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Influence of sea ice and subglacial hydrology
on the dynamics of outlet glaciers flowing into
Lützow-Holm Bay, East Antarctica
東南極リュツォ・ホルム湾に流入する溢流水河の変動に
海水と底面水文環境が与える影響

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PhD Dissertation

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Abstract

In this thesis, satellite data analysis and field observations were conducted with the aim of understanding the mechanisms of flow changes of outlet glaciers in the Antarctic ice sheet. In Chapter 3, the relationship between flow changes and sea ice variations over the past two decades in five outlet glaciers (Shirase, Skallen, Telen, Honnør and Langhovde glaciers) flowing into Lützow-Holm Bay, East Antarctica, was analyzed using satellite data. In Chapter 4, short-term flow changes of Langhovde Glacier were observed by glacier surface measurements, and its mechanisms were investigated by direct observations at the glacier base using hot-water drilling.

Satellite data were analyzed for the five studied glaciers to measure terminus positions, flow velocities, surface elevations, and sea ice extent in front of the glaciers for the period 2000–2020. The terminus positions were measured by manually delineating the front using geographic information system (GIS) software on satellite imagery (JERS-1, ASTER, Landsat 4/7/8). Flow velocities were obtained using the annual mean flow velocity dataset of the Antarctic ice sheet distributed by NASA MEaSUREs (ITS_LIVE). I also analyzed Landsat 7/8 imagery to obtain the flow fields with high spatiotemporal resolution by applying the feature-tracking method. Furthermore, the strain distribution of the glacier was analyzed by processing the obtained glacier flow field. Changes in surface elevations were measured using digital elevation models from Reference Elevation Model for Antarctica (REMA) and laser altimetry data (ICESat-1/2). To measure sea ice variations near the glacier fronts, the reflectance of visible satellite imagery was analyzed within the regions located in front of the glacier. The pixels within the regions were classified into sea ice, open water, or clouds. The analysis was performed with Google Earth Engine to efficiently process the large amount of data acquired during 2000–2020. Satellite data analysis revealed that the terminal positions of the glaciers advanced by 0.6–38.7 km from 2000–2016 and then retreated by 0.4–6.0 km by 2018. The flow speeds of the largest three glaciers (Shirase, Skarlen and Thelen glaciers) showed deceleration during 2010–2015 and acceleration during 2015–2018. Elevation increased by 6.5 ± 0.2 m and 2.4 ± 0.2 m near the grounding line of Shirase and Skarlen glaciers in 2012–2016, followed by decrease by -0.8 ± 0.1 m and -3.03 ± 0.07 m. These results indicate that the glaciers retreated, accelerated, and thinned after the sea ice broke up in 2016. These results suggest that the land-fast sea ice in front of the glacier termini suppressed iceberg calving

and glacier flow. Furthermore, the longitudinal strain rate of the glaciers decreased during 2010–2015 and increased during 2015–2018, indicating that the elevation changes observed near the grounding line were due to changes in strain rate caused by flow changes. These results indicate that relatively small changes in the stress balance due to sea ice breakup caused glacier acceleration and calving, which in turn caused changes in ice thickness.

Field observations were carried out on Langhovde Glacier from December 2021 to February 2022 to observe short-term flow changes that are difficult to detect by satellite data. In order to directly observe the relationship between glacier basal sliding and subglacial water pressure variations, a device equipped with a three-axis accelerometer and a water pressure sensor was developed. The device was installed at the base of Langhovde Glacier, using a 550-m long borehole drilled approximately 1 km upstream from the grounding line during a hot-water drilling campaign in the austral summer 2021/22. On the glacier surface, global navigation satellite system (GNSS) and an automatic weather station were operated to observe glacier flow changes and meteorological conditions. As a result of the observations, flow acceleration of up to 20% was measured when the subglacial water pressure rapidly increased between 2 and 4 January. During the same period, impulsive and strong horizontal vibration events were observed at the subglacial accelerometer. The occurrence of the events was twice greater than the background value, suggesting active basal sliding during the speedup. The increase in the subglacial water pressure occurred a day after the rainfall of 30 mm d^{-1} over the glacier. The correspondence strongly suggests that the drainage of the surface water into the glacier bed enhanced basal sliding and caused short-term glacier flow acceleration. These results are the first observation showing that the surface-to-bed hydrological connections in Antarctica caused pressurization of the drainage system and speedup, which have traditionally been considered to have limited or no impacts on the subglacial conditions.

The study revealed the variations of the five outlet glaciers flowing into Lützow-Holm Bay, East Antarctica, during 2000–2020 for the first time. Satellite observations revealed that the stress changes due to sea ice variations caused glacier flow changes. Field observations on Langhovde Glacier captured short-term summer flow acceleration for the first time in East Antarctica. Direct observations at the glacier bed revealed that the flow acceleration was caused by the enhanced sliding due to the subglacial water pressure elevated by the drainage of surface water to the glacier bed. The results of the satellite and field observations indicate that small changes in the glacier stress balance at the ice front or the base resulted in significant flow changes. In other words, outlet glaciers of the Antarctic ice sheet are highly sensitive to changes in the surrounding environment such as the atmosphere and the ocean, which results in glacier flow changes. These results contribute to an improved understanding of glacier variations in Lützow-Holm Bay and provide new insights into the mechanisms of outlet glacier changes in the Antarctic ice sheet.

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List of Abbreviations

AWS	Automatic Weather Station
ASTER	Advanced Spaceborne Thermal Emission and Reflection radiometer
CMIP6	Coupled Model Intercomparison Project Phase 6
DEM	Digital Elevation Model
ETM+	Enhanced Thematic Mapper Plus
GIS	Geographic Information System
GNSS	Global Navigation Satellite System
ICESat	Ice, Cloud, and land Elevation Satellite
IPCC	Intergovernmental Panel on Climate Change
ISMIP6	Ice Sheet Model Intercomparison Project for CMIP6
JARE	Japanese Antarctic Research Expedition
JERS	Japanese Earth Resources Satellite
mCDW	modified Circumpolar Deep Water
NDSI	Normalized Difference Snow Index
NOAA	National Oceanic and Atmospheric Administration
NSIDC	National Snow and Ice Data Center
OLI	Operational Land Imager
PDF	Probability Density Function
PSD	Power Spectral Density
REMA	Reference Elevation Model for Antarctica
ROBOTICA	Research of Ocean-ice BOUNDary INteraction and Change around Antarctica
ROI	Region of Interest
SAR	Synthetic Aperture Radar
SCAR	Scientific Committee on Antarctic Research
SIC	Sea Ice Concentration
SMB	Surface Mass Balance
SWIR	Short Wave Infrared Radiometer
TM	Thematic Mapper
TOA	Top Of Atmosphere

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General introduction

1.1 Mass loss of the Antarctic ice sheet

The Antarctic ice sheet forms the largest ice mass on the earth, holding $26.0 \pm 0.4 \times 10^6 \text{ km}^3$ of the volume of ice which accounts for 60% of the earth's freshwater, equivalent to 57.9 ± 0.9 m of sea level (Morlighem and others, 2020). The ice sheet has an area of $13.9 \times 10^6 \text{ km}^2$, covering 98% of the Antarctic continent. The ice sheet gains ice mass from snowfall, and the accumulated snow forms glacial ice by densification processes due to its own weight, which flows from inland toward the ice sheet margins. At the coastal margins, the ice flow converges into rapidly flowing outlet glaciers and ice streams, which eventually feed into the ocean. The outlet glaciers separate from the bed and form ice shelves floating in the ocean. Ice shelves cover 74% of the coastal margins of the ice sheet (Bindshadler and others, 2011), and the boundary between the ice shelf and the grounded ice is called the grounding line, which is reported to be $5.4 \times 10^4 \text{ km}$ in total length around Antarctica. The ice sheet lost $2671 \pm 530 \text{ Gt}$ of ice from 1992 to 2020, thereby contributing to a rise in sea level of $7.4 \pm 1.5 \text{ mm}$ over the past three decades (Otosaka and others, 2023). The rates of the mass loss are accelerating from $-19 \pm 39 \text{ Gt a}^{-1}$ between 1997 and 2001 to a maximum rate of $-150 \pm 43 \text{ Gt a}^{-1}$ between 2012 and 2016 (Otosaka and others, 2023). The dominant contributor to the mass loss is West Antarctica, which raised the sea level by $6.6 \pm 0.7 \text{ mm}$ during 1992–2020, accounting for 89% of the total ice loss from Antarctica.

Projections of the ice sheet mass changes are crucial for the human population and earth's climate since its significant loss would result in flooding of coastal regions, modification of the terrestrial environments and ocean circulations (IPCC, 2021). Estimation of the future contribution of the Antarctic ice sheet to sea-level rise until 2100 has been reported by the Ice Sheet Model Intercomparison Project for CMIP6 (ISMIP6), as a part of the Coupled Model Intercomparison Project Phase 6 (CMIP6). They projected the contribution of the global sea-level rise within the range of -76 to 300 mm under unabated warming pathway, Representative Concentration Pathway 8.5 (RCP8.5) (Payne and others, 2021). The current projections lie

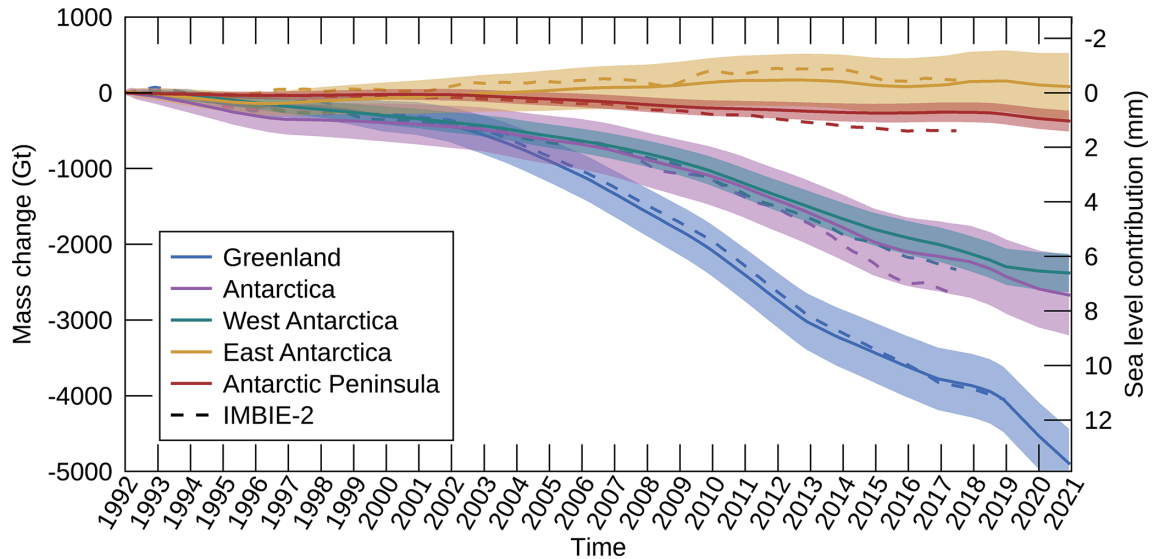


Figure 1.1: Cumulative ice sheet mass changes compiled by Ice Sheet Mass Balance Inter-comparison Exercise (IMBIE) (solid lines). The shadings represent the associated uncertainties. The dashed lines show the results from the previous assessments by IMBIE (The IMBIE team, 2018; The IMBIE Team, 2020). Figure from Otosaka and others (2023)

both in the positive and negative contribution with the absolute range of ~ 380 mm, highlighting the uncertain future of the ice sheet. This uncertainty is because of the counteracting effects of mass loss due to ocean warming and mass gain from increased snowfall. However, when this projection was extended to the year 2300, a severe mass loss equivalent to ~ 1.5 m of sea-level rise was projected for the Antarctic ice sheet as a mean of fourteen-experiment ensemble (Greve and others, 2023). Furthermore, the projection by the year 3000 showed an even greater amount of mass loss reaching ~ 3.5 m, despite the climate forcing after 2100 was set to the late-21st-century climate without any further trend by 3000 (Chambers and others, 2022). These modelling studies demonstrate that the ice sheet has the potential to be a substantial contributor to sea-level rise in the coming centuries, thus, there is an urgent need to quantify recent mass changes and understand their mechanisms for accurate prediction.

The mass changes of the grounded part of the Antarctic ice sheet are determined by a balance between surface mass balance (SMB) and ice discharge. SMB is the mass change on the surface of the ice sheet and is calculated by accumulation minus ablation. Accumulation on the Antarctic ice sheet is mainly given by snowfall. Since the climate over the Antarctic ice sheet is exceptionally cold and arid, only limited snowfall (< 100 mm w.e. a^{-1}) deposits on the surface (Mottram and others, 2021). The snow accumulates on the ice sheet because of the limited ablation due to ice sheet surface melt, which only occurs only part of the the region near the ice sheet margin (Trusel and others, 2013). Ice discharge is the mass flux passing through the grounding line that drains into the ocean from the ice sheet. The discharge accounts for most of the ablation from the Antarctic ice sheet, which has been a driving process of the recent ice sheet mass changes (Rignot and others, 2019). An increase in ice discharge is caused by the acceleration of outlet glaciers that transport ice mass from the inland to the coast, thus

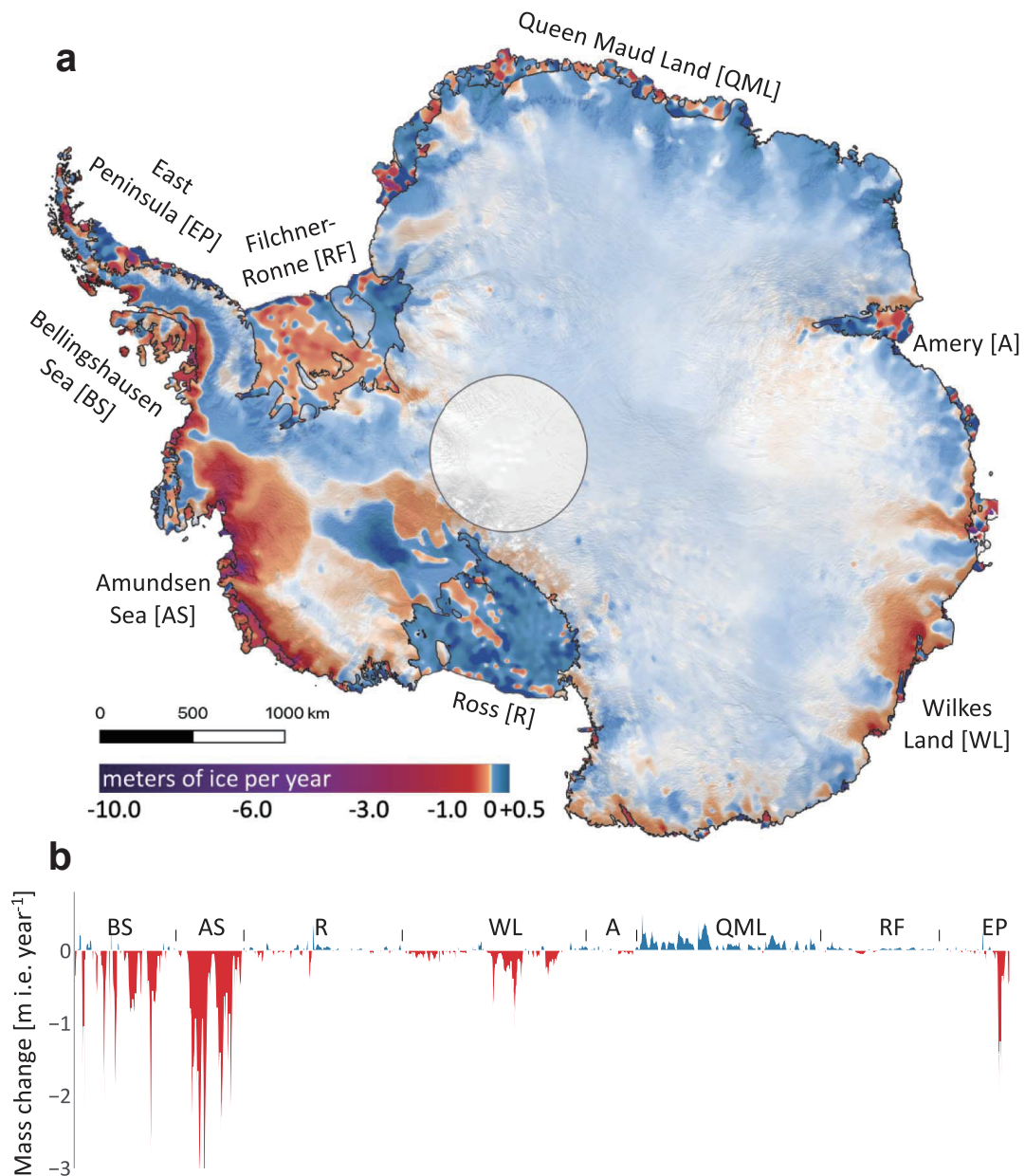


Figure 1.2: (a) Elevation change for Antarctica. (b) Elevation changes at the grounding line. The highest mass loss rates are observed in West Antarctica and Wilkes Land, East Antarctica. Map and grounding line mass change have been smoothed with a 35-km median filter for improved visualization. Figure from Smith and others (2020b)

ice flow of these glaciers is crucial for ice sheet mass balance. The glacier acceleration and subsequent changes in the ice flow regime have resulted in glacier thinning due to enhanced horizontal stretching of ice, a process known as the dynamic thinning (Pritchard and others, 2009). The thinning of the ice has been observed as a lowering of the surface elevation over the ice sheet (Smith and others, 2020b) (Fig. 1.2).

Extensive mass loss as a result of increased ice mass discharge and associated decrease in the surface elevation have been observed in West Antarctica, the Antarctic Peninsula, and Wilkes Land in East Antarctica (e.g. Pritchard and others, 2009; Smith and others, 2020b). In these regions, subglacial bedrock topography lies below sea level and deepens considerably toward the inland. Numerical studies proposed that the vulnerability of the ice sheet lies on such bedrock geometry, which showed sustained retreat of the grounding line without reaching a steady state (Weertman, 1974; Schoof, 2007). The inherent instability of the ice sheet is hypothesized as one of the factors controlling the observed significant mass loss in West Antarctica (Joughin and others, 2014). In contrast to the rapid changes in West Antarctica and the Antarctic Peninsula, East Antarctica has been relatively stable according to observations over the past three decades. The mass budget of East Antarctica has shown a slight increase of $3 \pm 15 \text{ Gt a}^{-1}$ during 1992–2020 (Otosaka and others, 2023). The mass increase is thought to be due to increased SMB, although it is the most uncertain component for the estimate for the Antarctic mass changes (Otosaka and others, 2023).

1.2 Flow changes of outlet glaciers

Processes acting at the interface of the ocean and the ice sheet margins are crucial for the ice sheet mass budget since the processes control flow and thickness changes of the outlet glaciers discharging ice into the ocean. Major glaciers contributing to the ice sheet mass loss in West Antarctica have shown a significant acceleration upstream from the grounding line, which is attributed to a loss of buttressing force from ice shelves (Pritchard and others, 2012). The key processes are the thinning of the ice shelves and retreat of the grounding lines, which have been caused by increased basal melting by the intrusion of warm and dense modified Circumpolar Deep Water (mCDW) into the sub-shelf cavity (e.g. Shepherd and others, 2004; Dutrieux and others, 2014; Christianson and others, 2016). These processes have been studied on outlet glaciers flowing into the Amundsen Sea Embayment sector in West Antarctica, a dominant contributor to sea level rise from the Antarctic ice sheet (Rignot and others, 2019; Shepherd and others, 2019). Grounding line retreat (Rignot and others, 2014), flow acceleration (Mouginot and others, 2014), and thinning (Pritchard and others, 2009; Smith and others, 2020b) have been observed in association with a strengthening of CDW inflow to the bay (Jacobs and others, 2011). Such glacier acceleration is caused by changes in stress balance triggered by ocean-induced ice shelf thinning. Previous numerical studies demonstrated the impact of the ice shelf thinning. For example, ice flow modelling

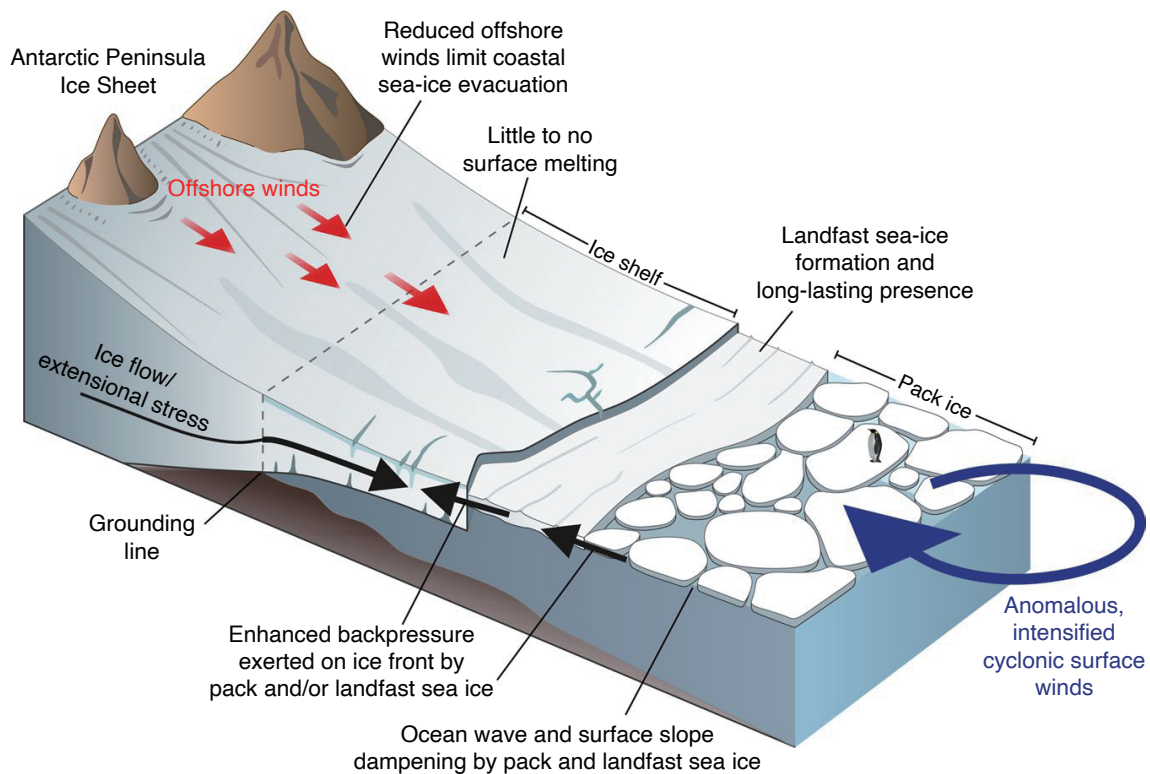


Figure 1.3: Schematic diagrams showing the key atmospheric and sea-ice processes controlling the stability of the eastern Antarctic Peninsula's ice shelves. Figure modified from Christie and others (2022).

experiments on Pine Island Glacier in the Amundsen Sea Embayment showed that reduction of basal drag due to grounding line retreat could result in ice thinning ~200 km inland from the grounding line via transmitted longitudinal stress gradient (Payne and others, 2004). Another study simulated the instant response of ice flow to the loss of buttressing force due to observed ice shelf thinning in Antarctica during 1994–2012. The results predicted the pattern of glacier acceleration and associated ice loss consistent with the current observation (Gudmundsson and others, 2019). These studies suggest that the thinning or loss of ice shelves could cause flow acceleration of the outlet glaciers, which results in rapid thinning over the inland ice sheet.

Other studies suggested that glacier dynamics are also influenced by changes in stress balance induced by iceberg calving from the glacier or ice shelf front. In the Antarctic Peninsula, for example, the rapid disintegration of Larsen A and B Ice Shelves triggered a significant speedup of the outlet glaciers feeding into the ice shelves (Rignot, 2004; Scambos, 2004), and resulted in increased ice discharge and mass loss (Wuite and others, 2015). Another study demonstrated that the recent terminus retreat by 20 km at Pine Island Glacier caused ice speedup from 4000 to 4600 m a^{-1} near the grounding line during 2017–2020 (Joughin and others, 2021). For an accurate understanding of the mechanisms of the Antarctic ice mass loss, detailed observations are needed on the ice speed, the front position, and the ice thickness near the glacier front and the grounding line, since these factors control the discharge from the ice sheet (Dupont and Alley, 2005; Fürst and others, 2016).

In contrast to the rapid changes in West Antarctica and the Antarctic Peninsula, outlet glaciers in East Antarctica have been relatively stable. An exception is Totten Glacier located in Aurora Subglacial Basin in East Antarctica, where the ice lies on the bedrock well below sea level and warm water intrusion has been causing grounding line retreat and mass loss (Rignot and others, 2019; Hirano and others, 2023; Li and others, 2023). In East Antarctica, previous studies showed a link between glacier terminus variations and sea-ice concentrations (SIC) as well as air temperature (Miles and others, 2013, 2016; Baumhoer and others, 2021). Land-fast sea ice plays a role in terminus variations of glaciers in Antarctica, as it affects the stability of ice shelves by exerting buttressing force on the glacier front (Fig. 1.3). A change in the buttressing effect has the potential to modify the glacier flow speed as well as the front position. For example, ice shelves in Porpoise Bay simultaneously disintegrated after the breakup of land-fast sea ice in front of the glaciers (Miles and others, 2017). Other studies of East Antarctica (Miles and others, 2013, 2016) and Antarctica as a whole (Baumhoer and others, 2021) showed that long-term glacier advance and retreat coincided with positive and negative anomalies of SIC, respectively. Sea ice in front of glaciers also stabilizes glacier termini by damping ocean swell reaching the glacier fronts. In the Antarctic Peninsula, the disintegration of Larsen B Ice Shelf was deemed to be a loss of sea ice because open water allowed storm-generated ocean swell to trigger extensive collapse of the ice shelf (Massom and others, 2018).

An established link between sea ice and glacier change is important for the response of the Antarctic ice sheet to the rapidly changing climate, since land-fast sea ice is highly sensitive to climate change (Heil, 2006; Fraser and others, 2012; Aoki, 2017). Indeed, recent observations showed a substantial decline in Antarctic sea-ice extent with a record-breaking minimum level during the winter of 2023, which was attributed to the recent Southern Ocean warming (Purich and Doddridge, 2023). The observations suggest that the sea ice exhibits abrupt changes as a result of influences from the climate. However, studies focused on the impact of sea ice on East Antarctic glacier dynamics are scarce. Further studies are needed to improve understanding of ongoing glacier changes and their mechanisms in this region.

1.3 Basal motion and subglacial conditions

Fast flow of the outlet glaciers is mainly driven by basal motion, i.e., sliding of ice over its bed and deformation of the underlying unlithified sediment. Thus, understanding their mechanisms is critical for projecting mass changes of the ice sheet (Ritz and others, 2015; Brondex and others, 2019). The processes of basal motion include regelation and enhanced ice creep around bedrock bumps (Weertman, 1957), ice-bed separation due to cavity formation (Lliboutry, 1968), and deformation of underlying sediment (Boulton and Hindmarsh, 1987). All these processes occur where the bedrock has reached its pressure melting point, and would account for a significant portion of ice flow. Therefore, the presence of subglacial water at the

ice-bed interface is an important factor in facilitating the fast flow of glaciers.

One of the most important factors controlling the basal motion is subglacial water and its pressure. For the glaciers lying on hard bedrock, theories have been developed to propose the role of elevated subglacial water pressure, which facilitates the growth of subglacial cavities on the lee side of bedrock undulations (Lliboutry, 1968; Iken, 1981). Expanded subglacial cavities lead to ice-bed separation and reduction in basal drag, which result in higher basal sliding rate (Iken, 1981; Bindshadler, 1983). The investigation by Iken (1981) predicted that the maximum sliding speed would occur when the subglacial cavities were expanding, thus, the rate of the cavity increase is more relevant for basal sliding speed than the cavity volume itself. The processes of cavity formation and associated active basal sliding have been confirmed by field observations, which showed that the maximum horizontal speed was achieved during the surface moved upward (Iken and others, 1983; Kamb and Engelhardt, 1987). The uplift of glacier surface can be caused by two processes, i.e., subglacial cavity expansion and ice thickening due to vertical straining, and a following study has shown that the uplift was indeed caused by cavity expansion at the glacier base (Sugiyama and Gudmundsson, 2004). Such ice-bed separation caused by water pressure increase could result in significant glacier speedup within short periods (less than a day or a week) reaching several times larger velocity than the background value.

Another important process that facilitates fast glacier flow is the deformation of underlying unlithified sediment (i.e., till). Active seismic experiments carried out on Willans Ice Stream revealed the existence of a thick (~5–6 m) water-saturated sediment layer beneath the ice stream (Blankenship and others, 1986). Seismic analysis showed that the porosity of the underlying sediment was ~0.4, and saturated with subglacial water at a pore pressure only $\sim 50 \pm 40$ kPa less than the ice overburden pressure, suggesting the sediment was very weak and actively deforming under the ice. The deformation within the sediment layer drives the fast flow of the ice stream, thus its mechanical properties should play a key role in the processes of the basal motion (Alley and others, 1986). In Whillans Ice Stream, subglacial sediments were sampled using boreholes drilled to the bed, and laboratory experiments were conducted to investigate their mechanical and rheological properties (Kamb, 1991; Tulaczyk and others, 2000). The till exhibited Coulmb-plastic rheology, with the yield strength highly dependent on the normal effective stress (Tulaczyk and others, 2000). Experiments showed that the sediment deforms very easily under sustained high water pressure. Such deformation has been proposed as a mechanism of the fast flow of West Antarctic glaciers with relatively flat surfaces and thus low driving stress.

Hot-water drilling is an essential technique for direct observation at the glacier bed. The observations have provided important results contributing to our knowledge about subglacial conditions and mechanisms of basal motion. In Antarctica, extensive hot-water drilling observations were conducted in Siple Coast, West Antarctica, during the 1990s. The drilling provided wide variety of observation records that directly obtained under the 1000-m-thick

ice (e.g. Kamb, 2001), i.e., subglacial pressure variations (Engelhardt and Kamb, 1997), till samples (Tulaczyk and others, 2000), and glacier sliding rate (Engelhardt and Kamb, 1998). The observations revealed that the glacier base was melting, and the water pressure (p_w) was close to the ice overburden pressure (p_i), i.e., the effective pressure defined by $N = p_i - p_w$ was close to zero. These conditions are favorable for both rapid sliding and sediment deformation, thus result in basal lubrication and fast glacier flow. However, the lack of subglacial measurements in other regions of the ice sheets limits our understanding of the mechanisms of the fast glacier flow.

Direct observations of basal motion at the glacier bed are essential for understanding its mechanisms. Studies were conducted by installing sensing devices at the glacier bottom using boreholes drilled by hot water (e.g. Clarke, 2005). For example, Iverson and others (1994) developed a dragometer, which measures the tension of a cable as it tows an anchor placed in the sediment, and quantifies sediment strength (Fig. 1.4A). Blake and others (1994) installed a drag spool at the glacier bottom, which places an anchor in subglacial sediment and measures the length of a wire as it pays out from a spool fixed in a borehole (Fig. 1.4B). The measurements directly quantified glacier basal sliding rate on alpine glaciers (Blake and others, 1994) and ice streams (Engelhardt and Kamb, 1998). Ploughmeter was developed by Fischer and Clarke (1994) which measures the bending moment on a steel rod that penetrates the glacier substrate, and quantifies force variations acting of the device as it is dragged on the glacier bed (Fig. 1.4C). Deformation of subglacial sediment has been measured using tilt cells installed in the sediment (Fig. 1.4D) (Blake and others, 1992; Truffer and others, 2000). Such in-situ observations are crucial for revealing the processes and mechanisms of glacier basal motion, but only limited observations have been made in Antarctica, (Engelhardt and Kamb, 1998; Kamb, 2001) due to the difficulty of drilling operations and installation of the devices in the hundreds of meters-thick ice sheet.

Because of the high sensitivity of basal motion to subglacial water pressure, glacier flow could vary significantly under the influence of water input from the glacier surface. Iken and Bindshadler (1986) found that Findelengletcher accelerated by 50% when subglacial water pressure elevated after drainage of a large amount of surface water into the glacier bed. They found that the velocity increases with increasing water pressure. Flow velocity showed a relationship with an asymptote where the water pressure equals the ice overburden pressure, meaning that the velocity significantly increases when water pressure approaches the overburden pressure. The observed relationship is widely used to describe the variations in the basal sliding speed (u_b) associated with the subglacial water pressure (p_w) with the formulation of Weertman-Budd sliding law (Weertman, 1957; Budd and others, 1979) as:

$$u_b = k \frac{\tau_b^p}{(p_i - p_w)^q} = k \frac{\tau_b^p}{N^q}, \quad (1.1)$$

where τ_b is the basal shear stress, p_i is the ice overburden pressure, N is the effective pressure,

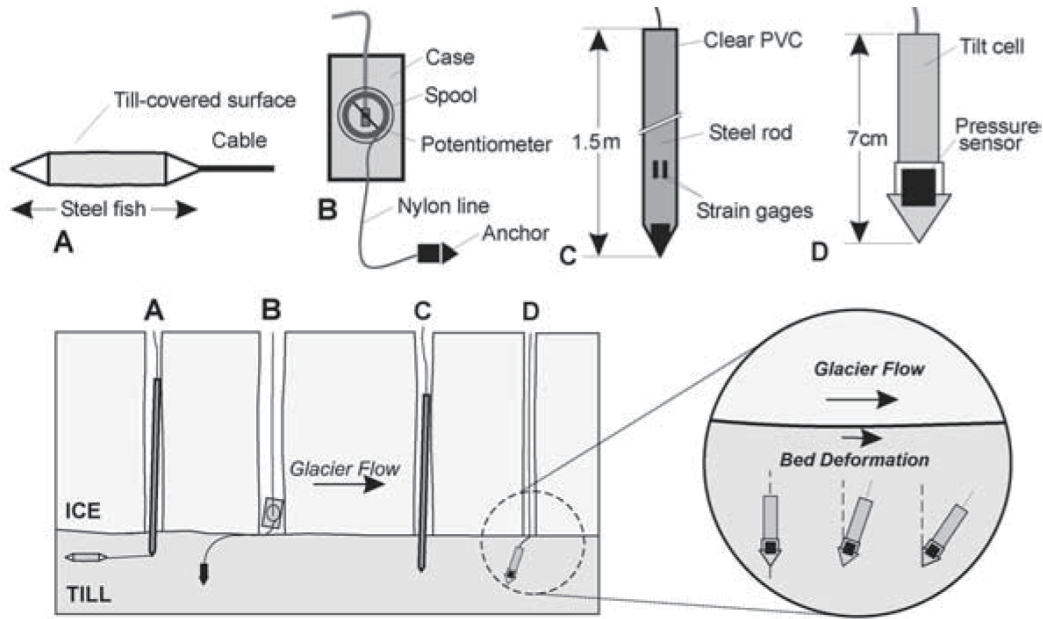


Figure 1.4: Approaches to measuring subglacial phenomena. In situ instruments from left to right: dragometer (A), drag spool (B), ploughmeter (C), tilt cells (D). Figure from Clarke (2005)

k , p , q are empirical parameters. Such controls of subglacial water pressure on glacier flow acceleration have been widely observed in alpine glaciers and the Greenland Ice Sheet from diurnal to seasonal time scales, as well as short-term speedup events after a large amount of water was injected into the basal hydrological system (e.g. Meier and others, 1994; Kamb and others, 1994; Sugiyama and others, 2011; Doyle and others, 2018; Andrews and others, 2018). However, researchers have assumed that such surface-to-bed hydraulic connections do not exist in Antarctica, owing to low ice temperature and limited amount of glacier surface melt (e.g. Dow, 2023). Indeed, the only previous observation of subglacial water pressure conducted in Willans Ice Stream has shown no short-term variations associated with meltwater supply from the surface (Engelhardt and Kamb, 1997).

