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Key Points:

- Feedback between reduced winter sea ice and summer hydrography
- Low stratification persists into summer after winter mixing and cooling
- Increased summer heat uptake in low ice years exceeds extra winter cooling

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Feedbacks between ice cover, ocean stratification, and heat content in Ryder Bay, western Antarctic Peninsula

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Abstract A multiyear, all-season time series of water column physical properties and sea ice conditions in Ryder Bay, at the western Antarctic Peninsula (WAP), is used to assess the effects on the ocean of varying ice cover. Reduced ice cover leads to increased mixing and heat loss in the winter. The reduction in stratification persists into the following summer, preconditioning the water column to a greater vertical extent of surface-driven mixing. This leads to an increased amount of heat from insolation being mixed down, affecting approximately the top 100 m. The increased heat uptake in summer exceeds the heat lost the preceding winter, giving the initially counter-intuitive effect that enhanced winter cooling generates warmer temperatures in the following summer and autumn. This process is therefore a positive feedback on sea ice, as reduced sea ice leads to increased heat content in the ocean the following autumn. It also causes increased winter atmospheric temperatures due to the increased winter heat loss from the ocean. In the deeper part of the water column, heat and carbon stored in the Circumpolar Deep Water (CDW) layer are released by deep mixing events. At these depths, conditions are restored by advection and vertical mixing on multiyear time scales. In recent years, stronger deep mixing events in winter have led to a persistent reduction in CDW temperatures at the study site. Ocean glider data demonstrate the representativeness of these results across the wider region of Marguerite Bay, within which Ryder Bay is situated.

1. Introduction

Sea ice duration along the western Antarctic Peninsula (WAP) has reduced significantly in the last few decades, mostly through a later advance in autumn [*Stammerjohn et al.*, 2008a]. This reduction is likely due to changes in wind patterns that have also caused increases in ice cover elsewhere around Antarctica [*Holland and Kwok*, 2012]. The changes in the local wind forcing have been linked to the Southern Annular Mode (SAM), El Niño-Southern Oscillation (ENSO), and to the Atlantic Multidecadal Oscillation [*Marshall*, 2003; *Stammerjohn et al.*, 2008b; *Turner et al.*, 2009; *Li et al.*, 2014].

The intensification of the Southern Annular Mode, the dominant mode of climate variability in the extratropical Southern Hemisphere, has caused an increase in northerly winds in the region of the Antarctic Peninsula [*Marshall*, 2003]. This is a nonzonally symmetric component of the otherwise mostly zonal structure of the SAM as it overlies the Southern Ocean [*Lefebvre et al.*, 2004; *Thompson and Solomon*, 2002]. The El Niño-Southern Oscillation phenomenon also exerts a marked influence on this region [*Meredith et al.*, 2004; *Stammerjohn et al.*, 2008b]. While meridional winds at the WAP are known to be sensitive to the phasing of ENSO, the extent to which recent trends in the WAP region are ascribable to ENSO-related variability is not yet determined.

The regional reduction is counter to the overall trend around Antarctica of a slight increase in sea ice [*Comiso and Nishio*, 2008]. At the Antarctic Peninsula, the ice declines significantly influence the ecosystem [*Ducklow et al.*, 2007] with the associated changes in hydrography leading to sensitive effects between ice and phytoplankton along the Antarctic margins [*Venables et al.*, 2013; *Williams et al.*, 2008]. Linked to the ice loss, atmospheric winter temperatures are increasing at $\approx 1^{\circ}C$ per decade [*Turner et al.*, 2005]. Ocean temperature increases have also been observed along the Antarctic Peninsula in the top 100 m in summer [*Meredith and King*, 2005]. In addition, over 80% of glaciers on the Antarctic Peninsula are retreating [*Cook et al.*, 2005], and there has been increased precipitation at the WAP in recent decades [*Thomas et al.*, 2008].

The oceanography of the WAP continental shelf is important for the regional cryosphere (sea ice, floating ice shelves, and glaciers grounded beneath sea level) as warm Circumpolar Deep Water (CDW) floods onto the shelf at depths below 200 m delivering large quantities of heat to the coastal environment [*Martinson and McKee*, 2012; *Moffat et al*., 2009]. The proximal source of this water is the southern edge of the Antarctic Circumpolar Current (ACC), which flows close to the shelf break, but it originates in North Atlantic deep water formation and hence is old and has high concentrations of dissolved carbon [*Key et al*., 2004]. In contrast to other Antarctic shelf regions, the flooding of the WAP shelf by comparatively unmodified CDW means that the deep water is relatively warm at 1–2°C (3–4°C above in situ melting temperatures). Surface waters are fresher than CDW and colder in winter when mixing reaches 50–150 m. In spring and summer, the melting of both sea ice and glacial ice [*Meredith et al*., 2010] combine with surface warming to produce a warm and very fresh surface layer above the remnant winter water (WW). Mixing with thermocline waters modifies CDW characteristics as it flows toward the coast, reducing the salinity and temperature [*Martinson et al*., 2008]. Modeling studies have shown that increased wind stress leads to changes in the lateral flow of CDW onto the shelf and greater heat loss from CDW due to vertical mixing [*Dinniman et al*., 2012].

There are many potential feedback processes between atmospheric, oceanographic, and cryospheric processes at the WAP, and these may have differing time scales of operation and hence different temporal impacts on the water column. Sea ice acts as a mechanical and thermal barrier and thus reduces surface mixing in winter [*Hyatt et al*., 2011]. A loss or reduction in sea ice cover can lead to polynya-type environments where convection driven by brine rejection may increase or decrease, depending on whether the exposed surface water continues to freeze, with the additional ice advected away [*Comiso et al*., 2011]. Increased mixing releases heat from the ocean, either taken up the previous summer into the upper layers or stored in old CDW [*Goosse and Zunz*, 2014]. Similarly, in spring the presence of sea ice has a cooling effect on the ocean through increased albedo [*Perovich et al*., 2007] leading to reduced ocean heat uptake [*Screen and Simmonds*, 2010]. The uptake of latent heat for melting the sea ice, if it melts in situ, further reduces the sensible heat. The complexity generated by these interlinked and spatially and temporally variable processes leads to significant unknowns concerning how the marine system will evolve as climatic change progresses.

