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Slowdown of Shirase Glacier caused by strengthening alongshore winds

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Abstract

Observations have confirmed basal melt rates of up to 16 m a^{-1} underneath the Shirase ice tongue in East Antarctica. These high basal melt rates are caused by intrusions of warm modified Circumpolar Deep Water (mCDW) onto the continental shelf, a mechanism responsible for widespread mass loss in West Antarctica, together with parts of Wilkes Land. In contrast to those regions, the catchment of Shirase Glacier has been gaining mass, a trend attributed to increased precipitation. Here, we report on the dynamical ocean-driven slowdown, thickening and grounding line advance of Shirase Glacier, in response to strengthening easterly winds that reduce mCDW inflow and decrease basal melt rates. Our findings are significant because they demonstrate that warm water regimes are not universally associated with glacier acceleration and mass loss in Antarctica, and they highlight the overlooked role of ocean forcing in the recent mass gain of the Dronning Maud Land sector.

1. Introduction

Shirase Glacier is the fastest flowing outlet glacier in East Antarctica, reaching speeds in excess of 2,200 m a^{-1} across its grounding line, before flowing into Lützow-Holm Bay (Fig. 1). Its annual ice discharge approaches 15 Gt a^{-1} (Rignot et al., 2019) and it drains a catchment containing 1.2 m of sea level equivalent (Fig.1, Morlighem et al., 2020). This rapid ice flow speed is associated with vigorous

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melt underneath its floating tongue, where basal melt rates have been observed to vary seasonally between 7 and 16 m a^{-1} in 2018, 16 km downstream of the glacier's grounding line (Hirano et al., 2020). These high melt rates are caused by warm modified Circumpolar Deep Water (mCDW) intruding onto the continental shelf and being transported directly to the glacier via bathymetric troughs (Fig. 1; Moriwaki & Yoshida, 1983; Hirano et al., 2020), a process referred to as Mode 2 melting (Jacobs et al., 1992). Elsewhere in Antarctica, most regions that experience this mode of oceanic melt have been losing mass, as enhanced melt has thinned ice shelves, causing glacier acceleration and mass loss e.g. the Amundsen Sea (Jenkins et al., 2018; Mouginot et al., 2014), the Western Antarctic Peninsula (Cook et al., 2016) and Wilkes Land (Rintoul et al., 2016; Greene et al., 2017). However, mass loss has not been observed at Shirase Glacier and, between 2003 and 2019, its wider catchment (sometimes referred to as drainage basin 7) gained mass at a rate of +25 ± 6 Gt a^{-1}, which is the largest imbalance of all drainage basins in East Antarctica (Smith et al., 2020), including the comparatively well studied Wilkes Land (drainage basin 13; -20 ± 14 Gt a^{-1}).

The mass gain and thickening in the Shirase catchment over the past two decades has been hypothesized to have been caused by increased precipitation (Schröder et al., 2019; Smith et al., 2020) from extreme snowfall events within the wider region (Boening et al., 2012; Lenaerts et al., 2013). Although, earlier field-based estimates, using repeat triangulation surveys in 1969 and 1973, demonstrated ice surface lowering of around 0.7 m a^{-1} around 100-200 km inland of the Shirase Glacier grounding line (Mae & Naruse, 1978; Naruse, 1979; Nishio et al., 1989). Furthermore, repeat GPS surveys in 1980 and 1988 revealed a thinning rate of around 0.5 m a^{-1} 100-150 km inland of the grounding line (Toh et al., 1992). These rates of surface lowering during that time are comparable with some of the fastest rates of thinning observed across Antarctica over the past decade at similar distances inland of the grounding line (Smith et al., 2020). This surface lowering may have a much longer-term signal with ice core records estimating a surface lowering of 350 m over the past 2000 years of the Mizuho Plateau (Kameda et al., 1990), which is located around 200 km inland of the Shirase coastline. The surface lowering over the past 2000 years is coincident with an increase in ice discharge from the Lützow-Holm Bay, which has been estimated from subglacial erosion rates (Sproson et al., 2021).

Oceanographic observations in Lützow-Holm Bay in 2018 have revealed a two-layered structure with a cool and relatively fresh layer of Winter Water overlying a warm and saline mCDW layer, where temperatures near the ice front seasonally exceed the *in situ* melting point by 2.7 °C (Hirano et al., 2020). Observations and modelling demonstrate a strong seasonal variation in the basal melt rate of the Shirase ice tongue (Hirano et al., 2020; Kusahara et al., 2021), which is caused by seasonal variations in the depth of the thermocline forced by the strength of the alongshore easterly winds near

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the continental shelf (Ohshima et al., 1996). There is no evidence of large seasonal variations in ice flow speed at the grounding line, but observations do show some seasonal variation in ice flow speed on the floating tongue that could be connected to external forcing (Nakamura et al., 2007; 2010).

