Ice Shelf Water-influenced Fast Ice and Sub-Ice Platelet Layer Thickness Distributions beside the Campbell Ice Tongue in Terra Nova Bay, Antarctica

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Abstract

Prior to the present study, no dedicated in situ measurement of the thickness distributions of fast ice and the sub-ice platelet layer, formed by supercooled Ice Shelf Water in the near-surface ocean, had been carried out in Terra Nova Bay. Previous studies have recognised the biological importance of the sub-ice platelet layer observed beneath fast ice beside the Campbell Ice Tongue. Furthermore, a recent airborne survey of fast ice in the western Ross Sea implied that smaller ice bodies (ice tongues and outlet glaciers) contribute to the formation of supercooled Ice Shelf Water. With the objective of inferring source regions and circulation of Ice Shelf Water, we measured fast ice and sub-ice platelet layer thickness distributions near the Campbell Ice Tongue in late spring of 2021, using drill hole surveys and high-resolution ground-based electromagnetic induction soundings. We observed thicker fast ice and sub-ice platelet layer near the ice tongue with very thick and narrow sub-ice platelet layer maxima identifying highly channelled outflow of supercooled Ice Shelf Water from beneath the ice tongue through ice mélange, subglacial formations, and grounded regions. We conclude that a significant volume of supercooled Ice Shelf Water is locally sourced from the Campbell Ice Tongue through basal melting and affirm that the icescape in north Terra Nova Bay results from a complex interplay of glacial morphology, bathymetry, polynya dynamics, and ocean circulation.
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Key Points:

- Basal melt of the Campbell Ice Tongue is a significant source of Ice Shelf Water in north Terra Nova Bay.
- In late spring, thick fast ice and sub-ice platelet layer identified Ice Shelf Water outflow from beneath the ice tongue.
- Fast ice persistence is enhanced by the Campbell Ice Tongue and diminished by the Terra Nova Bay Polynya.
Abstract

Prior to the present study, no dedicated in situ measurement of the thickness distributions of fast ice and the sub-ice platelet layer, formed by supercooled Ice Shelf Water in the near-surface ocean, had been carried out in Terra Nova Bay. Previous studies have recognised the biological importance of the sub-ice platelet layer observed beneath fast ice beside the Campbell Ice Tongue. Furthermore, a recent airborne survey of fast ice in the western Ross Sea implied that smaller ice bodies (ice tongues and outlet glaciers) contribute to the formation of supercooled Ice Shelf Water. With the objective of inferring source regions and circulation of Ice Shelf Water, we measured fast ice and sub-ice platelet layer thickness distributions near the Campbell Ice Tongue in late spring of 2021, using drill hole surveys and high-resolution ground-based electromagnetic induction soundings. We observed thicker fast ice and sub-ice platelet layer near the ice tongue with very thick and narrow sub-ice platelet layer maxima identifying highly channelled outflow of supercooled Ice Shelf Water from beneath the ice tongue through ice mélangé, subglacial formations, and grounded regions. We conclude that a significant volume of supercooled Ice Shelf Water is locally sourced from the Campbell Ice Tongue through basal melting and affirm that the icescape in north Terra Nova Bay results from a complex interplay of glacial morphology, bathymetry, polynya dynamics, and ocean circulation.

Plain Language Summary

Fresh meltwater from glacial ice on the Antarctic continent can influence coastal sea ice formation. If the meltwater originates from deep in the ocean, it can be supercooled and form characteristic platelet ice crystals, which contribute to sea ice formation and form thick layers beneath called sub-ice platelet layers (SIPL). This crystallographic signature of supercooled glacial meltwater provides important information on difficult to observe interactive processes occurring between the atmosphere, glacial ice, the ocean, and sea ice along the Antarctic coast. In late spring of 2021, we carried out detailed surveys of glacially-influenced coastal sea ice and SIPL beside the Campbell Ice Tongue in north Terra Nova Bay, Ross Sea, Antarctica with high-resolution geophysical surveying. Our objective was to use sea ice and SIPL distributions to infer where the glacial meltwater was coming from and where it circulates. Our surveys revealed thicker sea ice and SIPL near the ice tongue with thick bands of SIPL showing highly channelled outflow of supercooled glacial meltwater from beneath the Campbell Ice Tongue through glacial formations. We conclude that a significant volume of supercooled glacial meltwater in north Terra Nova Bay is locally sourced from the Campbell Ice Tongue through basal melting.

1 Introduction

On the Antarctic coastline, interactive processes between the atmosphere, ocean, land ice, and sea ice affect the mass balance of the grounded ice sheet, global ocean circulation, and sea level. Land-fast sea ice (henceforth referred to as fast ice) is an important interface between land ice and the open ocean (Giles et al., 2008; Massom et al., 2018), which affects ice sheet mass balance by stabilizing and buttressing ice tongues (Gomez-Fell et al., 2022, 2024; Massom et al., 2010) and ice shelves (Greene et al., 2018) from ocean swell (Christie et al., 2022; Massom et al., 2018). Land ice, in turn, provides a margin for fast ice to attach to, and affects fast ice formation through melt processes and freshwater input to the ocean (Fraser et al., 2023). Fast ice,
ice shelves, and ice tongues are key features for the formation of coastal polynyas (Fraser et al., 2019; Nihashi and Oshima, 2015).

Coastal latent heat polynyas are open water regions formed by strong offshore winds, where intense sea ice and High Salinity Shelf Water (HSSW) formation occurs (Nakata et al., 2015). HSSW is an important precursor to Antarctic Bottom Water (AABW) (Williams et al., 2010) and can circulate heat energy from the ocean surface to the grounding zones of ice shelves and outlet glaciers, where it causes basal melting (Silvano et al., 2018). A potentially supercooled water mass called Ice Shelf Water (ISW) is formed (Jacobs et al., 1985), which can freshen the water column and affect deep ocean convection (Williams et al., 2016), rendering ice shelves more susceptible to warm water mass intrusions (Silvano et al., 2018).