This study reports on 15 years of continuous oceanographic sampling from the British Antarctic Survey base at Rothera, just inside the Antarctic Circle on the western Antarctic Peninsula [*Clarke et al*., 2008; *Meredith et al*., 2010; *Venables et al*., 2013]. We present year-round water column data and analyze the effects of changing sea ice cover through the course of the time series. Complementing previous studies of feedback effects from reductions in Arctic sea ice extent in summer [*Screen and Simmonds*, 2010], we present evidence of a positive feedback loop from decreased winter sea ice duration in Antarctica that starts with increased ocean mixing and cooling in winter that then allows increased heat uptake the following summer. The sampling site within Ryder Bay is sheltered and close to glacial sources of freshwater, so the representativeness of the local oceanographic in the broader-scale context of northern/central Marguerite Bay is established using data from autonomous ocean gliders.

2. Data and Methods

Water column sampling has been carried out at Rothera Research Station (Figure 1) since 1998 [*Venables et al*., 2013], as part of the Rothera oceanographic and biological time series (RaTS). The primary RaTS site (67.570°S, 68.225°W) is 520 m deep (Figure 1) and located at the deepest point of Ryder Bay, about 4 km from the base. When the standard RaTS site cannot be occupied (e.g., due to partial sea ice cover), a secondary site, 300 m deep, is used (67.581°S, 68.156°W).

Sampling is undertaken from a small boat when conditions allow, or through a hole cut in fast ice. Sampling is targeted to be twice weekly in summer and as often as weather and ice conditions allow in winter (ideally weekly). There is a gap in late 2000, caused by heavy pack and brash ice that did not allow passage by boat or sledge. This problem is also responsible for a number of other smaller gaps in the time series, mostly in early spring. A second major gap occurred after equipment was lost in a fire in September 2001. Any gaps of greater than 30 days are treated as breaks in the time series.

A RaTS sampling event consists of a full-depth vertical profile made using a SeaBird 19+ conductivity-temperature-depth instrument (CTD), a WetLabs in-line fluorometer, and a LiCor photosynthetically available radiation (PAR) sensor. Prior to January 2003, a Chelsea Instruments Aquapak CTD and fluorometer were used, limited to 200 m by the pressure rating of the instrument. An annual joint cast is also made with

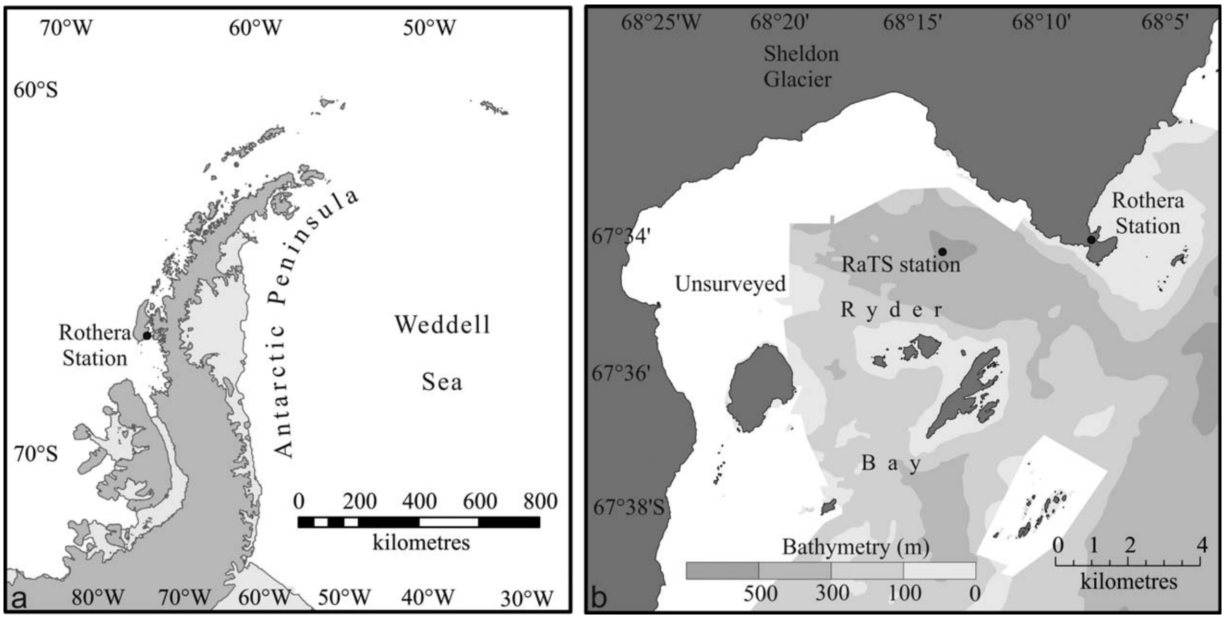

Figure 1. Maps of the location of the sampling location on the western Antarctic Peninsula.

the RaTS profiling system and the SeaBird 911+ CTD package on the R/V *Laurence M. Gould* during the Palmer Long Term Ecological Research (LTER) grid survey [*Ducklow et al.*, 2007]. After applying the offsets identified, we estimate that the temperature is accurate to 0.002°C and salinity to 0.005.

Data were also collected from Marguerite Bay using two deep (1000 m) Teledyne Webb Slocum G2 gliders deployed and recovered at Rothera. The sensors included a pumped CTD and the temperature and salinity values from concurrent deep profiles in Ryder Bay agreed between the RaTS and glider CTDs to within the uncertainty of the RaTS data. Comparisons between glider up and subsequent down casts showed no significant net offset.