There have been several studies analysing the ice flow dynamics of the Shirase Glacier, largely covering short sub-decadal time periods (Pattyn & Derauw, 2002; Pattyn & Naruse, 2003; Nakamura et al., 2010; Aoyama et al., 2013). However, the longer-term geological signal of ice sheet thinning and increased ice discharge, along with observations of thinning in the 1970s and 1980s, followed by thickening from the 2000s raise some important questions into the processes causing this switch. In this study, we produce a time series of ice flow speed that spans 47 years and show that long-term ice speed trends coincide with alongshore wind speeds. We then discuss how these observations may relate to wider hemispheric trends in atmospheric circulation and what this may mean for the future mass balance of the Shirase catchment and the wider Dronning Maud Land sector.

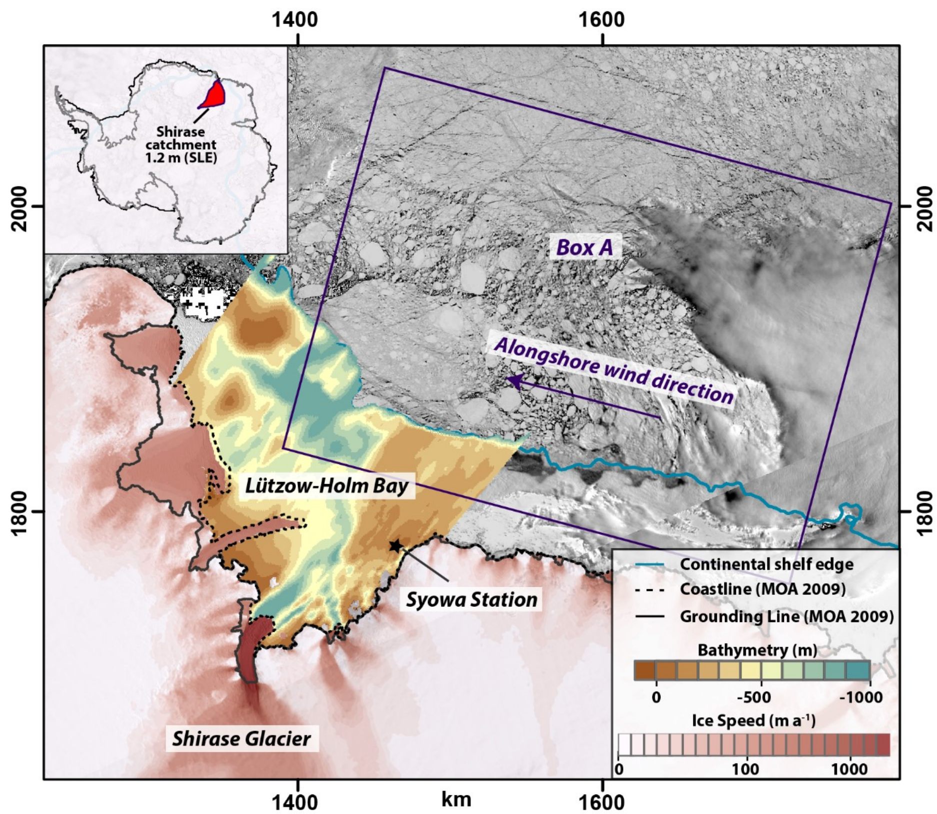

Figure 1: MODIS image of Lützow-Holm Bay and Shirase Glacier from the 4th November 2019 obtained from NASA WorldView. Overlain is the ITS_LIVE composite velocity product in logarithmic scale (Gardner et al., 2018; 2020), the MODIS 2009 grounding line and coastline (black line; Scambos et al., 2007; Harran et al., 2019), 1000 m bathymetric contour obtained from

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BedMachine (Morlighem et al., 2020) which is taken as the continental shelf boundary and bathymetry of the Lützow-Holm Bay (Kusahara et al., 2021). Note the deep trough connecting Shirase Glacier to the open ocean. The location of the Syowa research station and Box A, the region where ERA5 derived winds were extracted are also shown. The ~~coordinate~~ grid is in km and it is projected in Polar Stereographic.

2. Data and Methods

2.1 Ice-front position, ice speed, grounding line and ice thickness

We create a time series of ice-front positions between 1963 and 2020 using a variety of different sources including: ARGON imagery from 1963, Landsat-1 imagery from 1973, Landsat-5 imagery from 1984, Landsat-4 imagery from 1988, RADARSAT RAMP mosaic from 1997 (Jezek et al., 2013), and MODIS imagery from 2000-2020, with the spatial resolution of the satellite data ranging from 15-250 m. In each case, we map the outer limit of the collection of loosely bound icebergs that form the Shirase ice tongue that are typically surrounded by a smoother surface of fast ice (Fig. 1 & 2a). Errors associated with this mapping are insignificant in the context of the ice tongue typically advancing 2,500 m a^{-1}, or retreating in short-lived calving events typically >10 km.

