1 Recent Precipitation Decrease Across the Western Greenland Ice Sheet
2 Percolation Zone

3 Lewis, Gabriel[1]; Osterberg, Erich[1]; Hawley, Robert[1]; Marshall, Hans Peter[2]; Meehan, Tate[2]; Graeter, Karina[3];
4 McCarthy, Forrest[4]; Overly, Thomas[5,6]; Thundercloud, Zayta[1]; Ferris, David[1]

5 [1]Department of Earth Sciences, Dartmouth College, Hanover, NH, USA
6 [2]Geosciences Department, Boise State University, Boise, ID, USA
7 [3]Office of Sustainability, University of Maine, Orono, ME, USA
8 [4]College of Fisheries and Ocean Sciences, University of Alaska Fairbanks, Fairbanks, AK, USA
9 [5]NASA Cryospheric Sciences Laboratory, NASA Goddard Space Flight Center, Greenbelt, MD, USA
10 [6]Earth System Science Interdisciplinary Center (ESSIC), University of Maryland, College Park, MD, USA

11 *Correspondence to*: Gabriel Lewis (Gabriel.M.Lewis.GR@dartmouth.edu)

12 **Abstract**

13 The mass balance of the Greenland Ice Sheet (GrIS) in a warming climate is of critical interest to scientists

14 and the general public in the context of future sea-level rise. Increased melting in the GrIS percolation zone

15 due to atmospheric warming over the past several decades has led to increased mass loss at lower elevations.

16 Previous studies have hypothesized that this warming is accompanied by a precipitation increase, as would

17 be expected from the Clausius-Clapeyron relationship, negating some of the melt-induced mass loss

18 throughout the Western GrIS. This study tests that hypothesis by calculating snow accumulation rates and

19 trends across the Western GrIS percolation zone, providing new critical accumulation estimates in regions

20 with sparse and/or dated *in situ* data for calibration of future regional climate models. We present

21 accumulation records from sixteen 22 – 32 m long firn cores and 4436 km of ground penetrating-radar,

22 covering the past 20 – 60 years of accumulation, collected across the Western GrIS percolation zone as part

23 of the Greenland Traverse for Accumulation and Climate Studies (GreenTrACS) project. Trends from both

24 radar and firn cores, as well as commonly used regional climate models, show decreasing accumulation and

25 precipitation over the 1996 – 2016 period, which we attribute to shifting storm-tracks related to stronger

26 atmospheric summer blocking over Greenland. Changes in atmospheric circulation over the past 20 years,

27 specifically anomalously high summertime blocking, have reduced GrIS surface mass balance through both

28 an increase in surface melting and a decrease in accumulation.

29 1. Introduction

30 Greenland Ice Sheet (GrIS) mass loss has accelerated over the past few decades, with modern mass loss rates

31 more than double that from Antarctica (van den Broeke et al., 2016). The 2010-2018 GrIS mass loss is 286

32 $\pm$ 20 Gt a^{-1} (Mouginot et al., 2019), contributing 0.7 $\pm$ 0.2 mm a^{-1} of sea level rise. Over the past 20 years, the

33 largest warming rates (Hanna et al., 2012) and fastest mass loss have occurred in Western Greenland (26 $\pm$ 7

34   GT a^{-2} in basins F + G of Sasgen et al., 2012), where surface mass balance (SMB) has decreased between

35   31.1% (European Centre for Medium Range Weather Forecasting downscaled; ECMWFd) and 76.5%

36   (Modèle Atmosphérique Régional; MAR) over the 1996 – 2008 period (Vernon et al., 2013) due to higher

37   surface melt and runoff (van den Broeke et al., 2009, 2016). Modern surface melt rates are at their highest

38   levels of at least the last 450 years across Western Greenland (Graeter et al., 2018) and more broadly

39   throughout Greenland (Trusel et al., 2018). In particular, ice core records from Western Greenland show an

40   abrupt increase in surface melt rates beginning in the middle-late 1990's due to a combination of higher North

41   Atlantic sea surface temperatures, enhanced summertime blocking highs, and anthropogenic warming

42   (Graeter et al., 2018).

43

44   Enhanced GrIS surface melt is driven fundamentally by rising Greenland temperatures as recorded by

45   automated weather stations from the Greenland Climate Network (GC-Net; Steffen and Box, 2001), coastal

46   weather stations (Box, 2002), borehole thermometry (Polashenski et al., 2014), remote sensing (Hall et al.,

47   2008), and ice core stable isotopes (Buchardt et al., 2012). Average annual temperature across interior

48   Greenland increased by 0.055 ± 0.044 °C a^{-1} from 2000 – 2012 (Hall et al., 2013), with summer trends

49   upwards of 0.135 ± 0.047 °C a^{-1} (Hall et al., 2013; Reeves Eyre and Zeng, 2017). These warming trends

50   extend to the highest elevations of the ice sheet, with warming at Summit Station of 0.09 ± 0.01 °C a^{-1} from

51   1982 – 2011 (McGrath et al., 2013). Nearly every Greenland dataset shows statistically significant positive

52   temperature trends in recent decades, especially during the summers (Reeves Eyre and Zeng, 2017).

53

54   Basic physics implies that rising temperatures should cause an increase in accumulation over the ice sheet

55   due to the Clausius-Clapeyron relationship – warmer air has a higher saturation vapor pressure, potentially

56   leading to more precipitation (Box et al., 2006; Buchardt et al., 2012). The Coupled Model Intercomparison

57   Project, phase 5 (CMIP5) predicts precipitation increases of 20 – 50% over the GrIS by the end of the 21st

58   century (Bintanja and Selten, 2014), partially offsetting mass loss and sea-level rise from enhanced summer

59   melt and runoff. However, most *in situ* records of Greenland snow accumulation do not span the modern

60   period of rapid warming and accelerating mass loss since the mid-1990s. It is difficult to determine whether

61   accumulation has been increasing with the observed warming temperatures as predicted. For example, the

62   Program for Arctic Regional Climate Assessment (PARCA) campaign collected 46 shallow ice and firn cores

63   (<100 m long) and three deeper cores (120 – 152 m long) to capture the spatial and temporal accumulation

64   variability over the ice sheet (Mosley-Thompson et al., 2001). However, the PARCA firn cores were collected

65   in 1997 – 1998, just at the onset of accelerated surface melting (Graeter et al., 2018), and the precipitation

66   record from automated weather stations is too brief and localized to analyze significant precipitation trends

67   (Rennermalm et al., 2013). The most recently analyzed deep ice cores (over 100 m long) were collected in

68   2003 – 2004 (D4, D5, Sandy, Katie; Banta and McConnell, 2007) and there have been no published *in situ*

69   accumulation records from the Western GrIS percolation zone for the past decade. Updated *in situ* snow

70   accumulation data are needed from this region to assess recent changes in accumulation during this period of

71   warming and SMB loss from melt and runoff.

72

73   In addition to measuring snow accumulation with ice cores and automated snow depth sensors, several

74   previous studies have used ground-based and airborne radar to calculate GrIS accumulation and trends (e.g.

75   Medley et al., 2013; Spikes et al., 2004). For example, Hawley et al. (2014) found a 10% increase in northwest

76   coastal Greenland accumulation over the past 52 years using Ground Penetrating Radar (GPR) along the

77   Greenland Inland Traverse (GrIT; see Figure 1 for location), although they did not find any statistically

78   significant trends further inland between the North Greenland Eemian Ice Drilling (NEEM) site and Summit

79   Station. Similarly, Wong et al. (2015) show a coastal increase in precipitation near Thule over 1981 – 2012,

80   but no statistically significant change in precipitation rate further inland at the Camp Century, B26, or 2Barrel

81   ice core sites. Overly et al. (2016) found a 20% accumulation increase below 3000 m a.s.l. on the historic

82   Expéditions Glaciologiques Internationales au Groenland (EGIG) line from 1994 – 2004 vs. 1985 – 1994

83   using the Airborne SAR/Interferometric Radar Altimeter System (ASIRAS) radar. We build upon these

84   previous studies by collecting GPR data across the lower percolation zone of Western Greenland, where

85   airborne radargrams are often obscured by refrozen melt percolation (Nghiem et al., 2005). By having our

86   GPR antenna coupled with the snow, we avoid losing energy, and, therefore, penetration depth, from a strong

87   reflection off of the snow-air interface.

88

89   In addition to temperature-precipitation relationships through the Clausius-Clapeyron relationship, previous

90   studies have analyzed the dynamic climate controls on Greenland precipitation. Mernild et al. (2014), Auger

91   et al. (2017), and Lewis et al. (2017) have hypothesized that a positive Atlantic Multidecadal Oscillation

92   (AMO) index correlates with rising accumulation over most of the GrIS interior, since higher sea surface

93   temperatures increase moisture flux over the GrIS and induce greater snowfall. High pressure (blocking)

94   systems east of Greenland tend to deflect eastward-moving storms over central Greenland and increase

95   precipitation, whereas blocking directly over Greenland or in Baffin Bay has the potential to prevent storms

96   from crossing the ice sheet (Auger et al., 2017). Through reanalysis data, Auger et al. (2017) showed that

97   persistent blocking highs increase precipitation in southwest Greenland and reduce precipitation in the

98   southeast.

99

100   The Greenland Blocking Index (GBI) quantifies blocking directly over Greenland and is defined as the mean

101   500 hPa geopotential height for the 60 – 80°N, 20 – 80°W region (Hanna et al., 2016). Over the 1991 – 2015
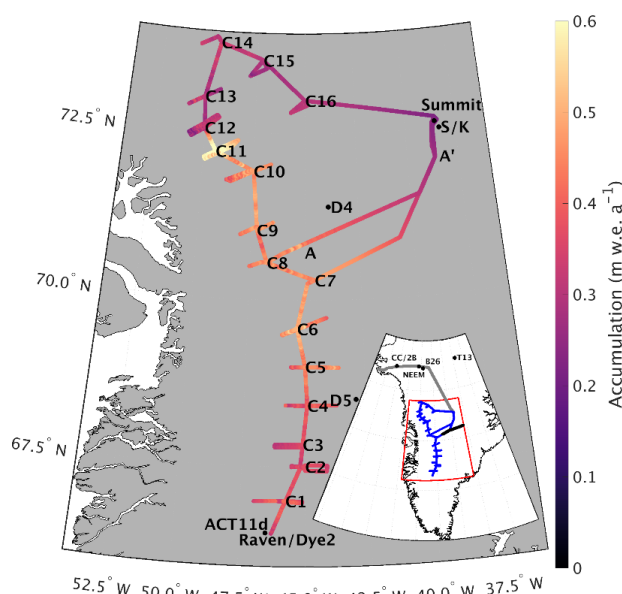
102   period there has been an especially high Greenland Blocking Index sustained throughout the summers (Hanna

103   et al., 2016). Alternatively, persistent blocking episodes have the potential to reduce snowfall accumulation

104   over the GrIS by displacing the polar jet stream and corresponding storm tracks equatorward, although this

105   relationship has not yet been documented *in situ*.