Mixed layer depth (MLD) is calculated as the depth at which density is 0.05 kg m^{-3} greater than at surface. The choice of a density criterion is inherently arbitrary and the method necessarily includes a small region of stratified water. It is however more practical to use a density criterion than to consider just the depth of no stratification (the "mixing layer"). Such conditions are short-lived and likely to be missed by subsampling, whereas a small density-difference criterion records the effect of recent mixing events and so is a more robust measure. Sensitivity tests with varying density criteria lead to no qualitative changes in the conclusions.

Stratification is quantified here as the additional potential energy that would need adding to homogenize the water column from the surface to a given depth. This is the negative of the potential energy anomaly [*Simpson et al.*, 1978]. This is used in preference to the buoyancy (Brunt-Väisälä) frequency N^2 as it provides a bulk measurement between two depths and the units (Joules) enable comparisons with mechanical energy input, though these are only conducted qualitatively in this instance.

Sea ice coverage and type is assessed each day in the three bays around the Rothera station by direct observation. The necessary changeover in base personnel leads to some internal heterogeneity in this somewhat subjective measurement. However, the seasonal and interannual changes in sea ice coverage are far greater than the interobserver differences. Sea ice coverage has been previously discussed in relation to assessing freshwater inputs into Ryder Bay [*Meredith et al.*, 2008; *Meredith et al.*, 2010].

The effects of advection past the sampling site and vertical motions associated with internal waves add to short-term variability. This effect is an inevitable feature of all single-point time series sites, but the length of the series means that the positive and negative anomalies thus generated are thought to largely compensate.

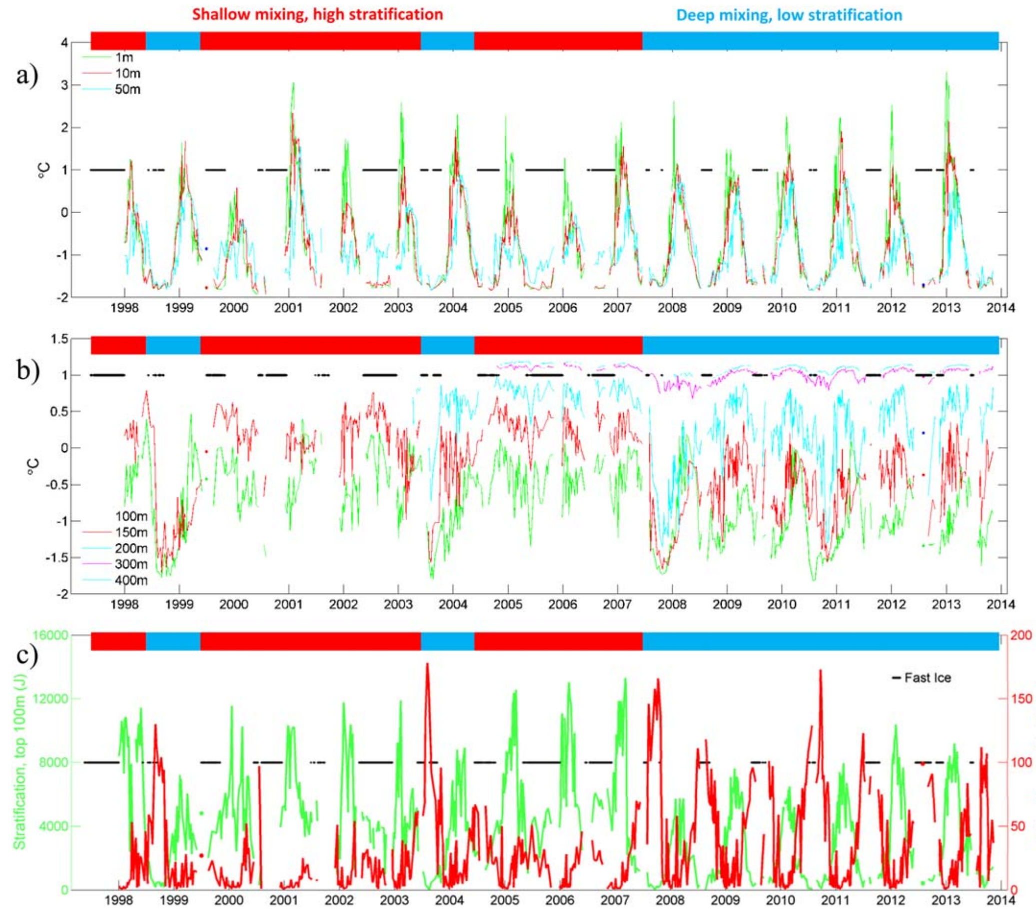

Figure 2. (a, b) Time series of temperature at selected depths. (c) Stratification in the top100 m and mixed layer depth. Daily presence of fast ice is marked on each plot by black dots.

3. Results

3.1. Interannual Variability of Temperature

The full time series of temperature measurements at RaTS is shown in Figures 2a and 2b. It is clear that there is significant interannual variability through the time series, including deep reaching cooling events in some years (such as 1998, 2003, and 2007), particularly warm (2001, 2013) and cold (2000, 2006) sea surface temperatures and varying amount of heat retained through winter at 50 m. The deep cooling events correspond to the formation of deep mixed layers in some winters (Figure 2c) as these bring deeper water into contact with the surface where heat is lost rapidly to the atmosphere.

Above the depth of the summertime temperature minimum that marks the WW (around 100 m), heat loss in winter is approximately balanced by surface warming the following summer. Below the temperature minimum, the processes that can increase temperatures at a given depth are upward mixing of heat from below and horizontal advection of water that has been modified differently elsewhere. These are slow processes, taking over a year to replenish the heat content in the water column that is lost during deep mixing events (Figure 2b); this is most evident after the deep mixing in 2007. The multiyear timescale of the recovery means that if very deep mixing happens frequently (two successive winters), the latter releases of heat will be smaller than the initial occurrence.