We calculate 18 ice speed estimates for Shirase Glacier between 1973 and 2020. For 1973 we use a pair of Landsat-1 (band 7) images from the 25th January 1973 and the 21st January 1974 that we manually co-register to each other, before co-registering to a Landsat-8 image. The combination of the relatively coarse Landsat-1 imagery (60 m) and the development of surface melt ponding over the fast flowing section of the glaciers between the two images prevented the automatic extraction of ice speed. Instead, we extract an ice speed estimate by manually tracking the displacement of a prominent rift ~24 km downstream of the grounding line (Fig. 2c). Errors associated with the manual tracking of this rift stem from the co-registration between the two image pairs which we estimate to be one pixel (60 m; Animation S1). For 1988, we use a pair of Landsat-4 (band 3) images from the 14th January 1988 and the 15th February 1988 that we also co-register to a Landsat-8 image. The quality of the Landsat-5 images (30 m resolution) is superior to that of the Landsat-1 imagery and, in the absence of significant surface melt ponding, we use the feature tracking software COSI-CORR (Leprince et al., 2007; Scherler et al., 2008) to extract ice speed. For these images co-registration error is negligible (Animation S2) and error in the feature tracking is estimated at <0.5 pixels (e.g. Heid and Kääb, 2012), because of the close time separation of the image pairs this results in a larger error of ±171 m a^{-1}. For 2000-2018 we use 14 annual ice speed mosaics from the ITS_LIVE dataset which cover Shirase Glacier (Gardner et al., 2018) and use the corresponding error grids for error values, which range from 1 to 32 m a^{-1}. For 2019 (n=27) and 2020 (n=19) we take an average for all

GoLIVE generated ice speed fields (Fahnestock et al., 2016; Scambos et al., 2016) with a time separation of 16-320 days from scene ID's 149_109 and 150_109. Taking an average of multiple ice speed grids reduces error and, as such, we prescribe a nominal error of 16 m a^{-1}. We extract ice speed profiles from each time period across a transect, T1 (Fig. 2a), and also produce a time-series of ice speed change where T1 crosses the grounding line. In 1973, the only possible observation of ice speed was extracted 24 km downstream of the grounding line (Point x; Fig. 2a) and there are no observations directly at the grounding line. To account for this, we estimate ice speed at the grounding line in 1973 using the average difference between point x, 24 km downstream of the grounding line, and where T1 crosses the grounding line in each of the other 17 ice speed profiles (1988-2020). Across these profiles, ice speed was on average 2% slower (ranging from 1% to 4%) at the grounding line, compared to ice speed at point x. Therefore, to estimate ice speed at the grounding line in 1973 we reduce the ice speed observed 24 km downstream of the grounding line by 2 ±1%. We also include the measurements of ice speed from Nakamura et al. (2007) at the grounding line derived from the JERS-1 satellite in 1996, 1997 and 1998.

To estimate the direction and magnitude of any migration in the Shirase Glacier grounding line we compare time stamped digital elevation model (DEM) strips with a spatial resolution of 2 m from the 6th January 2013 and the 8th October 2015 from the REMA project (Howat et al., 2019). We select these strips because they cover the complete Shirase Glacier grounding line and represent the longest time gap in the record. This is in addition to a SPOT5-HRS DEM from the SPIRIT project (Korona et al., 2009) from the 8th February 2008, with a spatial resolution of 40 m. Elevation uncertainty is estimated at around 4 m by comparing derived elevations from exposed bedrock between the two REMA DEM's and a larger uncertainty of around 7 m between the SPOT5-HRS and REMA DEM's. The tidal amplitude of the region is limited to 0.2 m (Aoki et al., 2000) and is deemed insignificant. We extract elevation profiles along transect T1 (Fig. 2a) from these dates. A comparison of elevation profiles cannot provide a location of the true grounding line position, but any horizontal migration of these elevation slopes can provide reasonable estimates in both the direction and rate of grounding line migration (Fricker et al., 2009; Brunt et al., 2010).

We also extract an ice thickness change time-series from the dataset presented in Schröder et al. (2019) from point IT, which is around 20 km inland of the grounding line (Fig. 2a). This multi-mission dataset spans between 1978 and 2017 and contains data from a variety of satellites. We use the accompanying uncertainty estimates described in Schröder et al. (2019). We also utilize modelled melt rate anomalies of the Shirase ice tongue that are derived by an ocean model that is forced by ERA5 wind reanalysis between 2008 and 2018 by Kusahara et al. (2021). The melt rate dataset contains melt anomalies that have been simulated with and without fast ice cover, we use the melt

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rates with fast ice cover, but note that there is little difference between the two melt rate datasets (see Fig. 20; Kusahara et al., 2021)

2.2 Climatological data

We extract mean monthly ERA5 (Hersbach et al., 2020) 10 m zonal (U) and meridional winds (V) speeds with a gridded 30 km spatial resolution between 1979 and 2020 from an approximately 340 x 250 km box adjacent to the coastline (Box A; Fig. 1). We do not extend the box all the way into Lützow-Holm Bay because it is semi-permanently covered with landfast sea-ice (Fig. S1) that dampens the impact of winds on ocean circulation. We also utilize an observational wind direction and wind speed time-series (1966-2020) from the Syowa research station (See Fig.1) that is available via the SCAR Met READER (Turner et al., 2004). We then calculate alongshore easterly wind speed for both the ERA5 reanalysis data extracted from Box A and the observational time-series from Syowa research station, using an alongshore angle of 80° from due north:

$$A = W\cos(\theta - 80)$$

Where W is wind speed and θ is wind direction. Some differences between the ERA5 and observational record from Syowa are expected because the ERA5 record represents an average from a much larger region, while the observational data from Syowa research station are from a single spatial location.