In recent years, summer flow acceleration of outlet glaciers has been reported in the Antarctic ice sheet, based on the recently available high-resolution satellite dataset. Several studies showed seasonal flow changes on the grounded part of the glaciers in the Antarctic Peninsula (Tuckett and others, 2019; Boxall and others, 2022; Wallis and others, 2023). For example, Tuckett and others (2019) reported short-lived (<6 days) acceleration up to 100% greater than the annual mean on five outlet glaciers in the Antarctic Peninsula. The flow changes corresponded with sporadic increases in surface melt caused by Föhn. Therefore, subsequent delivery of meltwater into the glacier bed was assumed as a cause of basal lubrication and speedup. However, other factors have also been proposed and debated, such as artifacts in the satellite data analysis and the oceanic conditions in front of the glaciers (Rott and others, 2020). Boxall and others (2022) and Wallis and others (2023) found seasonal flow changes on outlet glaciers in the Antarctic Peninsula with summer acceleration by ~10–20% of the annual mean speed. Both surface meltwater and ocean-induced terminal position changes were

discussed as the factors controlling the seasonality, but no conclusions have been reached. All of the abovementioned studies have been conducted in the Antarctic Peninsula region, where the relatively warm climate generates a substantial amount of surface meltwater (Trusel and others, 2013). Recently, the widespread formation of a supraglacial hydrological system such as active channels and lakes has been found on the Antarctic ice sheet, and their impacts on the ice sheet stability have received increasing attention. Thus, there is a growing interest in whether such seasonality of glacier flow can be seen elsewhere in Antarctica.

1.4 Objectives of this study

The objective of this thesis is to understand the mechanisms of flow changes of outlet glaciers in the Antarctic ice sheet. To achieve this objective, satellite and field observations were conducted on five outlet glaciers flowing into Lützow-Holm Bay in East Antarctica.

In Chapter 3, satellite data analysis was carried out to investigate the glacier flow changes over the past two decades in the five outlet glaciers (Shirase, Skallen, Telen, Honnør, and Langhovde glaciers) and evaluate their link with land-fast sea-ice variations in Lützow-Holm Bay. To achieve this aim, terminus positions, flow velocity, surface elevations, and sea-ice conditions in front of the glaciers were analyzed using a satellite dataset.

In Chapter 4, field observations were conducted in Langhovde Glacier to observe short-term flow changes and investigate their mechanisms by subglacial measurements. To study a possible relationship between glacier basal motion and subglacial water pressure variations, a device equipped with a three-axis accelerometer and a water pressure sensor was developed. The device was installed at the glacier base using a 550-m long borehole drilled approximately 1 km upstream from the grounding line during a hot-water drilling campaign in the austral summer 2021/22.

Study site

2.1 Lützow-Holm Bay

This study focuses on five marine-terminating outlet glaciers flowing into Lützow-Holm Bay in East Antarctica: Shirase, Skallen, Telen, Honnør, and Langhovde glaciers (69.20–70.17° S and 39.8–38.95° E) (Fig. 2.2). The studied glaciers are located on the Soya Coast at about 20–110 km south of Syowa Station, a Japanese research base, in East Ongul Island. At the grounding line locations, the maximum ice speed ranges from 70 m a⁻¹ (Langhovde Glacier) to 2200 m a⁻¹ (Shirase Glacier), and the glacier width ranges from 2 km (Honnør Glacier) to 10 km (Shirase Glacier). Since the Japanese Antarctic Research Expedition began its activity in 1956 at East Ongul Island, a variety of research has been performed in the region. For glaciological studies, SMB (Endo and Fujiwara, 1973; Takahashi and others, 1994; Motoyama and others, 2008), ice thickness (Wada and Mae, 1981; Mae and Yoshida, 1987), and glacier flow (Nakamura and others, 2016, 2007, 2010) have been measured on the ice sheets and the glaciers flowing into Lützow-Holm Bay. Past ice sheet variations have also been inferred from geomorphological and geological studies on the exposed bedrock distributed at the ice sheet margin, suggesting the ice sheet thinned and retreated rapidly during the Early to Mid-Holocene (9–5 ka before present) (Kawamata and others, 2020; Suganuma and others, 2022). The streamlined landforms on the exposed bedrock were assumed to have been formed by erosion of subglacial water flow under the past ice sheet (Sawagaki and Hirakawa, 1997). A study also pointed out the possibility of outbursts from subglacial lakes as a source of the water (Sawagaki and Hirakawa, 2002).

Lützow-Holm Bay is usually covered with land-fast sea ice, which has shown quasiperiodic breakup events (Ushio, 2006) since the first observation in 1980 (Higashi and others, 1982). Atmospheric and oceanic conditions, including air temperature, southerly winds, sea surface temperature, SIC in the pack-ice zone, and snowfall on the sea ice, are deemed to be responsible for the breakup (Enomoto and others, 2002; Ushio, 2006; Aoki, 2017). The most recent breakup was observed in April 2016, when a large portion of the fast ice in the bay

disintegrated (Fig. 2.1) (Aoki, 2017). The event was attributed to relatively warm sea surface temperature and reduction of pack ice offshore from Lützow-Holm Bay, which resulted in the penetration of ocean swell into the fast-ice region in the bay (Aoki, 2017). Previous observations showed the influence of sea ice on outlet glaciers in Lützow-Holm Bay. During the retreat of the land-fast sea ice, an iceberg conglomerate detached from Shirase Glacier, suggesting that the icebergs were fastened by the sea ice (Nakamura and others, 2010; Aoki, 2017). Therefore, the breakup of land-fast sea ice in Lützow-Holm Bay possibly affected other outlet glaciers in the bay.

Hydrographic observations have also been conducted by the JARE activities in Lützow-Holm Bay (e.g. Ohshima and others, 1996). Recent ship-based ocean measurements found an inflow of mCDW through a submarine trough connecting the continental shelf break and the calving front of Shirase Glacier (Hirano and others, 2020). Phase-sensitive radar survey showed active basal melt (7.2 m a^{-1}) of the ice shelf of Shirase Glacier associated with the intrusion of the warm water. The warm water flux was shown to be controlled by the variability of the alongshore easterly wind over the continental slope (Hirano and others, 2020; Kusahara and others, 2021).

2.2 Shirase Glacier

Shirase Glacier is the fastest-flowing glacier in Lützow-Holm Bay. The ice shelf of Shirase Glacier disintegrates into large iceberg pieces, which aggregate and form a continuous floating ice tongue. The floating tongue significantly retreated six times in the period 1956–2016, and four of the events were associated with major breakups of land-fast sea ice in Lützow-Holm Bay (Pattyn and Derauw, 2002; Ushio, 2006; Nakamura and others, 2010; Aoki, 2017), suggesting the influence of sea-ice conditions on the glacier front position. Previous observations indicated that the breakup of land-fast sea ice also affected the ice flow of Shirase Glacier. Nakamura and others (2010) found a flow acceleration of the floating tongue in April–July 1998 that coincided with a major retreat of the land-fast sea ice in April 1998. After another sea-ice breakup in 2016, the glacier flow accelerated by 30% (Nakamura and others, 2022). The authors attributed the speedup to the loss of buttressing force exerted by the sea ice. These observations covered only the floating part of the glacier over periods of less than three consecutive years (1996–98 and 2016–18). Thus, a possible link between sea ice and the grounded-ice dynamics is not addressed. The glacier has shown a trend of mass gain for the last two decades, which was attributed to the reduction in the ice shelf basal melting associated with a weakening of mCDW intrusion (Hirano and others, 2020; Kusahara and others, 2021; Miles and others, 2023). The grounding line of the glacier was reported to have advanced at a rate of 14.7 m a^{-1} between 2010 and 2016 (Konrad and others, 2018).

Airborne and in-situ observations were conducted at Shirase Glacier by the Japanese Antarctic Research Expedition. The bedrock elevation of Shirase Glacier was measured by airborne

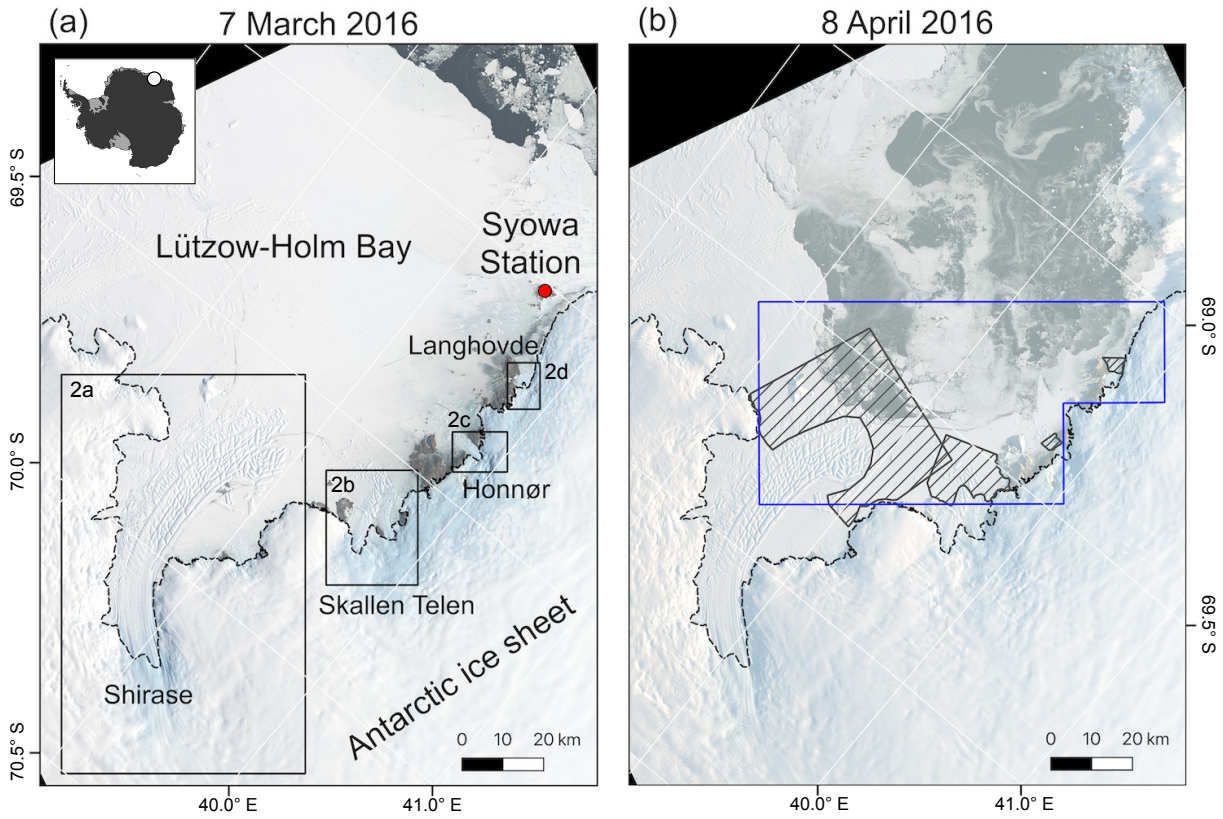


Figure 2.1: Landsat 8 OLI images of Lützow-Holm Bay acquired on (a) 7 March 2016 and (b) 8 April 2016, showing the studied area before and after the breakup of land-fast sea ice. The inset in (a) shows the location of the study site in Antarctica. The boxes in (a) indicate the studied glaciers and areas shown in Fig. 2. The hatched polygons in (b) show the regions of interest (ROIs) in front of the glaciers, where sea-ice conditions were analyzed using satellite images. Sea-ice concentration data from Climate Data Record of Passive Microwave Sea Ice Concentration, Version 4 (Meier and others, 2021) were analyzed in the region indicated by the blue polygon in (b). The dashed line indicates the grounding line estimated by Fukuda (2014) for Langhovde Glacier and Bindschadler and others (2011) for the other regions.

radar-echo soundings in the 1980s (Wada and Mae, 1981; Mae and Yoshida, 1987). The survey results showed that the ice thickness was 880 m near the grounding line, above which the bed elevation was below sea level for 30 km. At about 200 km from the grounding line, rapid thinning ($0.5\text{--}1.0\text{ m a}^{-1}$) was observed by triangulation chain surveys conducted in the period 1969–74 (Naruse, 1979). The thinning was attributed to stretching ice flow enhanced by acceleration due to increased basal sliding (Mae and Naruse, 1978). This observation demonstrated the importance of glacier dynamics on the mass budget of the ice sheet, since ice-flow changes of the outlet glaciers have a significant impact on the thickness of inland ice (e.g. Nick and others, 2009). Phase-sensitive radar measurement on the ice shelf revealed a basal melting of 7.2 m a^{-1} (Hirano and others, 2020). The rapid melting and concurrent hydrographic observations suggested an inflow of mCDW into the sub-shelf cavity through a submarine trough connecting the continental shelf break and the calving front of the glacier.

2.3 Langhovde Glacier

Among the five studied glaciers, we conducted field observations at Langhovde Glacier located at 20 km south from Syowa Station. The glacier front is 3 km wide and the flat surface extending to the 1.3–2.5 km upstream suggests the formation of an ice shelf (Fukuda, 2014). The lower 10 km forms a channel flowing at a rate of 130 m a^{-1} at the calving front and 70 m a^{-1} at the grounding line. Ice motion of the ice shelf was modulated by ocean tide (Sugiyama and others, 2014; Minowa and others, 2019), showing maximum/minimum horizontal speed during falling/rising tide. A number of supraglacial lakes and streams form throughout summer seasons on Langhovde Glacier (Langley and others, 2016). Rapid drainage of lake water through englacial channels has been observed on the ice shelf, but such englacial drainage has not been reported on the grounded ice.

Hot-water drilling has been carried out in austral summers 2011/12 and 2017/18 at Langhovde Glacier. Sugiyama and others (2014) explored 10–24-m-deep ocean cavity beneath 398–431-m-thick ice using boreholes drilled at the grounding zone in January 2012. The sub-shelf observations revealed a warm ocean-water (0.7°C above freezing) transport towards the grounding line. In the following hot-water drilling campaign in 2017/18, four boreholes were drilled on the ice shelf along its flowline. Sub-shelf ocean measurements revealed density-driven circulation under the ice shelf, leading to a mean basal melt rate of 1.42 m a^{-1} (Minowa and others, 2021). Despite the intensive drilling campaigns, no borehole observations had been conducted on the grounded part of Langhovde Glacier.

Satellite observations revealed that the glacier retreated by $\sim 200 \text{ m}$ in the summer of 2011 during a breakup event of land-fast sea ice in Lützow-Holm Bay, while no significant change was observed in the surface elevation from 2006 to 2012 (Fukuda and others, 2014).

2.4 Skallen, Telen, and Honnør glaciers

Glaciological studies of Honnør, Skallen, and Telen glaciers are scarce. The ice flow speeds of Skallen and Telen glaciers were measured during the period 2007–11 using satellite data (Shiramizu, 2018). Interannual velocity variations were reported, but the mechanisms of the changes remained unclear.

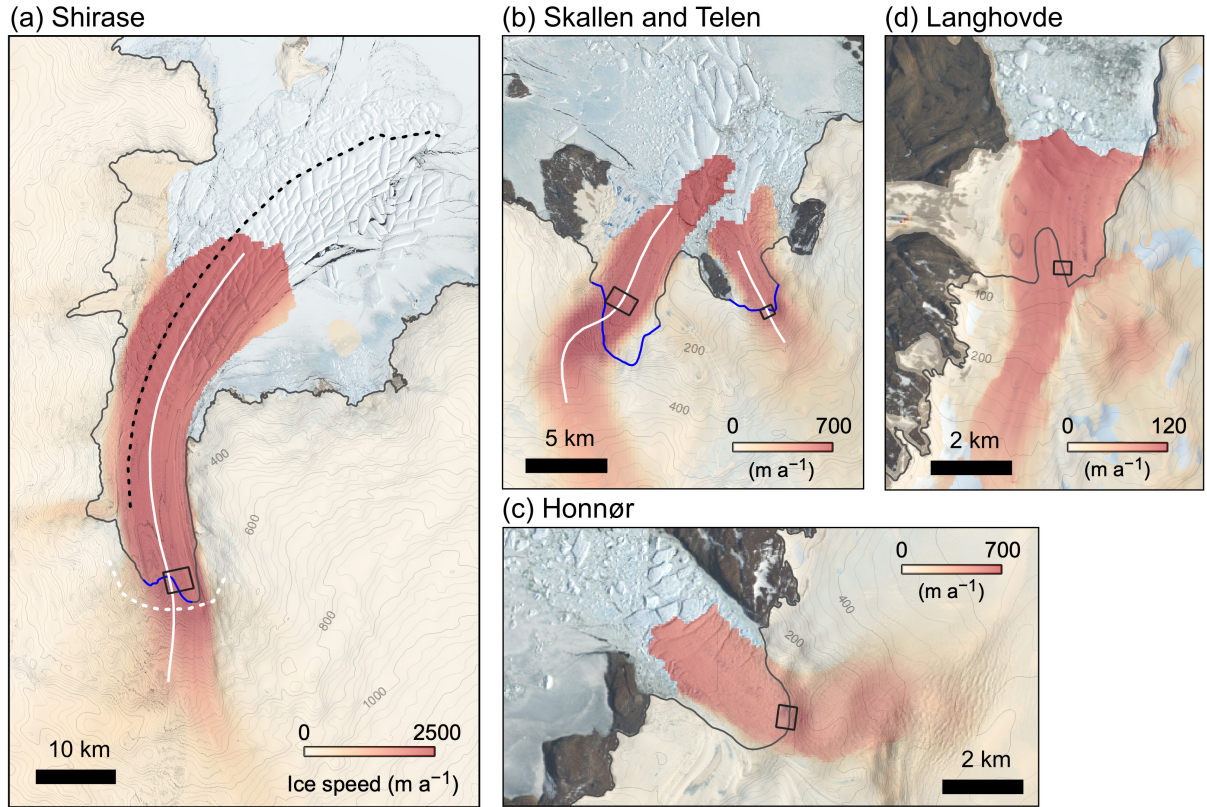


Figure 2.2: Landsat 8 OLI images acquired on 3 January 2016 showing the five studied glaciers flowing into Lützow-Holm Bay. The color scale shows the mean ice speed from 2001 to 2020. The solid black lines indicate the grounding line estimated by (a–c) Bindshadler and others (2011) and (d) Fukuda (2014). The blue lines in (a) and (b) indicate the grounding line estimated in this study, based on the change in surface slope observed with the Reference Elevation Model of Antarctica Strip digital elevation models acquired in 2013 and 2012, respectively. The boxes near the grounding line indicate areas used for the calculation of the ice speed shown in Fig. 3.4. The dotted black line in (a) shows an example of a flowline used for the length measurement of Shirase Glacier. The solid white lines in (a) and (b) indicate the central flowlines used to plot the ice speeds shown in Fig. 3.14. The dotted white line in (a) shows the flux gate used for the calculation of ice mass discharge from Shirase Glacier.

Satellite observations: Influence of sea ice on glacier changes

The content of this chapter was published in Journal of Glaciology (Kondo and Sugiyama, 2023):

Kondo K and Sugiyama S (2023) Calving, ice flow, and thickness of outlet glaciers controlled by land-fast sea ice in Lützow-Holm Bay, East Antarctica. Journal of Glaciology, First View, 1–13. doi: 10.1017/jog.2023.59

In this chapter, satellite data analysis was carried out to investigate glacier changes over the past two decades in the five outlet glaciers (Shirase, Skallen, Telen, Honnør, and Langhovde glaciers) and evaluate their link with land-fast sea-ice variations in Lützow-Holm Bay. To achieve this aim, terminus positions, flow velocity, surface elevations, and sea-ice conditions in front of the glaciers were analyzed using satellite dataset.

3.1 Data and Methods

3.1.1 Terminus position

The frontal margins of the studied glaciers were mapped during the period 1988–2020, using satellite images obtained by the Japanese Earth Resources Satellite-1 (JERS-1) synthetic aperture radar (SAR), Advanced Spaceborne Thermal Emission and Reflection (ASTER) radiometer, Landsat 4 Thematic Mapper (TM), Landsat 7 Enhanced Thematic Mapper Plus (ETM+), and Landsat 8 Operational Land Imager (OLI). Relatively large geolocation errors due to inaccurate co-registration were found in the JERS-1 SAR and Landsat 4 TM images. These images were corrected against the Landsat 8 OLI image acquired on 11 March 2020, by adjusting the

coordinates of commonly observed bedrock features with the horizontal shift of 0.3–0.7 km. I utilized backscatter images from JERS-1 SAR, the visible and near-infrared band 3N from ASTER, true-color composites from Landsat 4 TM (bands 1–3), Landsat 7 ETM+ (bands 1–3), and Landsat 8 OLI (bands 2–4).

In each image, the glacier fronts were manually delineated using geographic information software QGIS or Google Earth Engine Digitization Tool (Lea, 2018). The mapping was performed 40–65 times for each glacier, by analyzing 125 images in total. To obtain mean frontal displacement, a change in the ice surface area near the front was divided by the glacier width. To take into account the influence of curved flowline, I employed the curvilinear box method proposed by Lea and others (2014). Because it was difficult to distinguish the calving front of Shirase Glacier from the icebergs aggregated near the front, I measured the length of the floating ice tongue following Ushio and others (2006); Aoki (2017). The distance from the grounding line to the forefront of the iceberg aggregate was measured along a flowline as shown by the example in Fig. 2.2a.

Temporal resolution of the measurement was dependent on data availability. After a 5-year gap from 1989 to 1994, the resolution was approximately one year from 1994 to 2003 and finer from 2003 to 2020. To precisely measure the timing and magnitude of glacier retreat, all available images were used for mapping during the periods of retreat. The uncertainty in the measurement was estimated by mapping an ice-free coastline using 3–18 images for each sensor. The standard deviation of the coordinates along the coastline was 27 m for JERS-1, 21 m for ASTER, 21 m for Landsat 4, 24 m for Landsat 7, and 7 m for Landsat 8.

3.1.2 Ice flow velocity

The annual flow velocity of Shirase, Skallen, and Telen glaciers was obtained for the period 2001–18 from the ITS_LIVE dataset (Gardner and others, 2019). The dataset provides velocity fields with a spatial resolution of 240 m, derived by applying a feature tracking algorithm auto-RIFT (Gardner and others, 2018) to the Landsat 4, 5, 7, and 8 imagery. Mean annual speeds near the grounding lines were obtained as a mean over the boxes shown in Fig. 2.2. Speed profiles along the central flowlines shown in Figs 2a and b were also analyzed.

To study the details of the variations in velocity, I applied a feature tracking algorithm (Sakakibara and Sugiyama, 2014) to 125 image pairs of Landsat 7 ETM+ and Landsat 8 OLI imagery acquired over the period 2006–20. The objectives of this measurement were (1) to obtain annual velocity fields of the smaller Langhovde and Honnør glaciers, which could not be resolved by the ITS_LIVE dataset, (2) to improve the spatial and temporal resolutions of the velocity of Shirase, Skallen, and Telen glaciers, and (3) to measure sea-ice displacement near the termini of Shirase, Skallen, and Telen glaciers. The temporal separations of the image pairs ranged from 16 to 416 days. I used an automatic image matching scheme known as the orientation correlation method in a frequency domain (Heid and Käab, 2012; Sakakibara and

Sugiyama, 2020). Horizontal biases on the image pairs were removed by minimizing displacement vectors obtained in off-ice areas. I excluded vectors obtained with a low signal-to-noise ratio (<30) and those that deviated by more than 30° or 100 m a^{-1} from a median vector within 3×3 neighboring pixels (Heid and Kääb, 2012; Sakakibara and Sugiyama, 2014). Errors in velocity measurement are associated with (1) ambiguities in the cross-correlation peak, (2) co-registration errors and (3) false correlations (Howat and others, 2010). I evaluated (1) and (2) from the root-mean-square errors of horizontal displacement in off-ice areas where displacement was assumed to be zero. During the calculation of the root-mean-square errors, displacement greater than twice the pixel size was attributed to a false correlation and therefore excluded. Errors caused by (3) were minimized by applying a 3×3 median low-pass filter. Error estimates ranged between ± 1 and $\pm 115 \text{ m a}^{-1}$ with a mean value of $\pm 31 \text{ m a}^{-1}$. The error was dependent on the temporal separation of the image pairs, i.e. uncertainty was greater for shorter temporal separation. The magnitude of the velocity obtained near the grounding line of Langhovde Glacier was consistent with that of the ITS_LIVE dataset (within $\pm 5\%$). Ice speed was averaged in space and time within the boxes near the grounding lines shown in Fig. 2.2 to obtain mean annual speeds during the periods 2014–20 for Langhovde Glacier and 2006–20 for Honnør Glacier. I also utilized satellite-derived flow speeds reported by Fukuda (2014) for the region near the grounding line of Langhovde Glacier from 2003 to 2012.

Ice discharge from Shirase Glacier across a flux gate near the grounding line (Fig. 2.2a) was calculated for the years 2013 and 2015. Ice speed from the ITS_LIVE dataset was used with the assumption of a full slip condition. Ice thickness was computed using surface elevation from DEMs on 6 January 2013 and 15 March 2016 (see Section 3.1.3 for details) and bed elevation from BedMachine Antarctica, Version 3 (Morlighem and others, 2020; Morlighem, 2022). An ice density of 910 kg m^{-3} was assumed to convert the volume to mass flux. The uncertainty in the discharge was estimated at 10% according to the errors expected in the ITS_LIVE ($<1\%$) and BedMachine datasets (10%).

Changes in flow regime affect vertical strain rate. To investigate this effect, the logarithmic strain rate (Nye, 1959) was calculated for Shirase, Skallen, and Telen glaciers, using the annual glacier velocity field from the ITS_LIVE dataset. The calculation was performed at each velocity pixel, by employing the method and code provided by (Alley and others, 2018).

3.1.3 Glacier surface elevation

Changes in the surface elevation on Shirase, Skallen, and Telen glaciers were measured with digital elevation models (DEMs) from the Reference Elevation Model of Antarctica (REMA) Strip (Howat and others, 2019) using all available DEMs during 2012–16. The REMA Strip is time-stamped high-resolution DEMs generated from stereophotogrammetry with submeter (0.32–0.5 m) resolution optical satellite imagery such as GeoEye-1 and WorldView-1/2/3 acquired between 2009 and 2021 (Howat and others, 2019). I compared DEMs representing the

Table 3.1: REMA Strip DEMs used in this study. The standard deviations of the DEMs from the reference (the REMA Mosaic DEM and CryoSat-2) are shown (σ_e).

Region	Period	Acquisition date	σ_e (m)
Shirase	2012/13	25 December 2012	0.9
		6 January 2013	2.4
	2015/16	15 March 2016	1.4
Skallen/Telen	2012/13	7 December 2012	0.3
	2015/16	12 October 2015	0.9
		16 October 2015	0.8

austral summers 2012/13 and 2015/16, which were generated by averaging DEMs available from October 2012–March 2013 and October 2015–March 2016, respectively (Table 3.1).

To homogenize the REMA Strips distributed with a grid spacing of 2 m or 8 m, those with a 2-m grid were resampled to 8 m by taking the average of the neighboring pixels. Horizontal and vertical shift, and elevation dependent biases were minimized by the co-registration and bias correction procedures introduced by Nuth and Kääb (2011) and Levinsen and others (2013). For this procedure, I used the REMA Mosaic DEM in off-ice areas and CryoSat-2 altimetry data over the ice sheet. The REMA Mosaic DEM was used as the elevation reference after excluding pixels from steep terrain ($>30^\circ$). The CryoSat-2 altimetry data obtained at the point of closest approach is distributed by the European Space Agency CryoTEMPO project (Gourmelen and others, 2018). I utilized the altimetry data obtained within ± 100 days from the DEM acquisition dates. All the elevation data was referenced to WGS-84 ellipsoid. The co-registration was performed in the region below 800 m. The standard deviation of the elevation of the REMA Strip DEMs against the REMA Mosaic DEM and CryoSat-2 altimetry data was between 0.3 and 2.4 m (Table 3.1) (Fig. 3.1).

I also utilized the Ice, Cloud, and land Elevation Satellite (ICESat-1) and ICESat-2 data for measurements of the glacier surface elevation in 2003/04 and 2019/20, respectively. The instruments have been operated to measure the surface elevation of the ice sheet by laser altimetry from 2003 to 2009 for ICESat-1 and from 2018 to the present for ICESat-2. The GLA06 version 34 product of ICESat-1 (Zwally and others, 2014) obtained between October 2003 and October 2004 and the ATL06 Land Ice Height version 3 product of ICESat-2 (Smith and others, 2020a) obtained between May 2019 and June 2020 were downloaded from OpenAltimetry (Khalsa and others, 2020). The accuracy of the altimetry was 0.15 m for ICESat-1 (Shuman and others, 2006) and 0.03 m for ICESat-2 (Brunt and others, 2019).

The DEMs were analyzed to estimate the grounding line positions. As it has previously been proposed that the surface elevation profile near the grounding line is accompanied by a slope inflection point (Fricker and Padman, 2006), I delineated the inflection points of the surface slope on the DEMs using QGIS. The mean displacement of the grounding line between 2012/13 and 2015/16 was calculated by the box method as described in Section 3.1.1.

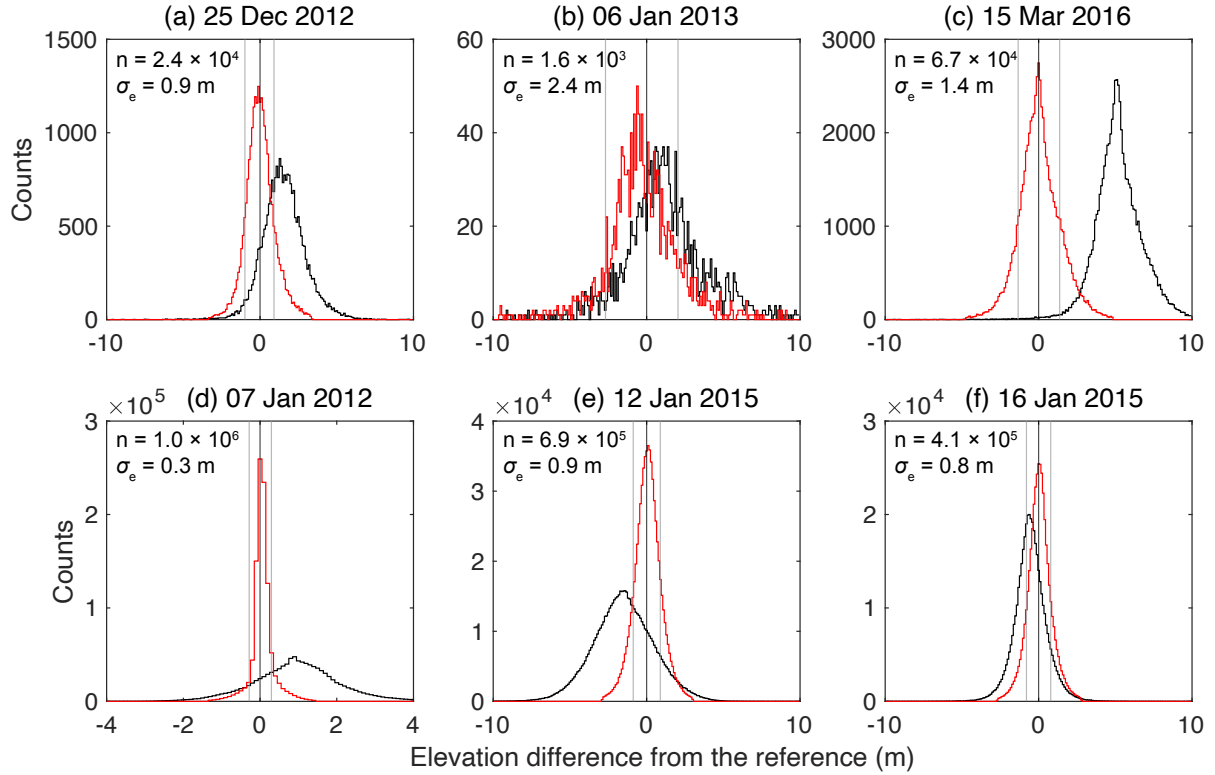


Figure 3.1: Elevation differences of the REMA Strip DEMs from the reference before (black) and after (red) the co-registration. The pixels numbers (n) and the standard deviations (σ_e) were shown. The vertical gray lines show the standard deviations of the DEMs from the reference after the co-registration.

3.1.4 Sea ice variations

Open water and ice-covered ocean areas were measured from 2000 to 2020 in the regions of interest (ROIs) defined near the glacier fronts as shown in Fig. 2.1b. The same ROI was given to Skallen and Telen glaciers, due to their proximity. The ASTER L1T Radiance and Landsat 7/8 Collection 1 top of atmosphere (TOA) reflectance products with a spatial resolution of 30 m were analyzed in Google Earth Engine for this purpose. The ASTER L1T Radiance was converted to TOA reflectance using coefficients provided by the distributors. I analyzed between 538 and 778 images for each glacier to classify pixels within the ROIs as being open water, ice, or cloud. The analysis used optical satellite imagery, thus covering only the summer period from September to March. Thresholds for the classification were determined for each satellite, based on a manually classified dataset (Fig. 3.2). The manual classification was performed for 27 images for ASTER (10.0×10^6 pixels), 17 for Landsat 7 (19.7×10^6 pixels), and 17 for Landsat 8 (20.5×10^6 pixels), which covered a wide range of illumination conditions and seasons.

Before the ocean surface classification took place, cloud pixels were eliminated based on the TOA reflectance at the shortwave infrared (SWIR) band and the Normalized Difference Snow Index (NDSI). Cloud pixels were selected for their high reflectance in the SWIR band (Moussavi and others, 2020), whereas NDSI was able to distinguish clouds from snow surface

(Hall and others, 1995; Moussavi and others, 2020). Pixels were regarded as cloud when the TOA SWIR reflectance and NDSI were >0.1 and <0.65 for ASTER and >0.06 and <0.8 for Landsat 7/8, respectively. This procedure eliminated 98% of the pixels manually classified as cloud (Fig. 3.2). After removing scenes with $>20\%$ cloud cover over the ROIs, the remaining pixels were classified either as open water or as ice, based on TOA reflectance at the green band for ASTER and the panchromatic band for Landsat 7/8, given the distinct differences in the reflectance between water and ice in the optical band. For all the sensors, pixels with a reflectance ≥ 0.25 were classified as ice (sea ice, iceberg, or ice mélange), whereas those with <0.25 were classified as open water. The results obtained by this procedure concurred with manual classification in the case of 99% of the pixels (Fig. 3.2).

Sea-ice conditions in Lützow-Holm Bay were analyzed by the Climate Data Record of Passive Microwave Sea Ice Concentration, Version 4 (Meier and others, 2021) provided by the National Oceanic and Atmospheric Administration's (NOAA) National Snow and Ice Data Center (NSIDC). In this dataset, SIC measured by passive microwave sensors has been distributed with a 25-km grid from 1978 to the present. We averaged monthly SIC data obtained at seven data cells, which covered the area indicated by the blue line in Fig. 2.1b. Monthly anomalies were calculated using the mean from 1988 to 2020 as a reference.

3.1.5 Meteorological data

The mean summer air temperature (for December, January, and February) was calculated for the period 1988–2020, using hourly data measured at Syowa Station (69.00° S, 39.58° E, 29 m a.s.l.) and distributed by the Japan Meteorological Agency. The station is located on the coast of Lützow-Holm Bay, 20–110 km north of the studied glaciers (Fig. 2.1a).

