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Key Points:
• We present a method to map Greenland firm aquifers using Sentinel-1 for the first time.
• The spaceborne aquifer detection agrees well with that of airborne Operation IceBridge.
• The firm aquifer area is estimated at 54,800 km² across the Greenland ice sheet.

Supporting Information:
• Supporting Information SI
• Data Set SI
• Figure S1

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Abstract Fim aquifers in Greenland store liquid water within the upper ice sheet and impact the hydrological system. Their location and area have been estimated with airborne radar sounder surveys (Operation IceBridge, OIB). However, the OIB coverage is limited to narrow flight lines, offering an incomplete view. Here, we show the ability of satellite radar measurements from Sentinel-1 to map fim aquifers across all of Greenland at 1 km² resolution. The detection of aquifers relies on a delay in the freezing of meltwater within the firm above the water table, causing a distinctive pattern in the radar backscatter. The Sentinel-1 aquifer locations are in very good agreement with those detected along the OIB flight lines (Cohen’s $\kappa = 0.84$). The total aquifer area is estimated at 54,800 km². With continuity of Sentinel-1 ensured until 2030, our study lays a foundation for monitoring the future response of fim aquifers to climate change.

Plain Language Summary The Greenland ice sheet encloses vast amounts of liquid water beneath its surface, within a layer of old compacted snow referred to as firm. These water reservoirs, named fim aquifers, are a peculiar feature with potentially widespread impacts on ice sheet temperature, hydrology, and contribution to sea level. Monitoring aquifer areas is therefore critical in assessing the vulnerability of the Greenland ice sheet to climate change. Current knowledge on aquifer locations relies on aircraft flights of NASA Operation IceBridge, surveying only narrow stretches of the ice sheet along its flight lines. Satellite observations are necessary to provide full coverage. Here, we present the first satellite-based map of Greenland’s fim aquifers, using radar observations from the Sentinel-1 constellation. The detection with Sentinel-1 relies on a strong and prolonged absorption of the emitted radar signal by the liquid water within the firm, yielding low values in the measured return signal. Very good agreement is found when comparing Sentinel-1 aquifer locations with those of Operation IceBridge. The total aquifer area estimated by Sentinel-1 is 54,800 km². The long-term continuity provided by the satellite observations offers great potential for the monitoring of changes in aquifer location and area related to climate change.

1. Introduction

In some areas, the margins of the Greenland ice sheet (GrIS) contain water reservoirs within their firm layer (partially compacted snow from previous years), which can persist in the liquid phase year-round (Forster et al., 2014; Miège et al., 2016). This phenomenon, referred to as firm aquifers, was first discovered in 2011 by analyzing firm cores collected in southeast Greenland during the Arctic Circle Traverse expedition (Forster et al., 2014). Likely, a firm aquifer can be sustained year-round when meltwater produced at the surface of the ice sheet percolates into the firm layer and is insulated from winter freezing temperatures by sufficient snowfall (Kuipers Munneke et al., 2014). Therefore, firm aquifers are typically found in areas subject to moderate or strong surface melt during the summer and high snow accumulation during the autumn through early spring (Forster et al., 2014). The depth to the aquifer water table typically ranges from 5 to 50 m. Most aquifers are found in the wet snow zone of the southeastern and southern periphery of the GrIS and in small patches (~5% of the total aquifer extent) of northwestern Greenland (Miège et al., 2016).

Firm aquifers affect the thermal state of the underlying ice, play a role in the ice sheet mass balance, and might impact ice dynamics via hydrofracturing. The freezing of liquid water releases latent heat that warms...
the surrounding firn and ice, thereby delaying further freezing (Forster et al., 2014; Humphrey et al., 2012; Phillips et al., 2010). Freezing in the shallow firn layer will also limit water access to the deeper firn and increase lateral runoff (Machguth et al., 2016; van Angelen et al., 2013). Mainly through hydrofracture in crevassed areas, aquifers can cause liquid water to flow to the bed of the ice sheet, providing a potential mechanism for aquifer drainage and impacting ice dynamics (Fountain & Walder, 1998; Nienow et al., 2017; Poinar et al., 2017). For example, water at the base of a glacier bed can decrease friction and increase glacier sliding velocity (Fountain & Walder, 1998). The storage in, or release of, meltwater from the aquifer into the englacial hydrology system can modify Greenland’s contribution to sea level rise (e.g., Chu et al., 2018; Flowers, 2018; Harper et al., 2012; Koenig et al., 2014, and others).

