

# 1 Seasonal and interannual variability of landfast sea ice in Atka Bay, 2 Weddell Sea, Antarctica

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8 **Abstract.** Landfast sea ice (fast ice) attached to Antarctic (near-)coastal elements is a critical component of the local physical  
9 and ecological systems. Through its direct coupling with the atmosphere and ocean, fast ice properties are also a potential  
10 indicator of processes related to a changing climate. However, in-situ fast-ice observations in Antarctica are extremely sparse  
11 because of logistical challenges and harsh environmental conditions. Since 2010, a monitoring program observing the seasonal  
12 evolution of fast ice in Atka Bay has been conducted as part of the Antarctic Fast Ice Network (AFIN). The bay is located on  
13 the north-eastern edge of Ekström Ice Shelf in the eastern Weddell Sea, close to the German wintering station Neumayer III.  
14 A number of sampling sites have been regularly revisited each year between annual ice formation and breakup to obtain a  
15 continuous record of sea-ice and sub-ice platelet-layer thickness, as well as snow depth and freeboard across the bay.  
16 Here, we present the time series of these measurements over the last nine years. Combining them with observations from the  
17 nearby Neumayer III meteorological observatory as well as auxiliary satellite images enables us to relate the seasonal and  
18 interannual fast-ice cycle to the factors that influence their evolution.

19 On average, the annual consolidated fast-ice thickness at the end of the growth season is about two meters, with a loose platelet  
20 layer of four meter thickness beneath, and 0.70 meter thick snow on top. Results highlight the predominately seasonal character  
21 of the fast-ice regime in Atka Bay without a significant interannual trend in any of the observed variables over the nine-year  
22 observation period. Also, no changes are evident when comparing with sporadic measurements in the 1980s and 90s. It is  
23 shown that strong easterly winds in the area govern the year-round snow distribution and also trigger the breakup of fast ice  
24 in the bay during summer months.

25 Due to the substantial snow accumulation on the fast ice, a characteristic feature is frequent negative freeboard, associated  
26 flooding of the snow/ice interface, and a likely subsequent snow ice formation. The buoyant platelet layer beneath negates the  
27 snow weight to some extent, but snow thermodynamics is identified as the main driver of the energy and mass budgets for the  
28 fast-ice cover in Atka Bay.

29 The new knowledge of the seasonal and interannual variability of fast-ice properties from the present study helps to improve  
30 our understanding of interactions between atmosphere, fast ice, ocean and ice shelves in one of the key regions of Antarctica,  
31 and calls for intensified multi-disciplinary studies in this region.

## 32 **1 Introduction**

33 The highly dynamic pack ice of the open polar oceans is continuously in motion under the influence of winds and ocean  
34 currents (Kwok et al., 2017). In contrast, landfast sea ice (short: fast ice) is attached to the coast or associated geographical  
35 features, such as for example a shallow seafloor (especially in Arctic regions) or grounded icebergs, and is therefore immobile  
36 (JCOMM Expert Team on Sea Ice, 2015). Fast ice is a predominant and characteristic feature of the Arctic (Dammann et al.,  
37 2019; Yu et al., 2014) and Antarctic coasts (Fraser et al., 2012), especially in winter. Its seaward edge may vary between just  
38 a few meters and several hundred kilometers from where it is attached to, mostly depending on the local topography and  
39 coastline morphology. The main processes for fast-ice formation are either in-situ thermodynamic growth, or dynamic  
40 thickening and subsequent attachment of ice floes of any age to the shore (Mahoney et al., 2007b).

41 In the Arctic, coastal regions that are characterized by an extensive fast-ice cover in winter are for example found in the  
42 Chukchi Sea and Beaufort Sea (Druckenmiller et al., 2009; Mahoney et al., 2014; Mahoney et al., 2007a), the Canadian Arctic  
43 (Galley et al., 2012), the East Siberian and Laptev Seas (e.g. Selyuzhenok et al., 2017), and the Kara Sea (Olason, 2016).  
44 While the fast-ice cover in these regions comes with its own particular impacts on the respective coastal systems, a common  
45 feature is that they have undergone substantial changes in recent decades (Yu et al., 2014). These include a reduction of fast-  
46 ice area (Divine et al., 2003), later formation and earlier disappearance (Selyuzhenok et al., 2015) and a reduction of thickness  
47 (Polyakov et al., 2003).

48 Along the Antarctic coastline, the fast-ice belt extends even further from the coast (Fraser et al., 2012; Giles et al., 2008) due  
49 to the presence of grounded icebergs in much deeper waters of up to several hundred meters (Massom et al., 2001a).  
50 Embayments and grounded icebergs provide additional protection against storms and currents, and are often favorable for the  
51 formation of a recurrent and persisting fast-ice cover (Giles et al., 2008). Fast-ice around Antarctica is still usually seasonal  
52 rather than perennial, and reaches thicknesses of around 2 meters (Jeffries et al., 1998; Leonard et al., 2006), although it may  
53 attain greater ages and thicknesses in some regions (Massom et al., 2010). It mostly forms and breaks up annually as a response  
54 to various environmental conditions, such as heavy storms (Fraser et al., 2012; Heil, 2006). Its immediate response to both  
55 local atmospheric conditions and lower latitude variability of atmospheric and oceanic circulation patterns via the respective  
56 teleconnections (Aoki, 2017; Heil, 2006; Mahoney et al., 2007b) make fast ice a sensitive indicator of climate variability and  
57 even climate change (Mahoney et al., 2007a; Murphy et al., 1995). Based on the complexity and significance of fast ice in the  
58 Antarctic climate system, there is an urgent need for prognostic Antarctic fast ice in regional models, and later in global climate  
59 models, to capture its potential major impacts on the global ocean circulation, as developed recently for the Arctic (Lemieux  
60 et al., 2016).

61 Although fast ice only represents a rather small fraction of the overall sea-ice area in Antarctica (Fraser et al., 2012), it may  
62 contribute significantly to the overall volume of Antarctic sea ice, especially in spring (Giles et al., 2008). The presence and  
63 evolution of Antarctic fast ice is often associated with the formation and persistence of coastal polynyas, regions of particularly  
64 high sea-ice production (Fraser et al., 2019; Massom et al., 2001a; Tamura et al., 2016; Tamura et al., 2012) and Antarctic

65 Bottom Water formation (Tamura et al., 2012; Williams et al., 2008). Also, it forms an important boundary between the  
66 Antarctic ice sheet and the pack ice/ocean, for example prolonging the residence times of icebergs (Massom et al., 2003),  
67 mechanically stabilizing floating glacier tongues and ice shelves, and delaying their calving (Massom et al., 2010; Massom et  
68 al., 2018). Therefore, one particularly interesting aspect of Antarctic fast ice is its interaction with nearby ice shelves, floating  
69 seaward extensions of the continental ice sheet that are present along nearly half of Antarctica's coastline. Under specific  
70 oceanographic conditions, supercooled Ice Shelf Water favors the formation of floating ice crystals within the water column  
71 (Foldvik, 1977; Dieckmann et al., 1986), as opposed to the regular process of sea-ice formation by heat transport from the  
72 ocean towards the colder atmosphere. These crystals may be advected out of an ice-shelf cavity and rise to the surface  
73 (Hoppmann et al., 2015b; Mahoney et al., 2011; Hughes et al., 2014). They are eventually trapped under a nearby fast-ice  
74 cover and may accumulate in a layer reaching several meters in thickness (Gough et al., 2012; Price et al., 2014; Brett et al.,  
75 2020). This sub-ice platelet layer has profound consequences for the local sea-ice system, and forms an entirely unique habitat.  
76 Thermodynamic growth of the overlying solid fast ice into this layer (by heat conduction from the ocean into the atmosphere)  
77 leads to subsequent consolidation, and the resulting incorporated platelet ice may contribute significantly to the local fast-ice  
78 mass and energy budgets. This phenomenon has been documented at various locations around Antarctica (Langhorne et al.,  
79 2015 and references therein), and where present, is a defining feature of the local coastal system. Refer to Hoppmann et al.  
80 (2020) for a comprehensive review of platelet ice.

81 The effects of fast ice on the exchange processes between ocean and atmosphere are further amplified by the accumulation of  
82 snow, as it forms a thick layer over large portions of the Antarctic sea ice (Massom et al., 2001b). However, the snow cover  
83 has opposing effects on the energy and mass budgets of sea ice in the region. On the one hand, due to its low thermal  
84 conductivity, snow acts as a barrier to heat transfer from sea ice to the atmosphere and effectively reduces ice growth at the  
85 bottom (Eicken et al., 1995). On the other hand, snow contributes significantly to sea-ice thickening at the surface through two  
86 distinct seasonal processes: snow-ice and superimposed ice formation. In winter/spring, the heavy snow load leads to the  
87 depression of the sea-ice surface below water level, causing flooding of the snow/ice interface. The subsequent refreezing of  
88 the snow/water mixture forms a salty layer of so-called snow-ice (e.g. Eicken et al., 1994; Jeffries et al., 1998; 2001). In  
89 contrast, in spring and summer, surface and internal snowmelt leads to melt water that can refreeze and form fresh  
90 superimposed ice, as it percolates through snow and eventually to the snow-ice interface (Haas, 2001; Haas et al., 2001;  
91 Kawamura et al., 2004). Both processes contribute significantly to sea-ice growth from the top, and thus to the overall sea-ice  
92 mass budget in the Southern Ocean.

93 Beyond its contribution to the general sea-ice mass and energy budget in the Southern Ocean, fast ice also plays an important  
94 role for the ice-associated ecosystem, as it provides a stable habitat for microorganisms (e.g. Günther and Dieckmann, 1999)  
95 and serves as a breeding ground for, e.g., Weddell seals and Emperor penguins (Massom et al., 2009).

96 Fast ice and its properties as described above have been studied around Antarctica for a long time, especially related to  
97 logistical work at the summer and overwintering bases close to the coast of the continent. In order to commonly coordinate  
98 and facilitate this research, and thus establish an international network of fast-ice monitoring stations around the Antarctic

99 coastline, the international Antarctic Fast Ice Network (AFIN) was initiated during the International Polar Year (IPY)  
00 2007/2008 (Heil et al., 2011). Active international partners are, e.g., Australia and China working at Davis Station and  
01 Zhongshan Station on the eastern rim of Prydz Bay in East Antarctica (Heil, 2006; Lei et al., 2010), New Zealand working out  
02 of Scott Base in McMurdo Sound in the Ross Sea (Langhorne et al., 2015, and references therein), Norway at the fast ice in  
03 front of Fimbul Ice Shelf at Troll Station (Heil et al., 2011) and Germany in Atka Bay at Neumayer III (Hoppmann, 2015),  
04 both in the vicinity of Dronning Maud Land. The regular, AFIN-related monitoring program at Neumayer III started in 2010  
05 in order to fill the observational gap in the Weddell Sea sector.

