An empirical algorithm to map perennial firn aquifers ~~and,~~ ice slabs, ~~and perched firn aquifers~~ within the Greenland Ice Sheet using satellite L-band microwave radiometry

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Abstract

Perennial firn aquifers are subsurface meltwater reservoirs consisting of a meters-thick ~~formed from a~~ water-saturated firn layer that can form on spatial scales as large as tens of kilometers. They have been observed within the percolation facies of glaciated regions experiencing intense seasonal surface melting and high snow accumulation. Widespread perennial firn aquifers have been identified within the Greenland Ice Sheet (GrIS) via field expeditions, airborne ice-penetrating radar surveys, and satellite microwave sensors. In contrast, ice slabs are nearly-continuous ice layers that can also form on spatial scales as large as tens of kilometers as a result of surface and subsurface water-saturated snow and firn layers sequentially refreezing following multiple melting seasons. They have been observed within the percolation facies of glaciated regions experiencing intense seasonal surface melting, but in areas where snow accumulation is at least ~25% lower as compared to perennial firn aquifer areas. Widespread ice slabs have recently been identified within the GrIS via field expeditions and airborne ice-penetrating radar surveys, specifically in areas where perennial firn aquifers typically do not form. However, ice slabs have yet to be ~~inferred~~ identified from space. Together, these two ice sheet features represent distinct, but related, sub-facies within the broader percolation facies of the GrIS that can be defined primarily by differences in snow accumulation, which influences the englacial hydrology and thermal characteristics of firn layers at depth.

Here, for the first time, we use ~~use~~ enhanced-resolution vertically-polarized L-band brightness temperature (T_V^B) imagery (2015-2019) generated using observations collected over the GrIS by NASA's Soil Moisture Active Passive (SMAP) satellite to map ~~both~~ perennial firn aquifer and

ice slab areas together as a continuous englacial hydrological system over the percolation facies. We also map 'perched' firn aquifer areas, which we define as areas where shallow water-saturated firn layers transiently form on top of buried ice slabs, or other semi-impermeable layers within the snow and firn. We use an An empirical algorithm previously developed to map the extent of Greenland's perennial firn aquifers via fitting exponentially decreasing temporal L-band signatures to a set of sigmoidal curves. Th is algorithm is recalibrated to also map the extent of ice slab and perched firn aquifer areas using airborne ice-penetrating radar surveys collected by NASA's Operation Ice Bridge (OIB) campaigns (2010-2017). Our SMAP-derived maps show that between 2015 and 2019, perennial firn aquifer areas extended over 6464,000 km², and ice slab areas extended over 76,000 km², and perched firn aquifer areas extended over ~37,000 km².. Combined together, these three sub-facies are the equivalent of 2424% of the percolation facies of the GrIS. As Greenland's climate continues to warm, and seasonal surface melting will increases in extent, intensity, and duration., Qquantifying the possible rapid expansion of each of these sub-facies using current and forthcoming satellite L-band microwave radiometry and synthetic aperture radar missions has significant implications for understanding ice sheet-wide variability in englacial firn hydrology resulting in meltwater-induced hydrofracturing and accelerated ice flow as well as high-elevation meltwater run-off that can impact the mass balance and stability of the GrIS.

1 Introduction

The recent launches of several satellite L-band microwave radiometry missions by NASA (Aquarius mission, Levine, et al., 2007; Soil Moisture Active Passive (SMAP) mission, Entekhabi et al., 2010) and ESA (Soil Moisture and Ocean Salinity (SMOS), Kerr et al., 2010) haves provided a new Earth-observation tool capable of detecting stored meltwatermeltwater storedage approximately tens of meters to kilometers beneath the surface of ice sheetsice sheet surface. Jezek et al. (2015) recently demonstrated that in the high-elevation (~3500 m a.s.l.) dry snow facies of the Antarctic Ice Sheet, meltwater stored in subglacial Lake Vostok can be detected as deep as ~4000 4 km beneath the ice sheet surface. Subglacial lakes represent radiometrically cold subsurface meltwater reservoirs. Upwelling L-band emissions from the radiometrically warm bedrock underlying the subglacial lakes are is effectively blocked by high reflectivity and attenuation at the interface between the bedrock and the overlying lake bottom. This results in a lower observed microwave brightness temperature (T^B) at the ice sheet surface as compared to other dry snow facies areas where bedrock contributes to L-band emission s depth-integrated over the entire ice sheet thickness.

Similar to subglacial lakes, perennial firn aquifers also represent radiometrically cold subsurface meltwater reservoirs (Miller et al., 2020) consisting of a ~4-25 m thick water-saturated firn layer (Koenig et al., 2014; Montgomery et al., 2017; Chu et al., 2018) that can form on spatial scales as large as tens of kilometers (Forster et al., 2014). Perennial firn aquifersformed from a ~4 m-25 m thick water-saturated firn

78 ~~layer (Koenig et al., 2014; Montgomery et al., 2017; Chu et al., 2018). They~~ have been ~~observed~~ identified
79 via field expeditions (Forster et al., 2014), airborne ice-penetrating radar surveys (Miege et al., 2016), and
80 satellite microwave sensors (Brangers et al., 2020; Miller et al., 2020)~~via field expeditions and airborne ice-~~
81 ~~penetrating radar surveys~~ in the lower-elevation (<~2000 m a.s.l.) percolation facies of the Greenland Ice
82 Sheet (GrIS) at depths from between ~1 m and 40 m beneath the ice sheet surface. They exist in areas
83 that experience intense seasonal surface melting and rain (>~650 mm w.e. yr^{-1}) during the melting season
84 and high snow accumulation (>~800 mm w.e. yr^{-1}) during the freezing season (Forster et al., 2014).~~at~~
85 ~~depths from between ~1 m and 40 m beneath the ice sheet surface (Miège et al. 2016), and in areas that~~
86 ~~experience intense seasonal surface melting (>650 mm yr^{-1}) during the melting season and high snow~~
87 ~~accumulation (>800 mm yr^{-1}) during the freezing season (Forster et al., 2014).~~ High snow accumulation in
88 perennial firn aquifer areas thermally insulates water-saturated firn layers from the cold atmosphere
89 allowing seasonal meltwater to be stored in liquid form year-round if the overlying seasonal snow layer is
90 sufficiently thick (Kuipers Munneke et al., 2014). Koenig et al. (2014) estimated that the volumetric fraction
91 of meltwater stored within the pore space of Greenland's perennial firn aquifers just prior to melt onset
92 ranges from between ~10% and 25%, which limits the upward propagation of electromagnetic energy from
93 greater depths within the ice sheet. Large volumetric fractions of meltwater within the firn pore space results
94 in high reflectivity and attenuation at the interface between water-saturated firn layers and the overlying
95 refrozen firn layers, and between glacial ice or a~~n~~ ~~semi-~~impermeable layer and the overlying water-
96 saturated firn layers. Upwelling L-band emission_s ~~from~~ deeper glacial ice and the underlying bedrock ~~are~~
97 is effectively blocked.

        While perennial firn aquifers are radiometrically cold, the slow refreezing of deeper firn layers
99 saturated with large volumetric fractions of meltwater represents a significant source of latent heat that is
100 continuously released throughout the freezing season. Refreezing of seasonal meltwater by the descending
101 winter cold wave (Pfeffer et al., 1991), and the subsequent formation of embedded ice structures (i.e.,
102 horizontally-oriented ice layers and ice lenses, and vertically-oriented ice pipes;~~;~~ Benson et al., 1960;
103 Humphrey et al., 2012; Harper et al., 2012) within the upper snow and firn layers represents a secondary
104 source of latent heat. These heat sources help maintain meltwater at depth. Perennial firn aquifer areas
105 are radiometrically warmer than other percolation facies areas where the single source of latent heat is via
106 refreezing of seasonal meltwater. This results in a higher observed T^B at the ice sheet surface during the
107 freezing season as compared to other percolation facies areas where seasonal meltwater is fully refrozen
108 and stored exclusively as embedded ice.

109 ~~Many open questions remain about Greenland's perennial firn aquifers, regarding initial~~
110 ~~formation, extent, depth, flow characteristics, timescales of refreezing and/or onglacial drainage, and~~
111 ~~connections to the subglacial hydrological system. Seasonal surface melting over the GrIS has increased~~
112 ~~in extent, intensity, and duration since the beginning of the satellite era (Steffen et al., 2004; Tedesco e al.,~~
113 ~~2008; Tedesco et al., 2011; Nghiem et al., 2012; Tedesco et al., 2016; Tedesco and Fettweis, 2020;~~
114 ~~Cullather et al., 2020). If this trend continues (Franco et al., 2013; Noël et al., 2021), subsequent increases~~

115 in the volume of meltwater stored within Greenland's perennial firn aquifers will increase the possibility of
116 crevasse-deepening via meltwater-induced hydrofracturing (Alley et al., 2005; van der Veen, 2007),
117 especially if crevasse fields laterally expand into perennial firn aquifer areas as a result of accelerated ice
118 flow (Colgan et al., 2016). Meltwater-induced hydrofracturing is an important component of supraglacial
119 lake drainage during the melting season (Das et al., 2008; Stevens et al., 2015) leading to at least temporary
120 accelerated flow velocities (Zwally et al., 2002; Joughin et al., 2013; Moon et al., 2014) and mass balance
121 changes (Joughin et al., 2008). Greenland's firn perennial aquifers may also support meltwater-induced
122 hydrofracturing, even during the freezing season (Poinar et al., 2017; 2019).

123 Recently, mapping the extent of Greenland's perennial firn aquifers from space was demonstrated using
124 satellite L-band microwave radiometry (Miller et al., 2020). Exponentially decreasing temporal L-band
125 signatures observed in enhanced-resolution vertically-polarized L-band brightness temperature (T_V^B)
126 imagery (2015-2016) generated using observations collected over the GrIS by the microwave radiometer
127 on NASA's SMAP satellite (Long et al., 2019) were correlated with a single year of perennial firn aquifer
128 detections (Miège et al. 2016). These detections were identified via the Center for Remote Sensing of Ice
129 Sheets (CReSIS) Multi-Channel Coherent Radar Depth Sounder (MCoRDS) flown by NASA's Operation
130 Ice Bridge (OIB) campaigns (Miège et al. 2016; Rodriguez-Morales et al, 2014). An empirical algorithm to
131 map extent was developed by fitting temporal L-band signatures to a set of sigmoidal curves derived from
132 the continuous logistic model.

133 The relationship between the radiometric, and thus the physical, temperature of perennial firn
134 aquifer areas, as compared to other percolation facies areas, forms the basis of the empirical algorithm.
135 Miller et al. (2020) hypothesized that the dominant control on the relatively slow exponential rate of T^B
136 decrease over perennial firn aquifer areas is physical temperature versus depth. L-band emissions from
137 the radiometrically warm upper snow and firn layers decreases during the freezing season as embedded
138 ice structures slowly refreeze at increased depths below the ice sheet surface. In the percolation facies,
139 refreezing of seasonal meltwater results in the formation of an intricate network of embedded ice structures
140 that are large (1-10-100 cm long, ~10-20 cm wide; Jezek et al., 1994) relative to the L-band wavelength
141 (~21 cm). Embedded ice structures induce strong volume scattering (Rignot et al., 1993; Rignot 1995) that
142 decreases T^B (Zwally, 1977; Swift et al. 1985; Jezek et al., 2018).

143

144 Ice slabs are ~1 m-16 m thick nearly-continuous ice layers that that can form on spatial scales as
145 large as tens of kilometers form on spatial scales of kilometers as a result of surface and subsurface water-
146 saturated snow and firn layers sequentially refreezing following multiple melting seasons (Machguth et al.,
147 2016; McFerrinMacFerrin et al., 2019). Over time, they become dense low-permeability solid-ice layers
148 overlying deeper permeable firn layers. Similar to perennial firn aquifers, Ice slabs ice slabs have been
149 observed identified via field expeditions and airborne ice-penetrating radar surveys ice-penetrating airborne
150 radar surveys in the lower-elevation (<~2000 m a.s.l.) percolation facies of the GrIS. They form at depths
151 from between ~1 m m and 20 m beneath the ice sheet surface (MacFerrin et al., 2019). They exist in in

152 areas that experience intense seasonal surface melting and rain (~266-573 mm w.e. ~~mm~~ yr−1) during the

153 melting season, and lower snow accumulation (< ~572 +/-32 mm w.e. yr−1) during the freezing season as

154 compared to perennial firn aquifer areas (MacFerrin et al., 2019). ~~Particularly in areas that experience~~

155 ~~intense seasonal surface melting (>600 mm yr−1) during the melting season, and lower snow accumulation~~

156 ~~(<600 mm yr−1) during the freezing season as compared to perennial firn aquifer areas (McFerrin et al.,~~

157 ~~2019).~~ Lower snow accumulation in ice slab areas results in a seasonal snow layer that is insufficiently thick

158 to thermally insulate water-saturated firn layers and seasonal meltwater is instead stored as embedded ice.

159 Refreezing of seasonal meltwater by the descending winter cold wave, and the subsequent formation of

160 ice slabs as well as other embedded ice structures within the upper snow and firn layers is the single source

161 of latent heat ~~in ice slab areas~~. While ice slab areas are radiometrically warmer than other percolation facies

162 areas with a lower volumetric fraction of embedded ice, they are radiometrically colder than perennial firn

163 aquifer areas. This results in typically higher observed T^B at the ice sheet surface during the freezing

164 season in ice slab areas, as compared to other percolation facies areas, however, typically lower observed

165 T^B as compared to perennial firn aquifer areas. Similar to temporal L-band signatures over perennial firn

166 aquifer areas, temporal L-band signatures over ice slab areas are exponentially decreasing during the

167 freezing season, however, the rate of T^B ~~is~~ decrease is slightly more rapid.~~Lower snow accumulation in ice~~

168 ~~slab areas results in a seasonal snow layer that is insufficiently thick to thermally insulate water-saturated~~

169 ~~firn layers and seasonal meltwater is instead stored as embedded ice. Refreezing of seasonal meltwater~~

170 ~~by the descending winter cold wave, and the subsequent formation of ice slabs as well as other embedded~~

171 ~~ice structures within the upper snow and firn layers is the single source of latent heat in ice slab areas.~~

172 ~~While ice slab areas are radiometrically warmer than other percolation facies areas with a lower volumetric~~

173 ~~fraction of embedded ice, they are radiometrically colder than perennial firn aquifer areas. This results in a~~

174 ~~lower observed T^B at the ice sheet surface during the freezing season.~~

175 In this study, we exploit the observed sensitivity of L-band emission to differences in the depth- and

176 time-integrated dielectric and geophysical properties of the percolation facies of the GrIS to map perennial

177 firn aquifer and ice slab areas together as a continuous englacial firn hydrological system using satellite L-

178 band microwave radiometry

179 ~~In this study, we exploit the observed sensitivity of L-band emissions to differences in the depth-~~

180 ~~and time- integrated dielectric and geophysical properties of the percolation facies of the GrIS to map~~

181 ~~perennial firn aquifer and ice slab areas together as a continuous englacial firn hydrological system using~~

182 ~~satellite L-band microwave radiometry.~~

183 **2   Methods**

184

185 ~~Consistent with recent seasonal surface melting trends, meltwater run-off has accelerated to~~

186 ~~become the dominant mass loss mechanism over the GrIS (van den Broeke et al., 2016). However,~~

187 ~~significant uncertainty remains in meltwater run-off estimates in the percolation facies as a result of the lack~~

188 ~~of knowledge of heterogeneous infiltration processes within the snow and firn layers (Pfeffer and Humphrey,~~

189 1996), the depths to which meltwater can descend beneath the ice sheet surface (Humphrey et al., 2012),
190 and the formation of englacial firn hydrological features (Benson et al., 1960; Humphrey et al., 2012; Forster
191 et al., 2014), especially ice layers and ice slabs (Machguth et al., 2016, McFerrin et al., 2019; Culberg et
192 al., 2021). A notable example of this lack of knowledge is the identification by Forster et al., (2014) of
193 widespread perennial firn aquifers within the percolation facies of the GrIS via airborne ice-penetrating radar
194 surveys collected by NASA's OIB campaigns (2010-2014; Rodriguez-Morales et al, 2014) that store large
195 volumes (~140 Gt; Koenig et al., 2014) of meltwater that was previously unknown. The mapped extent
196 (2010-2014) shown in Forster et al., (2014) can be distinctly observed in 1978 enhanced resolution Ku-
197 band radar backscatter imagery (Long and Drinkwater, 1994) collected by the radar scatterometer on
198 NASA's first Earth-observing satellite – the Seasat-A mission (Jones et al., 1982). This suggests that
199 Greenland's perennial firn aquifers have likely existed undetected in the deeper firn layers of the percolation
200 facies for decades. Meltwater storage in both solid (i.e., embedded ice structures) and liquid (i.e., perennial
201 firn and perched firn aquifers) form can buffer meltwater run-off in the percolation facies (Harper et al.,
202 2012). However, the formation of near-surface ice layers and ice slabs reduces the pore space within the
203 upper snow and firn layers and facilitates lateral meltwater flow with minimum vertical percolation into the
204 deeper firn layers, thus enhancing meltwater run-off downslope towards the periphery. Lateral meltwater
205 flow across ice layers overlying deeper permeable firn layers was first postulated by Müller (1962). The
206 theory was then further developed by Pfeffer et al., (1991) as an end-member case for meltwater run-off,
207 with the other end-member case being lateral meltwater flow across superimposed ice in the wet snow
208 facies and/or across glacial ice in the ablation facies. McFerrin et al., (2019) recently identified widespread
209 near-surface ice slabs within the percolation facies of the GrIS via airborne ice-penetrating radar surveys
210 collected by NASA's OIB campaigns (2010-2014; Rodriguez-Morales et al, 2014). Lateral meltwater flow
211 and high-elevation (~1850 m.a.s.l) meltwater run-off across the identified ice slabs was also observed in
212 visible satellite imagery collected by the NASA-USGS Landsat 7 mission (e.g. Goward et al., 2001). This
213 was also observed during the anomalous 2012 melting season (McFerrin et al., 2019) during which
214 seasonal surface melting extended over ~99% of the GrIS (Nghiem et al., 2012)

215 In this study, we use enhanced-resolution L-band T_V^B imagery (2015-2019) generated using
216 observations collected over the GrIS by the microwave radiometer on the SMAP satellite (Brodzik et al.,
217 2019) to map ice sheet-wide englacial firn hydrological features within the percolation facies. First, Wwe
218 adapt our previously developed empirical algorithm to map the extent of Greenland's perennial firn aquifers
219 (Miller et al., 2020) using a multi-year calibration technique.. We use enhanced-resolution L-band T_V^B
220 imagery (2015-201921) generated using observations collected over the GrIS by the microwave radiometer
221 on NASA's SMAP satellite (Long et al., 2019) and airborne ice-penetrating radar surveys collected by
222 NASA's OIB campaigns (Rodriguez-Morales et al, 2014). First, wWe correlate: (1) a 'firn saturation'
223 parameter derived from a simplecorrelate two-layer L-band brightness temperature model (Miller et al.,
224 2021, in press), (2) maximum and (3) minimum T_V^B values, and (4) exponentially decreasing temporal L-
225 band signatures, with five years of perennial firn aquifer detections (2010-2014) identified via the CReSIS

226 Accumulation Radar (AR) ~~flown by NASA's OIB campaigns~~ (Miège et al. 2016), and three years of

227 additional detections (2015-2017) more recently identified via MCoRDS (Miller et al., 2020). Next, we

228 extend our empirical algorithm to ~~also~~ map the extent of ice slab ~~and perched firn aquifer~~ areas. ~~We identify~~

229 ~~distinct temporal L-band signatures in T_V^B time series over ice slab detections (2010-2014) recently identified~~

230 ~~via AR (McFerrin et al., 2019). Similar to temporal L-band signatures over perennial firn aquifer areas,~~

231 ~~temporal L-band signatures over ice slab areas are exponentially decreasing during the freezing season,~~

232 ~~however, the rate of T_V^B decrease is slightly more rapid.~~ We correlate the SMAP-derived parameters ~~these~~

233 ~~relatively rapidly exponentially decreasing temporal L-band signatures~~ with five years of ice slab detections

234 (2010-2014) recently identified via AR (MacFerrin et al., 2019~~AR-derived ice slab detection)~~)s. ~~Additionally,~~

235 ~~we correlate exponentially decreasing temporal L-band signatures with AR- and MCoRDS derived~~

236 ~~detections where perennial firn aquifer and ice slab areas overlap. We identify these transitional areas as~~

237 ~~perched firn aquifer areas. We infer that, in these areas, shallow water-saturated firn layers transiently form~~

238 ~~on top of buried ice slabs or other semi-impermeable layers, such as spatially coherent melt layers that~~

239 ~~form in the higher elevations (> ~2000 m.a.s.l.) of the percolation facies and the dry snow facies that were~~

240 ~~recently identified via AR (Culberg et al., 2021). Perched firn aquifers likely form during some melting~~

241 ~~seasons as a result of interannual variability in surface melting and snow accumulation, and the formation~~

242 ~~of englacial firn hydrological features.~~ Finally, we~~re~~ re-calibrate our empirical model ~~re-calibrate the~~

243 ~~sigmoidal curves~~ to map the extent of perennial firn aquifer and~~,~~ ice slab~~, and perched firn aquifer~~ areas

244 over the percolation facies~~.~~. Interannual variability in extent is not resolved in this study, however, it will be

245 explored further in future work.

246

247

248 facies of the GrIS

250 **2   Methods**

251 **2.1   The Soil Moisture Active Passive (SMAP) Mission**

252 The key science objectives of NASA's SMAP mission () are to map terrestrial soil moisture and freeze/thaw
253 state over Earth's land surfaces from space. However, the global L-band T^B observations collected by the
254 SMAP satellite also have many cryospheric applications. Mapping ice sheet-wide englacial firn hydrological
255 features over Earth's polar ice sheets represents an interesting analog and an innovative extension of the
256 science objectives. Measurements of moisture (i.e., defined in this study in terms of the volumetric fraction
257 of meltwater within the upper snow and firn layers of the percolation facies) and freeze-thaw state (i.e.,
258 defined in this study in terms of the firn saturation parameter (see Section 2.4.3) and the refreezing rate
259 parameter (see Section 2.4.4)) are critical to understanding the hydrospheric state over Earth's polar ice
260 sheets. Perennial firn aquifers, ice slabs, and perched firn aquifers represent recently identified components
261 of the hydrosphere that are capable of storing large volumes of meltwater in both solid and liquid form that
262 can initiate meltwater-induced hydrofracturing and accelerated ice flow as well as high-elevation run-off,
263 and impact the mass balance and stability of the GrIS. Critically, the majority of meltwater is stored at depths
264 that only L-band satellite microwave sensors (i.e., radiometers, radar scatterometers, and synthetic
265 aperture radars) are capable of detecting.