Buoyant ISW can rise from deep under ice shelves and outlet glaciers becoming in situ supercooled (Holland & Feltham, 2005). ISW promotes sea ice growth by stabilising the upper water column (Hellmer, 2004), or if in situ supercooled, through heat loss to the ocean (Gough et al., 2012). Frazil ice grows into larger platelet ice crystals (Smith et al., 2012), which can be deposited beneath sea ice and incorporated with congelation growth, augmenting sea ice thickness (Purdie et al., 2006). Once sea ice growth sufficiently diminishes, an unconsolidated mass of crystals called a sub-ice platelet layer (SIPL) can form beneath (Gough et al., 2012; Wongpan et al., 2015, 2021). Platelet ice plays an important role in Antarctic biology by providing a habitat for algae and micro-organisms (Arrigo et al., 1993) and a nursery for the keystone species Antarctic silverfish Pleuragramma antarcticum (Vacchi et al., 2012). The freeboard of ISW-influenced fast ice stands higher inherently because it is thicker, and from additional buoyancy of a SIPL, if present (Price et al., 2014). This effect on freeboard has been detected in McMurdo Sound with satellite altimetry (Brett et al., 2021; Price et al. 2013).

Studies of ISW-influenced fast ice and SIPL have linked late spring volumes of platelet ice, within fast ice and SIPL, to supercooled ISW outflow in Atka Bay, Weddell Sea (Hoppmann et al., 2015a, 2015b), and McMurdo Sound (Langhorne et al., 2015). The thickness distribution of SIPL reflects the circulation and degree of in situ supercooling of the ocean beneath (Hughes et al., 2014; Wongpan et al., 2021). Multiple years of ground-based (Brett et al., 2020) and airborne (Haas et al., 2021) electromagnetic induction (EM) assessments of ISW-influenced fast ice and SIPL thickness distributions in McMurdo Sound were correlated with supercooled ISW circulation (Hughes et al., 2014; Lewis and Perkin, 1985; Robinson et al., 2014) from the McMurdo-Ross Ice Shelf Cavity. This relationship allowed interannual and diurnal variability of SIPL thickness to be linked to polynya activity in the western Ross Sea (Brett et al., 2020) and the oscillation of the tides (Brett et al., 2024a).

Crystallographic signatures of in situ supercooled ISW in the form of marine, frazil, and platelet ice have been observed beneath ice shelves, ice tongues, and fast ice in Terra Nova Bay, in the northwest Ross Sea (refer to Figure 1a for locations). South of the Drygalski Ice Tongue, platelet ice was observed in a sea ice core (Jeffries and Weeks, 1993) and beneath fast ice (Stevens et al., 2017). The edge of Hells Gate Ice Shelf is comprised entirely of marine ice (Souchez et al., 1991), with thick frazil and platelet ice layers observed beneath the ice shelf (Tison et al., 1993, Souchez et al., 1991), and adjoining fast ice (Langhorne et al., 2023; Tison et al., 1998). An extensive airborne electromagnetic induction (AEM) survey of fast ice along the Victoria Land Coastline identified that smaller ice bodies including ice tongues and marine-
terminating outlet glaciers also contribute to the formation of in situ supercooled ISW and platelet ice in the western Ross Sea (Langhorne et al., 2023).

In north Terra Nova Bay, marine ice (Souchez et al., 1995) and basal freezing (Han and Lee, 2015) occur beneath the Campbell Ice Tongue (Figure 1). Beneath adjoining fast ice, a SIPL was detected in biological studies in late spring of 2002, 2005, and 2006 (Vacchi et al., 2004, 2012) and in AEM surveys in late spring of 2017 (Langhorne et al., 2023). The presence of marine ice and SIPL signal that in situ supercooled ISW is circulating in the region, which the AEM surveys indicated could be locally sourced from the ice tongue (Langhorne et al., 2023). However, no dedicated in situ investigation of the influence of ISW on fast ice, the SIPL, and the Campbell Ice Tongue in north Terra Nova Bay has been carried out. This information is important given that the ice tongue is thinning (Han and Lee, 2015) and decreasing in area (Han et al., 2022) with major implications expected for adjoining fast ice and local ecosystems.

In this study, we assessed spatial distributions of ISW-influenced fast ice and SIPL near the Campbell Ice Tongue, to infer the pattern of supercooled ISW circulation in north Terra Nova Bay, and the effects of local geomorphological features including bathymetry and grounded regions of the ice tongue. To investigate the processes influencing fast ice in north Terra Nova Bay, we used satellite imagery to observe fast ice, ice tongue, and polynya interactions during the winter of 2021. In late spring of 2021, we assessed thickness distributions of ISW-influenced fast ice and SIPL with drill hole and single-frequency electromagnetic induction (EM) sounding surveys. As observed in McMurdo Sound, the spatial distributions of thicker ISW-influenced fast ice and SIPL in Terra Nova Bay should reflect supercooled ISW circulation. We describe the study area in section 2, methods in section 3, and results in section 4. In section 5, we discuss our findings in a regional context considering polynya dynamics and ocean circulation in Terra Nova Bay.

2 Study Area

2.1 Terra Nova Bay

Terra Nova Bay is bound by the Drygalski Ice Tongue in the south, the Nansen Ice Shelf in the west, and a large embayment in the north (Figure 1a). The Terra Nova Bay Polynya (Figure 1a) is a dynamic force in the region, producing 10% of sea ice (Kurtz and Bromwich 1985), 33% of HSSW (Fusco et al., 2009; Rusciano et al., 2013), and significantly contributing to AABW formation in the Ross Sea (Rusciano et al., 2013). HSSW formed within the polynya (Thompson et al., 2020) is thought to drive basal melting at depth in the grounding zones of the Nansen Ice Shelf and larger outlet glaciers in the region forming ISW (Budillon and Spezie, 2000; Yoon et al., 2020). Cyclonic circulation transports ISW to the northeast (Figure 1a) (Cappelletti et al, 2010, Yoon et al., 2020) with ISW observed extensively in Terra Nova Bay at depths of 30-600 m, being deepest and most abundant near the Nansen Ice Shelf (Budillon and Spezie, 2000; Yoon et al., 2020). In the northeast, a narrow tongue of ISW at shallower depths, extending southeast from Gerlache Inlet, was consistently observed in multiple years of oceanographic profiling (Budillon and Spezie, 2000).
2.2 Campbell Ice Tongue Embayment