106

107   Here we develop new accumulation records across the Western GrIS percolation zone using sixteen firn cores

108   and 4436 km of GPR data collected during an over-ice traverse spanning two field seasons. We evaluate the

109   veracity of the accumulation data through comparisons of our firn core time series with previous

110   measurements. We quantify multi-year trends in accumulation across Western Greenland to test the

111   hypothesis that precipitation has recently increased from the Clausius-Clapeyron relationship and higher GrIS

112   temperatures. Further, we assess the ability of RCMs to capture the year-to-year variability and multi-year

113   trends in Western GrIS accumulation. Finally, we evaluate relationships between recent accumulation trends

114   and atmospheric circulation patterns, particularly changes in storm tracks.

115

116

117   **Figure 1. Average accumulation across the GreenTrACS traverse for the length of each record showing the location of each firn core,**
118   **ACT11d, D4, D5, Katie (K), Raven/Dye-2, and Sandy (S) ice cores, and Summit Station. Transect A-A' discussed in Section 3.3. Inset**
119   **shows locations of Camp Century (CC), 2Barrel (2B), NEEM, B26, and TUNU2013 (T13) ice cores, as well as locations of EGIG (black),**
120   **GrIT (grey), and GreenTrACS (blue) traverses.**

121   **2. Methods**

122   This study uses data from the 2016 – 2017 Greenland Traverse for Accumulation and Climate Studies

123   (GreenTrACS), which measured accumulation and melt across the Western GrIS percolation zone over two

124 summer snowmobile traverses (closely following the 2150 m a.s.l. elevation contour). The May-June 2016

125 season traversed 860 km from Raven/Dye-2 northward to Summit Station and the May – June 2017 traverse

126 made a 1230 km clockwise loop starting and ending at Summit Station (Figure 1). This manuscript focuses

127 on accumulation rates derived from 400 MHz GPR data collected along the entire traverse path, as well as

128 sixteen shallow (22 – 32 m long) firn cores spaced 40 – 100 km apart along the backbone of the traverse

129 (Figure 1). Firn Cores 1 – 7 were collected in 2016 and Cores 8 – 16 were collected in 2017. We returned to

130 the Core 7 location at the beginning of the 2017 traverse to recover a weather station and to connect the two

131 season's GPR data. Additionally, we collected GPR data ~30 – 70 km east and west of each core site,

132 hereafter called "spurs", to measure accumulation changes along strong elevation gradients (see Figure 1).

133   **2.1. GPS Positioning**

134 During the 2016 traverse we collected GPS data using a Trimble NetR8 reference receiver with a Zephyr

135 Geodetic antenna mounted to a Nansen sled ~5 m in front of the GPR antenna. For each spur and the tail

136 ends of each transect between core sites we performed differential corrections to the GPS data using RTKLIB

137 2.4.1 and a Trimble NetR8 base station near the core site. Between spurs, when not operating a base station,

138 we post-processed GPS data in precise point positioning mode (Zumberge et al., 1997). Estimated root-mean-

139 square horizontal errors were generally between 13 and 18 cm from standard deviations calculated during

140 stationary periods at the end of spurs. To co-register GPR and GPS data, we used time stamps embedded in

141 the two data streams and locations where we stopped to save GPR files, approximately every 15 km. The

142 time drift in the GPR logger is negligible over these durations.

143

144 During the 2017 traverse we used GPS data from a Garmin 19x GPS receiver wired directly to the GPR

145 instrument, which recorded position data at every radar sample with RMS values of 3 m. During radar

146 processing we average 75 adjacent traces, corresponding to a distance of ~20 m, so errors in GPS positioning

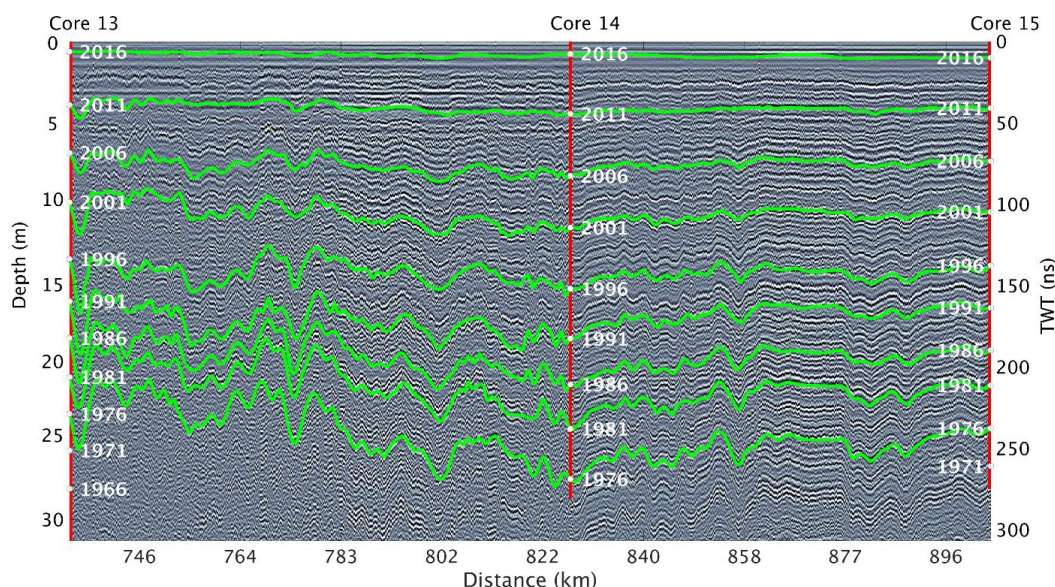147 have a negligible effect on the final dataset.

148

149   **2.2. Ground-penetrating radar**

150 We develop a spatially continuous record of accumulation using GPR profiles collected with Geophysical

151 Survey Systems Inc. (GSSI) SIR-3000 (during 2016) and SIR-30 (during 2017) radar units with a 400 MHz

152 antenna (following Hawley et al., 2014). The antenna was towed on the snow surface in a small plastic sled

153 ~5 m behind a wooden Nansen sled and ~15 m behind a snow machine. We recorded 2048 samples (2016)

154 and 4096 samples (2017) per trace over a range window of 800 ns (Figure 2). At a relative permittivity of

155 1.26 ± 0.07, typical of firn in the GrIS percolation zone, the range was ~111 – 114 m. The 400 MHz short-

156   pulse radar has a range resolution (ability to resolve distinct features) of 0.35 ± 0.1 m in firn, which is fine

157   enough to resolve Internal Reflecting Horizons (IRHs) that have been shown to represent isochrones (Medley

158   et al., 2013; Rodriguez-Morales et al., 2014; Spikes et al., 2004; Hawley et al., 2014). We recorded 10 traces

159   per second, which at the snowmobile's average travel speed of approximately 2.75 m s^{-1} results in ~3.6 traces

160   recorded per meter. Note that this spacing between traces varies with vehicle speed.
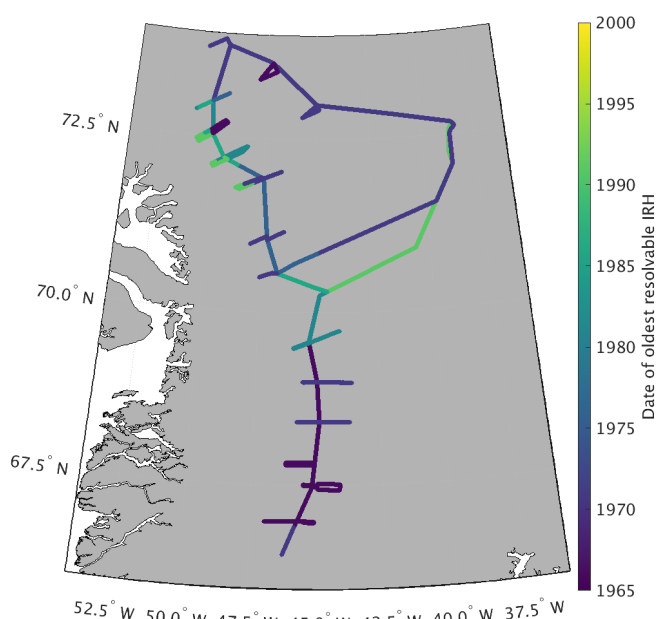

161

162   **Figure 2. Radargram showing the top 32 m of the transect along the main 2017 traverse from Core 13 to Core 15. Cores are indicated as**
163   **red lines down to their final depth, with dates plotted every 5 years at corresponding depths. Traced internal reflecting horizons are**
164   **shown as isochronous green lines. The depth scale on the vertical axis is calculated from the TWT-depth conversion (see Section 2.4) for**
165   **Core 13, although there is no visual difference in depth scale across this radargram.**

166   Depending on signal attenuation within the firn column, IRHs can be traced to a depth of 20 – 50 m (Figure

167   2), providing accumulation records over the past 20 – 60 years (Figure 3). For areas with high attenuation

168   (i.e. shallow penetration of the radar signal), such as lower elevation regions with more refrozen melt layers,

169   we calculate accumulation results for shorter time periods. We are not able to trace as many IRHs to the west

170   of Cores 10 – 13 compared to the east due to higher signal attenuation, resulting in slightly different average

171   accumulation values on either side of these cores (Figure 3). Likewise, we experienced an equipment

172   malfunction at the end of the 2016 traverse, reducing the number of observable IRHs from Core 7 to Summit

173   Station (Figure 3). We have less confidence in calculated accumulation throughout this section of the traverse

174   due to this malfunction, although the 2017 Summit to Core 8 interval overlaps nicely with the last 140 km of

175   the problematic 2016 interval, and provides high quality accumulation measurements for this section near

176   Summit Station.

177

178 We reduce the GPR data volume and signal noise by averaging 75 adjacent traces, which has the effect of

179 suppressing random noise by the principle of trace stacking (Yilmaz, 2001). We apply a combination of

180 median trace filtering, residual mean filtering (Gerlitz et al., 1993), and bandpass filtering using a butterworth

181 design (Selesnick and Sidney Burrus, 1998) between 200 – 800 MHz. For data visualization, we apply an

182 automatic gain control (Yilmaz, 2001) to give the interpreter more confidence when picking IRHs.

183

184

185

186 **Figure 3: Date of oldest resolvable internal reflecting horizon throughout the entire GreenTrACS traverse route. Anomalously young**
187 **ages from Core 7 to Summit are due to equipment malfunction.**

188

189   **2.3. Firn core processing and density profiles**

190 The amount of snow mass and the time span between IRHs are necessary to calculate accumulation rates.

191 The accumulation rate is a function of the depth-age scale, travel time-depth conversion rate, and the firn

192 density profile. We obtain the depth-age and depth-density scales from each of the shallow firn cores

193 collected along the GreenTrACS traverse, and from density models based on temperature and accumulation

194 rate data.