266   Previous and current satellite microwave radiometer, radar scatterometer, and synthetic aperture
267 radar missions that operate in the frequency range between 37 GHz (Ka-band) and 5.3 GHz (C-band) have
268 provided a multi-decadal (1978-present) record of multi-frequency T^B and radar backscatter observations
269 over Earth's polar ice sheets since the beginning of the satellite era. The most common geophysical
270 parameter mapped over ice sheets using these observations is the extent of seasonal surface melting. The
271 key difference between L-band and higher frequency satellite microwave sensors is penetration depth.
272 When the snow and firn layers are saturated with meltwater during the melting season, the penetration
273 depth of both L-band and higher frequency satellite microwave sensors is less than ~a meter. When surface
274 and subsurface water-saturated snow and firn layers and embedded ice structures subsequently refreeze,
275 the penetration depth of higher frequency satellite microwave sensors ranges from between ~centimeters
276 and meters. During the freezing season, water-saturated snow and firn layers either completely refreeze
277 (i.e., ice layers, ice slabs, spatially coherent melt layers) or underlay the refrozen upper snow and firn layers
278 of the percolation facies and descend to depths ranging from between ~1 m and 40 m (Miège et al., 2016)
279 beneath the ice sheet surface (i.e., perennial and perched firn aquifers). While the upper surface of stored
280 meltwater in some perennial and perched firn aquifers may remain at depths that are shallow enough to be
281 directly detected by C-band satellite microwave sensors, the mean depth just prior to melt onset (~22 m;

282 Miège et al., 2016) is too deep to be detected at this wavelength. L-band satellite microwave sensors can
283 detect perennial firn aquifers from as much as an order of magnitude deeper than can be observed by C-
284 band satellite microwave radiometers. Deep enough to directly detect the upper surface of stored meltwater
285 over the entire depth range mapped by airborne ice-penetrating radar surveys over the GrIS.
286
287 **2.1 SMAP EEnhanced-RResolution L-band T^B IImagery**

288 The key science objectives of NASA's SMAP mission (https://smap.jpl.nasa.gov/) are to map terrestrial soil
289 moisture and freeze/thaw state over Earth's land surfaces from space. However, the global L-band T^B
290 observations collected by the SMAP satellite also have cryospheric applications. Mapping perennial firn
291 aquifer and ice slab areas over Earth's polar ice sheets represents an interesting analog and an innovative
292 extension of thethe science objectives. NASA's The SMAP satellite was launched 31 January 2015 and
293 carries a microwave radiometer that operates at an L-band frequency of 1.41 GHz (L-band) (Enkentabi et
294 al., 2010). It is currently collecting observations of vertically and horizontally-polarized T^B over Greenland.
295 The surface incidence angle is ~40°, and the radiometric accuracy is approximately ~1.3 K (Piepmeier et
296 al., 2017).
297
298 The Scatterometer Image Reconstruction (SIR) algorithm was originally developed to reconstruct
299 coarse resolution satellite radar scatterometry imagery on a higher spatial resolution grid (Long et al., 1993;
300 Early and Long, 2001). The SIR algorithm has been adapted for coarse resolution satellite microwave
301 radiometry imagery (Long and Daum, 1998; Long and Brodzik, 2016; Long et al., 2019). The microwave
302 radiometer form of the SIR algorithm (rSIR) exploits uses the measurement response function (MRF) for
303 each observation, which is a smeared version of the antenna pattern. Using the overlapping MRFs, the
304 rSIR algorithm reconstructs T^B from the spatially filtered low-resolution sampling provided by the
305 observations. In effect, it generates an MRF-deconvolved T^B image. Combining multiple multiple orbital
306 passes increases the sampling density, which improves both the accuracy and resolution of the SMAP
307 enhanced-resolution T^B imagery (Long et al., 2019).
308 Over Greenland, the rSIR algorithm combines satellite orbital passes that occur between 8 a.m.
309 and 4 p.m. local time-of-day to reconstruct SMAP enhanced-resolution T^B imagery twice-daily (i.e., morning
310 and evening orbital pass interval, respectively). T^B imagery is projected on a Northern Hemisphere (NH)
311 Equal-Area Scalable Earth Grid (EASE-Grid 2.0; Brodzik et al., 2012) at a 3.125 km rSIR grid cell spacing
312 (e.g., Fig. 1). The effective resolution for each grid cell is dependent on the number of observations used
313 in the rSIR reconstruction and is coarser than the rSIR grid cell spacing. While the effective resolution of
314 conventionally processed SMAP T^B imagery posted on a 25 km grid is approximately ~30 km (e.g., Fig.
315 1a), the effective resolution of SMAP enhanced-resolution T^B imagery posted on a 3.125 km grid is
316 approximately ~18 km (e.g., Fig. 1b), an improvement of ~60% (Figs. 1; 2) (Long et al., 2020).
317 As previously noted, fFor our analysis of the percolation facies, we use SMAP enhanced-resolution

318 T_V^B imagery over the GrIS. Compared to the horizontally-polarized channel, the vertically-polarized channel

319 exhibits decreased sensitivity to variability in the volumetric fraction of meltwater, which is attributed to

320 reflection coefficient differences between channels (Miller et al., 2020). Using the vertically polarized

321 channel also results in a reduced chi-squared error statistic when fitting T_V^B time series to the sigmoid

322 function (see Section 2.34.45). We construct T_V^B imagery that alternate morning and evening orbital pass

323 observations annually, beginning and ending just prior to melt onset. The Greenland Ice Mapping Project

324 (GIMP) Land Ice and Ocean Classification Mask and Digital Elevation Model (Howat et al., 2014) are

325 differences between channels (Miller et al., 2020). Using the vertically polarized channel also results in a

326 reduced chi-squared error statistic when fitting T_V^B time series to the sigmoid function (see Section 2.4.5).

327 We construct T_V^B imagery that alternate morning and evening orbital pass observations annually, beginning

328 and ending just prior to melt onset. The Greenland Ice Mapping Project (GIMP) Land Ice and Ocean

329 Classification Mask and Digital Elevation Model (Howat et al., 2014) are projected on a NH EASE-Grid 2.0

330 at a 3.125 km rSIR grid cell spacing. T_V^B imagery between 1 April 2015 and 31 March 2019 are ice sheet-

331 masked, and an elevation for each rSIR grid cell is calculated.

332

333 **2.2   Airborne Ice-Penetrating Radar Surveys**

334 Miller et al., (2020) calibrated the empirical algorithm to map the extent of Greenland's perennial firn

335 aquifers by correlating a single year of exponentially decreasing temporal L-band signatures (2015-2016)

336 with coincident perennial firn aquifer detections (2016) identified via MCoRDS. Here, we extend and expand

337 the calibration of our adapted empirical algorithm to include four years of exponentially decreasing temporal

338 L-band signatures (2015-2019) correlated with eight years of perennial firn aquifer detections (2010-2017)

339 and five years of ice slab detections (2010-2014) identified via AR and MCoRDS (Fig. 1c). Our multi-year

340 calibration technique projects perennial firn aquifer and ice slab detections on three separate NH EASE-

341 Grids 2.0 at an rSIR grid cell spacing of 3.125 km, consistent with the rSIR grid cell spacing of the SMAP

342 enhanced-resolution L-band T_V^B imagery. Interannual variability is not resolved in this study, however, it will

343 be explored further in future work.

344   An advantage of the multi-year calibration technique as compared to the single-coincident year

345 calibration technique (Miller et al., 2020) is that it increases the number of rSIR grid cells that can be

346 assessed. It also provides repeat targets that can account for variability in the dielectric and geophysical

347 properties that seasonally influence the radiometric temperature and temporal L-band signatures in stable

348 perennial firn aquifer, ice slab, and perched firn aquifer areas. Uncertainty is introduced by correlating

349 exponentially-decreasing temporal L-band signatures with AR- and MCoRDS-derived detections that are

350 not coincident in time. The multi-year calibration technique assumes the extent of each area remains stable,

351 which is not necessarily the case as climate extremes (Cullather et al., 2020) can influence each of these

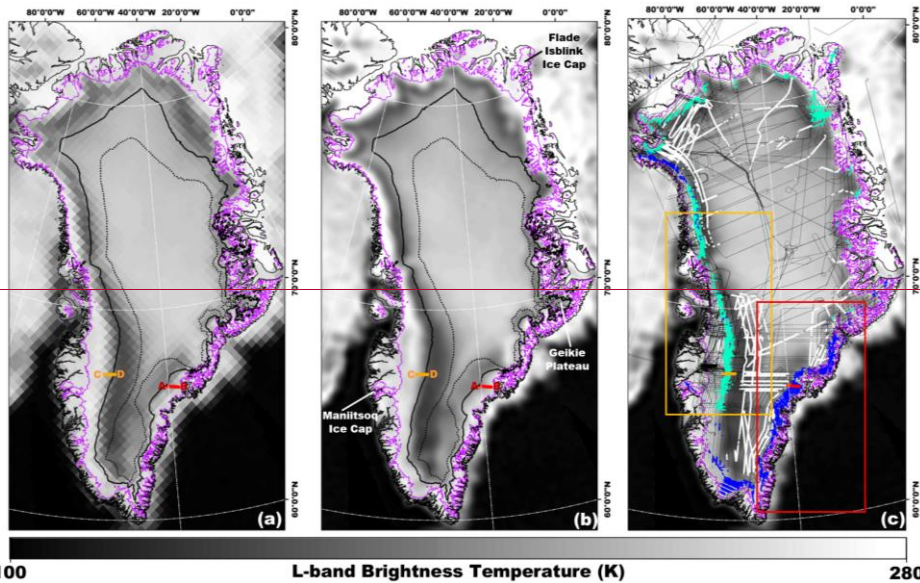
352 sub-facies. The assumption of stability neglects boundary transitions in the extent of perennial firn aquifer

353 areas associated with refreezing of shallow water-saturated firn layers, englacial drainage of meltwater into

354 crevasses at the periphery (Poinar et al., 2017; Poinar et al, 2019), and transient upslope expansion

355 (Montgomery et al., 2017). Once formed, ice slabs are essentially permanent features within the upper
356 snow and firn layers of the percolation facies until they are compressed into glacial ice. However, they may
357 transition into superimposed ice at the lower boundary of ice slab areas or rapidly expand upslope,
358 particularly following extreme melting seasons (McFerrin et al., 2019). By our definition, perched firn
359 aquifers are transient features. Thus, we simply consider our mapped extent a high-probability area for the
360 preferential formation of each of these sub-facies within the broader percolation facies, with continued
361 presence dependent on seasonal surface melting and snow accumulation in subsequent years.

362
363

364 ### ~~Figure 1~~

365 *(a) Gridded (25 km), and (b) enhanced-resolution (3.125 km) L-band T_V^B imagery generated using*
366 *observations collected 15 April 2016 by the microwave radiometer on the SMAP satellite during the evening*
367 *orbital pass interval over Greenland (Brodzik et al., 2019). The solid black line is the 2000 m.a.s.l. contour,*
368 *and the black dotted line is the 2500 m.a.s.l. contour (Howat et al., 2014). The purple line is the ice sheet*
369 *extent (Howat et al., 2014). The black peripheral line is the coast of Greenland and adjacent Ellesmere*
370 *Island (Wessel and Smith, 1996). The whiter regions of higher T_V^B over the high-elevation (> 2500 m.a.s.l.)*
371 *interior are the dry snow facies. The darker grey regions of lower T_V^B are the percolation facies, including*
372 *ice slabs and perched firn aquifer areas. The whiter regions of higher T_V^B over the coastal areas, peripheral*
373 *ice caps (e.g., Maniitsoq and Flade Isblink) and nearby islands are perennial firn aquifers, superimposed*
374 *or glacial ice, land, or spatially integrated L-band emissions. The whiter regions of higher T_V^B outside the ice*
375 *sheet extent are sea ice. (c) The SMAP enhanced-resolution L-band T_V^B imagery is overlaid with AR- and*
376 *MCoRDS-derived 2010-2017 perennial firn aquifer (blue shading; Miège et al., 2016), 2010-2014 ice slab*
377 *(cyan shading; ~~McFerrin~~MacFerrin et al., 2019), and 2012 spatially coherent melt layer (white shading;*
378 *Culberg et al., 2021) detections along OIB flight lines (black lines). Overlapping perennial firn aquifer and*
379 *ice slab detections are interpreted as perched firn aquifer areas. The red and orange boxes in (c) are zoom*
380 *areas over south eastern Greenland (Fig. 2a), and south western Greenland (Fig. 2b), respectively. The*
381 *red line is AR radargram profile along perennial firn aquifer transect A-B (Fig. 3a). The orange line is AR*

382 *radargram profile along ice slab transect C-D (Fig. 3b).*

Figure 2

*Enhanced-resolution (3.125 km) L-band T_V^B imagery generated using observations collected 15 April 2016 by the microwave radiometer on the SMAP satellite during the evening orbital pass interval over (a) south eastern Greenland (Fig. 1c; zoom area in red box), and (b) south western Greenland (Fig. 1c; zoom area in orange box) (Brodzik et al., 2019). The purple line is the ice sheet or ice cap extent (Howat et al., 2014). The black peripheral line is the coast (Wessel and Smith, 1996). (c) The SMAP enhanced-resolution L-band T_V^B imagery is overlaid with AR- and MCoRDS-derived 2010-2017 perennial firn aquifer (blue shading; Miège et al., 2016), 2010-2014 ice slab (cyan shading; McFerrin*MacFerrin *et al., 2019), and 2012 spatially coherent melt layer (white shading; Culberg et al., 2021) detections along OIB flight lines (black lines). Overlapping perennial firn aquifer and ice slab detections are interpreted as perched firn aquifer areas. The red line is AR radargram profile along perennial firn aquifer transect A-B (Figs. 1; 3a). The orange line is AR radargram profile along ice slab transect C-D (Figs. 1; 3b). The blue circle is a perennial firn aquifer area (Figs. 3a; 4a). The cyan circle is a perched firn aquifer area (Figs. 3b; 4b). The orange diamond is a percolation facies area (Fig. 4c). The red triangle is a high-elevation (~2500 m.a.s.l.) percolation facies area*

398   *(Fig. 4d). The yellow circle is a superimposed ice area (Fig. 4e).*

399 projected on the~~a~~ NH EASE-Grid 2.0 at a 3.125 km rSIR grid cell spacing. T_V^B imagery between 1 April 2015

400 and 31 March 2019 are ice sheet-masked, and an elevation for each rSIR grid cell is calculated.

2.2 Airborne Ice-Penetrating Radar Surveys

Miller et al., (2020) calibrated the empirical algorithm to map the extent of Greenland's perennial firn aquifers by correlating a single year of exponentially decreasing temporal L-band signatures (2015-2016) with coincident perennial firn aquifer detections (2016) identified via MCoRDS. Here, we extend and expand the calibration of our adapted empirical algorithm to include four years of exponentially decreasing temporal L-band signatures (2015-2019) correlated with eight years of perennial firn aquifer detections (2010-2017) and five years of ice slab detections (2010-2014) identified via AR and MCoRDS (Fig. 1c).

Our multi-year calibration technique projects perennial firn aquifer and ice slab detections on three separate NH EASE-Grids 2.0 at an rSIR grid cell spacing of 3.125 km, consistent with the rSIR grid cell spacing of the SMAP enhanced-resolution L-band T_V^B imagery. Interannual variability is not resolved in this study, however, it will be explored further in future work.

An advantage of the multi-year calibration technique as compared to the single-coincident year calibration technique (Miller et al., 2020) is that it increases the number of rSIR grid cells that can be assessed. It also provides repeat targets that can account for variability in the dielectric and geophysical properties that seasonally influence the radiometric temperature and temporal L-band signatures in stable perennial firn aquifer, ice slab, and perched firn aquifer areas. Uncertainty is introduced by correlating exponentially decreasing temporal L-band signatures with AR- and MCoRDS-derived detections that are not coincident in time. The multi-year calibration technique assumes the extent of each area remains stable, which is not necessarily the case as climate extremes (Cullather et al., 2020) can influence each of these sub-facies. The assumption of stability neglects boundary transitions in the extent of perennial firn aquifer areas associated with refreezing of shallow water-saturated firn layers, englacial drainage of meltwater into crevasses at the periphery (Poinar et al., 2017; Poinar et al, 2019), and transient upslope expansion (Montgomery et al., 2017). Once formed, ice slabs are essentially permanent features within the upper snow and firn layers of the percolation facies until they are compressed into glacial ice. However, they may transition into superimposed ice at the lower boundary of ice slab areas or rapidly expand upslope, particularly following extreme melting seasons (McFerrin~~MacFerrin~~ et al., 2019). By our definition, perched firn aquifers are transient features. Thus, we simply consider our mapped extent a high-probability area for the preferential formation of each of these sub-facies within the broader percolation facies, with continued presence dependent on seasonal surface melting and snow accumulation in subsequent years Annual perennial firn aquifer and ice slab detections that may introduce significant uncertainty into the multi-year calibration technique include those following the 2010 melting season, which was exceptionally long (Tedesco et al., 2010), the anomalous 2012 melting season (Nghiem et al., 2012), and the 2015 melting season which was especially intense in western and northern Greenland (Tedesco et al., 2016). Following these extreme melting seasons, significant changes in the dielectric and geophysical properties likely occurred across large portions of the GrIS, including perennial firn aquifer recharging resulting in increases in meltwater volume and decreases in the depth to the upper surface of stored meltwater. The formation of expansive near-surface ice slabs (McFerrin~~MacFerrin~~ et al., 2019) likely resulted in the formation of more extensive perched firn aquifers during subsequent melting seasons. The upper snow and firn layers of the dry snow facies and percolation facies were also saturated with relatively large volumetric fractions of meltwater as compared to the negligible to limited

volumetric fractions of meltwater that percolates during more typical seasonal surface melting on the GrIS. Seasonal meltwater was refrozen into spatially coherent melt layers following the 2010 and 2012 melting seasons (Culberg et al., 2021) as well as following the 2015 and 2018 melting seasons (i.e., identified as part of the temporal L-band signature analysis in this study; see Section 2.4.2).

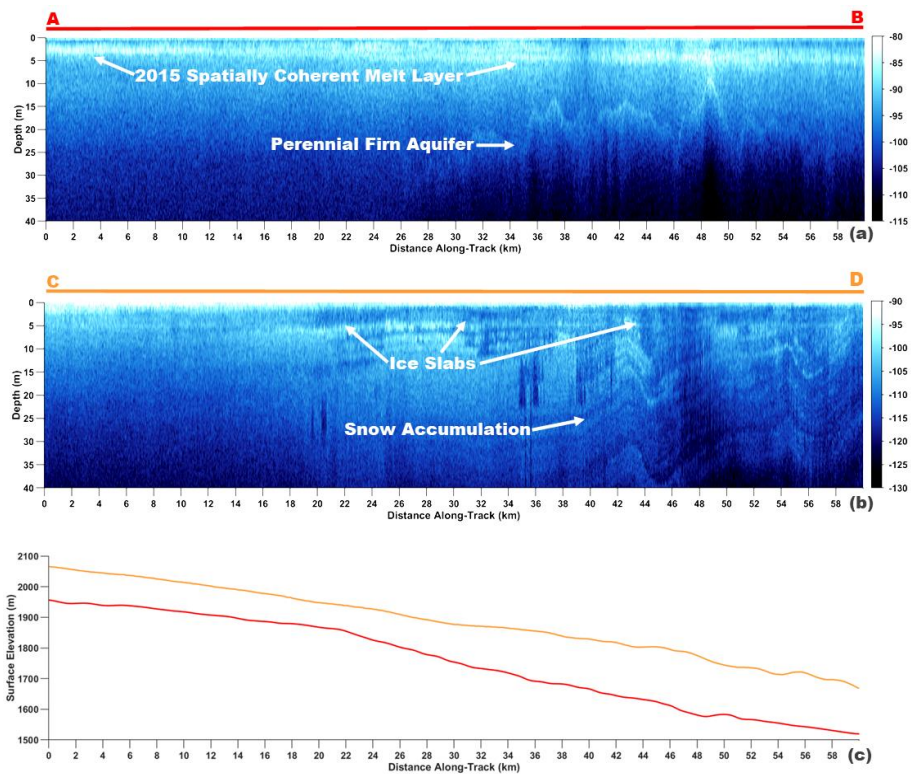
As compared to ice slabs, which are dense low-permeability solid-ice layers, spatially coherent melt layers are a network of embedded ice structures primarily consisting of discontinuous horizontally-oriented ice layers and ice lenses sparsely connected via vertical-oriented ice pipes (Culberg et al., 2021). Ice slabs are relatively thick (~1 m – 16 m) and form in the high-elevation percolation facies (~2100 m.a.s.l.) at depths of between ~1 m and 20 m beneath the ice sheet surface following intense seasonal surface melting over multiple melting seasons (McFerrinMacFerrin et al., 2019). Spatially coherent melt layers are relatively thin (~0.02 cm - 2 m) and can rapidly form across the entire high-elevation dry snow facies (~3200 m.a.s.l ; Nghiem et al., 2012) at depths of less than ~1 m beneath the ice sheet surface following a single extreme melting season. They can further merge together into thicker solid-ice layers following multiple extreme melting seasons (Culberg et al., 2021). Similar to ice slabs, the formation of spatially coherent melt layers reduces the pore space within the upper snow and firn layers and may also facilitate lateral meltwater flow with minimum vertical percolation into the deeper firn layers, thus enhancing meltwater run-off from significantly higher elevations downslope towards the periphery on accelerated time scales. The formation of spatially coherent melt layers overlying deeper perennial firn aquifers (e.g., Fig. 3a) will limit or terminate gravity-driven meltwater drainage and seasonal recharging (Fountain and Walder, 1998), which may eventually completely refreeze stored meltwater into decimeters thick solid-ice layers overlying deeper glacial ice. Spatially coherent melt layers are exceptionally bright in AR radargrams (e.g., Fig 3a). The large dielectric contrast between the spatially coherent melt layer and the overlying, underlying, and interior snow and firn layers results in high reflectivity at the interfaces. However, electromagnetic energy still propagates downward through the high reflectivity layer into the deeper firn layers. Culberg et al., 2021) recently demonstrated mapping the extent of the spatially coherent melt layer formed following the anomalous 2012 melting season (Nghiem et al., 2012) via AR (Figs. 1c; 2). AR and MCoRDS (Rodriguez-Morales et al, 2014) were flown over the GrIS on a P-3 aircraft in April and May between 2010 and 2017 (Figs. 1c; 2). The AR instrument operates at a center frequency of 750 MHz with a bandwidth of 300 MHz, resulting in a range resolution in firn of 0.53 m (Lewis et al., 2015). The collected data have an along-track resolution of approximately ~30 m with 15 m spacing between traces in the final processed radargrams. At a nominal flight altitude of 500 m above the ice sheet surface, the cross-track resolution varies between 20 m for a smooth surface, to 54 m for a rough surface with no appreciable layover. The MCoRDS instrument operated at three different frequency configurations: (1) a center with a bandwidth of 30 MHz (2010-2014, 2017, 2018), (2) a center frequency of 315 MHz with a band width of 270 MHz (2015), and (3) a center frequency of 300 MHz with a bandwidth of 300 MHz (2016). The vertical range resolution in firn for each of these frequency configurations is 5.3 m, 0.59 m, and 0.53m, respectively (CReSIS, 2016). The collected data have an along-track resolution of approximately 25 m with 14 m spacing between traces in the final processed radargrams. At a nominal flight altitude of 500 m, the cross-track resolution varies between 40

480 m for a smooth surface in the highest bandwidth configuration, to 175 m for a rough surface with no
481 appreciable layover in the lowest bandwidth configuration.
482   The multi-year calibration technique uses perennial firn aquifer detections previously identified