3. Results

We observe a total range of nearly 90 km in the ice-front position of the Shirase ice tongue between 1963 and 2020. Its maximum extent was in 1963, before retreating to its minimum extent in 1988 (Fig. 2a, b). Since 1988 there has been a general pattern of advance with a few sporadic calving events (Fig. 2a, b). Most of the variation in the extent of the Shirase ice tongue is in the heavily fractured and unconstrained section of the floating ice tongue. The only exception to this was in 1988 when the ice tongue retreated to the entrance of the narrow and more constrained section of its fjord, 24 km advanced of its 2009 grounding line (Fig. 2a).

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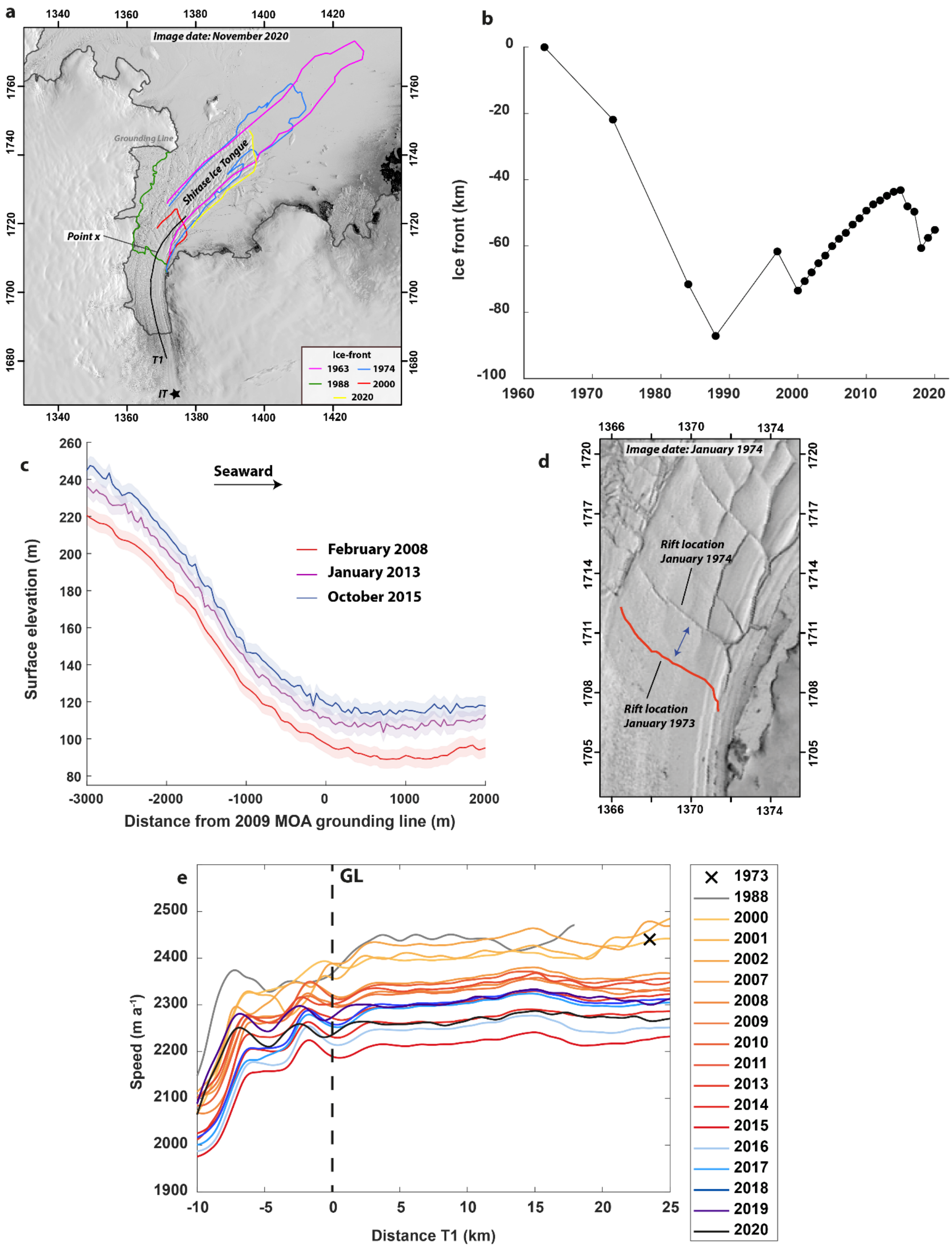