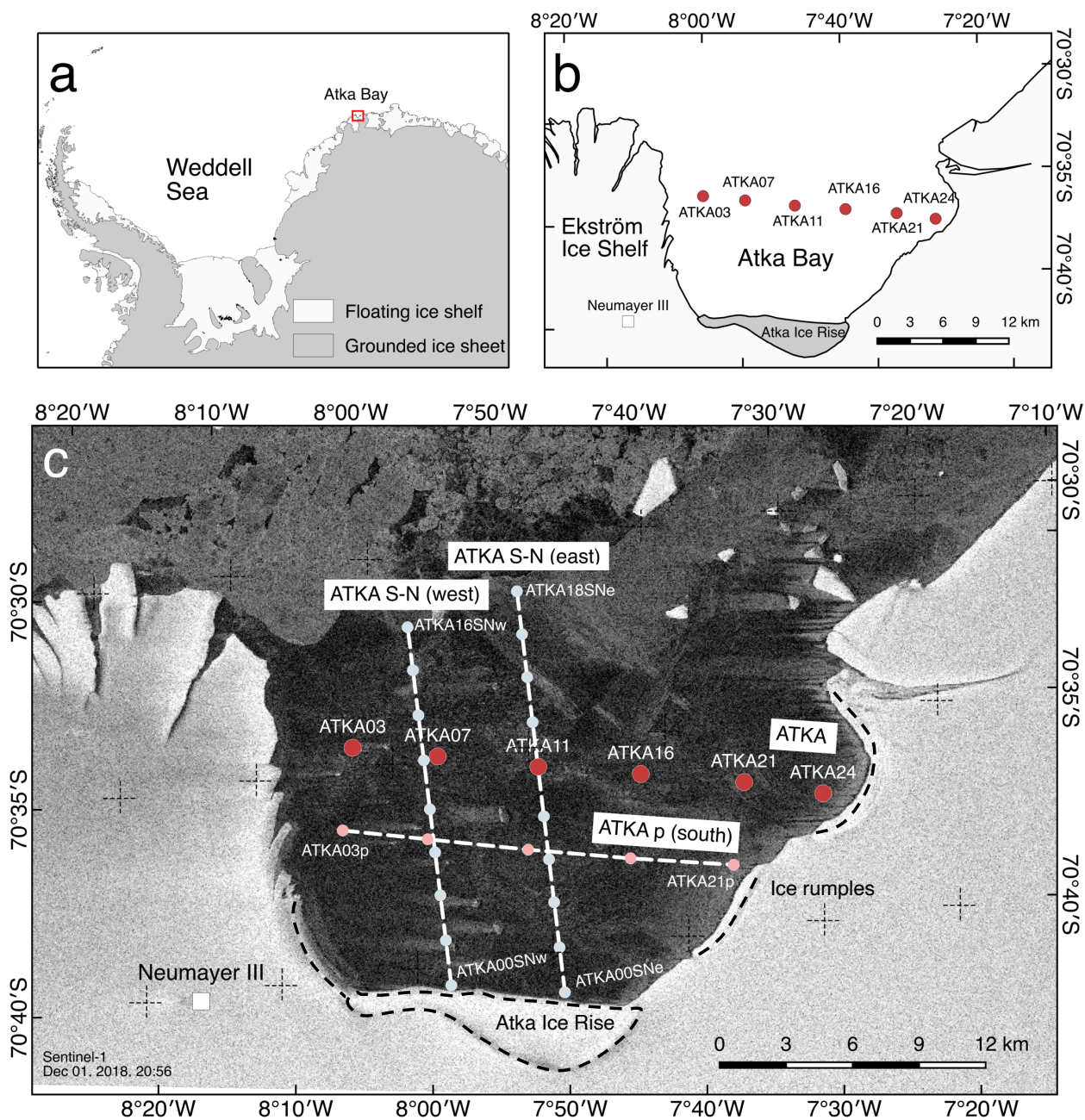
06 Here we present a decade of annual in-situ fast-ice observations in Atka Bay, which is one of the longest and most continuous  
07 time series within AFIN so far. The main dataset is a record of fast-ice thickness, snow depth, freeboard, and sub-ice platelet-  
08 layer thickness that was collected by a number of overwintering teams between 2010 and 2018. In addition to determining the  
09 spatio-temporal variability of the fast-ice cover, we co-analyze this data with meteorological observations and satellite imagery  
10 in order to determine how snow and platelet ice influence the local fast-ice mass budget. In doing so, we aim to improve our  
11 understanding of the interaction between the atmosphere, fast ice, ocean and ice shelves in one of the key regions in Antarctica.

## 12 **2 Study site and measurements**

### 13 **2.1 Study site: Atka Bay**

14 The main study area of this paper is Atka Bay, an 18 km-by-25 km embayment in front of the Ekström Ice Shelf located on  
15 the coast of Dronning Maud Land in the eastern Weddell Sea, Antarctica, at 70°35'S/ 7°35'W (Figure 1). Atka Bay is flanked  
16 towards the east, south and west by the edges of the ice shelf which rise as high as 20 meters above sea level. The cavity  
17 geometry of the Ekström Ice Shelf is one of the best known in Antarctica (Smith et al., 2020). Atka Bay is seasonally sea-ice  
18 covered, and the water depth ranges between 80 m and 275 m with maximum depth in the central bay (Kipfstuhl, 1991). Since  
19 the 1980s, when the first German research station Georg-von-Neumayer Station was established in the region, a variety of  
20 measurements has been carried out in the bay. Today's German research station Neumayer III is located at a distance of about  
21 8 km from the bay, where drifting snow regularly forms natural ramps between the sea ice and the ice-shelf surface. Prior  
22 investigations of the interactions between ice shelf, sea ice and ocean in the bay and its surroundings have been carried out by  
23 Kipfstuhl (1991) and Nicolaus and Grosfeld (2004), as well as more recently by Hoppmann et al. (2015a) and Hoppmann et  
24 al. (2015b). Ecosystem studies from the 1990's have been published by Günther and Dieckmann (1999) and Günther and  
25 Dieckmann (2001).

26



28

29 **Figure 1.** Overview of the study site and its surroundings. (a) Atka Bay (red marker) is located at the edge of the northeastern  
 30 Weddell Sea. Coastline data taken from SCAR Antarctic Digital Database. (b) Close-up of map (a) to focus on the study site  
 31 of Atka Bay. The sampling sites of the standard transect (ATKA) are marked with red circles. (c) Enlargement of (b) showing

32 in addition to the standard transect (red circles) the parallel transect in the south (ATKA p; light red circles) from ATKA03p  
33 to ATKA21p as well the eastern and western perpendicular transects ATKA S-N (east) from ATKA00SNe to ATKA18SNe  
34 and ATKA S-N (west) from ATKA00SNw to ATKA16SNw, each with a distance of 2 kilometers between adjacent sampling  
35 sites (light blue circles). The southern, eastern and western transects were sampled during a field campaign between November  
36 and December 2018. The dotted black curves indicate the locations of ice rises and rumples in Atka Bay. Background:  
37 Copernicus Sentinel data 01 December 2018, processed by ESA.

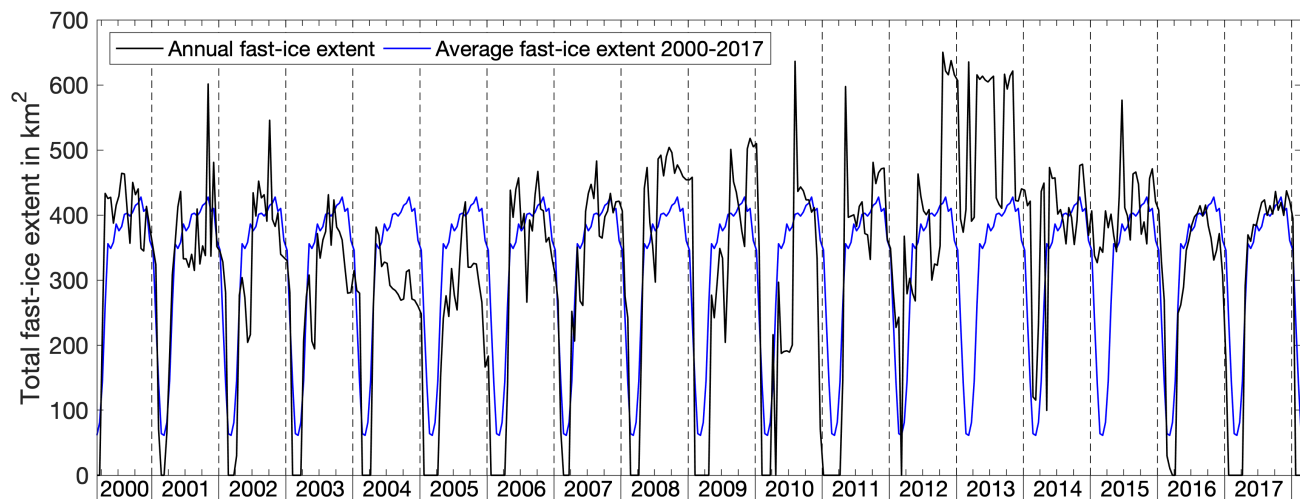
## 38 **2.2 Sea-ice conditions**

39 Atka Bay is seasonally covered with sea ice that is attached to the ice shelf to form immobile fast ice. Following the method  
40 of fast-ice time series retrieval detailed in Fraser et al. (2019), we obtained year-round estimates of fast-ice extent in Atka Bay  
41 from MODIS visible and thermal infrared satellite imagery. Hence, the fast-ice extent time series presented here in Figure 2 is  
42 a) produced at a 1 km spatial and 15 day temporal resolution, from 15-day MODIS cloud-free composite images (following  
43 Fraser et al., 2010) and edge-detected non-cloud-filtered composite images; b) spans the time period from March 2000 to  
44 March 2018; and c) is semi-automated in the sense that the fast-ice edge is automatically delineated during times of high  
45 contrast to offshore pack ice/open water, and manually delineated at other times.

46 Accordingly, the initial ice formation in the bay has started in March in recent years (Figure 2), with persistent easterly winds  
47 forcing increased dynamic sea-ice growth towards the western ice-shelf edge of the bay. Once the bay is completely covered  
48 by fast ice usually at the end of April (Figure 2), further in-situ ice growth takes place. In the following summer, the ice does  
49 not disappear by surface melting *in-situ*, but breaks up and drifts out of the bay once the conditions are sufficiently unstable.  
50 Stabilization and breakup of the ice-covered bay depend on the presence/absence of pack ice offshore of Atka Bay associated  
51 with changing ocean currents and winds, as well as stationary and passing icebergs. Thus, fast-ice breakup in the bay starts  
52 usually in December/January after the pack ice in front of the fast ice has retreated (Figure 2).

53 During our study period from 2010/2011 to 2018/19, there were two exceptions to this “typical” annual cycle: In September  
54 2012, a large iceberg (generally referred to as “B15G”) grounded in front of Atka Bay, sheltering the fast ice and consequently  
55 preventing sea-ice breakup in the following summer (Hoppmann et al., 2015b), resulting in second-year fast ice in the bay in  
56 2013. A year later, in August 2013, the iceberg dislodged itself, drifting westwards following the Antarctic Coastal Current.  
57 Fragments of the iceberg remained grounded in the northern part of the bay, causing it to be blocked again two years later, and  
58 therefore preventing sea-ice breakup in austral summer 2014/2015 for a second time within the study period. The iceberg  
59 fragments became mobile during the course of the following year, resulting in the bay to become ice-free again in the following  
60 summer.

61



62

63 **Figure 2:** Time series of fast-ice extent in Atka Bay between 8° 12'W and 7° 24'W derived from MODIS data between early  
 64 2000 to early 2018 (black line). The blue line shows the annual mean extent repeated each year over the same time period. The  
 65 average fast-ice extent over the entire time series is  $319.2 \pm 167.8 \text{ km}^2$ , with an uncertainty of  $86.6 \text{ km}^2$ .

66

### 67 2.3 Sea-ice measurements across Atka Bay

68 Since 2010, the AFIN monitoring protocol has been implemented to study the seasonal evolution of fast ice along a 24-km  
 69 long west-east transect in Atka Bay (“standard transect”, red circles in Figure 1). Here, six sampling sites have been regularly  
 70 revisited between annual sea-ice formation and breakup each year to obtain a continuous record of snow depth, freeboard, sea  
 71 ice- and sub-ice platelet-layer thickness across the bay (Arndt et al., 2019). Sampling sites on the standard transect are referred  
 72 to in this paper as ATKAx<sub>x</sub>, where xx represents the distance in kilometers to the ice-shelf edge in the west.