3.2. Ice and Stratification

As has been discussed previously [*Venables et al*., 2013], the variation in ice cover in Ryder Bay strongly influences winter MLD (Figure 3a). This is as expected due to reduced sea ice, especially fast ice, exposing the sea surface to increased mechanical mixing from wind stress. Reduced sea ice cover is caused locally in Ryder Bay by stronger and/or more persistent northerly winds blowing ice out of the south facing bay

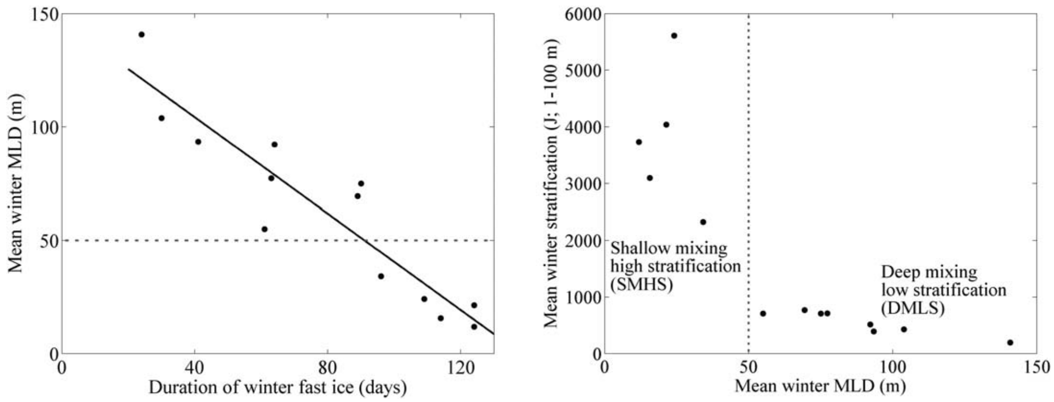

Figure 3. (a) Fast ice duration against mean winter mixed layer depth for each year of the RaTS time series. (b) Mean winter stratification against mean winter MLD. The dashed lines separate deep and shallow mixing years.

[*Meredith et al*., 2010]. The area can therefore exhibit some characteristics of a polynya [*Massom et al*., 2006], with reduced ice cover compared with the wider Marguerite Bay.

Near-surface stratification increases in the spring and summer of all years through glacial freshwater discharge and in situ melting of sea ice and icebergs [*Meredith et al*., 2008]. Reduced fast ice leads to deeper mixed layers in winter (Figure 3a). This in turn relates to notably different amounts of stratification retained in the top 100 m (Figure 3b). Lower winter stratification then leads to lower summer stratification in the top 100 m (Figure 4), preconditioning the surface layers to greater vertical mixing for the same energy input from wind stress and tides. The gradient of the relationship between summer and winter stratification is 0.8 ± 0.5; thus, the positive relationship is significant but the gradient is not significantly different from unity, which would represent all winter variation following through to the summer.

3.3. Annual Cycles of Temperature and Salinity

To analyze the time series in more detail, it has been partitioned into years with low ice, deep mixing and low stratification (DMLS) years, and high ice, shallow mixing and high stratification (SMHS) years. The actual criterion used is a mean winter MLD greater than 50 m for DMLS years (Figure 3b). As the deep winter mixed layers are due to significantly reduced sea ice, there are other criteria based on different variables that lead to the same partitioning. One partial exception to this is 2004, which had a very deep mixed layer in midwinter, but fast ice had returned by early spring. This year is included in the DMLS years, but the results are qualitatively unchanged if it is included with SMHS years.

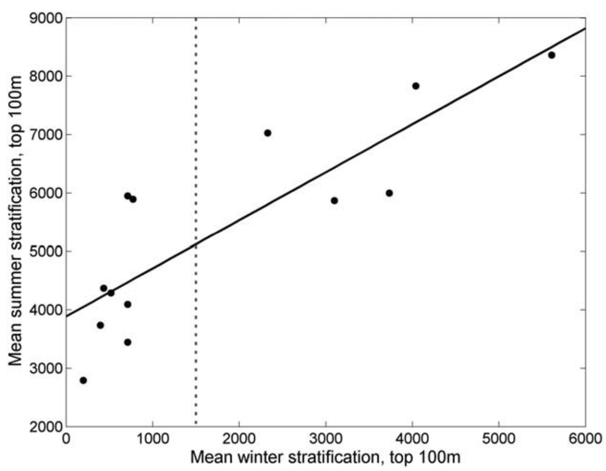

Figure 4. Winter stratification against summer stratification, with best fit linear regression ($r^2 = 0.71$). The dashed line separates years grouped as deep winter mixing (DMLS) and shallow winter mixing (SMHS).

Figure 5 shows annual cycles for temperature and salinity for SMHS and DMLS. The 7 and 8 years, respectively, of data have been averaged to create the annual cycles and then smoothed with a 30 day running mean. The different depths of sampling over the time series lead to a slight discontinuity at about 190 m. Starting in late autumn (about day of year 120, which is when the data are split at the boundaries between SMHS and DMLS periods), the differences between the modes is clear. In DMLS years, the cold surface layer extends much deeper through the winter, with the salinity largely mixed in those depths by late winter. This is followed in spring and summer by heat uptake extending to greater depth. The temporal progression of properties at selected depths is