483 along OIB flight lines via AR (2010-2014) and MCoRDS (2015-2017) radargram profiles and the
484 methodology described in Miège et al. (2016). Bright lower reflectors that undulate with the local
485 topographic gradient underneath which reflectors are absent in the percolation facies are interpreted as the
486 upper surface of meltwater stored within perennial firn aquifers (e.g., Fig. 3a). The large dielectric contrast
487 between refrozen and water-saturated firn layers results in high reflectivity at the interface. However, the

488 presence of meltwater increases attenuation, limiting the downward propagation of electromagnetic energy
489 through the water-saturated firn layer. The total number of AR derived perennial firn aquifer detections is
490 325,000, corresponding to a total extent of 98 km^2. The analysis assumes a smooth surface, which is typical
491 of much of the percolation facies, and a grid cell size of 15 m x 20 m. The total number of MCoRDS-derived
492 perennial firn aquifer detections is 142,000, corresponding to a total extent of 80 km^2. This analysis also
493 assumes a smooth surface, and a grid cell size of 14 m x 40 m. The combined total number of grid cells
494 (467,000) and total extent (178 km^2) is significantly larger than the total number of MCoRDS-derived grid
495 cells (78,000) and total extent (44 km^2) calculated for 2016 (Miller et al., 2020). Perennial firn aquifer
496 detections are mapped in western, southern, and south and central eastern Greenland as well as the
497 Maniitsoq and Flade Isblink Ice Caps (Figs. 1c; 2a). We project AR- and MCoRDS-derived perennial firn
498 aquifer detections on the NH EASE-Grid 2.0 at an rSIR grid cell spacing of 3.125 km. Each rSIR grid cell
499 has an extent of approximately 10 km^2. The total number of rSIR grid cells with at least one perennial firn
500 aquifer detection is 800, corresponding to a total extent of 8000 km^2. However, given the limited AR and
501 MCoRDS grid cell coverage, less than 1% of the rSIR grid cell extent has airborne ice-penetrating radar
502 survey coverage. As compared to the total resolution varies between 20 m for a smooth surface, to 54 m
503 for a rough surface with no appreciable layover. The

504

505 **Figure 3**

506 *AR radargram profile (Rodriguez-Morales et al, 2014) (a) along perennial firn aquifer transect A-B (red line,*
507 *Figs. 1; 2a) that was collected on 22 April 2017, and (b) ice slab transect C-D (orange line, Figs. 1; 2b) that*
508 *was collected on 5 May 2017. (c) The corresponding perennial firn aquifer transect A-B elevation profile*
509 *(red line), and ice slab transect C-D elevation profile (orange line). The exceptionally bright upper surface-*
510 *parallel reflector in (a) is interpreted as a spatially coherent melt layer that formed following the 2015 melting*
511 *season. The bright lower reflector in (a) is interpreted as the upper surface of meltwater stored within a*
512 *perennial firn aquifer. Thick dark surface-parallel regions of low-reflectivity in (b) are interpreted as ice slabs.*
513 *Alternating sequences of bright and dark surface-parallel reflectors in (b) are interpreted as seasonal snow*
514 *accumulation layers. A first maximum after maximum gradient re-tracker is used to identify the surface*
515 *return in each profile. Each profile is flattened so that the depth axis is measured relative to the local*
516 *elevation. Corresponding elevation profiles in (c) are calculated by subtracting the radar-measured flight*
517 *clearance over the ice sheet from the aircraft's global positioning system altitude measurements that were*
518 *coincidently collected along each transect.*

AR and MCoRDS (Rodriguez-Morales et al, 2014) were flown over the GrIS on a P-3 aircraft in April and May between 2010 and 2017. The AR instrument operates at a center frequency of 750 MHz with a bandwidth of 300 MHz, resulting in a range resolution in firn of 0.53 m (Lewis et al., 2015). The collected data have an along-track resolution of ~30 m with 15 m spacing between traces in the final processed radargrams. At a nominal flight altitude of 500 m above the ice sheet surface, the cross-track resolution varies between 20 m for a smooth surface, to 54 m for a rough surface with no appreciable layover. The MCoRDS instrument operated at three different frequency configurations: (1) a center frequency of 195 MHz with a bandwidth of 30 MHz (2010-2014, 2017, 2018), (2) a center frequency of 315 MHz with a band width of 270 MHz (2015), and (3) a center frequency of 300 MHz with a bandwidth of 300 MHz (2016). The vertical range resolution in firn for each of these frequency configurations is 5.3 m, 0.59 m, and 0.53m, respectively (CReSIS, 2016). The collected data have an along-track resolution of ~25 m with 14 m spacing between traces in the final processed radargrams. At the same nominal flight altitude of 500 m, the cross-track resolution varies between 40 m for a smooth surface in the highest bandwidth configuration, to 175 m for a rough surface with no appreciable layover in the lowest bandwidth configuration.

The multi-year calibration technique uses perennial firn aquifer detections previously identified along OIB flight lines via AR (2010-2014) and MCoRDS (2015-2017) radargram profiles and the methodology described in Miège et al. (2016). Bright lower reflectors that undulate with the local topographic gradient underneath which reflectors are absent in the percolation facies are interpreted as the upper surface of meltwater stored within perennial firn aquifers (e.g., Fig. 3a). The large dielectric contrast between refrozen and water-saturated firn layers results in high reflectivity at the interface. However, the presence of meltwater increases attenuation, limiting the downward propagation of electromagnetic energy through the water-saturated firn layer. The total number of AR derived perennial firn aquifer detections is ~325,000, corresponding to a total extent of ~98 km². The analysis assumes a smooth surface, which is typical of much of the percolation facies, and a grid cell size of 15 m x 20 m. The total number of MCoRDS-derived perennial firn aquifer detections is ~142,000, corresponding to a total extent of ~80 km². This also assumes a smooth surface, and a grid cell size of 14 m x 40 m. The combined total number of grid cells (~467,000) and total extent (~178 km²) is significantly larger than the total number of MCoRDS-derived grid cells (~78,000) and total extent (~44 km²) calculated for 2016 (Miller et al., 2020). Perennial firn aquifer detections are mapped in western, southern, and south and central eastern Greenland as well as the Maniitsoq and Flade Isblink Ice Caps (Figs. 1c; 2a). We project AR- and MCoRDS-derived perennial firn aquifer detections on the NH EASE-Grid 2.0 at an rSIR grid cell spacing of 3.125 km. Each rSIR grid cell has an extent of ~10 km². The total number of rSIR grid cells with at least one perennial firn aquifer detection is ~800, corresponding to a total extent of ~8000 km². However, given the limited AR and MCoRDS grid cell coverage, less than ~1% of the rSIR grid cell extent has radargram coverage. As compared to the total number of MCoRDS-derived perennial firn aquifer detections (~780) calculated for 2016 (Miller et al., 2020), the total number of rSIR grid cells with at least one detection is only increased by ~20 for the multi-year calibration technique, corresponding to an increased total extent of ~200 km².

We also use ice slab detections previously identified along OIB flight lines via AR (2010-2014) radargram profiles and the methodology described in McFerrinMacFerrin et al. (2019) in the multi-year calibration technique. Thick dark surface-parallel regions of low-reflectivity in the percolation facies are interpreted as ice slabs (e.g., Fig. 3b). The large dielectric contrast between ice slabs and the overlying and underlying snow and firn layers results in high reflectivity at the interfaces. However, electromagnetic energy is not scattered or absorbed within the homogeneous ice slab, it instead propagates downward through the layer and into the deeper firn layers. The total number of AR-derived ice slab detections is ~505,000, corresponding to a total extent of ~283 km^2. Ice slab detections are mapped in western, central and north eastern, and northern Greenland as well as the Flade Isblink Ice Cap (Figs. 1c; 2b). We project the AR-derived ice slab detections on the NH EASE-Grid 2.0 at an rSIR grid cell spacing of 3.125 km. The total number of rSIR grid cells with at least one ice slab detection is ~2000, corresponding to a total extent of ~20,000 km^2. However, less than ~2% of the rSIR grid cell extent has airborne ice-penetrating radar surveyradargram coverage.

An advantage of the multi-year calibration technique as compared to the single-coincident year calibration technique (Miller et al., 2020) is that it increases the number of rSIR grid cells that can be assessed. It also provides repeat targets that can account for variability in the depth- and time-integrated dielectric and geophysical properties that influence the radiometric temperature in stable perennial firn aquifer and ice slab areas. Uncertainty is introduced by correlating the SMAP-derived parameters with AR- and MCoRDS-derived detections that are not coincident in time. The multi-year calibration technique assumes the extent of each area remains stable, which is not necessarily the case as climate extremes (Cullather et al., 2020) can influence each of these sub-facies. The assumption of stability neglects boundary transitions in the extent of perennial firn aquifer areas associated with refreezing of shallow water-saturated firn layers, englacial drainage of meltwater into crevasses at the periphery (Poinar et al., 2017; Poinar et al, 2019), and transient upslope expansion (Montgomery et al., 2017). Once formed, ice slabs are essentially permanent features within the upper snow and firn layers of the percolation facies until they are compressed into glacial ice. However, they may transition into superimposed ice at the lower boundary of ice slab areas or rapidly expand upslope, particularly following extreme melting seasons (MacFerrin et al., 2019). Thus, we simply consider our mapped extent a high-probability area for the preferential formation of each of these sub-facies, with continued presence dependent on seasonal surface melting and snow accumulation in subsequent years.

Annual perennial firn aquifer and ice slab detections that may introduce significant uncertainty into the multi-year calibration technique include those following the 2010 melting season, which was exceptionally long (Tedesco et al., 2010), the anomalous 2012 melting season, during which seasonal surface melting extended across 99% of the GrIS (Nghiem et al., 2012), and the 2015 melting season, which was especially intense in western and northern Greenland (Tedesco et al., 2016). Following these extreme melting seasons, significant changes in the dielectric and geophysical properties likely occurred across large portions of the GrIS, including perennial firn aquifer recharging resulting in increases in

meltwater volume and decreases in the depth to the upper surface of stored meltwater. The upper snow and firn layers of the dry snow facies and percolation facies were also saturated with relatively large volumetric fractions of meltwater as compared to the negligible to limited volumetric fractions of meltwater that percolates during more typical seasonal surface melting on the GrIS.

Seasonal meltwater was refrozen into spatially coherent melt layers following the 2010 and 2012 melting seasons (Culberg et al., 2021) as well as more recently following the 2015, and 2018 melting seasons identified as part of the temporal L-band signature analysis in this study, (Section 2.3.1). As compared to ice slabs, which are dense low-permeability solid-ice layers, spatially coherent melt layers are a network of embedded ice structures primarily consisting of discontinuous horizontally-oriented ice layers and ice lenses sparsely connected via vertical-oriented ice pipes (Culberg et al., 2021). Spatially coherent melt layers are relatively thin (0.02 cm-2 m) and can rapidly form across the high-elevation (up to 3200 m a.s.l.) dry snow facies at depths of less than 1 m beneath the ice sheet surface following a single extreme melting season. They can further merge together into thicker solid-ice layers following multiple extreme melting seasons. Spatially coherent melt layers are exceptionally bright in AR radargrams (e.g., Fig 3a). The large dielectric contrast between the spatially coherent melt layer and the overlying, underlying, and interior snow and firn layers results in high reflectivity at the interfaces. However, electromagnetic energy still propagates downward through the high reflectivity layer into the deeper firn layers. Culberg et al., (2021) recently demonstrated mapping the extent of spatially coherent melt layers formed following the 2012 melting season (Nghiem et al., 2012) via AR (Figs. 1c; 2).

We detect perched firn aquifer areas by comparing the AR- and MCoRDS-derived perennial firn aquifer and ice slab detections projected on the NH EASE-Grid 2.0 and then identify overlapping rSIR grid cells. The total number of AR-derived perched firn aquifer detections is ~75,000, corresponding to a total extent of ~23 km². The total number of MCoRDS-derived perched firn aquifer detections is ~20, corresponding to a near-negligible extent (~0.006 km²). Perched firn aquifer detections are mapped in western, and central eastern Greenland as well as the Flade Isblink Ice Cap (Figs. 1c; 2b).

The total number of rSIR grid cells with at least one perched firn aquifer detection is ~200, corresponding to a total extent of ~2000 km². However, similar to the other sub-facies, less than ~1% of the rSIR grid cell extent has radargram coverage. The total number of AR- and MCoRDS-derived perennial firn aquifer, ice slab, and perched firn aquifer detections that we project on three separate NH EASE-Grids 2.0, the associated total number of rSIR grid cells that we use in the calibration of our adapted empirical algorithm, and the coverage of detections and rSIR grid cells over each of the three sub-facies within the broader percolation facies are summarized in Table 1.

Table 1. *The total number of airborne ice penetrating radar survey detections (2010-2017), the associated total number of rSIR grid cells, and the coverage of detections and rSIR grid cells over perennial firn aquifer, ice slab, and perched firn aquifer areas.*

	Detections	Coverage (km²)	rSIR Grid Cells	Coverage (km²)
Perennial Firn Aquifers	~467,000	~178	~80	~8000
Ice Slabs	~505,000	~283	~2000	~20,000
Perched Firn Aquifers	~75,000	~23	~200	~2000

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2.~~3~~4 Empirical ~~A~~Algorithm

~~2.4.1 Greenland's Ice Facies~~

~~Greenland's ice facies (i.e., dry snow facies – percolation facies – wet snow facies – ablation facies) were first described in detail by Benson et al., (1960), and were shown to represent the GrIS's response to climate. Evolution of the boundaries of Greenland's ice facies are often used as an indicator of climate change. Early studies using field-based (Jezek et al., 1994; Zabel et al., 1995), airborne (Swift et al., 1985; Bindschadler et al., 1987; Rignot et al., 1993; Jezek et al., 1993), and satellite (Fahnestock et al., 1993; Long and Drinkwater, 1994; Parrington, 1998) synthetic aperture radars and radar scatterometers operating at frequencies between Ku-band (13 GHz) and P-band (0.4 GHz) have demonstrated the exceptional capabilities of microwave sensors for mapping Greenland's ice facies. Early airborne studies using C-band microwave radiometry (Swift et al., 1985), and more recent studies using L-band microwave radiometry (Jezek et al. 2018) have demonstrated similar capabilities. In this study, we extend these capabilities to include satellite L-band microwave radiometry. We delineate the boundaries of the percolation facies relative to the adjacent dry snow facies (i.e., where negligible seasonal surface melting occurs) and wet snow facies (i.e., where snow layers are fully water-saturated during the melting season and subsequently refreeze as superimposed ice overlying deeper glacial ice). And, we further identify sub-facies (i.e., perennial firn aquifer, ice slabs, and perched firn aquifers) within the broader percolation facies that are currently experiencing rapid expansion (McFerrin et al., 2019; Culberg et al., 2021) as Greenland's climate continues to warm (Hanna et al., 2013; Cullather et al., 2020) and seasonal surface melting increases in extent, intensity, and duration (Steffen et al., 2004; Tedesco e al., 2008; Tedesco et al., 2011; Nghiem et al., 2012; Tedesco et al., 2016; Tedesco and Fettweis, 2020; Tedesco and Fettweis, 2020). Higher frequency microwave sensors provide shallower penetration depths, and an increased sensitivity to snow grain size, layering, embedded ice structures (Long and Drinkwater, 1994; Drinkwater et al., 2001) and stored meltwater (Jezek et al., 1993; Miller, 2019) within the upper snow and firn layers of the percolation facies. Lower frequencies provide deeper penetration depths and a range of sensitivities to embedded ice structures (Jezek et al., 1993; Jezek et al., 2018) and stored meltwater (Miller et al., 2020) in the deeper firn layers.~~

2.~~3~~4.1~~2~~ Temporal L-band ~~S~~signatures over the ~~P~~percolation ~~F~~facies

~~Microwave brightness temperature~~ (T^B) expresses the satellite-observed magnitude of thermal emission and is influenced by the microwave instrument's observation geometry as well as the depth- and time-integrated dielectric and geophysical properties of the ice sheet (Ulaby et al., 2014). The most significant geophysical property influencing T^B is the volumetric fraction of meltwater within the snow and firn pore space (Mätzler and Hüppi, 1989). During the melting season, the upper snow and firn layers of the percolation facies are saturated with large volumetric fractions of meltwater that percolates vertically into

the deeper firn layers (Benson, 1960; Humphrey et al., 2012). Increases in the volumetric fraction of meltwater results in rapid relative increases in the imaginary part of the complex dielectric constant (Tiuiri et al., 1984)~~.,~~ ~~with corresponding~~This typically increases ~~in~~ T^B~~, and.~~ This ~~increase~~decreases ~~is attributed to a decrease in~~ volume scattering~~,~~ and penetration depth. The L-band penetration depth can rapidly decrease from ~tens to hundreds of meters~~,~~ to less than ~a meter, dependent on the local snow and firn conditions~~, and englacial firn hydrological features~~. During the freezing season, ~~s~~Surface and subsurface water-saturated snow and firn layers and embedded ice structures subsequently refreeze. ~~De~~During the freezing season, ~~de~~creases in the volumetric fraction of meltwater results in rapid relative decreases in the imaginary part of the complex dielectric constant. This~~, with corresponding~~ decreases ~~in~~ T^B~~., This increase is attributed to an increase in~~ and increases volume scattering~~,~~ and penetration depth. The L-band penetration depth increases back to ~tens to hundreds of meters on variable time scales.

We analyze melting and freezing seasons in temporal L-band signatures exhibited in T_V^B time series ~~(1 April 2015 - 31 March 2019)~~ over and near the AR- and MCoRDS-derived perennial firn aquifer, ~~and,~~ ice slab~~., and perched firn aquifer~~ detections projected on the NH EASE-Grid~~s~~ 2.0 (Fig. 4; Table 1). We project ice surface temperature ~~data~~ observations calculated using thermal infrared brightness temperature collected by the Moderate Resolution Imaging Spectroradiometer (MODIS) on the Terra and Aqua satellites (~~i.e.,~~ Hall et al., 2012) on the ~~coincident~~ NH EASE-Grids 2.0 at a 3.125 km rSIR grid cell spacing. We then derive melt onset and surface freeze-up dates ~~(2015-2019)~~ for each rSIR grid cell using the methodology described in Miller et al., (2020). We set a threshold of ice surface temperature >−1°C for meltwater detection (Nghiem et al., 2012), consistent with the ±1°C accuracy of the ice surface temperature observations~~data~~. For temperatures that are close to 0°C, ice surface temperatures are closely compatible with contemporaneous NOAA near-surface air temperature ~~data~~ observations (Shuman et al., 2014). Melt onset and surface freeze-up dates are overlaid on T_V^B time series to partition the melting and freezing seasons. Melt onset dates typically occur between ~April and July, and surface freeze-up dates typically occur between ~July and September. The melting season increases in duration moving downslope from the dry snow facies, and ranges from a single day in the highest elevations (>2500 m) of the percolation facies, to ~150 days in the ablation facies. Similarly, the ~~associated~~ freezing season decreases in duration moving downslope~~.~~ and ranges from between ~215 days and 365 days.

~~Over perennial firn aquifer areas (e.g., Figs. 1c; 2a; 4a), T_V^B is radiometrically warm during the melting season. Vertically percolating meltwater and gravity-driven meltwater drainage seasonally recharges perennial aquifers at depth (Fountain and Walder et al., 1998). Maximum values range from between ~200 K and 275 K during seasonal surface melting. Temporal L-band signatures exhibit increases on time scales of ~days to weeks following the melt onset date, and melting seasons range from between ~75 and 100 days. T_V^B remains radiometrically warm during the freezing season as a result of latent heat continuously released by the slow refreezing of the deeper firn layers that are saturated with large volumetric fractions of meltwater (Miller et al, 2020). Minimum values range from between ~180 K and 250~~

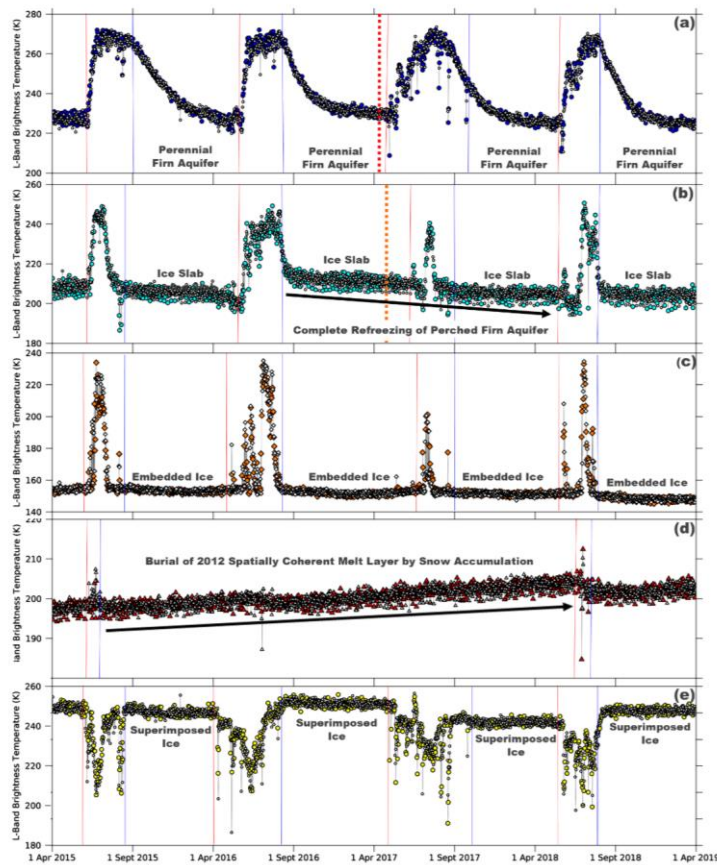
704   K following the surface freeze-up date. L-band emissions from the radiometrically warm upper snow and
705   firn layers decrease during the freezing season as embedded ice structures slowly refreeze at increased
706   depths below the ice sheet surface (Miller et al., 2020). Temporal L-band signatures exhibit exponential
707   decreases on time scales of ~months that approach and sometimes achieve relatively stable T_b values,

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Figure 4

Temporal L-band signatures that alternate morning (white symbols) and evening (colored symbols) orbital pass interval enhanced-resolution T_B^B generated using observations collected over the GrIS by the microwave radiometer on the SMAP satellite (Brodzik et al., 2019) over (a) perennial firn aquifer area (blue circles; Figs. 2a; 3a), (b) perched firn aquifer area (cyan circles; Figs. 2b; 3b), (c) percolation facies area (orange diamonds; Fig 2b), (d) high-elevation (~2500 m.a.s.l.) spatially coherent melt layer area (red triangles; Fig. 2a), and (e) superimposed ice area (yellow circles; Fig. 2b). Melt onset (red lines) and surface freeze-up (blue lines) dates are derived from thermal infrared T^B collected by MODIS on the Terra and Aqua satellites (Hall et al, 2012). AR radargram profile along perennial firn aquifer transect A-B (red dashed line; Figs. 1; 2a; 3a) that was collected on 22 April 2017, and ice slab transect C-D (orange dashed line; Figs. 1; 2b; 3b) that was collected on 5 May 2017.