The Campbell Glacier flows from the Transantarctic Mountains into an embayment in north Terra Nova Bay where it forms the southward protruding Campbell Ice Tongue (Figure 1). The ice tongue consists of a fast flowing ‘Main Flow’ stream (13.5 km long and 4.5 km wide) and a smaller ‘Branch Stream’ to the West (8 km long and 2.5 km wide) (Figure 1b) composed of ice mélange (Han and Lee, 2014). Interactions between the Campbell Ice Tongue and adjoining fast ice have been observed where stresses induced by ice flow affected the distribution and strength of the fast ice, which in turn provided stabilization to the ice tongue (Han and Lee, 2018). The U-shaped grounding line of the ice tongue is ~14 km long (Figure 1b) (Han and Lee, 2014). The Main Flow stream is grounded in the southwest near the ice tongue terminus (Figure 2) (Han and Lee, 2014) and forms an ice rumple (Han and Lee, 2022). The Branch Stream is grounded in the east and west (Figure 2) (Han and Lee 2014) on bedrock (Souchez et al., 1995). Radio Echo Sounding of the ice tongue in 1999 showed a rippled undersurface and thicknesses of 200-300 m for most of its length, deepening to 700 m in the grounding zone (Bianchi et al., 2001).

Figure 1b shows the ice tongue splits north Terra Nova Bay into Gerlache Inlet in the west and Silverfish Bay in the east. Gerlache Inlet is 10-12 km from east to west and 13 km from north to south and includes Tethys Bay. Silverfish Bay is 25 km from east to west and 12 km from north to south. Multi-beam bathymetry (Figure 2a) collected in Gerlache Inlet (Jung et al., 2021; Lee et al., 2022) revealed complex seafloor morphology which shoals to depths of 25-200 m near the coast with deep troughs, ridges and a seabed rise near the ice tongue. Numerous smaller deep (~500 m) channels run from beneath the Main Flow stream of the ice tongue. We were unable to source any detailed oceanographic assessments in the study region. However, in the winter of 2000, near supercooled ocean temperatures were observed to the south of Tethys Bay (Cappelletti et al 2010). Tides in the region are diurnal with a maximum range of 0.60 m (Han and Lee, 2014) with strong tidal currents expected for the coastal morphology and bathymetry in Gerlache Inlet (Han and Lee, 2018). The Terra Nova Bay Polynya forms the southern boundary and should influence ocean circulation in the study region.

2.3 Fast ice and sub-ice platelet layer in north Terra Nova Bay

Fast ice consistently forms in Gerlache Inlet and Silverfish Bay with a SIPL first observed in biological studies (Vacchi et al., 2004, 2012). In November 2002, abundant Antarctic silverfish eggs and larvae were found within the SIPL in Gerlache Inlet identifying the first, and to date, only known nursery for this species in Antarctica (Vacchi et al., 2004). In late spring of 2005 and 2006, extensive surveys of silverfish egg distribution with respect to the occurrence of platelet ice were carried out (Vacchi et al., 2012), and revealed high abundance near the ice tongue in both bays where sea ice was thick (~2.5 m). In November 2017, AEM surveys detected thicker fast ice and SIPL near the ice tongue and northern coastline in both bays (Langhorne et al., 2023). AEM transects to the east and west of the ice tongue observed maximum SIPL thicknesses near the coast of 2.1 m and 1.6 m, respectively, which rapidly thinned to the south, i.e., away from the coast. Elsewhere, SIPL was detected with AEM in the centre and west of Silverfish Bay and in north Gerlache Inlet. Other point observations in the region recorded fast ice thicknesses of 2.4-2.5 m near the coast in Gerlache Inlet (Guglielmo et al., 2007; Vacchi et al, 2004). Near Tethys Bay, fast ice and SIPL thickness were respectively 1.4 m and 0.8-1.2 m in spring of 1997 (Guglielmo et al., 2007), and 2.4 m and 0-2 m in spring of 1999 (Lazzara et al., 2007).
Figure 1. a) Inset map of Antarctica (Gerrish et al., 2022), and Terra Nova Bay in the western Ross Sea showing locations of prior observations (red triangles) of marine ice and SIPL near ice shelves and ice tongues on the Victoria Land Coastline displayed on a MODIS image (4 November 2021). The Terra Nova Bay Polynya is active and shown with cyclonic circulation of ISW (blue lines) conjectured from Budillon and Spezie (2003). The red box shows the location of b) the study area with outlines of Campbell Ice Tongue area on 10 October 2020 (orange) and 11 October 2021 (dark grey). Fast ice sections are shown with drill sites (white circles), EM survey tracks (black line), and the Italian ‘Mario Zucchelli’ and Korean ‘Jang Bogo’ research stations (red crossed circles) on a Landsat-8 panchromatic image (5 November 2021).
Figure 2. a) Multi-beam bathymetry and depth measurements (Jung et al., 2021; Lee et al., 2022) with the North and NS- EM transects also shown in Figures 4, 5, and 6, b) EM consolidated ice thickness, c) EM SIPL thickness, and d) snow depth interpolated from drill sites displayed on a Landsat-8 panchromatic image (5 November 2021). Grounded regions of the ice tongue (Han and Lee, 2022) are shown as white lines and conjectured ISW outflow from the ice tongue illustrated as black arrows in a) and c). Black solid (dashed) line denotes the edge of fast ice in place for 6.5 (7) months.

3 Methods

To characterise the late spring fast ice conditions, we monitored fast ice formation in Gerlache Inlet and Silverfish Bay and polynya dynamics in Terra Nova Bay during winter and spring (March-November) of 2021 with Synthetic Aperture Radar (SAR) (Sentinel-1) and optical (Landsat-8 and Moderate Resolution Imaging Spectroradiometer (MODIS)) images, when available. In late spring of 2021 (3-7 November), the thickness distributions of snow, fast ice, and the SIPL were surveyed with drill hole and single-frequency electromagnetic induction (EM) measurements (Brett et al., 2024b, 2024c) over ~60 km$^2$ of fast ice in Gerlache Inlet as shown in Figure 1b.