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177 **Figure 2: a)** Landsat-8 image from November 2020 showing the Shirase ice tongue. Overlain are selected ice-front positions from 1963, 1974, 1988, 2000 and 2020; along with the transect, T1, used to extract ice speed profiles and point x, which is the location of the 1973/74 ice speed estimate on the floating tongue. Point IT is the location of the ice thickness time-series. The grey line is the MODIS 2009 grounding line (Scambos et al., 2007, Harran et al., 2021). The coordinate grid is in

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km. **b)** Change in ice-front extent relative to 1963. **c)** Surface elevation profiles along a small section of T1 as it intersects the grounding line from February 2008, January 2013 and October 2015 showing a seaward migration of the surface slope d) The rift used to estimate ice speed in 1973/74. The red line is the digitized rift from January 1973. **e)** Ice speed profiles from transect T1 between 1973 and 2020. The black cross represents the ice speed measurement from 1973/74. Landsat images are courtesy of the U.S. Geological Survey.

Ice speed profiles along the transect (T1, Fig. 2a) show a uniform pattern of change across both the grounded and floating sections of Shirase Glacier (Fig. 2e). At the grounding line, we observe little change in ice speed between 1973 and 1988, although we note the larger uncertainty in the 1988 estimate of ±171 m a^{-1} (Fig. 3b). Between 1988 and 1996 we observe a 2 ±7% slowdown and a 2 ±1% increase in ice speed between 1997 and 2000 (Fig. 3a). Post-2000 we observe a slowdown, with an 8 ±1% decrease in ice speed between 2000 and 2015 (Fig. 3b). Between 2015 and 2019 ice speed increased by 4 ±1% (Fig. 3a). Elevation profiles along a section of T1 in 2008, 2013 and 2015 show a seaward migration of the surface slope as it approaches the grounding line (Fig. 2d), which is indicative of grounding line advance and can be visualised in animation S3. Between February 2008 and October 2015 we estimate that the grounding line has advanced around 400 m (~50 m yr^{-1}) from measuring the seaward displacement of the surface slope, an estimate that is consistent with CryoSat based observations of seaward grounding line migration between 2010 and 2016 (~30 m a^{-1}; Konrad et al., 2018). Observations of ice thickness change 20 km inland of the grounding line show a thinning trend of 0.27 ±0.33 m a^{-1} between 1987 and 1997, before reversing to a thickening trend of 0.19 ±0.10 m a^{-1} between 1997 and 2017 (Fig. 3b).

ERA5 derived estimates of alongshore wind speed between 1979 and 2020 show limited variation between 1984 and the early 1990s (Fig. 3a). In the early 1990s there was a small increase in alongshore wind speed, before a much more marked increase from 2000-2010 where alongshore wind speed increased from around 4 m s^{-1} to 4.8 m s^{-1} (Fig. 3a). This is before falling slightly to around 4.5 m s^{-1} between 2010 and 2018, which is coincident with an increase in basal melt rate anomalies (Fig. 3a). Long-term increased alongshore wind speeds at the Syowa research station in Lützow-Holm Bay between 1966 and 2020 are also observed with a small peak in alongshore wind speeds in the early 1990s before a sharp increase in along shore wind speed between the late 1990s and 2005 (Fig. S2). However, the observational record at Syowa also records a peak in alongshore wind speed in the early 1980s (Fig. S2) that is not present in the ERA5 offshore reanalysis record (Fig. 3a).