Perennial Firn Aquifers and , Ice Slabs, Perched Firn Aquifers   27

Over perennial firn aquifer areas (e.g., Fig. 4a, SMAP Test Site A: 66.2115°N, 39.1795°W, 1625 m a.s.l), maximum T_V^B ($T_{V,max}^B$) values are radiometrically warm during the melting season. Vertically percolating meltwater and gravity-driven meltwater drainage seasonally recharges perennial firn aquifers at depth (Fountain and Walder et al., 1998). Minimum Maximum values range from between ~200 K and 275 K during seasonal surface melting. Temporal L-band signatures exhibit increases on time scales of ~days to weeks following the melt onset date, and melting seasons range from between ~75 and 100 days. T_V^B ($T_{V,min}^B$) values remainremains radiometrically warm during the freezing season as a result of latent heat continuously released by the slow refreezing of the deeper firn layers that are saturated with large volumetric fractions of meltwater (Miller et al, 2020). Minimum values range from between ~180 K and 250 K following the surface freeze-up date. L-band emissions from the radiometrically warm upper snow and firn layers decrease during the freezing season as embedded ice structures slowly refreeze at increased depths below the ice sheet surface (Miller et al., 2020). Temporal L-band signatures exhibit slow exponential decreases, and on time scales of ~months that approach, and sometimes achievve, relatively stable T_V^B valuesvalues. and freezing seasons range from between ~265-290 days. T_V^B can often decreases by more than ~50 K during the freezing seasonn (e.g., Fig. 4a), representingwhich represents the descent of the upper surface of stored meltwater by meters to ~tens of meters (Miège et al., 2016).

Over ice slab and perched firn aquifer areas (e.g., Fig. 4b, SMAP Test Site B: 66.8850°N, 42.7765°W, 1817 m a.s.le.g., Figs. 1c; 2b; 5b), $T_{V,max}^B$ T_V^B values are is typically radiometrically colder than over perennial firn aquifer areas during the melting season. The presence of dense low-permeability solid-ice layerss (e.g., Fig. 3b) reduces the snow and firn pore space available to store seasonal meltwater at depth. Meltwater may alternatively run-off ice slabs downslope towards the wet snow facies. $T_{V,min}^B$ Maximum values range from between ~170 K and 260 K during seasonal surface melting. Temporal L-band signatures exhibit increases on time scales of ~days to weeks following the melt onset date, and melting seasons range from between ~60 and 90 days. T_V^B values are is also typically radiometrically colder than over perennial firn aquifer areas during the freezing season as a result of the absence of meltwater stored at depth. (i.e. ice slab areas), or the presence of limited volumetric fractions of meltwater stored at depth in shallow water-saturated firn layers (i.e. perched firn aquifer areas). Minimum values range from between ~130 K and 240 K following the surface freeze-up date. Temporal L-band signatures exhibit exponential decreases that are slightly more rapid than over perennial firn aquifer areas, on time scales of ~weeks to months thaand t often achieve relatively stable T_V^B values., and freezing seasons range from between ~275-305 days. Exponentially decreasing temporal L-band signatures sometimes transition to linearly decreasing on time scales of ~years following the surface freeze-up date (e.g., between ~September 2016 and May 2018 in Fig. 4b). We infer this indicates the formation and subsequent refreezing of a shallow perched firn aquifer on top of a buried ice slab or other semi-impermeable layer. As compared to the large T_V^B decreases in percolation facies areas, T_V^B decreases over perched firn aquifer areas are as small as ~a few K annually, which represents the descent of the upper surface of stored meltwater by ~meters rather than by ~tens of meters.

Over other percolation facies areas (e.g., Fig. 4c, SMAP Test Site C: 66.9024°N, 44.7528°W, 2350 m a.s.l).~~,~~ where seasonal meltwater is fully refrozen and stored exclusively as embedded ice. $T^B_{V,max}$ ~~(e.g., Fig. 4c),~~ T^B_V values ~~are~~is typically radiometrically colder than over perennial firn aquifer and ~~,~~ice slab ~~, and perched firn aquifer~~ areas during the melting season. $T^B_{V,min}$ ~~Maximum values range from between ~150 K and 200 K during seasonal surface melting. Temporal L-band signatures exhibit increases on time scales of ~days to weeks following the melt onset date, and melting seasons range from between ~1 and 60 days.~~ T^B_V values ~~are~~is also typically radiometrically cold during the freezing season. ~~Minimum values range from between ~130 K and 180 K following the surface freeze-up date.~~ Temporal L-band signatures exhibit rapid exponential decreases, ~~on time scales of ~days to weeks~~ and achieve ~~relatively~~ stable T^B_V values~~, and freezing seasons range from between ~305-364 days~~. However, over the highest elevations (> ~2500 m ~~.~~a.s.l.) of the percolation facies approaching the dry snow line, where seasonal surface melting and the formation of embedded ice structures is limited, $T^B_{V,min}$ ~~T^B_V~~values remain ~~remains~~ radiometrically warm during the freezing season. ~~. Minimum values range from between ~180 K and 220 K following the surface freeze-up date.~~ We infer T^B_V decreases~~decreases~~, often ~~sometimes~~step-responses exceeding ~10 K, ~~that follow the surface freeze-up date (e.g., between April 2018 and September 2018 in Fig. 4c)~~ are a~~a~~ result of an increase in volume scattering from newly formed embedded ice structures within a spatially coherent melt layer. ~~We also infer that temporal~~Temporal L-band signatures that increase several K on time scales of ~years ~~(e.g., between ~April 2015 and April 2018 in Fig. 4c)~~ indicate the burial of spatially coherent melt layers formed following the 2010, 2012, 2015, and 2018 melting seasons by snow accumulation.~~formed following the 2010, 2012, 2015, and 2018 melting seasons by snow accumulation.~~

Exponentially decreasing temporal L-band signatures transition smoothly between perennial firn aquifer~~,~~ ice slab~~, perched firn aquifer,~~, and ~~and~~ other percolation facies areas – there are no distinct temporal L-band signatures that delineate boundaries between these sub-facies. Boundary transitions between ~~other facies~~the dry snow facies and the wet snow facies, however, are delineated ~~both~~above and below the percolation facies. Over the dry snow facies (e.g., Fig. 4d, SMAP Test Site D: 66.3649°N, 43.2115°W, 2497 m a.s.l)~~e.g., Fig. 4d)~~, $T^B_{V,max}$ ~~T^B_V~~ and $T^B_{V,min}$ values ~~are~~is radiometrically warm during the melting and freezing seasons. ~~Values range from between ~200 K and 240 K. While T^B_V is known to be relatively stable in the dry snow facies,~~ ~~T~~temporal L-band signatures that increase on time scales of ~years are observed throughout ~~this region~~the dry snow facies at elevations as high as Summit Station (~3200 m ~~.~~a.s.l) and ~~, similar to those observed in the highest elevations (> ~2500 m.a.s.l.) of the percolation facies.~~ indicate the burial of the spatially coherent melt layer formed following the 2012 melting season (Nghiem et al., 2012) by snow accumulation (Culberg et al., 2021). ~~We infer increasing temporal L-band signatures indicate the burial of the spatially coherent melt layer formed following the anomalous 2012 melting season (Nghiem et al., 2012) by snow accumulation (Culberg et al., 2021).~~ Over the wet snow facies (e.g., Fig. 4e, SMAP Test Site E: 67.3454°N, 48.4789°W, 1469 m a.s.l)~~e.g., Fig. 4e)~~, where seasonal meltwater is fully refrozen and stored as superimposed ice, $T^B_{V,max}$ ~~T^B_V~~ values are ~~is~~radiometrically warm during the melting

794 seasonseason. Maximum values range from between ~230 K and 250 K during seasonal surface melting.
795 As compared to the percolation facies, where temporal L-band signatures exhibit rapid increases following
796 melt onset, temporal L-band signatures reverse and exhibit rapid decreases on time scales of ~days to
797 weeks, and melting seasons that range between ~90-120 days. We infer theseThese reversals are the a
798 result of high reflectivity and attenuation at the fully water-saturated snow layer and/or at the wet, rough
799 superimposed ice-air interface. Meltwater runs-off superimposed ice downslope towards the ablation facies
800 in the wet snowablation facies. $T_{V,min}^B$ T_V^B values remains radiometrically warm during the freezing season.
801 Minimum values range from between ~230 K and 250 K following seasonal surface melting. Temporal L-
802 band signatures exhibit rapid increases, on time scales of ~days that achieveand achieve relatively stable
803 T_V^B values., and freezing seasons range from between ~245 and 275 days.
804 The MODIS-derived total number of days in the melting and freezing seasons estimated from melt
805 onset and surface freeze-up dates, the SMAP-derived maximum and minimum vertically-polarized L-band
806 brightness temperature, and the time scales of exponential decrease following the surface freeze-up date
807 estimated for each T_V^B time series for rSIR grid cells over perennial firn aquifer, ice slab, perched firn aquifer,
808 and other percolation facies areas as well as for the dry snow facies, and the wet snow facies are
809 summarized in Table 2.
810
811
812
813
814
815
816 **Table 2.** *The MODIS-derived total number of days in the melting and freezing seasons (2015-2019), the*
817 *SMAP-derived maximum vertically-polarized L-band brightness temperature ($T_{V,max}^B$), the minimum*
818 *vertically-polarized L-band brightness temperature ($T_{V,min}^B$), and the time scale scales of exponential*
819 *decrease following the surface freeze-up date (1 April 2015 - 31 March 2019) for perennial firn aquifer, ice*
820 *slab, perched firn aquifer, and other percolation facies areas as well as for the dry snow facies and the wet*
821 *snow facies.*

	Melting Season (days)	Freezing Season (days)	$T_{V,max}^B$ (K)	$T_{V,min}^B$ (K)	Exponential Decrease (time scale)
Perennial Firn Aquifers	~75 - 100	~265 - 290	~200 - 275	~180 - 250	~weeks - months
Ice Slabs / Perched Firn Aquifers	~60 -90	~275 - 305	~170 - 260	~130 – 240	~days - Weeks
Percolation Facies	~1 - 60	~305 - 364	~150 - 200	~130 – 220	~days

| Dry Snow Facies | - | 365 | ~200 - 240 | ~200 - 240 | - |
| Wet Snow Facies | ~90 - 120 | ~245 - 275 | ~230 - 250 | ~230 – 250 | - |

2.3 Two-Layer L-band Brightness Temperature Model

Based on our analysis of $T_{V,max}^B$ and $T_{V,min}^B$ values in temporal L-band signatures over the percolation facies (Section 2.3.1), we derive a 'firn saturation' parameter using the simple two-layer L-band brightness temperature model (Ashcraft and Long, 2006; Miller et al., 2021, in press). The 'firn saturation' parameter is similar to the 'melt intensity' parameter derived in Hicks and Long (2011) that uses enhanced resolution vertically-polarized Ku-band radar backscatter imagery (2003) collected by the SeaWinds radar scatterometer that was flown in tandem on NASA's Quick SCATterometer (QuikSCAT) satellite (Tsai et al., 2000) and JAXA's Advanced Earth Observing Satellite 2 (ADEOS-II) (Freilich et al., 1994). We use the firn saturation parameter to estimate the maximum seasonal volumetric fraction of meltwater within the saturated upper snow and firn layers of the percolation facies using $T_{V,max}^B$ and $T_{V,min}^B$ values extracted from T_V^B time series (1 April 2015 - 31 March 2019). We calculate the firn saturation parameter for each rSIR grid cell within the ice sheet-masked extent of the GrIS as part of our adapted empirical algorithm (see Section 2.4.5).

We assume a base layer underlying a water-saturated firn layer with a given depth and volumetric fraction of meltwater. Each of the layers is homogenous. The ice sheet is discretely layered to calculate T_V^B at an oblique incidence angle (Eq. 1). Emission from the base layer is a function of both the macroscopic roughness and the dielectric properties of the layer. It occurs in conjunction with volume scattering at depth, and is locally dependent on embedded ice structures, spatially coherent melt layers, ice slabs, and perennial firn aquifers. Reflectivity at depth (i.e., at the base layer-water-saturated firn layer interface), and at the ice sheet surface (i.e., at the water-saturated firn layer-air interface) is neglected. The contribution from each layer is individually calculated.

The two-layer L-band brightness temperature model is represented analytically by

$$T_{V,max}^B = T\left(1 - e^{-\kappa_e d sec\theta}\right) + T_{V,min}^B e^{-\kappa_e d sec\theta} , \qquad \text{(Eq. 1)}$$

where $T_{V,max}^B$ is the maximum vertically-polarized L-band brightness temperature at the ice sheet surface, and represents emission from the maximum seasonal volumetric fraction of meltwater stored within the water-saturated firn layer. $T_{V,min}^B$ is the minimum vertically-polarized L-band brightness temperature emitted from the base layer. T is the physical temperature of the water-saturated firn layer, θ is the transmission angle, κ_e is the extinction coefficient, and d is depth.

855

856   We invert Eq. 1 and solve for the firn saturation parameter (ξ)

857

858    $\xi = ln\left(\frac{T_{V,max}^{B}-T}{T_{V,min}^{B}-T}\right)cos\theta$,   ———   (Eq. 2)

859

860 where $\xi = \kappa_e d$. The maximum vertically-polarized L-band brightness temperature asymptotically approaches

861 the physical temperature of the water-saturated firn layer as the extinction coefficient and the depth of the

862 water-saturated firn layer increases. For simplicity, we follow Jezek et al., (2015) and define the extinction

863 coefficient as the sum of the Raleigh scattering coefficient (κ_s) and the absorption coefficient (κ_a). This

864 assumes scattering from snow grains, which are small (millimeter scale) relative to the L-band wavelength

865 (21 cm), and neglects Mie scattering from large (centimeter scale) embedded ice structures. However, The

866 extinction coefficient is defined as the sum of the Raleigh scattering coefficient (κ_s) and the absorption

867 coefficient (κ_a). fFor water-saturated firn, absorption dominates over scattering, and increases in the

868 extinction coefficient are controlled by the volumetric fraction of meltwater (m_v).

869   We assume that thicker water-saturated firn layers with larger volumetric fractions of meltwater

870 generate higher firn saturation parameter values. However, the thickness of the water-saturated firn layer

871 is limited by the L-band penetration depth. Theoretical L-band penetration depths calculated for a water-

872 saturated firn layer range from between ~10 m for small volumetric fractions of meltwater (m_v<1%), and ~1

873 cm for large volumetric fractions of meltwater (m_v=20%) (Fig. 5). Large volumetric fractions of meltwater

874 results in high reflectivity and attenuation at at the water-saturated firn layer-air interface the interface

875 between water-saturated firn layers and the overlying refrozen firn layers, and between glacial ice or a semi-

876 impermeable layer and the overlying water-saturated firn layers, and a radiometrically cold firn layer (e.g.,

877 Fig 5e).

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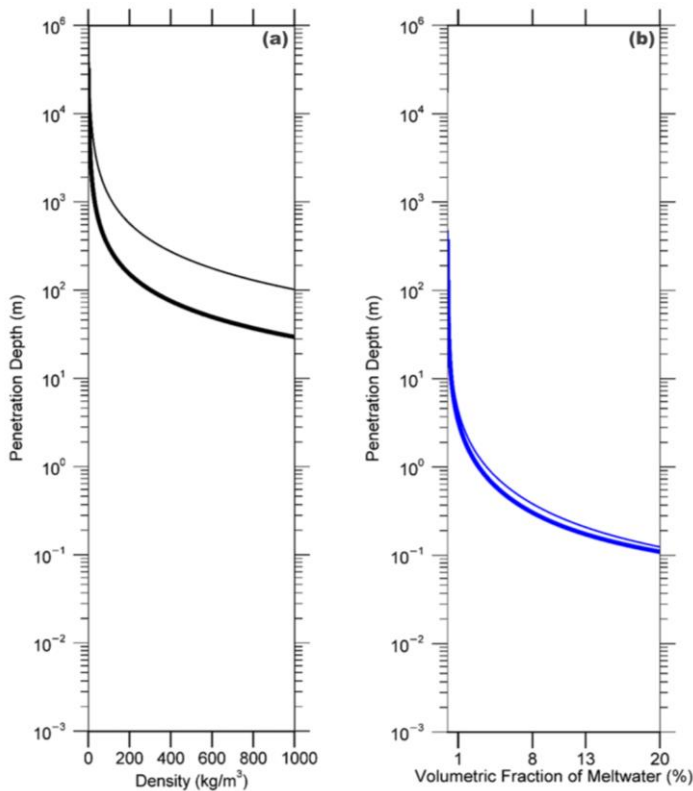
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880 ## ~~Figure 5~~

881 *~~Theoretical L-band penetration depths for of uniform layer of (a) refrozen, and (b) water-saturated firn.~~*

882 *~~Penetration depths $\left(\frac{1}{\kappa_s + \kappa_a}\right)$ are calculated as a function of the Raleigh scattering coefficient (κ_s; Eq. 8) and~~*

883 *~~the absorption coefficient (κ_a; Eq. 10), which are functions of the dielectric and geophysical properties of~~*

884 *~~the GrIS. The complex dielectric constant is calculated using the empirically derived models described in~~*

885 *~~Tiuri et al., (1984). Refrozen firn penetration depths are calculated as a function of firn density (ρ_{firn}), and~~*

886 *~~the curves are plotted for snow grain radii (r) set to r =0.5 mm (upper curve), and r =4 mm (lower curve).~~*

887 *~~Water-saturated firn penetration depths are calculated as a function of the volumetric fraction of meltwater~~*

888 *~~(m_v), and the curves are plotted for firn density set to ρ_{firn}=400 kg/m³ (upper curve), and ρ_{firn}=917 kg/m³~~*

889 *~~(lower curve). Given the complexity of modeling embedded ice structures, they are excluded from the~~*

890 *~~penetration depth calculation. Increases in the volumetric fraction of embedded ice in the firn will result in~~*

891 *~~an increase in volume scattering, which will decrease and compress the distance between the penetration~~*

892 *~~depth curves for both refrozen and water-saturated firn.~~*

2.34.34   Continuous Logistic Model

We adapt our previously developed empirical algorithm to map the extent of Greenland's perennial firn aquifers (Miller et al., 2020) to also map the extent of ice slab and perched firn aquifer areas. The empirical algorithm is derived from the continuous logistic model, which is based on a differential equation that models the decrease in physical systems as a function of time using a set of sigmoidal curves. These curves begin at a maximum value with an initial interval of decrease that is approximately exponential. Then, as the function approaches its minimum value, the decrease slows to approximately linear. Finally, as the function asymptotically reaches its minimum value, the decrease exponentially tails off and achieves stable values. We use the continuous logistic model to parametrize the refreezing rate within the water-saturated upper snow and firn layers of the percolation facies using T_V^B time series (1 April 2015 - 31 March 2019) that are partitioned using $T_{V,max}^B$ and $T_{V,min}^B$ values. We calculate the refreezing rate for each rSIR grid cell within the percolation facies extent as part of our adapted empirical algorithm (see Section 2.34.45).

The continuous logistic model is described by a differential equation known as the logistic equation

$$\frac{dx}{dt} = \zeta x(1-x) \qquad \text{(Eq. 3)}$$

that has the solution

$$x(t) = \frac{1}{1+\left(\frac{1}{x_o}-1\right)e^{-\zeta t}}, \qquad \text{(Eq. 4)}$$

where x_o is the function's initial value, ζ is the function's exponential rate of decrease, and t is time. The function $x(t)$ is also known as the sigmoid function. We use the sigmoid function to model the exponentially decreasing temporal L-band signatures observed over the percolation facies as a set of decreasing sigmoidal curves.

We first normalize T_V^B time series for each rSIR grid cell

$$T_{V,N}^B(t) = \frac{T_V^B(t)-T_{V,min}^B}{T_{V,max}^B-T_{V,min}^B}, \qquad \text{(Eq. 5)}$$

where $T_{V,min}^B$ is the minimum vertically-polarized L-band brightness temperature, and $T_{V,max}^B$ is the maximum vertically-polarized L-band brightness temperature. We then apply the sigmoid fit

$$T_{V,N}^B\left(t \in [t_{max}, t_{min}]\right) = \frac{1}{1+\left(\frac{1}{T_{V,N}^B(t_{max})}-1\right)e^{-\zeta t}}. \qquad \text{(Eq. 6)}$$

$T_{V,N}^B\left(t \in [t_{max}, t_{min}]\right)$ is the normalized vertically-polarized L-band brightness temperature on the time interval $t \in [t_{max}, t_{min}]$, where t_{max} is the time the function achieves a maximum value, and t_{min} is the time the function achieves a minimum value. The initial normalized vertically-polarized L-band brightness temperature ($T_{V,N}^B(t_{max})$) is the function's maximum value. The final normalized vertically-polarized L-band brightness temperature ($T_{V,N}^B(t_{min})$) is the function's minimum value. The function's exponential rate of decrease represents the refreezing rate parameter (ζ). An example set of simulated sigmoidal curves is shown in Fig. 6.

2.34.45   SMAP-Dderived Pperennial Ffirn Aaquifer and, Iice Sslab , and perched firn aquifer Mmappings

Our adapted empirical algorithm uses ice sheet-masked SMAP enhanced-resolution T_V^B imagery over the GrIS that alternates morning and evening orbital pass observations annually, beginning and ending just prior to melt onset. Our adapted empirical algorithm is implemented in two steps: (1) mapping the extent of the percolation facies using the firn saturation parameter derived from the simple two-layer L-band geophysical brightness temperature model ((see Section 2.34.23), and (2) mapping the extent of perennial firn aquifer and , ice slab, and perched firn aquifer areas over the percolation facies using the continuous logistic model (see Section 2.34.34) we calibrate using airborne ice-penetrating radar detectionssurveys (Section 2.2). projected on three separate NH EASE-Grids 2.0 (see Section 2.2).

Using Eq. 2, we first set a threshold for the firn saturation parameter (ξ_T) defined by the relationship

$$\xi_T = (\kappa_s + \kappa_a)d \leq \xi . \qquad \text{(Eq. 7)}$$

We calculate the Raleigh scattering coefficient (κ_s) in Eq. 7 using

$$\kappa_s = N_d \frac{8}{3} k_o^4 r^6 \left|\frac{\varepsilon_r - 1}{\varepsilon_r + 2}\right|^2 , \qquad \text{(Eq. 8)}$$

where N_d is the particle density, k_o is the wave number of the background medium of air, r is the snow grain radius set to r=2 mm, and ε_r is the complex dielectric constant. The particle density is defined by

$$N_d = \frac{\rho_{firn}}{\rho_{ice}} \frac{1}{\frac{4}{3}\pi r^3} , \qquad \text{(Eq. 9)}$$

958 where ρ_{firn} is firn density, ~~which we~~ set to ρ_{firn}=400 kg/m³, and ρ_{ice} is ice density, ~~which we~~ set to ρ_{ice}=917

959 kg/m³. Our grain radius and firn density estimates are consistent with measurements within the upper snow

960 and firn layers of the percolation facies of south eastern Greenland at the Helheim Glacier field site (Fig.