At each drill site, two 30 m cross-profiles were laid out in north-south and east-west directions. Sea ice and SIPL thicknesses were measured in five drill holes, made at the centre and end points of each line, using a tape measure with weighted bar, and applying the resistance method (Price et al., 2014). Snow depth was measured at 0.5 m intervals along each cross-profile using a ruler (cm accuracy). The ~120 snow depth measurements were averaged to provide a mean value at each drill site and then spline interpolated using first-derivative minimum-curvature (no smoothing) to approximate snow distribution in the region (Figure 2d).

EM surveys of fast ice and SIPL thickness distributions were carried out using a single-frequency (9.8 kHz and 3.66 m coil-spacing) Geonics Ltd. EM31-MK2 instrument mounted on a sledge and towed by a vehicle. We configured the EM31 to sample at 1 Hz resulting in a geolocated measurement every ~5 m at typical travel speeds. EM measurements were made for 20 seconds over each of the 28 drill holes. A total distance of 39 km was surveyed with EM. We used the processing method (Irvin, 2018) applied in McMurdo Sound (Brett et al., 2020, 2024) to obtain sea ice and SIPL thicknesses from the in phase and quadrature components of single-frequency EM.

An electromagnetic forward model was run to compute the EM response over a layered subsurface. The model included three horizontal conductive layers with a range of thicknesses and conductivities: 1) consolidated ice thickness (i.e., sea ice plus the snow layer) (0.5-6 m; 0 mS m$^{-1}$), 2) SIPL (0-15 m; 100-1500 mS m$^{-1}$), and 3) seawater (2400-3000 mS m$^{-1}$). A ‘brute force’ inversion was applied, which compared EM inphase and quadrature readings at drill holes with theoretical forward modeled values, to obtain the best-matching bulk SIPL conductivity model, as determined by Root Mean Square Error (RMSE) between drill hole and inverted thicknesses. EM inverted sea ice (Figure 2b) and SIPL thicknesses (Figure 2c) were then linearly interpolated.

Given that seawater conductivity in the region was unconstrained, we ran multiple inversions with fixed seawater conductivities, and looped through a range of SIPL conductivities
from 100-1500 mS m$^{-1}$ in 50 mS m$^{-1}$ increments. We did this individually for each seawater conductivity from 2400-3000 mS m$^{-1}$ in 100 mS m$^{-1}$ increments. An optimum SIPL bulk conductivity of 600 mS m$^{-1}$ was consistently returned. We then ran inversions with a fixed SIPL bulk conductivity of 600 mS m$^{-1}$ for the same range of seawater conductivities, resulting in an optimum value of 2700 mS m$^{-1}$ determined from the RMSE for the entire study region (Figure 3a).

In this study, snow measurements covered a smaller area than EM surveys, and we could not account for the contribution of the snow layer to EM consolidated ice thickness. However, snow measured at drill sites and sighted along EM tracks was generally thin and loosely-packed. The mean snow depth of the 28 drill holes was 0.08±0.05, 3.7% of the mean drill hole measured consolidated ice thickness (2.17±0.25 m). We thus expect the addition of snow to be ≤0.10 m or <5% of EM consolidated ice thickness and negligible.

The combined thicknesses of consolidated ice and SIPL affect the error of the inversion to a mean relative value of ~10%, for thicknesses of either layer over 2 m (Irvin, 2018). When consolidated ice and SIPL are very thin or thick, they occupy a sensitive region of the forward model where a small change in the EM response can produce large changes in inverted thicknesses. This limitation and its implications for inversion of consolidated ice when a thin SIPL is present are described in detail in section 3.1 and Figure 2c of Brett et al., (2024a).

Figures 3b and 3c show a 1:1 comparison of drill hole measured consolidated ice and SIPL thicknesses with spatially coincident EM thicknesses inverted from the model. R$^2$ values for linear fits of EM versus drill hole thicknesses were 0.83 for consolidated ice and 0.90 for SIPL, the latter having a wider range. The mean drill hole (EM) measured thicknesses at drill holes were 2.17±0.25 m (2.09±0.27 m) for consolidated ice and 1.77±0.53 m (2.04±1.22 m) for SIPL. A mean deviation of -3.7% for consolidated ice and +10.7% for SIPL, was observed with EM relative to drill hole thickness. Drill hole point measurements will not capture SIPL thickness variability over metre-scale distances and an estimated error of ±0.10 m can be incurred when identifying the bottom of the SIPL with the resistance method (Price et al., 2014).

Generally, the data plotted near the 1:1 lines (Figures 3b and 3c) and deviation was small for most drill sites, which demonstrated that the inverted seawater (2700 mS m$^{-1}$) and SIPL (600 mS m$^{-1}$) conductivities were representative of conditions in the study region. However, five EM SIPL thicknesses (circled in Figure 3c) measured at a drill site nearest the ice tongue in the northeast (Figure 1b) were 1.4-2 m greater than the coincident drill hole measurements. In contrast, consolidated ice thicknesses at this site matched closely with a small mean deviation of +0.04 m for EM versus drill hole measurements.

The relatively large deviation between EM and drill hole measured SIPL thicknesses suggested a shift in the subsurface conductivity near the ice tongue, caused by a change in seawater conductivity, which we discuss further in section 5.1. We thus interpreted absolute magnitudes of EM SIPL thickness in the northeast region near the ice tongue with caution because they could be overestimated. However, the main objective of this study was to constrain and quantify the late spring distributions of ISW-influenced fast ice and SIPL thickness and we are confident that the EM method did this well.
Figure 3. a) RMSE values from EM inversions for a fixed SIPL bulk conductivity of 600 mS m$^{-1}$ and a range of seawater conductivities (2400-3000 mS m$^{-1}$). Comparison of coincident inverted EM (at drill holes) and drill hole measurements of (b) consolidated ice (sea ice plus snow) and (c) SIPL thickness with R$^2$ values from linear fits.

4 Results

4.1 Fast Ice Formation

Satellite observations revealed that the fast ice from the previous year broke out from Gerlache Inlet and Silverfish Bay. This was immediately followed by calving of the ice tongue from the terminus and eastern flank as highlighted by the difference in ice tongue area in October 2020 and 2021 in Figure 1b. In early March, sea ice began to form in the study region and polynya activity increased. During calm periods, thermodynamic sea ice growth was observed with fast ice attaching to the coastline, ice tongue, and Cape Washington. Thin fast ice was frequently broken up and advected eastwards by the polynya where it coalesced along the western margins of the ice tongue and Cape Washington.