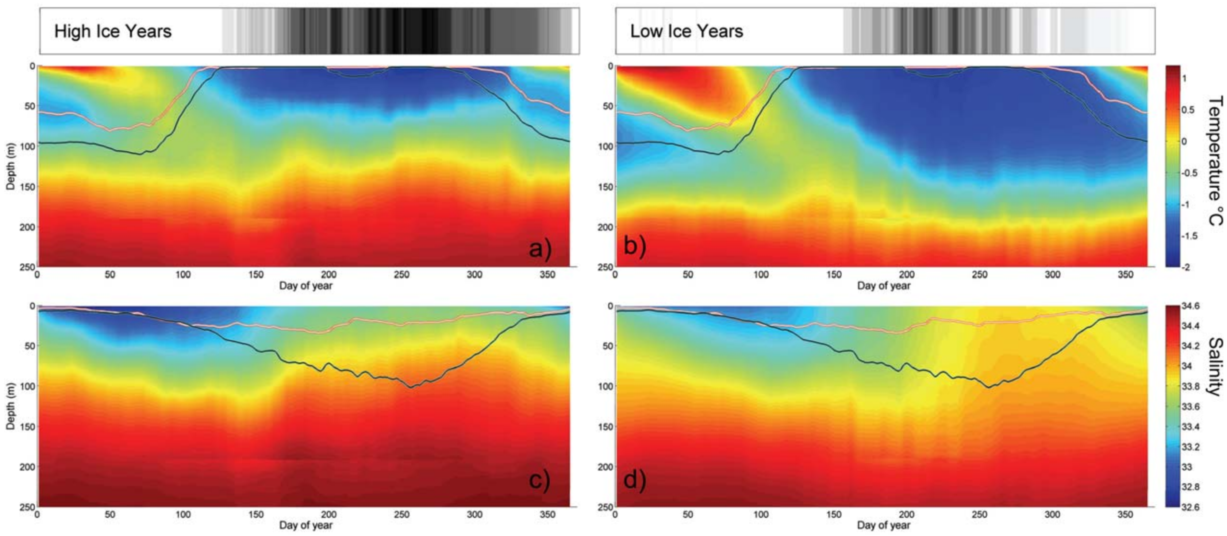

Figure 5. Time/depth plots for potential temperature and salinity. (a) Average temperature in SMHS years, (b) average temperature in DMLS years. The average annual cycle of temperature minima are plotted for DMLS years (blue/black) and SMHS years (red/white). (c) Average salinity in SMHS years, (d) average salinity in DMLS years. Mixed layer depths for DMLS year (blue/black) and SMHS years (red/white) plotted. The average ice cover in Ryder Bay, excluding brash ice, is plotted at the top.

discussed next in order to identify the processes that cause DMLS and SMHS years to differ in the manner observed.

3.3.1. Temperature and Salinity at 150–250 m

Very deep mixing events result in a significant loss of heat from 200 m, with slight effects at 300 m and deeper (Figure 2b). Below the depth of the autumn temperature minimum (Figure 5b) heat cannot be replaced by surface heating. Instead there is a slower process of replenishment by advection of less strongly modified CDW. This process can take over a year (Figure 2b), which, combined with typical advective speeds at the WAP (typically several cm s^{-1} up to a few tens of cm s^{-1} [*Savidge and Amft*, 2009]), indicates that the deep mixing occurred over a wide area (tens to hundreds of kilometres). The deep salinities are reduced slightly, due to the increased mixing with surface waters.

3.3.2. Temperature and Salinity at 100 m

In SMHS years the extensive fast ice cover limits mechanical mixing, and there is no seasonality in the temperature at 100 m (Figure 6a). There is a slight increase in salinity as ice is formed and brine is rejected, but this does not lead to deep convective mixing (Figure 6a). In DMLS years, the reduced winter ice extent allows greater wind stress to act on the water surface, leading to deep mixing. This releases heat from 100 m to the atmosphere. The deep mixing also reduces the near-surface stratification (Figure 6c) with salinity at 100 m being similar to that at the surface (Figure 6d). The lower stratification persists into the spring and summer, which allows more heat to be mixed downward the following summer. This leads to the lost heat at 100 m being replaced, with parity between SMHS and DMLS temperatures being restored in late autumn.

3.3.3. Temperature and Salinity at 50 m

The pattern at 50 m is similar to 100 m (Figures 6a and 6b). There is some seasonality in SMHS years, though temperatures only start to increase in autumn once surface temperatures have peaked. In DMLS years, with mixed layer depths typically >50 m, the winter temperature at this depth is just above freezing point whereas in SMHS years the average temperature remains at about $-1.2°C$. The reduced stratification leads to temperatures increasing gradually from spring through autumn, starting at the same time as the increase in surface temperatures. The peak in the average temperature is significantly above that in SMHS years, despite the lower winter temperatures. The amplitude of the seasonality is therefore increased, together with a qualitative change in the annual cycle.