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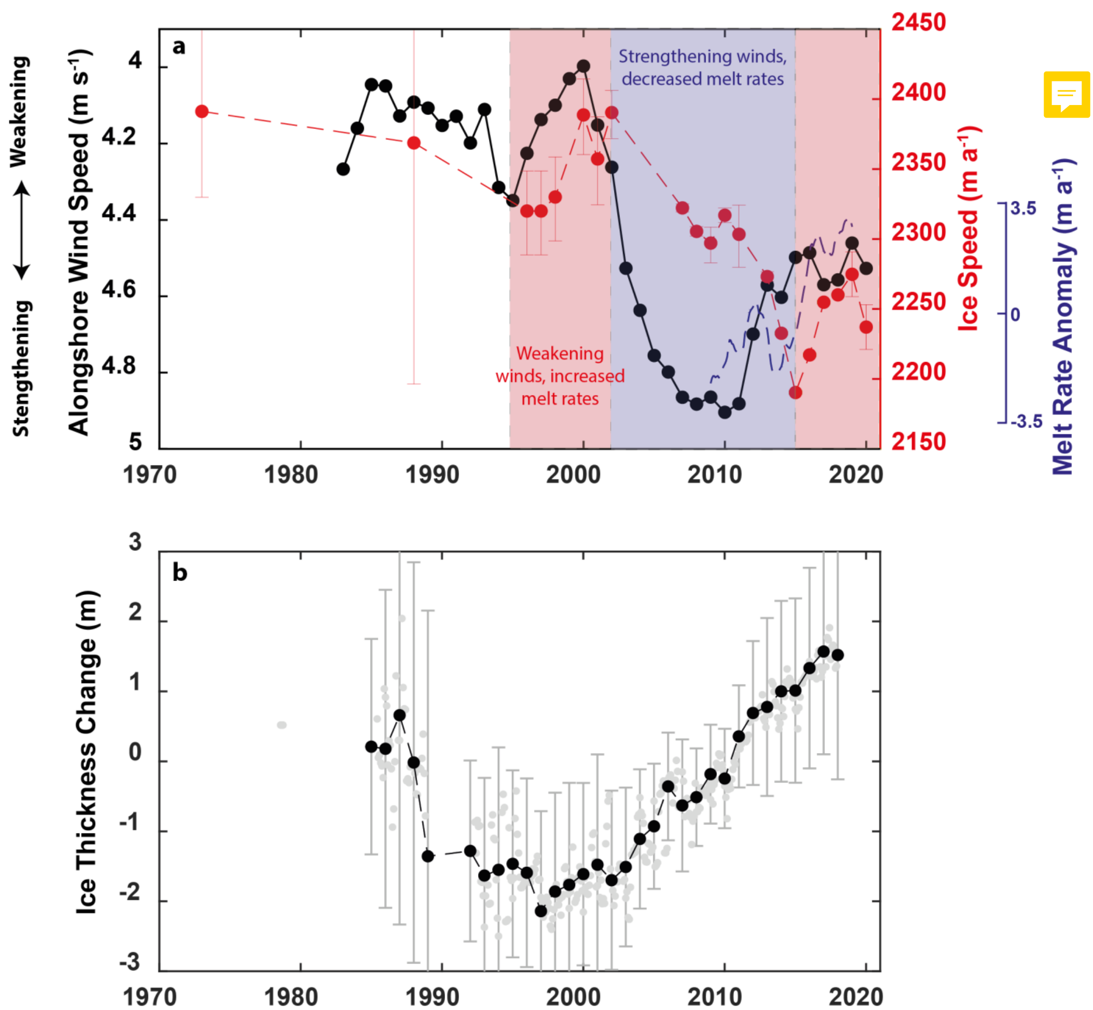

Figure 3: a) Annually averaged ERA5 derived alongshore wind speed from Box A (See Fig. 1) and plotted as a 5-year rolling mean (black), ice speed at the Shirase Glacier grounding line along T1 (red) and modelled melt rate anomaly of the Shirase ice tongue between 2008 and 2018 (blue; Kusahara et al., 2021). Periods of weakening winds cause increased mCDW transport, increased basal melt and acceleration. Periods of strengthening winds result in relatively less mCDW transport, decreased basal melt rates and glacier slowdown. Note that alongshore wind seed is plotted with an inverted axis. **b)** Annually averaged ice thickness change at point IT (see Fig. 2a) extracted from the Schröder et al. (2019) dataset, where there are at least 6 data points in the calendar year. The error bars are annually averaged errors. The background grey points are the raw monthly data points.

4. Discussion

4.1 Slowdown and thickening caused by strengthening alongshore winds

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Theoretical considerations indicate that the heavily fractured and unconstrained section of the Shirase ice tongue offers little buttressing force (Fürst et al., 2016). Therefore, it is unlikely that any varations in the extent of the Shirase ice tongue have had a direct effect on the ice speed trends we have observed (Fig. 4a). The only exception to this is in 1988 where the ice tongue briefly retreated to the edge of its more confined embayment (Fig. 2a), closer to where the extent of the ice tongue might be expected to exert buttressing and impact on inland flow speed, were it to be removed (Fig. 4a).

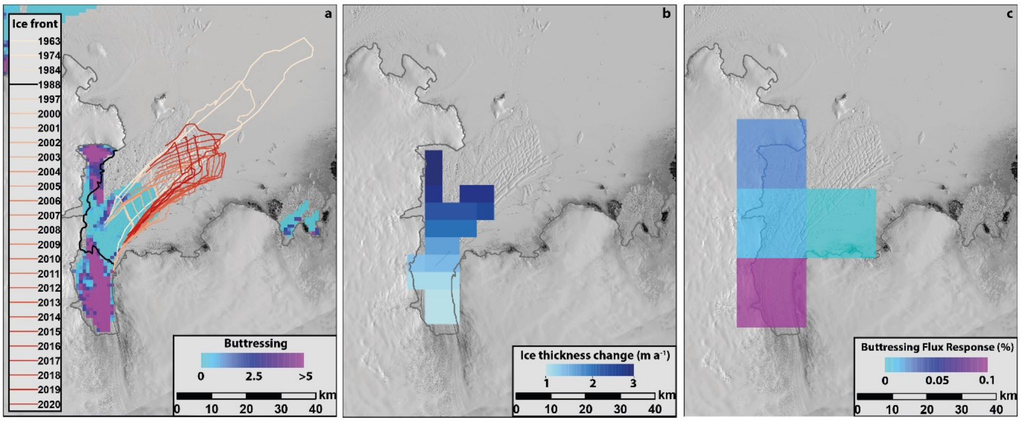

Figure 4: No change in ice tongue buttressing in response to observed changes in ice tongue extent, but increased buttressing expected in response to observed ice tongue thickening. **a)** Simulated maximum buttressing potential of the Shirase ice tongue (Durand et al., 2016). Light blues mean the ice is passive, purples mean the floating ice is dynamically important. No dynamically important ice has calved ~~over the past~~ over the past 57 years. **b)** Ice tongue thickness change between 2003 and 2019 showing thickening of the Shirase ice tongue (Smith et al., 2020). **c)** Simulated response of ice flux to thinning of floating ice in each grid cell by 1 m (Reese et al., 2018), floating ice near the grounding line is important for buttressing.