In early November 2021, the fast ice in north Terra Nova Bay was extensive and covered an area of 168 km$^2$ in Gerlache Inlet and 547 km$^2$ in Silverfish Bay. In Gerlache Inlet, the fast ice was entirely first-year in composition and made up of three sections (ages) as shown in Figure 1b: Tethys Bay and northeast section (7-months), coastal fringe (6.5-months), and southern section (4.5-months). Fast ice in Gerlache Inlet first persisted from 28 March 2021 in Tethys Bay and in the northeast between the Branch and Main Flow streams (i.e., north of the 7-month line).
From mid-April, fast ice persisted in both bays, and extended in a straight line from the ice tongue terminus to the western coastline, and tip of Cape Washington. Polynya activity in mid-June caused an extensive breakout of fast ice that had persisted in Gerlache Inlet for 2 months. The remaining fast ice consisted of a 3-6 km wide fringe attached to the coastline and ice tongue (Figure 1b) which persisted over shallower bathymetry (0-300 m) (Figure 2a) (i.e., north of the 6.5-month line in Figure 1b). Fast ice to the south of this coastal fringe persisted from late-June and had been established for 4.5-months when surveyed in early November.

4.2 Thickness Distributions of Fast Ice and Sub-ice Platelet Layer

Figure 2 shows the thickness distributions of snow depth measured at drill sites, and EM measured consolidated ice and SIPL in early November 2021. Snow coverage in Gerlache Inlet (Figure 2d) was thin, wind-distributed, and had a flat surface. Mean snow depth at drill sites was 0.08±0.05 m and ranged from bare ice in the west, deeper snow in the centre (0.16 m), and thinner coverage (0.09 m) in the east and northeast. Mean drill hole-measured fast ice thickness was 2.09±0.27 m for the entire survey region, and 2.26±0.27 m and 1.76±0.00 m, for the 6.5 and 4.5-month fast ice sections respectively. The thickest fast ice (2.40 m) was measured in the northeast near the ice tongue (i.e., 6.5-month old fast ice) and the west (2.25 m) where the ice was bare of snow. Mean drill hole measured SIPL thickness was 1.70±0.54 m, varying from 1.30 m at the western drill hole sites to 2.67 m in the northwest. Mean SIPL thicknesses for the 4.5 and 6.5-month fast ice were 1.47±0.12 m and 1.81±0.66 m, respectively.

The more extensive and higher resolution EM surveys (~5 m sample spacing) detected the thickest consolidated ice (Figure 2b) and SIPL (Figure 2c) in the northeast near the Branch and Main Flow streams, with thicknesses decreasing to the south and west. EM consolidated ice (Figure 2b) was also thicker in the far west where the ice was bare of snow. Mean EM consolidated ice and SIPL thicknesses were respectively 2.14±0.31 m and 2.17±1.52 m from all EM surveys. Differences in thickness with respect to ice age were apparent, with thinner mean consolidated ice (1.82 m) and SIPL (1.39 m) measured on the younger 4.5-month fast ice versus the older 6.5-month section, with respective thicknesses of 2.35 m and 2.72 m. We observed the thinnest consolidated ice (~1.6 m) on the 4.5-month fast ice in the southeast and the thinnest SIPL in west Gerlache Inlet and the small inlet south of Jang Bogo.

Refer to Figure 2 for the location of EM transects in relation to bathymetry, grounded regions of the ice tongue, EM consolidated ice and EM SIPL thicknesses, and interpolated snow distributions. For reasons discussed in sections 3 and 5.1, we interpreted absolute magnitudes of SIPL thicknesses in the northeast near the ice tongue with caution. Additionally, on all EM transects, the effect of the inversion over thin SIPL (<0.5 m) (as described in section 3) was evident in EM inverted consolidated ice thickness. This manifested as sharp spikes in consolidated ice thickness and co-variance with SIPL thickness when the SIPL decreased to <0.5 m. From previous experience of this effect in McMurdo Sound (refer to Brett et al., 2024a), we interpreted the spikes in EM consolidated ice thickness as an unwanted effect and EM SIPL thickness variability as a real observed effect. True EM consolidated ice thicknesses generally corresponded to smaller magnitude EM thickness measurements as indicated by coincident drill hole measurements (Figures 4 and 6), when available.

The EM transects in Figures 4 and 5 show a pattern of SIPL thickness increasing from west to east, on both the 6.5 and 4.5-month sections. Consolidated ice thickness on the NS-2 to NS-3 transects (Figure 4) displayed the same increasing trend towards the ice tongue. The NS-
CIT (Figure 5) transect nearest the ice tongue had similar consolidated ice and SIPL thicknesses as NS-3 (Figure 4c). The North transect (Figure 6) had the thickest consolidated ice and SIPL with substantial thicknesses observed in front of the ice tongue in the east.

The NS-I transect in Figure 4a revealed thick fast ice (2.3 m) and thin SIPL (0-1.9 m) in the west. The thickest SIPL was observed near the northern coastline with isolated pockets of SIPL (1.4-1.9 m) along the profile.

The NS-2 transect (Figure 4b) showed very thick consolidated ice of ~3 m with no SIPL over shallow bathymetry (0-50 m) in the north, in the small inlet south of Jang Bogo (Figure 2). However, this thick 3 m consolidated ice may have resulted from the inversion where SIPL is less than 0.5 m. At latitude -74.632° (Figure 4b), consolidated ice thickness decreased sharply to ~2-2.2 m with a concurrent increase in SIPL thickness to >2 m. The SIPL was consistently thick (2-4 m) under the 6.5-month fast ice section. A 5-6 m SIPL maxima centred at latitude -74.636° (Figure 4b) was observed east of the Branch Stream (Figure 2c). Sharp decreases in consolidated ice and SIPL thickness were observed on the transition from the 6.5 to 4.5-month fast ice (Figure 4b), which occurred over deepening bathymetry (Figure 2a) (220-500 m depth over 1 km). On the 4.5-month fast ice, consolidated ice was ~1.7 m and SIPL thicknesses varied from 0-1.5 m.