73 Generally, measurements along that standard transect are carried out once a month by the wintering team usually between June  
 74 and January, when safe access to the sea ice is possible. At each sampling site, up to five measurements are taken in an  
 75 undisturbed area, one as the center measurement and four more at a distance of approx. 5 meters in each direction (north, east,  
 76 south, and west), in order to account for the spatial variability of sea-ice and snow properties. In years of prevailing second-  
 77 year ice in the bay (2012/2013, 2014/2015), the number of observations per sampling site was reduced to one (the center  
 78 measurement) due to exceptionally thick snow and ice. Throughout this manuscript, we mainly present the mean values from  
 79 those up to five single measurements per sampling site. While all measurements along the standard transect from 2010/2011  
 80 to 2018/2019 are included in this study, the sea-ice monitoring activities will be continued beyond this work.

81 In November and December 2018, additional measurements in both, parallel and perpendicular transect lines to the standard  
 82 transect, have been performed (Figure 1c). Sampling sites on parallel transects are referred to in this paper as ATKAx<sub>pp</sub>, where  
 83 xx represents the distance in kilometers to the ice shelf edge in the west and “p” refers to “parallel”. Along the perpendicular

84 western (w) and eastern (e) transects from south to north, sampling sites are referred to in this paper as ATKAYySNw and  
85 ATKAYySNe, where yy represents the distance in kilometers to the ice-shelf edge in the south.

86 Sea-ice and platelet-layer thickness as well as freeboard are measured with a (modified) thickness tape. In order to enable the  
87 penetration of the usually semi-consolidated platelet layer, the regular metal plates at the bottom of the thickness tape were  
88 replaced by a metal bar of ~2kg. The underside of the platelet layer is determined by gently pulling up the tape and attempting  
89 to feel the first resistance to the pulling. Sea-ice thickness was measured either by pulling this modified tape through the entire  
90 platelet layer until the solid sea-ice bottom is reached (with a high risk of it getting stuck), or using a regular ice thickness tape.  
91 The modified tape is retrieved by pulling a small rope attached to one side of the metal bar. Snow depth was measured using  
92 ruler sticks. Freeboard is defined as the distance between the snow/ice interface and the sea-water level, while the snow/ice  
93 interface above (below) sea-water level is referred to as positive (negative) freeboard.

94 In order to determine the influence of snow and the underlying platelet layer to the observed freeboard (F), we also calculated  
95 the freeboard, assuming a hydrostatic equilibrium for floating snow-covered sea ice with an additional buoyancy (the platelet  
96 layer below), using Archimedes' principle:

$$97 \quad F = - \frac{I \cdot (\rho_I - \rho_W) + S \cdot \rho_S + P \cdot (\rho_P - \rho_W)}{\rho_W} . \quad (\text{Eq. 1.1})$$

98 As soon as F becomes negative, the involved components of the above-mentioned hydrostatic equilibrium are assumed to be  
99 balanced after the flooding of the snow/ice interface. Here, the depth of the wet soaked snow is considered as equal to the  
100 absolute value of the freeboard. Thus, the latter is calculated for the flooded case as

$$101 \quad F = - \frac{I \cdot (\rho_I - \rho_W) + S \cdot \rho_S + P \cdot (\rho_P - \rho_W)}{\rho_W - \rho_{swet}} . \quad (\text{Eq. 1.2})$$

102 In Equation 1.1 and 1.2, I refers to sea-ice thickness, S to the dry snow depth, P to platelet-layer thickness, the indices I refers  
103 to sea ice, S to dry snow, P to the platelet layer, and W to water. Constant typical densities of  $\rho_W = 1032.3 \text{ kg m}^{-3}$ ,  $\rho_S = 330$   
104  $\text{kg m}^{-3}$  and  $\rho_I = 925 \text{ kg m}^{-3}$  are assumed in this study. The density of the soaked snow, which is a mixture of water, snow, ice  
105 and air bubbles, is here assumed as  $\rho_{swet} = 920 \text{ kg m}^{-3}$  (e.g. Wang et al., 2015). The platelet-layer density  $\rho_P$  is calculated by  
106 means of the platelet-layer ice volume fraction  $\beta$  as  $\rho_P = \beta \cdot \rho_I + (1 - \beta) \cdot \rho_W$ . In this study, we used a constant ice-volume  
107 fraction of  $\beta = 0.25$ , as suggested by Hoppmann et al. (2015b).

108 The described bore-hole measurements are occasionally complemented by additional total (sea-ice plus snow) thickness  
109 measurements with a ground-based electromagnetic induction instrument (e.g. Hunkeler et al., 2016) as well as autonomous  
110 ice-based systems, such as Ice Mass balance or Snow Buoys (Grosfeld et al., 2015; Hoppmann et al., 2015a). However, this  
111 paper focusses on the regular bore-hole measurements only, as the additional observations address scientific questions beyond  
112 the scope of this paper.

113

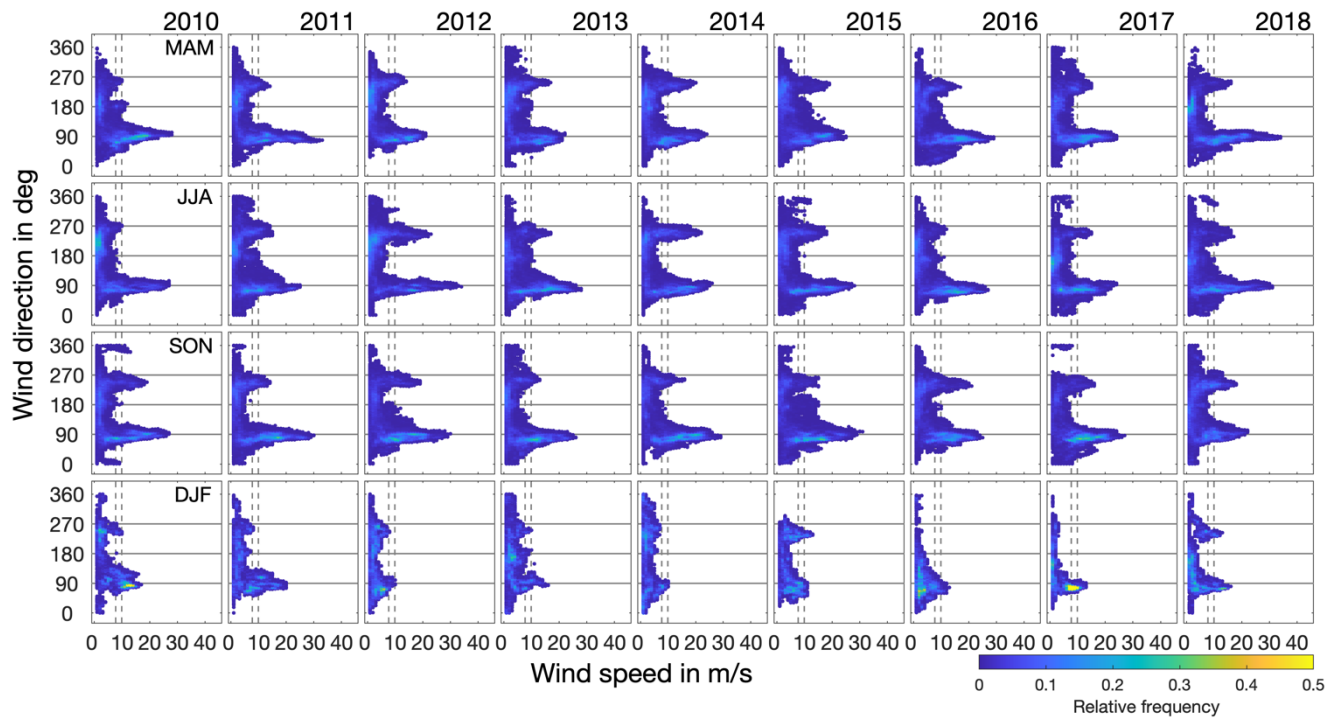


## 14 **2.4 Meteorological conditions and observations at Neumayer III**

15 At the meteorological observatory of the nearby wintering base Neumayer III, atmospheric conditions have been recorded  
16 since 1981 (König-Langlo and Loose, 2007), including the study period from 2010/2011 to 2018/19, and continuing beyond  
17 it (Schmithüsen et al., 2019). Occasionally, automatic weather stations (AWS) were temporarily installed on the sea ice to  
18 record the meteorological conditions directly on the sea ice (Hoppmann et al., 2015a). Since the 2m air temperature and the  
19 wind velocity at the meteorological observatory and the AWS on the ice showed a fairly good agreement in prior studies  
20 (Hoppmann et al., 2015a; Hoppmann et al., 2013), we use in this paper the more continuous records of the meteorological  
21 observatory in order to investigate the links between sea-ice conditions and atmospheric conditions. The Neumayer III data is  
22 recorded as minutely averages of typically 10 values per averaging interval. The instrumentation is checked on a daily basis,  
23 any erroneous values, e.g. caused by riming or instrument failure, are removed from the record. Therefore, the data quality can  
24 be considered high, even though there might be gaps in the records due to the validation routines. Nevertheless, data availability  
25 is 99.4% for wind direction, 99.0% for wind speed and 99.7% for air temperature. Uncertainties are essentially those classified  
26 by the manufacturers. Instrument details are given in the metadata of the datasets since February 2017 in Schmithüsen et al.  
27 (2019), earlier data is documented in König-Langlo and Loose (2007).

28 Generally, in the vicinity of Neumayer III the weather is strongly influenced by cyclonic activities which are dominated by  
29 easterly moving cyclones north of the station. This leads to prevailing persistent and strong easterly winds which exhibit a  
30 seasonal cycle with strongest winds during winter time (Figure 3). The second strongest mode in the wind direction distribution  
31 at 270° (westward) is associated with super geostrophic flows resulting from a high-pressure ridge north of Neumayer III  
32 (König-Langlo and Loose, 2007). These strong winds lead to frequent drifting and blowing snow. Here, we expect snow  
33 transport for 10-m wind velocities exceeding 7.7 m/s for dry snow and exceeding 9.9 m/s for wet snow (Li and Pomeroy,  
34 1997).

35



36

37 **Figure 3:** Distribution of wind speed related to wind directions separated for austral fall (March, April, May; MAM), winter  
 38 (June, July, August; JJA), spring (September, October, November; SON), and summer (December, January, February; DJF)  
 39 for the study period from 2010 to 2018. Colors indicate the relative frequency of each shown wind direction to wind speed  
 40 pair. Dashed vertical lines denote thresholds for 10-m wind speeds for snow transport of dry (7.7 m/s) and wet snow (9.9 m/s)  
 41 (Li and Pomeroy, 1997).