195

196 The sixteen firn cores were drilled using an Ice Drilling Program hand auger with a Kyne sidewinder

197 attachment (see Graeter et al., 2018). We sampled the firn cores for chemical measurements using a

7

198   continuous ice core melter system with discrete sampling (Osterberg et al., 2006). We used an Abakus (Klotz)

199   laser particle detector to measure microparticle concentrations and size distribution from the continuous ice

200   core meltwater stream, a Dionex Model ICS5000 capillary ion chromatograph to measure major ion (Na^+,

201   Mg^{2+}, Ca^{2+}, K^+, NH_4^+, Cl^-, NO_3^-, SO_4^{2-}) and methanesulfonic acid concentrations, and a Picarro L1102-I and

202   a Los Gatos Research Liquid Water Isotope Analyzer to measure oxygen and hydrogen isotope ratios ($\delta^{18}O$,

203   δD; Graeter et al., 2018).

204

227   We determine depth-age curves for each core by identifying annual layers based on seasonal oscillations in

228   $\delta^{18}O$ and the concentrations of major ions and dust, consistent with previous ice core studies in this region

229   (Graeter et al., 2018; Mosley-Thompson et al., 2001; Osterberg et al., 2015). We combine the depth-age

230   scales with measured density to calculate annual accumulation rates at the firn core sites.

231

232   At each firn core and at the ends of each spur, we measured the density in the top meter of snow using a 1000

233   cm^3 SnowMetrics cutter. To calculate firn core density profiles, we measured the mass, length, and diameter

234   of 0.03–1 m long core segments in the field and again after transporting the cores to the Dartmouth College

235   Ice Core Laboratory. To calculate accumulation rates at Raven/Dye-2, we use density data from a 119.6 m

236   long firn core collected in 1997 (Bales et al., 2009) and a 19.3 m long core collected from the same location

237   in 2015, which did not include accumulation data (Vandecrux et al., 2018). For this location we use the most

238   recent density data for the near-surface and the older densities for depths below the 2015 core. Likewise, we

239   use a density profile from a 109 m long firn core collected from Summit in 2010 (Mary Albert, personal

240   communication, 2015). We also incorporate density data from measurements along the EGIG traverse at T19,

241   T21, T23, T27, and T31 to improve the density profile between Core 7 and Summit (Morris and Wingham,

242   2014).

243

244   After collecting each firn core, we measured borehole temperature for 24 – 48 hours using a 20 m long

245   thermistor string. We estimate mean annual temperature from the deepest thermistor on the twenty-

246   thermistor-string. For the location of each firn core, we use the depth-density data from that core and then

247   calculate a Herron and Langway (1980) depth-density model for depths below the core using our measured

248   mean annual temperature, firn core mean annual accumulation, and top-meter snow density. Likewise, we

249   calculate Herron-Langway profiles for the ends of each spur using MODIS satellite derived mean annual

250   temperature (Hall et al., 2012), MAR modeled accumulation (Burgess et al., 2010), and the measured snow

251   density in the upper meter of each of the spur's snow pits. Finally, we interpolate depth-density profiles both

252   between firn cores and along radar spurs to estimate the depth-density matrix everywhere along our traverse
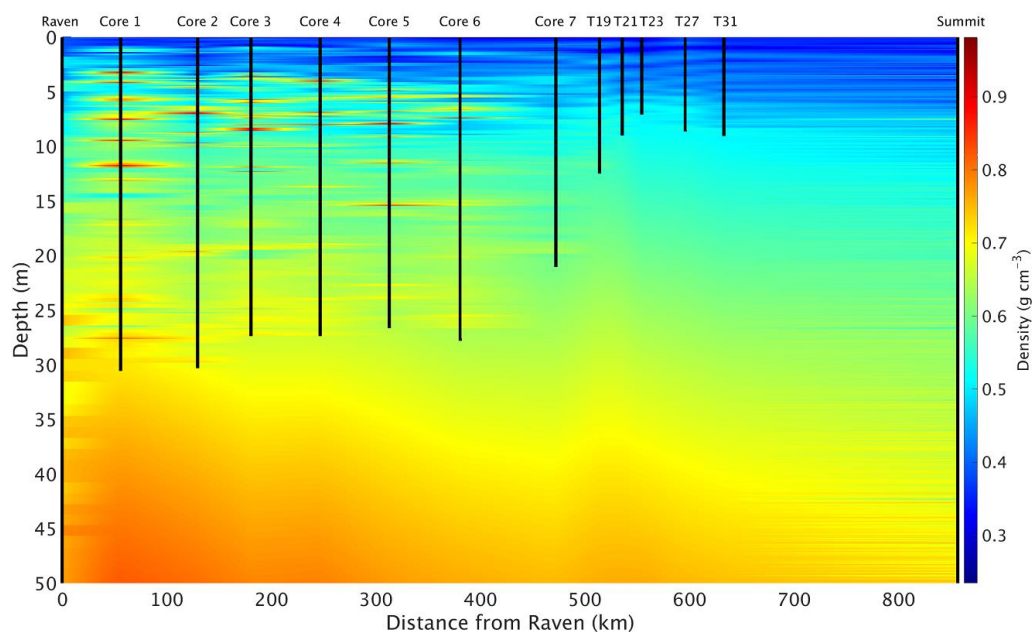
253   (Figure 4). Final accumulation rates are insensitive to the accumulation we use to calculate our Herron-

254   Langway models (Lewis et al., 2017).

255

256   As shown in Figure 4, ice layers within several firn cores are extrapolated laterally along the traverse,

257   although these dense lenses are typically both localized and heterogeneous in nature (Brown et al., 2011;

258   Rennermalm et al., 2013). This ice lens density interpolation is as accurate as possible between firn cores

259   without additional *in situ* data, and this estimation does not significantly alter our results, as discussed in

260   Section 2.6, since the ice layers represent a small fraction of the total depth to IRHs.

261

Figure 4. Depth-density profile along the main 2016 traverse used for calculation of electromagnetic wave velocity and accumulation in this study. Densities are linearly interpolated between the two nearest cores and are modeled using Herron-Langway profiles below the depth of each core. The left and right boundary data come from the Raven/Dye-2 and Summit firn cores, respectively. Ice layers in Cores 1 – 5 are clearly visible as red lenses, but their extent is, in reality, likely more localized.

266

2.4. Travel-time to depth conversion

268   We convert the radar travel time to depth by iteratively multiplying the velocity of the electromagnetic wave

269   by the signal's one-way travel time to each IRH. The electromagnetic speed of the radar wave, v (m s⁻¹), is

270   calculated from the dielectric permittivity, ε_r (dimensionless), and the speed of light in a vacuum, c ($3*10^8$ m

271   s⁻¹), from

272    $v = \frac{c}{\sqrt{\epsilon_r}}$ **(1).**

273 In turn, we calculate the dielectric permittivity for each radar trace from the density, ρ (g cm^{-3}), of snow and

274 ice at depth, as shown in Figure 4, for each radar trace at every range bin (following Kovacs et al., 1995) by

275 $\epsilon_r = (1.0 + 0.845 * \rho)^2$ (2).

276 We calculate the depth of each subsequent radar sample for each trace in the profile using the radar travel

277 time and velocity profile from equations 1 and 2, following Hawley et al. (2014) and Lewis et al. (2017).

278

2.5. Internal reflecting horizons

280 We manually select 10 clear, strong IRHs spaced approximately 5 years apart to consistently trace from

281 Raven/Dye-2 to Summit Station and throughout the 2017 main traverse (Figure 2). We trace each layer

282 manually by visually identifying strong amplitude peaks throughout the radargram, starting with the 2016

283 layer and working downwards. We use a spline interpolation between manual picks to trace each layer along

284 large amplitude reflections every ~500 – 700 m along the traverse. When a layer appears to bifurcate due to

285 changes in accumulation, we continue to trace the layer based on the trajectory of surrounding IRHs.

286 Horizons are not traced in areas where the attenuated signal makes them too difficult to interpret (Figure 3).

287 We trace layers for each spur starting at the depth of each layer at the corresponding firn core location. We

288 can trace layers below the depth of some firn cores by tracing them from cores that are deeper or have lower

289 accumulation rates.

290

291 We trace layers between cores using a connect-the-dots approach using the depth-age scale at each firn core.

292 We trace layers from one firn core to the next before checking that we intersect that core location at the

293 proper depth for the age of our traced IRH. Note that the depths of several layers at Cores 2 – 16 are located

294 below the bottom depth of those cores. Since these layers are isochronous, they are used to calculate

295 accumulation over appropriate time epochs by using dates obtained from intersections with other cores (see

296 Figure 3).

297

2.6. Accumulation calculations and uncertainty

299 Finally, we calculate snow accumulation using the firn core depth-age scales, measured and interpolated

300 depth-density profiles (Figure 4), and traced IRHs (Figure 2). We calculate the water equivalent accumulation

301 $\dot{b}$ (m w.e. a^{-1}) between adjacent IRHs from the depth z (m) and age t (year) of each layer, the average density

302 ρ (kg m^{-3}) between layers, and the density of water ρ_w (1000 kg m^{-3}):

303 $\dot{b} = \frac{1}{t_2 - t_1} \int_{z_1}^{z_2} \frac{\rho(z)}{\rho_w} \partial z$ (3).

304

305   We correct for layer thinning using a Nye (1963) model. The thinning factor has an average value of 0.9993

306   ± 0.0003 and is multiplied by the accumulation rate for each radar trace. For each radar trace, the thinning

307   factor, $\lambda(z)$, is calculated from the average accumulation $\dot{b}$ (m w.e. a^{-1}) of each epoch, average age of the

308   epoch a (year), and water equivalent thickness of the GrIS H (m), from Morlighem et al. (2014):

309    $\lambda(z) = e^{-\frac{\dot{b}}{H}a}$ **(4).**

310

311   Accumulation uncertainty can arise from independent errors in tracing IRHs, errors from incorrectly dating

312   firn cores, and/or errors in the densities used for converting from separation distance to water equivalent

313   accumulation. To reduce tracing errors, we retraced each IRH along the two main traverse paths four times

314   each. Close inspection of the IRHs reveals that the peaks defining IRHs are within ± 2 radar samples (within

315   at most ±0.12 m), and incorrectly jumping to the next IRH would result in an error of at most ± 10 samples

316   (within ±0.55 m). Our average epoch between IRHs is 5.0 years from the firn core depth-age scales, which

317   corresponds to a maximum tracing error of ~±0.11 m a^{-1} for each epoch, or a maximum error of ±0.061 m

318   w.e. a^{-1}, for an average density of 0.55 g cm^3 over this dataset.