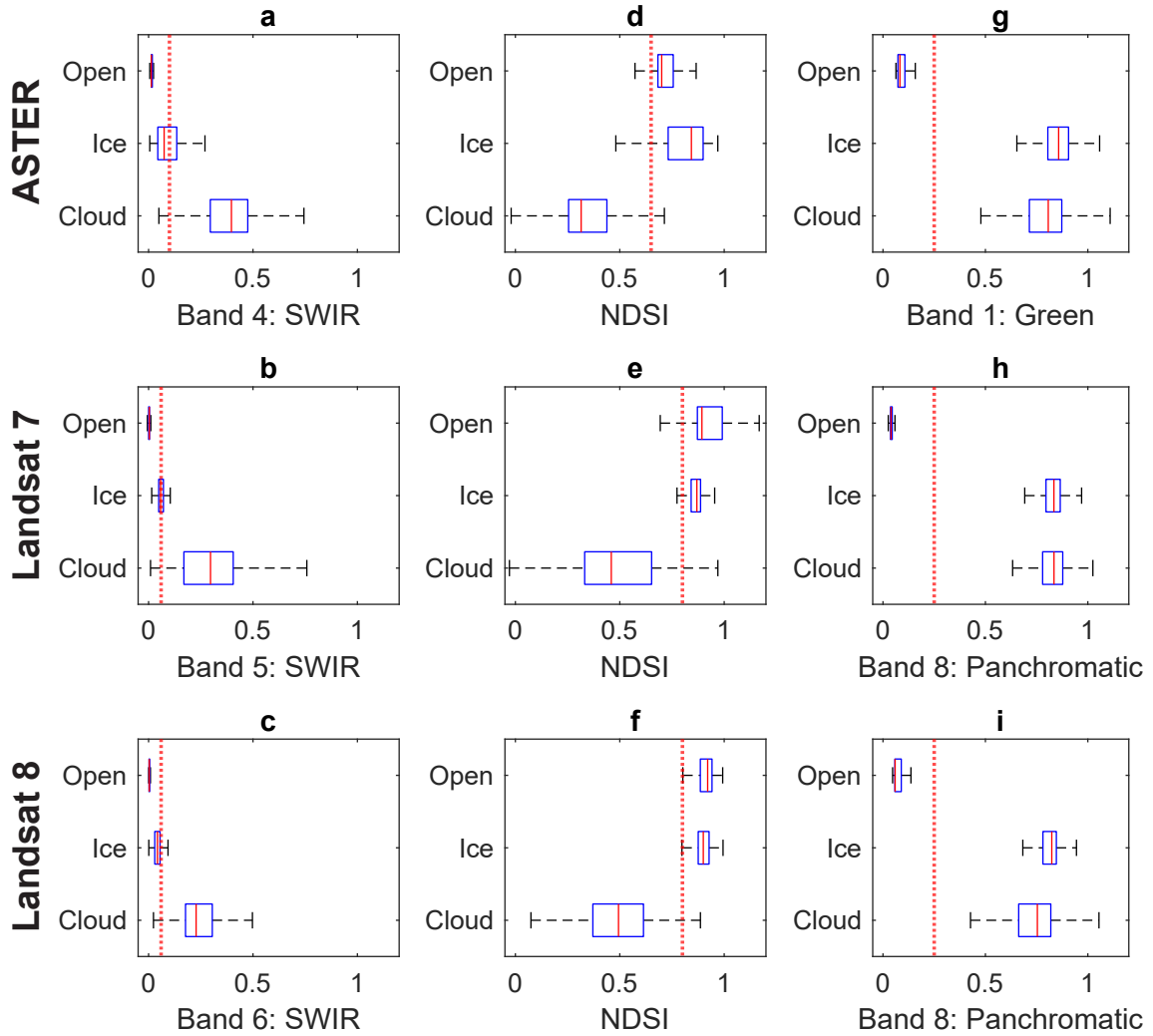


Figure 3.2: Box plots of the values from the training dataset for (a–c) shortwave infrared bands, (d–f) NDSI, and (g–i) visible (green or panchromatic) bands for ASTER, Landsat 7, and Landsat 8 imagery. The vertical dotted lines show the threshold used to eliminate cloud pixels (a–f) and classify ice and open water pixels (g–i).

3.2 Results

3.2.1 Terminus position and sea ice conditions

The studied glaciers showed similar frontal variations, characterized by a steady advance interrupted by at least three retreat events (Figs 3.3a–e). From 1994 to 1997, the studied glaciers advanced steadily without noticeable retreat, with the exception of Honnør Glacier, which retreated by 0.9 km in 1995–96. During 1997–98, Langhovde Glacier retreated by 0.5 km, and the other glaciers retreated in the period 1998–2000, resulting in the removal of 130 km² of glacier ice in total. After the retreat between 1997 and 2000, all the glaciers advanced from 2000 to 2016. Shirase Glacier advanced until the disintegration of a 1.5-km-long floating tongue in the period December 2012–November 2013 (Fig. 3.3a). Skallen Glacier steadily advanced until 2016, except for a retreat of 2.1 km in the period February 2003–March 2006 (Fig. 3.3b). The other three glaciers were relatively stable between 2000 and 2008, followed by an advance from 2008 to 2016 without significant loss of ice due to calving (Figs 3.3c–e). In 2016–18, all the studied glaciers retreated simultaneously at the greatest rate recorded during the observation period. Between March and October 2016, Shirase and Langhovde glaciers retreated by 6.0 and 0.4 km, respectively (Figs 3.3a and e). Skallen and Honnør glaciers retreated between October and December 2016 by 1.8 and 0.8 km, respectively (Figs 3.3b and d). These four glaciers continued to retreat until the summer of 2018. Telen Glacier retreated by 2.2 km slightly later - between February and September 2018 (Fig. 3.3e). In total, 200 km² of ice was lost by the studied glaciers over the period 2016–18. In 2019–20, Skallen and Langhovde glaciers further retreated by 0.9 and 0.3 km, respectively (Figs 3.3b and e).

The observed glacier front variations were associated with variations in SIC (Figs 3a–e). In general, the open water area dropped below 10% between September and November and peaked between January and March. During the summer months of 2000–08, when the glacier front positions were relatively stable, the fraction of the open water area increased to >20% near the front of Honnør and Langhovde glaciers (Figs 3d and e). From January to February 2005, open water in front of Shirase, Skallen/Telen, Honnør, and Langhovde glaciers increased to 8, 36, 99, and 99%, respectively. Sea ice in Lützow-Holm Bay recovered after 2008. From 2009 to 2015, the observation of open water areas was consistently lower than 5%, except for an increase at Honnør Glacier in the summer 2010 (21%) and 2013 (40%). The period of stable sea ice corresponded with a simultaneous advance of the glacier fronts (Figs 3a–e). The most significant sea-ice loss during the period of this analysis was the breakup of land-fast ice in April 2016 (Fig. 3.3f). The greatest extent of open water over the studied period was recorded in March 2016 at Shirase Glacier (10%) and February 2017 at the other glaciers (91% at Skallen/Telen, 100% at Honnør and Langhovde). The period of sea-ice loss coincided with the onset of rapid glacier retreat since 2016. Near Shirase Glacier, the open water area increased to 6% in the summer immediately after the sea-ice breakup (2016/17), followed by

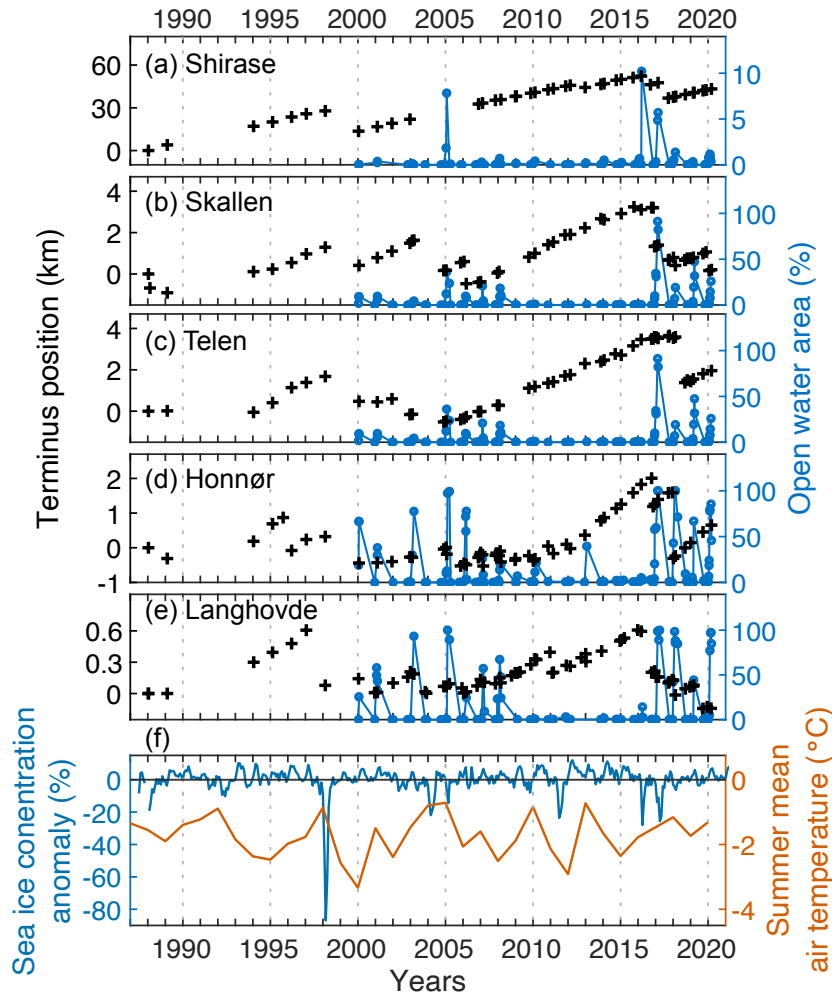


Figure 3.3: Terminus position relative to 1988 (crosses) and fraction of open water area within the ROIs shown in Fig. 1b (circles) for (a) Shirase, (b) Skallen, (c) Telen, (d) Honnør, and (e) Langhovde glaciers. (f) Monthly sea-ice concentration anomaly in Lützow-Holm Bay (blue), mean summer (December, January, and February) air temperature at Syowa Station (orange).

more stable sea-ice conditions (<2% open water) from 2018 to 2020 (Fig. 3a). Near the ice front of Honnør and Langhovde glaciers, open water expanded to >85% in the summer of 2017, 2018, and 2020, indicating that the impact of the breakup of land-fast sea ice in 2016 have persisted over the following four years (Figs 3.3d and e).

3.2.2 Glacier flow velocity

The annual ice speed of each glacier varied within 7–18% of its mean over the period of the study. Shirase, Skallen, and Telen glaciers showed similar flow variations in terms of timing and magnitude, with a slowdown from 2011 to 2015 and a speedup after 2016 (Figs 3.4a–c), whereas smaller, more frequent variations were observed at Honnør and Langhovde glaciers (Figs 3.4d and e). The ice speed at the grounding line of Shirase Glacier decreased from 2002 to 2015 by 180 m a^{-1} (Fig. 3.4a), with the most substantial slowdown of 80 m a^{-1} from 2013 to 2015. Skallen and Telen glaciers decelerated in 2011–15 (-90 and -80 m a^{-1}) after a short

period of acceleration in 2009–10. At Shirase, Skallen and Telen glaciers, the slowest speeds over the study period were observed in 2015 (2180, 620, and 520 m a^{-1} , respectively). By 2018, these three glaciers had accelerated to 2250, 670, and 560 m a^{-1} , respectively. Honnør Glacier decelerated from 2009 to 2010, after a relatively stable speed during 2007–09 (Fig. 4d). The speed increased from 420 to 470 m a^{-1} from 2012 to 2014. Thereafter, the glacier speed was stable within 450–480 m a^{-1} until 2020. The speed of Langhovde Glacier showed fluctuations with relatively fast flow ($>100 \text{ m a}^{-1}$) in the years 2003–04, 2009–10, 2014–15, and 2018–19 (Fig. 3.4e).

Spatial patterns of flow acceleration show further details of the synchronous ice speed changes observed at Shirase, Skallen, and Telen glaciers (Fig. 3.5). From 2007 to 2011, only small changes were observed in limited areas near the glacier fronts (Figs 3.5a and d). During this period, the rates of mean speed changes along the central flowline of each glacier (white lines in Figs 3.5a and d) were within $\pm 5 \text{ m a}^{-2}$. Substantial changes were observed in 2011–15 and 2015–18 in the areas downstream of the grounding lines. Shirase Glacier decelerated in 2011–15, with the most pronounced changes (-60 m a^{-2}) observed near the glacier front (Fig. 3.5b). Thereafter, the terminus area accelerated in the period 2015–18 at a rate of 80 m a^{-2} (Fig. 3.5c). These speed changes extended upstream of the grounding lines. For example, the speed 10 km upstream of the grounding line decreased by -28 m a^{-2} in 2011–15 and increased by 23 m a^{-2} in the period 2015–18. Similarly, Skallen and Telen glaciers showed the greatest speed changes downstream of the grounding lines, and large variations were also observed upstream of the grounding line (Figs 3.5e and f). Along the central flowlines, a deceleration in the years 2011–15 (-17 and -15 m a^{-2} for Skallen and Telen, respectively) was followed by an acceleration in the period 2015–18 (11 m a^{-2} for both glaciers).

To better understand glacier acceleration observed after 2015, displacement of sea ice near the fronts of Shirase, Skallen, and Telen glaciers were analysed by applying the feature tracking algorithm to the Landsat 8 OLI images acquired from October 2014 to February 2016 (Fig. 3.6). Sea-ice displacement prior to the breakup in 2016 was substantially smaller than the forward motion of the glacier terminus. For example, the displacement of the sea ice 10 km off Shirase Glacier was 50% of the glacier front motion (Fig. 6b). From January to February 2016, the displacement rate of the sea ice increased by 1000 m a^{-1} , reaching approximately 90% of the glacier speed (Fig. 3.6c). A similar speedup was observed on the sea ice near the front of Skallen Glacier, showing a two-fold increase from the summer 2014/15 to 2015/16 (Figs 3.6e and f).

3.2.3 Surface elevation

Near the grounding line of Shirase Glacier, surface elevation increased from 2003 to 2012/13 (Figs 3.7a and 8a). The mean elevation change along a track of ICESat-1 located 0–5 km downstream of the grounding line (AA' in Fig. 3.7b) was $5.9 \pm 0.3 \text{ m}$. The elevation increase

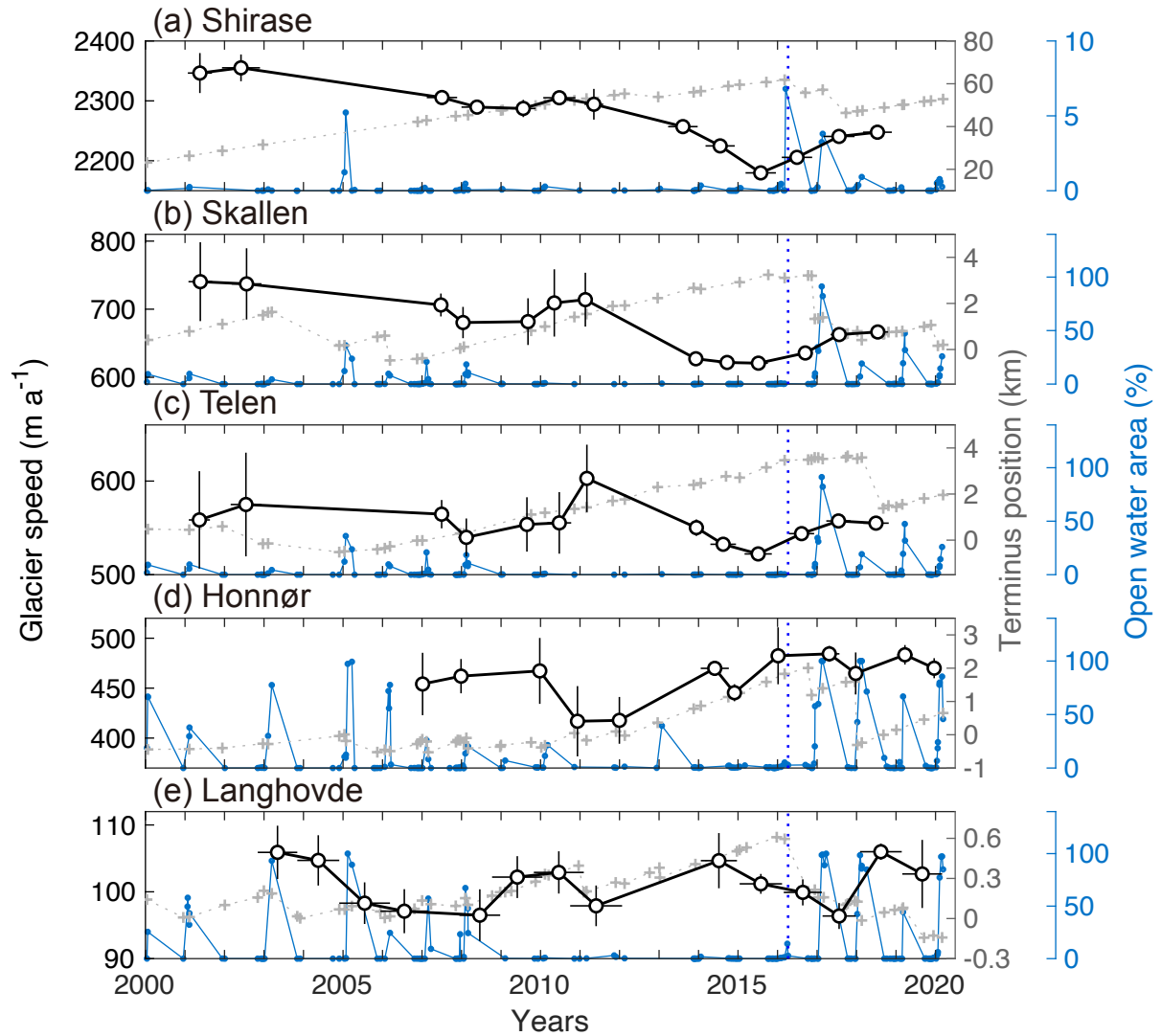


Figure 3.4: Mean annual glacier speeds (open circles) at the grounding line (averaged in the polygons in Fig. 2.2), terminus positions relative to 1988 (crosses), and the fraction of open water areas within the ROIs shown in Fig. 2.1b (dots) at (a) Shirase, (b) Skallen, (c) Telen, (d) Honnør, and (e) Langhovde glaciers. The length of the horizontal and vertical lines associated with the markers show the mean time separation of image pairs used for velocity calculations and uncertainty in glacier speed, respectively. The vertical dotted line indicates the timing of the land-fast sea ice breakup in April 2016.

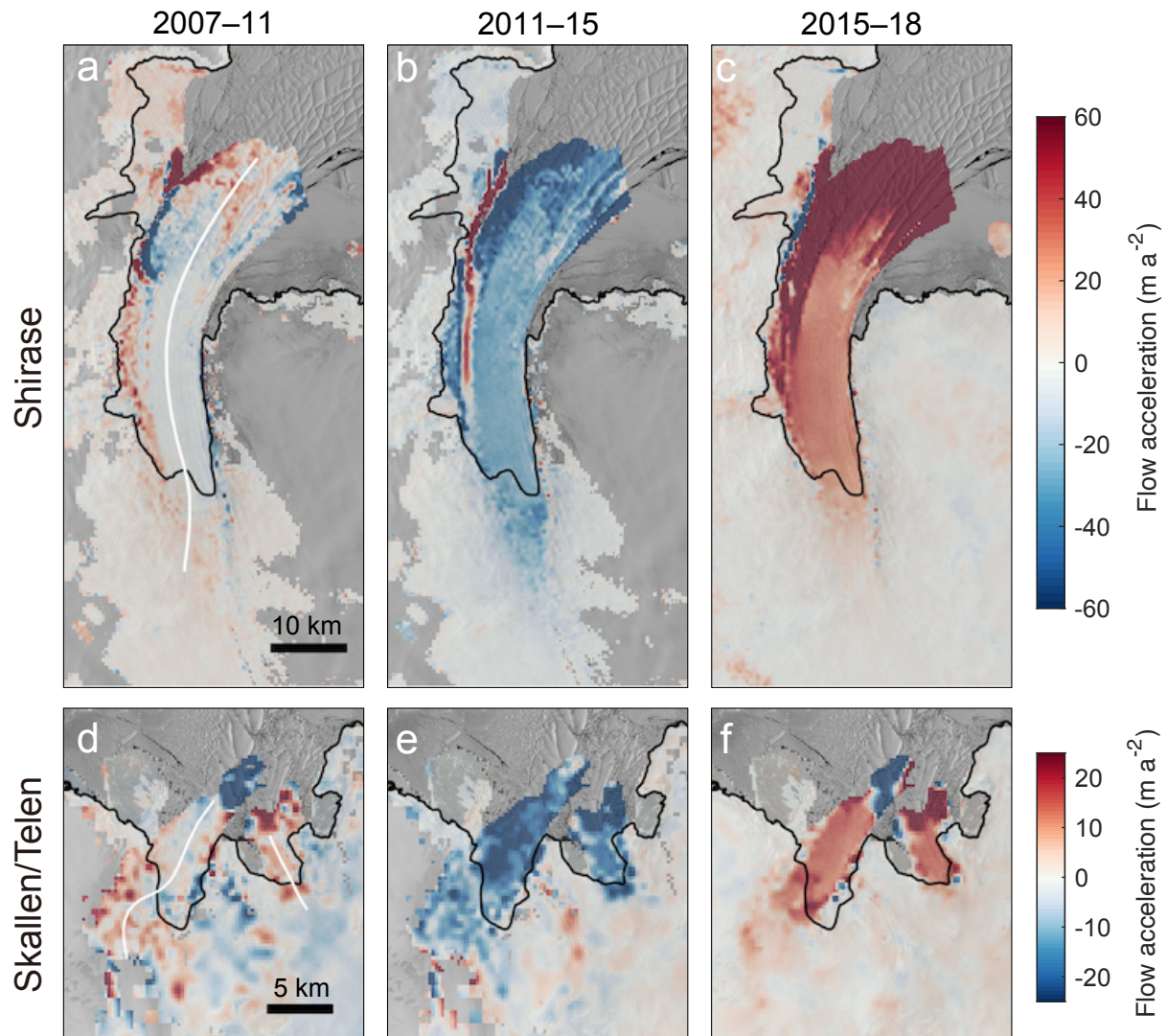


Figure 3.5: Ice speed acceleration from the ITS_LIVE dataset at (a–c) Shirase and (d–f) Skallen/Telen glaciers during the periods (a and d) 2007–11, (b and e) 2011–15, and (c and f) 2015–18. The background is a Landsat 8 OLI image acquired on 3 January 2016. White lines (a and d) show central flowlines used to measure the ice speed changes described in the text.

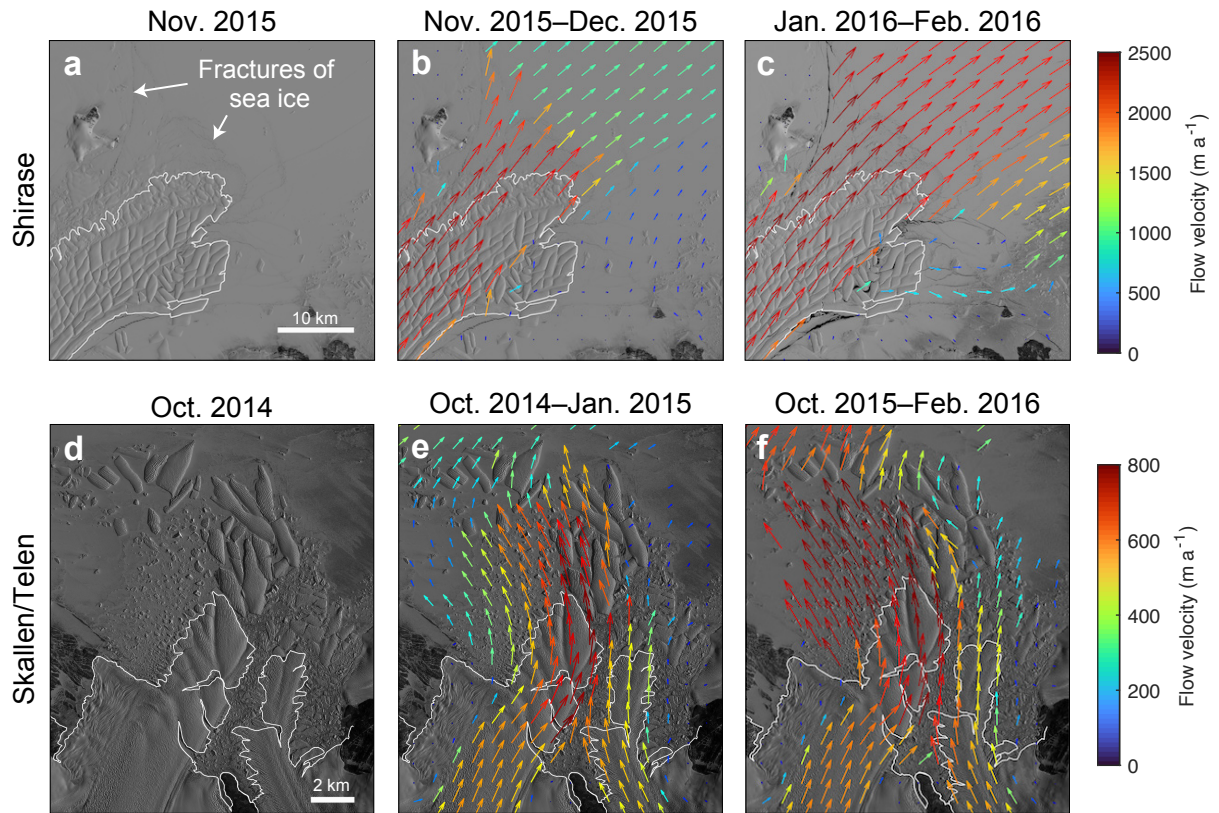


Figure 3.6: Landsat 8 OLI images showing sea ice and glacier fronts of (a–c) Shirase and (d–f) Skallen/Telen Glaciers. The images were acquired on (a and b) 16 November 2015, (c) 3 January 2016, (d and e) 3 October 2014, and (f) 8 October 2015. The color-coded vectors (b, c, e, and f) are glacier velocity fields from the indicated periods. The flow vectors are twice as large as the actual distance. The white lines show the glacier margins.

was more rapid in the following period (2012/13–16). The magnitude of the mean changes was 3.1 ± 0.3 m along AA' and 5.5 ± 0.1 m along the ICESat-2 track BB' located 0–6.5 km downstream of the grounding line (Fig. 3.7b and 3.8). An even greater increase of 6.5 ± 0.2 m was observed as a mean along the ICESat-2 track CC' crossing the grounding line. The greatest elevation change along the track AA' from 2003 to 2016 was 27 ± 1 m (2.2 ± 0.1 m a^{-1}) at 3.5 km from the western margin (Fig. 3.8a). In the following period (2016–20), the elevation increase continued at a reduced rate along BB' (mean of 1.60 ± 0.06 m) (Figs 3.7c and 3.8b), whereas the surface dropped along CC' (mean of -0.8 ± 0.1 m) (Fig. 3.7c). A large amount of positive and negative elevation changes were observed 15 km downstream of the grounding line. These changes were attributed to the advection of surface topographic features (e.g. bumps and depressions).

A substantial change in surface elevation was also observed on Skallen Glacier. Along the ICESat-1 track crossing the grounding line (DD' in Fig. 3.7), the mean surface elevation change from 2004 to 2012 was 0.87 ± 0.06 m (Fig. 3.7d). Greater changes were observed in the period 2012–15, reaching 2.4 ± 0.2 m as a mean along DD' (Fig. 3.7e). In the following period (2015–19), mean elevation decreased by 3.03 ± 0.07 m along EE' (Fig. 3.7f). Telen Glacier showed a similar trend to Skallen Glacier. Along the ICESat-2 track FF' located 0–1 km downglacier from the grounding line, the surface elevation increased in the period 2012–15 and decreased in the period 2015–19 (Figs. 3.7e and f). The changes observed on Telen Glacier were smaller in magnitude than those observed on Shirase and Skallen glaciers.

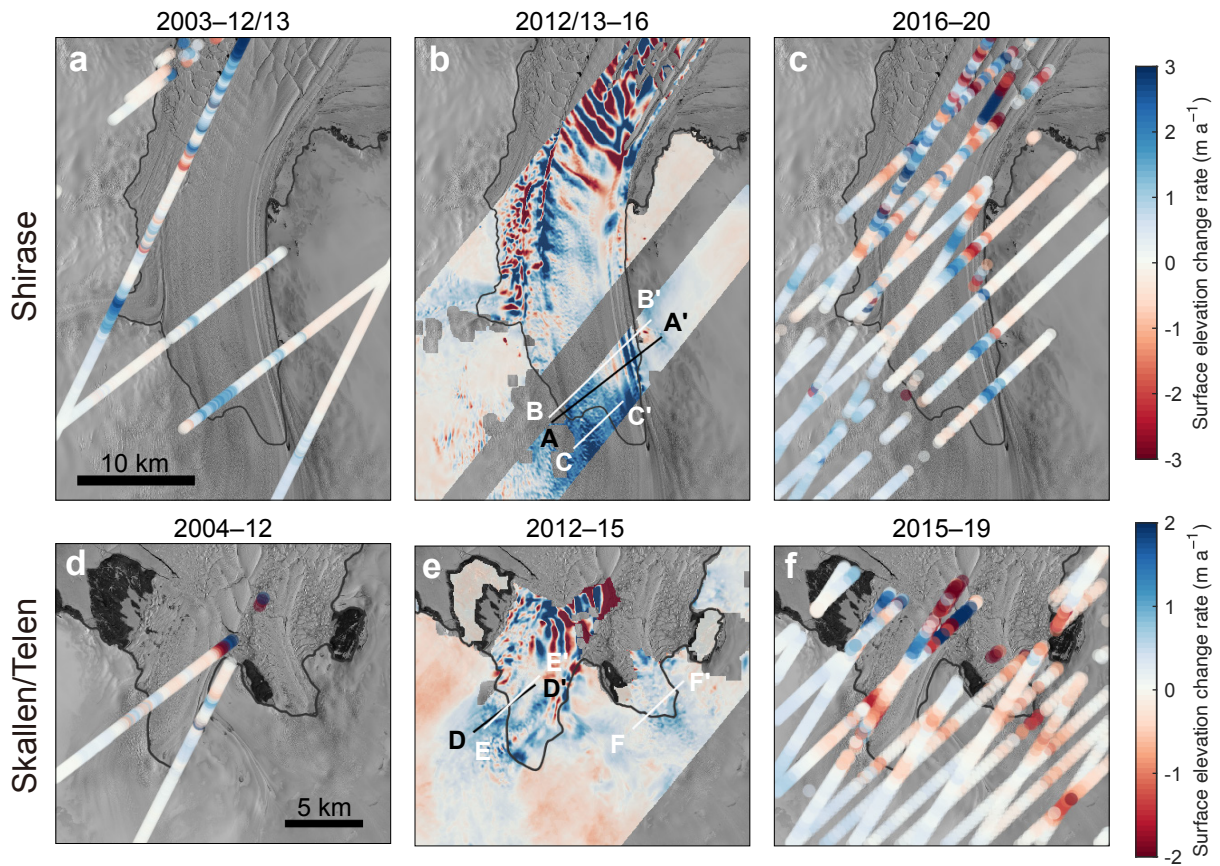


Figure 3.7: Surface elevation change rates at (a–c) Shirase, (d–f) Skallen/Telen glaciers during the periods (a and d) 2003/04–12/13, (b and e) 2012/13–15/16, and (c and f) 2015/16–19/20. The rates were obtained using elevation data acquired from (a and d) ICESat-1 and REMA, (b and e) REMA, and (c and f) REMA and ICESat-2. Elevation changes along the ICESat-1 (AA' and DD') and ICESat-2 tracks (BB', CC', EE', FF') are described in the text and shown in Fig. 3.8. The background is a Landsat 8 OLI image acquired on 3 January 2016.

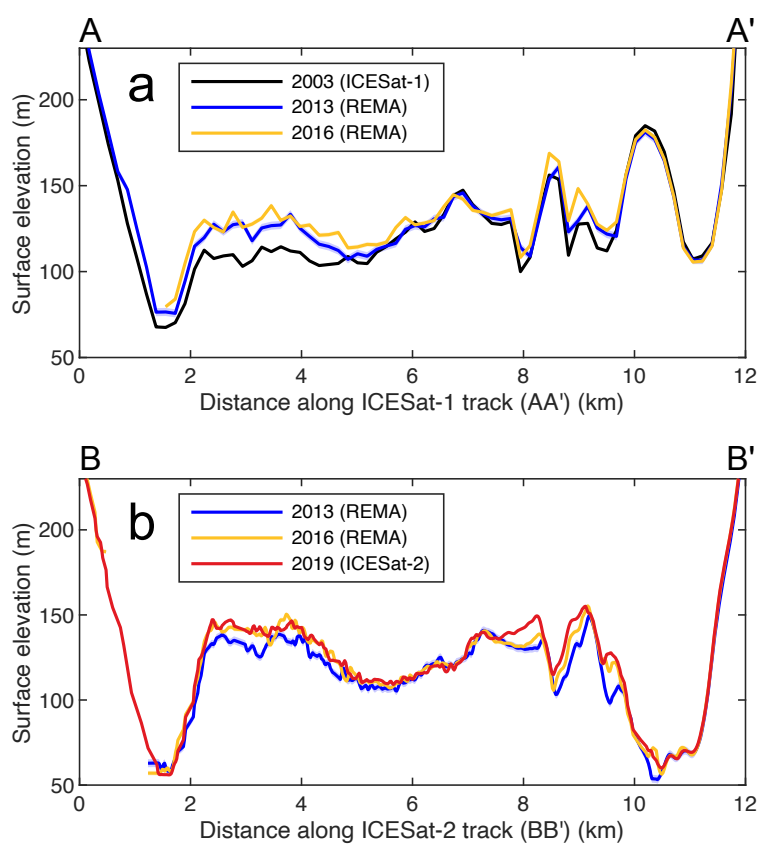


Figure 3.8: Surface elevation measured near the grounding line of Shirase Glacier along (a) AA' and (b) BB' shown in Fig. 3.7b.

3.3 Discussion

3.3.1 Driver of terminus position changes

The obtained data revealed synchronous frontal variations of the glaciers flowing into Lützow-Holm Bay. In general, the glaciers advanced from 1989 to 1997, followed by a period of retreat observed from 1997 to 2000 (Fig. 3.3). After a period of relatively stable terminus position from 2000 to 2008, the glaciers progressively advanced until a large retreat was initiated in the summers of 2015/16 (Shirase and Langhovde), 2016/17 (Skallen and Honnør), and 2017/18 (Telen) (Fig. 3.3). The trend observed in this study concurs with previously reported decadal glacier variability in the studied region, i.e. a retreating trend during the period 1974–90 followed by an advance from 1990 to 2012 (Miles and others, 2016).

The data suggests that the glacier front variations were influenced by sea-ice conditions in Lützow-Holm Bay (Fig. 3.3). To investigate a possible link between sea ice and the stability of the glacier front, the frontal displacement rate was compared with the SIC anomaly near each glacier front, by taking the average over every four-year period from 1992 to 2020. Taking the variability in the flow rates among the glaciers into account, the frontal displacement rates were normalized by the mean ice speed near each glacier terminus. The results illustrate that the glaciers synchronously retreated when the SIC anomaly was substantially negative, whereas they generally advanced during the positive phase (Fig. 3.9). The displacement rates were dependent on the SIC anomaly with substantially different sensitivities in the positive and negative SIC anomaly ranges ($0.30\%^{-1}$ for negative and $0.01\%^{-1}$ for positive). Excluding the advancing period of 2012–16, there was a significant correlation ($r = 0.77, p < 10^{-4}$) between the SIC anomaly and the normalized frontal displacement rate. During the period 2012–16 (SIC anomaly $> 3.5\%$), all the glaciers advanced without significant ice loss by calving (Fig. 3.3). The normalized displacement rates during this period were within the range of 0.49–1.00 (Fig. 3.9), meaning that the glaciers advanced at rates between 49 and 100% of the glacier speed. The most rapid glacier retreat was observed in the period 2016–20, when the SIC anomalies were below -2% after the breakup of land-fast sea ice in 2016 (Fig. 3.9). Open water was consistently observed near the glacier fronts during the summer months of this period (Figs 3.3a–e). These observations suggest that stable land-fast sea ice suppressed glacier calving and facilitated the terminus advance in 2008–16. The breakup in 2016 destabilized the glacier termini, resulting in the loss of frontal ice. Based on the simultaneous retreat of the glaciers coincided with the sea-ice loss in 1998, 2004, and 2016, we assume that multidecadal glacier front variations on Lützow-Holm Bay are governed by the periodic breakup of land-fast sea ice occurring at intervals of 10–20 years (Ushio, 2006; Aoki, 2017).