Presently, mapping firn aquifers relies on airborne radar sounder measurements from NASA’s Operation IceBridge (OIB). Combining flight lines from several years across the GrIS and assuming temporal stability of the aquifers, Miège et al. (2016) estimated a total aquifer area of \( \sim 21,900 \, \text{km}^2 \). Alternatively, Forster et al. (2014) estimated the Greenland firn aquifer area at \( \sim 70,000 \, \text{km}^2 \) based on model simulations with requirements of sufficient surface melt and high snow accumulation rates. Steger et al. (2017) obtained area estimates of 55,700 and 90,200 km\(^2\) based on two firn models with different complexities. While these airborne and modeling approaches provide essential information on firn aquifer area, they have shortcomings. Airborne observations provide only coverage directly below the flight lines, typically once per year. Model simulations are often coarse-scale (\( \sim 10–20 \, \text{km} \) resolution) and prone to uncertainties in their meteorological forcings. Therefore, the available estimates of the firn aquifer area on the GrIS could benefit from further constraints. Whereas some satellite observations hold promise to indirectly constrain firn aquifer area estimates, there are currently no satellite-based observations of the aquifer area itself.

Here, we demonstrate the ability of Sentinel-1 (S1) synthetic aperture radar (SAR) measurements to detect firn aquifers and present a firn aquifer mosaic over Greenland for 2014–2019. The detection mechanism relies on the strong absorption of the S1 radar signal by liquid water, potentially at the aquifer water table (when close to the surface; e.g., less than 10 m), but more likely within the firn layer above it. The limited penetration depth at C-band does not always allow sensing the top of the water table itself, which on average is located 22 m below the surface (Miège et al., 2016). However, the time-delayed increase in radar backscatter caused by the delayed refreezing of water in the firn at aquifer locations yields a unique signature in the S1 backscatter data. We compare aquifer locations derived from S1 with those of OIB, and provide a new estimate of the aquifer area across the GrIS. The long-term continuity in C-band SAR observations with the S1 and RADARSAT constellation missions warrants the monitoring of firn aquifers at the decadal time scale. The mapping of firn aquifers from S1 may contribute to a better understanding of melt dynamics (including storage) over time and space.

2. Materials and Methods

2.1. S1 Data

We processed C-band (5.4 GHz) radar backscatter measurements (\( \sigma^0; \) in dB) from the S1 constellation (two satellites: 1A and 1B) obtained in extra-wide swath mode, for the period October 2014, that is, the start of the Sentinel-1A measurements, through June 2019. Over polar regions, measurements in this mode are collected in dual polarization, that is, horizontal-horizontal (HH) and horizontal-vertical (HV) transmit-receive, respectively, within a 410-km swath width at 20 by 40 m spatial resolution. The preprocessing of the data was performed using the Google Earth Engine Python application programming interface (Gorelick et al., 2017). Thereby, standard S1 processing steps were applied, including border noise removal, thermal noise removal, radiometric calibration and range-Doppler terrain correction. The data were aggregated (by averaging in linear scale) and projected onto the 1-km\(^2\) resolution global cylindrical Equal-Area Scalable Earth Grid version 2 (EASE-2) (Brodzik et al., 2012). The aggregation to 1 km\(^2\) (considered sufficient for an initial mapping of firn aquifers) improves the signal-to-noise ratio, which is typically low in cross-polarized \( \sigma_{HV}^0 \) measurements. Future research shall investigate the mapping at higher spatial resolutions. Each of the Sentinel-1A and Sentinel-1B satellites has an exact 12-day repeat cycle, with 175 orbits per cycle. The impact of varying incidence and azimuth angles on \( \sigma^0 \) (e.g., between ascending and descending orbits) was corrected by removing the static mean bias between each of the orbits. For each copolarized and cross-polarized S1 data set separately, four subsets of S1 data (i.e., ascending (6 p.m.) and descending (6 a.m.) data from Sentinel-1A and Sentinel-1B) were preprocessed separately and combined (by averaging in linear scale) daily. Both S1 satellites share the same orbital plane with a 6-day offset, so that the two-satellite
constellation offers an exact 6-day repeat cycle. However, the observation frequency during the period considered (October 2014 through June 2019) varies from ~daily close to the GrIS margins, where most firn aquifers are located, to about every 2 weeks in the interior. The frequency primarily depends on the latitude and the (evolving) S1 observation acquisition strategy, which prioritizes the GrIS margins for the monitoring of outlet glaciers.

