0.37   | 0.33    | 0.58    | InS           |

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