The association of glacier front position with sea-ice conditions is in line with previous studies in East Antarctica. Data suggests that tabular iceberg calving is suppressed by the formation of multi-year land-fast sea ice and the development of dense ice mélange in front of

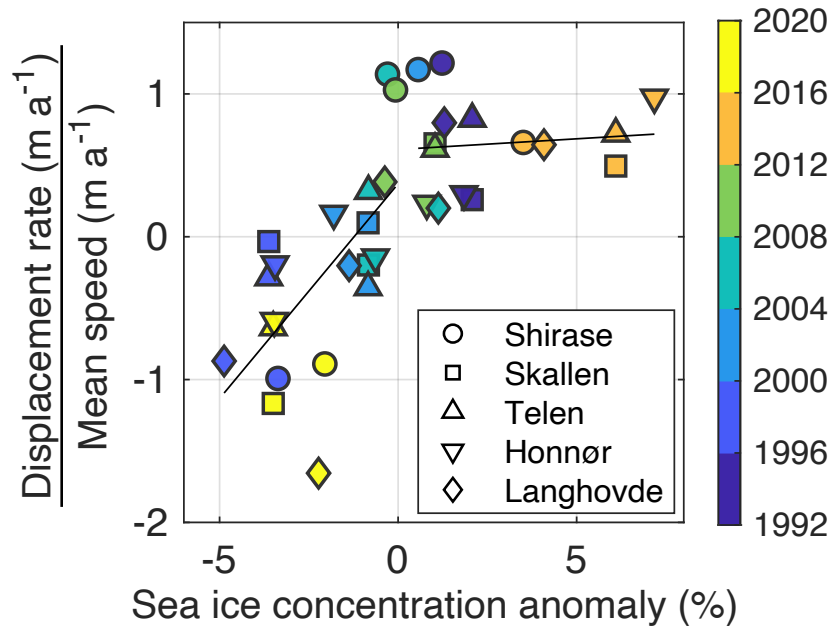


Figure 3.9: Scatter plot of glacier front displacement rates vs the SIC anomaly near each glacier front. The values for each glacier (symbols) are averaged over every 4-year period (colors). The solid lines show the linear regression of the data in positive and negative SIC anomaly ranges. The displacement rates are normalized by ice speed near the terminus.

the glacier (Massom and others, 2010; Miles and others, 2017; Arthur and others, 2021). A correlation between sea-ice conditions and decadal glacier front variations has been also reported in East Antarctica (Miles and others, 2013, 2016), on the Antarctic Peninsula (Christie and others, 2022), and in the whole of Antarctica (Baumhoer and others, 2021). Our data confirmed the influence of sea ice on glacier fronts based on the data acquired in a finer temporal resolution.

Observations on the glacier fronts during the retreat suggest that structural weakening of the ice shelves has preconditioned the calving before it was triggered by the sea ice breakup. The importance of rifts, crevasses, and troughs on iceberg calving has been emphasized in recent studies since they result in ice break-off (Miles and others, 2017; Arthur and others, 2021). For example, the ice shelves of Skallen and Telen glaciers in 2016 were extensively damaged by collision with each other as the termini advanced (Fig. 3.10a). The subsequent retreat of Skallen Glacier occurred due to the detachment of the fractured ice when the sea ice broke up (Figs 3.10a and b). The ice shelf of Telen Glacier was still surrounded by ice mélange in 2017 (Fig. 3.10b), and most likely, this is the reason why the retreat of Telen Glacier has lagged behind Skallen for two years (Figs 3.3b and c). At Langhovde Glacier, large surface troughs have been developed near the western margin of the glacier front before 2016 (Fig. 3.10d). The glacier retreated to the location of the trough during the largest retreat from 2016–2020 (Figs 3.10e), suggesting their importance on iceberg calving. Retreat at Honor Glacier in 2018 occurred when a large rift was formed on the ice shelf after the removal of sea ice in 2017 (Figs 3.10h and i). The rifts were initiated from the side margins of the ice shelf,

and icebergs detached due to the propagation of a rift across the ice shelf. At Shirase Glacier, the terminus of the ice tongue disintegrated associated with the sea-ice breakup as described by Aoki et al. (2017) (Figs 3.10k and l), suggesting they have been fastened by the land-fast sea ice. These observations indicate that the development of fractures, rifts, and troughs likely determines the location of iceberg calving when it was triggered by the removal of sea ice.

3.3.2 Mechanisms of glacier flow changes

Similar changes in ice speeds were observed at Shirase, Skallen, and Telen glaciers, i.e. generally decelerating trends from 2011 to 2015, which were followed by acceleration (Figs 3.4a–c and 5). Because the onset of the speedup in 2015–16 began slightly before the significant glacier retreat occurred during 2016–20 (Fig. 3.4), the speedup was unlikely to have been caused by the retreat. For example, Telen Glacier retreated by 2.2 km over the period February–September 2018, lagging two years behind the onset of the speedup.

Since the glacier speed changes were greater on the ice shelves than on the grounded ice (Fig. 3.5), I assume that a change near the front affected the force balance of the glacier. Given that the onset of the flow acceleration coincided with the breakup of land-fast sea ice in 2016, a potential driver of the glacier speedup could be a loss of back stress exerted by sea ice. As shown in Figs 3.6b and d, sea-ice displacement prior to the breakup in 2016 was substantially smaller than the forward motion of the glacier terminus. During this period, cracks and fractures were observed on the land-fast sea ice in front of Shirase Glacier (satellite image on 16 November 2015, Fig. 3.6a). These observations imply that a significantly large stress was exerted on the sea ice by the advancing glacier and that this stress impeded the forward glacier motion. The following summer (2016), sea ice sped up to approximately 90% of the glacier speed (Fig. 3.6c), implying that the sea ice lost support from the land and began to move with the glacier front. After the summer of 2015/16, an acceleration of the grounded ice and an even greater speedup of the ice shelves was observed at Shirase, Skallen, and Telen glaciers. These observations suggest that the multi-year land-fast sea ice was firmly anchored to the coast and resisted the glacier flow before 2016, whereas the increased mobility of the sea ice resulted in a reduction in resistive stress and subsequent glacier acceleration. Thus, I assume that the land-fast sea ice near the glacier front affected the glacier dynamics. Furthermore, the observed fractures and cracks on the land-fast sea ice can affect the mechanical rigidity of the sea ice, which might influence the eventual breakup of the sea ice. Such an interaction between the variations of sea ice and glaciers should be expected.

The process of the speed changes of the land-fast sea ice can be also found in the flow velocity measurements with high temporal resolution. The velocity of the sea ice near the terminus increased from 2050 to 2450 m a⁻¹ within five months during the summer of 2015/16, reaching the same level as the velocity of the floating ice tongue (Fig. 3.11). The increase in the flow velocity of the land-fast sea ice suggests that the sea ice lost support from the

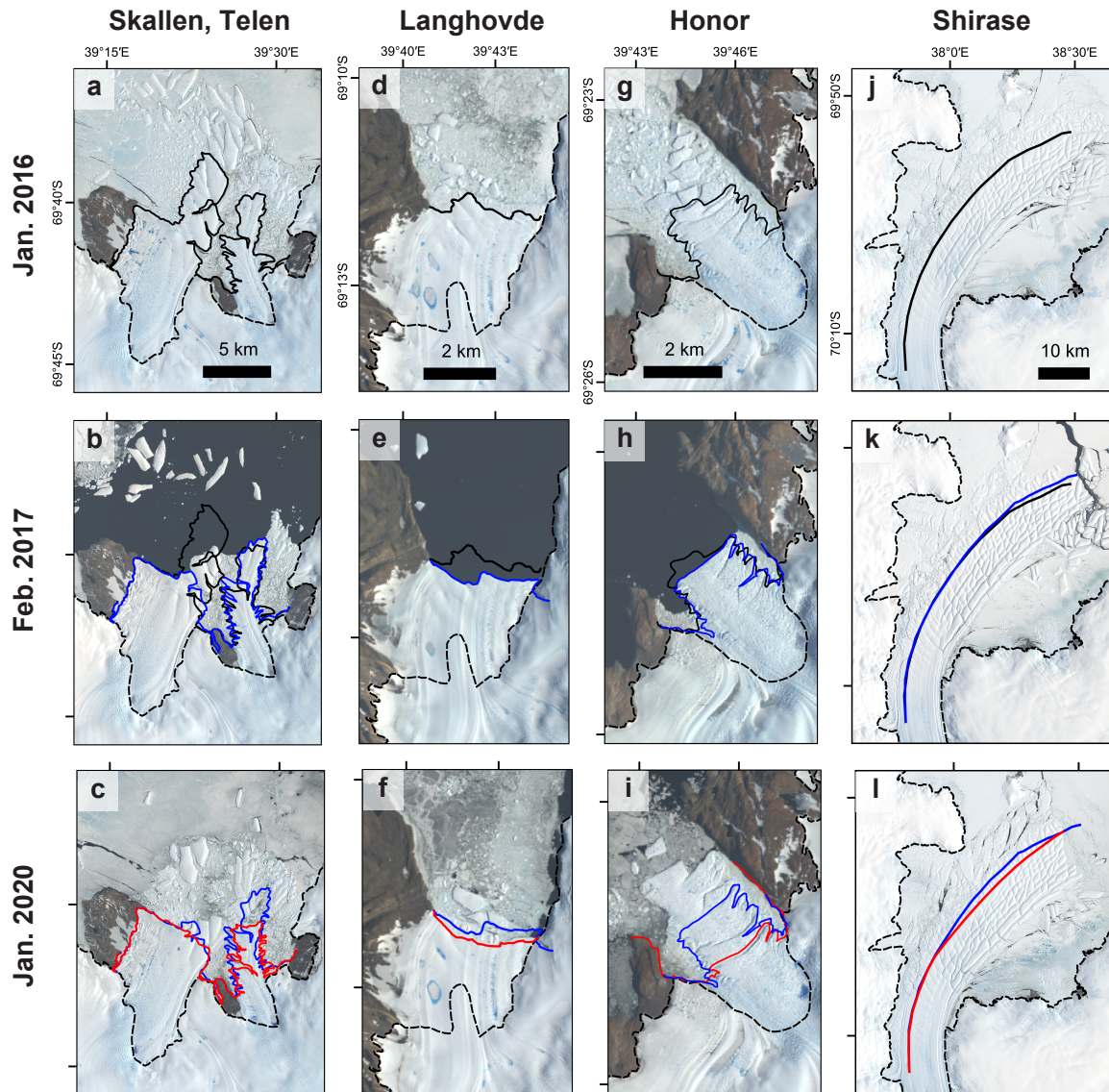


Figure 3.10: Landsat 8 OLI imagery in 2016–2020 at (a–c) Skallen and Telen, (d–e) Langhovde, (g–i) Honor, and (j–l) Shirase Glaciers. Terminus positions in January 2016 (black), February 2017 (blue), and January 2020 (red) are shown.

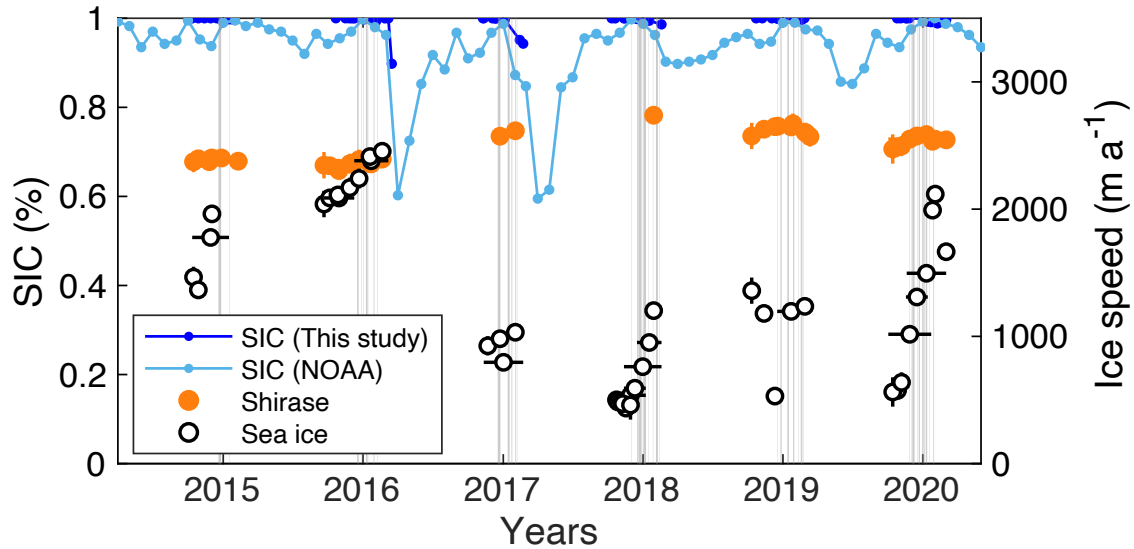


Figure 3.11: Speeds at the floating ice tongue of Shirase Glacier (orange) and on the sea ice 5 km away from the terminus (white). Sea ice concentration near the glacier terminus obtained in this study (blue) and from NOAA NSIDC Climate Data Records (light blue) are shown. Dates with daily mean air temperature above 0°C are indicated by shaded gray.

land. In the following summers in 2017/18 and 2019/20, substantial summer speedup was observed on the sea ice near the terminus. The sea ice showed 2.7- and 3.7-fold increase in the speeds during November 2017–January 2018 and October 2019–February 2020, respectively (Fig. 3.11). The onset of the summer speedup corresponded with the periods when the air temperature rose above 0°C , thus meltwater percolation to sea ice is a possible factor affecting the sea ice rigidity, which softens and weakens the ice. The above results suggest that the sea ice was highly mobile and not enough rigid to generate stress against the glacier flow during the period. This is consistent with the acceleration of the glaciers lasting after 2016 (Figs 3.4), despite the recovery in SIC above 80% at Shirase and 50% at Skallen/Telen after the summer 2017/18.

It is likely that multi-year land-fast sea ice and ice mélange in Lützow-Holm Bay had thickened since 2008 and had become rigid enough to generate stress which added resistance to the glacier flow. During the period of deceleration from 2010 to 2015, the condition of the sea ice in front of the studied glaciers was probably similar to those reported at Brunt/Stancomb-Wills Ice Shelf and the Larsen B embayment (Khazendar and others, 2009; Rott and others, 2018), where the flow speeds were thought to have been reduced by sea ice/ice mélange accumulated near the glacier fronts.

A possible impact of the ocean on the glacier flow is the change in ice thickness due to the basal melting of the ice shelves. Continuous ice radar survey and concurrent ship-based hydrographic observations found active basal melt (7.2 m a^{-1}) of the ice shelf of Shirase Glacier associated with mCDW intrusion (Hirano and others, 2020). Since the bathymetric survey confirmed the ocean trough connected from the continental shelf break to the calving

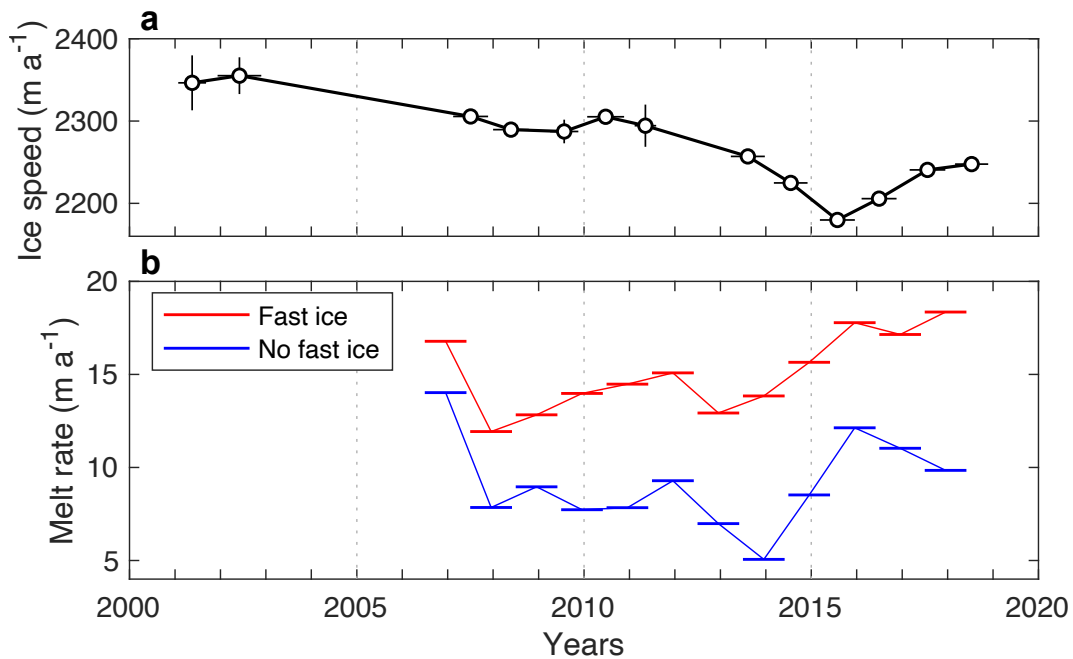


Figure 3.12: (a) Mean annual speeds at the grounding line of Shirase Glacier. (b) Mean annual basal melt rate on the ice shelf of Shirase Glacier simulated by ocean-sea ice-ice shelf model with (red) and without (blue) sea-ice cover on the ocean surface (Kusahara and others, 2021).

front of Shirase Glacier (Hirano and others, 2020), variability of warm water inflow possibly affects the ice thickness near the grounding line. However, a simulation of ocean-sea ice-ice shelf model suggested that the basal melt rate at Shirase Glacier increased by 50% from 2007 to 2016 (Fig. 3.12). The increase in the melt rate coincides with a period when the glacier decelerated, thus it is unlikely that the variability of basal melting was a factor causing the glacier flow changes. The simulation also showed that the melt rate decreased by 60% when the land-fast sea ice in the bay disappeared because the removal of sea ice thickens the surface cold water layer and inhibits the inflow of mCDW. The results are also contradictory to the glacier flow acceleration that began in 2016, when the land-fast sea ice broke up.

The observed variations of Shirase Glacier resulted in a decrease in the ice discharge from 12.7 to 12.5 Gt a^{-1} during the period of deceleration from 2013 to 2015. The change expected from the deceleration was -0.4 Gt a^{-1} , but its 33% was canceled by a thickening of the grounded ice during the same period. The analysis highlights the significant influence of both ice speed and ice thickness on glacier discharge, as well as the impact of sea ice variations on ice sheet mass balance.

3.3.3 Elevation change of grounded ice

Surface elevation change indicated thickening of Shirase, Skallen, and Telen glaciers in 2003/04–15/16 (Figs 3.7 and 3.8) during the period of glacier deceleration (Fig. 3.4). The greatest elevation change occurred near the grounding line of Shirase Glacier ($2.2 \pm 0.1 \text{ m a}^{-1}$) (Figs 3.7 and 3.8) and was far greater than the surface mass balance reported along the Soya

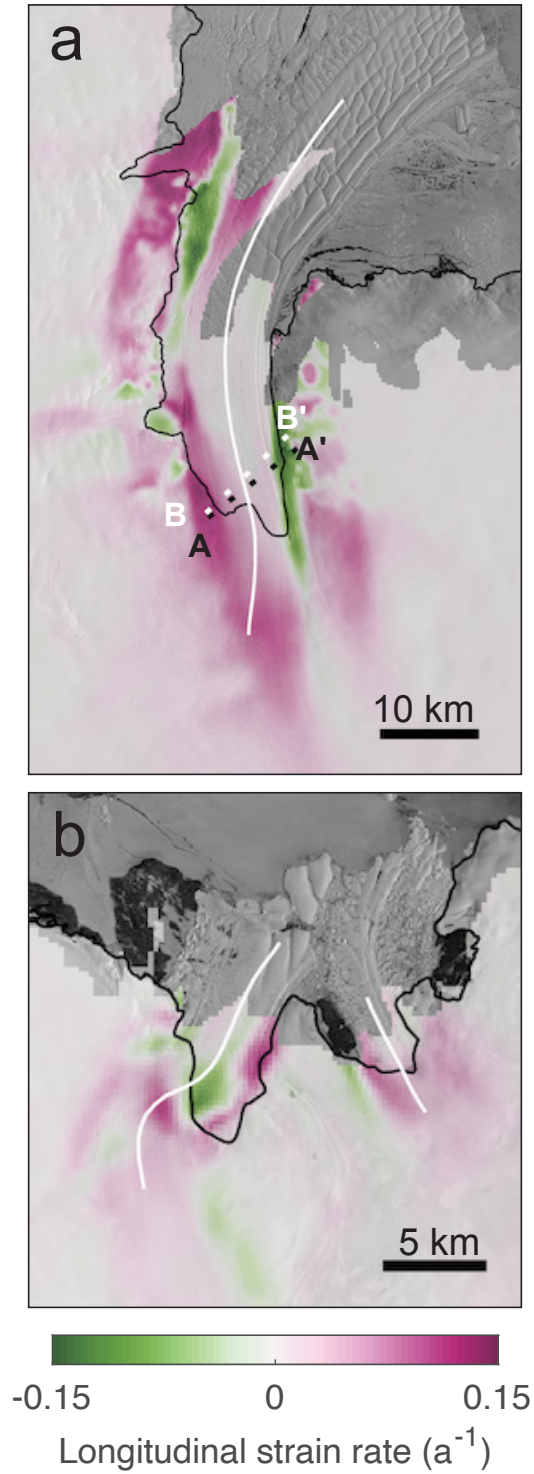


Figure 3.13: Longitudinal strain rate distributions calculated from the glacier flow velocity in 2018 from the ITS_LIVE dataset at (a) Shirase and (b) Skallen/Telen glaciers. The white lines show the central flowlines used for Fig. 3.14. The background is a Landsat 8 OLI image acquired on 8 January 2018.

Coast (<0.4 m w.e. a^{-1}) (Motoyama and others, 2008, 2015; Wang and others, 2015). Shirase, Skallen, and Telen glaciers began thinning in 2016, in response to the acceleration of the glaciers, suggesting a link between flow regime and ice thickness (e.g. Pritchard and others, 2009). Since the observed ice speed changes were greater on the ice shelves as compared to the grounded ice (Fig. 3.5), the stretching ice flow along the glacier should have been reduced/enhanced by deceleration/acceleration. According to the strain rate analysis, the longitudinal (parallel to the flow direction) component of the strain rate tensor from the velocity fields in 2018 shows stretching flow regimes near the grounding line of Shirase, Skallen, and Telen glaciers (Fig. 3.13). A positive (stretching) strain rate exceeding 0.1 a^{-1} was observed near the grounding line of the three glaciers. At Shirase Glacier, the greatest strain rate of 0.14 a^{-1} occurred near the western margin (Fig. 3.13a, near A and B), where the largest surface elevation change was observed in the period 2003–16 (Figs 3.7 and 3.8).

Further analysis demonstrated the impact of the ice speed change on the strain rate regime. The magnitude of the deceleration from 2007 to 2015 was greater on the ice shelf than on the grounded area (Figs 3.14a–c), which resulted in a reduction in the stretching flow and an increase in the vertical strain rate near the grounding line and the upper reaches (Figs 3.14d–f). The increase in the vertical strain rate implies a thickening, consistent with the surface uplift observed near the grounding line at Shirase, Skallen, and Telen glaciers (Figs 3.14g–i). In the following period (2015–18), the vertical strain rate decreased as a result of the speedup, which was more pronounced over the ice shelf (Figs 3.14a–f). The enhanced stretching flow regime is consistent with the decrease in surface elevation at Shirase, Skallen, and Telen glaciers (Figs 3.14g–i). These results strongly suggest that the increase and decrease in elevation during the periods of deceleration and acceleration was driven by the changes in the vertical strain rate.

The data also indicated that the observed change in ice thickness affected the grounding line positions. From 2012/13 to 2016, during the period of deceleration and thickening, the grounding line of Shirase Glacier estimated from surface slope showed seaward advance by 65 m (20 m a^{-1}) (Figs 3.15a and b). This observation is consistent with a previous measurement with CryoSat-2 altimetry data, i.e. an advance at a rate of 14.7 m a^{-1} between 2010 and 2016 (Konrad and others, 2018). Skallen and Telen glaciers also showed seaward grounding line migration of 7 and 18 m between 2012 and 2015 (Figs 3.15c and d). The changes in grounding line position affect the ice motion near the grounding line because of changes in basal stress (e.g. Payne and others, 2004), thus a complex feedback mechanism is expected between the changes in ice speed and ice thickness triggered by variations in land-fast sea ice.

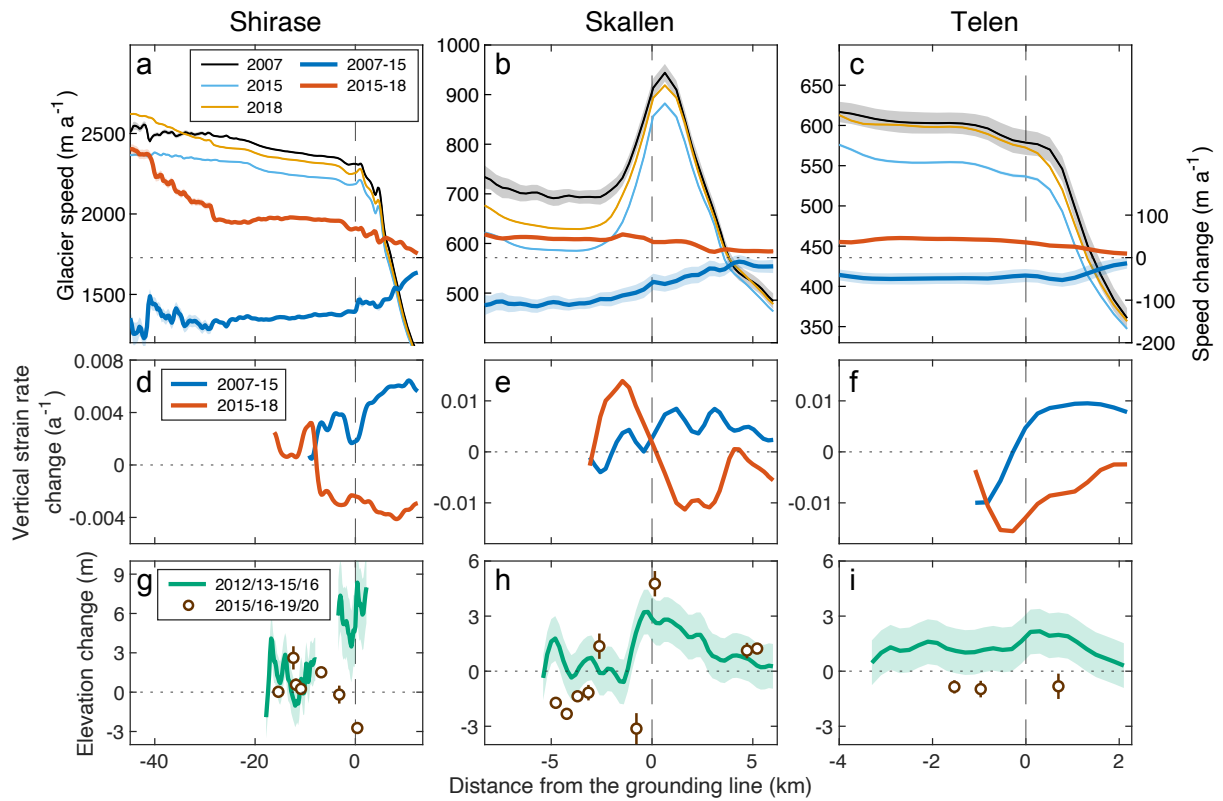


Figure 3.14: (a–c) Glacier speed in 2007 (black), 2015 (light blue), and 2018 (orange) along the central flowline shown in Fig. 3.13. Speed changes across the periods 2007 to 2015 (blue) and 2015 to 2018 (red) are also shown. (d–f) Changes in the vertical strain rate computed from the glacier velocity field. (g–i) Elevation changes in 2012/13–15/16 (green) and 2015/16–19/20 (circles). Vertical dashed lines indicate the location of the grounding line.

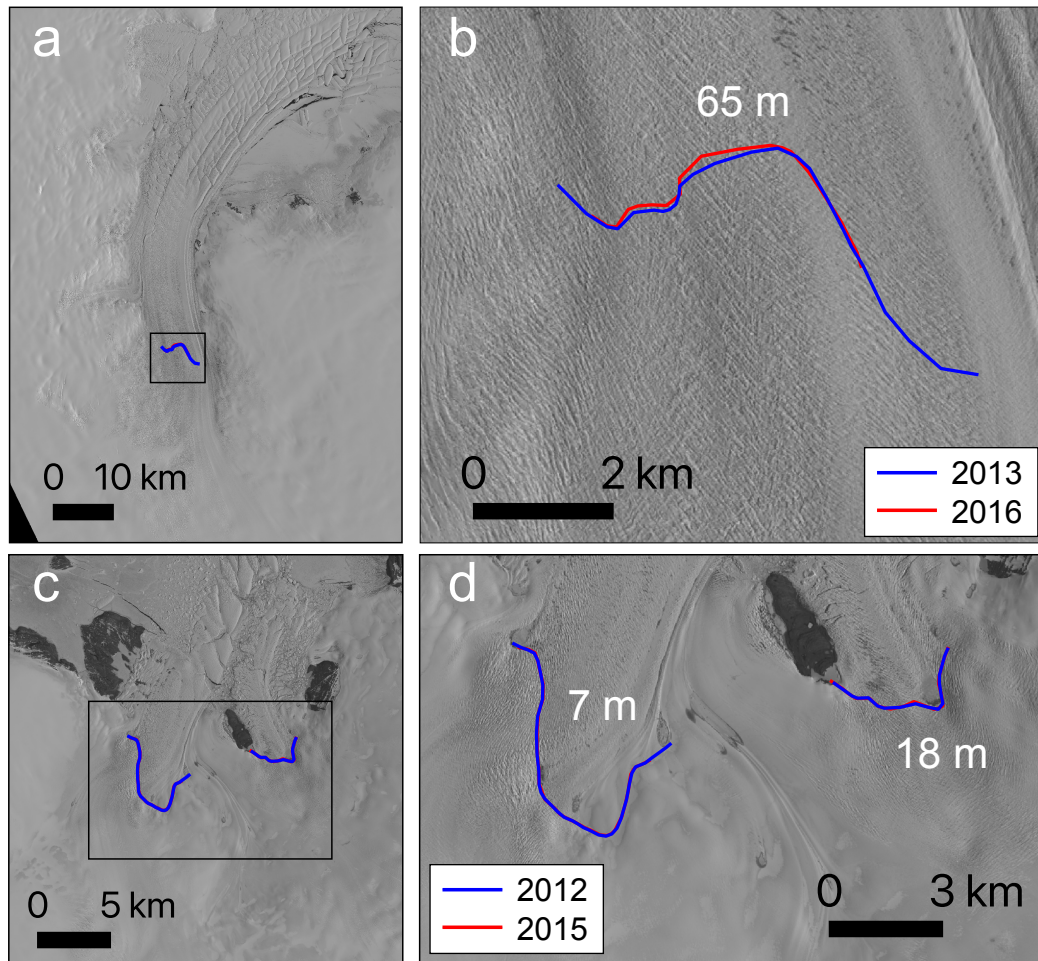


Figure 3.15: Groundline positions estimated from break in slope in the REMA Strip DEMs in 2012/13 (blue) and 2015/16 (red) at (a and b) Shirase and (c and d) Skallen/Telen glaciers. (b) and (c) show close-up views in the polygons in (a) and (c), respectively. The numbers indicate changes in the groundline positions between 2012/13 and 2015/16.

3.4 Summary

To investigate recent glacier variations in Lützow-Holm Bay, East Antarctica, I measured the front position, flow velocity, and surface elevation of five outlet glaciers from 1988 to 2020. The glacier variations were compared with sea ice conditions near the glacier fronts to investigate a possible link between glacier and sea ice variations.

The fronts of the studied glaciers advanced from 2006 to 2015 when the ocean was largely covered with sea ice. All the glaciers rapidly retreated by 0.4–6.0 km in the period 2016–18, which coincided with a major breakup of land-fast sea ice in Lützow-Holm Bay in 2016. The breakup was the most significant sea ice loss in the region since 1998. Shirase, Skallen, and Telen glaciers simultaneously decelerated in the period 2011–15 and accelerated in 2016–18. These flow speed changes were most pronounced on the ice shelves but were also observed upstream of the grounding line. Near the grounding lines of Shirase, Skallen and Telen glaciers, the surface elevations increased during 2003/04–15/16 and decreased during 2015/16–19/20, corresponding with the flow deceleration and acceleration periods, respectively.

I found that the retreat rate of the glaciers correlated with the SIC anomaly near the glacier fronts when the SIC anomaly dropped below 0, suggesting that the sea ice is an important factor controlling the terminus variations. The displacement rate of the sea ice before the breakup in 2016 was substantially smaller than the forward motion of the glacier terminus. During this period, cracks and fractures were observed on the land-fast sea ice in front of the glaciers, implying that a significantly large stress was exerted on the sea ice by the advancing glacier and this stress impeded the forward glacier motion. The glacier flow accelerated after the sea ice sped up and began to move with the glacier front, suggesting a reduction in the resistive stress resulted in the glacier acceleration after 2016. Glacier thickness was also affected by the sea-ice breakup and the associated ice speed changes. Strain rate analysis suggested that the stretching ice flow along the glacier was reduced/enhanced by the observed flow deceleration/acceleration. The results suggest that the observed elevation changes were driven by the variations in the vertical strain rate near the grounding line, caused by changes in the glacier flow regime.

The results demonstrate that the presence of land-fast sea ice affects glacier speed and ice thickness, as well as glacier front variations. Consequently, sea ice variations in front of glaciers may trigger a positive feedback mechanism between glacier flow and ice thickness. Given that sea ice is highly sensitive to climate (Heil, 2006; Fraser and others, 2012; Aoki, 2017), an accurate understanding of land-fast sea ice is crucial for the projection of glacier changes in Antarctica under a changing climate.

Field measurements: Influence of subglacial conditions on the glacier flow changes

In this chapter, field observations were conducted with the aim of observing short-term flow changes of Langhovde Glacier and investigating its mechanisms by subglacial measurements. In order to elucidate the relationship between glacier basal motion and subglacial water pressure variations, a device equipped with a three-axis accelerometer and a water pressure sensor was developed. The device was installed at the glacier bottom using a borehole drilled by hot water, and the acquired data was analyzed together with the velocity time series observed on the glacier surface.

4.1 Overview of the field campaign

The field observations were carried out between 16 December 2021 and 6 February 2022 as a part of the JARE 65th summer activity. A wide variety of observations was carried out on the surface and the bottom of the glacier. Hot-water drilling was performed at three locations, of which I focus on the two sites named LGX and LG5 in this thesis. The drilling sites were located at 1.5 and 0.5 km upstream from the grounding line estimated by the surface slope (Fig. 4.1) (Fukuda, 2014). Using three boreholes reaching the bed, subglacial water pressure, vibrations and tremors at the glacier base, and englacial ice temperature were measured. On the glacier surface, ice flow, icequakes, snow/ice melt, and weather were measured, as well as ice radar surveys and supraglacial water sampling.

During the field campaign, three boreholes (BH01, BH02, and BH03) were drilled to the glacier bed by hot-water drilling method. Among the boreholes, BH01 and BH02 were drilled at LGX, whereas BH03 was drilled at LG5 (Fig. 4.1). The first drilling operation attempting

to the glacier bed was carried out on 25 December 2021 for the borehole BH01 at LGX. The drilling succeeded after 15 hours of operations over four days during 25–31 December. At the bottom of the borehole BH01, a ploughmeter equipped with a three-axis accelerometer and a water pressure sensor was installed. After the successful drilling of BH01, the second borehole (BH02) was drilled at LGX 10 m apart from BH01 on 6 January 2022. The drilling was completed within a day in 15 hours without trouble, and a water pressure sensor and a thermistor string were installed in the borehole. The thermistor strings were equipped with seven thermistors, which were placed at depths of 100, 200, 300, 400, 500, 525, and 550 m. The final drilling took place on 22–23 January at LG5, 1 km downstream from LGX. LG5 was located 0.6 km upstream from the grounding line estimated from the break in the surface slope (Fukuda, 2014). We installed a water pressure sensor here to observe variations in water pressure. As a result, two boreholes (BH01 and BH02) were drilled at LGX and a borehole (BH03) was drilled at LG5. In this thesis, I focus on the data retrieved from BH01 and BH02 to investigate the subglacial conditions and their dynamics well upstream from the grounding line. The data from the thermistor string were not present here because the temperature had not been converged yet at the end of the field campaign due to the influence of the drilling.