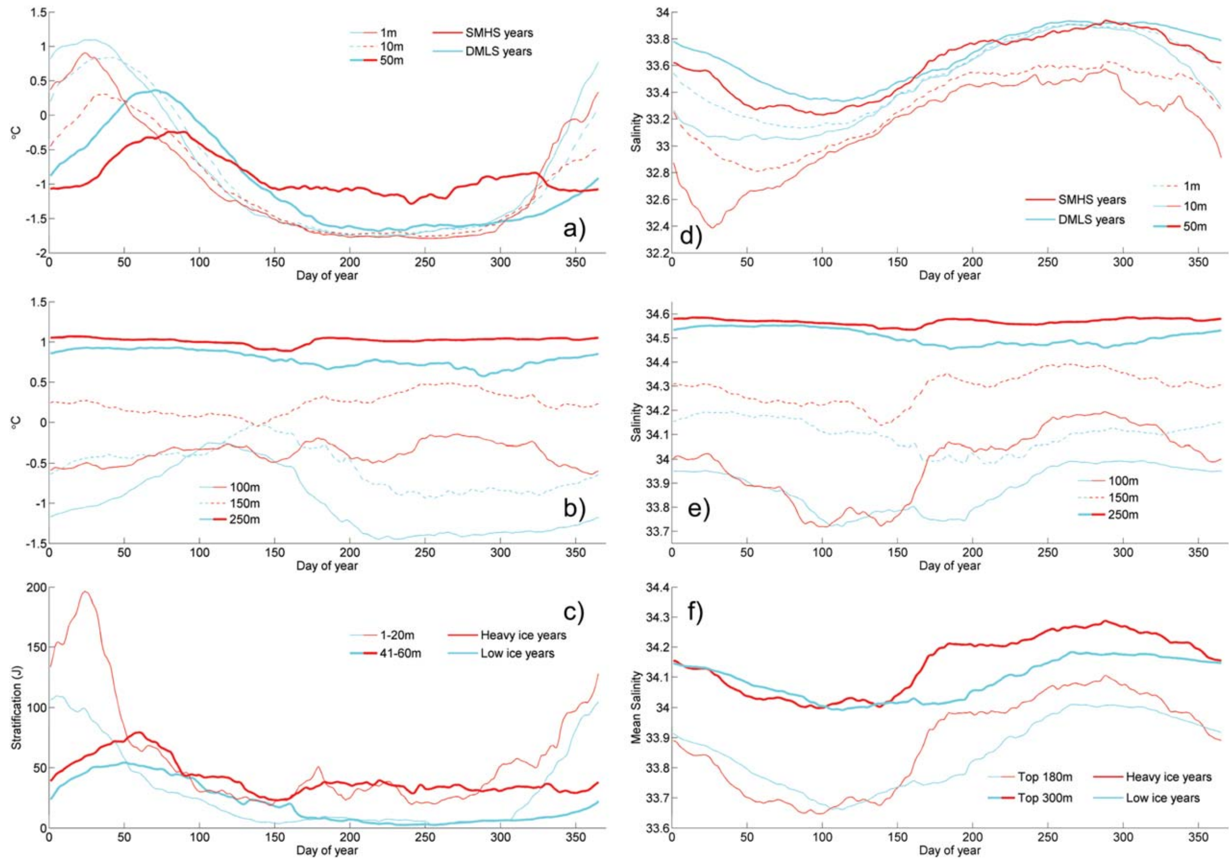

Figure 6. (a, b) Annual cycles of potential temperature, (c) stratification over 20 m, centered on stated depth, (d, e) salinity at selected depths, and (f) depth averaged salinity.

3.3.4. Temperature and Salinity at 1 and 10 m

Near-surface temperatures reduce to freezing point in winter as ice forms in both DMLS and SMHS regimes. They then increase in a similar manner, despite differences in albedo and latent heat fluxes in years with differing amount of ice. The reasons for this are that the greater heat uptake associated with reduced ice cover is spread over a greater depth and that in SMHS years the upward sloping isotherms (Figure 5a) indicate that some heat that is retained at relatively shallow depths (30–60 m) is mixed to the surface as ice retreats. The increased stratification in SMHS years leads to a greater temperature gradient between 1 and 10 m, which will contribute to the more rapid decrease in temperatures in February.

Freshwater is added to the surface, from melting of sea ice and glacial ice, in both regimes but the higher winter surface salinity in DMLS years persists to higher near-surface salinity in summer. Similarly, the near-surface stratification remains lower in the spring and summer following deep winter mixing (Figure 6c).

3.3.5. Regional Representativeness of Ryder Bay

As Ryder Bay is relatively enclosed and close to glacial sources of freshwater, the buoyancy and mixing there could be significantly different from more open water areas further from land. To assess this, glider data from Marguerite Bay are presented (Figure 7a). The glider track reaches the mouth of Marguerite Bay, which is 60 km from the nearest land. It is therefore significantly more remote from coastal influences than Ryder Bay, both in terms of freshwater inputs and shelter from wind mixing, though it is still marginally more sheltered than the fully open regions closer to the shelf edge.

Figure 7 shows the temperature, salinity, and stratification values in Marguerite Bay, as measured by ocean gliders in 2013 and 2014, together with the corresponding values from RaTS for the same periods. The values are generally similar for all parameters and depths. Local and short-term processes such as internal