On the NS-3 transect (Figure 4c), a thick SIPL maxima (centred at -74.643°) with thicker overlying consolidated ice (2.6-2.8 m) was observed. In the far north (towards latitude -74.63°), SIPL increased in thickness towards the ice tongue, and between the Branch and Main Flow streams (Figure 2c). On the younger ice, consolidated ice and SIPL thicknesses were thinner with several SIPL maxima of 2.2-2.4 m thickness and 100-130 m width observed.

The NS-CIT transect in Figure 5 was carried out ~0.5-1 km from the ice tongue and partially adjacent to the NS-3 transect in the north (Figure 2a). On the 6.5-month fast ice, consolidated ice thickness was ~2.5 m in the north and up to 3 m over narrow SIPL maxima. Over deeper bathymetry (300-400 m), the SIPL was thicker with 3 maxima of ~50 m width (centred at latitude -74.65°) observed over a short distance (~200 m). Both consolidated ice and SIPL thickness decreased in the north between latitudes -74.643° and -74.633° where bathymetry shoaled to 250 m. The transition to the 4.5-month old fast ice was detected and occurred north of where the ice tongue was grounded (location indicated on Figure 5). The younger fast ice had consolidated ice thickness of ~1.7 m and SIPL thickness of ~2 m. Prominent SIPL maxima were observed in narrow bands of ~100-150 m width with thicker overlying fast ice (2.3-2.4 m).

The 6.5 km North EM transect in Figure 6 detected SIPL along the entire northern coast with very thick and narrow accumulations in front of the Branch Stream of the ice tongue. On the western side of this transect, mean consolidated ice thickness was 2.2 m and SIPL varied from 2-3 m. In the centre, SIPL was thicker with a very narrow (~15 m) and thick SIPL maxima (4-12 m) observed over shallow depths (<25 m) east of the Branch Stream. In the east, the SIPL was significantly thicker than the west with four SIPL maxima with thicker overlying total ice observed due south of the Branch Stream. We estimated the widths of the SIPL maxima to be 100-500 m, which indicated highly channeled outflow from the ice mélange of the Branch Stream. To the south of the 0.5 km grounded region in the east (location indicated on Figure 6), the SIPL was substantially thinner.
Figure 4. Parallel north-south EM transects of consolidated ice (positive y-axis) and SIPL thickness (negative y-axis) spaced from west to east at 3.5 km distance, with corresponding maximum, minimum, and mean drill hole measurements (illustrated by dashes), on a) NS-1, b) NS-2, and c) NS-3. Dashed vertical lines show the transition between the 4.5 (south and left) and 6.5-month (north and right) old fast ice sections.
Figure 5. The NS-CIT EM transect (upper panel) of consolidated ice (positive y-axis) and SIPL thickness (negative y-axis) with bathymetric depth (lower panel) extracted from Jung et al., (2021). Dashed vertical lines show the transition between the 4.5 and 6.5-month old fast ice sections with the estimated extent of the adjacent grounded region of the Main Flow stream to the east from Han and Lee, (2022).

Figure 6. The North alongshore EM transect (upper panel) of consolidated ice (positive y-axis) and SIPL thickness (negative y-axis) with bathymetric depth (lower panel) extracted from Jung et al., (2021). The dashed vertical arrow shows the western extent of the Branch Stream and horizontal grey lines the estimated locations of grounded regions of the ice tongue to the north from Han and Lee, (2022).
5 Discussion

5.1 EM inverted consolidated ice and SIPL thicknesses

The mean deviations of coincident EM and drill hole measured consolidated (-3.7%) and SIPL (+10.7%) thicknesses (Figure 3c and 3b) were within the expected error of the inversion (Irvin, 2018). Given that the mean snow depth at drill sites was <5% of the mean drill hole consolidated ice thickness, we considered EM measured consolidated ice thickness as a good approximation of true fast ice thickness, except where the SIPL was less than 0.5 m in thickness. The close agreement of EM and drill hole measurements provided confidence that the conductivities of the SIPL (600 mS m$^{-1}$) and seawater (2700 mS m$^{-1}$) were generally representative of conditions in Gerlache Inlet.

However, the large deviation of inverted SIPL thicknesses at the drill site in the northeast (location in Figure 1b and circled in Figure 3c) suggested a different subsurface conductivity near the ice tongue. We queried this by running multiple forward models, with seawater conductivities ranging from 1000-2400 mS m$^{-1}$ and SIPL conductivities from 100-1000 mS m$^{-1}$, for the five drill holes made at this site. The optimum SIPL bulk conductivity was consistently 600 mS m$^{-1}$ for this seawater range. We then ran multiple forward models for a fixed SIPL bulk conductivity of 600 mS m$^{-1}$ and a seawater conductivity range of 1000-2400 mS m$^{-1}$. An optimum seawater conductivity of 1300 mS m$^{-1}$ was inverted, which corresponds to a salinity of 15.5 (for a temperature -1.9°C and pressure of 15 dbar) (McDougall and Barker, 2011). This brackish water suggests abundant freshwater in the upper surface ocean in the northeast, which corroborates high basal melt rates of the ice tongue deduced from SAR interferometry (Han and Lee, 2015). However, this low seawater conductivity was inverted from only five coincident drill hole and EM measurements and requires oceanographic assessment to be conclusive.

Applying the same technique in McMurdo Sound, Brett et al., (2020, 2024) used a seawater conductivity of 2700 mS m$^{-1}$ obtained from extensive oceanographic observations in the region (e.g., Robinson et al., 2014). In Gerlache Inlet, Brogioni et al., (2023) measured seawater salinity of 31.2-32.1 g kg$^{-1}$ in the centre and near Tethys Bay on 18 November 2018, corresponding to a conductivity of ~2500 mS m$^{-1}$ for a temperature of -1.91°C and pressure of 15 dbar (McDougall and Barker, 2011). This is slightly lower than the seawater conductivity of 2700 mS m$^{-1}$ we obtained from EM measurements of the full study region. However, seawater conductivity in Gerlache Inlet could be spatially and temporally variable with incursion of currently unknown water masses, polynya dynamics, freshwater input from the ice tongue, and sea ice melt later in the season. Without coincident oceanographic observations, we could only assume a mean seawater conductivity for the entire study region.