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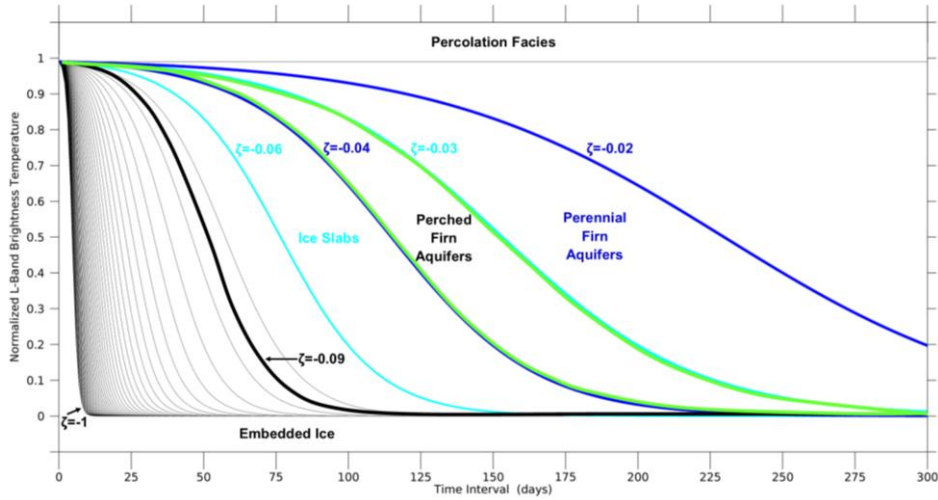
962

963 ~~**Figure 6**~~

964 ~~*Example set of simulated sigmoidal curves that represent our model of the exponentially decreasing*~~

965 ~~*temporal L-band signatures predicted over the percolation facies. The initial normalized vertically-polarized*~~

966 ~~*L-band brightness temperature was fixed at a value of $T^B_{V,fn}(t_{max}) = 0.99$, and the time interval was set to a*~~

967 ~~*value of $t \in [t_{max}, t_{min}] = 300$ observations. The refreezing rate parameter was set to values between $\zeta =$*~~

968 ~~*[-1, 0] incremented by steps of 0.02. The blue lines correspond to the interval $\zeta \in [-0.04, -0.02]$ and*~~

969 ~~*produce curves similar to those observed over perennial firn aquifer areas. The cyan lines correspond to*~~

970 ~~*the interval $\zeta \in [-0.06, -0.03]$ and produce curves similar to those observed over ice slab areas. The green*~~

971 ~~*lines correspond to the interval $\zeta \in [-0.04, -0.03]$ and produce curves similar to those observed over*~~

972 ~~*perched firn aquifer areas. The black line is the observed lower bound ($\zeta = -0.09$) of the refreezing rate*~~

973 ~~*parameter of partitioned T^B_V time series (1 April 2015 - 31 March 2019) iteratively fit to the sigmoid function*~~

974 ~~*(see Section 3).*~~

975

2a.; blue circle; Fig. 4a), where in situ perennial firn aquifer measurements have recently been collected (Miller et al., 2017).

We calculate the absorption coefficient (κ_a) in Eq. 7 using

$$\kappa_a = -2k_o \Im\{\sqrt{\varepsilon_r}\} \,, \qquad\qquad\qquad\qquad \text{(Eq. 10)}$$

where $\Im\{\}$ represents the imaginary part. We calculate the complex dielectric constant of the water-saturated firn layer in Eq. 8 and Eq. 10 using the empirically derived models described in Tiuri et al., (1984). We set the volumetric fraction of meltwater to m_v=1%. We set the depth of the water-saturated firn layer in Eq. 7 to d=1 m. These values are consistent with typical lower frequency (e.g., 37 GHz, 13.4 GHz, 19 GHz) passive (e.g., Mote, et al. 1995; Abdalati and Steffen, 1997; Ashcraft and Long, 2006) and active (e.g., Hicks and Long, 2011) microwave algorithms used to detect seasonal surface melting over the GrIS. Using the results of Eq. 7, 8, 9, and 10, we calculate the firn saturation parameter threshold at ξ_T=0.1.

The first step in our adapted empirical algorithm is to map the extent of the percolation facies. For each rSIR grid cell within the ice sheet-masked extent of the GrIS, we smooth the corresponding T_V^B time series s (1 April 2015 - 31 March 2019) using a 14-observation (1 week) moving window. We extract the minimum vertically-polarized L-band brightness temperature ($T_{V,min}^B$), and the maximum vertically-polarized L-band brightness temperature ($T_{V,max}^B$). We set the physical temperature of the water-saturated firn layer to T=273.15 K, and the transmission angle to θ=40°. We then calculate the firn saturation parameter (ξ) using Eq. 2. If the calculated firn saturation parameter exceeds the firn saturation parameter threshold, the rSIR grid cell is converted to a binary parameter to map the total extent of the percolation facies.

We note that smoothing T_V^B time series will mask brief low-intensity seasonal surface melting that occurs in the high-elevation (> ~2500 m) percolation facies, where seasonal meltwater is rapidly refrozen within the colder snow and firn layers (e.g., Fig. 4d). Thus, the calculated firn saturated parameter will not exceed the firn saturation parameter threshold, and these rSIR grid cells will be excluded from the algorithm. The exclusion of rSIR grid cells in the high-elevation percolation facies is not expected to have a significant impact on our results as our algorithm targets rSIR grid cells in areas that experience intense seasonal surface melting. The exclusion of rSIR grid cells will, however, slightly underestimate the mapped percolation facies extent.

The second step in our adapted empirical algorithm is to map the extent of perennial firn aquifer and, ice slab, and perched firn aquifer areas over the percolation facies. For each rSIR grid cell within the mapped percolation facies extent, we normalize the corresponding T_V^B time series time series (1 April 2015 - 31 March 2019) using Eq. 5 ($T_{V,N}^B(t)$) using Eq. 5. We then extract the initial normalized vertically-polarized L-band brightness temperature, ($T_{V,N}^B(t_{max})$) and the final normalized vertically-polarized L-band brightness temperature ($T_{V,N}^B(t_{min})$), and partition $T_{V,N}^B(t)$ on the time interval $t \in [t_{max}, t_{min}]$. We smooth $T_{V,N}^B(t \in [t_{max}, t_{min}])$ using a 56-observation (4 week) moving window. The sigmoid fit is then iteratively

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1012 applied using Eq. 6. Smoothing reduces the chi-squared error statistic when fitting $T_{V,N}^B(t \in [t_{max}, t_{min}])$ to
1013 the sigmoid function. We fix the initial normalized vertically-polarized L-band brightness temperature at
1014 $T_{V,N}^B(t_{max})$=0.99, which provides a uniform parameter space in which the refreezing rate parameter (ζ) can
1015 be analyzed. Variability in $T_{V,N}^B(t_{max})$ is controlled by the volumetric fraction of meltwater within the upper
1016 snow and firn layers of the percolation facies, and is accounted for in the firn saturation parameter (ξ), which
1017 is analyzed separately. $T_{V,N}^B(t \in [t_{max}, t_{min}])$ iteratively fit to the sigmoid function converge quickly (i.e.,
1018 algorithm iterations I ∈ [5, 15]), and observations are a good fit (i.e., chi squared error statistic is $\chi2 \in$ [0,
1019 0.1]).

1020   Using the SMAP-derived $T_{V,N}^B(t_{max})$ and $T_{V,N}^B(t_{min})$, rather than the MODIS-derived initial
1021 normalized vertically-polarized L-band brightness temperature at the surface freeze-up date ($T_{V,N}^B(t_{sfu})$),
1022 and final normalized vertically-polarized L-band brightness temperature at the melt onset date ($T_{V,N}^B(t_{mo})$)
1023 that were used in the empirical algorithm described in Miller et al., 2020 ~~(e.g., Fig. 4),~~ has several
1024 advantages. They key advantage of this approach is that maps can be generated using T^B imagery
1025 collected from a single satellite, which simplifies ~~the~~ our adapted empirical algorithm. Another advantage is
1026 that unlike T^B collected at shorter-wavelength thermal infrared frequencies (e.g., MODIS), T^B collected at
1027 longer wavelength microwave frequencies (e.g., SMAP) is not sensitive to clouds, which eliminates
1028 observational gaps and cloud contamination, and provides more accurate time series partitioning and more
1029 robust curve fitting. ~~The mapped extent of Greenland's perennial firn aquifers generated by our adapted~~
1030 ~~empirical algorithm and by our empirical algorithm (Miller et al., 2020) are consistent (see Section 3).~~

1031   We calibrate our adapted empirical algorithm using the AR- and MCoRDS-derived perennial firn
1032 aquifer~~r (2010-2017)~~ and~~,~~ ice slab ~~(2010-2014), and perched firn aquifer (2010-2017)~~ detections projected
1033 ~~separately~~ on ~~the three~~ NH EASE-Grid~~s~~ 2.0. For each rSIR grid cell with at least one detection, we extract
1034 the ~~corresponding~~ correlated maximum vertically-polarized L-band brightness temperature ($T_{V,max}^B$), the
1035 minimum vertically-polarized L-band brightness temperature ($T_{V,min}^B$), the firn saturation parameter (ξ), and
1036 the refreezing rate parameter (ζ)~~, and for~~For each of the extracted ~~SMAP-derived~~ calibration parameters,
1037 we calculate the standard deviation (σ). ~~Thresholds~~Similar to Miller et al., 2020, thresholds of ±2σ are set
1038 ~~for each of the extracted SMAP-derived calibration parameters~~ in an attempt to eliminate peripheral rSIR
1039 grid cells near the ice sheet edge and near ~~the~~ the upper and lower boundaries of each sub-faci~~ee~~, where
1040 ~~, where~~ L-band emission~~s~~ can be influenced by morphological features, ~~,~~ such as crevasses, superimposed
1041 and glacial ice, and spatially integrated with emission~~s~~ from rock, land, the ocean, and adjacent percolation
1042 facies and wet snow facies areas.~~.~~ The ~~SMAP-derived~~ calibration parameter ~~threshold~~ intervals ~~extracted~~
1043 ~~from T_V^B time series that we use to map perennial firn aquifer, ice slab, and perched firn aquifer areas~~ are
1044 given in Table 2~~3~~. We apply the calibration to each rSIR grid cell within the percolation facies extent. If the
1045 extracted ~~SMAP-derived~~ calibration parameters are within the ~~threshold~~ intervals, the rSIR grid cell is
1046 converted to a binary parameter to map the total extent of each of these sub-facies.

1047

1048

1049 Table 3. *SMAP-derived calibration parameter threshold intervals (1 April 2015 - 31 March 2019) used for*
1050 *mapping perennial firn aquifer, ice slab, and perched firn aquifer areas.*

	ξ	$T_{V,max}^{B}$ (K)	$T_{V,min}^{B}$ (K)	ζ
Perennial Firn Aquifers	0.2 – 4	200 – 275	180 – 250	-0.04 – 0.02
0.1 – 2	170 – 260	130 – 240	-0.03 – 0.06	
Perched Firn Aquifers	0.2 – 1.2	200 – 260	180 – 240	-0.03 – 0.04

1051

1052 Iteratively applying the sigmoid fit to $T_{V,N}^{B}(t \in [t_{max}, t_{min}])$ over perched firn aquifer areas is a
1053 source of uncertainty in our adapted empirical algorithm. While the continuous logistic model is reasonable
1054 for the majority of exponentially decreasing temporal L-band signatures over the percolation facies, it is not
1055 optimal for exponentially decreasing temporal L-band signatures that transition to linearly decreasing on
1056 time scales of ~years following the surface freeze-up date. Especially multi-year linearly decreasing
1057 temporal L-band signatures over areas where perching occurs following intense seasonal surface melting
1058 and shallow water-saturated firn layers persist throughout the following freezing season as well as
1059 throughout weaker seasonal surface melting the following melting season (e.g., between ~September 2016
1060 and May 2018 in Fig. 4b). Although $T_{V,N}^{B}(t \in [t_{max}, t_{min}])$ over perched firn aquifer areas iteratively fit to
1061 the sigmoid function converge quickly (i.e., algorithm iterations $I \in [8, 15]$), and observations appear to be
1062 a good fit (i.e., chi squared error statistic is $\chi^2 \in [0.06, 0.1]$), simulated sigmoidal curves often
1063 asymptotically approach $T_{V,min}^{B}$ too quickly, which underestimates the refreezing rate parameter (ζ) to
1064 values outside the SMAP-derived calibration parameter threshold intervals. Perched firn aquifer areas may
1065 alternatively be mapped as ice slab areas or percolation facies areas, which will underestimate or
1066 overestimate the mapped extent of each of these sub-facies.
1067 Miller et al., 2020 cited significant uncertainty in the SMAP-derived perennial firn aquifer extent as
1068 a result of the lack of a distinct temporal L-band signature delineating the boundary between perennial firn
1069 aquifer areas and adjacent percolation facies areas. In this study, similar uncertainty exists in the SMAP-
1070 derived perennial firn aquifer and, ice slab, and perched firn aquifer extents. This uncertainty could, at least
1071 in part, be a result of the rSIR algorithm. An rSIR grid cell corresponds to the weighted average of T^B over
1072 SMAP's antenna footprint (Long et al., 2020). The weighting is the grid cell's spatial response function
1073 (SRF), which is approximately ~18 km (i.e., the effective resolution) in diameter. The SRF is centered on
1074 the rSIR grid cell. Since the effective resolution (i.e., the size of the 3 dB3 dB contour of the SRF) is less
1075 than the rSIR grid cell spacing, rSIR grid cell SRF's overlap and the grid cells T^B values are not statistically
1076 independent. This uncertainty, however, could also have a geophysical basis, as it is unlikely that the
1077 boundaries between sub-facies (perennial firn aquifers, ice slabs, and perched firn aquifers) as well as
1078 between facies (percolation facies, dry snow facies, wet snow facies) are distinct. The thickness of the

Perennial Firn Aquifers and , Ice Slabs, Perched Firn Aquifers 39

water-saturated firn layer or ice slab may thin and taper-off at the periphery, and sub-facies and facies may become spatially scattered and merge together. ~~Over SMAP's ~18 km footprint, spatially integrated L-band emissions may also result in a smooth transition between temporal L-band signatures.~~

The limited extent (AR, 15 m x 20 m; MCoRDS, 14 m x 40 m) of the airborne ice-penetrating radar ~~detections~~ surveys as compared to the rSIR grid cell extent (3.125 km ~~x 3.125 km~~) and the effective resolution ~~(~18 km)~~ of the SMAP enhanced-resolution T_{V}^{B} imagery is also cited in Miller et al., 2020 as a source of uncertainty in the empirical algorithm. In this study, similar uncertainty exists in our adapted empirical algorithm. The total rSIR grid cell extent with airborne ice-penetrating radar survey ~~radargram~~ coverage is less than 2%~~,~~. Thus,~~which means that~~ ~98% of the total rSIR grid cell extent ~~with radargram coverage,~~ from which the SMAP-derived calibration parameter ~~threshold~~ intervals are extracted~~,~~ is unknown. Calculating the total rSIR grid cell extent where detections are absent along OIB flight lines and statistically integrating this calculation into the multi-year calibration technique may help reduce the uncertainty, particularly the significant uncertainty in the interannual variability, in extent, which we have yet to resolve. A sensitivity analysis suggests that even small changes in ~~any of~~ the SMAP-derived calibration parameter ~~threshold~~ intervals (i.e., several K for $T_{V,min}^{B}$, and $T_{V,max}^{B}$, several tenths of a percentage point for ξ, and several hundredths of a percentage point for ζ) can result in variability in the mapped extents of hundreds of square kilometers, and boundary transitions between perennial firn aquifer, and, ice slab, ~~and perched firn aquifer~~ areas. Thus, the mapped extent of each of these sub-facies ~~of the broader percolation facies~~ should simply be considered an initial result demonstrating the potential of our adapted empirical algorithm for future work.

~~Similar to ice slabs, the formation of spatially coherent melt layers reduces the pore space within the upper snow and firn layers and may also facilitate lateral meltwater flow with minimum vertical percolation into the deeper firn layers, thus enhancing meltwater run-off from significantly higher elevations downslope towards the periphery on accelerated time scales. The formation of spatially coherent melt layers overlying deeper perennial firn aquifers (e.g., Fig. 3a) will limit or terminate gravity-driven meltwater drainage and seasonal recharging (Fountain and Walder, 1998), which may eventually completely refreeze stored meltwater into decimeters thick solid-ice layers overlying deeper glacial ice.~~

3. Results and Discussion

The SMAP-derived maximum vertically-polarized L-band brightness temperature values generated by our adapted empirical algorithm range from between $T_{V,max}^{B}$=150 K and 275 K, and the minimum vertically-

polarized L-band brightness temperature values range from between $T^B_{V,min}$=130 K and 250 K. These values are consistent with the range of $T^B_{V,max}$ and $T^B_{V,min}$ values given in the temporal L-band signature analysis (Section 2.4.2; Table 12). Firn saturation parameter values range from between ξ=0.1 and 4.0. Refreezing rate parameter values range from between ζ=-0.09 and -0.01. The observed lower bound (ζ=-0.09) of the refreezing rate parameter observed over the percolation facies is significantly higher than the predicted lower bound (ζ=-1) in our example set of simulated sigmoidal curves (black line, Fig. 6).

The SMAP-derived perennial firn aquifer, (blue shading), ice slab (cyan shading), perched firn aquifer (green shading), and percolation facies (purple shading) extents (2015-2019) are generated by our adapted empirical algorithm are shown in Figs. 7a-9a, and are summarized in Table 4. The percolation facies extent (~5.8 x 10^5 km^2) generated by our adapted empirical algorithm is mapped at elevations between ~500 m a.s.l. and 3500 m a.s.l., and extends over ~32 % of the GrIS extent (~1.8 x 10^6 km^2). The perennial firn aquifer extent (64,000 km^2) is mapped at elevations between ~600 m a.s.l and 2600 m a.s.l., and extends over ~11% of the percolation facies extent, and ~4% of the GrIS extent. Predominately high $T^B_{V,max}$, $T^B_{V,min}$,
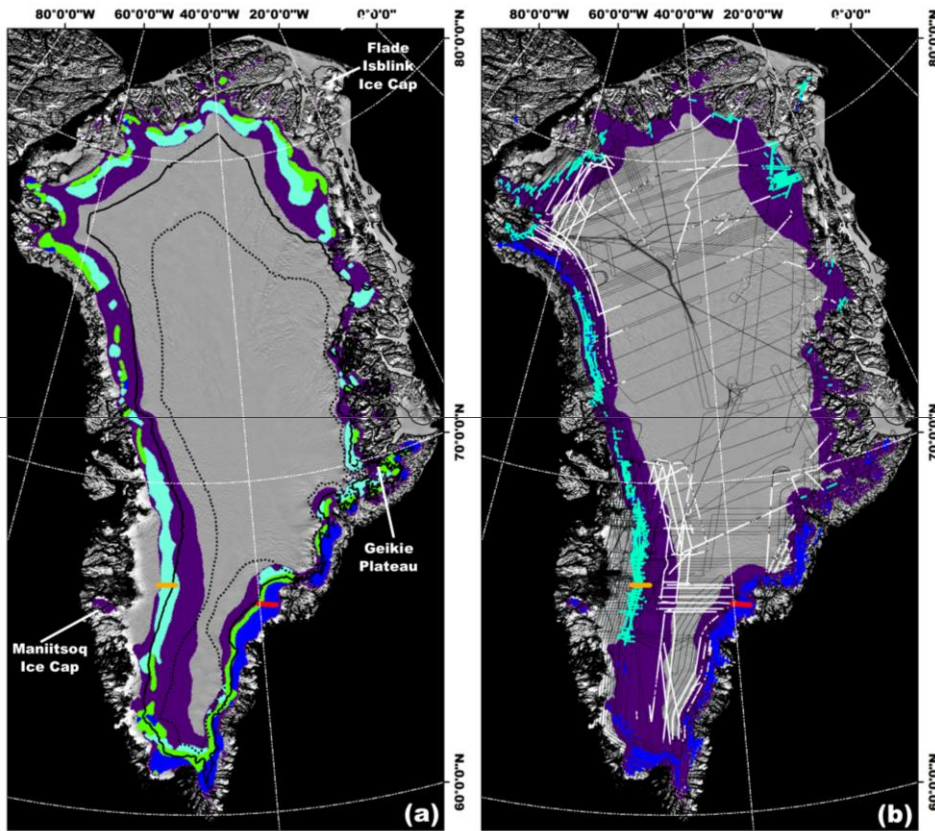
Figure 7

(a) The SMAP-derived perennial firn aquifer (blue shading), ice slab (cyan shading), perched firn aquifer (green shading), and percolation facies (purple shading) extents (2015-2019) generated by the adapted empirical algorithm overlaid on the 2015 MODIS Mosaic of Greenland image map (Haran et al., 2018). The black line is the 2000 m.a.s.l. contour, and the black dotted line is the 2500 m.a.s.l. contour (Howat et al., 2014). (b) The SMAP-derived extents are overlaid with AR- and MCoRDS-derived 2010-2017 perennial firn aquifer (blue shading; Miège et al., 2016), 2010-2014 ice slab (cyan shading; McFerrinMacFerrin et al., 2019), and 2012 spatially coherent melt layer (white shading; Culberg et al., 2021) detections along OIB flight lines (black lines). Overlapping perennial firn aquifer and ice slab detections are interpreted as perched firn aquifer areas. The red line is AR radargram profile along perennial firn aquifer transect A-B (Fig. 3a). The orange line is AR radargram profile along ice slab transect C-D (Fig. 3b).