42

### 3 Results

43

#### 3.1 Nine-year record of sea-ice and plateau-layer thickness, snow depth and freeboard along a 24-km W-E transect

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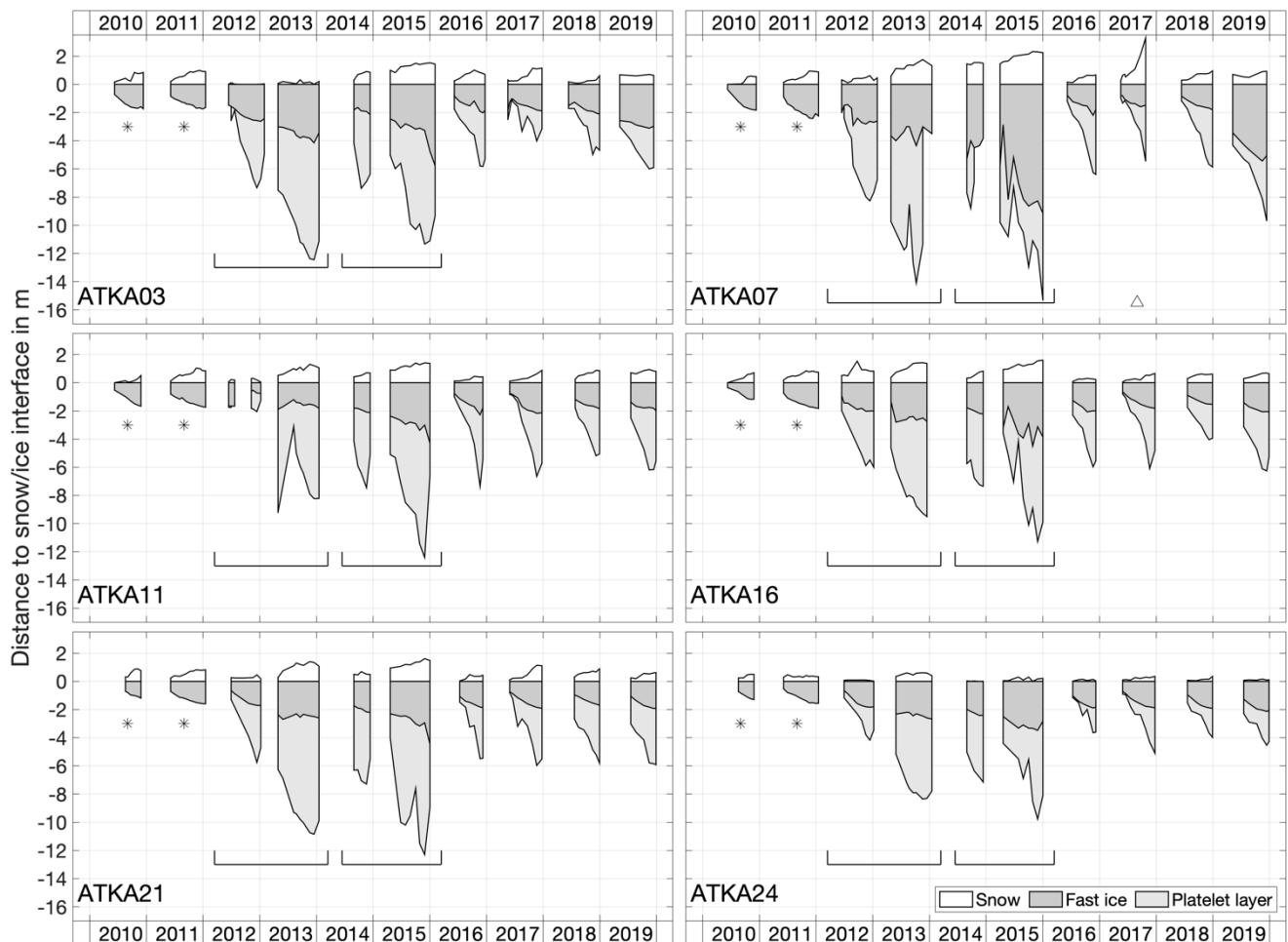
Figure 4 summarizes all conducted measurements of snow depth, sea-ice and plateau-layer thickness on the standard transect  
 45 from bore-hole measurements for each ATKA sampling site in the study period from 2010 to 2018. In the seven months when  
 46 sea-ice conditions allowed safe access (usually from May/June to December), about eight sets of measurements were taken  
 47 along the standard transect crossing Atka Bay, i.e. once every three to four weeks.

48

Analyzing the average annual maximum values of the investigated parameters (Table 1) for years of seasonal fast ice only  
 49 (excluding 2013 and 2015) and neglecting local iceberg disturbances (ATKA07 in 2017), the highest annual snow  
 50 accumulation of  $0.89 \pm 0.36$  m was measured at ATKA07, while the smallest by far was measured at ATKA24 at the  
 51 easternmost sampling site, with only  $0.28 \pm 0.19$  m. Averaged over the entire bay, the lowest snow accumulation of  $0.51 \pm$   
 52  $0.30$  m was observed in 2016. In contrast, 2011 was the year with the most snow and an average snow depth of  $0.85 \pm 0.20$  m

53 across the bay. The average seasonal fast-ice thickness based on the measurements during the observation period varied  
 54 between  $1.74 \pm 0.31$  m (ATKA21) and  $2.58 \pm 1.28$  m (ATKA07) with a mean value of  $1.99 \pm 0.63$  m. The underlying seasonal  
 55 platelet layer reached an average annual thickness of 3.91 m, which, however, shows a strong gradient in the average annual  
 56 maximum values (Table 1) from  $4.62 \pm 0.67$  m at ATKA07 in the west of the bay to  $2.82 \pm 1.20$  m at ATKA24 in the east.  
 57 In 2013 and 2015, the fast ice in Atka -Bay became second-year ice due to grounded icebergs in front of the bay. Within the  
 58 respective second year, snow depth increased further by an additional  $0.88 \pm 0.43$  in 2013 and by  $0.74 \pm 0.27$  m in 2015. In  
 59 2013, the average fast-ice thickness across the bay increased by an additional  $1.21 \pm 0.42$  m, while in 2015, it increased by an  
 60 additional  $2.79$  m  $\pm 1.48$  m. In the years of prevalent second-year ice in the bay, the thickness of the platelet layer increased  
 61 on average by  $5.13$  m  $\pm 1.43$  m in 2013 (compared to the end of 2012), and  $4.11$  m  $\pm 1.86$  m in 2015 (compared to the end of  
 62 2014). During these periods, ATKA11 experienced the highest annual platelet-layer thickness increase of  $6.82$  m and  $6.44$  m,  
 63 respectively.

64

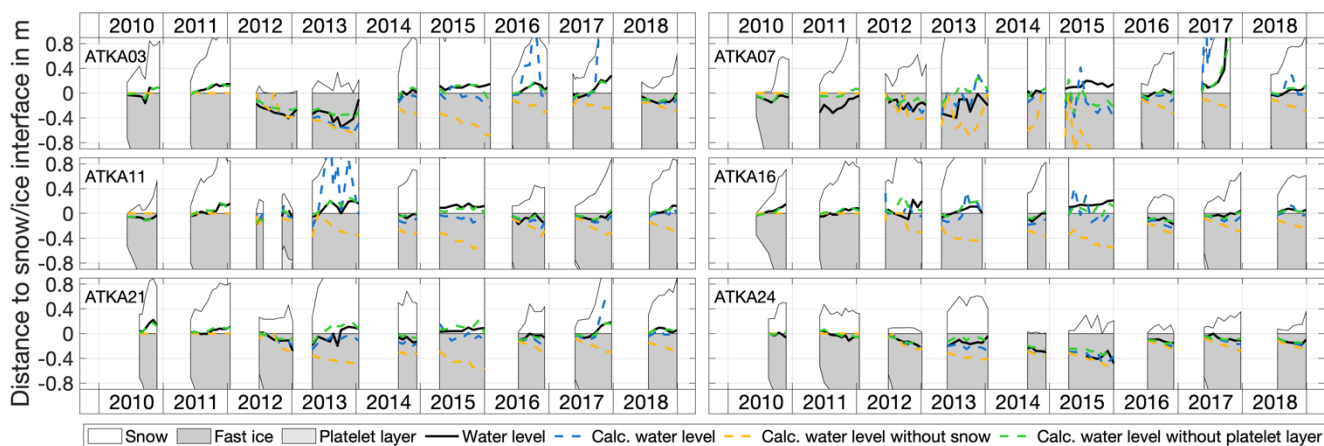


65

66 **Figure 4:** Time series of snow depth, fast-ice and platelet-layer thickness from bore-hole measurements along the standard  
 67 transect for each ATKA sampling site (Figure 1) for the time period from 2010 to 2018. Note: In 2010 and 2011, the platelet-  
 68 layer thickness was not measured (\*). In 2012/2013 and 2014/2015 Atka Bay was blocked by icebergs, so the fast ice did not  
 69 break up and turned into second-year ice instead (–). In 2017, a small iceberg in the vicinity of ATKA07 strongly influenced  
 70 the snow measurements ( $\Delta$ ). Reference depth of 0 meters is the snow/ice interface.

71  
 72 Figure 5 depicts the evolution of the water level with respect to the snow/ice interface (which is the freeboard with an opposite  
 73 sign) along the standard transect in the study period from 2010 to 2018. Taking all conducted freeboard measurements from  
 74 seasonal fast ice into account, 38% reveal negative data, i.e. flooding can be assumed, with an average negative freeboard of -  
 75  $0.10 \pm 0.08$  m. In contrast, considering freeboard measurements from second-year ice only, 55% of the data indicate a negative  
 76 freeboard, with an average of  $-0.22 \pm 0.15$  m. Analyzing the average annual maximum of the negative freeboard values (Table  
 77 1) for years of seasonal fast ice only, and neglecting local iceberg disturbances (ATKA07 in 2017), there is no distinct gradient  
 78 across Atka Bay, but higher average negative freeboard values ( $-0.07$  to  $-0.08$  m) are recorded both in the far west (ATKA03)  
 79 and in the east (ATKA16 and ATKA21), whereas the lowest average negative freeboard of  $-0.01 \pm 0.08$  m was measured at  
 80 ATKA07. According to Equation 1.1, 70% of the calculated freeboard values are smaller than the measured values. The  
 81 difference between measured and calculated freeboard values ranges from  $-0.54$  to  $1.26$  m with an average of  $-0.02 \pm 0.18$  m.  
 82 Neglecting the underlying buoyant platelet layer in the calculation reduces the freeboard by  $0.03 \pm 0.17$  m, whereas neglecting  
 83 the snow layer on top of the sea ice increases the freeboard by  $0.17 \pm 0.25$  m (Figure 5).

84



85

86 **Figure 5:** Close-up of Figure 4 which highlights the location of the water level with respect to the snow/ice interface (which  
 87 has the opposite sign of the freeboard) as measured in the field (black solid line) and as calculated according to Equation 1.1  
 88 and 1.2 including snow and platelet-layer thickness (blue dashed line), neglecting the snow cover (dashed yellow line) and  
 89 platelet layer (dashed green line), respectively. Please note that, for the purpose of better illustration, we depict here the actual

90 location of the water level rather than the freeboard (the only difference being the opposite sign). This means that, if the water  
 91 level is above the snow/ice interface, this is depicted in the figure accordingly, while the actual freeboard carries a negative  
 92 sign, and vice versa. The reference depth of 0 represents the snow/ice interface.

93

94 **Table 1:** Average annual maximum of snow depth, sea-ice and platelet-layer thickness, as well as freeboard (negative equals  
 95 potential flooding) on the standard transect from bore-hole measurements for each ATKA sampling site (Figure 1) for the time  
 96 period from 2010 to 2018, excluding years of second-year ice due to blocking of the bay (i.e. 2013 and 2015). <sup>1</sup>At ATKA11  
 97 all measurements of the year 2012 are also neglected as the ice has temporarily broken up again. <sup>2</sup>At ATKA07 the snow  
 98 measurements of the year 2017 are also neglected as a small iceberg has strongly influenced the accumulation rates. Standard  
 99 deviations are given in parentheses.