5.2 Fast ice formation

Fast ice formation during winter in Gerlache Inlet and Silverfish Bay was influenced by coastal morphology and the Terra Nova Bay Polynya. Throughout winter, the polynya frequently broke up recently formed fast ice and advected it eastward where it accumulated along the western boundaries of the Campbell Ice Tongue and Cape Washington. We mostly observed smooth and level fast ice in Gerlache Inlet, indicating predominantly thermodynamic growth. However, rougher dynamically-formed fast ice could be present near Cape Washington and the fast ice edge, as detected with AEM in both bays in 2017 (Langhorne et al., 2023).
The coastline, including the southward protruding ice tongue and Cape Washington provided anchorage for fast ice as it forms, and once established, protection from katabatic winds and polynya dynamics. This was indicated by fast ice extent in winter, which formed a straight line from west to east, across the ice tongue terminus, which was also observed in winter of 2018 (Brogioni et al. 2023). The coastal fringe, as described in section 4.1, reoccurred over winter in 2021, and was also observed in late spring of 2005 (Vacchi et al. 2012), and in winter of 2018 (Brogioni et al., 2023). The coastline is likely the dominant influence on the reoccurrence of the fast ice fringe, but shallower bathymetry (<300 m) could also aid fast ice persistence by providing shelter from polynya-driven and open ocean circulation. This is supported by the breakout of well-established 2-month old fast ice in mid-June, which left coastal-bound fast ice over shallower depths (<300 m).

We observed the earliest establishment of fast ice in northeast Gerlache Inlet near the ice tongue where the thickest fast ice and SIPL were measured in late spring (Figures 2b and 2c). Fast ice near the coast likely grows and strengthens more rapidly from heat loss to supercooled ISW and platelet ice consolidation (Gough et al., 2012). Thicker consolidated ice on the North, NS-3, and NS-CIT transects over SIPL maxima near the ice tongue, indicated sustained ISW outflow as the fast ice was forming over winter, implying that supercooled ISW outflow played an early and consistent role in fast ice formation. This is also supported by thicker consolidated ice in the northeast where snow was present. Although, the snow layer was thin (<0.10 m), it would still diminish sea ice growth (Wongpan et al. 2021), indicating that ISW had significantly influenced fast ice growth and thickness in proximity to the ice tongue. In future campaigns, ice core texture analysis would both constrain the timing and quantify the influence of in situ supercooling on consolidated fast ice growth in the region (Dempsey et al., 2010; Langhorne et al., 2015).

5.3 Sub-ice platelet layer distribution

Although, EM inverted SIPL thicknesses in the northeast region near the ice tongue were likely overestimated by 1-2 m, we are confident that the distributions of ISW-influenced fast ice and SIPL captured with EM were representative of ice conditions in the region in late spring of 2021. The SIPL was thickest in the north, northeast, and east near the Branch and Main Flow streams of the ice tongue and decreased in thickness to the west and southwest (Figure 2c). A similar pattern was observed with the AEM surveys in November 2017 (Langhorne et al., 2023). We thus confidently infer that the in situ supercooled ISW that forms the SIPL in Gerlache Inlet is outflowing from beneath the Campbell Ice Tongue. Multiple narrow (~100’s m) SIPL maxima south of the Branch Stream in Figure 6 identified highly channelled outflow from the ice mélange. This is supported by satellite images which showed gaps between icebergs within the mélange, directly north of the SIPL maxima.

Fast ice persistence determines how long platelet ice crystals (once present) can accumulate and grow in situ to form the SIPL. Two months of additional SIPL growth is apparent on all north-south transects (Figures 4 and 5). SIPL thickness decreased significantly from the 6.5 to the 4.5-month fast ice, which strongly indicates that the SIPL was well-established, and of significant thickness in mid-June, prior to the seaward fast ice breaking out. The fast ice breakout in mid-June occurred north of where the Main Flow stream is grounded on a seafloor rise (Figures 2 and 5), perhaps due to a mechanical effect of ice tongue motion on adjacent fast ice. The grounded region of the Branch Stream correlated with thinner SIPL (Figure
suggesting hindrance of ISW outflow. These effects support the location of grounded regions obtained from SAR interferometry (Han et al., 2022).

In comparison to the AEM surveys in November 2017 by Langhorne et al., (2023), the SIPL we measured in Gerlache Inlet was thicker and more extensive. However, our overestimate in EM inverted SIPL thickness could be a contributing factor especially near the ice tongue. Additionally, the fast ice was less extensive in November 2017 and had partially broken out from Gerlache Inlet in early September 2017. Interannual variability in SIPL in this region is expected and could be driven by a combination of factors including wind-forcing on fast ice persistence, and ocean circulation affecting the volumes of in situ supercooled ISW forming and circulating in the region. The snow layer contributes to SIPL formation by diminishing congelation growth of sea ice and allowing platelet ice crystal deposition to overtake sea ice growth (Dempsey et al. 2010; Wongpan et al., 2021). The SIPL was thickest where snow was present in the east.

5.4 Campbell Ice Tongue interactions

As described, the Campbell Ice Tongue provided anchorage for fast ice formation and protection from katabatic winds and polynya-driven circulation. The presence of fast ice, in turn, stabilised the ice tongue as indicated by the calving that occurred immediately after the fast ice broke out in January 2021. In February 2021, dark open water regions were observed between icebergs within the Branch Stream, indicating that fast ice is present within the ice mélange. Given significant outflow of supercooled ISW from the Branch Stream as shown by thick SIPL maxima downstream of gaps between icebergs, platelet and marine ice likely contribute to the integrity of the Branch Stream mélange. We speculate that the U-shaped grounding line of the ice tongue guides ISW to the upper surface ocean from deeper source regions. Thicker SIPL maxima observed along the Main Flow stream in transect NS-CIT in Figure 5 also suggest that ripple formations in the base of the ice tongue, as observed by Bianchi et al., 2001, could channel ISW outflow. Stevens et al., (2017) conjectured that similar basal channels in the Drygalski Ice Tongue would affect water mass outflow and variability of properties.