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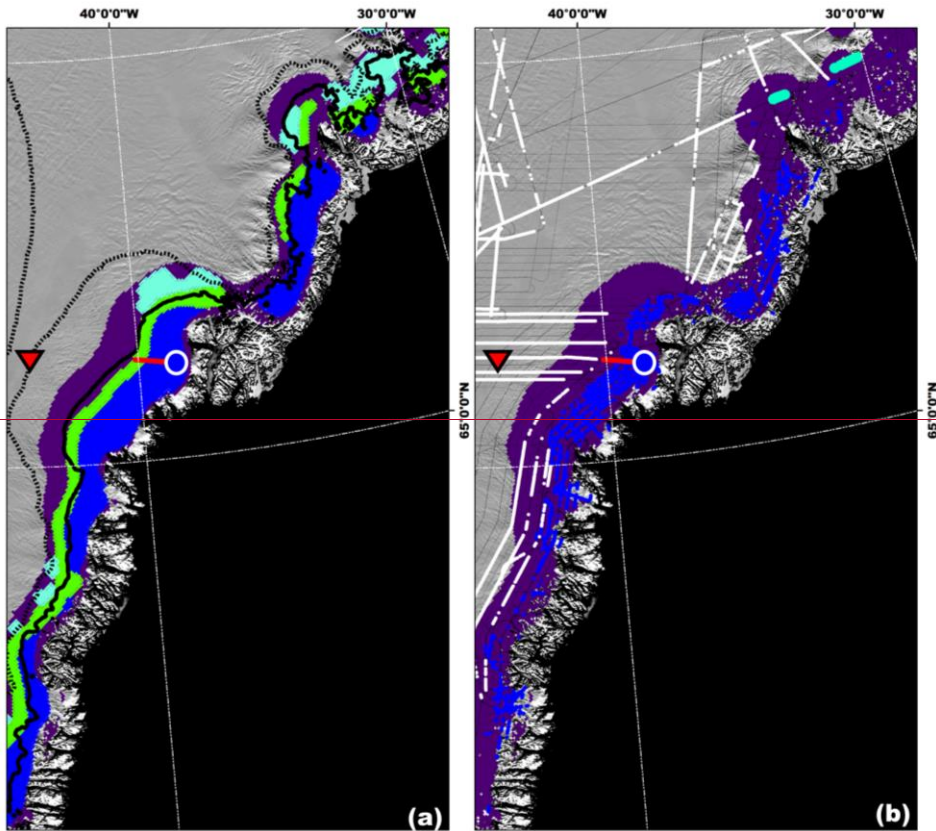
Figure 8

(a) The SMAP-derived perennial firn aquifer (blue shading), ice slab (cyan shading), perched firn aquifer (green shading), and percolation facies (purple shading) extents (2015-2019) generated by the adapted empirical algorithm over south eastern Greenland (Fig. 1c; zoom area in red box) overlaid on the 2015 MODIS Mosaic of Greenland image map (Haran et al., 2018). The solid black line is the 2000 m.a.s.l. contour, and the black dotted line is the 2500 m.a.s.l. contour (Howat et al., 2014). (b) The SMAP-derived percolation facies extent is overlaid with AR- and MCoRDS-derived 2010-2017 perennial firn aquifer (blue shading; Miège et al., 2016), 2010-2014 ice slab (cyan shading; McFerrinMacFerrin et al., 2019), and 2012 spatially coherent melt layer (white shading; Culberg et al., 2021) detections along OIB flight lines (black lines). Overlapping perennial firn aquifer and ice slab detections are interpreted as perched firn aquifer areas. The red line is AR radargram profile along perennial firn aquifer transect A-B (Figs. 1; 3a). The blue circle is a perennial firn aquifer area (Figs. 3a; 4a). The red triangle is a high-elevation (~2500 m.a.s.l.) percolation facies area (Figs. 4d).
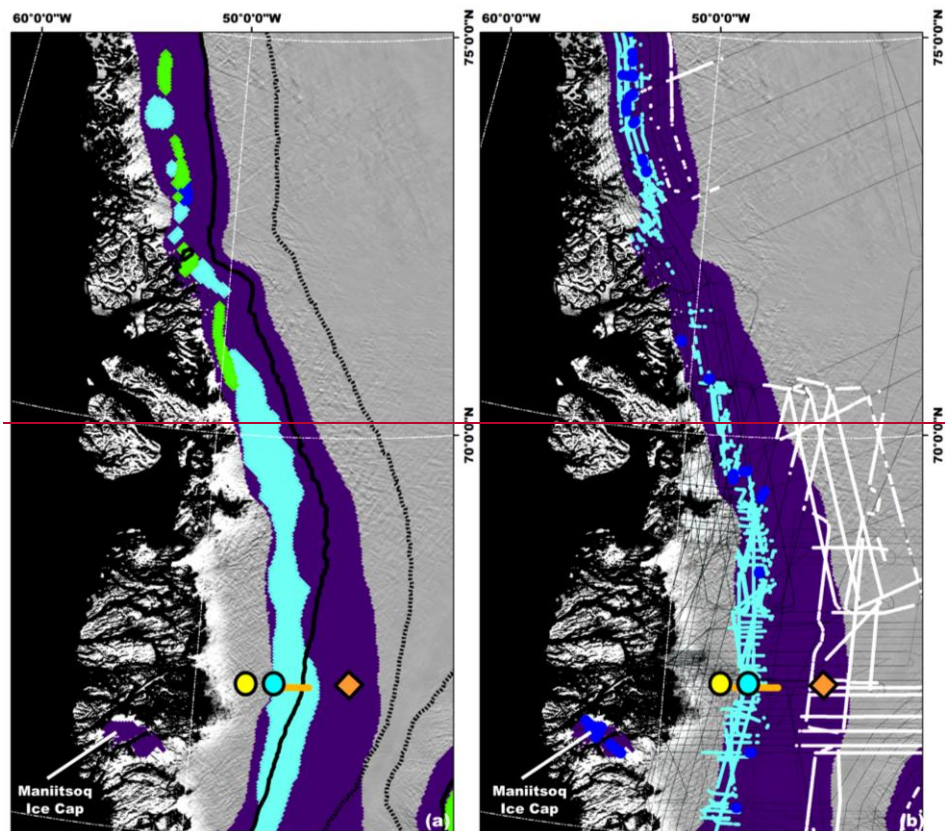
Figure 9

(a) The SMAP-derived perennial firn aquifer (blue shading), ice slab (cyan shading), perched firn aquifer (green shading), and percolation facies (purple shading) extents (2015-2019) generated by the adapted empirical algorithm over south western Greenland (Fig. 1c; zoom area in red box) overlaid on the 2015 MODIS Mosaic of Greenland image map (Haran et al., 2018). The solid black line is the 2000 m.a.s.l. contour, and the black dotted line is the 2500 m.a.s.l. contour (Howat et al., 2014). (b) The SMAP-derived percolation facies extent is overlaid with AR- and MCoRDS-derived 2010-2017 perennial firn aquifer (blue shading; Miège et al., 2016), 2010-2014 ice slab (cyan shading; McFerrin~~MacFerrin~~ et al., 2019), and 2012 spatially coherent melt layer (white shading; Culberg et al., 2021) detections along OIB flight lines (black lines). Overlapping perennial firn aquifer and ice slab detections are interpreted as perched firn aquifer areas. The orange line is AR radargram profile along ice slab transect C-D (Figs 1; 3b). The cyan circle is a perched firn aquifer area (Figs. 3b; 4b). The orange diamond is a percolation facies area (Fig. 4c). The yellow circle is a superimposed ice area (Fig. 4e).

ξ, and ζ values mapped within the perennial firn aquifer extent indicates the widespread presence of thicker water-saturated firn layers with larger volumetric fractions of meltwater that are radiometrically warm during both the melting and freezing seasons, and have extended refreezing rates. The ice slab extent (76,000 km²) is mapped at elevations between ~800 m a.s.l and 2700 m a.s.l., and extends over ~13 % of the percolation facies extent, and ~4 % of the GrIS extent. As compared to perennial firn aquifer areas, decreased $T_{V,max}^B$, $T_{V,min}^B$, ξ and ζ values in ice slabs areas indicates the presence of thinner water-saturated firn layers with lower volumetric fractions of meltwater that are radiometrically colder, and have slightly more rapid refreezing rates. ~~The SMAP-derived perched firn aquifer calibration parameter intervals are within the perennial firn aquifer and the ice slab calibration parameter intervals (see Fig. 6, Table 3). The extents of these three sub-facies within the broader percolation facies are overlapping, such that perched firn aquifers typically represent the upper boundary of perennial firn aquifer areas that subsequently transition to percolation facies as well as the lower boundary of ice slab areas that subsequently transition to wet snow facies. However, in several areas in south and southeastern Greenland, the full progression of facies and sub-facies (dry snow facies - percolation facies - ice slab - perched firn aquifer - perennial firn aquifer – ablation facies) occur (e.g., Fig. 7a). The perched firn aquifer extent (37,000 km²) is mapped at elevations between ~600 m.a.s.l and 2700 m.a.s.l., and extends over ~30% of perennial firn aquifer extent, ~24% of ice slab extent, ~2% of the percolation facies extent, and less than ~ 1% of the GrIS extent.~~ Combined together, the total extent (~140,000 km²) is the equivalent of ~24% of the percolation facies extent, and 10% of the GrIS extent. The extents of these sub-facies are generally isolated and somewhat scattered within the percolation facies. However, in several areas in south, south and central eastern, and northern Greenland, the sequential formation of facies and sub-facies (dry snow facies - percolation facies - ice slab - perennial firn aquifer – ablation facies) is mapped. ~~This increases previous AR- and MCoRDS-derived elevation estimates upslope ~600 m.a.s.l. in both perennial firn aquifer areas (Miège et al., 2016) and ice slab areas (McFerrin~~MacFerrin ~~et al., 2019).~~

~~The highest perennial firn aquifer, ice slab, and perched firn aquifer elevations (>2500 m.a.s.l.) are mapped in southern Greenland and on the Geikie plateau, in central eastern Greenland.~~

~~**Table 4.** *The SMAP-derived perennial firn aquifer, ice slab, and perched firn aquifer extents (2015-2019) over the percolation facies and the GrIS, and the elevation range at which they are mapped.*~~

	~~Percolation Facies Extent (%)~~	~~Ice Sheet Extent (%)~~	~~Elevation Range (m.a.s.l.)~~
~~Perennial Firn Aquifers~~	~~11~~	~~4~~	~~600 – 2600~~
~~Ice Slabs~~	~~13~~	~~4~~	~~800 – 2700~~
~~Perched Firn Aquifers~~	~~5~~	~~<1~~	~~600 – 2700~~

1198 Figs. 7b-9b shows perennial firn aquifers ~~(blue shading), ,~~ ice slabs~~, and spatially coherent melt~~
1199 ~~layers (cyan shading), and spatially coherent melt layers (white shading)~~ detected by airborne ice-
1200 penetrating radar surveys ~~(2010-2017)~~ overlaid on the SMAP-derived percolation facies extent ~~(2015-~~
1201 ~~2019).~~ Perched firn aquifer areas are inferred where perennial firn aquifer and ice slab detections overlap.
1202 The SMAP-derived perennial firn aquifer extent mapped in southern, and south and central eastern
1203 Greenland is consistent with the AR- and MCoRDS ~~geophysical~~ derived perennial firn aquifer detections
1204 ~~(2010-2017), except in north western Greenland where perched firn aquifers are alternatively mapped.~~
1205 Additional smaller perennial firn aquifer areas are mapped in northern Greenland. The SMAP-derived ice
1206 slab extent mapped in south western, and central eastern, ~~central and north eastern, and northern~~
1207 Greenland is generally consistent with the spatial patterns of the AR-derived ice slab detection~~ss (2010-~~
1208 ~~2014)~~, however, is significantly ~~significantly~~ expanded upslope in ~~each of~~each of these ~~these~~ area~~s~~. In
1209 northern Greenland, perennial firn aquifers areas are alternatively mapped, and additional expansive ice
1210 slab areas are mapped upslope of perennial firn aquifer areas. Additional smaller ice slab areas are mapped
1211 in south and south eastern (Figs. 9a; 9b) Greenland. ~~We note that the AR- and MCoRDS-derived perennial~~
1212 ~~firn aquifer and ice slab detections are limited in space and time, particularly in northern Greenland, with a~~
1213 ~~time interval as large as nine years between the airborne ice-penetrating radar surveys and the SMAP~~
1214 ~~enhanced-resolution T_V^B imagery we use in our adapted empirical algorithm. In western and northern~~
1215 ~~Greenland, the 2015 melting season was especially intense (Tedesco et al., 2016). And, in northern~~
1216 ~~Greenland, the ablation facies have recently (2010-2019) increased in extent (Noël et al., 2019), and~~
1217 ~~supraglacial lakes have recently (2014-2019) advanced inland (Turton et al., 2021), indicating a possible~~
1218 ~~geophysical basis for the observed formation, boundary transitions, and expansion.~~ Neither perennial firn
1219 aquifer or ice slab areas are mapped on the Maniitsoq and Flade Isblink Ice Caps, where spatially integrated
1220 L-band emission results in calibration parameter values outside the defined intervals for each of these sub-
1221 facies.
1222 Although the AR-derived spatially coherent melt layers detections are often observed to be adjacent
1223 to perennial firn aquifer and ice slab areas, these sub-facies were masked in the original airborne ice
1224 penetrating radar survey analysis by Culberg et al., (2021). Spatially coherent melt layers often overlay
1225 perennial firn aquifers (e.g., Fig. 3a), and merge with ice slabs (Culberg et al., 2021; Fig.4).~~We note that~~
1226 ~~the AR-derived ice slab detections are limited in space and time, particularly in northern Greenland, with a~~
1227 ~~time interval as large as nine years between the airborne ice penetrating radar surveys and the SMAP~~
1228 ~~enhanced-resolution T_V^B imagery used in the adapted empirical algorithm (i.e., 2010 to 2019). In western~~
1229 ~~and northern Greenland, the 2015 melting season was especially intense (Tedesco et al., 2016). And, in~~
1230 ~~northern Greenland, the ablation facies have recently increased in extent (2010-2019; Noël et al., 2019),~~
1231 ~~and supraglacial lakes have recently advanced inland (2014-2019; Turton et al., 2021), indicating a likley~~
1232 ~~geophysical basis for the observed upslope expansion. In central and north eastern, and northern~~
1233 ~~Greenland, perched firn aquifers are often alternatively mapped.~~ Additional smaller ice slab areas are
1234 ~~mapped in south and south eastern (Figs. 9a; 9b) Greenland. The scattered SMAP-derived perched firn~~

aquifer extent mapped in north western and central eastern Greenland is fairly consistent with the sparse AR- and MCoRDS-derived perched firn aquifer detections (2010-2017), however, in central western Greenland (Figs. 9a; 9b) ice slab areas are alternatively mapped. Expansive additional perched firn aquifer areas are mapped in southern, and south and central eastern Greenland. These areas are often coincident with spatially coherent melt layer detections, particularly in south eastern Greenland (Figs. 8a, 8b). Neither perennial firn aquifer, ice slab, nor perched firn aquifer areas are mapped on the Maniitsoq and Flade Isblink Ice Caps. Over these two small ice caps, L-band emissions spatially integrated with emissions from rock, land, the ocean, and adjacent percolation facies and wet snow facies areas result in SMAP-derived calibration parameter values outside the defined intervals for each of these sub-facies.

We infer that the SMAP-derived perched firn aquifer extent represents L-band emissions from: (1) spatially expansive, relatively shallow water-saturated firn layers with lower volumetric fractions of meltwater as compared to perennial firn aquifer areas. These shallow water-saturated firn layers transiently form on top of buried ice slabs, spatially coherent melt layers, or other semi-impermeable layers that have previously formed within the upper snow and firn layers of the percolation facies, as shown in Figs. 7-8. Or, (2) spatially scattered deeper water-saturated firn layers with larger volumetric fractions of meltwater (i.e., perennial firn aquifers) that are spatially integrated with L-band emissions from adjacent ice slabs, percolation facies, and/or wet snow facies areas. These areas are observed as shallow water-saturated firn layers with lower volumetric fractions of meltwater over SMAP's ~18 km footprint (i.e., the effective resolution). Or, (3) a combination of these englacial firn hydrological features, which is a likely scenario over many perched firn aquifers areas. This is particularly likely in north western Greenland, where airborne ice penetrating radar surveys consistently detect perennial firn aquifers; however, the SMAP-derived extent indicates perched firn aquifer areas (Fig. 7).

Shallow buried supraglacial lakes have recently been identified within the percolation facies of western, northern, and north and central eastern Greenland using airborne ice -penetrating radar surveys (Koenig et al., 2015) and satellite synthetic aperture radar imagery (Miles et al., 2017; Schröder et al., 2020; Dunmire et al., 2021). These buried supraglacial lakes are within the SMAP-derived perennial firn aquifer and, ice slab, and perched firn aquifer extents, however, they are not expected to significantly influence L-band emissions in these areas for two reasons.: (1) Aas compared to SMAP's ~18 km footprint, the mean extent of buried supraglacial lakes is limited (less than ~1 km^2), and they are sparsely distributed in perennial firn aquifer and, ice slab, and perched firn aquifer areas (Dunmire et al., 2021). (2) Supraglacial lakes form during the melting season as a result of meltwater storage withinmeltwater storage in topographic depressions at the ice sheet surface (Echelmeyer et al. 1991). Similar to subglacial lakes (Jezek et al., 2015) and perennial firn aquifers (Miller et al., 2020), supraglacial lakes represent radiometrically cold subsurface near-surface meltwater reservoirs. Upwelling L-band emissions from deeper firn layers, superimposed and/or glacial ice, and the underlying bedrock are effectively blocked by

high reflectivity and attenuation at the interface between the lake bottom and the underlying impermeable layer. This results in a low observed T^B at the upper surface of meltwater stored within supraglacial lakes. During the freezing season, the upper surface of meltwater refreezes and forms a partial or solid-ice cap that is sometimes buried by snow accumulation (Koenig et al., 2015). Airborne ice-penetrating radar surveys in April and May between 2009 and 2012 suggest the mean depth to the upper surface of meltwater stored within buried supraglacial lakes is approximately 2 m (Koenig et al., 2015). Over buried supraglacial lakes, L-band emission from the refreezing partial or solid-ice cap, which is smooth relative to the L-band wavelength (~21 cm), likely induces surface scattering. As a result, T_V^B decreases over buried supraglacial lakes are likely negligible. Thus, over SMAP's ~18 km footprint, we postulate water-saturated firn layers dominate L-band emission over the percolation facies of the GrIS.

The SMAP-derived perennial firn aquifer extent (64,000 km^2) generated by our adapted empirical algorithm and the multi-year calibration technique (2015-2019) is consistent with the extent (66,000 km^2) generated by the previously developed empirical algorithm and the single-coincident year calibration technique (2016) described in Miller et al., 2020. The SMAP-derived perennial firn aquifer extent is generally consistent with previous C-band (5.3 GHz) satellite radar scatterometer-derived perennial firn aquifer extents mapped using the Advanced SCATterometer (ASCAT) on the European Organization for the Exploitation of Meteorological Satellites (EUMETSAT) Meteorological Operational A (MetOp-A) satellite (2009-2016, 52 000-153 000 km^2; Miller, 2019), and the Active Microwave Instrument in radar scatterometer mode (ESCAT) on ESA's European Remote Sensing (ERS) satellite series (1992-2001, 37 000-64 000 km^2; Miller, 2019) as well as the C-band (5.4 GHz) synthetic aperture radar-derived extent mapped using ESA's Sentinel-1 satellite (2014-2019, 54 000 km^2; Brangers et al., 2020). The exception is the ASCAT-derived perennial firn aquifer extent (2012-2013, 153,000 km^2; Miller, 2019) mapped following the 2012 melting season (Nghiem et al., 2012) in which significant changes in the dielectric and geophysical properties that influence radar backscatter likely occurred. The unreasonably expansive (i.e., more than twice the mean) mapped extent is a result of ASCAT'S shallow (several meters) C-band penetration depth (Jezek et al., 1994), and the simple threshold-based algorithm, which was not calibrated for an extreme melting season that included saturation of the upper snow and firn layers of the dry snow facies and percolation facies with relatively large volumetric fractions of meltwater (Miller et al., 2019). Water-saturated firn layers had extended refreezing rates, however, seasonal meltwater was not stored at depth. Widespread spatially coherent melt layers were alternatively formed in many of the mapped areas (Culberg et al., 2021). The

SMAP-derived ice slab extent (76,000 km^2) is also consistent with previous AR-derived ice slab extents (2010-2014, 64,800 km^2-69,400 km^2; McFerrin et al., 2019).

The SMAP-derived perennial firn aquifer extent (~64,000 km^2) generated by our adapted empirical algorithm and the multi-year (2010-2017) calibration technique is consistent with the extent (~66,000 km^2) generated by the previously developed empirical algorithm and the single-coincident year (2016) calibration technique described in Miller et al., 2020. The SMAP-derived perennial firn aquifer extent is generally consistent with previous C-band (5.3 GHz) satellite radar scatterometer-derived perennial firn aquifer extents mapped using the Advanced SCATterometer (ASCAT) on the European Organization for the Exploitation of Meteorological Satellites (EUMETSAT) Meteorological Operational A (MetOp-A) satellite (2009-2016, ~52 000-153 000 km^2; Miller, 2019), and the Active Microwave Instrument in radar scatterometer mode (ESCAT) on ESA's European Remote Sensing (ERS) satellite series (1992-2001, ~37 000-64 000 km^2; Miller, 2019) as well as the C-band (5.4 GHz) synthetic aperture radar-derived extent mapped using ESA's Sentinel-1 satellite (2014-2019, ~54 000 km^2; Brangers et al., 2020). The exception is the ASCAT-derived perennial firn aquifer extent (2012-2013, ~153,000 km^2; Miller et al., 2019) mapped following the anomalous 2012 melting season (Nghiem et al., 2012) in which significant changes in the dielectric and geophysical properties that influence radar backscatter and the temporal C-band signatures occurred. The unreasonably expansive (i.e., more than twice the mean) mapped extent is a result of ASCAT'S shallow (~several meters) C-band penetration depth (Jezek et al., 1994), and the simple threshold-based algorithm that was not calibrated for an extreme melting season that included saturation of the upper snow and firn layers of the dry snow facies and percolation facies with relatively large volumetric fractions of meltwater (Miller et al., 2019). Water-saturated firn layers had extended refreezing rates, however, seasonal meltwater was not stored at depth. Spatially coherent melt layers were alternatively formed in many of the mapped areas (Culberg et al., 2021). The SMAP-derived ice slab extent (~76,000 km^2) is also consistent with previous AR-derived ice slab extents (2010-2014, ~64,800 km^2-69,400 km^2; McFerrinMacFerrin et al., 2019).

Although we simply consider our mapped extents a high-probability area for preferential formation, the maps generated by our adapted empirical algorithm and the multi-year (2010-2017) calibration technique for individual years suggest there reasonable interannual variability in perennial firn aquifer and , ice and slab, and perched firn aquifer extents (, which is summarized in Table 35). Our results demonstrate reasonable sensitivity to the variability in the depth- and time-integrated dielectric and geophysical properties of the percolation facies that dielectric and geophysical properties that influence the radiometric temperature and temporal L-band signatures, even during the extreme 2015 melting season (Tedesco et al., 2016).

4   Implications

Seasonal surface melting over the GrIS has increased in extent, intensity, and duration since early in the satellite era (Steffen et al., 2004; Tedesco e al., 2008; Tedesco et al., 2011; Nghiem et al., 2012; Tedesco et al., 2016; Tedesco and Fettweis, 2020; Cullather et al., 2020). Consistent with recent seasonal surface

melting trends, meltwater run-off has accelerated to become the dominant mass loss mechanism over the GrIS (van den Broeke et al., 2016). Meltwater storage in both solid (i.e., embedded ice structures, including ice slabs, spatially coherent melt layers) and liquid (i.e., perennial firn aquifers) form can buffer meltwater run-off in the percolation facies and delay its eventual release into the ocean (Harper et al., 2012). However, significant uncertainty remains in meltwater run-off estimates as a result of the lack of knowledge of heterogeneous infiltration and refreezing processes within the snow and firn layers (Pfeffer and Humphrey, 1996), and the depths to which meltwater can descend beneath the ice sheet surface (Humphrey et al., 2012).

If the increasing seasonal surface melting trend continues (Franco et al., 2013; Noël et al., 2021), perennial firn aquifer formation and expansion may increase the possibility of crevasse-deepening via meltwater-induced hydrofracturing (Alley et al., 2005; van der Veen, 2007), especially if crevasse fields expand into perennial firn aquifer areas as a result of accelerated ice flow (Colgan et al., 2016). Meltwater-induced hydrofracturing is an important component of supraglacial lake drainage during the melting season (Das et al., 2008; Stevens et al., 2015) leading to at least temporary localized accelerated ice flow velocities (Zwally et al., 2002; Joughin et al., 2013; Moon et al., 2014) as well as ice discharge from outlet glaciers (Chudley et al., 2019), and mass balance changes (Joughin et al., 2008). Perennial firn aquifers may also support meltwater-induced hydrofracturing, even during the freezing season (Poinar et al., 2017; Poinar et al., 2019).

The formation and expansion of ice slabs reduces permeability within the upper snow and firn layers and facilitates lateral meltwater flow with minimum vertical percolation into the deeper firn layers, thereby enhancing meltwater run-off and mass loss at the periphery (MacFerrin et al., 2019). Lateral meltwater flow across ice layers overlying deeper permeable firn layers was first postulated by Müller (1962). The theory was then further developed by Pfeffer et al., (1991) as an end-member case for meltwater run-off in the percolation facies, with the other end member case being lateral meltwater flow across superimposed ice in the wet snow facies and/or across glacial ice in the ablation facies. Lateral meltwater flow and high-elevation (1850 m a.s.l) meltwater run-off across ice slabs in the percolation facies was recently observed in visible satellite imagery collected by the NASA-USGS Landsat 7 mission during the 2012 melting season (MacFerrin et al., 2019).