319

320   We perform a leave-one-out cross validation to calculate accumulation errors at locations where we do not

321   have firn core density profiles. Here we choose one of the sixteen firn cores, in addition to the Raven/Dye-2

322   and Summit cores, to omit from our density interpolation (Figure 4), so that we interpolate density profiles

323   between adjacent firn cores and a Herron-Langway profile at the missing core location. We find maximum

324   single-epoch errors of 0.079 m w.e. a^{-1} and maximum RMS (1971 – 2016) errors of 0.046 m w.e. a^{-1} (Table

325   1) at the location of missing cores, corresponding to 20.1% of the accumulation at that location. These

326   differences are approximately twice as large at Cores 1 – 6 than Cores 7 – 16 due to larger differences between

327   measured and interpolated density profiles, likely a result of meltwater percolation and ice lenses (Graeter et

328   al., 2018).

329

330   Similarly, we perform a leave-out-out validation by omitting a firn core density profile location entirely and

331   interpolating density profiles over a larger distance (e.g. between Core 1 and Core 3). In this case we find

332   maximum single-epoch errors of 0.057 m w.e. a^{-1} and maximum RMS (1971 – 2016) errors of 0.033 m w.e.

333   a^{-1}.

334

335   We conservatively take our accumulation error from missing density measurements to be 0.079 m w.e. a^{-1}.

336   This error highlights the importance of our firn core spacing between 40 – 100 km along the traverse and

337   confirms that the accuracy of future remotely sensed radar accumulation (e.g. IceBridge snow and

338   accumulation radars) estimates depend on precise field-based *in situ* density profiles for accurate

339   accumulation history in the percolation zone. Overly et al. (2016) calculated accumulation in the dry snow

340   zone using Herron-Langway profiles within 3.5% of accumulation calculated using neutron-probe density

341   profiles. However, here we show that *in situ* measurements, or accurate meltwater percolation modeling

342   (Meyer and Hewitt, 2017), are required to correctly calculate SMB in the percolation zone.

343

344   **Table 1. Difference between accumulation rates at each GreenTrACS core site calculated using Herron-Langway profiles and firn core**
345   **density information.**

Core	RMS average difference (m w.e. a^{-1})	Max epoch difference (m w.e. a^{-1})	Max Epoch difference (% of acc.)
1	0.046	0.079	20.1
2	0.025	0.061	16.2
3	0.037	0.074	19.9
4	0.028	0.045	10.7
5	0.026	0.054	11.5
6	0.038	0.052	10.0
7	0.015	0.026	5.4
8	0.026	0.045	10.3
9	0.030	0.049	10.9
10	0.019	0.039	8.5
11	0.023	0.035	5.0
12	0.018	0.027	8.2
13	0.025	0.031	10.7
14	0.019	0.027	8.2
15	0.010	0.016	5.3
16	0.014	0.025	8.2

346

347   We assume uncertainty in dating the firn cores from annual layer counting to be ±0.5 years (Buchardt et al.,

348   2012). At the lowest accumulation locations, the smallest distance between layers is 0.15 m w.e. over an

349   epoch of 4.91 years. This gives an uncertainty in accumulation due to dating of at most ~±0.03 m w.e. a^{-1}.

350   The error associated with measuring *in situ* firn density has been estimated to be 1.4% (Karlöf et al., 2005).

351   However, following Hawley et al. (2014) and Lewis et al. (2017), we conservatively assume that our

352   measurements have a density measurement error of up to twice this large, corresponding to a maximum

353   accumulation error of ±0.014 m w.e. a^{-1}.

354

355   We calculate the total uncertainty from formal error propagation (following Bevington and Robinson, 1992)

356   from the average accumulation rate $\dot{b}$ = 0.385 m w.e. a^{-1}, average thickness between IRHs Δh = 3.56,

12

357   uncertainty in tracing δh, average firn density ρ, uncertainty in density measurements $\delta\rho$, average time

358   period between IRHs Δt, and uncertainty in core dating δt. We find a total accumulation rate uncertainty of

359   0.0709 m w.e. a^{-1} from equation 5.

360   $$\sigma_b = \sqrt{b^2\left(\left(\frac{\delta h}{\Delta h}\right)^2 + \left(\frac{\delta t}{\Delta t}\right)^2 + \left(\frac{\delta\rho}{\rho}\right)^2\right)} \qquad (5)$$

361

362   Due to the random and non-systematic nature of these errors, we can assume that they are unlikely to

363   contribute to a regional or temporal accumulation bias. To calculate uncertainty for accumulation averaged

364   over multiple epochs ($\sigma_{n-epochs}$) we divide our uncertainty σ_{epoch} by the square root of the number of traced

365   layers (n) at that location.

366   $$\sigma_{n-epochs} = \frac{\sigma_{epoch}}{\sqrt{n}} \qquad (6).$$

367

368   2.7. Model comparison

369   We compare our GreenTrACS accumulation results with annual outputs from Box et al. (2013; hereafter

370   "Box13"; 1840 – 1999), the Fifth Generation Mesoscale Model (Polar MM5; 1958 – 2008; Burgess et al.,

371   2010), MAR (1948 – 2015; Fettweis et al., 2016), and the Regional Atmospheric Climate Model (RACMO2;

372   1958 – 2015; Noël et al., 2018) over common time periods. Grid cell sizes for these model outputs are 5 km,

373   3 km, 5 km, and 1 km, respectively. For each radar trace we calculate statistically significant differences (at

374   $\alpha = 0.05$) using a two sample t-test with the GreenTrACS accumulation records for each epoch and RCM

375   accumulation for each common year. Additionally, we compare our GreenTrACS accumulation with an

376   accumulation map kriged from 295 firn cores and 20 coastal weather stations (Bales et al., 2009; hereafter

377   "Bales09"). We perform the same two sample t-test with the reported Bales09 uncertainty of 0.092 m w.e. a^{-}

378   ^{1} (Bales et al., 2009).

379

380   2.8. Accumulation trends

381   To investigate recent changes in GrIS accumulation, we calculate trends in accumulation across our GPR and

382   GreenTrACS firn core dataset. We fit a linear model to the accumulation time series for each radar trace and

383   analyze the trend for both slope and statistical significance. Likewise, we calculate trends and their statistical

384   significance for total precipitation (snowfall + rainfall) for MAR and RACMO2 grid cells from 1996 through

385   the end of both models' temporal coverage. We can compare these results with our accumulation trends since

386   precipitation and accumulation are nearly identical above the equilibrium line altitude, due to zero runoff and

387   negligible sublimation within the percolation zone.

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2.9. Storm track changes

To investigate the potential role of changing storm tracks in precipitation changes over the Western GrIS, we utilize the updated Serreze (2009) storm track database. This database contains six-hour interval positions of extratropical cyclone storm centers on a 2.5° grid. These centers are defined when a gridpoint sea level pressure is surrounded by gridpoints at least 2 mb higher than the central point (Serreze, 2009). We calculate the total number of days in which a storm center is located within our region of interest for each season. To determine statistical significance, we run a two sample t-test on the number of storms in our region of interest between 1958 – 1996 compared with 1996 – 2016.

3. Results and discussion

3.1. Firn core and GPR accumulation records

Figure 1 displays the mean accumulation at each location along the traverse route, with higher accumulation rates in the southwest and lower accumulation rates at higher elevations of the ice sheet interior, broadly consistent with previously published accumulation compilations (e.g. Bales et al., 2009) and RCM output (Box et al., 2013; Burgess et al., 2010; Fettweis et al., 2016; Noël et al., 2018). We analyze localized differences between GPR derived accumulation and these RCMs in Section 3.3. There is an especially high accumulation zone near Core 11 (0.685 m w.e. a^{-1}), nearly double the accumulation at Core 10 (0.453 m w.e. a^{-1}) and Core 12 (0.327 m w.e. a^{-1}), respectively situated only 43 km northwest and 73 km southwest of Core 11. In the GPR data, the number of traceable IRHs is highest towards the interior of the ice sheet and lowest in warmer areas towards the coast and in the south, where refrozen percolated melt water from enhanced surface melt attenuates the radar signal and reduces the number of observable IRHs (Brown et al., 2011; Figure 3).

3.2. Validation with past measurements

We validate our accumulation record with published core records from the PARCA campaign and accumulation data from the NASA IceBridge program. The locations of GreenTrACS Core sites 2, 5, 9, 10, 11, 14, 15, and 16 were chosen to reoccupy PARCA core locations 6745, 6945, 7147, 7247, 7249, NASA-U, 7347, and 7345, respectively. These GreenTrACS cores overlap with the accumulation history of each PARCA core and extend the record from 1997/1998 to 2016/2017. Accumulation rates derived from GreenTrACS firn cores are within error of those determined from corresponding PARCA cores during the period of overlap. Figure 5 compares the accumulation records from PARCA sites 6745, 6945, 7345, and NASA-U to their corresponding GreenTrACS cores, demonstrating that each pair of cores has similar long-term mean accumulation and nearly identical decadal variability. Thus, we have confidence in firn core
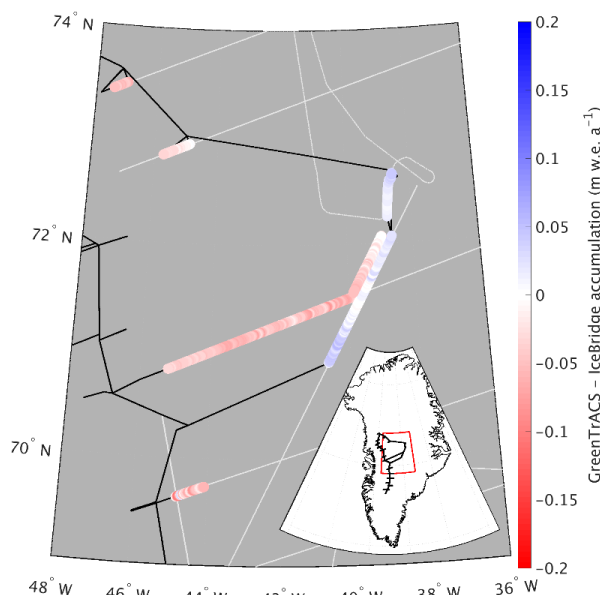
419   derived accumulation rates that are used in subsequent GPR calculations of accumulation rates throughout

420   the GreenTrACS traverse.