	ATKA03	ATKA07	ATKA11	ATKA16	ATKA21	ATKA24
Snow depth in m	0.81 (0.35)	0.89 (0.36) <sup>2</sup>	0.74 (0.23) <sup>1</sup>	0.79 (0.37)	0.77 (0.24)	0.28 (0.19)
Ice thickness in m	2.04 (0.31)	2.58 (1.28)	1.97 (0.25) <sup>1</sup>	1.81 (0.36)	1.74 (0.31)	1.83 (0.35)
Platelet-layer thickness in m	3.88 (1.31)	4.62 (0.47)	4.59 (0.83) <sup>1</sup>	3.99 (0.94)	4.21 (0.54)	2.82 (1.20)
Freeboard in m	-0.08 (0.14)	-0.01(0.08) <sup>2</sup>	-0.05(0.08) <sup>1</sup>	-0.07 (0.09)	-0.08 (0.10)	-0.05 (0.09)

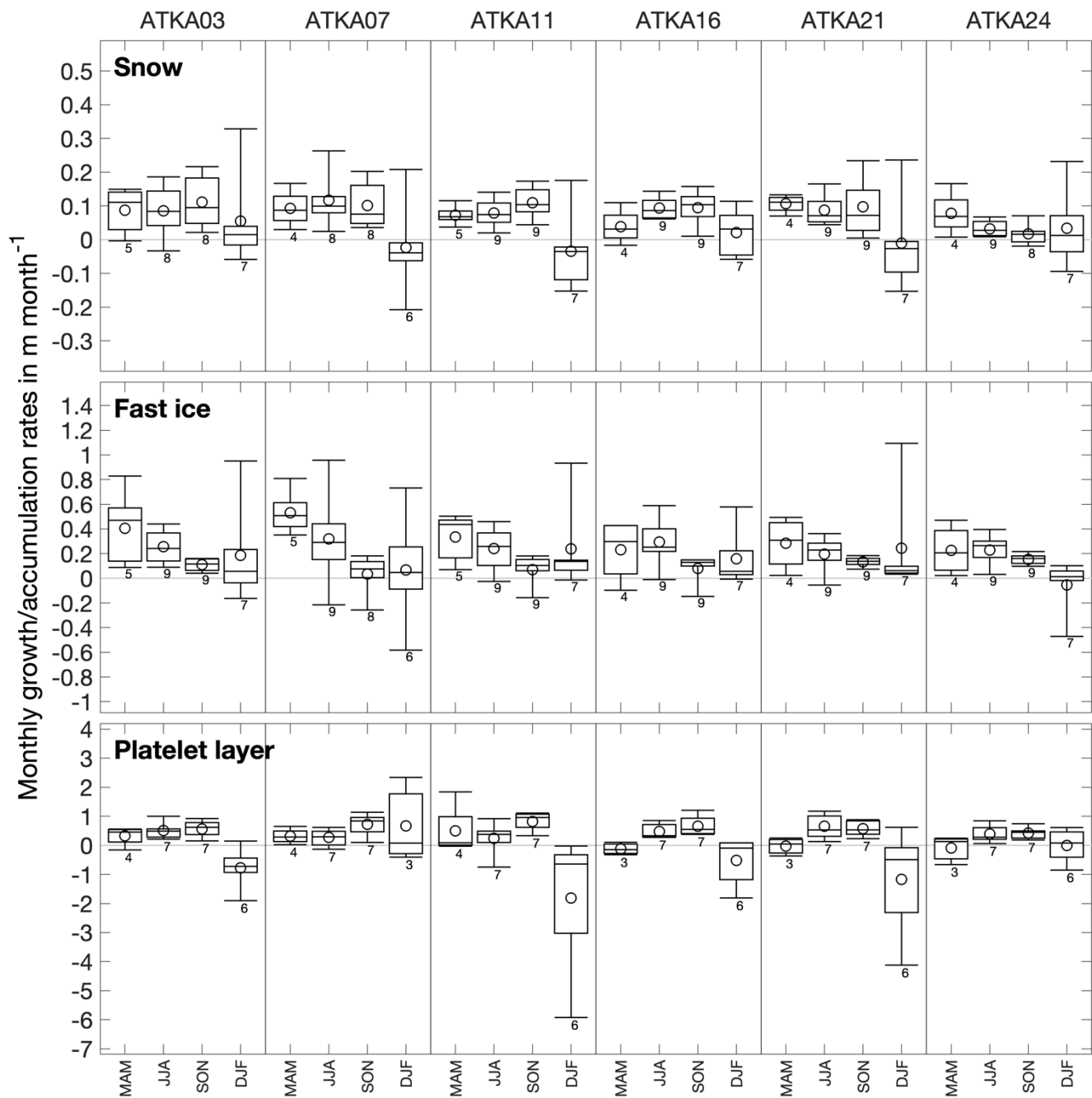
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 01

### 02 **3.2 Seasonal snow, sea-ice and platelet-layer accumulation/growth and melt rates**

03 Figure 6 summarizes the seasonal snow depth, sea-ice and platelet-layer thickness evolution separated for austral fall (March,  
 04 April, May; MAM), winter (June, July, August; JJA), spring (September, October, November; SON), and summer (December,  
 05 January, February; DJF) averaged for each ATKA sampling point over the duration of the whole study period from 2010 to  
 06 2018.

07 Considering the average monthly snow accumulation rates, a slight increase from fall (from 0.04 to 0.09 m per month, across  
 08 stations) to spring (0.09 to 0.11 m per month) becomes apparent, if excluding the eastern sampling sites at ATKA21 and  
 09 ATKA24. Latter sampling sites show the highest monthly averaged accumulation rates during austral fall (0.11 and 0.08 m  
 10 per month), which subsequently decrease to 0.10 and 0.02 m per month, respectively. In contrast, a clear snow loss with a  
 11 maximum monthly average of up to  $0.03 \pm 0.12$  m at ATKA11 and a maximum snow loss rate of 0.21 m per month at ATKA07  
 12 (80<sup>th</sup> percentile), can be seen mostly during summer months. The seasonal evolution of the platelet layer shows a similar  
 13 pattern: between austral autumn and spring, an average monthly thickness increase of up to  $0.82 \pm 0.30$  m at ATKA11 is

14 observed. Excluding ATKA07, afterwards an average monthly platelet-layer thickness decrease of 0.85 m is calculated for  
15 summer. The maximum decrease of 6.25 m per month occurred at ATKA11 in 2013 (80<sup>th</sup> percentile). However, it is highly  
16 likely that this is a measurement error. In contrast, ATKA07 also reveals an increase in platelet-layer thickness during the  
17 summer months with a monthly average of  $0.67 \pm 1.20$  m. With regard to the growth rates of fast ice in Atka Bay, a contrasting  
18 but expected seasonal development is observed: The highest average monthly fast-ice growth rates of up to approx. 1 m per  
19 month (80<sup>th</sup> percentile) are measured in autumn, and decrease in the following month until spring. These exceptionally high  
20 growth rates result from rapid growth of the solid fast ice into the (unconsolidated) sub-ice platelet layer, i.e. from the  
21 subsequent freezing of the interstitial water between the platelets in the top part of the platelet layer. In other words, some of  
22 the heat within the newly growing ice was already extracted earlier by the ocean during the process of platelet crystal formation  
23 in the supercooled Ice Shelf Water plume. In the subsequent summer months, average monthly sea-ice growth rates increased  
24 again to values between 0.07 m (ATKA07) and 0.24 m (ATKA21), except for ATKA24, where sea-ice melt dominates with  
25 an average monthly melt rate of  $-0.05 \pm 0.22$  m and a maximum monthly sea-ice melt rate of -0.58 m.  
26



27

28 **Figure 6:** Seasonal snow, sea-ice and platelet-layer accumulation/growth and melt rates separated for austral fall (March,  
 29 April, May; MAM), winter (June, July, August; JJA), spring (September, October, November; SON), and summer (December,  
 30 January, February; DJF) for each ATKA sampling point for the study period from 2010 to 2018. Boxes are the first and third

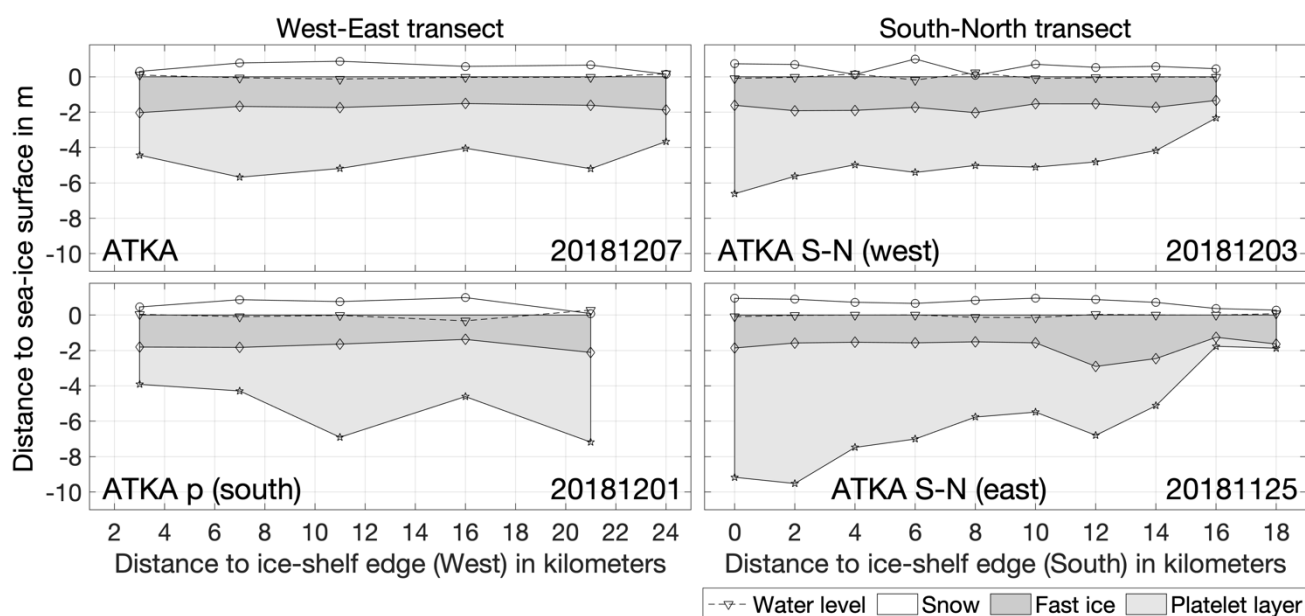
31 quartiles; whiskers the 20<sup>th</sup> and 80<sup>th</sup> percentile. Circles indicate the mean, horizontal lines in the boxes the median. Numbers  
 32 below the whiskers indicate the respective sampling size, i.e. the number of included years, with a maximum of nine.

33

### 34 3.3 Spatial variability of snow depth, sea-ice and platelet-layer thickness

35 In order to describe the spatial variability of snow depth, sea-ice and platelet-layer thickness in west-to-east as well as in south-  
 36 to-north direction across Atka Bay, additional parallel and perpendicular transects to the standard transect have been sampled  
 37 in November/December 2018 (Figure 7). Considering the solid sea ice only, the complementary transect data show that sea-  
 38 ice thickness over the bay in south-north and west-east direction is rather constant with an average of  $1.68 \pm 0.21$  m. In contrast,  
 39 neglecting the measurements in iceberg-affected areas, snow depth data show higher values in the south and in the center of  
 40 the bay of up to  $1.00 \pm 0.04$  m, while decreasing significantly towards the towards the ice-shelf edges in the east and west and  
 41 northern fast-ice edge to  $0.08 \pm 0.01$  m and  $0.28 \pm 0.09$  m, respectively. The platelet-layer thickness beneath the fast ice shows  
 42 a large spatial variability. While all measurements on the standard transect reveal the lowest platelet-layer thickness in the east  
 43 of the bay at ATKA24 (see Section 3.1), on the parallel transect in the south a maximum platelet-layer thickness of  $7.18 \pm 0.26$   
 44 m at the easternmost sampling point (ATKA21p) is observed. For the perpendicular transects in south-to-north direction, a  
 45 significantly decreasing gradient in platelet-layer thickness from the ice-shelf edge towards the northern fast-ice edge is  
 46 evident. On the western south-to-north transect, a decrease from  $6.62 \pm 0.25$  m to  $2.33 \pm 0.08$  m was observed, whereas for  
 47 the eastern transect this strong gradient is even more apparent with a decrease from  $9.17 \pm 0.11$  m to  $1.88 \pm 0.20$  m.

48



49



50 **Figure 7:** Overview of measurements on the standard transect from west to east (upper left), the parallel one (lower left), the  
51 western perpendicular transect from south to north (upper right) and the respective parallel one to the east (lower right) showing  
52 the water level, snow depth, fast-ice and platelet-layer thickness across Atka Bay. All measurements were conducted between  
53 November 25, 2018 and December 07, 2018. For the parallel west-east transect (December 01, 2018), the platelet-layer  
54 thickness evolution is influenced by a nearby iceberg (see Figure 1c). Also, for the western north-south transect (December  
55 03, 2018), snow measurements are influenced by several small icebergs in the vicinity between kilometers 4 and 8 (see Figure  
56 1 c).