Spatially coherent melt layers represent a newly identified end-member case for meltwater runoff in the dry snow facies (Culberg et al., 2021). Similar to ice slabs, the formation and expansion of spatially coherent melt layers reduces the pore space within the upper snow and firn layers, and may also facilitate lateral meltwater flow with minimum vertical percolation into the deeper firn layers, thereby enhancing meltwater run-off from significantly higher elevations on accelerated time scales. The rapid formation of spatially coherent melt layers that merge with ice slabs upslope of perennial firn aquifers areas may simultaneously accelerate both meltwater run-off and meltwater-induced hydrofracturing during extreme melting seasons. The formation of spatially coherent melt layers overlying deeper perennial firn aquifers may result in the formation of shallow perched firn aquifers (Culberg et al., 2021), or may terminate gravity-

driven meltwater drainage and seasonal recharging (Fountain and Walder, 1998), which may eventually completely refreeze stored meltwater into ice slabs or decimeters thick solid-ice layers overlying deeper glacial ice.

Nghiem et al., (2003) previously demonstrated mapping spatially coherent melt layers that were formed following the anomalous 2002 melting season (Steffen et al., 2004) using similar signatures observed in Ku-band radar backscatter time series collected by the SeaWinds radar scatterometer that was flown on NASA's QuikSCAT satellite (Tsai et al., 2000).

Table 5 *The SMAP-derived perennial firn aquifer, ice slab, and perched firn aquifer extents (2015-2019).*

	Perennial Firn Aquifer Extent (km²)	Ice Slab Extent (km²)	Perched Firn Aquifer Extent (km²)
2015-2019	66,000	76,000	37,000
2015-2016	63,000	23,000	17,000
2016-2017	69,000	48,000	38,000
2017-2018	73,000	27,000	20,000
2018-2019	70,000	38,000	26,000

Consistent with recent seasonal surface melting trends, meltwater run-off has accelerated to become the dominant mass loss mechanism over the GrIS (van den Broeke et al., 2016). However, significant uncertainty remains in meltwater run-off estimates in the percolation facies as a result of the lack of knowledge of heterogeneous infiltration processes within the snow and firn layers (Pfeffer and Humphrey, 1996), the depths to which meltwater can descend beneath the ice sheet surface (Humphrey et al., 2012), and the formation of englacial firn hydrological features (Benson et al., 1960; Humphrey et al., 2012; Forster et al., 2014), especially ice layers and ice slabs (Machguth et al., 2016, McFerrinMacFerrin et al., 2019; Culberg et al., 2021). A notable example of this lack of knowledge is the identification by Forster et al., (2014) of widespread perennial firn aquifers within the percolation facies of the GrIS via airborne ice-penetrating radar surveys collected by NASA's OIB campaigns (2010-2014; Rodriguez-Morales et al 2014) that store large volumes (~140 Gt; Koenig et al., 2014) of meltwater that was previously unknown. The

mapped extent (2010-2014) shown in Forster et al., (2014) can be distinctly observed in 1978 enhanced resolution Ku-band radar backscatter imagery (Long and Drinkwater, 1994) collected by the radar scatterometer on NASA's first Earth-observing satellite - the Seasat-A mission (Jones et al., 1982). This suggests that Greenland's perennial firn aquifers have likely existed undetected in the deeper firn layers of the percolation facies for decades. Meltwater storage in both solid (i.e., embedded ice structures) and liquid (i.e., perennial firn and perched firn aquifers) form can buffer meltwater run-off in the percolation facies (Harper et al., 2012). However, the formation of near-surface ice layers and ice slabs reduces the pore space within the upper snow and firn layers and facilitates lateral meltwater flow with minimum vertical percolation into the deeper firn layers, thus enhancing meltwater run-off downslope towards the periphery. Lateral meltwater flow across ice layers overlying deeper permeable firn layers was first postulated by Müller (1962). The theory was then further developed by Pfeffer et al., (1991) as an end-member case for meltwater run-off, with the other end member case being lateral meltwater flow across superimposed ice in the wet snow facies and/or across glacial ice in the ablation facies. McFerrinMacFerrin et al., (2019) recently identified widespread near-surface ice slabs within the percolation facies of the GrIS via airborne ice-penetrating radar surveys collected by NASA's OIB campaigns (2010-2014; Rodriguez-Morales et al. 2014). Lateral meltwater flow and high-elevation (~1850 m.a.s.l) meltwater run-off across the identified ice slabs was also observed in visible satellite imagery collected by the NASA-USGS Landsat 7 mission (e.g. Goward et al., 2001). This was also observed during the anomalous 2012 melting season (McFerrinMacFerrin et al., 2019) during which seasonal surface melting extended over ~99% of the GrIS (Nghiem et al., 2012)

Additionally, we correlate exponentially decreasing temporal L-band signatures with AR- and MCoRDS-derived detections where perennial firn aquifer and ice slab areas overlap. We identify these transitional areas as perched firn aquifer areas. We infer that, in these areas, shallow water-saturated firn layers transiently form on top of buried ice slabs or other semi-impermeable layers, such as spatially coherent melt layers that form in the higher elevations (> ~2000 m.a.s.l.) of the percolation facies and the dry snow facies that were recently identified via AR (Culberg et al., 2021). Perched firn aquifers likely form during some melting seasons as a result of interannual variability in surface melting and snow accumulation, and the formation of englacial firn hydrological features.

5   Summary and Future Work

Many open questions remain about Greenland's perennial firn aquifers, regarding initial formation, extent, depth, flow characteristics, timescales of refreezing and/or englacial drainage, and connections to the subglacial hydrological system. Seasonal surface melting over the GrIS has increased in extent, intensity, and duration since the beginning of the satellite era (Steffen et al., 2004; Tedesco e al., 2008; Tedesco et al., 2011; Nghiem et al., 2012; Tedesco et al., 2016; Tedesco and Fettweis, 2020; Cullather et al., 2020). If this trend continues (Franco et al., 2013; Noël et al., 2021), subsequent increases in the volume of meltwater stored within Greenland's perennial firn aquifers will increase the possibility of

crevasse-deepening via meltwater-induced hydrofracturing (Alley et al., 2005; van der Veen, 2007), especially if crevasse fields laterally expand into perennial firn aquifer areas as a result of accelerated ice flow (Colgan et al., 2016). Meltwater-induced hydrofracturing is an important component of supraglacial lake drainage during the melting season (Das et al., 2008; Stevens et al., 2015) leading to at least temporary accelerated flow velocities (Zwally et al., 2002; Joughin et al., 2013; Moon et al., 2014) and mass balance changes (Joughin et al., 2008). Greenland's firn perennial aquifers may also support meltwater-induced hydrofracturing, even during the freezing season (Poinar et al., 2017; 2019).

L-band satellite microwave sensors – including NASA's L-band SMAP mission - represent a relatively new Earth-observation tool that has exceptional capabilities for cryospheric applications. Especially, mapping englacial and subglacial hydrological features at depths of ~tens to hundreds of meters beneath the surface of Earth's polar ice sheets. In this study, for the first time, we have exploited this capability and demonstrated the novel use of the L-band microwave radiometer on NASA's SMAP satellite for mapping perennial firn aquifers and, ice slabs, and perched firn aquifers together as a continuous system over the percolation facies of the GrIS. We have also demonstrated that SMAP enhanced-resolution L-band T_V^B imagery can effectively resolve percolation facies features that are not effectively resolved in conventionally processed SMAP L-band T_V^B imagery (e.g., Fig. 1). We have adapted our previously developed empirical algorithm (Miller et al., 2020) by expanding our analysis of spatiotemporal differences in SMAP enhanced-resolution T_V^B imagery and temporal L-band signatures over the GrIS. We have used this analysis to derive a firn saturation parameter from a simple two-layer L-band geophysical brightness temperature model (Miller et al., 2021, in press). And, we have used the firn saturation parameter to map the extent of the percolation facies. We have found that by correlating maximum and minimum T_V^B values, the firn saturation parameter, and the refreezing rate parameter with perennial firn aquifer and, ice slab, and perched firn aquifer detections identified via via the CReSIS AR and MCoRDS instruments flown by NASA's OIB campaigns that we can calibrate our previously developed empirical algorithm (Miller et al., 2020) to map plausible extents.

We note that significant uncertainty exists in the mapped extents as a result of (1); (1) correlating the SMAP-derived parameters with airborne ice-penetrating radar detections that are not coincident in time, (2) (2) the lack of a distinct temporal L-band signature delineating the boundary between perennial firn aquifer areas, ice slabs areas, and adjacent percolation facies areas delineating the boundary between each of the mapped sub-facies within the broader percolation facies, and (3) (3) the much more limited extent of the airborne ice-penetrating radar detections as compared to the rSIR grid cell extent, as well as and the effective resolution of the SMAP enhanced-resolution T_V^B imagery. Additional uncertainty exists in the perched firn aquifer extent as a result of fitting L-band signatures to the continuous logistic model, which is not optimal for these specific sub-facies.

Miller et al., (2020) normalized SMAP enhanced-resolution T_V^B time series and converted the exponential rate of T_V^B decrease over perennial firn aquifer areas to a binary parameter to map extent. In this study, we have converted the SMAP-derived parameters to binary parameters to map the extent of both perennial firn aquifer, and, ice slab, and perched firn aquifer areas. Moreover, we have included additional analysis of the spatiotemporal differences in maximum and minimum T_V^B values, the firn saturation parameter, and the refreezing rate parameter. We have shown that spatiotemporal differences in the SMAP-derived parameters are consistent with our assumption of spatiotemporal differences in the englacial hydrology and thermal characteristics of firn layers at depth.

~~Particularly, our assumption that latent heat release influences temporal L-band signatures within the percolation facies of the GrIS. This includes continuous latent heat release via the slow refreezing of the deeper firn layers in perennial and perched firn aquifer areas that are saturated with large volumetric fractions of meltwater. And, latent heat release that occurs throughout the percolation facies via more rapid refreezing of seasonal meltwater by the descending winter cold wave, and the subsequent formation of embedded ice structures, including ice slabs and spatially coherent melt layers, within the upper snow and firn layers.~~

Future work will focus on simulating ~~maximum and minimum T_V^B, the firn saturation parameter, and the refreezing rate parameter as well as~~ temporal L-band signatures observed over perennial firn aquifer and, ice slab, and perched firn aquifer areas ~~within the percolation facies of the GrIS~~ for a wide range of geophysical properties. ~~Significant interannual variability in the dielectric and geophysical properties that seasonally influence the radiometric temperature and temporal L-band signatures can occur, particularly following extreme melting seasons, such that it is critical that these properties are understood and considered in any given year.~~ To better interannual variability, in extent ~~as well as other geophysical properties~~, we will interpret our results together with climatological parameters, such as snow accumulation, liquid water content, temperature, and surface mass balance simulated, and over the GrIS simulated using the by the Regional Atmospheric Climate Model (RACMO2.3p2; Noël et al., 2018). Additionally, we will simulate the distinct temporal L-band signatures observed over spatially coherent melt layers ~~in the upper snow and firn layers of the dry snow facies and percolation facies of the GrIS recently identified via MCoRDS flown by NASA's OIB campaigns (Culberg et al., 2021) following the anomalous 2012 melting season (Nghiem et al., 2012) and~~ and ~~as well as~~ explore ~~the potential for~~ mapping the the extent ~~of these near-surface englacial hydrological features using satellite L-band microwave radiometry.~~ ~~Nghiem et al., (2003) previously demonstrated mapping spatially coherent melt layers that were formed following the anomalous 2002 melting season (Steffen et al., 2004) using similar signatures observed in Ku-band radar backscatter time series collected by the SeaWinds radar scatterometer that was flown on NASA's QuikSCAT satellite (Tsai et al., 2000).~~

Combining multi-layer depth-integrated L-band ~~geophysical~~ brightness temperature models (e.g., Jezek et al., 2015) that include embedded ice structure parametrizations (e.g., Jezek et al., 2018) with models of depth-dependent geophysical parameters can lead to an improved understanding of the

1518 extremely complex and ~~very~~ poorly described physics controlling L-band emission~~s~~ over the percolation
1519 facies ~~of the GrIS. For L-band emissions over perennial firn aquifer, ice slab, perched firn aquifer, and~~
1520 ~~spatially coherent melt layer areas, the~~ ~~Key~~key geophysical parameters include atmospheric temperature
1521 forcing, physical temperature versus depth, latent heat, snow accumulation, the volumetric fraction and
1522 depth of meltwater, and the volumetric fraction and geometric configuration of embedded ice structures.
1523 The development of more sophisticated empirical algorithms that incorporate multi-layer depth-integrated
1524 L-band ~~geophysical~~ brightness temperature models that are constrained by in situ measurements can help
1525 reduce the significant uncertainty in the current mapped extents, and provide more accurate boundary
1526 ~~delineation~~ ~~ies delineating each of these sub-facies within the broader percolation facies~~ that can be used
1527 to further quantify the interannual variability in future mapped extents of perennial firn aquifer, ice slabs and
1528 spatially coherent melt layer areas. ~~in extent. As Greenland's climate continues to warm, and seasonal~~
1529 ~~surface melting increases in extent, intensity, and duration, quantifying the possible rapid expansion of each~~
1530 ~~of these sub-facies using satellite L-band microwave radiometry has significant implications for~~
1531 ~~understanding ice sheet-wide variability in englacial firn hydrology resulting in meltwater-induced~~
1532 ~~hydrofracturing and accelerated ice flow as well as high-elevation run-off that can impact the mass balance~~
1533 ~~and stability of the GrIS.~~
1534 ~~The results presented in this study demonstrate the outstanding potential of L-band satellite~~
1535 ~~microwave sensors for mapping englacial firn hydrological features within the percolation facies of the GrIS~~
1536 ~~that can be extended to forthcoming satellite missions, such as the NASA-ISRO SAR mission (NISAR),~~
1537 ~~ESA's Copernicus Imaging Microwave Radiometer (CIMR) mission, ESA's Copernicus Radar Observation~~
1538 ~~System for Europe in L-band (ROSE-L) mission, and candidate missions, such as ESA's Earth Explorer 10~~
1539 ~~Cryorad mission.~~
1540

1541 Data Availability

1542 SMAP Radiometer Twice-Daily rSIR-Enhanced EASE-Grid 2.0 Brightness Temperatures, Version 1 ~~SMAP~~
1543 ~~enhanced-resolution L-band~~ T_{L}^{B} ~~imagery~~ (2015-2019) have been produced as part of the NASA Science
1544 Utilization of SMAP project and are available at https://doi.org/10.5067/QZ3WJNOUZLFK (Brodzik et al.,
1545 2019). The NASA MEaSUREs Greenland Ice Mapping Project (GIMP) Land Ice and Ocean Classification
1546 Mask, Version 1, is available at https://doi.org/10.5067/B8X58MQBFUPA (Howat, 2017), and the Digital
1547 Elevation Model, Version 1, is available at https://nsidc.org/data/nsidc-0645/versions/1 (Howat et al., 2015).
1548 The coastline data are available from GSHHG – A Global Self-consistent, Hierarchical, High-resolution
1549 Geography Database https://doi.org/10.1029/96JB00104 (Wessel and Smith, 1996). Ice surface
1550 temperature imagery (2015-2019) have been produced as part of the Multilayer Greenland Ice Surface
1551 Temperature, Surface Albedo, and Water Vapor from MODIS V001 data set and are available at
1552 https://doi.org/10.5067/7THUWT9NMPDK (Hall and DiGirolamo, 2019). OIB AR- and MCoRDS-derived
1553 perennial firn aquifers detections (2010-2017) are available at

https://arcticdata.io/catalog/view/doi:10.18739/A2985M (Miège et al., 2016). OIB AR-derived ice slab detections (2010-2014) are available at https://doi.org/10.6084/m9.figshare.8309777 (MacFerrin et al., 2019). OIB AR-derived spatially coherent melt layer detections (2017) are available at (https://doi.org/10.18739/A2736M33W) (Culberg et al., 2021). OIB AR L1B Geolocated Radar Echo Strength Profiles, Version 2, are available at, https://doi.org/10.5067/0ZY1XYHNIQNY (Paden et al., 2018). NASA MEaSUREs MODIS Mosaic of Greenland (MOG) 2015 Image Map, Version 2, is available at https://nsidc.org/data/NSIDC-0547/versions/2 (Haran et al., 2018). SMAP-derived perennial firn aquifer, ice slab, and perched firn aquifer extents are available from JZM upon request.

Author Contributions

JZM initiated the study, adapted the empirical model, performed the analyses, and wrote the manuscript. RC processed and interpreted the OIB AR radargram profiles. RC and DMS provided the spatially coherent melt layer detections. All authors participated in discussions and reviewed manuscript drafts.

Competing Interests

The authors declare that they have no conflict of interest.

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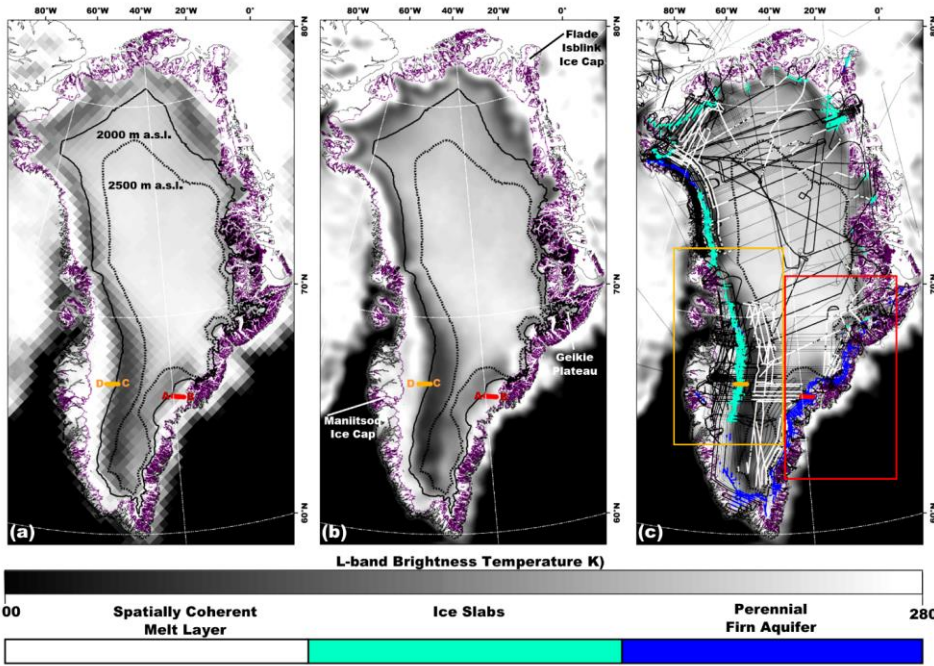
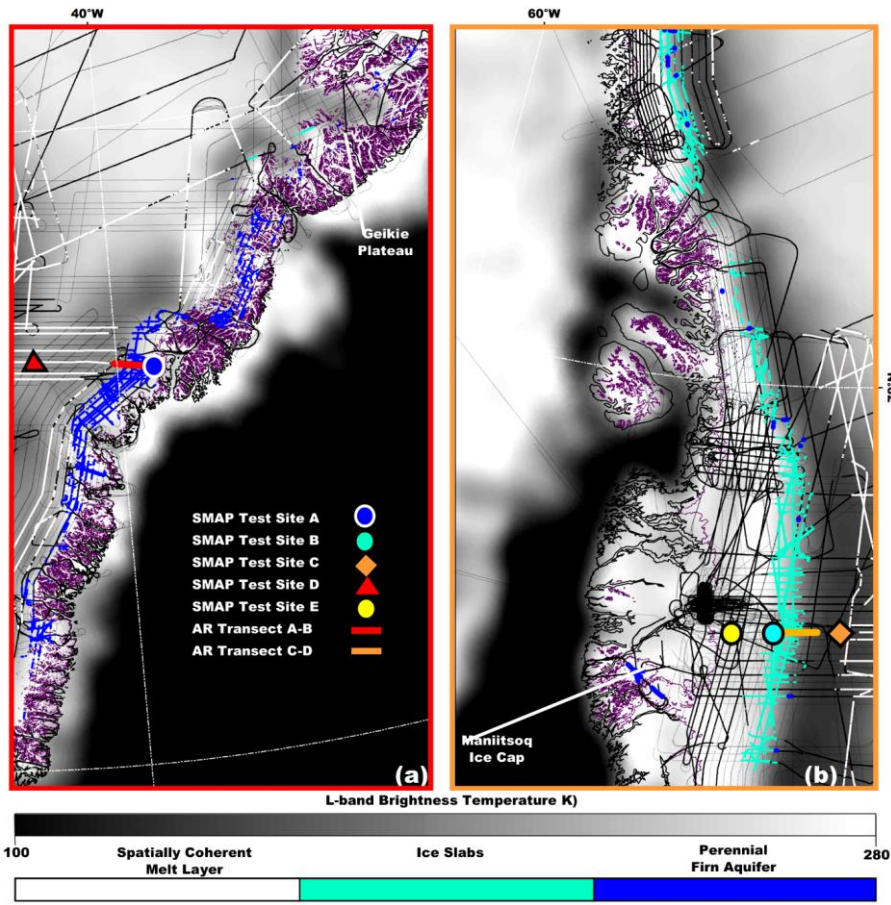


Figure 1

(a) Gridded (25 km gridding, 30 km effective resolution), and (b) enhanced-resolution (3.125 km gridding, 18 km effective resolution) L-band T_V^B imagery generated using observations collected 15 April 2016 by the microwave radiometer on the SMAP satellite during the evening orbital pass interval over Greenland (Long et al., 2019), overlaid with the 2000 m a.s.l. contour (black line), and the 2500 m a.s.l. contour (dotted black line; Howat et al., 2014); the ice sheet extent (purple line; Howat et al., 2014); and the coastline (black peripheral line; Wessel and Smith, 1996). (c) SMAP enhanced-resolution L-band T_V^B imagery overlaid with AR- and MCoRDS-derived 2010-2017 perennial firn aquifer (blue shading; Miège et al., 2016), 2010-2014 ice slab (cyan shading; MacFerrin et al., 2019), and 2012 spatially coherent melt layer (white shading; Culberg et al., 2021) detections along OIB flight lines (black interior lines); zoom areas over south eastern Greenland (red box; Fig. 2a), and south western Greenland (orange box; Fig. 2b); and AR radargram transect A-B (red line; Fig. 3a) and C-D (orange line; Fig. 3b).