2.2. S1 Signatures of GrIS Zones

Figure 1a shows the temporal mean $\sigma^0_{HV}$ (dB) from S1 over Greenland. Similar results were obtained for $\sigma^0_{HH}$. The spatial patterns in the $\sigma^0_{HV}$ measurements clearly differentiate between ice sheet zones in response to their seasonal melt dynamics (Benson, 1962) and confirm previous studies (e.g., Bindschadler & Vornberger, 1992; Drinkwater et al., 2001; Fahnestock et al., 1993; Jezek et al., 1993; Joughin et al., 2016; König et al., 2001; Wismann, 2000, and others), which mostly focused on copolarization (VV or HH) measurements from Ku- to L-band frequencies. At high elevations in the interior of the ice sheet, the dry snow zone lacks regular seasonal melt. This causes a gradual vertical transition from snow to ice and thus a low reflection of the radar signal ($\sigma^0_{HV}$ mostly below $-12.5$ dB at C-band). The percolation zone features occasional meltwater that refreezes and forms ice lenses and pipes within the firn (Benson, 1962). These refrozen percolation structures cause a high reflectivity ($\sigma^0_{HV}$ often above $-5$ dB) and dominate the signal in one of the strongest scattering regions on Earth (Long & Drinkwater, 1994; Rignot, 1995; Nghiem et al., 2005). At lower elevations of the GrIS, the wet snow zone experiences more intense seasonal melt that saturates the firn layer. When the wet layer refreezes, a higher density layer forms (Long & Drinkwater, 1994). The higher density, and thus stronger dielectric contrast at the air-snow interface, can reduce the radar signal penetration depth (Fahnestock et al., 1993; Jezek et al., 1993). While the melting here is more pronounced, the more uniform refreezing produces less refrozen ice structures and causes relatively lower $\sigma^0_{HV}$ ($\sim -10$ dB) compared to the percolation zone (Joughin et al., 2016). Finally, at the lowest elevations in the bare ice zone, the entire snowpack that accumulated over the winter disappears during summer melt. Consequently, bare ice is exposed at the surface during summer, allowing little volume scattering and reflecting most of the signal away from the radar (Bindschadler & Vornberger, 1992; Fahnestock et al., 1993; König et al., 2001), causing low $\sigma^0_{HV}$ ($< -10$ dB).

2.3. Firn Aquifer Detection and Mapping

Liquid water within the snow and firn layers absorbs and reflects most of the radar signal at C-band and decreases the radar signal penetration depth, causing low $\sigma^0$. Consequently, melt events can easily be detected in time series as strong decreases in $\sigma^0$ (e.g., Fahnestock et al., 1993; Joughin et al., 2016, and others). Maps of time differences in $\sigma^0_{HV}$ are shown in Figures 1b and 1c, and $\sigma^0_{HV}$ time series at select locations within the wet snow zone are shown in Figure 2. Firn aquifers are typically found in the wet snow zone, where strong summer melt saturates the firn with water and causes a low summer $\sigma^0$. 
Figure 2. Time series of S1 backscatter ($\sigma_0^{HV}$; dB) for three representative aquifer and nearby nonaquifer grid cells, for the period from October 2014 to June 2019, in northwest, southeast, and south Greenland. The locations are shown in Figure 3 and correspond to latitude/longitude (a) 76.51°N/−61.75°E and 76.61°N/−61.28°E, (b) 65.30°N/−41.65°E and 65.51°N/−41.98°E, and (c) 61.77°N/−46.38°E and 62.03°N/−46.27°E for aquifer and nonaquifer locations, respectively.

Figure 1b shows the reduction in $\sigma_0^{HV}$ for July (when melt is ongoing) relative to April (when dry snow accumulation is close to the maximum and melt is mostly absent), highlighting the presence of liquid water in the wet snow zone. The decrease in liquid water content due to refreezing increases $\sigma_0^{HV}$, as the signal reaches deeper into the firn layer and the contribution of snow and firn volume scattering increases. In most regions of the wet snow zone, the firn has refrozen by the end of August (Figures 1c and 2). However, in aquifer locations, the refreezing of liquid water in the firn is typically delayed due to the insulating effect of autumn snow accumulation, and the substantial amounts of latent heat released upon refreezing (Forster et al., 2014; Phillips et al., 2010; Winsvold et al., 2018). This causes aquifer locations to show a marked delay in the increase of $\sigma_0^{HV}$ throughout autumn (Figures 1c and 2). An important hypothesis here is that the radar will likely not directly sense the water table (only for shallow perched water tables). Instead, it is likely that the slowdown in refreezing of water in the upper profile (above the water table) provides a distinct signature and serves as a proxy for the detection of the aquifers. Furthermore, aquifer locations typically correspond to areas that receive significant amounts of snowfall (Forster et al., 2014). This high accumulation likely causes another (more subtle) increase in $\sigma_0^{HV}$ over winter, particularly in the cross-polarized measurements (Lievens et al., 2019). The strong contrast between the low autumn $\sigma_0^{HV}$ and the high end-of-winter $\sigma_0^{HV}$ offers a mechanism for firn aquifer detection.