## 57 **4 Discussion**

### 58 **4.1 Seasonal and interannual variability of snow depth, sea-ice and platelet-layer thickness**

59 The fast-ice regime in Atka Bay is primarily seasonal and the sea-ice cover usually only remains in the bay if a breakup is  
60 prevented by grounded icebergs in front of it. For example, while in 2013 a 17km-by-10km iceberg (B15G) blocked the entire  
61 bay, in 2015, only small iceberg fragments of B15G in front of the bay were sufficient to ensure that the sea ice in the bay did  
62 not break up, but became perennial. It may also occasionally happen that small areas of fast ice remain attached to the ice-  
63 shelf edge, or that individual ice floes remain in the bay and are incorporated into the newly growing ice in the following  
64 winter. Not only does the presence and size of the icebergs play a role in the fast-ice seasonality, but also the location and  
65 associated influence of atmospheric circulation patterns and ocean processes.

66 Considering first of all the seasonal sea ice only, the presented measurements along the standard transect across Atka Bay  
67 indicate a clear seasonal cycle in all investigated variables, i.e. snow depth, sea-ice and platelet-layer thickness: The initial  
68 sea-ice formation in Atka Bay starts in March and proceeds towards a completely fast-ice-covered bay at the end of April. The  
69 continuous sea-ice growth (i.e. ocean-atmosphere heat flux) proceeds with decreasing growth rate through fall and winter until  
70 the thickening snow cover more and more reduces the heat flux between the upper ocean and the atmosphere, preventing  
71 further thermodynamic sea-ice growth. However, the fast-ice thickness still increases in spring and even during austral summer  
72 months (albeit very slowly). This can be explained by the measurement uncertainty with respect to the large spatial variability  
73 of sea-ice thickness even on very small (centimeter) scales, but the consistency in the data suggests that it could also be caused  
74 by consolidation processes within the platelet layer below, i.e. in-situ sea-ice growth by heat transport into a supercooled  
75 plume residing right beneath the solid fast ice similar to observations in McMurdo Sound (Smith et al., 2012; Leonard et al.,  
76 2011; Dempsey et al., 2010; Robinson et al., 2014). So far, in Atka Bay there is only evidence that platelets grow quite large  
77 already while still suspended in the water column (Hoppmann et al., 2015b). To what degree an in-situ growth of platelet  
78 crystals and consolidation processes that go beyond regular freeze-in of the topmost part of the platelet layer by heat conduction  
79 to the atmosphere play a role at Atka Bay still needs to be investigated. In any case the platelet layer is an efficient buffer  
80 between the fast ice and the incoming warmer water in summer (Eicken and Lange, 1989), so the lack of noticeable fast-ice  
81 bottom melt is generally expected. Oceanographic (winter) data is sparse, and the monitoring at Atka Bay has recently been

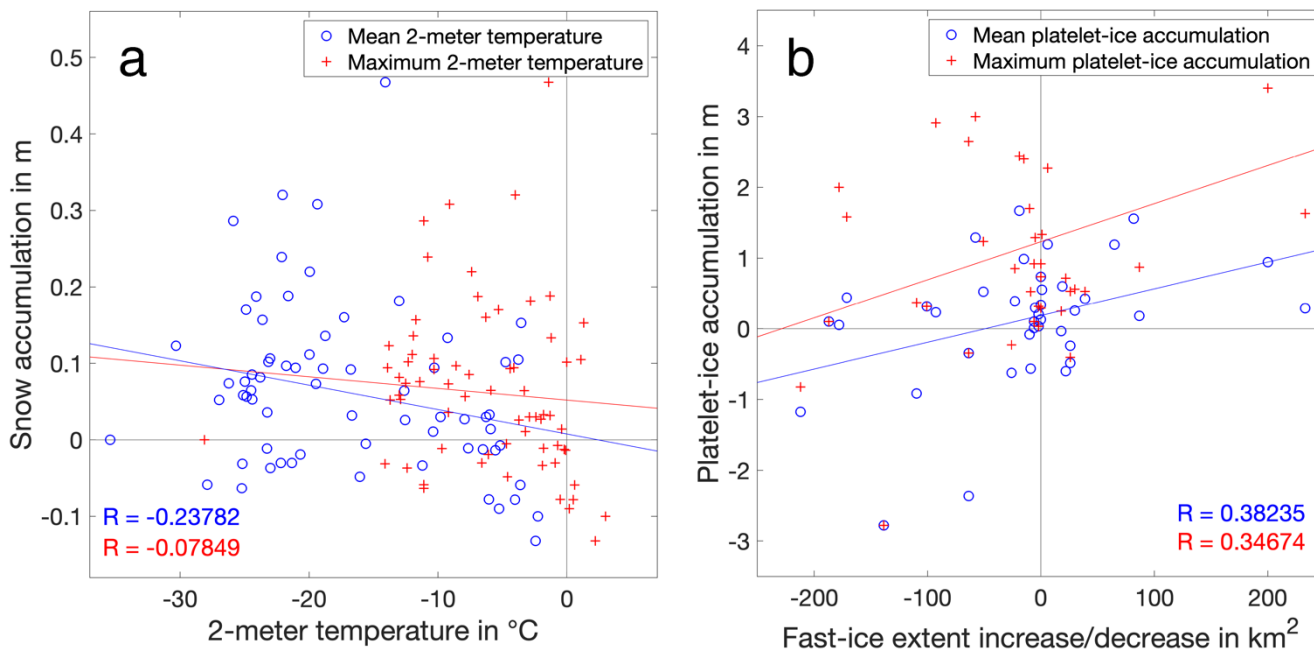
82 extended to also include regular CTD casts, whenever the (challenging) conditions and time constraints allow. An analysis of  
83 available CTD data in Atka Bay is currently ongoing, and will be shown in a future dedicated study to close the above  
84 observational gaps with respect to the ocean.

85 Destabilization of the fast ice and the platelet layer below in summer is to a large extent driven by the presence/absence of  
86 pack ice offshore Atka Bay. Thus, the initial breakup and subsequent retreat of the pack ice in front of the bay allow for locally  
87 increasing ocean currents beneath the fast ice and the inflow of warm Antarctic Surface Water from the east (Hoppmann et al.,  
88 2015a; Hattermann et al., 2012) causing both washing out of the platelet layer as well as mixing warm water into the water  
89 column associated with a thinning rate of the platelet layer of approx. one meter per month from December onwards. The  
90 retreating fast ice also initializes the sea-ice breakup in the bay starting usually in December/January (Figure 2). The  
91 diminishing fast-ice zone potentially goes along with an additional thinning of the platelet layer (Figure 8 b) by, e.g., further  
92 washing out mechanisms. Even though the correlation between the change of fast-ice extent and the mean (maximum) platelet-  
93 layer thickness between two consecutive surveys with an r-coefficient of 0.38 (0.35) is relatively low, Figure 8b indeed  
94 suggests that decreasing platelet-layer thicknesses are generally associated with retreating fast ice in Atka Bay. Also, Massom  
95 et al. (2018) have also shown that pack ice has a stabilizing effect as a buffer against ocean swells. The deployment of complex  
96 oceanographic moorings, either ice- or seafloor-based, would greatly help to further investigate ocean properties and currents  
97 and their effect on sea ice and the ice shelf, but their deployment, and especially their recovery, is extremely difficult and risky  
98 in the dynamic and harsh conditions of Atka Bay. While such deployments are logistically not feasible at the moment, it is  
99 planned to include suitable instrumentation in the monitoring within the next years.

00 In contrast to the decrease in platelet-layer thickness beneath the fast ice in summer over the entire bay, the snow cover on top  
01 does not show a clear seasonal pattern, but indicates a decrease in snow depth with increasing air temperature. However, even  
02 in summer, no consistent snow melt with associated strong mass loss is observed over the entire bay. Rather, a strong variability  
03 in snow depth over all sampling sites and all sampling years with a weak snow loss during summer months is observed (Figure  
04 6). The latter pattern is a result of both, temporary temperatures above freezing which favor surface melting (Figure 8a), and  
05 comparatively low wind speeds (Figure 3) preventing the accumulation of additional snow blown over from the surrounding  
06 ice shelf. These results match well with results from studies on the seasonal cycle of snow properties in the inner pack ice zone  
07 of the Weddell Sea as for example performed by Arndt et al. (2016), who also showed missing persistent summer melt as  
08 highlighted above.

09 Overall, the 9-year time series for in-situ snow depth, sea-ice and platelet-layer thickness in Atka Bay do not show any trend  
10 over the analyzed study period, whereas their inter-annual variability is dominated by local or temporary effects such as the  
11 presence of icebergs, which may for example lead to small-scale strong snow accumulations (Figure 4) or occasionally even  
12 to a perennial fast-ice regime. It is particularly remarkable that the average annual maximum platelet-layer thickness of 4 m  
13 (Table 1) is consistent with an earlier investigation at Atka Bay performed in 1982 by Kipfstuhl (1991) between the western  
14 ice-shelf edge and ATKA03, and, at the same time, much higher than in results from another study in 1995, where a maximum  
15 platelet-layer thickness of 1.5 m was measured in a similar location (Günther and Dieckmann, 1999). Considering a) the fact

16 that these two studies only sampled one location, and b) the generally large spatial and temporal variabilities of the platelet-  
 17 layer thickness, and in the present more detailed study, we infer that there seems to be no clear trend over the past decades.  
 18 Our results are also in line with a recent study of Brett et al. (2020), who also found spatially highly variable platelet layers of  
 19 4+m under fast ice in McMurdo Sound. Thereby, our results suggest that it is likely that the relevant (sub ice-shelf) processes  
 20 in this region have not changed much either. However, as already stated above, it is crucial to further look into all the available  
 21 oceanographic data that are available from the Atka Bay region in order to support (disprove) these indications provided by  
 22 the fast-ice monitoring. Another benefit of such comparison would be to strengthen (weaken) the hypothesis that fast-ice  
 23 properties can serve as an indicator for the status of an ice-shelf, as suggested by Langhorne et al. (2015).



24  
 25 **Figure 8:** Scatter plot comparing (a) the average 2-meter air temperature (see Section 2.4) and the snow accumulation between  
 26 two consecutive surveys, and (b) increasing (positive values) and decreasing (negative values) fast-ice extent and platelet-  
 27 layer thickness between two consecutive surveys. The analysis includes all measurements at all sampling sites throughout the  
 28 study period from 2010 to 2018. Blue circles and red crossed denote the respective mean and maximum values within the time  
 29 frame between the consecutive measurements. Colored solid lines in Figure (b) show the linear regression between both  
 30 parameters with the respective correlation coefficients R.  
 31

## 32 4.2 Spatial variability of fast-ice properties related to the distance to the ice-shelf edges around the bay

33 When neglecting local disturbances, such as icebergs, our results clearly indicate differences in the evolution of platelet-layer  
 34 thickness and snow depth with respect to the distance to the adjacent ice-shelf edges around Atka Bay. In contrast, the fast ice

35 itself does not exhibit any large spatial variability, with at the end of the season a nearly uniform thickness of 2 meters across  
36 the bay, both in west-east and south-north direction (Figure 7).