5.5 Ice Shelf Water source regions and circulation

Here, we discuss local and regional processes that could contribute to the formation of supercooled ISW and platelet ice in north Terra Nova Bay. The distribution of thicker fast ice and SIPL in Gerlache Inlet clearly show that in situ supercooled ISW is outflowing from beneath the Campbell Ice Tongue (conjectured as black arrows in Figures 2a and 2c). In Silverfish Bay, the AEM surveys detected a substantially thicker and more extensive SIPL in November 2017 (Langhorne et al., 2023) signalling that higher volumes of in situ supercooled ISW are outflowing to the east of the ice tongue. These observations show that the ice tongue is a significant source of ISW in north Terra Nova Bay.

The tongue of ISW previously observed near Gerlache Inlet by Budillon and Spezie, (2000) had properties that indicated that it was formed by HSSW interaction with glacial ice and pointed to the Campbell Ice Tongue as the source region. Given that the study area is small, and the polynya dominates the fast ice edge in the south, it is plausible that HSSW formed within the polynya could be causing basal melt of the ice tongue at depth and forming ISW. Basal freezing (mean rates of 0.75 m a$^{-1}$ and up to 20 m a$^{-1}$) has been deduced at ~400 m depth from SAR interferometry (Han and Lee, 2015). To be in situ supercooled at this depth, ISW would need to be of deeper origin. Bianchi et al., (2001) estimated the grounding line depth of the Campbell Ice
Tongue to be ~700 m and comparable to the Nansen Ice Shelf (660 m) (Frezzotti et al., 2000), which is posited to be the main source of ISW in Terra Nova Bay. In addition, the ice tongue thins over a short distance (Bianchi et al., 2000) which provides a steep slope for rapid pressure relief, inducing in situ supercooling, and promoting frazil/platelet ice formation (Jenkins & Bombosch, 1995; Lewis & Perkin, 1986).

HSSW has been observed in the water column from the seafloor to 400-700 m depth in summer (Yoon et al. 2020) and is more prevalent in winter when sea ice production in the polynya is intense (Ackley et al., 2020). This is deep enough for HSSW to interact with the Campbell Ice Tongue at the 700 m grounding zone. The complex bathymetry of north Terra Nova with its multiple deep troughs and channels could allow HSSW to circulate beneath the ice tongue. However, other unidentified water masses could also cause basal melting and we highlight that the region requires coincident oceanographic surveying in future fast ice observation campaigns. We conclude that basal melting of the Campbell Ice Tongue at depth is a significant contributor to ISW in Terra Nova Bay. This supports the finding of Langhorne et al., (2023) that smaller ice bodies along the Victoria Land Coastline are important contributors to ISW in the western Ross Sea.

5.6 Consequences of Campbell Ice Tongue degradation

The Campbell Ice Tongue is in a state of change with increased basal melting and decreased ice tongue area (Han et al, 2022). Without anchorage and protection provided by the ice tongue, fast ice establishment and persistence would diminish. Reciprocally, less extensive or thinner fast ice would negatively impact the ice tongue by providing less mechanical reinforcement and increased exposure to winds, ocean swell, and waves. Collapse of the ice tongue would result in more glacial ice discharge to the ocean and a single wide bay with less shelter from winds and ocean circulation. A reduction in fast ice persistence would reduce SIPL formation as shown by the thinner SIPL observed beneath the 4.5 versus the 6.5-month fast ice. Less ISW formation as a result of ice tongue disintegration would significantly impact the formation and stability of fast ice and SIPL. A reduction or disappearance of a SIPL would have major implications for Antarctic silverfish and other marine species reliant on this unique habitat. Antarctic silverfish is a key species in the polar marine food web, and the region is highly important for their survival in the Ross Sea. Complex interactions between the ice tongue and fast ice are clearly important for the stability of the icescape and reliant ecosystems in this region.

6 Conclusions

In north Terra Nova Bay, a sub-ice platelet layer (SIPL) occurs beneath fast ice beside the Campbell Ice Tongue, signalling the presence of in situ supercooled ISW in the upper surface ocean. Prior to this study, no dedicated in situ measurement of the thickness distributions of fast ice and SIPL, formed by supercooled ISW circulation, had been carried out and the source of ISW and platelet ice in this region was not well constrained. In late spring of 2021, we carried out drill hole and high-resolution ground-based electromagnetic induction surveys of fast ice and SIPL thickness distributions, to infer the pattern of ISW circulation in Gerlache Inlet in north Terra Nova Bay. To characterise the fast ice composition, we monitored fast ice formation throughout the winter of 2021 with satellite observations. We observed significant negative
forcing of the Terra Nova Bay Polynya on fast ice persistence, which was countered by the stabilising effect of the western coastline, Campbell Ice Tongue, and Cape Washington, and potentially sheltering by shallower bathymetry near the coast. In late spring, thicker fast ice and SIPL were observed near the ice tongue, with very thick and narrow SIPL maxima indicating highly channelled outflow of supercooled ISW, from beneath the ice tongue through ice mélange, subglacial formations, and grounded regions. We conclude that significant volumes of ISW are locally sourced from basal melt of the Campbell Ice Tongue. We identify the need for combined in situ fast ice and oceanographic surveying to fully elucidate the processes at play in this important region. Reciprocal effects observed between the ice tongue, fast ice, and SIPL highlighted the implications of Campbell Ice Tongue degradation for fast ice and SIPL formation, and marine species that are highly specialised to this region.

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Open Research

The drill hole (Brett et al., 2024b) and EM (Brett et al., 2024c) data are available at (DOI pending) at the World Data Centre PANGAEA (https://www.pangaea.de). Satellite images were obtained from the European Space Agency (ESA) (through https://search.asf.alaska.edu), the NASA Worldview application (https://worldview.earthdata.nasa.gov/) operated by the NASA/Goddard Space Flight Center Earth Science Data and Information System (EOSDIS) project, and Landsat-8 images courtesy of the U.S. Geological Survey. Multi-beam bathymetric data was contributed by Lee et al., (2022) and grounding line data by Han and Lee, (2022).
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