In agreement with previous work, we note that the observed fluctuations in ice tongue extent are correlated with landfast sea-ice conditions in Lützow-Holm Bay (Aoki, 2017). Long periods of ice tongue advance are associated with persistent landfast sea-ice in Lützow-Holm Bay, while ice tongue retreat is associated with landfast sea-ice break-out events removing parts of the ice tongue conglomerate (Aoki et al., 2017). It is unlikely that the fast ice has any major role in providing buttressing for Shirase Glacier. We note that there was no obvious spike in ice speeds on the grounding line or ice tongue in 1988 when the fast ice was completely removed from the bay (Fig. 2a, S3) and that Nakamura et al. (2010) recorded only a very modest 20 ± 30 m a^{-1} (0.8 ± 1.3 %) change in ice speed at the grounding line after a partial fast ice break-out event in 1998.

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Point IT, 20 km inland of the Shirase Glacier grounding line was thinning at a rate of -0.27 ±0.33 m a^{-1} between 1987 and 1997 (Fig. 3c), a pattern consistent with field observations up to 200 km further inland in the 1960s, 1970s and 1980s (Mae and Naruse, 1978; Naruse, 1979; Nisho et al., 1989; Toh and Shibuya, 1992). However, at ~2000 there was a slowdown in Shirase Glacier (Fig. 3a) and this thinning trend reversed to thickening (Fig. 3b). This slowdown and thickening coincides with an increase in alongshore wind speed adjacent to the Shirase coastline (Fig. 3a). The seasonal strenghtening in alongshore winds offshore of the Shirase coastline has been observed to deepen the thermocline in Lützow-Holm Bay, limiting the inflow of mCDW onto the continental shelf and reduce basal melt rates (Hirano et al., 2020). We suggest that this same process over annual to decadal timescales has caused the slowdown of Shirase Glacier.

Increased alongshore wind speed from ~2000 enhanced Ekman convergence at the coast, deepening the thermocline with a short lag and inhibited the inflow of warm mCDW into Lützow-Holm Bay. The subsequent cooling of Lützow-Holm Bay reduced the basal melt rate of the Shirase ice tongue. This reduction in basal melt caused the ice tongue to dynamically thicken and is confirmed by ICESat and ICESat-2 observations of the of the Shirase ice tongue that show a mean thickening of 1.87 m yr^{-1} from 2003-2019 (Fig. 4b; Smith et al., 2020). The dynamic thickening of the ice tongue increases buttressing through time (Reese et al., 2018; Fig. 4c) and ultimately drives the overall slowdown in ice speed, grounding line advance and inland thickening that we observe. Importantly, our results show that wind-driven ocean forcing is also contributing to the mass gain in the Dronning Maud Land regions in addition to surface mass balance processes (Boening et al., 2012; Lenaerts et al., 2013).

Within the longer term slowdown of Shirase Glacier between 1973 and 2020 we observe brief periods of acceleration in response to short-lived periods of weakening alongshore winds. For example, both of the accelerations in ice speed from 1997-2000 and 2015-2019 are preceded by brief periods of weakening alongshore winds (Fig. 3a). These periods of weakening alongshore winds cause relatively higher basal melt rates because they raise the thermocline closer to the ocean surface and enable a greater influx of mCDW into Lützow-Holm Bay (Hirano et al., 2020). For 2015-2019, this is supported by increasingly higher modelled melt rate anomalies (Fig. 3a). Because we observe coincidental acceleration in ice speed during this period of increased melt rates, it is indicative of a reduction in buttressing in response to ice tongue thinning. The interannual variability in ice flow speed at Shirase Glacier in response to wind-forced ocean variability is analogous to other regions of Antarctica where mCDW periodically floods the continental shelf e.g. Pine Island (Christianson et al., 2016), Thwaites (Miles et al., 2020) and Totten Glaciers (Greene et al., 2017). Although, the pattern of change at Shirase Glacier is unique because it is the only outlet glacier in Antarctica with a warm water regime, that has been observed to be slowing down and thickening during the 21st century, as opposed to accelerating and thinning (e.g. Mouginot et al., 2014; Greene et al., 2017). Therefore, this result highlights that this oceanic mode of ice melt is not universally associated with mass loss in Antarctica.