On the glacier surface, GNSS receivers were installed at five sites, LG1, LGLK, LG4, LG5, and LGX, located 0.3–4.2 km from the glacier front (Fig. 4.1). A GNSS receiver was placed as a base station on exposed bedrock at the western flank of the glacier. Among the installed GNSS stations, the data from LG4 and LGX were analyzed and shown in this thesis to focus on the dynamics of the grounded ice. At LG4, an automatic weather station (AWS) was operated to observe meteorological conditions. For icequake observations, six seismometers were installed on the glacier surface and two borehole seismometers were placed in boreholes at LGX at depths of 4 and 200 m, which were not shown in this thesis.

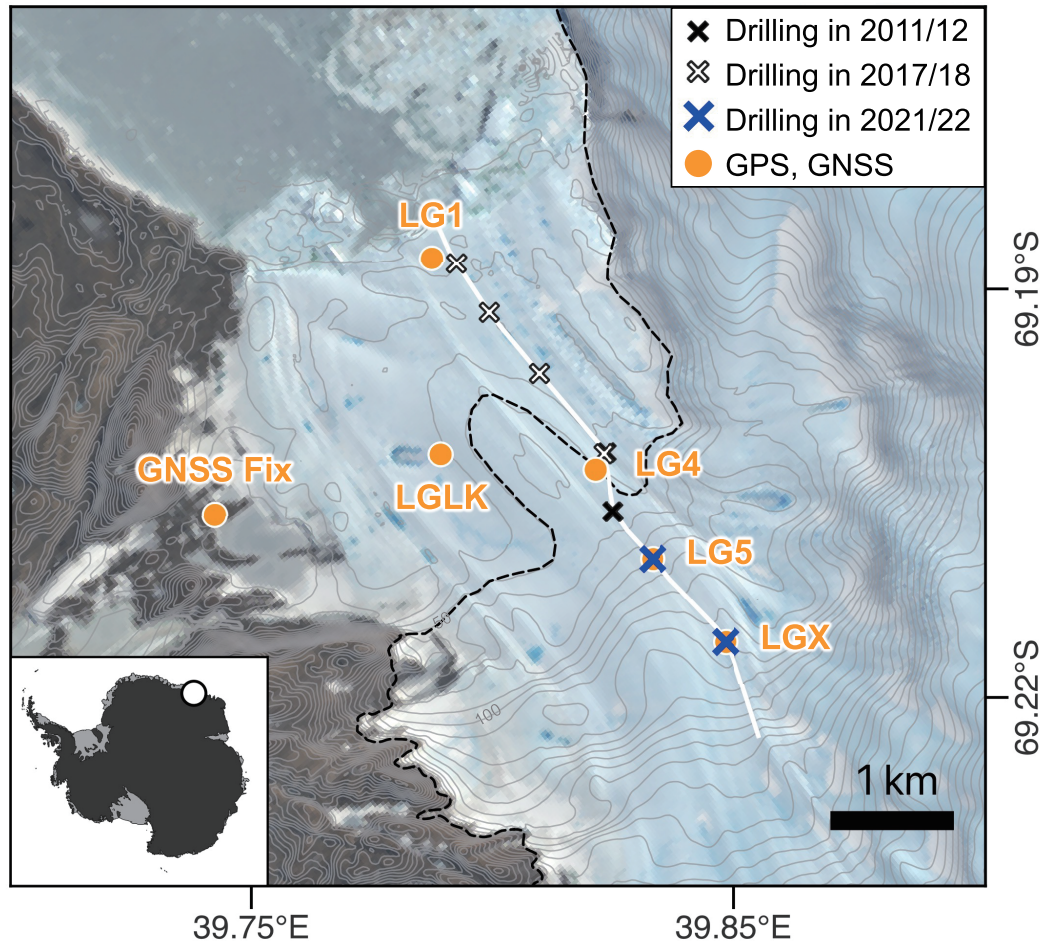


Figure 4.1: Landsat 8 OLI imagery (12 December 2021) showing hot-water drilling sites during 2011–2022 (crosses) and GNSS locations (circles). The dashed line shows the grounding line estimated from the break in the surface slope (Fukuda, 2014). The white line shows the profile shown in Fig. 4.5. The inset shows the location of the study site in Antarctica.

4.2 Methods

4.2.1 Hot-water drilling

In December 2021 and January 2022, three boreholes were drilled at sites LGX and LG5 on Langhovde Glacier (Fig. 4.1). Drilling was performed twice at LGX (BH01 and BH02) and once at LG5 (BH03). The drilling was performed at a mean rate of 40 m h^{-1} with a hot-water drilling system, which had been previously used on Langhovde Glacier (Sugiyama and others, 2014). The system consists of three high-pressure, hot-water machines (Kärcher HDS1000BE), a 1/2-inch diameter hose, and a winch system (Tsutaki and Sugiyama, 2009; Sugiyama and others, 2010b). Meltwater was supplied to the system from a supraglacial stream or pond. Ice thicknesses and ocean bed elevation were determined with an accuracy of $\pm 1 \text{ m}$, using a CTD (conductivity, temperature, and depth) profiler (IDRONAUT, Ocean Seven 304). At each borehole site, ice-surface elevation was measured using a dual-frequency GNSS (Leica Geosystems System 1200) with the static method using data from a reference station (Enabler, GEM-2) located on the western flank of the glacier (Fig. 4.1). To observe the wall and the bottom of the boreholes, a video camera (Panasonic, HX-A1H) enclosed in a pressure vessel was lowered into the boreholes with a water-pressure resistant light (Seatool, SCL-262) and the CTD profiler. The depths of the video images were determined by comparing the elapsed time with the depth obtained by the CTD profiler.

A variety of englacial and subglacial measurements were conducted using the boreholes. At LGX, a pressure sensor and a subglacial ploughmeter were installed at BH01 and BH02, respectively. At BH03, a pressure sensor was installed, but was not presented in this thesis.

4.2.2 Subglacial ploughmeter

Using BH01 drilled at LGX, I observed glacier basal motion directly by using a "subglacial ploughmeter" (Kondo and others, 2022). The device consisted of a water pressure sensor and an accelerometer enclosed in a stainless-steel rod. The accelerometer records three-dimensional acceleration with a sampling rate of up to 100 Hz. Subglacial acceleration measurements were performed by installing the ploughmeter at the bottom of the borehole BH01. In this thesis, the analysis of the acceleration data focused on the two observation targets, i.e., glacier basal sliding and subglacial tremor. For the observation of the basal sliding, the tip of the device was placed on the borehole bottom. As a result of the glacier sliding, the tip is dragged through the glacier bed, so that the accelerometer records signals associated with the basal motion. In addition to the signals associated with the movement of the ploughmeter, the high-frequency sampling enables to record subglacial tremors. In this section, I introduce the design of the device and the method to process the obtained acceleration data.

Design of the ploughmeter and sensors

The ploughmeter was equipped with an accelerometer and a water pressure sensor. The sensors and a control circuit were enclosed in a stainless-steel pipe of 34 mm in diameter and 657 mm in length (Fig. 4.2a). The accelerometer and the water pressure sensor were placed at 70 and 400 mm from the probe tip. To measure the surrounding pressure, a 2-mm-diameter hole is drilled in the tip of the probe and connected to the pressure sensor. Subglacial water pressure was measured by a piezo-resistive pressure sensor (Valeport, miniIPS) equipped in the ploughmeter with a sampling rate of 1 Hz. The accuracy and the resolution of the measurements were 1 kPa and 0.1 kPa, respectively. Four units of brass weights (34 mm diameter, 1830 mm length, and 6.7 kg in total) were attached to the top of the casing to insert the probe into the subglacial sediment by its own weight. The weights were also important to provide sufficient tension on a cable during the installation. A 600-m sensor cable was connected to the probe via a water-proof connector, and the other end was connected to the data logger (Clovertexh Inc.) on the glacier surface.

Three-axis component of acceleration was measured by a digital microelectromechanical system (MEMS) accelerometer with a quartz oscillator sensor (EPSON, M-A352) with a sampling rate of up to 100 Hz. The natural frequency of the quartz oscillator has a large temperature dependency. The measurements can be made with an accuracy less than $3 \times 10^{-5} \text{ m s}^{-2}$ ($3 \times 10^{-6} \text{ g}$) if appropriate temperature corrections were applied (Matsuda and others, 2018). The acceleration data was corrected by the control circuit equipped in the probe using the temperature measured by a built-in thermistor. The corrected acceleration was digitally transferred from the control circuit to the data logger on the glacier surface. Thus, the temperature dependence of the sensor has been removed by in-situ temperature. The sensors were powered by the surface data logger via a 600-m-long cable installed in the borehole. The data logger was powered by 24 V, 220 Ah lead-acid batteries and a 60 W solar panel.

Analysis of the acceleration data

The obtained acceleration data was analyzed for (1) the tilt of the ploughmeter, (2) impulsive vibration events, and (3) spectral characteristics. In the following paragraphs, details of each method are shown.

1. The tilt of the ploughmeter was computed from the observed three-axis acceleration. After removing high-frequency signals greater than 1 Hz by applying a low-pass filter, remaining acceleration was assumed to be due to gravity. Roll (α) and pitch (β) of the sensor were calculated from the acceleration (a) measured along the x, y, and z-axis as:

$$\alpha = \tan^{-1} \left(\frac{a_y}{\sqrt{a_x^2 + a_z^2}} \right), \quad (4.1)$$

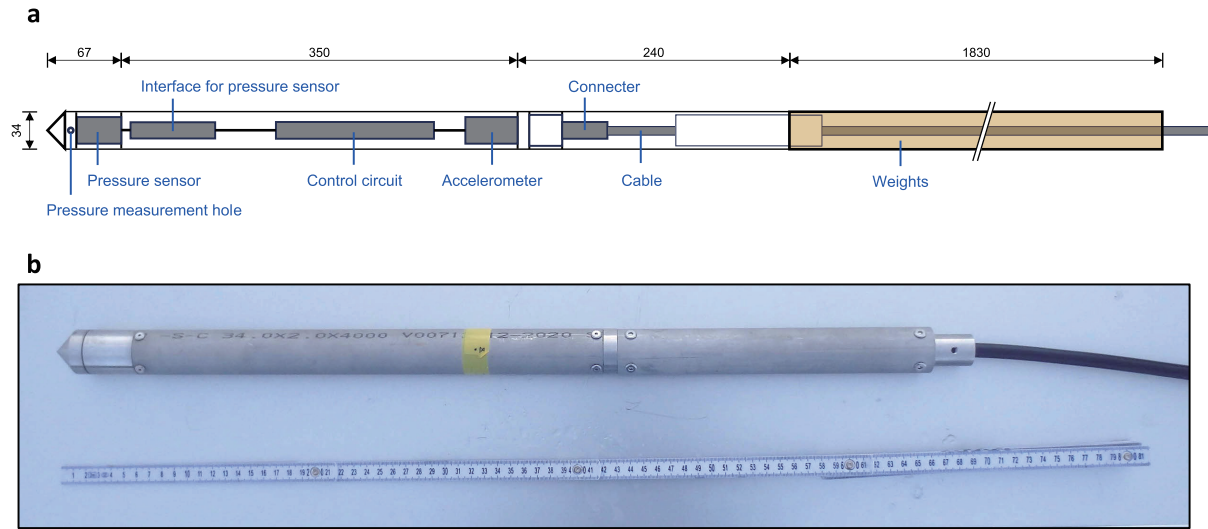


Figure 4.2: (a) Schematic diagram of the subglacial ploughmeter. The numbers indicate the length (mm) of each unit. (b) A picture showing the sensor-mounted probe of the ploughmeter. The length of the scale is 820 mm.

$$\beta = \tan^{-1} \left(\frac{a_x}{\sqrt{a_y^2 + a_z^2}} \right). \quad (4.2)$$

Tilt from the vertical (θ) was computed using roll (α) and pitch (β) as

$$\theta = \cos^{-1}(\cos \alpha \cos \beta). \quad (4.3)$$

Uncertainty in the tilt would be caused by the noise of the acceleration records. The precision of the tilt measurements was estimated using the tilt variations obtained from the ploughmeter left on the flat glacier surface for 14 hours on 24–25 December 2021 before the installation. After removing a trend and vibration events from the tilt data, the resulting time series was regarded as noise of the measurements. The precision of the tilt measurement was less than 10^{-4} degrees, thus considered to be negligible.

To determine the absolute accuracy of the tilt measurement, the ploughmeter was tilted at 30, 45, and 60° using an angle meter with a reading accuracy of $\pm 0.5^\circ$ (Niigata Seiki Co., Ltd., Dial Slantrule). The mean difference between the computed tilt and the reading of the angle meter was 0.7° on average, indicating that the accuracy was less than $\pm 0.9^\circ$ as a result of error propagation.

2. Vibration events were detected by the classic short-term average/long-term average (STA/LTA) algorithm implemented in Obspy, a Python toolbox for seismology (Beyreuther and others, 2010). I used 0.2 and 5.0 s window lengths for the STA and LTA, respectively, and detected an event when the STA/LTA ratio exceeded a factor of 6, following the method used in the previous studies (Barruol and others, 2013; Podolskiy and others, 2016; Minowa and others, 2019; Gimbert and others, 2021). Before the detection, a bandpass

filter was applied to the acceleration data between 1 and 20 or 50 Hz to remove the trend line as well as limit the detection up to the Nyquist frequency. For the detected vibration events, inter-event time and peak amplitude were computed to investigate the event characteristics.

3. To investigate spectral characteristics of the acceleration data, the continuous time series were processed in two ways. Firstly, I calculated the temporal variations of the power spectral densities (PSDs) for 1-minute segments of the acceleration data to obtain a long-term spectrogram. Secondly, temporal variations of seismic tremor were calculated. For this purpose, PSDs were computed for the acceleration data over a shorter time window of 2 s with 50% overlap. To maximize the sensitivity to the low amplitude seismic tremor and minimize the influences from energetic impulsive events, I chose the short time window of 2 s for the calculation. To quantify the temporal evolution of the seismic tremor, median power was calculated over the obtained PSDs with a moving time window of 2 hours. The PSDs were also used to generate probability density functions (PDFs) of the PSDs using the method introduced by McNamara (2004).

Installation and operation

The ploughmeter was successfully installed at the bottom of BH01 on 31 December. The tilt of the ploughmeter immediately after the installation was $1.9 \pm 0.9^\circ$, indicating the device was installed almost vertically. Given the initial tilt angle, the length of the ploughmeter of 2487 mm, and the visual inspection made by the borehole camera (Section 4.3.2), the initial borehole diameter was estimated to be ~80–150 mm. The accelerometer and the water pressure sensor were operated from 31 December 2021 to 6 February 2022. The accelerometer was operated with a sampling rate of 100 Hz between 31 December 2021 and 26 January 2022. Thereafter, the sampling rate was changed to 20 Hz to save memory space for year-round observations. During the field campaign, the data was last downloaded on 6 February 2022. Although the memory card was retrieved a year after on 20 January 2023, the data inside had been lost. Thus, I processed the acceleration data from 31 December 2021 to 6 February 2022 with a focus on the subglacial dynamics during the summer of 2021/22.

4.2.3 Subglacial water pressure

Subglacial water pressure was measured at the bottom of BH01 and BH02 drilled at LGX. At the bottom of BH01, water pressure was measured by the water pressure sensor equipped in the ploughmeter as described in Sec. 4.2.2. At the bottom of BH02, subglacial water pressure was measured using a vibrating-wire pressure sensor (Geokon 4500SH). The sensor was installed at 0.5 m above the bed on 7 January 2022 and operated until the device failure on 4 April 2022. The sensor measured water pressure with an accuracy and a resolution of 7.5 and 1.9 kPa with sampling intervals of 5 minutes during 7–26 January and 30 minutes during 26 January–4

April. A 570-m long sensor cable was connected to a data logger (Campbell Scientific, Inc., CR1000X) via a vibrating-wire interface module (AVW200) enclosed in a waterproof casing (Pelican 1600). The logger was powered with 12 V 24Ah lead-acid batteries and a 24 W solar panel. The ice overburden pressure (p_i) at the drilling site was calculated as $p_i = \rho_i g H$, where $\rho = 910 \text{ kg m}^{-3}$ is the density of ice, $g = 9.81 \text{ m s}^{-2}$ is gravitational acceleration and H is ice thickness. The study sites on Langhovde Glacier are located in an area with exposed ice, thus firn air content was not considered in the calculation of the ice overburden pressure. In this thesis, the pressure variations between 7 January and 6 February were analyzed.

4.2.4 GNSS survey

Dual-frequency GNSS (Enabler, GEM-1/2) receivers were continuously operated to measure glacier surface movement. The receivers were placed at five locations on the glacier spanning 0.3-4.2 km from the calving front (LG1, LGLK, LG4, LG5, and LGX). Among the locations, the data from LG4 and LGX are presented in this thesis (Fig. 4.1).

According to the previous drilling campaigns, LG4 is located on the ice shelf with ice thickness of 398 and 412 m measured in 2012 and 2018, respectively, with the ocean cavities of 13 and 24 m below (Sugiyama and others, 2014; Minowa and others, 2021). LGX was located 4.2 km upstream from the grounding line estimated from the surface slope, thus assumed to be grounded. The GNSS antennae were mounted on the top of 2-m-long aluminum poles drilled into the glacier surface. The receivers were kept running to collect GNSS satellite signals with a sampling rate of 1 Hz from 19 December 2021 to 6 February 2022 at LGx and from 23 December 2021 to 6 February 2022 at LG4. Each GNSS receiver was enclosed in a waterproof box with 12 V lead-acid batteries powered by a 24 W solar panel. At LG4, there is a data gap between 13:00 UTC on 23 and 7:00 UTC on 25 January due to a power shortage resulting from wind-induced overturning of the solar panel. At LGX, measurement failed due to a cable failure between 22:15 UTC on 23 and 12:20 UTC on 25 December 2021, 8:45 UTC on 12 and 14:50 UTC on 19 January 2022, and 4:30 UTC on 22 and 9:30 UTC on 26 January 2022.

The data was processed to obtain three-dimensional coordinates every 15 minutes using the GPS software RTKLIB in a static mode with a reference station placed on the exposed bedrock at the western flank of the glacier (Fig. 4.1). The horizontal and vertical accuracies are expected to be 2–3 mm and 3–5 mm, respectively, for the ~4 km baseline length (Sugiyama and others, 2010a, 2014; Minowa and others, 2019). The obtained coordinates were filtered using a Gaussian smoothing routine with ± 4 hour time window.

4.2.5 Meteorological measurements

We operated an automatic weather station (AWS) at LG4 to observe meteorological conditions on the glacier. The AWS measures meteorological variables (air temperature, wind speed and direction, relative humidity, atmospheric pressure) with an integrated weather sensors (Visala,

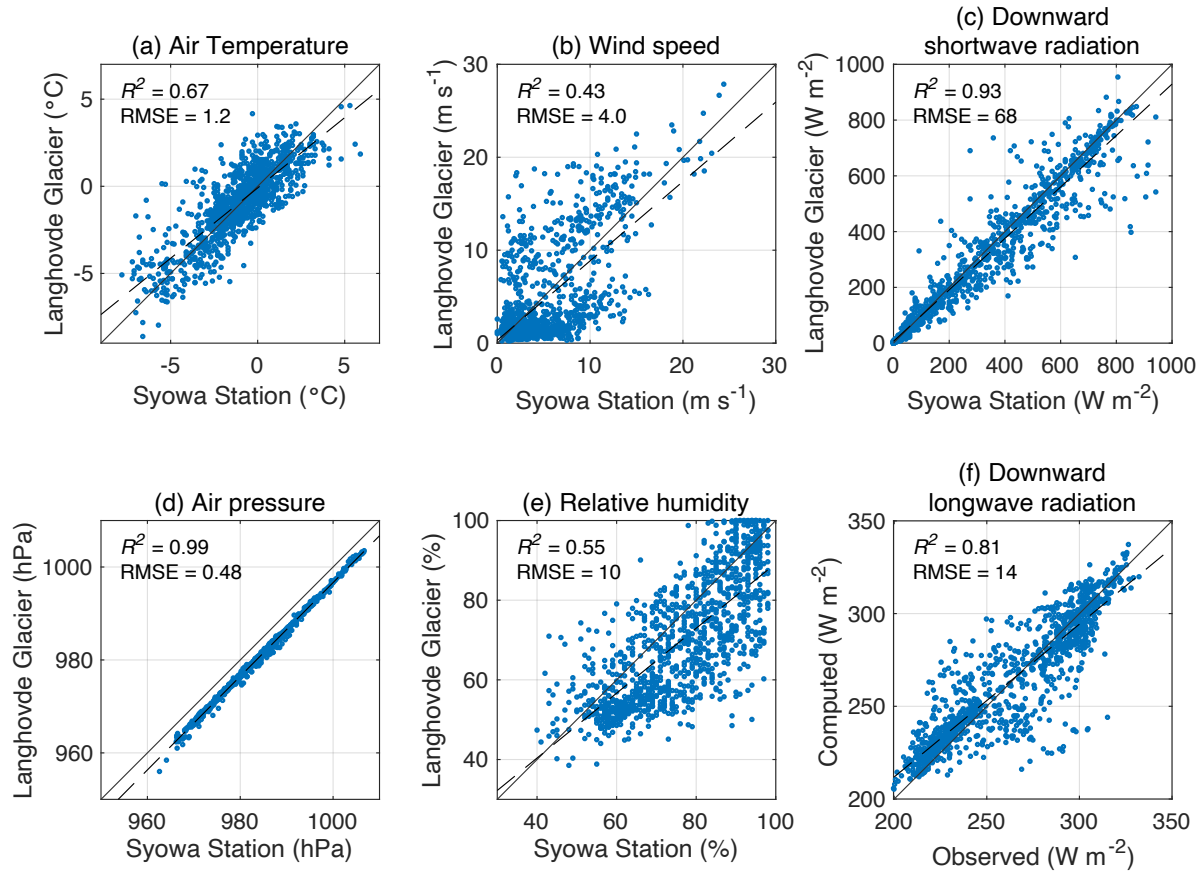


Figure 4.3: Scatter plot showing meteorological data observed at Syowa Station and at Langhovde Glacier. Coefficient of determination (R^2) and root mean square errors between the data observed at Syowa Station and Langhovde Glacier were given.

WXT536), as well as radiation components (downward and upward shortwave and longwave radiations) by a net radiometer (Kipp & Zonen, CNR-4). Glacier surface height was also measured by a supersonic distance meter (Campbell Scientific, Inc., SR50A). The weather sensors, the radiometer, and the distance meter were mounted at the heights of 2.0, 1.6, and 1.0 m from the ice surface, respectively, using a 4-m long aluminum pole drilled into the ice. The AWS was operated from 19 December 2021 to 6 February 2022 with an interval of 10 minutes. Sensor cables were connected to a data logger (Campbell Scientific, Inc., CR1000X) enclosed in a waterproof box powered by 12 V lead-acid batteries.

During 17–19 January, the measurement failed because the AWS fell down due to strong winds during a blizzard. The data gap was filled based on a linear relationship between the meteorological variables observed on the glacier and at Syowa Station (Fig. 4.3), except for longwave radiation which was not observed at Syowa Station. To fill the data gap, downward longwave radiation was estimated following the method using hours of sunlight (Kondo and others, 1991). Upward longwave radiation was computed using ice/snow surface temperature calculated by the surface energy balance model described in the following section.

4.2.6 Surface energy balance model

Glacier surface melt was estimated by the surface energy balance model using the meteorological variables observed on the glacier. The model was developed in the author's previous study (Kondo and others, 2021) based on equations that are commonly used for ice/snow surface melt (Kondo, 1994; Oerlemans, 2001). These formulations have been previously employed in studies of glacier mass balance (e.g. Fujita and Ageta, 2000; (Lisette) Klok and Oerlemans, 2002; Van AS, 2011; van Pelt and others, 2012) and glacier runoff (e.g. Liston and Mernild, 2012; Fujita and Sakai, 2014; Kondo and others, 2021). The energy flux available for melting (H_M) was expressed as the residual of the net shortwave radiation (S_{net}), net longwave radiation (L_{net}), sensible heat flux (H_S), latent heat flux (H_L), conductive heat flux due to rainfall (R), and conductive heat flux from subsurface (G), i.e.,

$$H_M = S_{\text{net}} + L_{\text{net}} + H_S + H_L + R + G. \quad (4.4)$$

The energy flux into the glacier was considered to be positive. Given that the contribution of the conductive heat fluxes from the subsurface and rainfall were relatively small during the melt season, they were ignored (Kondo and Yamazaki, 1990; Van AS, 2011). The sensible and latent heat fluxes were estimated using the bulk method and were expressed respectively as follows:

$$H_S = c_p \rho_a C_S U (T_a - T_s), \quad (4.5)$$

$$H_L = l \rho_a C_S U [h_r q(T_a) - q(T_s)], \quad (4.6)$$

where $c_p = 1006 \text{ J kg}^{-1} \text{ K}^{-1}$ represents the specific heat capacity of air under constant pressure, $C_S = 0.002$ represents the bulk coefficient for snow and ice surfaces proposed based on field data (Kondo and Yamazawa, 1986; Kondo, 1994) and previously used for glacier runoff modeling (e.g. Fujita and Sakai, 2014; Kondo and others, 2021), l represents the latent heat of evaporation of water ($2.5 \times 10^6 \text{ J kg}^{-1}$) when the ice surface is melting ($T_s = 0^\circ\text{C}$) or the latent heat of sublimation of ice ($2.83 \times 10^6 \text{ J kg}^{-1}$) during freezing ($T_s < 0^\circ\text{C}$). U , T_s , T_a , and h_r represent wind speed, surface temperature, air temperature, and relative humidity, respectively. In equation (4.4), as long as $H_M > 0$, the surface temperature, T_s was equal to 0°C . Otherwise, equation (4.4) was solved for T_s by substituting $H_M = 0$. The density of air (ρ_a) and the saturated specific humidity (q) were given by the following equations (Kondo, 1994):

$$\rho_a = 1.293 \frac{273.15}{273.15 + T_a} \left(\frac{p}{p_0} \right) \left(1 - 0.378 \frac{h_r e}{p} \right), \quad (4.7)$$

$$q = \frac{0.622e}{p - 0.378e}, \quad (4.8)$$

where p and $p_0 = 1013.2 \text{ hPa}$ are atmospheric pressure and its standard value. The saturated vapor water pressure (e) was calculated from air temperature values using Tetens formulation

(Tetens, 1930). The net shortwave radiation (S_{net}) and the net longwave radiation (L_{net}) were expressed as:

$$S_{\text{net}} = S_{\text{down}} - S_{\text{up}} \quad (4.9)$$

$$L_{\text{net}} = L_{\text{down}} - \varepsilon \sigma (T_s + 273.15)^4 \quad (4.10)$$

where S_{down} and L_{down} represent the downward shortwave and longwave radiation, respectively; σ , which represents the Stefan–Boltzmann constant, was equal to $5.67 \times 10^{-8} \text{ W m}^{-2} \text{ K}^{-4}$; and ε , which represents longwave emissivity, was assumed to be 0.98.

The surface melt rate, M was calculated in the water equivalent thickness using equation 4.11.

$$M = \frac{H_M}{\rho_w l_m} \quad (4.11)$$

where ρ_w and l_m represent the density of water (1000 kg m^{-3}) and the latent heat of fusion of ice ($3.34 \times 10^5 \text{ J kg}^{-1}$), respectively.

The model input variables were hourly data of air temperature, relative humidity, wind speed, atmospheric pressure, and upward and downward shortwave and longwave radiations measured at LG4. The model calculation was performed for the period from 20 December 2021 to 6 February 2022.

4.2.7 Subglacial drainage model

To better understand the variations in the subglacial hydrological system, the development of subglacial conduit was modelled based on the observed water pressure. The assumptions and equations were based on that described in Schoof (2010). A subglacial conduit was assumed to develop in two modes of a Röthlisberger channel and a cavity. The Röthlisberger channel was proposed by Röthlisberger (Röthlisberger, 1972) that assumes semi-circular shape drainage channels formed at the ice-bed interface. The channel evolves by a balance between the enlargement by melting of the ice wall due to heat dissipation of flowing water, and closure due to ice creep (Fig. 4.4a). The cavity forms on a lee side of the protrusion of bedrock as a result of horizontal glacier basal sliding, which opens a gap between the glacier ice and the bedrock (Fig. 4.4b). The gap size is controlled by the opening rate due to the glacier sliding rate and by the closure rate controlled by ice creep. These two forms of the development theories for subglacial conduit were synthesized into a single equation by Schoof (2010) as:

$$\frac{dS}{dt} = c_1 Q \Psi + u_b h - c_2 N^n S, \quad (4.12)$$

where Q is discharge through conduit, S is the cross-sectional area of the conduit, Ψ is hydraulic potential gradient along the conduit, u_b is sliding rate, h is a representative height of the protrusions on the bedrock, N is effective pressure, and c_1 and c_2 are constants given by material properties. The evolution of the conduit size is expressed by equation 4.12 as a bal-

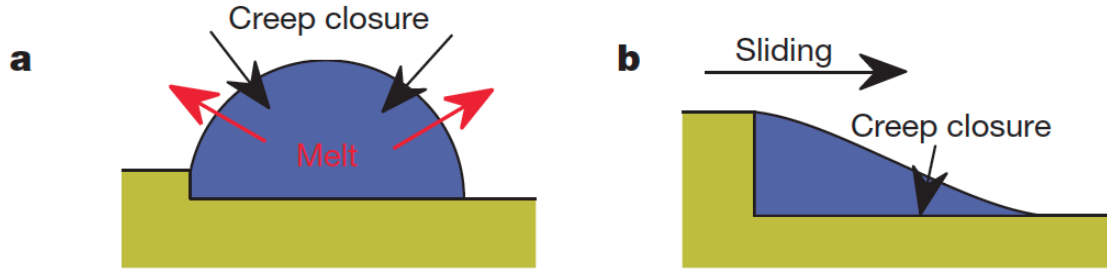


Figure 4.4: Schematic diagram showing processes controlling the development of (a) channels and (b) cavities. The step shows a bump in the bed topography. The basal ice flows from left to right.

ance between opening and closure rate. The first term of the right-hand side of the equation represents the opening rate of the conduit due to ice wall melting, the second is the opening rate of the cavities due to glacier sliding, and the third is the closure rate due to ice creep. The constants c_1 and c_2 are given by:

$$c_1 = \frac{1}{\rho_i L}, \quad (4.13)$$

$$c_2 = 2An^{-n}, \quad (4.14)$$

where $L = 3.35 \times 10^5 \text{ J kg}^{-1}$ is the latent heat of fusion of ice, $A = 6 \times 10^{-24} \text{ Pa}^{-3} \text{ s}^{-1}$ and $n = 3$ is rheological parameter of ice defined by Glen's flow law (Glen, 1955; Cuffey and Paterson, 2010). Effective pressure is given by ice overburden pressure (p_i) and water pressure in the conduit (p_w) as:

$$N = p_i - p_w. \quad (4.15)$$

Hydraulic gradient along the conduit is defined by glacier geometry and water pressure in the conduit, i.e.,

$$\Psi = -\rho_w g \frac{\partial b}{\partial s} - \frac{\partial p_w}{\partial s}, \quad (4.16)$$

where $\rho_w = 1000 \text{ kg m}^{-3}$ is density of water, $g = 9.81 \text{ m s}^{-2}$ is gravitational acceleration, b is bed elevation, and s is distance along the conduit. In terms of N , this can be written as:

$$\Psi = \Psi_0 + \frac{\partial N}{\partial s} \quad (4.17)$$

$$\Psi_0 = -\frac{\partial p_i}{\partial s} - \rho_w g \frac{\partial b}{\partial s}, \quad (4.18)$$

where Ψ_0 is the hydraulic gradient controlled by glacier geometry.

To relate hydraulic potential gradient to turbulent discharge, the Darcy-Weisbach law was used (Clarke, 1996) as:

$$Q = c_3 S^\alpha |\Psi^{-1/2}| \Psi, \quad (4.19)$$

$$\alpha = 5/4, \quad (4.20)$$

$$c_3 = 2^{1/4} \frac{\sqrt{\pi+2}}{\pi^{1/4} \sqrt{\rho_w f}}. \quad (4.21)$$

The conduit develops as a channel and a cavity with high and low discharges, respectively, depending on the dominant term in equation 4.12. The switching between the two modes occurs with the critical discharge Q_c expressed as:

$$Q_c = \frac{u_b h}{c_1(\alpha - 1)\Psi}. \quad (4.22)$$

The conduit behaves as a channel when $Q > Q_c$ due to the dominant term of the melt enlargement, whereas behaves as a cavity when $Q < Q_c$ because of the enlargement mainly due to glacier sliding.

Based on the glacier bed geometry (b) and the time series of the observed subglacial water pressure (p_w) revealed by borehole measurement, hydraulic gradient along the conduit (Ψ) was computed by equation 4.16. Equations 4.12 and 4.19 were solved forward in time to obtain unknown variables S and Q using Ψ and the prescribed initial channel size of $S(0) = 0.1 \text{ m}^2$.

4.3 Results

4.3.1 Hot-water drilling

Successful drilling at LGX resulted in two boreholes BH01 and BH02 penetrating through the entire ice and reaching the glacier bed. Observations with the CTD profiler and the borehole camera revealed that 553-m ice was grounded on the bedrock at this location. During the drilling, the water level in the borehole rapidly dropped at depths of 497 and 464 m at BH01 and BH02, respectively, suggesting that a connection with the subglacial hydrological system was established at 56 and 89 m above the glacier bed in each borehole. After the drilling, the water level at BH01 was at 83 m below the ice surface, indicating 22 m above sea level (Fig. 4.5).

After the drilling of BH03 at LG5, the water level in the borehole decreased to a level equivalent to the pressure exerted by the sea level. Subsequent water pressure measurements showed fluctuations of the water level in the borehole corresponded with the tidal variations observed at Syowa Station. The observations with the CTD profiler and the borehole camera showed that the 3-m-thick ocean layer underlain the 475-m-thick ice. Thus, the ice at LG05 was not grounded despite the location being 0.6 km upstream from the grounding line estimated by the break in surface slope. Given that the surface elevation at BH03 was well above the level expected for floating ice (Table 4.1), the ice was not in the hydrostatic equilibrium and supported by the surrounding ice, forming a complex geometry in the grounding zone as previously reported (Smith, 1991; Fricker and Padman, 2006; Sugiyama and others, 2014).

Hot-water drilling revealed ice and bed geometries along the longitudinal cross-section of Langhovde Glacier. Combined with the geometry over the ice shelf revealed by the previous drilling in 2011/12 (Sugiyama and others, 2014) and 2017/18 (Minowa and others, 2021), I obtained 4.2-km-long longitudinal geometry profile extending from the calving front to LGX, located at 1.5 km upstream from the grounding line (Fig. 4.5) (Table 4.1). The ice was grounded on the bedrock at LGX and a 3-m-thick ocean layer was found beneath the ice at BH03, suggesting the grounding line location within the 0.9 km between the two drilling sites. The bed topography deepens upglacier by 38 m from LG5 to LGX at a slope of 2.4° . Taking into account the previously reported sub-shelf topography, the bed progressively deepens from LG4 to LGX by 59 m within the 1.8-km distance at a mean slope of 1.8° . The bedrock near the grounding zone formed a bump within 1.5–4 km from the calving front (Fig. 4.5).

4.3.2 Subglacial environment

After the drilling operations were completed, the stainless-steel nozzles used for the drilling at BH01 and BH02 were damaged with scratches on the tip (Fig. 4.6). The nozzles were impacted against the borehole bottom several times to confirm it reached the glacier bed, thus,

Table 4.1: Glacier geometry at the drilling sites. The data in 2011/12 (BH1201–3) and 2017/18 (BH1801–4) were taken after Sugiyama and others (2014) and (Minowa and others, 2021), respectively.