To map firn aquifers, we apply the following criterion: the reduction in $\sigma_0^{HV}$ of the early autumn (i.e., September, when refreezing is delayed) relative to that near the end of winter (i.e., April, when dry
snow accumulation is close to the maximum) needs to exceed a difference threshold of 9.4 dB. This criterion is based on the observation that the early autumn $\sigma_0$HV is typically low in aquifer locations (due to persistence of liquid water) and relatively higher in nonaquifer locations (due to earlier refreezing). The firn aquifer mapping was limited to the GrIS, using a resampled 1-km² version of the Greenland Ice Mapping Project ice sheet mask (Howat et al., 2014). The threshold value of 9.4 dB was defined by maximizing the Cohen’s Kappa coefficient (Cohen, 1960) between the binary (0/1 for absence/presence) firn aquifer locations derived from S1 and OIB (section 2.4). The calibration was done with a randomly sampled half subset of the OIB data set, whereas the other half of the data set was used for the validation. The same procedure was tested with HH-polarized data. A similar but slightly less pronounced $\sigma_0$HH signal was observed which typically yielded lower $\kappa$ values (see section 3).

In the present study, a single map of Greenland firn aquifers is obtained from ~5 years of S1 data. That is, the thresholding was applied to multiyear monthly averages of $\sigma_0$HV. There are two main reasons to support this. First, in some years under certain meteorological circumstances, the delay in the increase of $\sigma_0$HV can be less pronounced (discussed in section 3). Second, the OIB aquifer detection (section 2.4) was similarly pooled together to produce a single aquifer map, based on flights from several subsequent years, assuming aquifers did not migrate or drain over time (Miège et al., 2016).

2.4. OIB Aquifer Detection
Firn aquifer locations were derived from airborne radars built and operated by the Center of Remote Sensing for Ice Sheets on board the NASA OIB aircraft (Rodriguez-Morales et al., 2014). Up to four radars operate simultaneously with different frequency ranges for different applications, from estimating snow depth over sea ice to ice sheet thickness (Rodriguez-Morales et al., 2014). The aquifer detections in spring 2010–2014 and 2017 were done using the Accumulation Radar (our preferred option), with a center frequency of 750 MHz and frequency bandwidth from 565 to 885 MHz. This radar system is used to generate echograms with a vertical sample interval (referred to as bin size) of ~0.3 m in firn (Lewis et al., 2015; Paden et al., 2014a). The surveys in spring 2015 and 2016 did not have the Accumulation Radar on board, and we instead used the Multichannel Coherent Radar Depth Sounder (MCoRDS) (Paden et al., 2014b). In 2015, MCoRDS operated at two alternating frequency bandwidths: a narrow bandwidth of 180–230 MHz (with bin size of 1.94 m), and a wide bandwidth of 180–450 MHz (improving the bin size to 0.35 m). Supplementary Figure S1 illustrates the impact of the different bandwidths for the same north-south transect along the southeast region of the GrIS, taken 15 days apart (on 13 and 28 April 2015, respectively). Firn aquifer water tables were better detected in wideband mode (which is more similar to the Accumulation Radar). In spring 2016, OIB measurements were similar to 2015 using both a narrow frequency bandwidth of 150–250 MHz (bin size of 0.96 m) and wide frequency bandwidth of 150–450 MHz (bin size of 0.32 m).

The firn aquifer locations were obtained for each spring using the same methodology as in Miège et al. (2016). We merged the available firm aquifer data for 2010–2014 (total aquifer area of 21,900 km²; Miège et al., 2016) with the more recent data from 2015–2017. The 2010–2017 polygon extents increased the total area to 29,268 km². This corresponds to 28,271 grid cells on the 1 km EASE-2 grid after reprojecting and limiting the data to the Greenland Ice Mapping Project ice sheet mask. Miège et al. (2016) assessed the uncertainty in the detection by processing the spring 2011 data set twice by two different operators, with 75% agreement. We anticipate similar levels of agreement for other years. However, the lower vertical resolution in the MCoRDS narrow bandwidth data poses additional challenges to the water table detection and may increase the uncertainty near the aquifer edges.