4.2 Wider links to climate forcing and future implications

In response to both increased greenhouse gas emissions and ozone depletion (Thompson et al., 2011; Wang et al., 2014; Perren et al., 2020) the band of mid-latitude westerly winds that encircle Antarctica have both strengthened and migrated southwards towards the ice sheet over recent decades (Thompson & Solomon, 2002; Marshall, 2003; Turner, 2005; Bracegirdle et al., 2018). In the Amundsen Sea sector this anthropogenic driven migration has been linked to westerly wind anomalies over the continental shelf (Holland et al., 2019), which have enabled a greater influx of warm mCDW onto the continental shelf and have driven enhanced localized ice sheet mass loss (Thoma et al., 2008). At Shirase Glacier, our observations of strengthening alongshore easterly winds suggest any southward encroachment of the mid-latitude westerlies has yet to impact the Shirase coastline. This may also be the case for parts of the wider Dronning Maud Land coastlines where alongshore easterlies have also been observed to have strengthened (Hazel & Stewart, 2019). This trend in strengthening alongshore easterlies is linked to enhanced katabatic winds as low pressure systems track progressively further south and enhance the pole to coast pressure gradient (Hazel & Stewart, 2019). It remains undetermined if the enhancement of the pole to coast pressure gradient has been influenced by the anthropogenically driven southerly migration of the mid-latitude westerlies, or if it has been caused by inherent natural decadal variability within the system.

Over the course of the 21st century, the southerly migration of the mid-latitude westerlies is projected to continue in line with a warming climate (Yin, 2005; Perren et al., 2020). Along the Shirase coastline, it remains unclear if this continued southerly migration will ultimately result in a similar situation to the Amundsen Sea, such that westerly wind anomalies offshore would result in enhanced mCDW transport into Lützow-Holm Bay and cause mass loss. Alternatively, the westerly winds may never migrate close enough to the Shirase coastline to impact alongshore winds, and instead, alongshore winds may continue to strengthen as the pole to coast pressure gradient increases. This would result in further cooling of Lützow-Holm Bay and ice tongue thickening, and further mass gain. In a wider context a greater understanding of the potential changes in ocean forcing in response to broader atmospheric patterns expected over the coming decades is needed in the Dronning Maud Land sector.

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5. Conclusion

Our observations of Shirase Glacier are a rare example of a glacier reversing a trend of mass loss from at least the 1970s-1990s to mass gain over the last two decades. As far as we are aware this is the only major fast flowing Antarctic outler glacier to display this pattern of behaviour. This reversal has been driven by a slowdown of the Shirase Glacier upstream of the grounding line in response to strengthening alongshore winds that have limited the inflow of warm mCDW into Lützow-Holm Bay, reduced basal melt rates and caused its ice tongue to dynamically thicken. This means that ocean forcing has contributed to some of the observed mass gain in Dronning Maud Land. It is not certain if these strengthening alongshore winds represent some form of inherent decadal-scale variability or if they are part of a multi-decadal trend, but we hypothesise that they are at least consistent with an anthropogenically-driven poleward migration of the mid-latitude westerly winds. Should this strengthening of alongshore easterly winds continue into the future, the Shirase catchment will continue to experience a positive mass balance due to both the slow-down in ice discharge, and to the predicted increase in precipitation in response to atmopsheric warming (e.g. Ligtenberg et al., 2013; Kittel et al., 2021). Our results highlight the need for a greater consideration in the potential role of ocean forcing in both the current and future mass balance of the wider Dronning Maud Land region.

Data Availability

Landsat and ARGON imagery was provided free of charge by the US Geological Survey Earth Resources Observation Science Center (https://earthexplorer.usgs.gov/). For the MODIS imagery we also acknowledge the use of imagery from the NASA Worldview application (https://worldview.earthdata.nasa.gov), part of the NASA Earth Observing System Data and Information System (EOSDIS). Cosi-corr is an ENVI plug-in and can be downloaded from http://www.tectonics.caltech.edu/slip_history/spot_coseis/download_software.html. The ITS_LIVE velocity products are available from https://doi.org/10.5067/IMR9D3PEI28U. GoLIVE velocity products are avaialble from http://dx.doi.org/10.7265/N5ZP442B. ERA5 data is available from https://doi.org/10.24381/cds.adbb2d47. Wind data from Syowa station is available via the SCAR READER at http://dx.doi.org/10.5285/569d53fb-9b90-47a6-b3ca-26306e696706. The MOA grounding line product is avaialble at https://doi.org/10.7265/N5KP8037. BedMachine is available at https://doi.org/10.5067/E1QL9HFQ7A8M. The ice shelf thickness change dataset from Smith et al. (2020) is available at http://hdl.handle.net/1773/45388. REMA DEM strips are avialble at https://www.pgc.umn.edu/data/rema/. Lützow-Holm Bay bathymetry is available at https://doi.org/10.17632/z6w4xd6s3s.1. The RAMP mosaic is available at https://doi.org/10.5067/8AF4ZRPULS4H. Ice shelf extent buttressing dataset from Durand et al., 2016 is available at https://doi.org/10.5067/FWHORAYVZCE7. SPIRIT DEM's are available from https://theia.cnes.fr/atdistrib/rocket/#/search?collection=Spirit

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