37 Analysis of the spatial distribution of platelet-layer thickness under the fast ice along the standard transect over the entire bay  
38 reveals that ATKA24 shows a significantly thinner platelet layer than all other sampling sites. In contrast, the parallel transect  
39 towards the south reveals a significantly higher platelet-layer thickness at the sampling point closest to the eastern shelf ice  
40 edge (ATKA21p). Perpendicular sampling transects from close to the southern ice-shelf edge towards the fast-ice edge in the  
41 north show a strong increase of platelet-layer thickness near the ice-shelf edge, followed by a moderate decrease in platelet-  
42 layer thickness towards the north, which rapidly decreases about 5 kilometers off the fast-ice edge. This thickness gradient is  
43 much more pronounced on the south-north transect in the central area of the bay compared to the western one. Moreover,  
44 considering the entire time series of the west-east transect (Figure 4), the highest platelet-layer thickness is observed in the  
45 central area of the bay (ATKA07 and ATKA11). Summarizing all these observations, we hypothesize that, on the one hand,  
46 the strongest outflow of supercooled water from the ice-shelf cavity (along with associated suspended platelet crystals) leads  
47 from the south centrally into the bay. On the other hand, local under-water topographic features of the ice shelf (i.e. ice rises)  
48 at the eastern boundary of Atka Bay (Figure 1c) might lead to a blocking of ocean currents and thus the advection of suspended  
49 platelet crystals, causing the high platelet-layer thickness at ATKA21p and the consistently low observed thickness north of  
50 this location at ATKA24 (Figure 4). The strongly decreasing gradient in the platelet-layer thickness towards the northern sea-  
51 ice edge is likely due to increasing distance from the source of suspended platelet crystals being advected from under the ice  
52 shelf and related washout effects. This is especially likely since at the time of the corresponding measurement, the pack ice in  
53 front of the bay was already broken open, allowing for wind-induced currents and locally solar-heated water production (the  
54 so-called mode 3 incursions, Jacobs et al. (1992), their Figure 1). Also, the fact that the northernmost sampling points are  
55 located close the edge of the bay or even already outside of it, raises the probability that the predominant coastal ocean current  
56 transports warm Antarctic Surface Water towards the fast-ice area, which would further intensify this effect (Hattermann et  
57 al., 2012; Hoppmann et al., 2015b). From this, it can be generalized that a smaller amount of platelet crystals can accumulate  
58 under narrow fast-ice areas, since these are exposed to stronger oceanic currents and associated washout effects, as well as  
59 warm water incursions. In that respect, Hoppmann et al. (2015b) used a subset of oceanographic data collected by the nearby  
60 PALAOA hydrographic observatory (Boebel et al., 2006) to link fast ice observations to ocean properties. A more recent study  
61 by Smith et al. (2020) helped to constrain the boundary conditions for Ice Shelf Water outflow by mapping in great detail the  
62 cavity geometry of the Ekström Ice Shelf. This study also shows data from repeated CTD casts through a borehole in the ice  
63 shelf, revealing the buoyant outflow of Ice Shelf Water in a relatively shallow surface layer. While these efforts help to better  
64 understand the complex system of ice shelf-ocean-sea ice interaction in this region, we suggest that a more comprehensive,  
65 year-round oceanographic study also implementing a dedicated survey program is still urgently needed as a complement to  
66 the ongoing sea ice monitoring. This would allow us to investigate in more detail the outflow of Ice Shelf Water and the  
67 complex processes involved in the redistribution of platelet crystals that emerge from the ice shelf cavity. Comparative

68 analyses to other study regions are not possible at this time, since, to our knowledge, no comparable transects were carried out  
69 so far in other Antarctic fast-ice regions with platelet layers beneath.

70 Examining the spatial distribution of snow over the bay, the considerably lower snow depth at ATKA24 compared to all other  
71 sampling sites is striking (Figure 4), and most likely related to the proximity to the ice-shelf edge in approximately 1 km  
72 distance. Due to the prevailing easterly winds in the bay (Figure 3), an east-west gradient in snow depth could have been  
73 expected over the rest of the bay. However, this gradient cannot be determined on average over the entire time series. This is  
74 mainly due to temporary local disturbance factors in the bay, such as icebergs and pressure ridges, which locally dominate the  
75 snow distribution and thus lead to a comparatively homogeneous distribution of snow depth over the central part of Atka Bay.  
76 A south-north survey across the bay at the beginning of austral summer 2018, however, revealed a trend of decreasing snow  
77 depth towards the northern fast-ice edge, with a stronger gradient approx. 5 km from the ice edge (Figure 7), which is in line  
78 with the northern boundary of the ice-shelf edge (Figure 1) and can therefore be explained by associated decreasing effects of  
79 the prevailing offshore winds. Consequently, also the measured snow depth is less in that part of the bay.

80 Due to the generally thick snow cover on Antarctic sea ice (Kern and Ozsoy-Çiçek, 2016; Markus and Cavalieri, 1998; Massom  
81 et al., 2001b), flooding of the snow/ice interface and the resulting formation of snow-ice is a widespread phenomenon in the  
82 Southern Ocean and contributes significantly to the sea-ice mass budget in the area (Eicken et al., 1995; Jeffries et al., 2001).  
83 While Günther and Dieckmann (1999) observed no flooding in Atka Bay during their study, Kipfstuhl (1991) reported flooding  
84 in relation to snow loads greater than 1 meter, an observation that we can largely confirm with our data. Exceptions are  
85 measurements on comparatively thin ice that already showed a sufficient snow layer, e.g. in the austral winter 2010 and 2011  
86 at ATKA21, leading to a negative freeboard and potential flooding already early in the season (Figure 5). Consequently, taking  
87 all conducted freeboard measurements on the seasonal fast ice into account, 55 % of the data indicate a negative freeboard, i.e.  
88 potential flooding and associated snow-ice formation can be assumed. While the snow cover reduces the buoyancy of the sea  
89 ice and accelerates flooding, the underlying platelet layer counteracts this by adding additional buoyancy. However, neglecting  
90 the platelet layer reduces the freeboard by  $0.09 \pm 0.06$  m, but still a negative freeboard is derived in half of the calculations.  
91 Thus, the spatial distribution of the sign of the freeboard, and therefore also the flooding of the snow/ice interface, is essentially  
92 controlled by the snow layer on top of the fast ice in Atka Bay. The thickness of the underlying platelet layer below, in turn,  
93 contributes to the resulting thickness of the flooded layer and consequently to the thickness of the potential snow-ice layer.

#### 94 **4.3 Impact of local disturbances on bay-wide properties and processes**

95 As already stated above, the largest overall effect on the fast-ice properties, the underlying platelet layer and the snow on top  
96 is due to the presence of icebergs in front of Atka Bay, which might entirely prevent a fast-ice breakup. At the same time,  
97 those grounded icebergs that are enclosed by sea ice within the bay add strong local effects. Thus, the large iceberg B15G  
98 grounded in front of Atka Bay sheltering the fast ice in the bay and consequently preventing sea-ice breakup in the following  
99 summer (Hoppmann et al., 2015b) led to second-year fast ice in the bay in 2013. Our measurements have shown that this  
00 hardly had any effect on the monthly snow accumulation rate and platelet-layer growth rate, but rather that these were within

01 the same range as in the years of seasonal sea-ice cover. Accordingly, for second-year sea ice, the total annual snow and  
02 platelet-layer thicknesses are approximately twice as thick as in the other years and average to  $1.30 \pm 0.60$  m and  $7.84 \pm 1.33$   
03 m, respectively. Higher snow loads do also increase the probability and extent of surface flooding. This is not only observed  
04 for years of second-year sea ice, but also for local disturbances as a result of the presence of small icebergs inside of the bay.  
05 In contrast, the evolution of the second-year fast-ice thickness shows a different pattern: Considering the large sea-ice thickness  
06 of around 2 m, as well as the insulating effect of the thick snow cover on top, the contribution of congelation growth is very  
07 limited. Instead, it is highly likely that dynamical growth as well as growth related to the consolidation of the platelet layer  
08 dominates the thickening of the perennial fast ice, adding up to an average thickness of  $4.19 \pm 1.90$  m, which is even more  
09 than double the thickness of seasonal sea ice in the bay.  
10

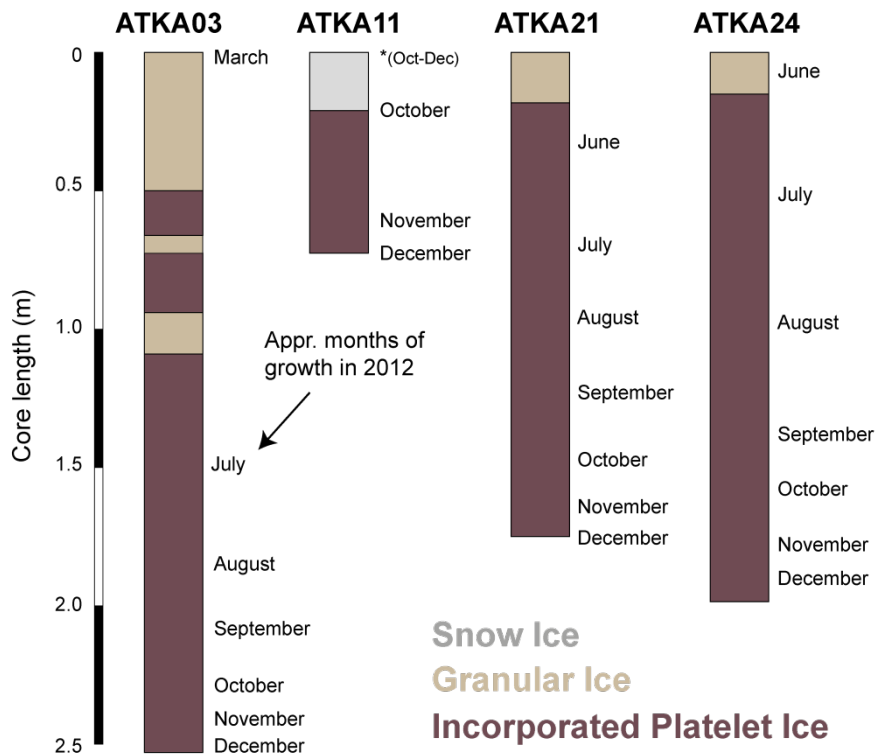
#### 11 **4.4 Sea-ice growth history**

12 A detailed study of sea ice crystal fabric by means of visual inspection of thick/thin sections or with the help of an automated  
13 fabric analyzer can help greatly to determine the dominant growth processes in a given area of interest. At the same time, the  
14 growth history of fast ice is to a large degree governed by the timing of the formation of a persistent ice cover, and can only  
15 be interpreted accurately by the help of as much auxiliary information as possible, most importantly from regular satellite  
16 imagery such as MODIS, Sentinel-1 or Radarsat.

17 It has been planned since the start of the AFIN monitoring at Atka Bay in 2010 to regularly obtain sea ice cores for crystal  
18 fabric analysis. A set of cores from the six main sampling sites (Figure 1) has been obtained in 2011, and again in 2012. Only  
19 4 out of these 12 cores have been processed so far (all from 2012), which is obviously only a very small sample size compared  
20 to the decade of measurements shown above. While the limited ice core data thereby is insufficient to make general statements  
21 about sea ice growth processes at Atka Bay, we provide this data here to highlight a few major aspects, some of which have  
22 already been discussed earlier.