Borehole	Ice-surface elevation (m a.s.l.)	Ice thickness (m)	Bed elevation (m)	Ice thickness under hydrostatic equilibrium (m)	Drilling date
BH1801	21.7	234	−515	241	16 Jan. 2018
BH1802	25.8	294	−508	287	9 Jan. 2018
BH1803	31.5	389	−468	349	31 Dec. 2017
BH1804	37.3	412	−388	414	22–23 Jan. 2017
BH1201	40.8	398	−382	453	6 Jan. 2012
BH1203	52.9	431	−390	588	16 Jan. 2012
BH03	69.3	475	−409	770	22–23 Jan. 2022
BH01	105.0	553	−447	1167	25–31 Dec. 2021

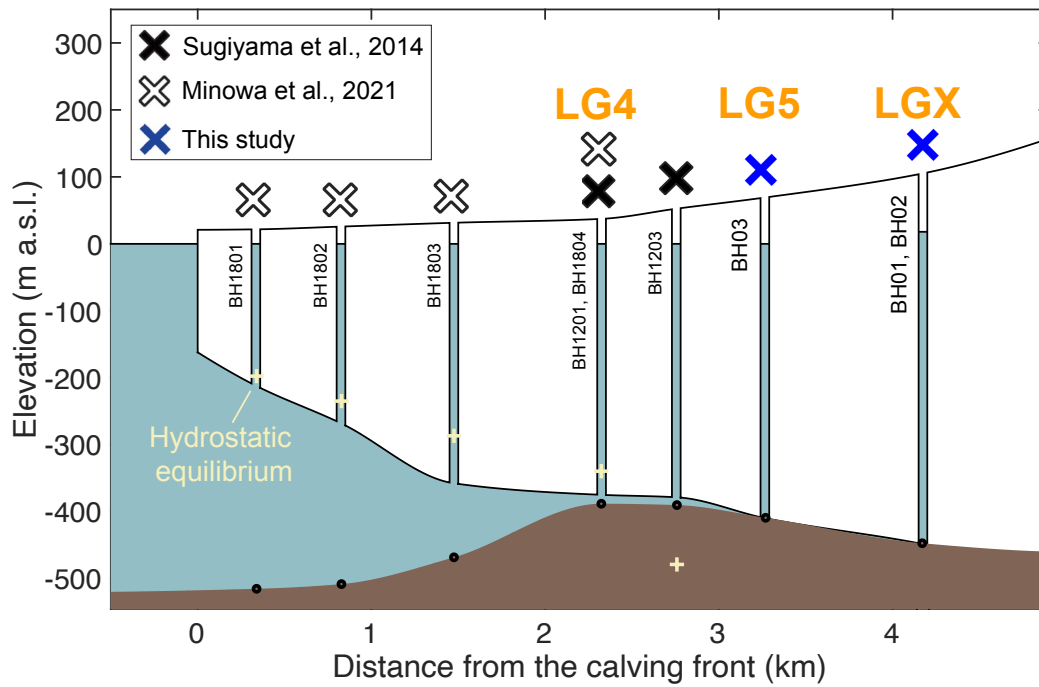


Figure 4.5: Vertical cross-section of Langhovde Glacier along the boreholes drilled in 2011/12 (black crosses), 2017/18 (white crosses), and 2021/22 (blue crosses). The yellow crosses show ice bottom elevation under the hydrostatic equilibrium assumption estimated from the surface elevation at the drilling sites.

the damages appear to have been made at the glacier bed. The damaged nozzle tip suggested that the bed consisted of a hard material that scratches stainless steel nozzles. The surfaces of the nozzles were abraded after the drilling (Figs 4.6b and d), which were assumed to be made by rubbing against the glacier bed.

Visual inspection with a video camera lowered into the borehole provided information on the englacial and subglacial environments. At the 553-m-long borehole BH01, the wall shallower than 500 m was clean and opaque due to bubble inclusions (Fig. 4.7a). The color of the wall began to change to transparent at the depth of 500 m (Fig. 4.7b) and became darker in the deeper region, probably due to the sediment inclusion (Fig. 4.7c–f). From this depth, sand or silt-like particles were observed in the ice wall (Fig. 4.7c). In some places, the particles that melted out from the wall were deposited on small steps formed on the borehole wall (e.g. Fig. 4.7d). From 525 to 540 m, many fractures and cracks were observed in the wall of the borehole, indicating the borehole passed through a region of extensive fracturing. In this zone, the largest englacial crack across the borehole was found at a depth of 530 m (Fig. 4.7d). The crack coincided with the depth at which the water level in the borehole rapidly dropped during the drilling, suggesting that the connection with the subglacial hydrological system was established through the intersection with the basal crack. Within 5 m from the borehole bottom, fine gravel and fine-grained sediment were entrained in the wall (Fig. 4.7e). The glacier bed was confirmed at the borehole bottom, which appeared to be covered by a fine-grained sediment layer (Fig. 4.7).

4.3.3 Glacier surface motion

Mean ice speeds during the observation period were 0.29 and 0.24 m d^{-1} at LG4 and LGX, respectively. These speeds are equivalent to 105 and 88 m a^{-1} , respectively, thus consistent with the speeds obtained from the satellite image analysis at each location (Figs 2.2d and 3.4e). Continuous GNSS survey revealed two short-term speedup events. Relatively fast ice flow was observed during 19–27 December, within a week after the installation of the GNSS (Fig. 4.8a). The speed reached 0.34 m d^{-1} and 0.28 m d^{-1} at LG4 and LGX, respectively, corresponding to 17 and 21% greater than the background values during the observation period. Another speedup event was observed during 2–7 January and the maximum ice speed reached 0.31 and 0.26 m d^{-1} at LG4 and LGX. At both sites, the magnitude of the acceleration (0.45 m h^{-1} at LGX) was greater than that of deceleration (-0.19 m h^{-1}).

Ice speed showed short-term fluctuations within 7% of the mean speed. Along-flow positions of the GNSS stations were in anti-phase with tides, showing the maximum/minimum detrended along-flow position during low/high tides at both sites (Fig. 4.9). The fluctuations in the ice speed represent speed increase/decrease during the rising/falling tides (Fig. 4.8a). During the glacier speedup from 2 to 8 January, the periodicity in the along-flow position deviated from the diurnal tidal signals (Figs 4.8a and 4.9).

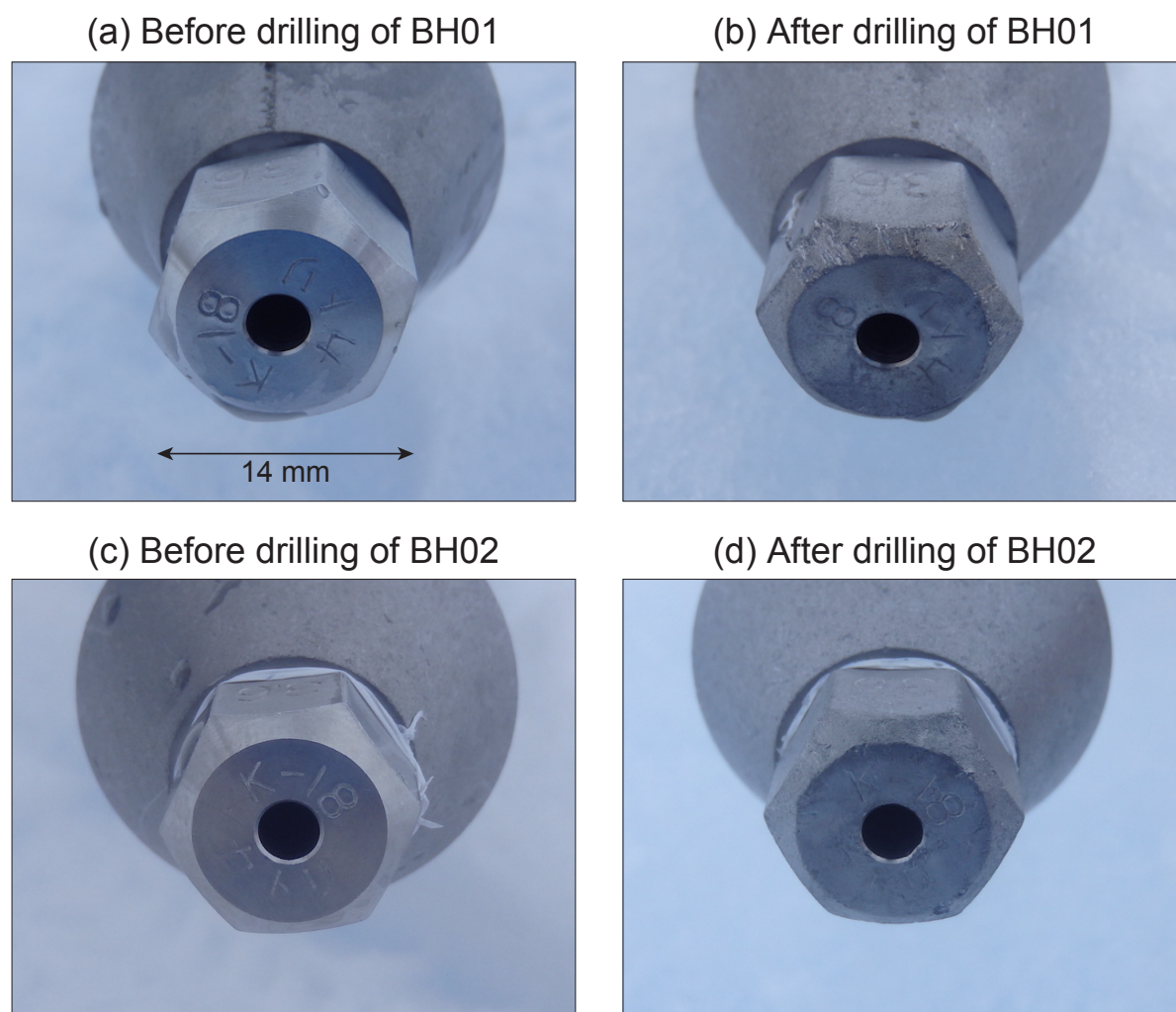


Figure 4.6: Photographs showing drilling nozzles used for the hot-water drilling at (a, b) BH01 and (c, d) BH02 (a, c) before and (b, d) after the operations.

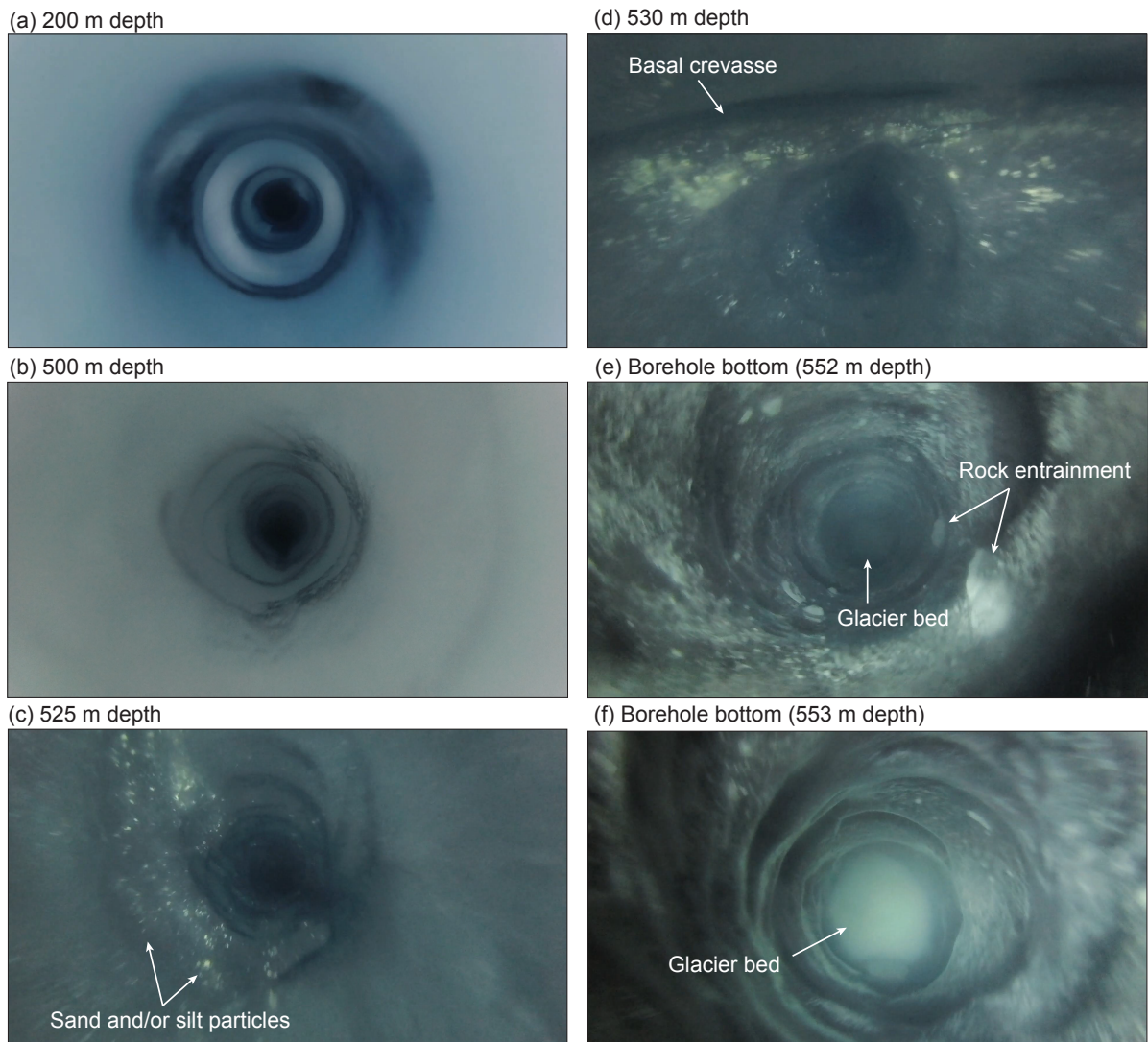


Figure 4.7: Images obtained by the video camera lowered into the borehole BH01 showing (a) 200 m depth, (b) 500 m depth, (c) 525 m depth, (d) 530 depth, (e) 1 m above the borehole bottom, and (f) the borehole bottom.

Upward glacier surface motion was observed twice at LGX, corresponding with the periods of the glacier speedup. The most significant uplift by 0.12 m was observed within a week during 20–26 January. After the uplift, the surface motion turned downward and was followed by another smaller surface uplift by 0.01 m during 3–5 January, which corresponded with the second speedup event observed between 2 and 7 January. The surface showed upward motion again by 0.03 m during 20–26 January. At LG4, located on the floating ice shelf, the vertical displacement was stable within 0.03 m.

4.3.4 Subglacial water pressure

Subglacial water pressure varied within a range of 4.43–4.77 MPa throughout the observation period (Fig. 4.8c), which is equivalent to a water level of 452–487 m above the bed and 16–51 m above sea level. These observations indicate that the water pressure was at 90–97% of the ice overburden pressure and the effective pressure was only 140–480 kPa throughout the period. After the rain event observed on 2 January, the water level of the borehole rapidly increased by 22 m within 38 hours, reaching the maximum pressure during the observation period (Fig. 4.8c). After the pressure peak, the water level gradually decreased until 13 January. Thereafter, the pressure showed fluctuation, which corresponded with the tidal variations observed at Syowa Station (Figs 4.8c and e). During the period, cross-correlation analysis showed no lag greater than the detection limit of one hour between the subglacial water pressure and the tide. The pressure at the bottom of BH02 quickly rose by 500 kPa from 22 to 23 January, and fluctuated at a similar level of the ice overburden pressure until the end of the observation period (Fig. 4.8c). The pressure at BH01 also fluctuated wildly from 29 January to 1 February, which was followed by a rapid increase of 980 kPa, exceeding the ice overburden until the end of the observation.

4.3.5 Ploughmeter data

Tilt variations

The tilt of the ploughmeter indicated that the device was first installed at the borehole bottom within 1.9° from the vertical axis (Fig. 4.8d). During the one-month observation period, the ploughmeter tilted by up to 30.9° . The variations in the tilt indicate that the ploughmeter moved in association with the glacier basal motion. The tilt variations were particularly complex during the first nine days. After the installation, the ploughmeter tilted rapidly by 2.4° within 5 seconds on 2 January. Thereafter, sawtooth variations were observed from 4 to 9 January, which were characterized by gradual tilting over ~ 5 hours as a mean and rapid reversal in several seconds (~ 2 s) (Figs 4.8d and f). Typical tilt rates differed between the two periods 4–6 and 6–9 January, with mean rates of 5.2 and 3.1° d^{-1} , respectively, showing faster tilt in the earlier period. After the sawtooth-shaped variations, the ploughmeter leaned steadily with a

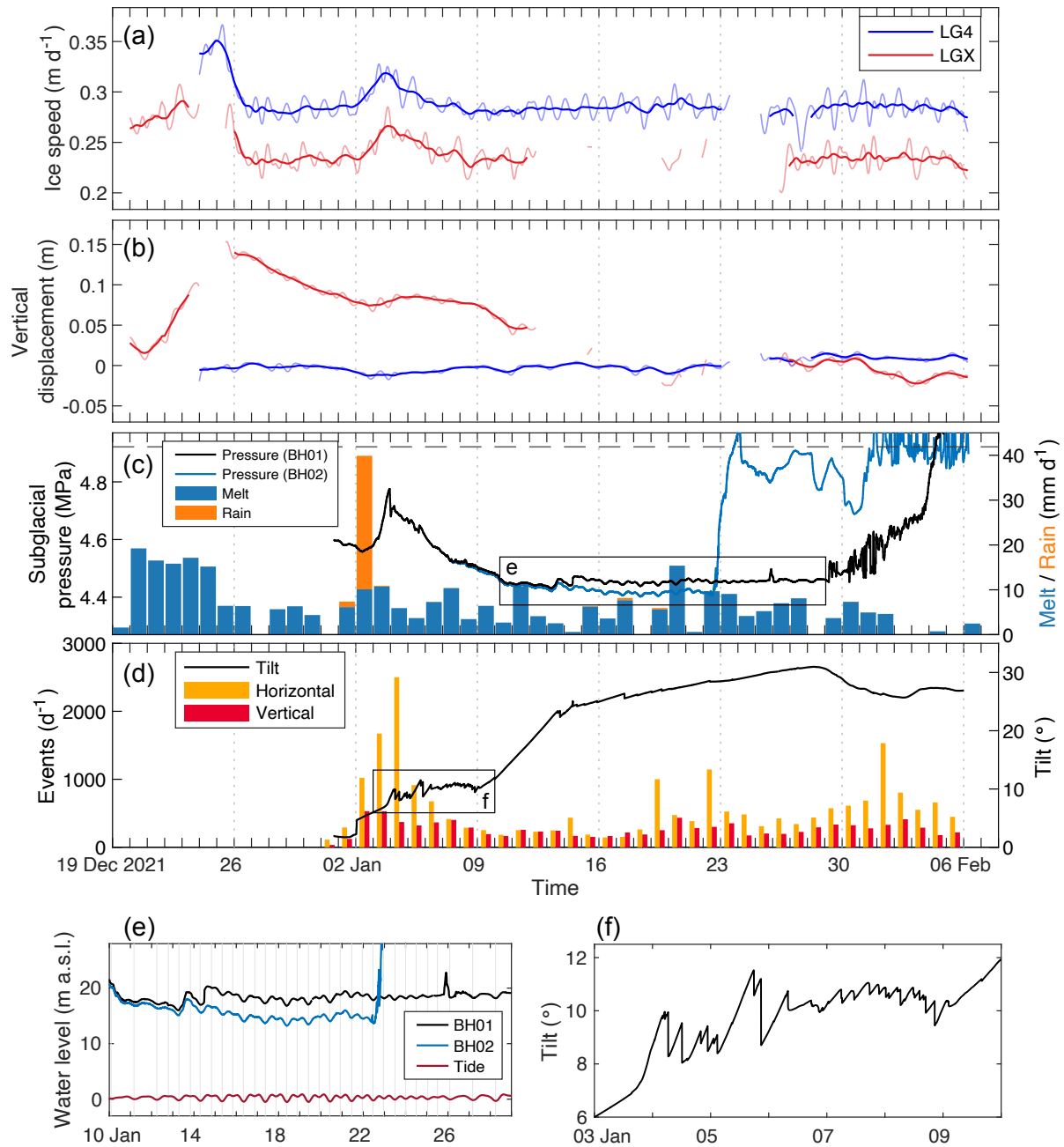


Figure 4.8: (a) Glacier surface speed observed at Sites 1 (light blue) and 2 (light red) and 1-day running mean speed (blue and red). (b) Vertical glacier surface displacement observed at Sites 1 (light blue) and 2 (light red) and 1-day running mean vertical displacement (blue and red). (c) Subglacial water pressure measured at BH01 (black line) and BH02 (blue line). The dashed line indicates the ice overburden pressure at LGX. Also shown are glacier surface melt computed by the energy balance model (blue bar) and rain observed on the glacier (orange). (d) The tilt of the ploughmeter (solid line) and the number of detected vibration events on horizontal (orange) and vertical (red) axes are shown. (e) Subglacial water pressure observed at BH01 (black) and BH02 (light blue) during 10–29 January were shown. Ocean tide observed at Syowa Station (red) was also shown. The vertical gray lines indicate the timing of low tide. (f) Tilt of the ploughmeter during 3–10 January was shown.

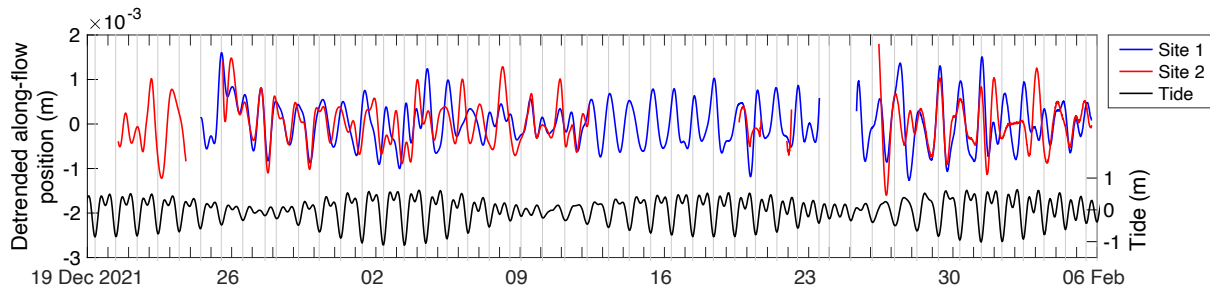


Figure 4.9: Detrended along-flow position observed at Site 1 (blue) and Site 2 (red). Ocean tide observed at Syowa Station (black) was also shown. The vertical gray lines indicate the timing of low tide.

continued rate of $3.1^{\circ} \text{ d}^{-1}$ from 9 January and reached 23.4° on 13 January. The ploughmeter tilted by 13.4° during the five-day period, accounting for 43% of the total tilt over the study period. Thereafter, the tilting continued steadily at a slower rate of $0.9^{\circ} \text{ d}^{-1}$ until 28 January, reaching its maximum of 30.9° . Thereafter, the increasing trend turned into a decreasing, and the tilt decreased by 5.2° , until 2 February, which was followed by a slight increase by the end of the observation.

Vibration events

The detected vibration events were characterized by impulsive signals spanning a wide frequency band up to the Nyquist frequency (e.g. Fig. 4.10). The characteristics of the signals substantially differ between before and after 3 January 13:00. Before 3 January, the impulsive events have strong signals across the full frequencies, with two elevated power peaks at 6 and 25 Hz. According to the spectrogram, these two peaks were also dominant in the ambient noise recorded between the impulsive events. The events have relatively long duration with the most frequent value of 1.2 s. After 3 January, impulsive signals tend to show shorter duration (~ 0.4 s) spanning the full frequency range with the dominant frequency of 16 Hz.

Vibration events occurred twice more frequently on the horizontal axes than on the vertical axes (Fig. 4.8d). The sensor detected 22,000 vibration events on the horizontal axes over the 27 days of observation, with a mean occurrence rate of 800 d^{-1} . A large number of events were observed between 2 and 7 January accounting for 33% of the total events during the 6 days period. The greatest occurrence of $2,500 \text{ d}^{-1}$ (11% of total events) was recorded on 4 January, corresponding with the rise in the subglacial water pressure and the glacier acceleration. The number of horizontal vibration events was smaller during 9–19 January (a mean rate of 234 d^{-1}), whereas increased between 29 January and 4 February (774 d^{-1}). The occurrence exceeded $1,000 \text{ events d}^{-1}$ on 19, 22 January, and 1 February. On the vertical axis, fewer vibration events and more stable variations were recorded compared to those on the horizontal axis.

Horizontal vibration events occurred most frequently during 4–5 January. The number of vibration events increased when the tilt angle rapidly decreased (Figs 4.11b and c). The num-

ber of the events decreased to $0\text{--}1\text{ min}^{-1}$ within several hours from the tilt drop, as the tilt progressively increased (Fig. 4.11b). Exceptionally high peak amplitudes reaching $>\sim 100\text{ mG}$ were recorded during the decrease in the tilt angle (Fig. 4.11c). This event was followed by numerous weaker events concentrated at $\sim 10^{-3}\text{ mG}$ (Fig. 4.11c). The interval of these events increased from $\sim 1\text{ s}$ to $\sim 100\text{ s}$ as the tilt of the ploughmeter increased and the peak amplitude decreased. Such variations were repeatedly observed corresponding with the sawtooth variations of the tilt of the ploughmeter. In the spectrogram, the elevated power of the events ranged within $15\text{--}22\text{ Hz}$ (Fig. 4.11d). After the decrease in the tilt angle, the dominant frequency was observed at 15 Hz . As the tilt increased, the frequency gradually increased to $\sim 22\text{ Hz}$. The shifts in the dominant frequency from 15 to 22 Hz were also repeatedly observed along with the sawtooth variations in the tilt.

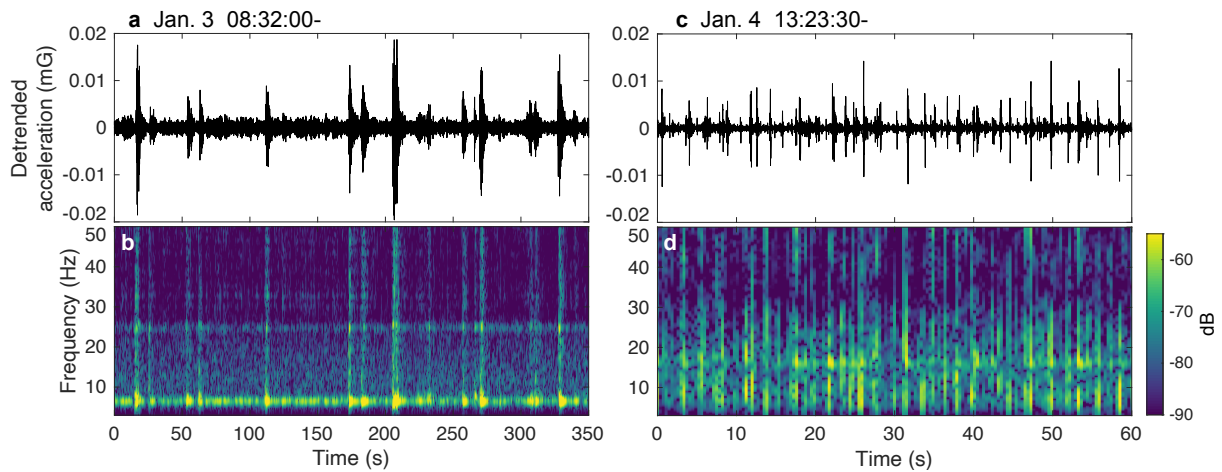


Figure 4.10: Example of the acceleration signals recorded by the ploughmeter (horizontal component). The data from (a) 3 January 08:32:00 and (c) 4 January 13:23:30, and its spectrogram (b and d) were shown after applying a band-pass filter between 1 and 50 Hz.

Spectral characteristics

The spectrum of the acceleration data throughout the observation period was shown as PSD-PDFs in Fig. 4.12. Both in the horizontal and vertical axes, a general increase in the ambient noise was recorded below 1 Hz , which can be expected from globally observable secondary microseisms such as noise induced by ocean waves (e.g. McNamara, 2004). Such low-frequency signals can be also influenced by changes in the gravitational acceleration due to transient variations in the ploughmeter attitude. Another similarity between horizontal and vertical axes was observed above 15 Hz which showed suppressed energy up to 50 Hz and assumed to be influenced by the measurement limit near the Nyquist frequency. The power on the horizontal component elevated within the frequency range of $3\text{--}15\text{ Hz}$ with a peak frequency of 7 Hz (Fig. 4.12a). Within the range, 3.1 dB higher median power was recorded on the horizontal axis than on the vertical axis (Fig. 4.12c). The spectrum of the vertical signals was more

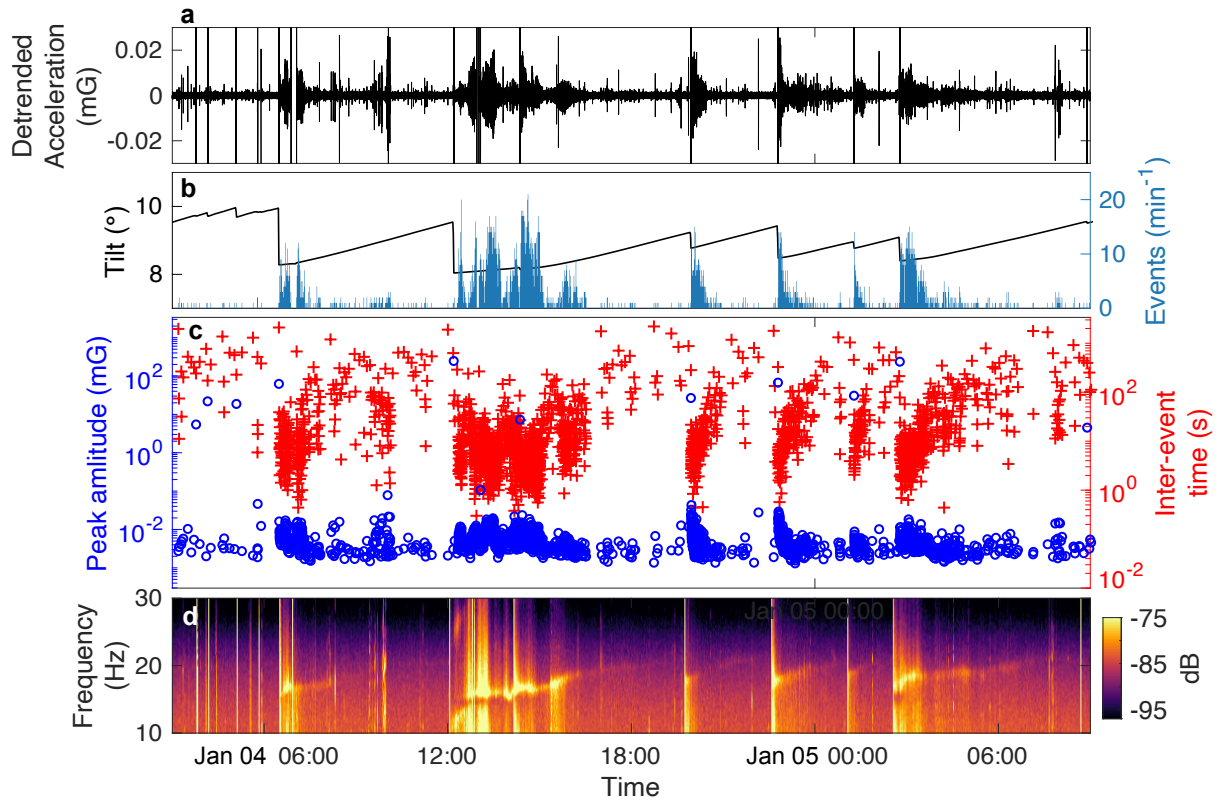


Figure 4.11: (a) Horizontal acceleration data after applying a band-pass filter between 1 and 50 Hz. (b) Tilt of the ploughmeter (solid line) and the number of the detected vibration events (bar). (c) Peak amplitude (blue) and inter-event time (red) of the vibration events. (d) Spectrogram of the high-pass filtered acceleration data.

stable, showing the less variable distribution concentrated on their median values (Figs 4.12b and c).

The spectrogram of the acceleration data showed distinct changes in the temporal variations of the power distribution during the observation. After the installation, relatively high power was recorded with dominant frequencies at 6 and 25 Hz on the horizontal axis until 4 January (Fig. 4.13a). The spectral characteristics changed from 3 January with the disappearance of the concentrated power at 6 and 25 Hz. Many impulsive increase in the power was observed during 4–8 January, corresponding with the large variability in the tilt of the device (Fig. 4.8d), indicating the vibrations of the device were recorded associated with the movement of the device. After the period of relatively stable ambient noise variations during 9–13 January, elevated power was observed at 6 and 15 Hz by the end of the observation. On the vertical axis, stable power distribution was observed were observed throughout the observation than on the horizontal axes (Fig. 4.13b). Impulsive vibration events were recorded over the full frequency range during 4–14 January as observed on the horizontal components associated with the movement of the device. Diurnal variations were observed during 8–15 January and 25 January–6 February in the background noise with typical fluctuations of ~ 1.0 dB (Fig. 4.13b). The variations were observed over the frequency range from 2 to 50 Hz, and the greatest amplitude was found within 4–20 Hz. The signals were strongest during the daytime (0–3

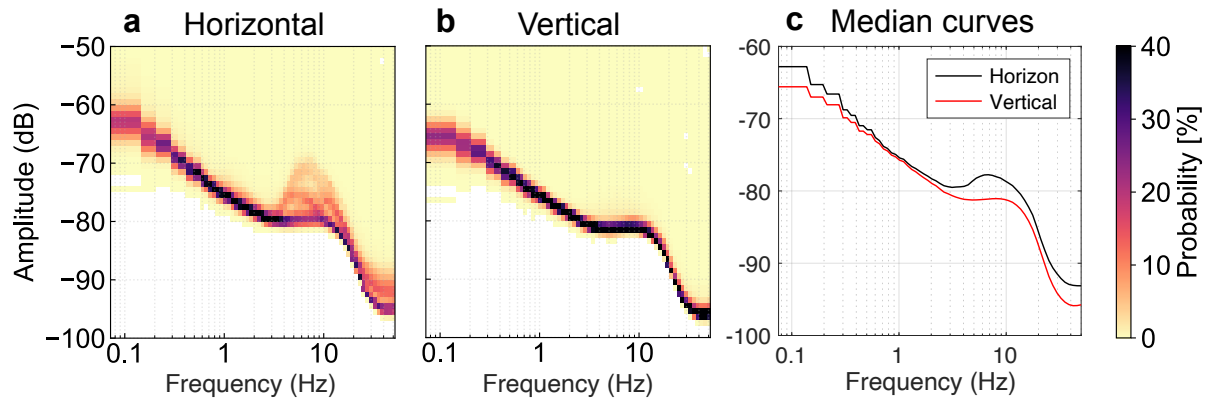


Figure 4.12: PSD-PDFs of the acceleration data recorded on the horizontal axis during (a) 31 December 2021–3 January 2022, (b) 3–14 January, and (c) 14–26 January, and (d) its median values.

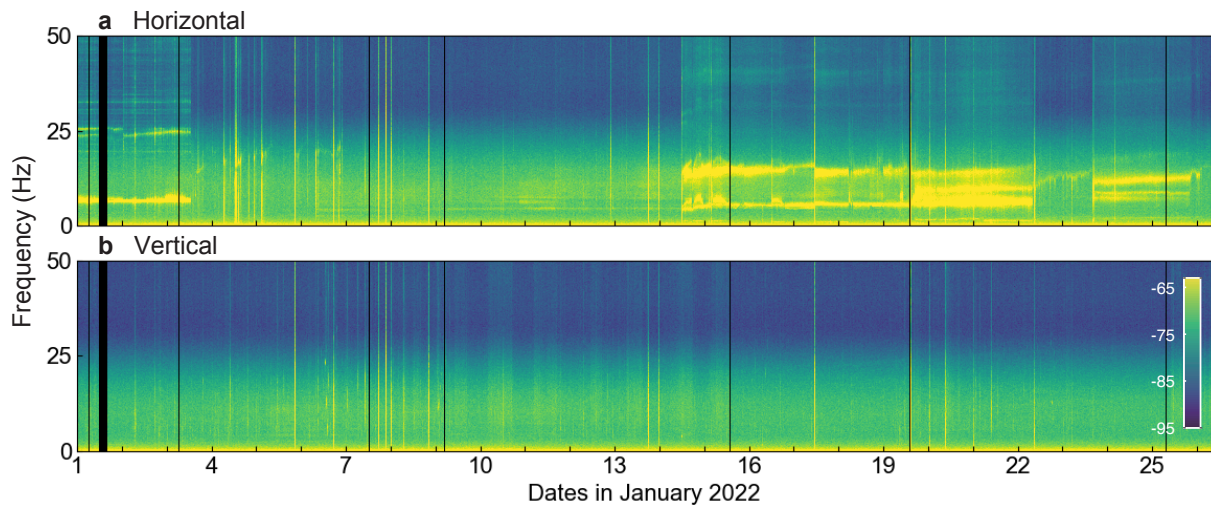


Figure 4.13: Spectrogram of the acceleration data observed on (a) horizontal and (c) vertical axes.

p.m. in local time, UTC+3) and took minimum during the night (1–7 a.m. in local time).