3. Results and Discussion
Figure 3 shows a mosaic of Greenland’s firn aquifers derived from S1 data of 2014–2019 and compares it with the aquifer locations obtained from OIB data of 2010–2017. Along the OIB flight lines, the S1 and OIB aquifer locations show strong agreement. The $\kappa$ indices equal 0.84 (with 95% confidence interval of ±0.004) for both the calibration and validation data sets. About 87% of the OIB aquifer locations (24,787 out of 28,271 grid cells of 1 km²) were correctly identified as aquifer by S1 (i.e., the true positive rate). Among locations that are not aquifer according to OIB, 2% (4,534 out of 209,216 grid cells of 1 km²) are identified as aquifer by S1 (i.e., the false positive rate). Part of the errors may be caused by the different sampling periods of S1 (2014–2019) and OIB (2010–2017). Our validation assumes steady-state behavior of firn aquifers and does not account for transient behavior such as the potential emergence, migration or drainage of aquifers over
Figure 3. (a) Aquifer locations according to S1 and OIB (1 = present, 0 = absent). (b) Enlargement over the south. (c) Enlargement over the northwest. (d) Enlargement over the southeast. Locations of the time series in Figure 2 are shown for aquifers (blue dots) and nonaquifers (red dots). The ice sheet mask is a resampled version (to 1 km²) of the Greenland Ice Mapping Project (Howat et al., 2014).

time observed in ground/airborne data (Miège et al., 2016). Future research based on additional OIB data including recent years will focus on investigating the temporal dynamics.

The total firn aquifer area obtained from S1 equals 55,444 km². That is substantially larger than the original OIB estimate of ~21,900 km² (Miège et al., 2016) or the herein updated estimate of ~29,268 km² and approximates the lower limit of model simulations ranging from 55,700–90,200 km² (Forster et al., 2014; Steger et al., 2017). The sensitivity of the S1 aquifer detection to the difference threshold value of 9.4 dB was investigated by modifying the latter by [−10%; +10%]. The corresponding values of $\kappa$ (for the entire calibration and validation data set) are [0.83; 0.83], and the total aquifer area ranges between [62,383 km²; 48,410 km²]. The present S1 aquifer detection uses $\sigma^0$ measurements in HV polarization. For reference, we also tested the detection using measurements in copolarization (HH), which are more abundantly available over Greenland from S1 and also from other (previous and current) SAR missions, such as ENVISAT (Environmental Satellite) and the RADARSAT Constellation. The use of HH polarization led to a slightly lower accuracy, with $\kappa = 0.82$ for the combined data set and an aquifer area of 63,133 km² (data not shown).

Uncertainties in the OIB data are caused by the manual identification of the aquifers, increasing the chances for misclassification such as (most likely) an underdetection near the aquifer edges (Miège et al., 2016). Similarly, the detection of aquifers with deeper water tables or those holding a small volume of water (less
than 200 kg/m$^3$ when integrated over the entire firn column) is more difficult, due to a stronger signal attenuation with depth, or due to the absence of a sharp dielectric contrast at the water table. In some areas, detection from OIB is challenging because of strong reflections at the ice sheet surface such as surface clutter due to crevasses. Further, the quality of the OIB aquifer detection depends on the airplane survey conditions, such as turbulence and turning geometry. For instance, Miège et al. (2016) mention that aircraft roll greater than 10$^\circ$ hampers the detection of internal layers and thus aquifers, at least when using Accumulation Radar.

Likewise, the S1 aquifer detection is prone to a number of uncertainties. The S1 mosaic may potentially include areas with nonperennial liquid water, that is, where liquid water is still present in September but refreezes afterward. Confusion may also occur in regions exposed to bare ice in summer (ablation zone), including glaciers. Bare ice typically shows low summer $\sigma_0^{HV}$ due to the often smooth ice surface. The low $\sigma_0^{HV}$ can be prolonged through early autumn (e.g., September) in case of limited snowfall. Subject to significant snowfall after the end of September, $\sigma_0^{HV}$ can increase and reveal a similar temporal evolution as in aquifer locations. When excluding glaciers (according to the Randolph Glacier Inventory 6.0 RGI Consortium, 2017) from the S1 aquifer mapping, $\kappa$ increases from 0.84 to 0.86; in contrast, when only considering glaciers, $\kappa$ decreases to 0.47. However, the latter accuracy estimate is more uncertain due to the significantly lower number of aquifer grid cells in the OIB data set (1,218 over glaciers, instead of 28,287 for the full GrIS).