421

422

423   **Figure 5. Accumulation from GPR and collected firn cores (this study) compared with cores from the PARCA Campaign. Thin lines**
424   **represent annual PARCA (blue) and GreenTrACS (black) firn core accumulation, while thick lines are 5-year averages over**
425   **corresponding GPR epochs. Error bars represent one standard deviation over each epoch. GPR and PARCA accumulation averages**
426   **and decadal trends are statistically indistinguishable.**

427

428   Average (1966 – 2016) GPR accumulation is statistically indistinguishable with average (1962 – 2014)

429   IceBridge Accumulation Radar measurements analyzed by Lewis et al. (2017), with an RMS difference of

430   0.0387 ± 0.0327 m w.e. a^{-1} along a total of 562.5 km of overlap (Figure 6). The disagreement is largest at

431   lower elevations, where Herron and Langway (1980) density profiles used in Lewis et al. (2017) differ the

432   most from GreenTrACS firn core density profiles in the upper 30 m of firn, demonstrating the importance of

433   field observations for calibration and validation. The close agreement at higher elevations is illustrated in

434   Figure 7a, where our GreenTrACS accumulation measurements are statistically indistinguishable from the

435   IceBridge radar-derived accumulation (Lewis et al., 2017) along the 285 km A – A' transect on Figure 1.

436   Notice that the uncertainty in GreenTrACS accumulation progressively decreases higher in the percolation

437   zone and into the dry snow zone (towards the right in Figure 7) along this transect as density becomes less

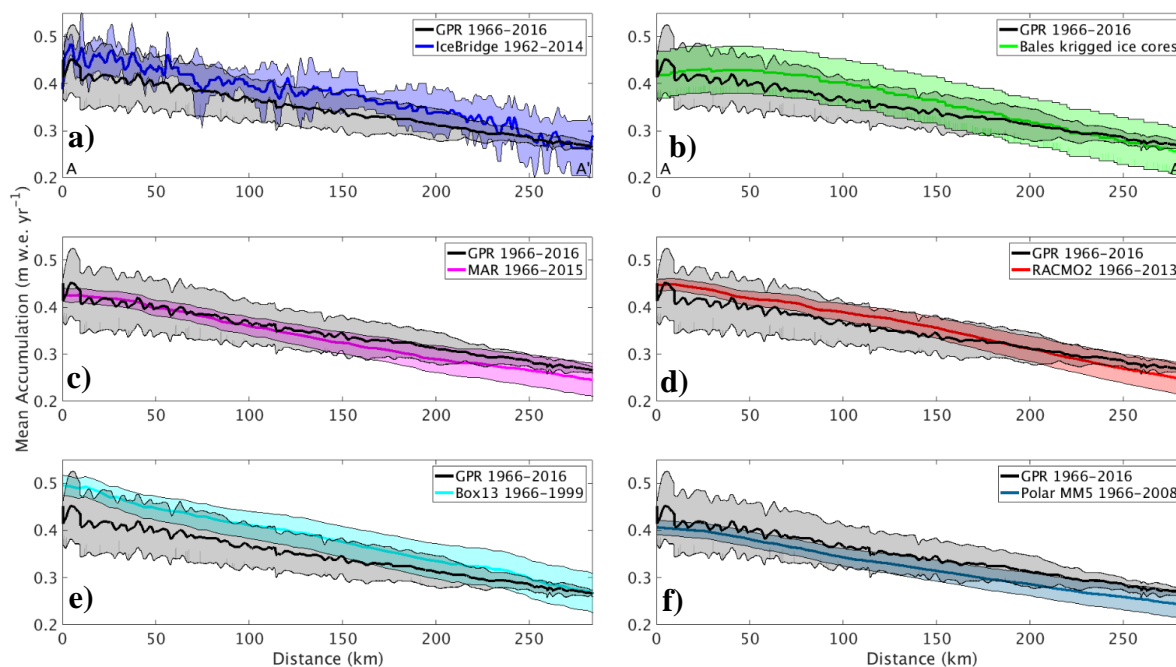438   heterogeneous from fewer melt layers (Graeter et al., 2018) and IRHs become easier to trace.
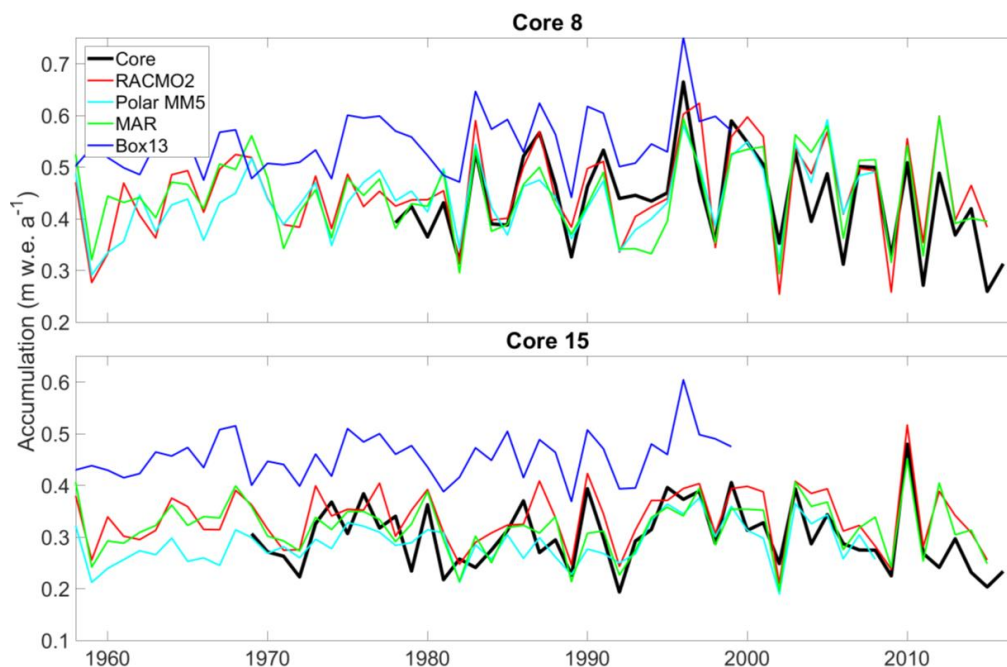
439

440

15

Figure 6. Difference between averaged (1966 – 2016) GreenTrACS accumulation and average (1962 – 2014) IceBridge Accumulation Radar rates from Lewis et al. (2017) across all 562.5 km of overlap. Spatially overlapping section of 2016 and 2017 traverses displayed as adjacent tracks. Also showing extent of GreenTrACS traverse (black) and IceBridge accumulation radar (grey). Inset shows map location with respect to GreenTrACS traverse (black).

3.3. Comparison to modelled accumulation

We assess differences between RCM accumulation output and GreenTrACS accumulation record at each firn core site, two of which are shown in Figure 8. In general, year-to-year correlations between GreenTrACS firn core accumulation records and RCM output for the corresponding grid cell are strong, positive, and statistically significant (Table 2). On average, GreenTrACS firn cores' correlation coefficient with MAR output is 0.718, with PolarMM5 is 0.701, with Box13 is 0.607, and with RACMO2 is 0.763. Every correlation is statistically significant at $p < 0.05$ except for Cores 7 and 11 with Box13. We do not report a correlation coefficient for Core 11 and Box13 because they only share two common years. Temporal correlation coefficients remain high even at locations with large magnitude differences between RCM output and firn core accumulation. For example, the Box13 model overestimates accumulation at Core 15 by 0.15 ± 0.05 m w.e. a^{-1}, on average, but the model output has a correlation coefficient of 0.48 with Core 15 (Table 2) and matches years of high accumulation (e.g. 1987, 1990, and 1996) and low accumulation (e.g. 1981, 1989, 1992).

16

Figure 7. Average GreenTrACS GPR accumulation (black) compared with a) IceBridge accumulation radar, b) Bales09 krigged ice core map, c) MAR, d) RACMO2, e) Box13, and f) Polar MM5. GPR measurements are statistically indistinguishable from each of the other measurements along this 285 km transect in the dry snow zone (A – A' on Figure 1).

Figure 8. Accumulation record at GreenTrACS Core 8 and Core 15 (black) compared with RCM output from RACMO2 (red), Polar MM5 (cyan), MAR (green), and Box13 (blue). We find statistically significant Pearson correlation coefficients between GreenTrACS and RCM accumulation rates for these cores (see Table 2).

17

469

470 **Table 2. Pearson correlation coefficients between accumulation rate time series from firn cores and co-located RCM output over their**
471 **common time period[#].**

	Available data period	MAR	PolarMM5	Box13	RACMO2
Core1	1966 – 2016	**0.70**	**0.66**	**0.56**	**0.73**
Core2	1969 – 2016	**0.75**	**0.77**	**0.62**	**0.79**
Core3	1971 – 2016	**0.72**	**0.69**	**0.63**	**0.74**
Core4	1977 – 2016	**0.79**	**0.74**	**0.72**	**0.72**
Core5	1984 – 2016	**0.81**	**0.80**	**0.60**	**0.79**
Core6	1985 – 2016	**0.76**	**0.76**	**0.65**	**0.83**
Core7	1993 – 2016	**0.81**	**0.82**	0.61	**0.73**
Core8	1978 – 2017	**0.78**	**0.77**	**0.69**	**0.81**
Core9	1984 – 2017	**0.68**	**0.75**	**0.74**	**0.79**
Core10	1984 – 2017	**0.88**	**0.80**	**0.80**	**0.80**
Core11	1997 – 2017	**0.75**	**0.59**	N/A	**0.75**
Core12	1962 – 2017	**0.6**	**0.54**	**0.53**	**0.64**
Core13	1955 – 2017	**0.51**	**0.62**	**0.37**	**0.76**
Core14	1974 – 2017	**0.70**	**0.62**	**0.46**	**0.74**
Core15	1969 – 2017	**0.68**	**0.63**	**0.48**	**0.75**
Core16	1979 – 2017	**0.79**	**0.77**	**0.66**	**0.88**