Temporal variations in the vertical component of the seismic tremor were further analyzed within the range of 3–10 Hz, where the magnitude of the diurnal variations was particularly large. The amplitude of the tremor varied within ~ 2 dB throughout the observation period (Fig. 4.14). The diurnal variations were observed from 8 January with a mean fluctuation of ~ 1 dB peaking during daytime. Such variations were absent during 16–25 January, except for an increase of the power on 20 January by 1.3 dB. In addition to the diurnal variations, a relatively high magnitude was observed during 3–7 January with an elevated signal floor by 1 dB compared to the background value. The onset of the signal increase coincided with the elevated subglacial water pressure after the rain event on the glacier (Fig. 4.8c). The increase in the amplitude was relatively rapid within 3 days from 4–7 January, compared to the decreasing phase over 7 days from 7–14 January.

The diurnal variations in the magnitude of the seismic tremor during 8–15 January and 25 January–6 February correlated well with the air temperature variations observed on the glacier surface (Fig. 4.14). An especially high correlation of $r = 0.77$ ($p < 10^{-5}$) was observed

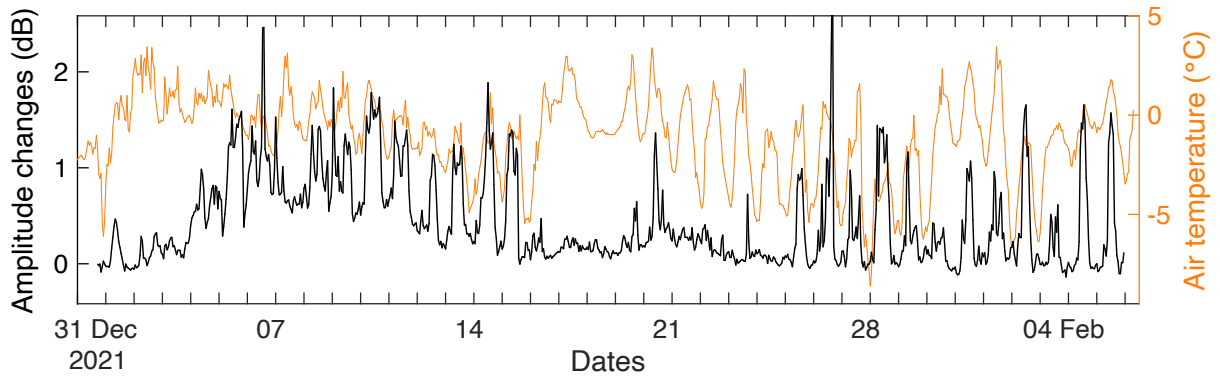


Figure 4.14: Seismic tremor amplitude observed by the subglacial accelerometer (black) and air temperature observed on the glacier surface (orange).

during 8–15 January, with similar variations in phase and magnitude. The correlation in 25 January–6 February was weaker ($r = 0.32$, $p < 10^{-5}$) than that in 8–15 January. Nevertheless, the phase of the variations closely coincided with the air temperature. The disappearance of the diurnal variations on 16 January corresponded with the blizzard as observed by the elevated air temperature (Fig. 4.14) and the wind speeds (Fig. 4.16b) on the glacier. The increase of the signal amplitude on 20 January corresponded with the increase in the glacier melt owing to increased net shortwave radiation flux as well as sensible heat flux toward the glacier surface (Fig. 4.17).

4.3.6 Modelled subglacial discharge

To better understand the evolution of the subglacial drainage system, the development of the subglacial conduit was modelled throughout the observation period based on the observed subglacial water pressure. The most substantial changes in the subglacial conduit were driven by the elevated subglacial water pressure peaking on 3 January after the rain event. The increased water pressure caused a steepening of the hydraulic gradient and resulted in a rise in the discharge through the conduit from 4 January. The model results indicated that the increased water flow caused enlargement of the channels by heat dissipation of the flowing water, which further increases discharge. Despite the pressure decreasing from 4 January, the rise in the discharge continued by 9 January, reaching the maximum of $7.4 \text{ m}^3 \text{ s}^{-1}$ owing to the sustained enlargement of the channel size. Subglacial water pressure decreased to the lowest level by 11 January, which resulted in a decrease in the discharge by 16 January. After the decrease, the discharge varied around $2.5 \text{ m}^3 \text{ s}^{-1}$ by the end of the calculation. The size of the conduit showed a gradual decrease from 11 January to the end due to the increased effective pressure and following sustained ice creep closure.

The modelled subglacial discharge was compared with the variations in the seismic tremor amplitude observed on the vertical acceleration (Fig. 4.15). The simulated increase in the subglacial discharge from 4 to 9 January corresponded with the increase in the tremor amplitude.

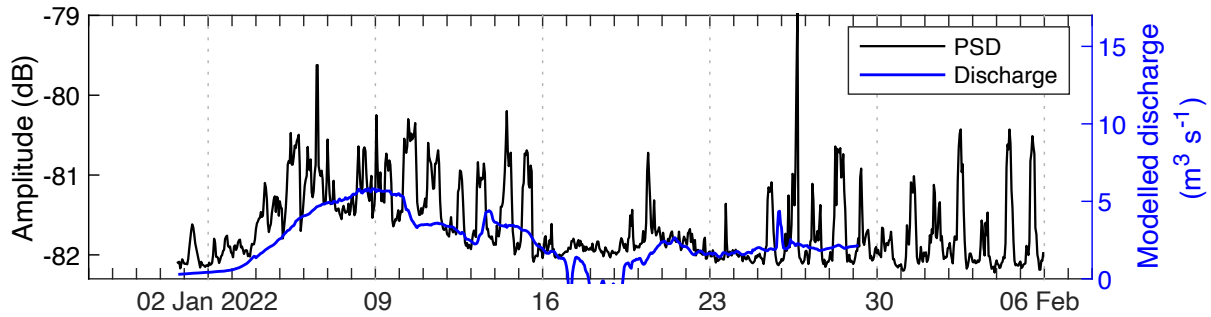


Figure 4.15: Seismic tremor amplitude of the vertical component of the acceleration observed by the subglacial accelerometer (black) and modelled subglacial discharge (blue).

Thereafter, the discharge decreased along with the daily minimum of the tremor amplitude by 16 January. Diurnal variations were not observed in the modelled discharge unlike the tremor amplitude. The tremor amplitude correlated well ($r = 0.91$, $p < 10^{-280}$) with the modelled discharge when a bandpass filter was applied to extract long-term fluctuations of more than three days.

4.3.7 Meteorological conditions

The mean air temperature on the glacier was -0.9°C during the observation period from 19 December 2021 to 6 February 2022. A relatively warm period was observed during 18–23 December, reaching the maximum air temperature of 6.2°C on 19 December with the mean of 0.6°C (Fig. 4.16a). Concurrent high downward shortwave radiation (Fig. 4.16c) and low downward longwave radiation (Fig. 4.16d) indicate a clear-sky condition during the period. On 2 January, rainfall of 30 mm d^{-1} was observed on the glacier (Fig. 4.16). This event was accompanied by relatively high temperature (2.8°C) and strong wind (15.9 m s^{-1}), indicating the passage of a low-pressure system over Lützow-Holm Bay. A blizzard was observed during 16–19 January with considerably large wind speeds reaching 33 m s^{-1} on 18 January which caused the failure of the AWS on Langhovde Glacier. The strong wind was associated with elevated air temperature, suppressed downward shortwave radiation, and a rise in downward longwave radiation. These meteorological conditions suggest that the severe weather was caused by a cyclonic atmospheric pressure pattern.

Surface energy balance showed the mean melt rate of 7.1 mm d^{-1} over the entire observation period. A large surface melt rate was simulated in the first five days during 20–24 January with the mean melt rate of 18.6 mm d^{-1} (Fig. 4.17). The intensive melt was caused by elevated heat flux toward the glacier surface due to the high downward shortwave radiation during the period (Fig. 4.17 and 4.16c). The heat flux showed a negative value on 18 January during the blizzard, followed by a relatively high melt rate of 17.1 mm d^{-1} on 19 January. Thereafter, the daily mean energy flux progressively decreased and became below 0 on 2 February.

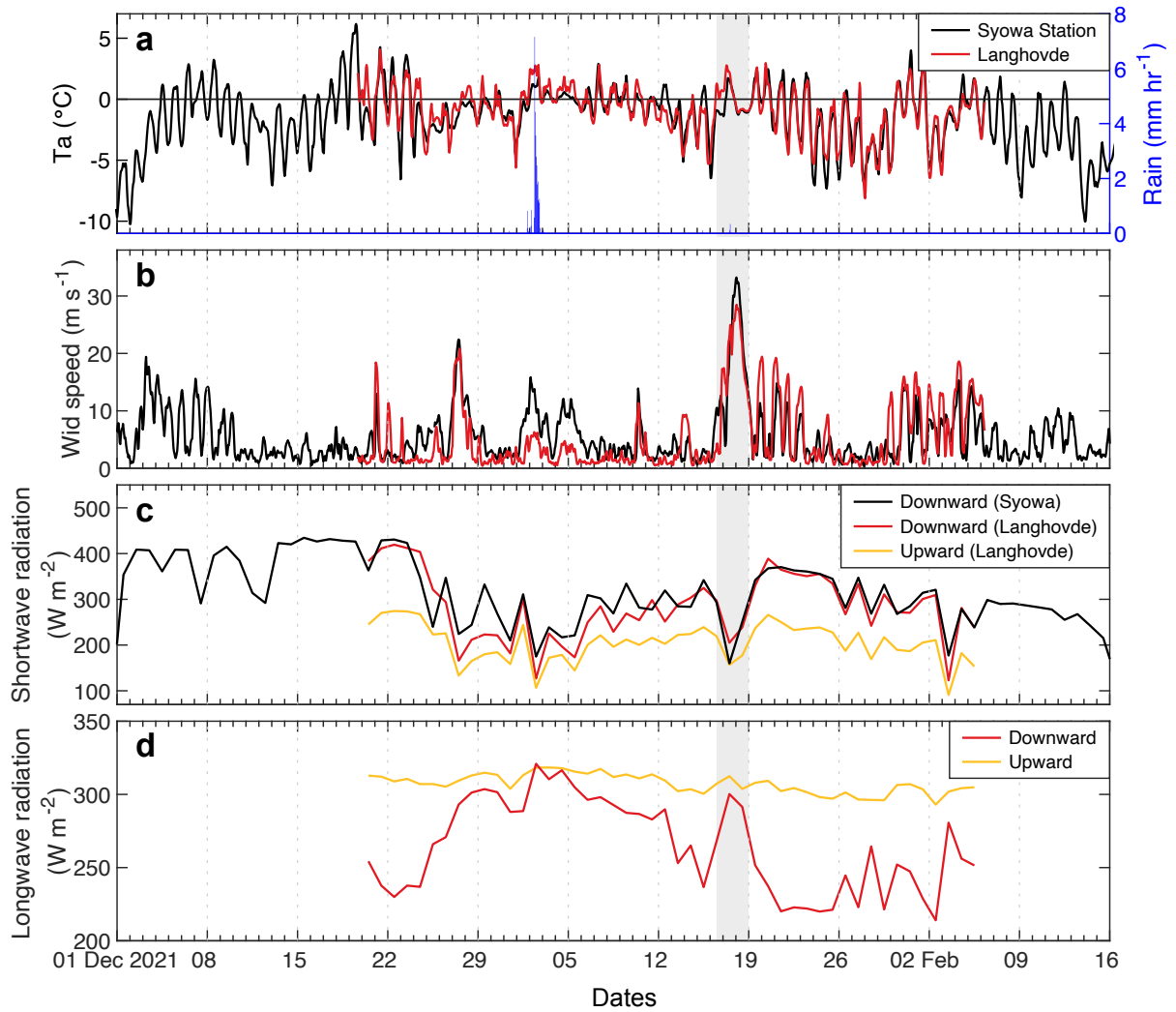


Figure 4.16: Hourly (a, b) and daily (c, d) meteorological data observed at Syowa station and Langhovde Glacier for (a) air temperature (solid line) and rainfall (bar), (b) wind speed, (c) shortwave radiation, and (d) longwave radiation. The shaded area indicates the period of the blizzard.

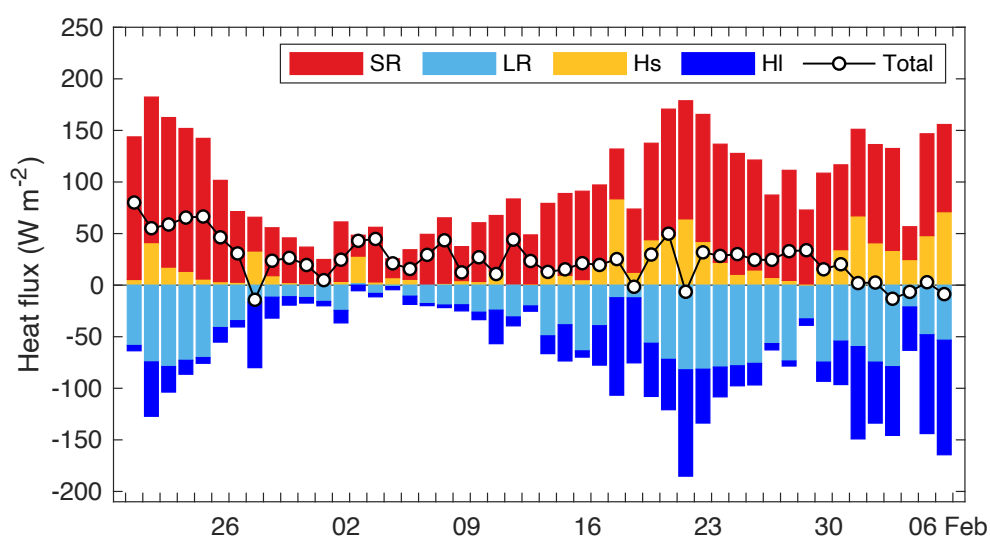


Figure 4.17: Energy balance of the glacier surface calculated from the surface energy balance model (circle). Components of the energy balance were shown as net shortwave radiation (red), net longwave radiation (light blue), sensible heat flux (yellow), and latent heat flux (blue).

4.4 Discussion

4.4.1 Subglacial and englacial conditions

Borehole observations at LGX revealed the nature of the subglacial bed conditions. After the drilling, the tip of the jet nozzles used for the drilling was damaged and scratched (Figs 4.6b and d), which was assumed to be made by the impact against the glacier bed. The observation suggests that the subglacial bed was composed of hard material that scratches stainless steel nozzles. In addition to the scar, the surfaces of the nozzles were abraded after the drilling (Figs 4.6b and d). The abrasion of the nozzles is assumed to be made by rubbing against the fine-grained sediments, suggesting the bed was covered with a thin sediment layer. Although observations by the borehole camera showed that the glacier bed was covered with fine-grained sediment, the scratches on the nozzles and clear water filling the borehole suggest a thick (e.g. >1–2 m) sediment layer is lacking.

Observations by the borehole camera revealed cracks along the borehole wall at depths of 525–540 m. The connection of the borehole with the subglacial drainage system was established through the intersection with the largest crack at 530-m depth (Fig. 4.7d). Given that the crack was connected to the basal hydrological system, it was extended more than 20 m from the bed and assumed to be a basal crevasse originating from the base. Such basal crevasses were proposed to be formed associated with strong longitudinal stretching of ice and/or high basal water pressure close to the ice overburden (van der Veen, 1998). Although the longitudinal strain rate is not substantially large at this location ($0.9 \times 10^{-2} \text{ a}^{-1}$ between LG4 and LGX), the subglacial water pressure never dropped below 90% of the ice overburden pressure. Such sustained low effective pressure is a favorable condition for the formation of basal crevasses (van der Veen, 1998). The existence of the basal drainage system connected by crevasses indicates that the hydrological system develops not only at the ice-bed interface, but also vertically in three dimensions into the ice. Such vertical extension of the drainage system due to crevasses was also found in a land-terminating glacier in Alaska, demonstrating considerable amount of water was accommodated in the crevasses which was equivalent to decimeter-thick water covering the glacier bed (Harper and others, 2010). Observations in Storglaciären also found water-filled fractures at depths from near the surface to 96% of the glacier thickness, which was assumed to form a large portion of the glacier hydrological system that conveys water (Fountain and others, 2005). The observations indicate that such hydrological conditions could be more widespread on calving glaciers where sustained high subglacial water pressure is maintained.

4.4.2 Subglacial water pressure variations

The drainage of the boreholes indicates a connection with a subglacial hydrological system. The subglacial water pressure varied within 4.43–4.77 MPa, equivalent to 90–97% of the ice overburden pressure. Here, I discuss the temporal variations of the water pressure and their mechanisms.

Hydrological connections from the glacier surface to the base

The increase in the subglacial water pressure after the rainfall on 2 January strongly suggests the drainage of the rainwater into the glacier bed. The water pressure had been decreasing between 31 December and 2 January, but rapidly increased during 3–4 January at a rate of 4.4 kPa h⁻¹. Here, the peak of the subglacial water pressure lagged by 38 hours behind the maximum intensity of rain, which can be attributed to the time required for water routing from the surface to the bed. This contrasts with glaciers in Greenland, for example, surface melt drives diurnal speedup with lags of 4–6 hours (e.g. Andrews and others, 2014; Davis and others, 2014; Stevens and others, 2021), indicating more direct surface-to-bed hydraulic connection resulted in quicker response of the glacier dynamics to surface meltwater production. The subglacial water pressure appears not affected by diurnal variations in the surface melt (Fig. 4.8d), suggesting the hydraulic connection is limited and the basal drainage system responded only to the large input of the water from the surface. Such basal drainage system is considered to be inefficient, and easily pressurized by impulsive water inputs and thus likely to be sensitive to short-lived water inputs as those observed during the rain event on 2 January. Such short-term variations of the subglacial water pressure have not been assumed in Antarctica, as it was thought to be largely close to a steady state with water input only from geothermal heating and ice-bed friction (e.g. Dow and others, 2022).

In Antarctica, the hydraulic connections from the surface to the base have been observed by borehole observations only in the Antarctic Peninsula region (Sugiyama and others, 2019) and speculated as a possible cause of seasonal speedup of tidewater glaciers in the region (Tuckett and others, 2019; Boxall and others, 2022; Wallis and others, 2023). This study confirmed such hydrological conditions for the first time in East Antarctica. Ice is exposed in the study area in Lanhovde Glacier, and many rivers and lakes develop during melt seasons (Langley and others, 2016). Supraglacial rivers draining into crevasses were observed during the field observations (Fig. 4.18), thus, such englacial drainage was assumed to be connected to the subglacial hydrological system. The englacial drainage through crevasses observed here has not been found in the previously conducted satellite observations over Lanhovde Glacier (Langley and others, 2016). Recent satellite observations indicated numerous lakes and streams formed on the ice sheet in East Antarctica (Kingslake and others, 2017; Arthur and others, 2022). My results suggest that surface-to-bed hydrological connections may have developed in the regions where such surface hydrological networks have been reported.



Figure 4.18: A photograph showing supraglacial stream flowing into a crevasse at LGX.

On 23 January, subglacial water pressure at BH02 rapidly increased to a level similar to the ice overburden pressure. The pressure at BH01 also increased from 29 January and exceeded the ice overburden on 4 February. The increase in the water pressure at the end of the melt season and the subsequent fluctuations near the overburden pressure can be attributed to the separation of the boreholes from the main drainage system (e.g. Fudge and others, 2008; Wright and others, 2016).

The obtained results revealed the hydrological connection between the glacier surface and the bed in East Antarctica. The finding indicates the subglacial conditions and the glacier dynamics in Antarctica are more sensitive to climate than previously thought. Surface melting in Antarctica is projected to increase under atmospheric warming over the next several decades (Trusel and others, 2015), thus, considerable impacts on the subglacial conditions should be assumed.

Tidal influences

The subglacial water pressure showed clear variations in phase with tides from 14 to 25 January (Fig. 4.8e), indicating the connection between the basal hydrological system and the ocean. The pressure variations show the influences of hydrostatic fluctuations by tides on the subglacial water pressure. Such tidal influences on the subglacial water pressure were previously hypothesized in numerical studies as a possible cause of tide-modulated glacier flow (Rosier and others, 2015) since the water pressure variations influence glacier basal lubrication. The results confirmed the tidal influences on the subglacial water pressure for the first time by the direct observations using boreholes. However, the tidal variations did not exhibit clear influences on glacier basal lubrication. If the tidal variations were affecting the lubrication, the glacier was expected to accelerate during high tides, however, GNSS observations showed flow acceleration during falling tides. This result suggests that the tidal influences on the glacier flow are likely due to changes in hydrostatic pressure at the ice-ocean boundary (see Section 4.4.3), indicating that the 1–2 m amplitude of the water pressure variations has

limited effects on glacier basal lubrication.

4.4.3 Ice speed variations

Short-term acceleration

GNSS observations revealed two speedup events during 19–27 December 2021 (the first speedup) and 2–9 January 2022 (the second speedup), reaching 17% and 21% increase in the ice speed, respectively. For the second speedup event, the borehole observations revealed that the speedup corresponded with the rise in the subglacial water pressure after the rain event (Fig. 4.19). The surface speed and the water pressure were well correlated ($r = 0.89$, $p < 0.005$), suggesting a significant control of the water pressure on the glacier basal lubrication (Fig. 4.20a). The speedup was associated with the upward motion of the glacier surface. The maximum horizontal ice speed coincided with the pressure peak, whereas slightly preceded the peak of the vertical displacement (Figs 4.19a and b). A possible explanation for the ice speedup is cavity formation at the ice-bed interface and enhanced basal sliding. The maximum ice speed during the uplift before the peak of the vertical coordinate is consistent with previous studies conducted in hard-bedded alpine glaciers (Iken and others, 1983). The rapid decrease in the pressure by 95 kPa (9.7 m of water level) within two hours after reaching its peak implies pressure reduction due to cavity expansion by enhanced basal sliding. Another possible process for the surface uplift is the vertical straining of ice and resulting ice thickening (Sugiyama and Gudmundsson, 2003; Young and others, 2019). However, it is not plausible to assume compressive flow at LGX during the speedup events owing to the similar magnitude of speed increases at LG4 (Fig. 4.8a). Thus, I interpret the vertical surface uplift as the result of basal separation, as it is assumed to be playing a key role in short-term acceleration in alpine glaciers (e.g. Iken and Bindenschadler, 1986; Bartholomäus and others, 2008) as well as outlet glaciers in Greenland Ice Sheet (e.g. Sugiyama and others, 2015; Doyle and others, 2018).

The speedup event during 19–27 December also corresponded with high water production on the glacier surface, thus its mechanisms can be attributed to basal lubrication due to elevated subglacial water pressure. During 20–24 January, intensive melting of the ice surface was estimated from the surface energy balance (Fig. 4.8c), dominated by elevated shortwave radiation (Fig. 4.16c). The mean melt rate during the 5-day period (16.8 mm d^{-1}) was larger than twice the average over the whole observation period (7.1 mm d^{-1}). Thus, similar to the acceleration during 2–9 January, the surface meltwater was assumed to drain into the subglacial hydrological system. Such a large input of water to the glacier base in the early melt season can cause a significant pressure increase and basal lubrication, since the drainage system has not been developed enough to accommodate the injected water. The greater magnitude of the acceleration and the surface uplift during the first speedup event than that observed during the second event can be assumed to this process.

The borehole data showed that the subglacial water pressure maintained consistently high

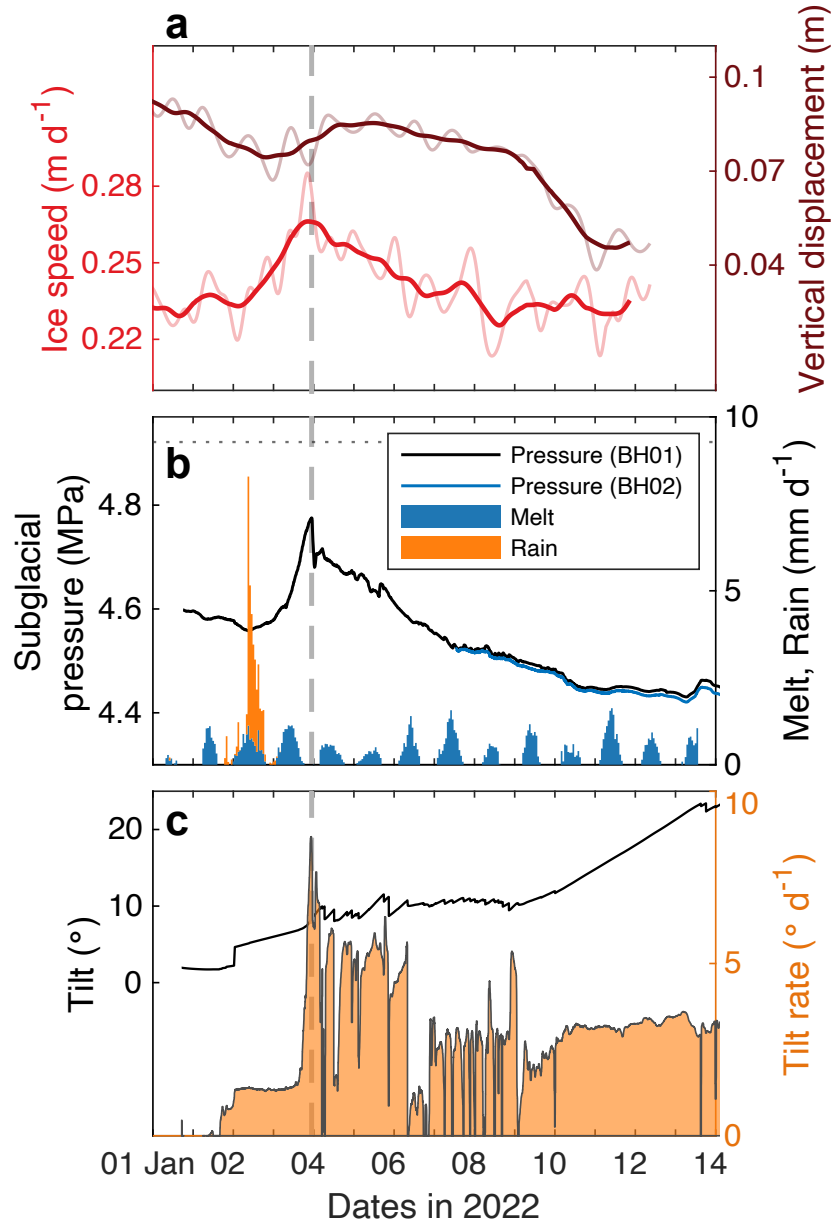


Figure 4.19: (a) Glacier surface speed and vertical displacement observed at Site 2 (light red and light dark red, respectively) and its 1-day running mean (red and dark red). (b) Subglacial water pressure measured at BH01 (black line) and BH02 (blue line). The dotted line indicates the ice overburden pressure at Site 2. Also shown are hourly glacier surface melt computed by the energy balance model (blue bar) and rain observed on the glacier (orange). (d) Tilt of the ploughmeter (solid line) and tilt rate (orange). The vertical dotted line indicates the timing of the maximum subglacial water pressure.

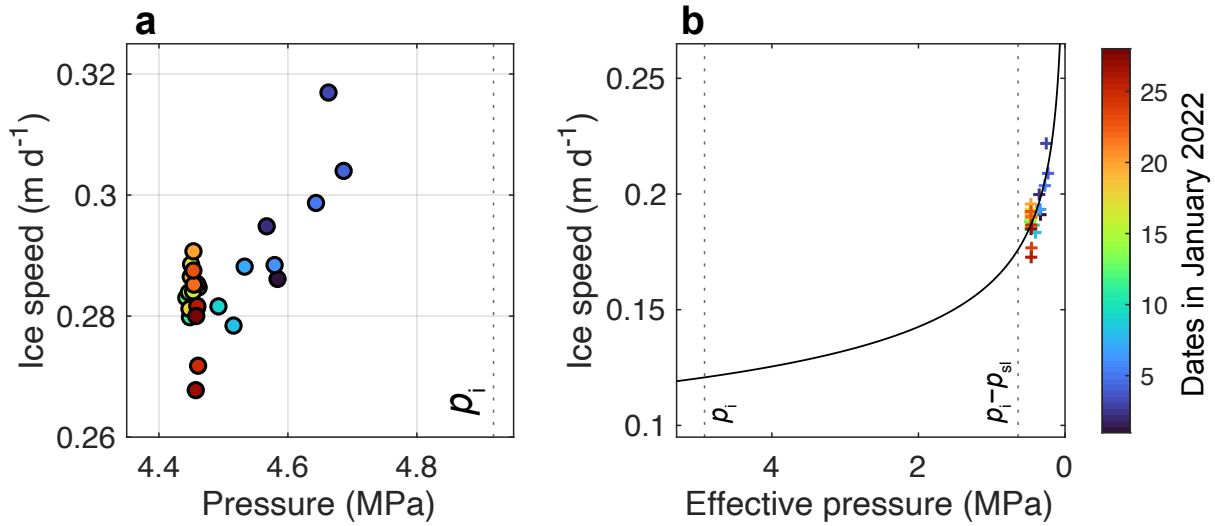


Figure 4.20: (a) Scatter plot showing the relationship between daily mean subglacial water pressure and surface ice speed. The vertical dotted line shows ice overburden pressure (p_i) (b) Scatter plot showing the relationship between daily mean effective pressure ($N = p_i - p_w$) and estimated basal ice speed. The solid line shows the fitting curve of the data, representing eq. 4.24. The vertical dotted lines show effective pressures for water levels equal to zero ($N = p_i$) and to the sea level ($N = p_i - p_{sl}$). The colors show data acquisition dates in January 2022.

pressure, and its rise caused substantial acceleration. To further evaluate the control of subglacial water pressure on glacier sliding speed, the observed data was compared with equation (1.1), which relates the sliding speed to the inverse power of effective pressure. The glacier sliding speed was estimated by subtracting the simulated ice deformation rate from the surface ice speed. The theoretical vertical gradient of ice deformation rate (du/dz) was given by Glen's flow law assuming a gravity-driven parallel-sided slab of ice at inclination angle α (e.g. Cuffey and Paterson, 2010) as:

$$\frac{du}{dz} = 2A(\rho gh \sin \alpha)^n, \quad (4.23)$$

with ice density ρ , gravitational acceleration g , and depth below the surface h . The power-law exponent $n = 3$ is conventionally assumed. $\alpha = 1.5^\circ$ is surface slope. A is a temperature-dependent rate factor, which was computed from the previously measured profile of englacial ice temperature at LG4 (Sugiyama and others, 2012). The contribution of the ice deformation rate to the surface ice speed was given by vertical integration of du/dz for the entire ice column. As a result of the calculation, the ice deformation rate was estimated to be 35 m a^{-1} , accounting for 40% of the mean surface ice speed at LGX (88 m a^{-1}). The dependence of the glacier sliding speed (u_b in m d^{-1}) on the effective pressure (N in MPa) was evaluated by fitting the data to equation (1.1). The least-square fitting yields

$$u_b = 0.16N^{-0.18}. \quad (4.24)$$

The obtained relationship indicates the glacier sliding speed is highly dependent on the effective pressure when the water level increases above sea level, thus small fluctuations in the

water pressure result in significant glacier acceleration.

Tidal influences

GNSS observations showed that the ice flow of Langhovde Glacier was influenced by ocean tide (Fig. 4.9). The tidal influences on the ice-shelf flow have been reported by the previous studies on Langhovde Glacier (Sugiyama and others, 2014; Minowa and others, 2019). Accordingly, the ice speed took maximum/minimum during falling/rising tide, and the phase and amplitude of the vertical motion corresponded with the tidal variations over the ice shelf. The greatest tidal effect has been observed near the ice-shelf front, and decays significantly in the upstream region at LG4 located close to the grounding line (Minowa and others, 2019). My observations showed that the along-flow ice motion was influenced by the tide at LGX as well as LG4, showing maximum/minimum detrended along-flow position during low/high tides (Fig. 4.9). The magnitude of the tidal influence on the along-flow displacement was similar between that observed at LG4 and LGX, despite LGX being located on the grounded ice 1 km upstream from LG4. These tidal influences were attributed to an elastic response of ice resulting from ocean-tide-induced hydrostatic stress fluctuations (Minowa and others, 2019), thus such elastic influence was assumed to propagate to LGX with only small decay in the magnitude.

The tidal influences on glacier flow have been reported on ice streams in West Antarctica (Anandakrishnan and others, 2003; Gudmundsson, 2006; Murray and others, 2007), which propagate several tens of kilometers inland from the grounding line. Such significant influences were attributed to the fact that these ice streams were located in extensive flat areas, with a limited degree of basal and lateral drag along the glacier boundaries. Langhovde Glacier has a relatively steep and narrow ice surface, indicating significant contributions of basal and lateral drags to the force balance. This observation suggests that the interior of Langhovde Glacier is less sensitive to ocean tides compared to such larger and flatter ice streams in West Antarctica.

4.4.4 Interpretation of the ploughmeter data

State of the ploughmeter

The ploughmeter tilted from 1.9° to $\sim 30^\circ$ during the observation period, indicating the device was moved along with the glacier basal motion. The most intriguing observation was the sawtooth variations during 4–9 January, corresponding with the short-term glacier acceleration (Figs 4.8d and f). The sawtooth variations were characterized by a gradual increase and a sharp drop in the tilt. A possible explanation is a stick-slip motion of the ploughmeter as a result of the movement pushed by the glacier basal motion. During the tip of the ploughmeter was stuck on the ice-bed interface, the ploughmeter would be pushed by the basal ice, causing

the tilt to gradually increase. The process results in the build up of strain in the ploughmeter, and when the limit is reached, the tip slips over the bedrock associated with the sharp drop of the tilt. The cyclicity of the sawtooth variations during 4–9 January suggests the ploughmeter was continuously moved along with the glacier basal motion by repeating the stick-slip. The sustained tilting between 9 and 13 January is likely to have occurred as a result of the ploughmeter tip becoming fixed in the bedrock. Possible factors of the stuck of the ploughmeter tip could be blocking by a bedrock bump or a large piece of gravel. As a result of the sustained strain buildup due to the basal motion, a part of the casing of the instrument was likely broken on 14 January and subsequently recovered from the stuck and dragged over the bedrock again.

The spectrogram of the horizontal component of the acceleration data also represents the state of the device. There are two clear changes in the spectral characteristics on 3 and 14 January. The differences in the spectrogram likely represent the changes in the state of the ploughmeter. From 1 to 3 January, the device was assumed to have loose contact with ice/sediments just after the installation, thus concentrated power on the horizontal component at 6 and 25 Hz was thought to be resulted from resonance. The following period during 3–14 January was characterized by the disappearance of the resonance and intermittent impulsive signals spanning the full frequencies associated with the large variability of the device tilt. Given that the ploughmeter was actively moving during this period due to the stick-slip of the instrument, the loss of resonance suggests that the tight coupling between the instrument and the ice/bed has occurred. After 14 January, the resonance recorded again at 6 and 15 Hz probably because the ploughmeter became loose again in the borehole as a result of the failure on the day.