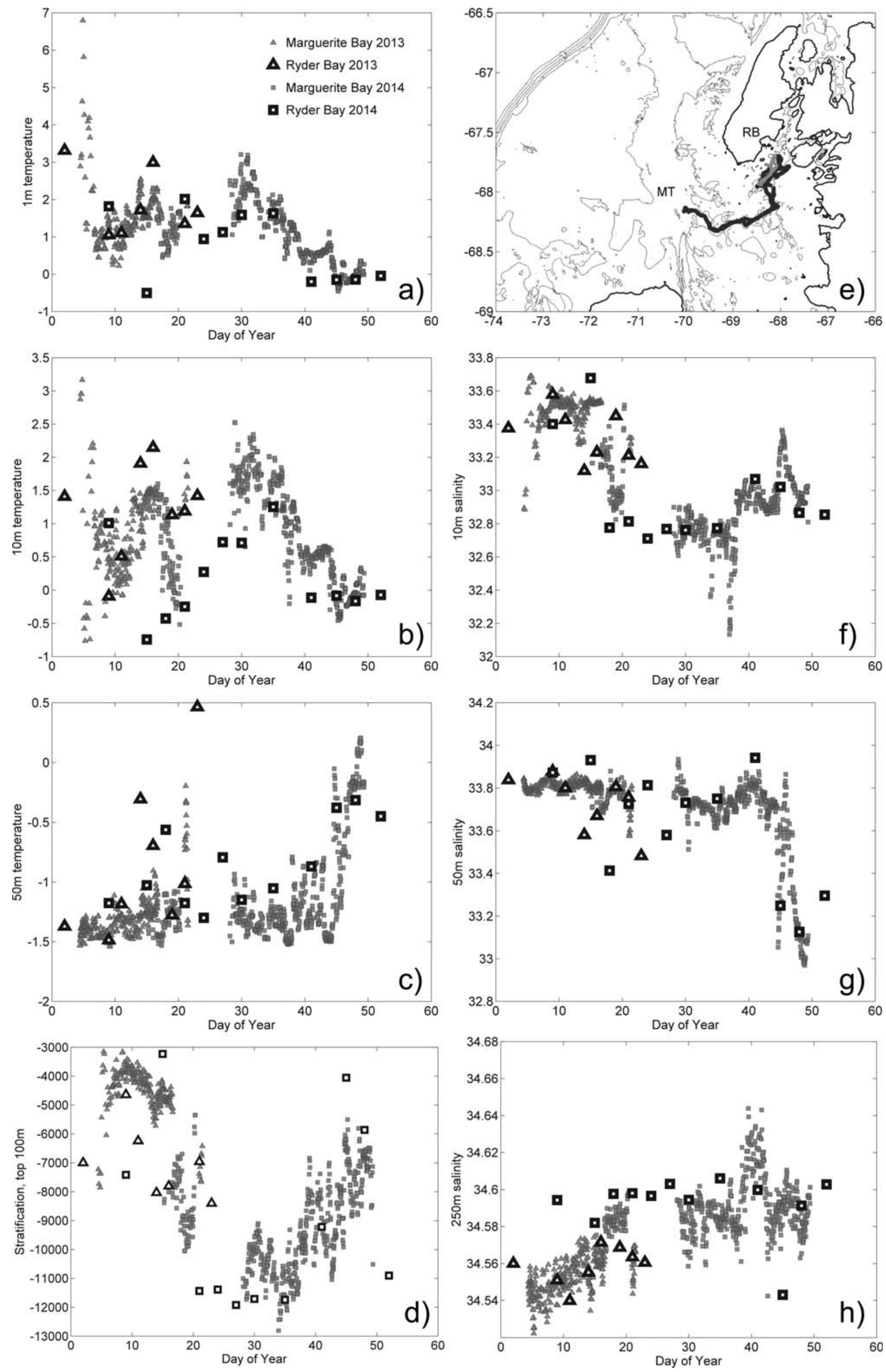

Figure 7. (a–c) Comparisons between Marguerite Bay and Ryder Bay temperature, (d) stratification, and (f–h) salinity. (e) A map showing the glider tracks in 2013 (pale gray) and 2014 (dark gray). In 2014, the glider reached the furthest point from Ryder Bay on day of year 40.

waves, glacial runoff, iceberg melting, and spatially varying weather lead to some inevitable differences, but there are no systematic offsets. Temporal changes are also broadly similar between Marguerite and Ryder Bay during these periods.

For shallow depths this strongly suggests that the same overall processes control the temperature and stratification and that the magnitude of these processes is broadly similar between Ryder Bay and Marguerite Bay. Surface drifters deployed in Ryder Bay (unpublished data) show variable and occasionally strong currents flowing between Ryder Bay and the adjacent part of Marguerite Bay.

At depth, the similar salinity values at 250 m (Figure 7h) show a connection between CDW in Marguerite and Ryder Bay for all depths where significant temperature changes occur. Below about 300 m, there are slight differences in temperature and salinity, linked to the blocking of deeper waters by the 350 m deep sill at the mouth of Ryder Bay.

3.4. Seasonal Water Column Structure Changes

The temperature changes between DMLS (low ice) and SMHS (heavy ice) years for the full water column are shown in Figure 8a. The step in the profiles just shallower than 200 m is a result of the shallow sampling depth early in the time series. The winter period is here taken to be calendar days 200–300, summer to be 335–50, and autumn to be 51–150. The full-year trend is derived using monthly averages to reduce problems from differing sampling density through the year.

The annually averaged temperature difference shows that reduced ice cover leads to warming of up to 0.2°C in the upper 40 m. Below this there is cooling at all depths, peaking at 0.8°C at 140 m. This represents an increased ventilation of CDW, which releases heat to the atmosphere, with the potential to influence sea ice processes at the time of the mixing. There is considerable variation around this overall structure in each season. In winter there is very limited warming near the surface, associated with the reduced ice cover, and a large cooling at depth, peaking at just above 1.2°C. In summer there is warming in the top 60 m, of up to 0.5°C. The deepest extent of warming is in autumn as the increased heat content from summer is mixed downward. Warming extends to 90 m in autumn but is reduced in magnitude relative to summer in the top 50 m. At 90 m the autumn warming has counteracted a winter reduction in temperature of 1°C. There is much smaller cooling below 300 m down to the seabed, with little variability between seasons.