472 [#]Statistically significant correlations ($p < 0.05$) are bold

473

474 We also assess spatial differences between GreenTrACS accumulation and mean RCM accumulation
475 averaged over several decades. Figure 9 shows that differences between GreenTrACS accumulation and
476 RCM output increase in magnitude, become more spatially heterogeneous, and vary by model at lower
477 elevations of the ice sheet where topographic variations are larger and surface melt increases. Averaged over
478 all 4436 km of the traverse, the RMS difference between each model and GreenTrACS accumulation is 0.068
479 ± 0.065 (MAR), 0.0562 ± 0.0548 (RACMO2), 0.0822 ± 0.0702 (Box13), 0.048 ± 0.045 (Polar MM5), and
480 0.0475 ± 0.0445 m w.e. a^{-1} (Bales09). We find that RCM differences from GreenTrACS accumulation are
481 small in the dry snow zone (Figure 9). For example, Figure 7 shows that average GreenTrACS accumulation
482 measurements from 1966 – 2016 along the A – A' transect in Figure 1 are statistically indistinguishable from
483 those derived from the Bales09 krigged ice core map (Figure 7b), MAR (1966 – 2015; Figure 7c), RACMO2
484 (1966 – 2013; Figure 7d), Box13 (1966 – 1999; Figure 7e), and Polar MM5 (1966 – 2008; Figure 7f).
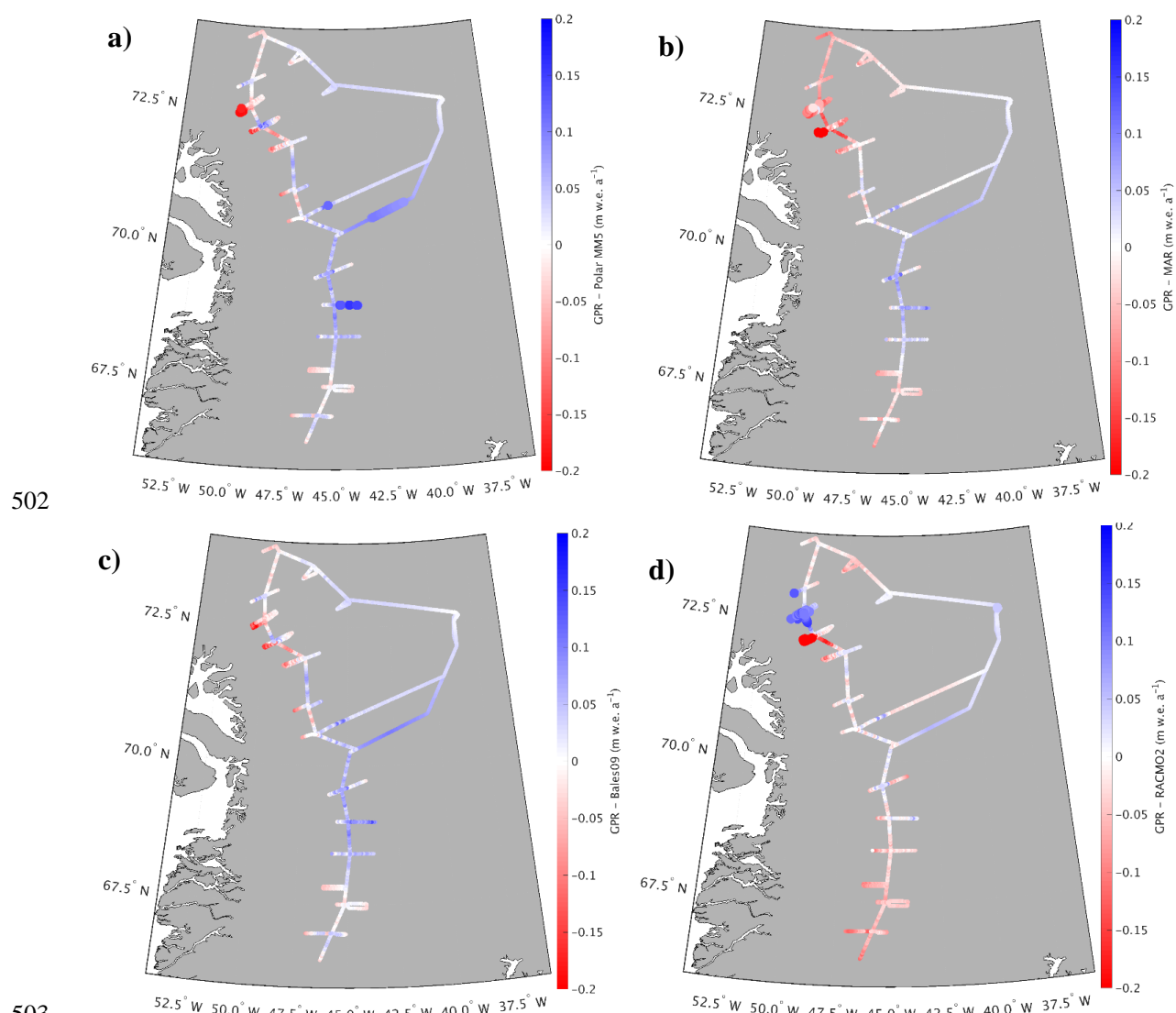
485

486 However, the high spatial resolution of our dataset shows significant accumulation variability not captured
487 in model output (Figure 9). For example, Polar MM5 and MAR both underestimate accumulation between
488 Core 4 and Summit, while overestimating accumulation to the west of Cores 10 – 12. Likewise, RACMO2
489 overestimates accumulation between Raven/Dye-2 and Core 5 by 0.03 to 0.08 m w.e. a^{-1} and shows
490 statistically significant differences east of Cores 11 and 12. Bales09 accurately calculates accumulation along
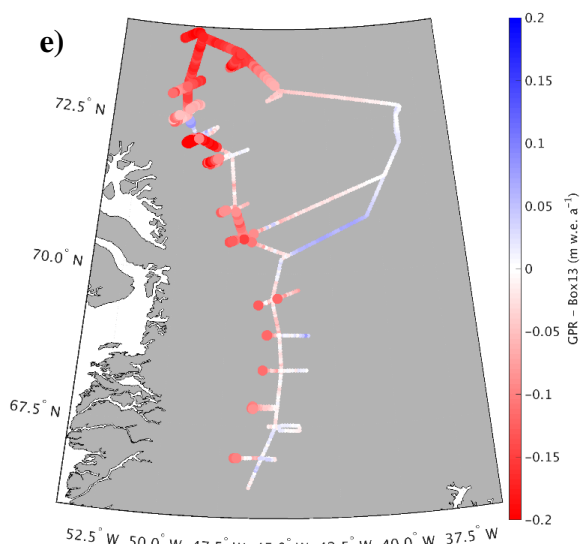491 most of the 2016 traverse, but overestimates accumulation west of Cores 11 and 12 by 0.135 ± 0.041 m w.e.

18

492   a^{-1}. Finally, Box13 overestimates accumulation along many of the western spurs and has statistically

493   significant overestimations of 0.1 to 0.4 m w.e. a^{-1} between Cores 10 and 16. Box13 overestimates 67.8% of

494   the data in the Core 10 – 16 region by at least 0.1 m w.e. a^{-1}, and 6.6% of that data by at least 0.2 m w.e. a^{-1}.

495

496   Our study is almost entirely contained within drainage basin E from Vernon et al. (2013), who note that basin

497   E is the only major Greenland drainage basin with no statistically significant differences in SMB between

498   the four RCMs. However, differences of 0.1 to 0.4 m w.e. a^{-1} exist when we look at a local (sub-drainage-

499   basin) scale for each model. All four of the RCMs overestimate accumulation along the western spur of Core

500   11 and they all underestimate accumulation along the eastern spur of Core 5 (Figure 9).

501

502

503

504

505

Figure 9. Differences between GreenTrACS accumulation and a) Polar MM5, b) MAR, c) Bales09, d) RACMO2, and e) Box13 accumulation averaged over the corresponding time periods. Large dots show statistically significant differences from GreenTrACS accumulation.

509

510   In summary, the RCMs do an excellent job of calculating accumulation averaged over this drainage basin,

511   with RMS values between 0.048 and 0.0822 m w.e. a^{-1}, but there are larger differences of 0.1 to 0.4 m w.e.

512   a^{-1} between model and GPR accumulation on local scales. Differences between GreenTrACS and RCM

513   accumulation are largest in areas concurrent with the fewest, shortest, and/or most outdated *in situ*

514   measurements. For example, the GPR vs. model differences near Cores 11, 12, and 13 are relatively large for

515   all RCMs, despite Core 11 being co-located with PARCA 7249. However, the PARCA cores were collected

516   over 20 years ago, and Core 11 only spanned 7 years because of the high accumulation rate at that site. This

517   highlights the importance of collecting updated field-based measurements to calibrate remotely sensed data

518   and RCM output.

519

520   **3.4. Accumulation temporal trends**

521   In most locations, there are no statistically significant trends in the GreenTrACS accumulation record from

522   1966 through the mid-1990s. However, a changepoint analysis (Lavielle, 2005) reveals that accumulation in

523   the Western GrIS percolation zone changed significantly after the 1995 – 1996 accumulation year. Since

524   1996, our record indicates a statistically significant average accumulation decrease of 0.009 ± 0.005 m w.e

525   a^{-2}, or 2.4 ± 1.5 % a^{-1}, from 1996 to 2017. Although we observe fewer statistically significant accumulation

20

526   trends when we expand this analysis to include the entire temporal duration for each firn core, the sign of the

527   trend at each core site does not change.

528

529   In Figure 10, we compare the negative accumulation trend in the GreenTrACS record (1996 – 2016) to best-

530   fit linear trends in total precipitation (rain + snowfall) across the ice sheet in MAR and RACMO2 simulations

531   over the 1996 – 2015 and 1996 – 2013 periods, respectively. Also shown in Figure 10 are 1996 – 2016

532   accumulation trends for all 16 GreenTrACS firn cores (squares), accumulation trends from ACT10A (1996

533   – 2010), ACT10B (1996 – 2010), ACT10C (1996 – 2010), D4 (1991 – 2002), D5 (1991 – 2002), Katie (1991

534   – 2002), Sandy (1991 – 2002), and Summit 2010 (1991 – 2010) ice/firn cores (stars on ice sheet), and

535   precipitation trends from coastal weather stations (Mernild et al., 2014; stars on coast). Statistically

536   significant trends ($p < 0.05$) in core data are indicated by black dots, while statistically significant trends in

537   the MAR and RACMO2 output are stippled in black.

538

539   We find strong agreement between the accumulation decrease in the GreenTrACS record and widespread

540   precipitation decreases in the RCMs over the study area (Figure 10). On average, the RCMs have a more

541   negative precipitation trend than the GreenTrACS record by 0.003 ± 0.005 for MAR and 0.0016 ± 0.0051 m

542   w.e. a^{-2} for RACMO2. Vernon et al. (2013) show a melt-driven decrease in SMB across this drainage basin

543   of 31.1% (ECMWFd), 61.6% (RACMO2), 76.5% (MAR), and 33.5% (Polar MM5) for the 1996 – 2008

544   period. The negative precipitation trends of 2.4 ± 1.5 % a^{-1} (Figure 10d) indicate a total of 2539.4 fewer Gt

545   of precipitation and a total of 5159.1 additional Gt of melt (not shown) over 1996 – 2013 across the GrIS.