Glacier basal motion

Assuming that the tilting of the ploughmeter occurs as a result of the instrument being pushed by the glacier basal sliding, the tilt rate represents the glacier sliding rate. To investigate the relationship between the basal ice dynamics and the movement of the ploughmeter, the tilt rate was calculated and compared with the surface ice speed. The tilt rate of the ploughmeter rapidly increased on 3 January and reached its maximum of $8.7^{\circ} \text{ d}^{-1}$ (Fig. 4.19c). The fastest tilting of the ploughmeter corresponded with the peak of the subglacial water pressure and the surface ice speed at LGX (Figs 4.19b and c). After reaching its maximum, relatively high tilt rates were sustained at $\sim 5\text{--}6^{\circ} \text{ d}^{-1}$ by 6 January, when the glacier accelerated. Thereafter, the tilt rate decreased to $\sim 3^{\circ} \text{ d}^{-1}$ and was kept stable by 14 January. The relatively high tilt rate observed during the period of glacier acceleration, along with its peak coinciding with the maximum surface ice speed, suggests that the elevated tilting speed of the ploughmeter is indicative of the active basal sliding during the glacier acceleration. The findings suggest that continuous inclination measurements taken at the ice-bed interface offer in-situ data on glacier basal sliding rates, that have conventionally been challenging to quantify.

In addition to the tilt rate, the number of vibration events on the horizontal axis also increased when the glacier accelerated (Fig. 4.8d), suggesting active glacier basal motion. Dur-

ing 4–5 January, the occurrence of the vibration events tends to increase after the decrease in the tilt angle, i.e., slip of the ploughmeter tip. The frequent event occurrence (Fig. 4.11c) and short interevent time (~ 1 s) (Fig. 4.11d) immediately after the device slip indicates that the device had relatively loose contact with bedrock and surrounded glacier ice. The first vibration event associated with the device slip with an amplitude reaching ~ 100 mG indicate releases of the strain in the ploughmeter that built up during the tilting of the device. As the tilt of the ploughmeter increased, the number of the vibration events decreased and their interevent time became longer (~ 100 s). The changes in the peak amplitudes and intervals of the vibration events likely represent the accumulation process of strain within the ploughmeter. As the ploughmeter gradually intensifies its contact with the ice and bedrock due to the glacier basal motion, the device becomes less mobile, resulting in a reduction in the vibration events and an increase in the inter-event time. The process of the strain build-up can be also seen in their spectrogram (Fig. 4.11b). After the device slip, dominant frequencies of the concentrated power slightly increased from 15 to 22 dB during the tilt increasing. This observation implies resonant frequency gradually increased as the strain accumulated on the ploughmeter due to increased tension acting on the device. The results indicate that the ploughmeter is actively moving along with the glacier basal ice, and suggest the potential utility of an accelerometer akin to inclinometers, seismometers, and strain gauges, all of which are useful for inferring glacier basal dynamics.

Subglacial water flow

Subglacial tremor of the vertical acceleration showed an increase after the rainfall on the glacier (Fig. 4.15), which corresponded with the increase in the modelled subglacial discharge. To better understand the relationship between the tremor amplitude and the subglacial discharge, a band-pass filter was applied to the time series to extract the variations with periods of more than three days. The filtered tremor amplitude and modelled discharge exhibited similar variations and correlated well with $r = 0.91$ (Fig. 4.21d). The seismic power within the frequency band of 3–10 Hz has been previously observed as an indicative of turbulent water flow (Bartholomaeus and others, 2015; Lindner and others, 2020; Nanni and others, 2020), thus their correspondence suggests that the subglacial water flow as a source of the subglacial tremor. It is likely that the increased subglacial drainage due to the rainwater drained into the glacier base caused enhanced turbulent water flow along the subglacial conduit, which resulted in the increase in the power of the tremor observed by the subglacial accelerometer.

Another possible factor of the subglacial tremor would be continuous seismic radiation by glacier basal sliding. A previous study on seismic tremors made by ocean bottom seismometer installed in front of an outlet glacier in Greenland showed a correspondence between the glacier flow speeds and tremor variations within the similar band of 3.5–14 Hz (Podolskiy and others, 2021). However, the ice speed measurements in this study showed the maximum ice speed on 3 January and subsequent decreasing by 10 January, which does not correspond with

the increase in the signal floor from 3 to 6 January. The increase in the seismic tremor amplitude was more correlated with the increase in the subglacial discharge resulting from elevated subglacial water pressure, rather than being directly synchronized with the ice speedup due to the elevated water pressure. Thus, I assume the seismic tremor observed by the accelerometer was caused by turbulent water flow along the subglacial conduits.

The diurnal variations of the seismic tremor amplitude correlated with the air temperature on the glacier surface (Fig. 4.14). Given that the increase in the seismic tremor amplitude during 3–14 January was induced by variations in the subglacial water flow, the diurnal variations also suggest the drainage of surface water into the glacier bed. To elucidate the underlying mechanisms of the diurnal variations in the subglacial tremor amplitude, a band-pass filter was applied to extract the tremor variations with 5–36-hour cycles and compared with the summation of computed glacier surface melt and observed rainfall on the glacier (Fig. 4.21c). Similar diurnal variations were observed in the filtered data during 8–15 January and 25 January–6 February, when the seismic tremor amplitude was well correlated with the air temperature (Fig. 4.14). The correspondence of the data was also observed during the blizzard observed during 17–19 January, with the disappearance of the diurnal signals. This relationship suggests that the daily fluctuations in subglacial tremor amplitude were influenced by the variations in the drainage of surface water into englacial or subglacial channels. Poor correlation during 1–7 January may be due to less efficient englacial drainage system, which has also been suggested from the fact that the maximum subglacial water pressure occurred after 38 hours of intensive rain.

Assuming the diurnal tremor variations were caused by the water flow from the glacier surface to the basal hydrological system, the lag between the surface water generation and the tremor variations represents the time required for routing through the englacial channel. Until 7 January, the production of the surface water preceded by more than 4 hours, but decreased ~0 hours by 14 January. Thereafter, the lag increased to ~6 hours on 19 January and decreased again to nearly zero from 26 to 29 January. The reduction in the lag observed during 1–15 January likely represents an increase in the drainage efficiency of englacial channels owing to the large input of the surface water on 2–4 January and subsequent development of the channels. Conversely, the lag notably increased on 19 January following a blizzard over the glacier that resulted in the complete freezing of surface water. The relatively large lag on 19 January is attributed to the narrowing of the englacial channels due to the cessation of meltwater supply from the glacier surface. The correlation in the phase from 28 January after the relatively large (~4 hours) lags between 19 and 26 January was probably due to the re-establishment of the efficient surface-to-bed hydraulic connections. These observations suggest that the diurnal tremor variations captured the process of the englacial channel development, which conveys water from the surface to the basal hydrological system on a daily basis.

These interpretations indicate that the drainage of water from the surface to the glacier base facilitated the development of subglacial channels. Variations in tremors induced by such

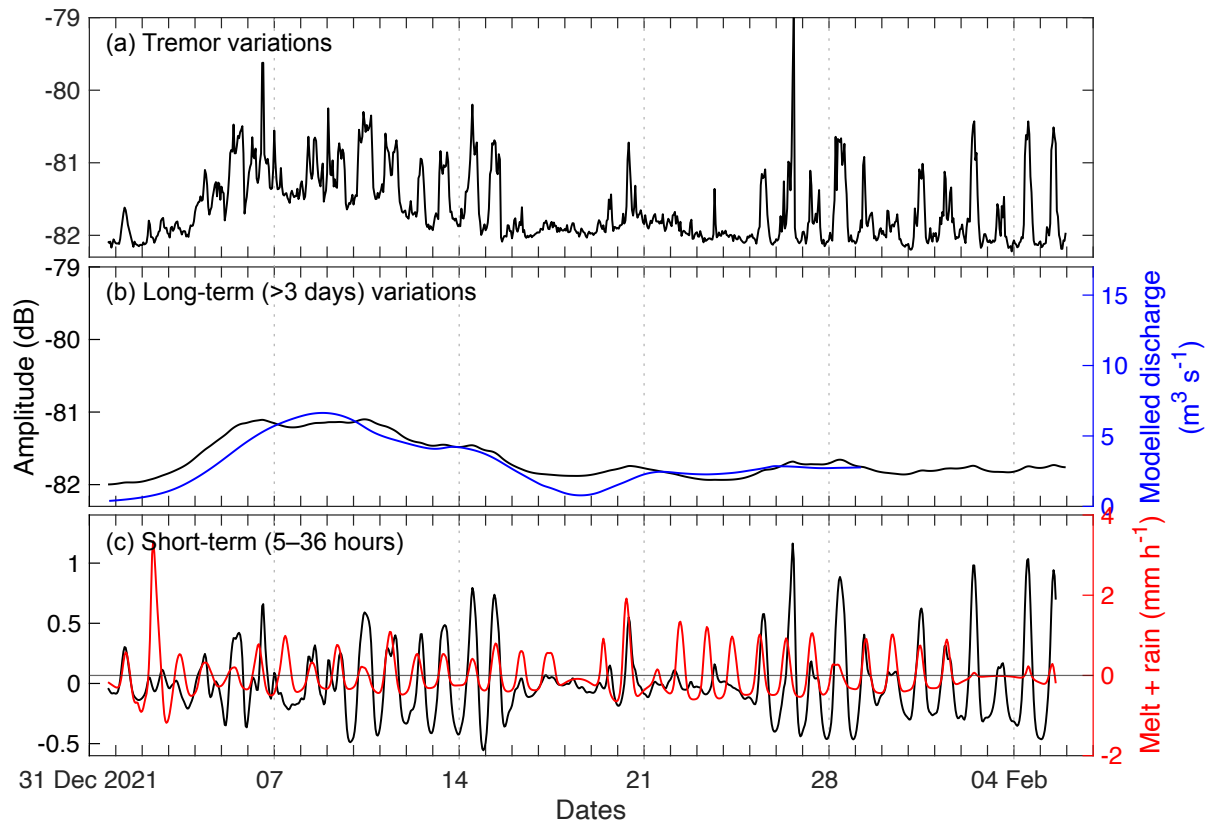


Figure 4.21: (a) Subglacial tremor variations observed by the accelerometer. (b) Tremor variations (black) and modelled subglacial discharge (blue) with a period of more than 3 days. (c) Tremor variations (black) and summation of estimated glacier surface melt and observed rain on the glacier (red) with a period of 5–36 hours.

subglacial water flow have been previously observed using seismometers installed on the surface of alpine glaciers (Bartholomaus and others, 2015; Lindner and others, 2020; Nanni and others, 2020), but this study represents the first direct observation of seismic tremor conducted at the glacier bed. Furthermore, this is the first documented observation inferring the development process of englacial and subglacial drainage systems in the Antarctic ice sheet. The results of this study showed the significant short-term fluctuations in the subglacial hydrological environment of the Antarctic ice sheet, influenced by the inflow of surface water.

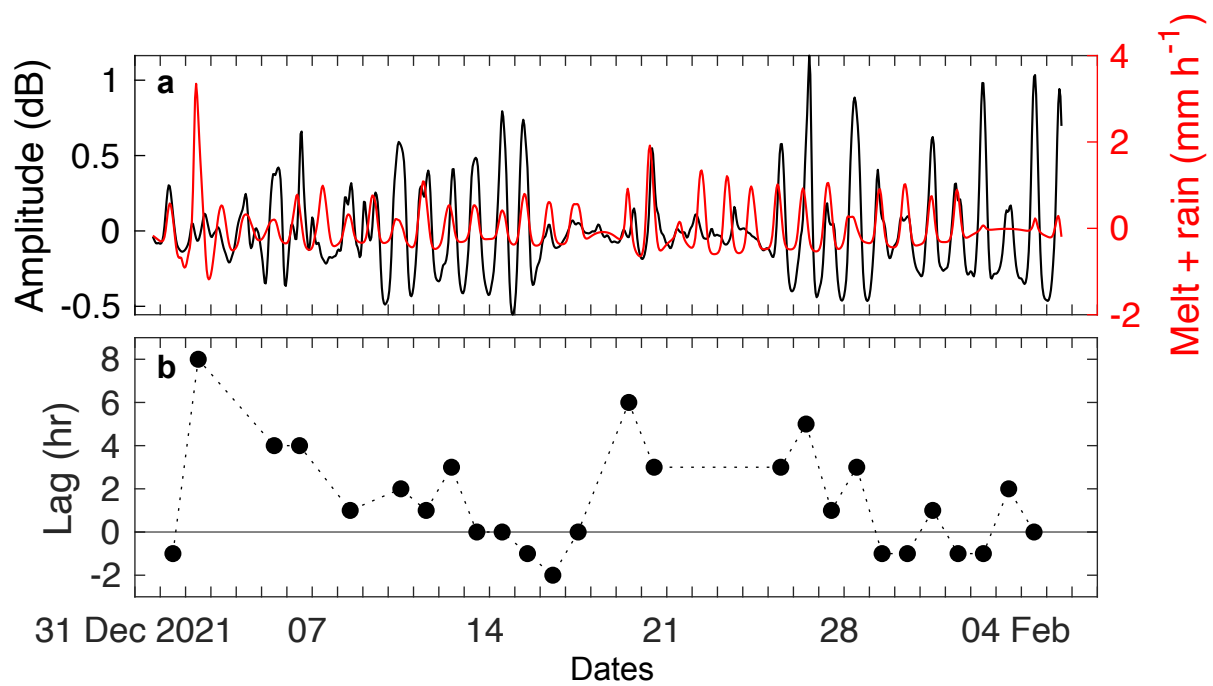


Figure 4.22: (a) Tremor variations (black) and summation of estimated glacier surface melt and observed rain on the glacier (red) with a period of 5–36 hours and (b) their lag computed from cross-correlation.

4.5 Summary

To investigate short-term flow changes on Langhovde Glacier feeding into Lützow-Holm Bay, East Antarctica, surface and subglacial measurements were conducted. In order to elucidate the relationship between glacier basal motion and subglacial water pressure variations, a device equipped with a three-axis accelerometer and a water pressure sensor was developed. The device was installed at the glacier bottom 1.5 km upstream from the grounding line using a 550-m long borehole drilled by hot water. The acquired data was analyzed together with the velocity time series observed on the glacier surface.

As a result of the observations, flow acceleration of up to 20% was measured when the subglacial water pressure rapidly increased between 2 and 4 January. During the same period, impulsive and strong horizontal vibration events were observed at the subglacial accelerometer with twice greater occurrence than the background value, suggesting enhanced basal sliding during the speedup. Subglacial water pressure increased a day after the rainfall of 30 mm d^{-1} , which strongly suggests that the drainage of the surface water pressurized the basal hydrological system enhanced basal sliding, and caused short-term glacier flow acceleration. These results are the first observations showing surface-to-bed hydrological connections in Antarctica. The drainage of the surface water caused pressurization of the drainage system and speedup, which have traditionally been considered to have limited or no impact on the subglacial conditions in Antarctica.

Seismic tremor observed at the accelerometer installed at the glacier base showed increased seismic amplitude after the elevated subglacial water pressure due to the input of the rainwater from the surface to the base. A numerical model of subglacial drainage showed increase in the water discharge along the subglacial channel associated with the elevated water pressure. The modelled subglacial discharge was closely correlated with the increased seismic tremor amplitude. The correspondence between the variations in modelled discharge and seismic tremor amplitude suggests subglacial turbulent water flow as a source of the seismic tremor. Furthermore, the tremor showed diurnal fluctuations corresponding with the simulated surface melt rate on the glacier surface. Thus, the daily fluctuations in the subglacial tremor were likely influenced by the variations in the drainage of surface water into englacial or subglacial channels. These findings indicate the significant short-term fluctuations in the subglacial hydrological environment of the Antarctic ice sheet, influenced by the inflow of surface water.

Implications and future prospects

In this thesis, I studied the variations of the outlet glaciers feeding into Lützow-Holm Bay, East Antarctica. The aim was to investigate and elucidate the long-term flow variations occurring over several decades as well as the short-term flow changes taking place over a period of a few weeks. In this section, I summarize the key findings and implications drawn from the observation results, and propose future research directions aimed at enhancing the understanding in this field.

5.1 Influence of sea ice on the glacier flow changes

The satellite observations demonstrated that land-fast sea ice exerts significant influences on the terminus position, ice flow, and ice thickness of the outlet glaciers in Antarctica. Limited studies have shown the impact of sea ice on glacier flow in Antarctica. For example, studies by Greene and others (2018); Liang and others (2019) have shown that seasonal flow variations of ice shelves in East Antarctica are controlled by the back stress exerted by sea ice in front of the ice shelves. However, these studies mainly focused on ice shelves, and their effects on the grounded ice have not been extensively studied. For grounded ice, Rott and others (2018) demonstrated a deceleration in glacier flow on the Antarctic Peninsula when land-fast sea ice persistently covered the ocean near the terminus, indicating a long-term influence of sea ice on the grounded ice flow. To my knowledge, the glacier flow changes revealed in this study represent the second report in Antarctica where grounded ice flow is influenced by sea ice variations. Moreover, this is the first observation highlighting ice thickness changes associated with dynamic changes in the outlet glaciers due to the stress perturbations induced by sea ice variations. These results demonstrate that small stress changes at the glacier terminus due to sea ice fluctuations can significantly impact the flow and thickness of outlet glaciers.

Land-fast sea ice is highly sensitive to climate variability. Previous studies in Lützow-

Holm Bay have indicated that various climate factors induce the breakup of the land-fast sea ice, i.e. air temperature, southerly winds, SST, SIC in the pack-ice zone, and snowfall on the sea ice (Enomoto and others, 2002; Ushio, 2006). Periodic climate variations of El Niño have also been identified as contributors to breakup (Aoki, 2017). Continuous monitoring of the land-fast sea ice revealed its periodic behavior taking place with a 10-20-year cycle of variation (Ushio, 2006; Aoki, 2017). Therefore, it is suggested that glacier variations in this region also exhibit periodic fluctuations influenced by the periodic land-fast sea-ice variations. Additionally, recent observations have revealed that sea ice can exhibit sudden and drastic changes. For example, a record-breaking reduction in sea ice extent occurred in the autumn of 2023, reaching the largest negative anomalies in the 44-year observational record (Fig. 5.1) (Purich and Doddridge, 2023). Such observations highlight the susceptibility of sea ice to climate conditions, thus substantial impact on glacier variations should be expected under the changing climate.

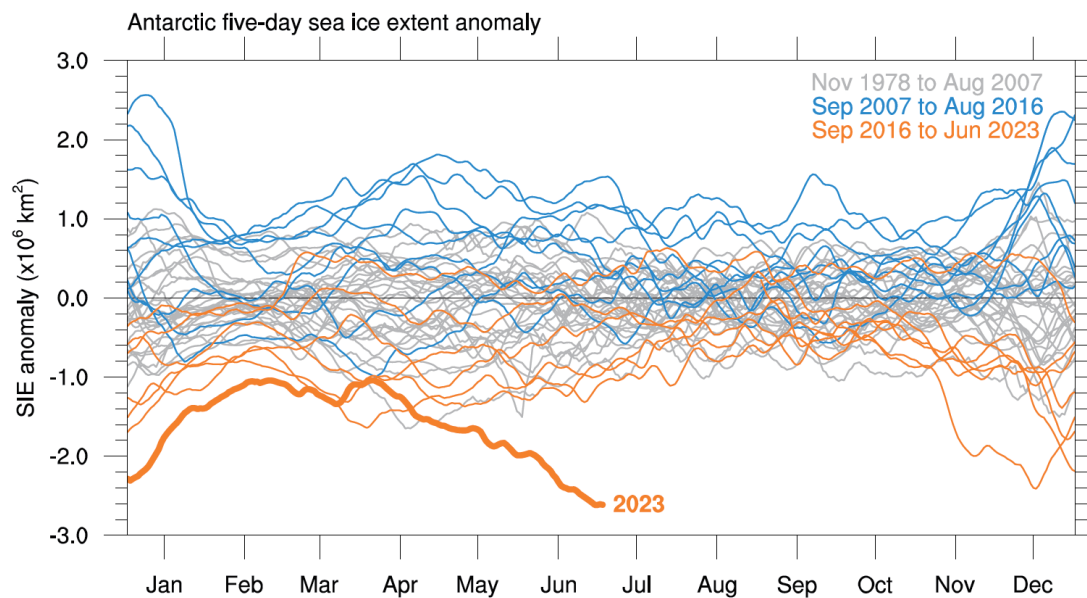


Figure 5.1: Antarctic five-day sea ice extent anomalies in millions of square kilometers for each year from the National Snow and Ice Data Center. Sea ice extent anomalies are calculated relative to the 1979–2022 climatology. Anomalies are colored by periods of November 1978 to August 2007 (grey), September 2007 to August 2016 (blue), and September 2016 to June 2023. Figure from Purich and Doddridge (2023).

5.2 Influence of subglacial conditions on the glacier flow changes

In the field observations, short-term flow acceleration over a week or shorter was found in Langhovde Glacier, one of the outlet glaciers feeding into Lützow-Holm Bay. The acceleration was attributed to elevated subglacial water pressure and subsequent basal lubrication caused by the drainage of surface water into the glacier bed. These findings imply the existence of the

hydraulic connectivity between the glacier surface and the base, highlighting the significant influence of surface water on the glacier basal sliding. While the acceleration of glacier flow induced by surface water has been widely recognized in mountain glaciers and the Greenland Ice Sheet, this is the first documented observation in Antarctica. This observation challenges the conventional assumption that the subglacial hydrological environment of the Antarctic ice sheet is characterized by steady-state conditions, primarily influenced by basal meltwater generated from geothermal and frictional heat (e.g. Dow, 2023). The presence of the surface-to-bed hydraulic connection in Antarctica suggests the coupling between the glacier basal motion and the atmospheric conditions. In the Antarctic ice sheet, a nonlinear increase of surface melt has been projected along with the atmospheric warming (Trusel and others, 2015), thus the impact of the atmospheric influences on the glacier dynamics may become more pronounced in the future. In this study, the flow acceleration was observed only after intensive surface melt or rainfall. The future warming in Antarctica has the potential to drive more pronounced seasonal flow variability as currently being observed in the Greenland Ice Sheet, thus substantial impact on the glacier dynamics should be expected.

The observed flow acceleration in this study occurred on a timescale of a week, shorter than the temporal resolution of satellite data analysis. This suggests that surface water-induced flow accelerations are more widespread than previously recognized, extending beyond Langhovde Glacier. In East Antarctica, active surface hydrological environments have been recognized and quantified in recent years (Kingslake and others, 2017; Bell and others, 2018; Arthur and others, 2022), and attention is drawn to their potential impact on ice sheet dynamics (Fig. 5.2). Hydraulic connections from the surface to the bed have not been recognized yet in these regions, but precise measurements by the GNSS technique have the potential to reveal their presence in East Antarctica.

Honor, Telen, Skallen, and Shirase glaciers are located along the Soya Coast under similar climatic conditions as Langhovde Glacier. Although short-term (<1 week) speed variations were not detected by the satellite observations, seasonal flow variations possibly exist on these glaciers as observed by the GNSS observations at Langhovde Glacier. As shown in Chapter 3, there has been no clear relationship between annual flow velocity and summer air temperatures (Fig. 3.3), therefore, long-term flow variations are considered to be primarily influenced by stress perturbations induced by sea ice variations for the last 20 years. However, if surface water-induced glacier acceleration exists on the glaciers, seasonal glacier flow variations may become a significant control on the glacier dynamics as a consequence of future atmospheric warming. In such a scenario, a potential change in the dominant factors governing the glacier flow should be expected from stress changes by sea ice to changes in subglacial conditions triggered by atmospheric variations.

In this study, I conducted the first direct observation of glacier basal sliding using accelerometers. The obtained data demonstrated the potential utility of accelerometers, akin to inclinometers, seismometers, and strain gauges, showing a potential to extract critical informa-

tion about subglacial flow dynamics that has been difficult to obtain. Specifically, variations in basal sliding speed were inferred from the tilt rate of the device, suggesting the value of using accelerometers for observing glacier basal motion.

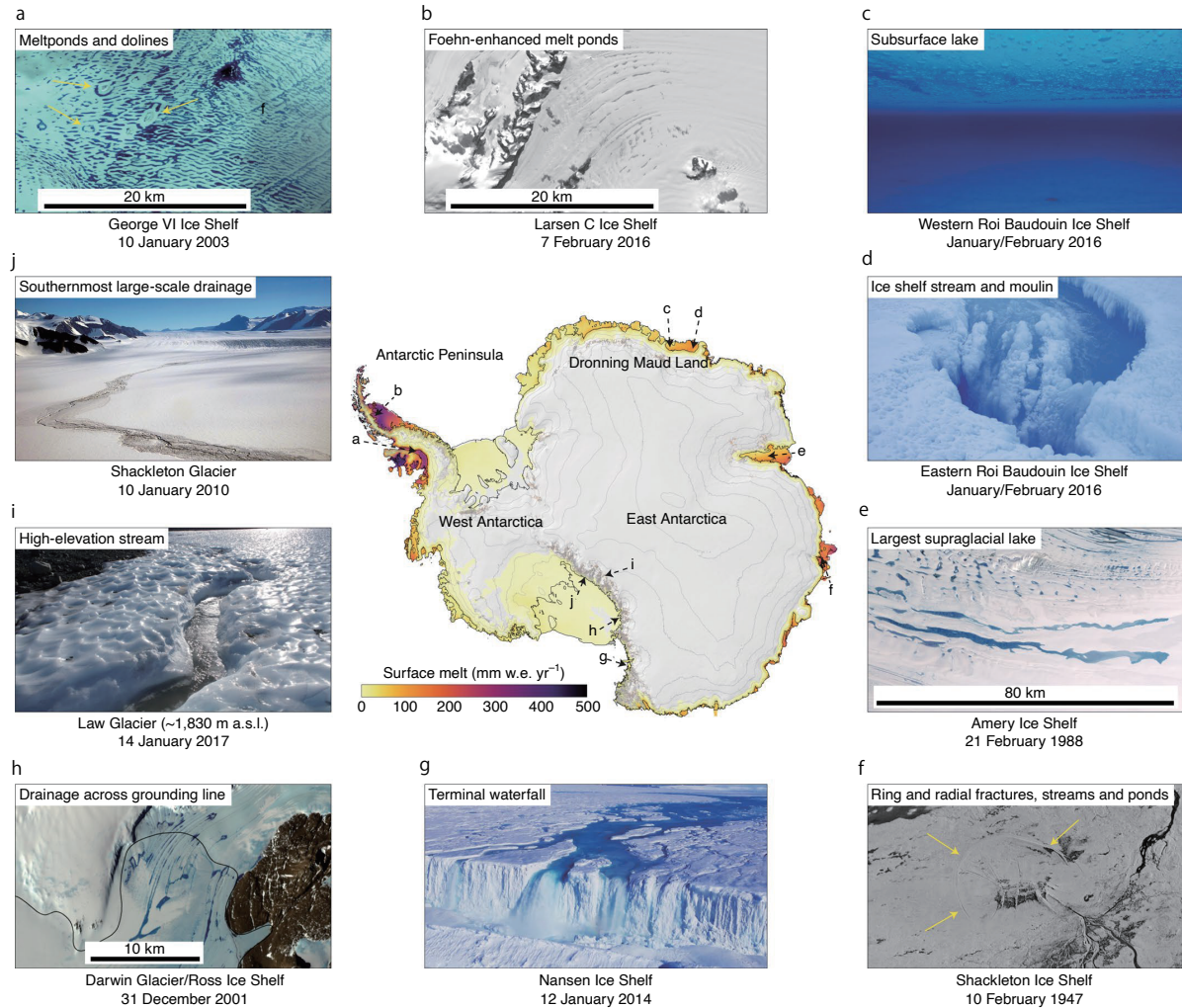


Figure 5.2: The central map shows 2000–2009 Antarctica surface melt from satellite observation (Trusel and others, 2013); the locations of the images in (a–j) are indicated. (a) Meltwater lakes and dolines (arrows), (b) Foehn wind-enhanced meltwater ponding. (c) Buried lake. (d) Moulin draining surface stream. (e) Elongate supraglacial lake. (f) Fractures around a drained lake. Scale unknown. (g) Persistent waterfall draining water. (h) Supraglacial streams transporting water across the grounding line of the Darwin Glacier onto the Ross Ice Shelf. (i) High-elevation (1,830 m) meltwater stream. (j) Meltwater stream crossing the grounding line. Figure from Bell and others (2018)

5.3 Future prospects of the study

This study has shown that the land-fast sea ice plays an essential role in the outlet glacier variations in Lützow-Holm Bay. Therefore, quantitative analysis is recommended for furthering our understanding. For example, the magnitude of back stress exerted by the sea ice at the glacier terminus remained unclear in this study. Numerical analysis for elucidating the stress balance

at the boundary of glacier termini and sea ice/ice mélange has been applied to several calving glaciers in the Greenland Ice Sheet (Thomas and others, 2004; Todd and Christoffersen, 2014). For accurate numerical analysis, glacier ice thickness is essential. However, the topography underneath the glaciers flowing into Lützow-Holm Bay remains largely unknown owing to the limited in-situ topographic survey. To observe the ice thickness, the Antarctic RINGS project conducted by the Scientific Committee on Antarctic Research (SCAR) will carry out airborne radar surveys along the coastal margin of the Antarctic ice sheet from the austral summer of 2023/24. Such international efforts are expected to provide important insights into the basal topography and subglacial conditions of the ice sheet. Further understanding of the effects of sea ice on glacier dynamics should be achieved by conducting numerical experiments on their flow variations, based on the observed bed topography.

The study results have shown that the significant controls of the subglacial water pressure on the glacier flow changes. The findings highlight the importance of understanding the mechanisms to develop the subglacial drainage system on the ice sheet. For this purpose, it is crucial to reveal the bed topography over the ice sheet, and elucidate the hydraulic head which determines water movement under the ice. Extensive radar surveys such as RINGS project led by SCAR are expected to provide an important dataset to achieve this goal. Additionally, confirming the presence of short-term flow variations in glaciers located along the Soya Coast, apart from the Langhovde Glacier, would be important to understand the mechanisms of the changes in the subglacial drainage system on a broader scale. To observe fine temporal-scale glacier flow variations, the use of radar satellites with short revisit times, such as Sentinel-1, and in-situ GNSS surveys are essential.

The relevance of the findings from this study should be confirmed for other regions in Antarctica. For example, Lützow-Holm Bay is characterized by extensive land-fast sea ice development, making it unique within Antarctica (e.g. Fraser and others, 2023). Moreover, the prevalence of bare ice and the active supraglacial hydrology in this region (Langley and others, 2016) suggest that the results obtained here may be specific to Lützow-Holm Bay. Nevertheless, considering previous research indicating the influence of land-fast sea ice on glaciers (Miles and others, 2017; Christie and others, 2022) and the presence of seasonal variations in glacier dynamics observed on the Antarctic Peninsula (Tuckett and others, 2019; Boxall and others, 2022; Wallis and others, 2023), there is substantial potential for similar phenomena occurring in other regions in Antarctica. Therefore, it is essential to conduct studies in areas beyond Lützow-Holm Bay, and verify the existence of regional characteristics or the robustness of the key findings in this study.

The ploughmeter developed in this study provided a variety of data for inferring the subglacial dynamics. To further enhance the utility of the ploughmeter, I propose the implementation of a magnetometer. This would enable analysis that relates the instrument's movement direction and the flow direction of the glacier, potentially providing quantitative insights into basal ice flow. Nevertheless, the results obtained in this study highlight the potential for de-

ployment on other glaciers. For example, the fast flow of the glaciers driving ice-sheet mass loss in Antarctica is known to result from the deformation of unlithified sediments beneath the glaciers. However, the deformation of these sediments and their contribution to ice flow remain poorly understood due to a lack of direct observational data. To investigate these unconstrained dynamics, the ploughmeter could be highly effective.

Conclusions

In this thesis, satellite data analysis and field observations were conducted to understand the mechanisms of flow changes of outlet glaciers in the Antarctic ice sheet. Satellite data analysis revealed the relationship between flow changes and sea ice variations over the past two decades in five outlet glaciers (Shirase, Skallen, Telen, Honnør and Langhovde glaciers) flowing into Lützow-Holm Bay, East Antarctica. Field observations revealed short-term flow changes of Langhovde Glacier. The mechanisms of the flow changes were investigated by direct observations at the glacier base using hot-water drilling.

The fronts of the studied glaciers advanced from 2006 to 2015 when the ocean was largely covered with sea ice. All the glaciers rapidly retreated by 0.4–6.0 km in the period 2016–18, which coincided with a major breakup of land-fast sea ice in Lützow-Holm Bay in 2016. The breakup was the most significant sea ice loss in the region since 1998. Shirase, Skallen, and Telen glaciers simultaneously decelerated in the period 2011–15 and accelerated in 2016–18. These flow speed changes were most pronounced on the ice shelves but were also observed up-stream of the grounding line. Near the grounding lines of Shirase, Skallen and Telen glaciers, the surface elevations increased during 2003/04–15/16 and decreased during 2015/16–19/20, corresponding with the flow deceleration and acceleration periods, respectively. I found that the retreat rate of the glaciers correlated with the SIC anomaly near the glacier fronts when the SIC anomaly dropped below 0, suggesting that the sea ice is an important factor controlling the terminus variations. The displacement rate of the sea ice before the breakup in 2016 was substantially smaller than the forward motion of the glacier terminus. During this period, cracks and fractures were observed on the land-fast sea ice in front of the glaciers, implying that a significantly large stress was exerted on the sea ice by the advancing glacier and this stress impeded the forward glacier motion. The glacier flow accelerated after the sea ice sped up and began to move with the glacier front, suggesting a reduction in the resistive stress resulted in the glacier acceleration after 2016. Glacier thickness was also affected by the sea-ice breakup and the associated ice speed changes. Strain rate analysis suggested that the stretching ice flow along the glacier was reduced/enhanced by the observed flow deceleration/acceleration. The

results suggest that the observed elevation changes were driven by the variations in the vertical strain rate near the grounding line, caused by changes in the glacier flow regime. These results indicate that relatively small changes in the stress balance due to sea ice breakup caused glacier acceleration and calving, which in turn caused changes in ice thickness.

As a result of the observations, flow acceleration of up to 20% was measured when the subglacial water pressure rapidly increased between 2 and 4 January. During the same period, impulsive horizontal vibration events were observed at the subglacial accelerometer with twice greater occurrence than the background value, suggesting enhanced basal sliding during the speedup. Subglacial water pressure increased a day after the rainfall of 30 mm d^{-1} , which strongly suggests that the drainage of the surface water pressurized the basal hydrological system enhanced basal sliding, and caused short-term glacier flow acceleration. These results are the first observations showing surface-to-bed hydrological connections in Antarctica. The drainage of the surface water caused pressurization of the drainage system and speedup, which have traditionally been considered to have limited or no impact on the subglacial conditions in Antarctica.

Seismic tremor observed at the accelerometer installed at the glacier base showed increased seismic amplitude after the elevated subglacial water pressure due to the input of the rainwater from the surface to the base. A numerical model of subglacial drainage showed an increase in the water discharge along the subglacial channel associated with the elevated water pressure. The modelled subglacial discharge was closely correlated with the increased seismic tremor amplitude. The correspondence between the variations in modelled discharge and seismic tremor amplitude suggests subglacial turbulent water flow as a source of the seismic tremor. Furthermore, the tremor showed diurnal fluctuations corresponding with the simulated surface melt rate on the glacier surface. Thus, the daily fluctuations in the subglacial tremor were likely influenced by the variations in the drainage of surface water into englacial or subglacial channels. These findings indicate the significant short-term fluctuations in the subglacial hydrological environment of the Antarctic ice sheet, influenced by the inflow of surface water.

The study revealed the variations of the five outlet glaciers flowing into Lützow-Holm Bay, East Antarctica, during 2000–2020 for the first time. Satellite observations revealed that the stress changes due to sea ice variations caused glacier flow changes. Field observations on Langhovde Glacier captured short-term summer flow acceleration for the first time in East Antarctica. Direct observations at the glacier bed revealed that the flow acceleration was caused by the enhanced sliding due to the subglacial water pressure elevated by the drainage of surface water to the glacier bed. The results of the satellite and field observations indicate that small changes in the glacier stress balance and basal conditions caused by sea ice retreat and surface-to-bed water input resulted in significant flow changes. In other words, outlet glaciers of the Antarctic ice sheet are highly sensitive to changes in the surrounding environment such as the atmosphere and the ocean, which results in glacier flow changes. These results contribute to a quantitative understanding of glacier variations in Lützow-Holm Bay and provide

new insights into the mechanisms of outlet glacier changes in the Antarctic ice sheet.

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