Similarly, when the bare ice region was excluded using the snowline product from Fausto and the PROMICE team (2018), the overestimation of aquifers in the bare ice zone is corrected and the $\kappa$ increases to 0.86 with a total aquifer area of 54,803 km$^2$. We expect this last area estimate to be the most accurate.

An important caveat to the S1 aquifer detection is that we here rely on multiyear average measurements and the assumption that aquifer locations are stable in time. Although the method has potential for monitoring temporal changes in aquifer presence and area, specific meteorological conditions may preclude the detection from a single year of S1 data. For instance, freezing temperatures in September with absence of significant dry snow accumulation may cause important amounts of meltwater to refreeze, increasing the September $\sigma_0^{HV}$. Limited summer melt leading to only small amounts of firm meltwater may also cause an insufficient reduction in September $\sigma_0^{HV}$ and a faster firn refreezing for the algorithm to detect a potential aquifer. As shown on the time series in Figure 2, aquifer locations in autumn 2015 and 2018 (Figure 2a, in the northwest) and autumn 2016 (Figure 2c, in the south) are more difficult to discern from nonaquifer locations, due to an earlier increase in $\sigma_0^{HV}$ that is likely caused by either of the above mentioned conditions. Conversely, nonaquifer areas with low surface melt production in September may reveal low $\sigma_0^{HV}$ and could therefore be falsely detected as aquifer by the S1 algorithm. As for OIB data, the presence of strong scatterers such as ice lenses, creating a heterogeneous medium above the aquifer, can lead to high September $\sigma_0^{HV}$ and reduce the discernibility from nonaquifer areas. However, it should be noted that the S1 algorithm is arguably less sensitive to these conditions than the OIB algorithm, as the former is driven by the change in $\sigma_0^{HV}$ over time (from September to April) rather than by the $\sigma_0^{HV}$ magnitude at a single instant in time. Finally, the relationship between the aquifer water table depth and the performance of the S1 aquifer mapping is currently not addressed and requires further research.

4. Conclusions

This study maps the extent of firn aquifers within the GrIS using S1 satellite radar observations. Aquifer areas show a unique radar backscatter signature of prolonged low backscatter values in the fall. This is caused by a time delay in the freezing of the water content within the firn layer, mainly due to the substantial amounts of fall/winter snowfall that provide sufficient insulation and prevent the firn from rapid cooling. The detection algorithm relies on a threshold exceedance of multiyear monthly backscatter differences between September (when the delay in refreezing is strongest) and April (when snow accumulation is close to the maximum). An important caveat is that S1 may not directly, or only weakly, sense the aquifer water table due to the limited penetration at C-band. It is more likely that the time delay in the refreezing of water in the unsaturated part of the firn profile above the aquifer creates a unique backscatter signal and thus serves as a proxy for aquifer detection.

The total aquifer area mapped by S1 equals 54,800 km$^2$, in between estimates from OIB observations ($\sim$29,268 km$^2$) and model simulations (55,700–90,200 km$^2$). Aquifer locations detected by S1 show very good agreement with those observed along the flight lines of OIB, with $\kappa = 0.84$ (0.86 when glaciers or bare ice are excluded). The OIB data reports a smaller total aquifer extent, most likely caused by an incomplete
coverage, underdetection in the manual procedure, weak aquifer signals returning from deep water tables, small amounts of water (less than ~200 kg/m$^3$), or heterogeneous firn and ice structures above the aquifer. Uncertainty in the S1 aquifer detection is larger in bare ice regions, where low backscatter due to smooth ice surfaces can be confused with signal absorption by firm water. Furthermore, meteorological conditions could potentially hamper detection when they impact the delayed increase in backscatter typical to aquifer areas. A specific example is the refreezing in early autumn due to cold temperatures with lack of snowfall. Future research is recommended to investigate these confounding factors in the aquifer detection more in depth. Nevertheless, S1 provides an unprecedented satellite-based mapping of the Greenland firm aquifers. A strong asset is the long-term continuity in C-band observations with the S1 and the RADARSAT constellation missions, allowing for the monitoring of changes in aquifer conditions at the decadal time scale, such as inland migration to higher elevations and drainage.

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References