546   Thus, our analysis suggests that a significant decline in snow accumulation contributes to declining SMB

547   throughout the Western GrIS over recent decades, in addition to increasing surface melt from rising

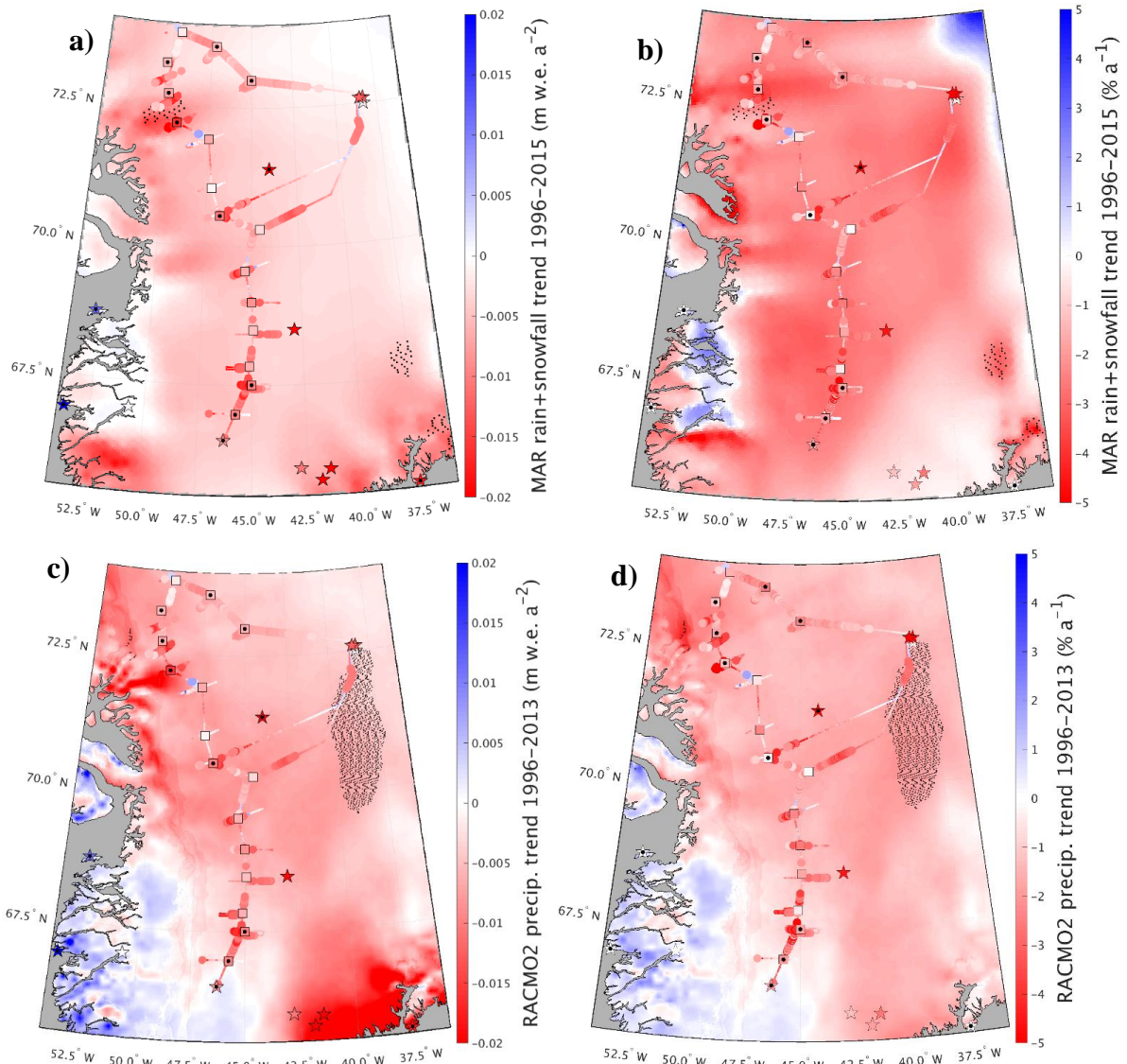548   temperatures (van den Broeke et al., 2009, 2016).

549

550

Figure 10. Best fit linear trends for each grid cell showing magnitude (left) and percent (right) changes in total precipitation for a) and b) MAR (1996 – 2015) and c) and d) RACMO2 (1996 – 2013). Statistically significant RCM grid cell trends are stippled black. Also shown are accumulation trends for GreenTrACS firn cores (squares), ACT10A, ACT10B, ACT10C, D4, D5, Katie, Sandy, Summit 2010, and Raven/Dye-2 cores (stars on ice sheet) and precipitation trends from Mernild et al (2014; stars on coast) with statistically significant trends indicated by black dots.

3.5. Effects of melt on accumulation trends

558 Increased melt throughout the 1996 – 2016 period is a confounding variable when analyzing trends in

559 accumulation. With increased melt over the past several decades in this region, meltwater percolates down

560 through several years of firn (Benson, 1962; Graeter et al., 2018; Harper et al., 2012; Wong et al., 2013).

561 This movement of mass into lower years can artificially increase the mass balance at depth and lower the

562 mass balance during the most recent years, which have not experienced as much meltwater percolation from

563 more recent annual layers. Therefore, it is necessary to evaluate the degree to which the recent accumulation

564 decrease in the GreenTrACS record is biased by the recent increase in surface melt and percolation.
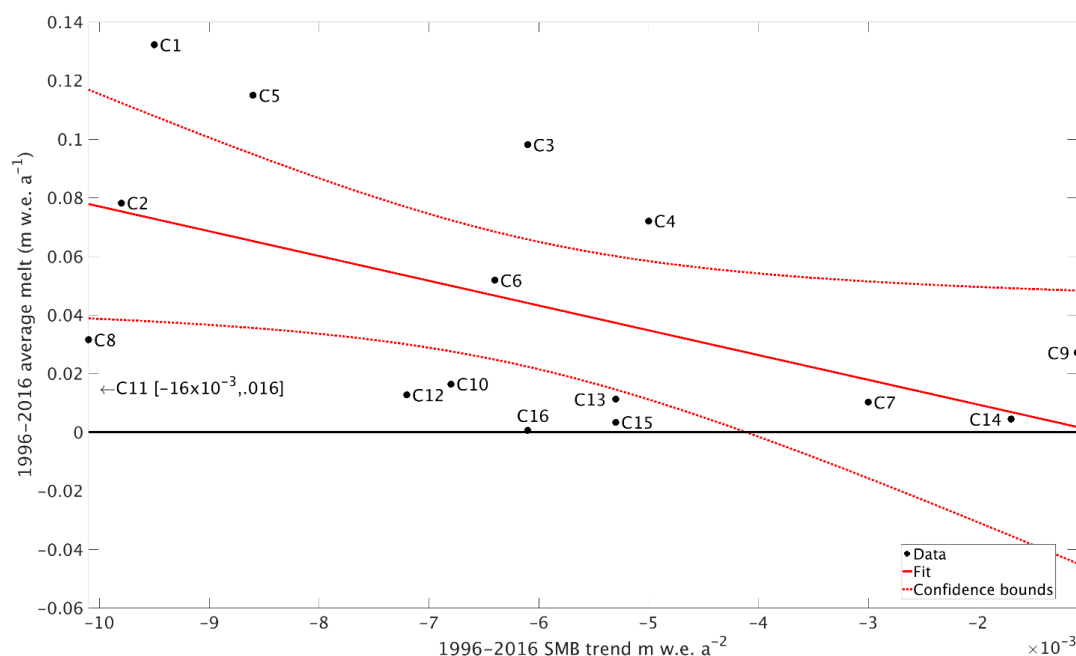
565

566 Figure 11 compares the 1996 – 2016 mass balance trends with 1996 – 2016 average melt for each of the

567 sixteen GreenTrACS firn cores. If we exclude Core 11 (which only dates back to 1997 and has a highly

568 negative SMB trend), the linear regression is statistically significant with $p = 0.04$ (Figure 11). Note that both

569 the measured Core 11 SMB trend and RCM trends at that location are so negative, with that small amount of

570 average melt, that the linear trend is no longer significant if that point is included in the calculations. On

571 average, we find larger negative accumulation trends (-7×10^{-3} to -10×10^{-3} m w.e. a^{-2}) at the lower latitude

572 cores that experience more melt, supporting the hypothesis that meltwater percolation and refreezing are

573 enhancing the negative accumulation trend.

574

575 However, several other lines of evidence support a negative accumulation trend in the study area since 1996.

576 First, we find statistically significant negative accumulation trends at Cores 10, 11, 12, 13, 15, and 16, each

577 of which experience <1.6 cm a^{-1} of meltwater percolation on average (Figure 11). Additionally, we have

578 confidence that GreenTrACS accumulation trends reported here are not artifacts of meltwater percolation

579 because both MAR and RACMO2 have similar trends in precipitation (Figure 10). Finally, we evaluate the

580 maximum effect meltwater percolation could have on GreenTrACS accumulation trends over 1996 – 2016.

581 The largest melt layer from our sixteen ice cores occurred during 2003 – 2004 in Core 1 and contains 0.364

582 m of ice, equivalent to 0.333 m w.e. (Graeter et al., 2018). We add this percolation to nine years' of

583 accumulation using a sine wave (percolation magnitude 0, 0.5, 1, 0.5, 0, -0.5, -1, -0.5, 0), square wave (0, 0,

584 0, 1, 1, 1, 0, 0, 0), and triangle wave (0, 0.25, 0.5, 0.75, 1, 0.75, 0.5, 0.25, 0) weighted kernel, before re-

585 computing hypothetical accumulation trends over the same time period with additional meltwater

586 percolation. Regardless of the wave-type choice, re-calculated trends remain within a factor of two of the

587 original SMB trends and do not change sign with additional percolation.

588

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3.6. Atmospheric circulation drivers of the recent accumulation decline

Our analysis indicates that snow accumulation has been declining in Western Greenland since 1996, despite significant warming and resulting increases in saturation vapor pressure from the Clausius-Clapeyron relationship. Instead, precipitation decreases over Western Greenland likely result from changes in atmospheric and/or oceanic circulation. Mernild et al. (2014) and Auger et al. (2017) found that the positive phase of the Atlantic Multidecadal Oscillation (AMO) is associated with a precipitation increase over interior and Southwestern Greenland based on ice core records and the Japanese Meteorological Agency 55 Year Reanalysis (JRA-55; Kobayashi et al., 2015), respectively. In direct contrast with these findings, the decline in Western Greenland accumulation documented in the GreenTrACS record began in the mid-1990s, contemporaneous with a switch to the AMO positive phase.

We hypothesize that the differences between our results and those of Auger et al. (2017) and Mernild et al. (2014) stem from different causes. Auger et al. (2017) validated the reanalysis data by demonstrating that JRA-55 precipitation at Nuuk, Greenland is significantly correlated with Nuuk station data from 1958 – 2013. Furthermore, coastal precipitation in Western Greenland is strongly and significantly ($p < 0.05$) correlated with precipitation over the interior Western GrIS in the JRA-55 dataset (not shown). However, Mernild et al. (2014) found that coastal Greenland precipitation is anti-correlated with ice core accumulation records from

609   the interior GrIS from 1900 to 2000. This suggests that JRA-55 precipitation data, which is not constrained

610   by ice core accumulation records, should be interpreted with caution over the interior GrIS. Mernild et al.