1965

Figure 2

1966

1967 *Enhanced-resolution (3.125 km gridding, 30 km effective resolution) L-band T_V^B imagery generated using*

1968 *observations collected 15 April 2016 by the microwave radiometer on the SMAP satellite during the evening*

1969 *orbital pass interval over (a) south eastern Greenland (red box, Fig. 1c), and (b) south western Greenland*

1970 *(orange box, Fig. 1c,) (Long et al., 2019) overlaid with the ice sheet extent (purple line; Howat et al., 2014);*

1971 *the coastline (black peripheral line; Wessel and Smith, 1996); the AR- and MCoRDS-derived 2010-2017*

1972 *perennial firn aquifer (blue shading; Miège et al., 2016), 2010-2014 ice slab (cyan shading; MacFerrin et*

1973 *al., 2019), and 2012 spatially coherent melt layer (white shading; Culberg et al., 2021) detections along OIB*

1974 *flight lines (black interior lines); AR transect A-B (red line; Fig. 3a), and C-D (orange line; Fig. 3b); and*

1975 *SMAP Test Site A (blue circle; Fig. 4a), B (cyan circle; Fig. 4b), C (orange diamond; Fig. 4c), D (red triangle;*

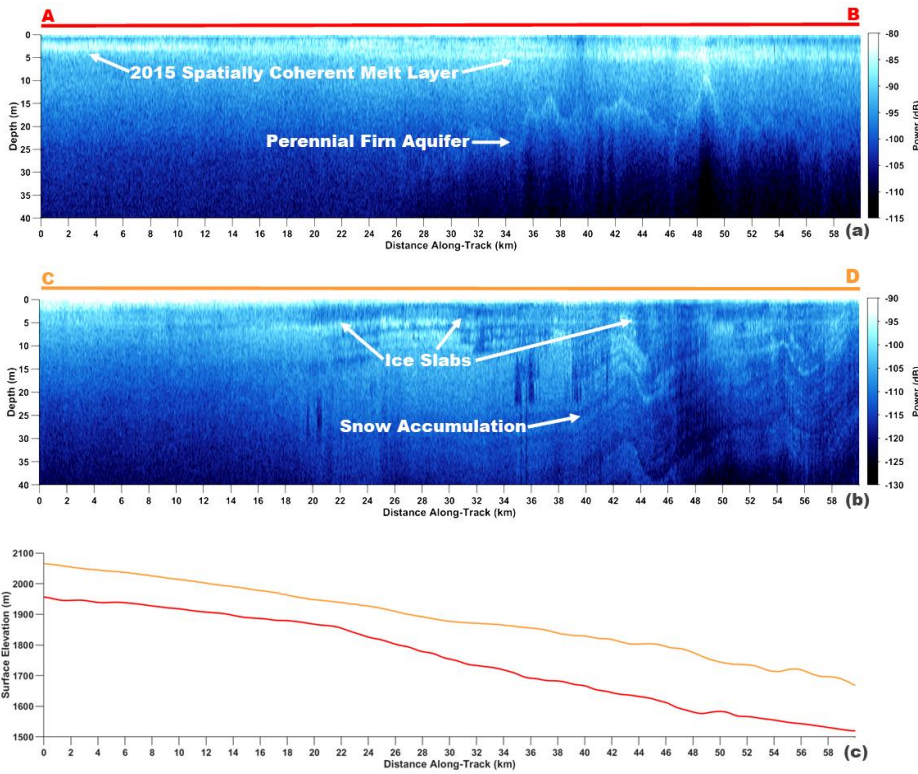
1976 *Fig. 4d), and E (yellow circle; Fig. 4e).*

1977

Figure 3

1978 **Figure 3**

1979 *AR transect radargram profile (Rodriguez-Morales et al, 2014) (a) along perennial firn aquifer transect A-B*

1980 *(red line, Figs. 1; 2a) that was collected on 22 April 2017, and and (b) ice slab transect C-D (orange line,*

1981 *Fig.s. 1; 2b) that was collected on 5 May 2017 (Rodriguez-Morales et al, 2014). (c) The corresponding*

1982 *perennial firn aquifer AR transect A-B elevation profile (red line), an and d ice slab transect C-D (orangle*

1983 *line) elevation profiles (orange line). The eThe exceptionally bright upper surface-parallel reflector in (a) is*

1984 *interpreted as a spatially coherent melt layer. The that formed following the 2015 melting season. The*

1985 *bright lower reflector in (a) is the interpreted as the upper surface of meltwater stored within a perennial firn*

1986 *aquifer. The t Thick dark surface-parallel regions of low-reflectivity in (b) are interpreted asare ice slabs.*

1987 *The aAlternating sequences of bright and dark surface-parallel reflectors in (b) are interpreted as*

1988 *seasonal snow accumulation layers. A first maximum after maximum gradient re-tracker is used to identify*

1989 *the surface return in each profile. Each profile is flattened so that the depth axis is measured relative to the*

1990 *local elevation. Corresponding elevation profiles in (c) are calculated by subtracting the radar-measured*

1991 *flight clearance over the ice sheet from the aircraft's global positioning system altitude measurements that*

1992 *were coincidently collected along each transect.*

1993

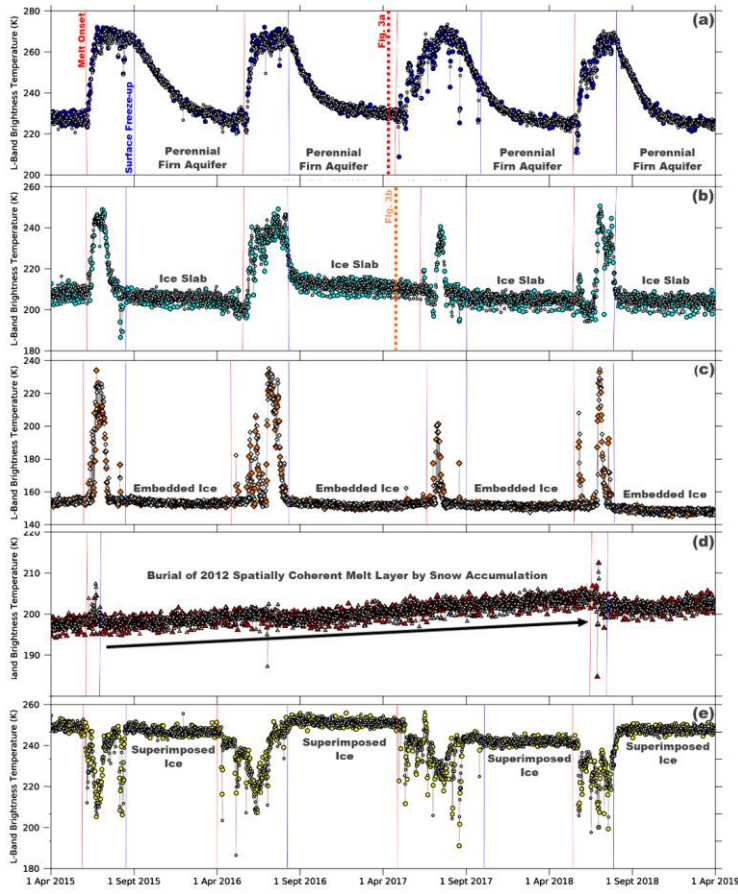



Figure 4

Temporal L-band signatures that alternate morning (white symbols) and evening (colored symbols) orbital pass interval enhanced-resolution T_V^B generated using observations collected over the GrIS by the microwave radiometer on the SMAP satellite (~~Brodzik et al~~Long et al., 20~~19~~19) over (a) ~~perennial firn aquifer~~SMAP Test Site A ~~area~~(blue circles; Figs. 2a; 3a), (b) ~~perched firn aquifer~~ B ~~area~~(cyan circles; ; Figs. 2b; 3b), (c) ~~percolation facies area~~C (orange diamonds; Fig 2b), (d) high-elevation (~2500 m.a.s.l.) ~~spatially coherent melt layer~~D ~~area~~(red triangles; Fig. 2a), and (e) ~~superimposed ice~~E ~~area~~(yellow circles; Fig. 2b). Melt onset (red lines) and surface freeze-up (blue lines) dates ~~are~~ derived from thermal infrared T^B collected by MODIS on the Terra and Aqua satellites (Hall et al, 2012). AR ~~radargram profile along perennial firn aquifer~~transect A-B (red dashed line;; Figs. 1; 2a; 3a) ~~that was~~ collected on 22 April 2017, and ~~ice slab transect~~ C-D (orange dashed line;; Fig. Figs. 1; 2b; 3b) ~~that was~~ collected on 5 May 2017.

2005

2006 **Figure 5**

2007 *Theoretical L-band penetration depths for of uniform layer of (a) refrozen, and (b) water-saturated firn.*

2008 *Penetration depths $\left(\frac{1}{\kappa_s + \kappa_a}\right)$ are calculated as a function of the Raleigh scattering coefficient (κ_s; Eq. 8), and*

2009 *the absorption coefficient (κ_a; Eq. 10)), which are functions of the dielectric and geophysical properties of*

2010 *the GrIS. The complex dielectric constant is calculated using the empirically derived models described in*

2011 *Tiuri et al., (1984). Refrozen firn penetration depths are calculated as a function of firn density (ρ_{firn}), and*

2012 *the curves are plotted for snow grain radii (r) set to r=0.5 mm (upper curve), and r=4 mm (lower curve).*

2013 *Water-saturated firn penetration depths are calculated as a function of the volumetric fraction of meltwater*

2014 *(m_v), and the curves are plotted for firn density set to ρ_{firn}=400 kg/m³ (upper curve), and ρ_{firn}=917 kg/m³*

2015 *(lower curve). Given the complexity of modeling embedded ice structures, they are excluded from the*

2016 *penetration depth calculation. Increases in the volumetric fraction of embedded ice within the firn will result*

2017   *in an increase in volume scattering, which will decrease and compress the distance between the penetration*
2018   *depth curves for both refrozen and water-saturated firn. Given the complexity of modeling embedded ice*
2019   *structures, they are excluded from the penetration depth calculation. Increases in the volumetric fraction of*
2020   *embedded ice in the firn will result in an increase in volume scattering, which will decrease and compress*
2021   *the distance between the penetration depth curves for both refrozen and water-saturated firn.*

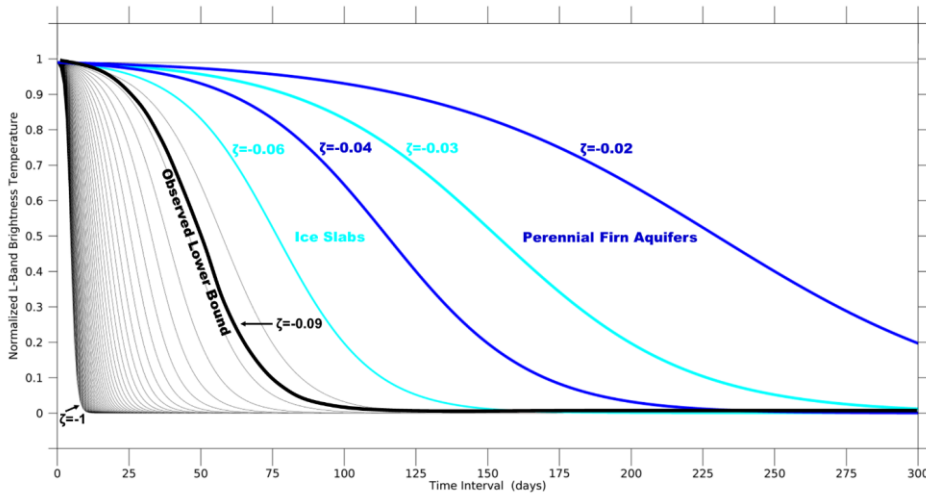
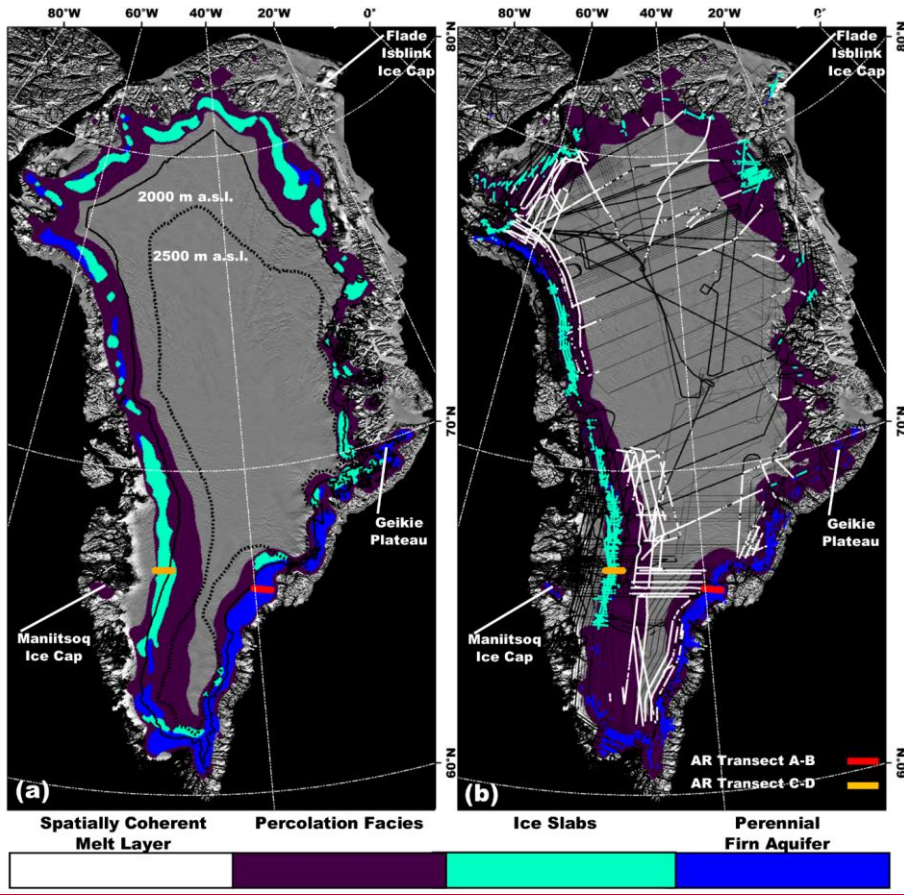



Figure 6

Example set of simulated sigmoidal curves that represent our model of the exponentially decreasing temporal L-band signatures predicted over the percolation facies. The initial normalized vertically-polarized L-band brightness temperature was fixed at a value of $T^{B}_{V,N}(t_{max})$ = 0.99, and the time interval was set to a value of $t \in [\,t_{max},t_{min}\,]$ = 300 observations. The refreezing rate parameter was set to values between $\zeta = [\,\text{-1, 0}\,]$ incremented by steps of 0.02. The blue lines correspond to the interval $\zeta \in [\,\text{-0.04, -0.02}\,]$ and produce curves similar to those observed over perennial firn aquifer areas. The cyan lines correspond to the interval $\zeta \in [\text{-0.06, -0.03}\,]$ and produce curves similar to those observed over ice slab areas. ~~The green lines correspond to the interval $\zeta \in [\,\text{-0.04 ,-0.03}\,]$ and produce curves similar to those observed over perched firn aquifer areas.~~ The black line is the observed lower bound ($\zeta = $ -0.09) of the refreezing rate parameter of partitioned T^{B}_{V} time series ~~(1 April 2015 – 31 March 2019)~~ iteratively fit to the sigmoid function (~~see~~ Section 2.3.43).

2039

2040

2041 **Figure 7**

2042 *(a) SMAP-derived perennial firn aquifer (blue shading), ice slab (cyan shading), and percolation facies*
2043 *(purple shading) extents (2015-2019) generated by the adapted empirical algorithm; and the 2000 m a.s.l.*
2044 *contour (black line), and the 2500 m a.s.l. contour (black dotted line; Howat et al., 2014) overlaid on the*
2045 *2015 MODIS Mosaic of Greenland (MOG) image map (Haran et al., 2018). (b) SMAP-derived extents are*
2046 *overlaid with AR- and MCoRDS-derived 2010-2017 perennial firn aquifer (blue shading; Miège et al., 2016),*
2047 *2010-2014 ice slab (cyan shading; MacFerrin et al., 2019), and 2012 spatially coherent melt layer (white*
2048 *shading; Culberg et al., 2021) detections along OIB flight lines (black interior lines); and AR transect A-B*
2049 *(red line; Fig. 3a), and C-D (orange line; Fig. 3b).*

2051 **Figure 8**

2052 *The SMAP-derived perennial firn aquifer (blue shading), ice slab (cyan shading), and percolation facies*
2053 *(purple shading) extents (2015-2019) generated by the adapted empirical algorithm over south eastern*
2054 *Greenland (red box; Fig. 1c); and the 2000 m a.s.l. contour (black line), and the 2500 m a.s.l. contour (black*
2055 *dotted line; Howat et al., 2014) overlaid on the 2015 MODIS MOG image map (Haran et al., 2018). (b) The*
2056 *SMAP-derived percolation facies extent is overlaid with AR- and MCoRDS-derived 2010-2017 perennial*
2057 *firn aquifer (blue shading; Miège et al., 2016), 2010-2014 ice slab (cyan shading; MacFerrin et al., 2019),*
2058 *and 2012 spatially coherent melt layer (white shading; Culberg et al., 2021) detections along OIB flight lines*
2059 *(black lines); AR transect A-B (red line; Fig. 3a); and SMAP Test Site A (blue circle; Fig. 4a), and D (red*
2060 *triangle; Fig 4d).*

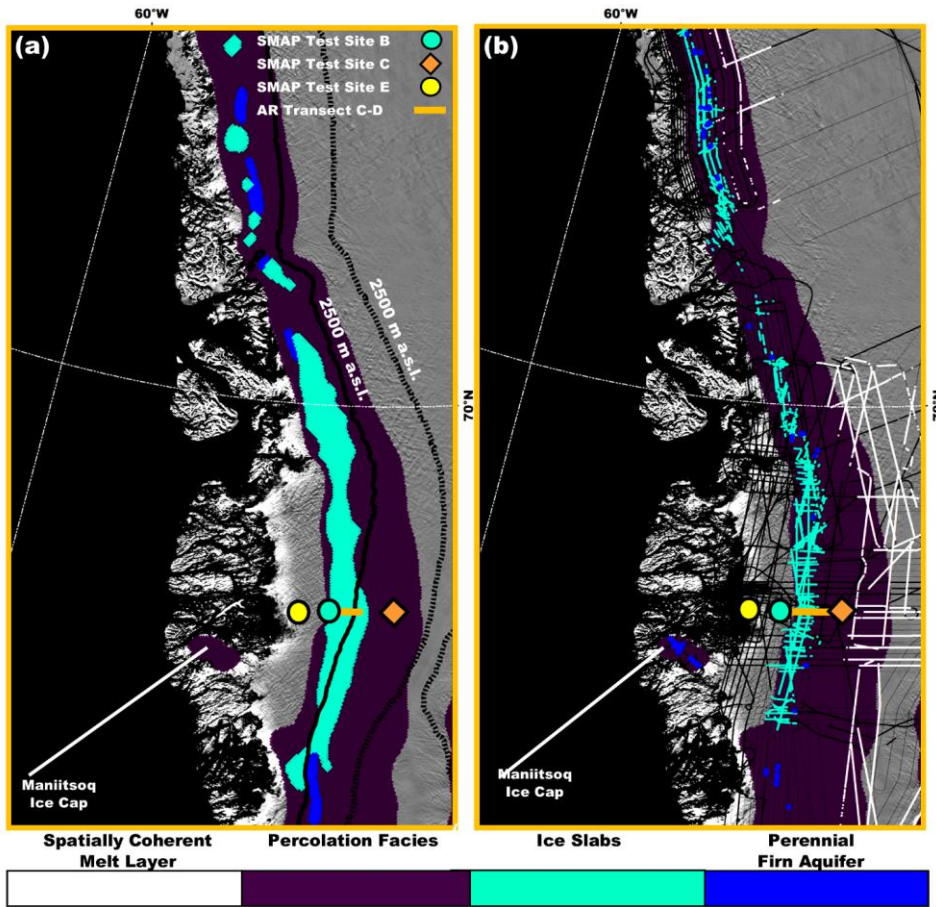



Figure 9

(a) SMAP-derived perennial firn aquifer (blue shading), ice slab (cyan shading), and percolation facies (purple shading) extents (2015-2019) generated by the adapted empirical algorithm over south western Greenland (orange box; Fig. 1c); and the 2000 m a.s.l. contour (black line), and the 2500 m a.s.l. contour (black dotted line; Howat et al., 2014) overlaid on the 2015 MODIS MOG image map (Haran et al., 2018). (b) SMAP-derived percolation facies extent is overlaid with AR- and MCoRDS-derived 2010-2017 perennial firn aquifer (blue shading; Miège et al., 2016), 2010-2014 ice slab (cyan shading; MacFerrin et al., 2019), and 2012 spatially coherent melt layer (white shading; Culberg et al., 2021) detections along OIB flight lines (black interior lines); AR transect C-D (orange line; Fig. 3b); and SMAP Test Site B (cyan circle; Fig. 4b), C (orange diamond; Fig. 4c), and E (yellow circle; Fig. 4e).

2073 **Table 1.**

2074 *MODIS-derived total number of days in the melting and freezing seasons; SMAP-derived maximum*

2075 *vertically-polarized L-band brightness temperature ($T^B_{V,max}$); minimum vertically-polarized L-band brightness*

2076 *temperature ($T^B_{V,min}$); time scale scales of exponential decrease following the surface freeze-up date for*

2077 *perennial firn aquifer, ice slab, percolation facies, dry snow facies, and wet snow facies areas.*

	Melting Season (days)	Freezing Season (days)	$T^B_{V,max}$ (K)	$T^B_{V,min}$ (K)	Exponential Decrease (time scale)
Perennial Firn Aquifers	75 - 100	265 - 290	200 - 275	180 – 250	weeks – months
Ice Slabs	60 -90	275 - 305	170 - 260	130 – 240	days - weeks
Percolation Facies	1 - 60	305 - 364	150 - 200	130 – 220	days
Dry Snow Facies	-	365	200 - 240	200 – 240	-
Wet Snow Facies	90 - 120	245 - 275	230 - 250	230 – 250	-

2078

Table 23.

SMAP-derived calibration parameter threshold intervals (1 April 2015 - 31 March 2019) used for mapping perennial firn aquifer and, ice slab, and perched firn aquifer extents areas.

	ξ	$T^B_{V,max}$ (K)	$T^B_{V,min}$ (K)	ξ
Perennial Firn Aquifers	0.2 – 4	200 – 275	180 – 250	-0.04 - -0.02
Ice Slabs	0.1 – 2	170 – 260	130 – 240	-0.03 - -0.06
Perched Firn Aquifers	0.2 - 1.2	200 – 260	180 – 240	-0.03 - -0.04
	ξ	$T^B_{V,max}$ (K)	$T^B_{V,min}$ (K)	ξ
Perennial Firn Aquifers	0.2 – 4	200 – 275	180 – 250	-0.04 – -0.02
Ice Slabs	0.1 – 2	170 – 260	130 – 240	-0.06 – -0.03

Table 4. *The SMAP-derived perennial firn aquifer, ice slab, and perched firn aquifer extents (2015-2019) over the percolation facies and the GrIS, and the elevation range at which they are mapped.*

	Percolation Facies Extent (%)	Ice Sheet Extent (%)	Elevation Range (m.a.s.l.)
Perennial Firn Aquifers	11	4	600 – 2600
Ice Slabs	13	4	800 – 2700
Perched Firn Aquifers	5	≤1	600 – 2700

2088 **Table 3.**

2089 *Interannual variability in SMAP-derived perennial firn aquifer and ice slab extents.*

	Perennial Firn Aquifers (km²)	Ice Slabs (km²)
2015-2019	66,000	76,000
2015-2016	63,000	23,000
2016-2017	69,000	48,000
2017-2018	73,000	27,000
2018-2019	70,000	38,000

2090