23 From the (limited) data we have from the four 2012 cores (Figure 9), it is evident that 1. there is no columnar texture at all; 2.  
24 there is a small fraction of granular ice in the top parts of three cores; 3. there is a small fraction of snow ice in one core and  
25 4. all cores are dominated by incorporated platelet ice. The core from the western part of Atka Bay (ATKA03) exhibits a  
26 comparably high fraction of granular ice: a 0.5m long section at the top, and 2 smaller sections a little bit deeper, with some  
27 incorporated platelet ice in between. This crystal fabric is a manifestation of the dynamic conditions under which the initial  
28 growth takes place, and supports the other datasets shown above. The strong easterly winds (Figure 3) keep pushing the initially  
29 forming thin ice towards the western ice shelf edge, which leads to a grinding of the fragile frazil crystals, and subsequently  
30 to a rafting of the newly formed ice. This process seems to be still relevant even after the ice has thickened to  $>0.5$  m, probably  
31 by very strong winds. In this way, the thickening rate of the sea ice is greatly accelerated initially (Figure 4). The absence of  
32 exclusively columnar ice is evidence that there are already platelet crystals emerging from the cavity very early in the season.  
33 While it has been suggested in an earlier study that such crystals would be present in the bay from June onwards (Hoppmann

34 et al., 2015b), there is a possibility that they might arrive even earlier, at least in parts of the bay close to the outflow of ISW.  
35 While the ice core taken at ATKA11 is not representative at all for sea ice in the bay due to an early breakup event and  
36 subsequent late refreezing, the presence of snow ice is an evidence for a process that we would argue plays an underestimated  
37 role in this region. However, we currently do not have any more direct evidence for the wide presence of snow ice at Atka Bay  
38 (due to the lack of ice core data) other than the observations of negative freeboard in our main dataset (Figure 5), and several  
39 observations of extensive surface flooding from summer campaigns (although the latter do not necessarily result in snow-ice  
40 formation due to the usually warm isothermal temperature profile in the summer ice cover). In order to fill this knowledge  
41 gap, a dedicated program of obtaining much more core sections from the top of the sea ice at different locations would have  
42 to be implemented, with a subsequent crystal fabric and/or oxygen isotope analysis. As indicated above, this is currently not  
43 feasible. The other ice cores taken at ATKA21 and ATKA24 are close to the “typical” sea ice thickness at Atka Bay of 2 m,  
44 and exhibit the expected granular ice at the top from wind and waves, and incorporated platelet ice throughout the rest of the  
45 core. Again, ATKA21 does not show snow-ice formation at the top which might be expected from the snow depth (Figure 4)  
46 and freeboard/water levels (Figure 5) measurements. No evidence from dynamic growth processes is found in these cores.  
47 This is in line with our knowledge so far, especially since the sea ice in that area of the bay typically forms later in the year  
48 and is less influenced by strong winds.  
49



50

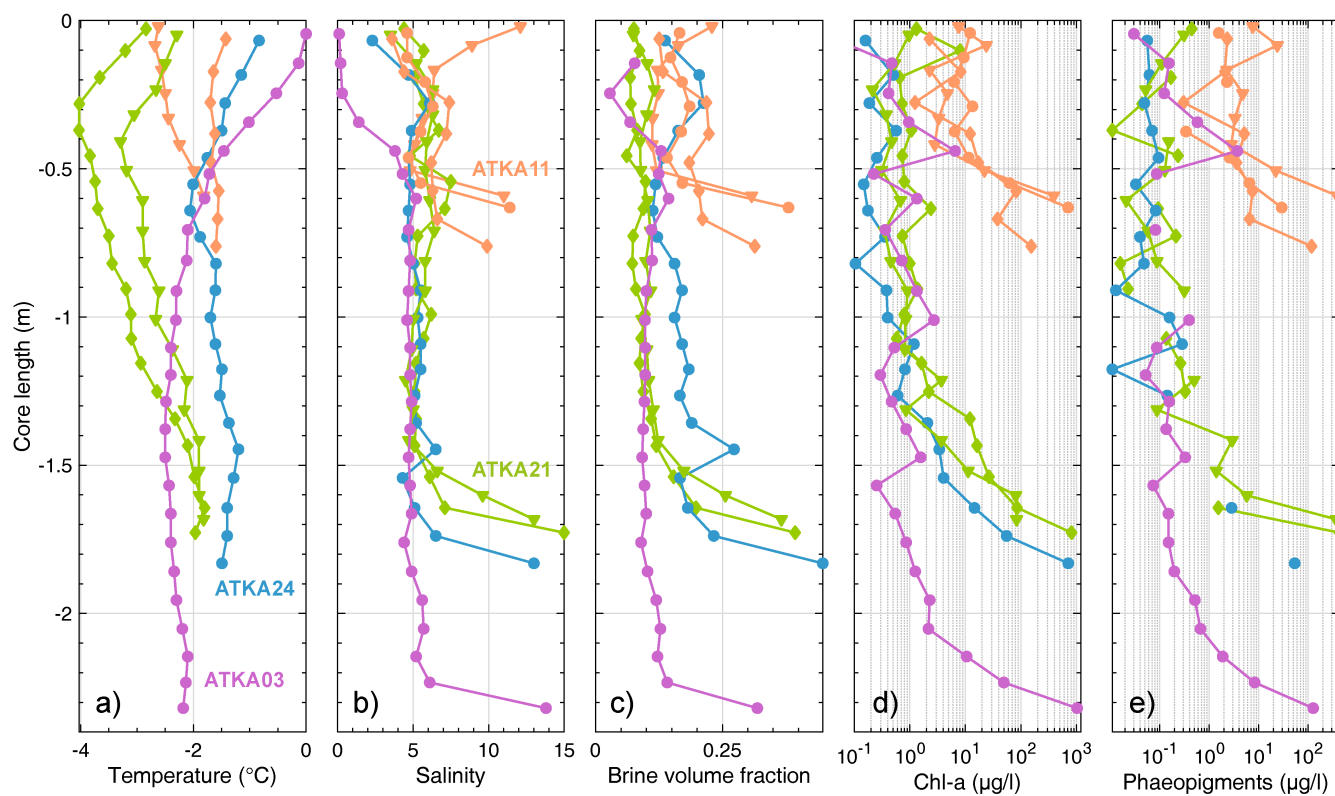
51 **Figure 9:** Sea-ice crystal fabric from ice cores obtained at four different fast ice sampling sites in December 2012, derived  
 52 from vertical and horizontal thin sections (0.1m spacing) along the full core length (see also Hoppmann et al., 2015a;  
 53 Hoppmann et al., 2015b; Hoppmann, 2015).

54

#### 55 **4.5 Implications for multi-disciplinary research**

56 Such a multi-layered, thick sea-ice cover not only very efficiently separates the atmosphere from the ocean with respect to ice  
 57 growth, but it also influences the exchange of any fluxes between the two climate system components. Thereby, it also strongly  
 58 impacts the ice-associated ecosystem, which is particularly unique in sub-ice platelet layers (Arrigo, 2014). Günther and  
 59 Dieckmann (1999) concluded from their study that about 99% of the total fast-ice biomass in Atka Bay originates from algae  
 60 initially growing in the sub-ice platelet layer. The maximum Chl-a concentration in their study was around 490 mg m<sup>-3</sup> in the  
 61 bottom of the fast ice, and 240 mg m<sup>-3</sup> in the platelet layer in summer, at a site that had up to 0.35 m of snow cover. The authors  
 62 argued that their total observed fast ice biomass was significantly lower compared to the mostly snow-free fast ice of the Ross  
 63 Sea. However, it was still in the very upper range of biomass usually found in Antarctic fast ice (Meiners et al., 2018). At the  
 64 same time, more recent results from 2012 at Atka Bay reveal that Chl-a concentrations can reach up to 900 mg m<sup>-3</sup> when there  
 65 is much less snow present (Figure 10).





67

68 **Figure 10:** Sea-ice physical and biological properties from cores obtained at different fast ice sampling sites in Nov/Dec 2012  
 69 (after Hoppmann et al., 2013).

70

71 While a few studies exist that investigate shade-adaptation in algae and link algal growth to snow depth on McMurdo Sound  
 72 fast ice (e.g. Sullivan et al., 1985; McGrath Grossi et al., 1987; Robinson et al., 1995), so far still comparably little is known  
 73 about the adaptation of the ecosystem in the upper ocean to perennial fast-ice conditions and sub-ice platelet layers. These and  
 74 similar knowledge gaps that exist with respect to ice-shelf influenced fast-ice regimes can only be addressed by integrated,  
 75 multi-disciplinary research in comparably easy to access locations in coastal Antarctica, one of which was introduced in this  
 76 physical study.

## 77 5 Conclusions

78 This study presents a unique, 9-year long record (2010 to 2018) of snow depth, freeboard, sea-ice and sub-ice platelet-layer  
 79 thickness observed at Atka Bay, a coastal Antarctic fast-ice regime in the southeastern Weddell Sea and key region in the  
 80 Southern Ocean. As one of the longest time series within the Antarctic Fast Ice Network, and complementary to similar records  
 81 in the Ross Sea (Brett et al., 2020; Langhorne et al., 2015 and references therein), this dataset is expected to serve as an

82 important baseline in the context of climate change and future sea-ice evolution in this region, and will contribute to an  
83 enhanced understanding of the complex interactions between the atmosphere, sea ice, ocean and ice shelves in the Southern  
84 Ocean.

85 For the period of the study presented, and considering individual observations from the 1980s and 1990s, a predominantly  
86 seasonal character of the fast-ice regime in Atka Bay is evident without a noticeable trend for any of the analyzed variables.  
87 The absence of any trend and the seasonality of surface characteristics associated with the year-round snow cover and  
88 negligible surface melting coincides with the prevailing conditions in the Antarctic pack-ice zone. Hence, the described  
89 observations in Atka Bay over the last nine years not only allow to document a baseline of the observed parameters, but also  
90 to capture processes and properties prior to expected future changes of pack ice in, e.g., the Weddell Sea, due to a changing  
91 climate.

92 Atka Bay is dominated by strong cyclonic events leading to easterly winds which determine not only the freeze-up of the bay  
93 in autumn and breakup during summer months, but also govern the year-round snow distribution on the ice. The consequent  
94 substantial annual snow accumulation determines both, the magnitude and duration of congelation sea-ice growth, as well as  
95 the magnitude and spatial distribution of the frequent negative freeboard and related flooding of the snow/ice interface, and  
96 thus potential snow-ice formation. In contrast, platelet ice contributes significantly to the total sea-ice mass balance in this  
97 region, both, in its unconsolidated form as an underlying (buoyant) platelet layer, as well as through its incorporation into the  
98 solid sea ice (see also Hoppmann et al., 2015a; Hoppmann et al., 2015b). However, our results indicate that, although the  
99 platelet layer partly offsets the negative freeboard, it is not buoyant enough to lift the snow/ice interface above sea level against  
00 the prevalent weight of the snow.

01 With regard to the platelet layer and its formation process, we conclude that, although the annual platelet-layer thickness  
02 increase of four meters seems to be independent of the age of the fast ice in the bay, the seasonal and inter-annual variability  
03 of this layer and thus the associated ocean properties and processes cannot be understood sufficiently by just considering the  
04 fast-ice properties alone. We therefore recommend to follow the approach of the New Zealand research program at Scott Base  
05 to generally include an oceanographic component into any fast-ice monitoring, especially in regions where ice shelves are  
06 present. This combination would allow for quantifying the seasonal interactions between sea ice, ocean and shelf ice even  
07 more precisely and thus to better understand current patterns and accumulation rates of platelet ice and associated biomass  
08 under the ice as a function of the distance to the shelf-ice and sea-ice edge. These results would provide a solid basis to be  
09 applied to all fast-ice areas around Antarctica, and thus make a fundamental contribution to the understanding of the Antarctic  
10 climate system.

## 11 **Data availability**

12 All presented meteorological data are archived in PANGAEA at <https://doi.pangaea.de/10.1594/PANGAEA.908826>. All fast-  
13 ice data are archived in PANGAEA at <https://doi.pangaea.de/10.1594/PANGAEA.908860>.

## 14 **Author contribution**

15 SA conducted most of the analyses for this paper and did the main writing with input from all co-authors. MH contributed the  
16 sea ice core data and helped SA to complete the revisions with input from all other authors. MN is the principal investigator  
17 of the AFIN work at Neumayer III. MH, MN and SA supervised the sea-ice measurements of the overwintering teams during  
18 the study period. MH and SA participated in field campaigns to collect parts of the presented data. HS contributed the  
19 meteorological datasets and the related analysis. AF contributed the fast-ice extent dataset and the related analysis.

## 20 **Competing interests**

21 The authors declare that they have no conflict of interest.

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