611   (2014) concluded that positive AMO conditions favor higher precipitation over the interior GrIS based on

612   the previous positive AMO phase (1920s to mid-1960s), contrasting with lower accumulation during the

613   negative AMO phases (mid-1960s to mid-1990s and prior to the 1920s). However, Mernild et al. (2014) state

614   that the ice core composite record in their analysis may be biased from 1995 – 2000, and they do not analyze

615   precipitation trends after 2000. Thus, the decline in Western GrIS accumulation documented in the

616   GreenTrACS cores during the latest positive AMO phase from 1996 to 2017 was not captured in the Mernild

617   et al. (2014) analysis. Our results suggest that factors other than the AMO are behind the decline in Western

618   GrIS accumulation since 1996.

619

620   We find that the decrease in accumulation over the Western GrIS is associated with a significant decrease in

621   the number of storm-days since 1996. The GreenTrACS region experienced an average of 115.8 ± 15.3 storm-

622   days per year over 1958 – 1996 and 96.2 ± 27.3 storm-days per year over 1996 – 2016. A two sample t-test

623   indicates that this 17% decline in storm-days is statistically significant ($p < 0.001$). The largest decrease in

624   storm-days (25%) over the GreenTrACS region occurred during summer, with 56.4 ± 6.1 storm-days per

625   summer from 1958 – 1996 and 42.3 ± 17.4 storm-days per summer from 1996 – 2016 ($p < 0.0001$; Figure

626   12b). We also find an increase in the number of storm-days in the Northwestern GrIS near Thule (not shown).
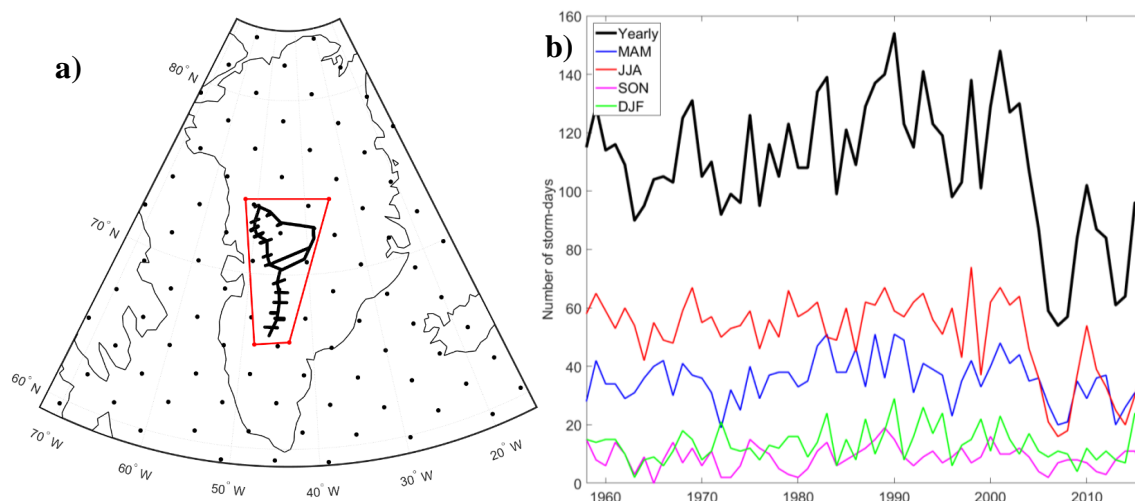
627

628   The decline in summer storm-days indicates a relationship with well-documented stronger summer blocking

629   over Greenland over the past two decades (Hanna et al., 2013; McLeod and Mote, 2016), with a positive

630   Greenland Blocking Index during 17 out of 21 summers between 1996 – 2016 (Hanna et al., 2016). The June

631   – August GBI had a statistically significant positive trend of 1.87 (unitless; normalized to 1951 – 2000) from

632   1991 – 2015 (Hanna et al., 2016). The summertime 500 mbar geopotential height increased 50 – 70 m over

633   the 1996 – 2016 period compared with the 1979 – 1996 baseline (Figure 12c), indicating stronger blocking

634   that we suggest likely reduced precipitation over the central GrIS by deflecting storms poleward from the

635   Greenland interior. This is consistent with an observed $0.9 \pm 0.3\%$ a^{-1} decrease in JJA cloud cover over

636   Greenland from 1995-2009, with the largest decreases in the GreenTrACS region (Hofer et al., 2017).

637   Furthermore, we find a strong negative correlation between ERA-Interim 1979 – 2015 June – August (JJA)

638   GBI and JJA precipitation in both MAR (Figure 12d) and RACMO2 (not shown) across the central and

639   southern GrIS. These results suggest that the blocking-induced accumulation decline observed in the

640   GreenTrACS region is representative of a broader pattern over the GrIS, with the exception of Northwest

641   Greenland where poleward blocking has increased storm-days (not shown) and accumulation (Figure 12d).

642

643

644 The effect of summertime Greenland blocking has been discussed primarily in the context of increasing

645 surface melt (Hanna et al.. 2013; Ballinger et al., 2017; Hanna et al., 2018; Hofer et al., 2017), while the

646 effect of blocking on precipitation has received less attention (Hanna et al., 2013; McLeod and Mote, 2016).

647 Our results highlight that stronger summer blocking reduces GrIS SMB through both an increase in surface

648 melting and a decrease in accumulation. Stronger summer blocking has been tied to an observed increase in

649 surface melt since 1996 across the Western GrIS percolation zone (Graeter et al., 2018), and to the July 2012

650 melt event, during which 98.6% of the GrIS experienced melt (Nghiem et al., 2012). We show here with *in*

651 *situ* data that snow accumulation has declined in this same region as strong blocking has decreased the

652 number of summer storm-days. Presently, none of the GBI outputs from the Coupled Model Intercomparison

653 Project 5 (CMIP5) suite of global climate models accurately capture the recent summer GBI increase (Hanna

654 et al., 2018). Improved predictions of summertime Greenland blocking under future anthropogenic forcing

655 scenarios are therefore critical for accurately predicting Greenland SMB and its contribution to sea level rise.
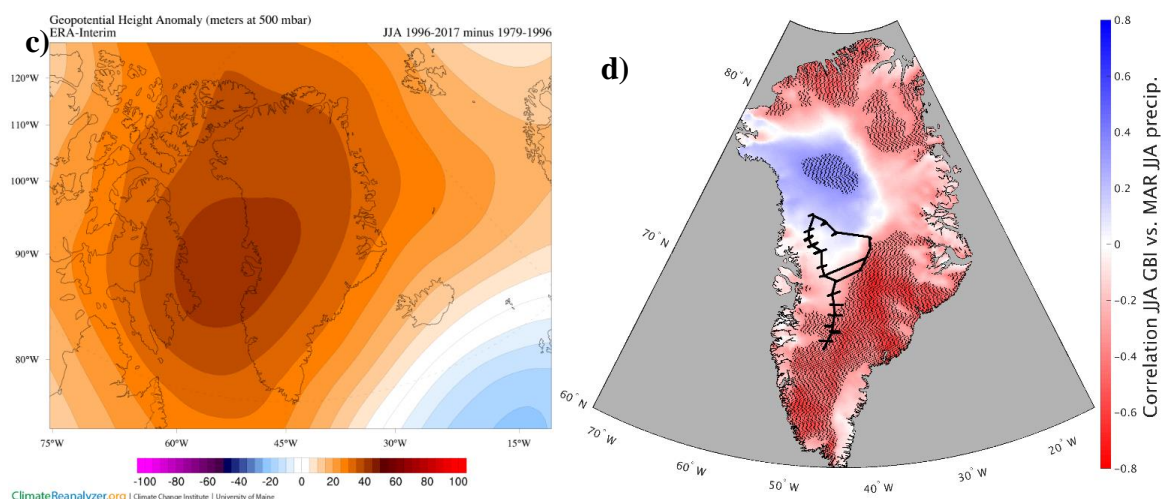
656

657

Figure 12. a) (Serreze, 2009) gridded storm track dataset showing location of GreenTrACS traverse and inquiry box. b) Total number of storm-days within inquiry box for annual and seasonal periods. Horizontal black lines show a decrease in 1958 – 1996 to 1996 – 2016 average number of storm-days within this region. c) 500 mbar geopotential height change over Greenland showing 1996 – 2016 minus 1979 – 1996 for the summer season. Image obtained using Climate Reanalyzer (http://cci-reanalyzer.org), Climate Change Institute, University of Maine, United States. d) Correlation between June – August Greenland Blocking Index and MAR June – August precipitation. Statistically significant RCM grid cell correlations are stippled black. GreenTrACS traverse is shown in black.

4. Conclusions

We have developed a new dataset of accumulation rates over the western interior of the Greenland ice sheet spanning the past 20 – 60 years, based on sixteen 22 – 32 m long firn cores and 4436 km of *in situ* GPR accumulation data. This accumulation record is internally consistent across the dataset and is validated by previous *in situ* field measurements and other radar-derived accumulation measurements (e.g Lewis et al., 2017).

Overall, the Polar MM5, MAR, Box13, and RACMO2 Regional Climate Models accurately capture large spatial patterns in accumulation over the GrIS, but show statistically significant differences from GPR accumulation on a regional basis. The average RMS difference between each model and GreenTrACS accumulation is 0.068 ± 0.065 (MAR), 0.048 ± 0.045 (Polar MM5), 0.0822 ± 0.0702 (Box13), 0.0562 ± 0.0548 (RACMO2), and 0.0475 ± 0.0445 m w.e. a^{-1} (Bales09). These differences are on the same order as the uncertainties in the GreenTrACS and RCM accumulation estimates. While these average differences are small, we find differences of 0.1 to 0.4 m w.e. a^{-1} when we investigate at a local scale for each model.

While global climate models predict a 21st-century increase in precipitation over the GrIS (e.g. Bintanja and Selten, 2014), we observe a decrease in precipitation across the Western GrIS from 1996 – 2016 using records

27

683 from firn cores, GPR, and published RCMs. We believe this study is the first to identify widespread negative

684 GrIS precipitation trends during this period of enhanced surface melt, evident in these RCMs and our field

685 observations (Graeter et al., 2018).

686

687 We attribute the decrease in accumulation over the Western GrIS between 1996 and 2016 to more persistently

688 positive Greenland blocking in the summer. We find a statistically significant 25% reduction in the number

689 of summer storms that precipitate over the GreenTrACS region since 1996. While warming temperatures

690 from anthropogenic forcing and enhanced summer blocking have increased melt across the western

691 percolation zone, here we show that summer blocking has also contributed to declining precipitation over the

692 past two decades. This has led to a strongly negative SMB trend on both the input and output sides of the

693 SMB equation that may not be accurately captured in global climate models that are currently unable to

694 reproduce the recent increase in blocking. This highlights the importance of improving GCM projections of

695 future summer blocking to accurately forecast Greenland precipitation and melt rates under stronger

696 greenhouse gas forcing.

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