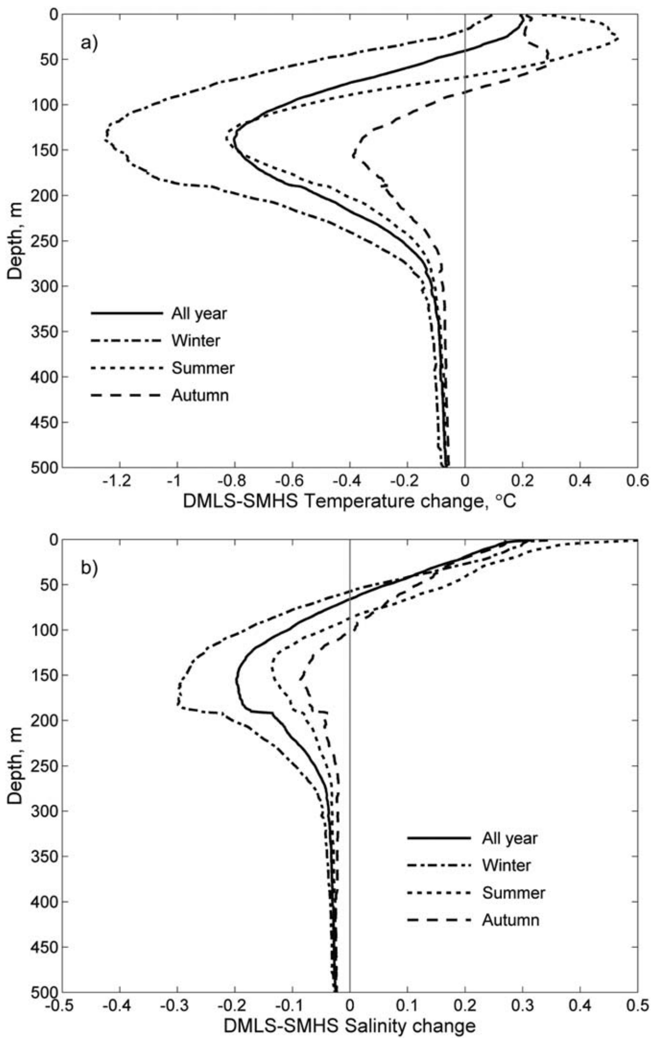

Figure 8. (a) Temperature and (b) salinity changes between deep winter mixing (DMLS) and shallow winter mixing (SMHS) years by depth and season.

The salinity changes between DMLS (low ice) and SMHS (heavy ice) years for the full water column are shown in Figure 8b. The possible causes for deep mixing after reduced ice cover are increased mechanical mixing, from increased wind stress at the surface, and convection following brine rejection from increased sea ice production. The reduction in salinity, below 55 m in winter, approximately balances the increase in salinity shallower than 55 m. Due

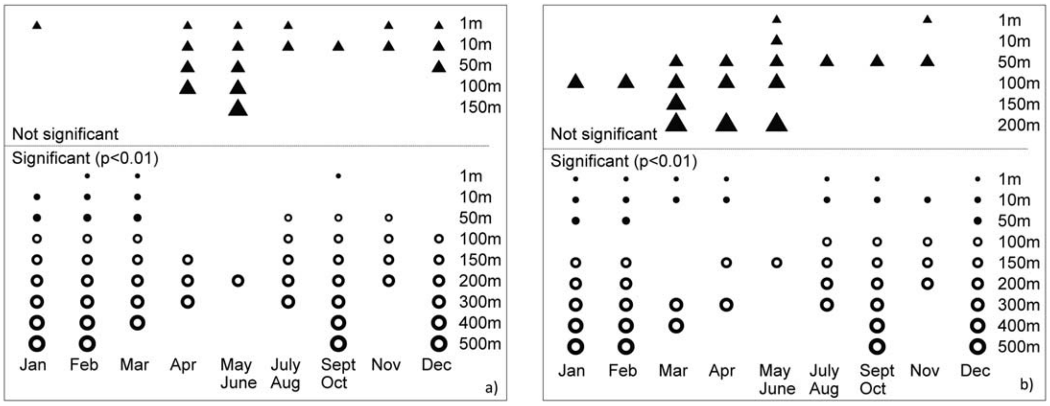

Figure 9. Plot showing when changes in (a) temperature and (b) salinity are statistically significant. Closed circles represent values higher in low ice, DMLS years.

to the dominance of salinity in setting density at low temperatures, the equivalent plot of density (not shown) is qualitatively similar, with a reduction of density below about 50 m. Mechanical mixing must therefore be the dominant process responsible for the mixing, as brine rejection cannot lead to the lower salinity and density observed at depth. This is in contrast to many polynya environments [*Gordon and Comiso*, 1988] where increased sea ice formation and brine rejection lead to an increase in salinity and density at all depths, such as the Weddell Sea [*Arthun and Nicholls*, 2013]. To demonstrate the level of statistical significance of these changes, the distributions of temperature and salinity were compared between SMHS and DMLS for each month, or pairs of months when sample sizes were small (Figure 9). If either DMLS or SMHS years had four or fewer samples in a period then it is left blank. Most differences are significant at the $p < 0.01$ level. Temperature changes that are not statistically significant occur when increased mixing pumps heat to greater depth, replacing the heat lost to winter mixing. At 50 m, this occurs in December before temperatures in DMLS years exceed SMHS years from January to March. At greater depths the increased mixing in autumn replaces heat lost to deep mixing the preceding winter. Shallow winter temperatures are close to freezing in both SMHS and DMLS years in winter, and hence are often not significantly different, though reduced ice cover allows surface temperatures to increase above the freezing point, leading to differences in September/October. They also increase at similar rates despite different ice conditions, as discussed in section 3.3.4. The significant differences in autumn are likely due to increased release of heat from below, maintaining surface temperatures further through autumn in DMLS years. For salinity there is less seasonal structure, with most nonsignificant changes occurring at 50 and 100 m, which is the boundary between increased salinity at shallower depths and decreased salinity at greater depths due to the increased mechanical mixing in DMLS years (Figure 8b).

3.5. Heat Content Changes

The different seasonal cycles described above show that the winter reduction in ocean heat content in years with low ice coverage is followed by an increase in the summer. This is due to both the reduced albedo from lower spring ice coverage and the reduced stratification that allows more heat to be mixed downward for the same mechanical energy input. Figure 10 shows that this increased heating in the top 100 m is strongly related to the winter temperature ($p < 0.001$). The gradient of this relationship is less than -1 ($p < 0.02$), indicating that the increased summertime warming is greater than the winter cooling that precedes it.

Latent heat can be estimated from the seasonal cycle of salinity in the upper water column. Figure 6f shows this cycle, over the upper 180 and 300 m. These two depths are used here as sampling was limited to about 190 m in the early part of the series, so the former enables the full series to be utilized. Most of these early years had shallow mixing with little seasonality in salinity below 150 m (Figure 2b); hence, 180 m is sufficient to capture the salinity changes. In years with deep mixing, seasonality extends to greater depths; thus, the 300 m level is more informative for these. Using shallower depths would increase the estimated ice production as salinity mixed upward from depth would be included as if it were from brine rejection. Later, deeper sampling covers most DMLS years (Figure 2b).

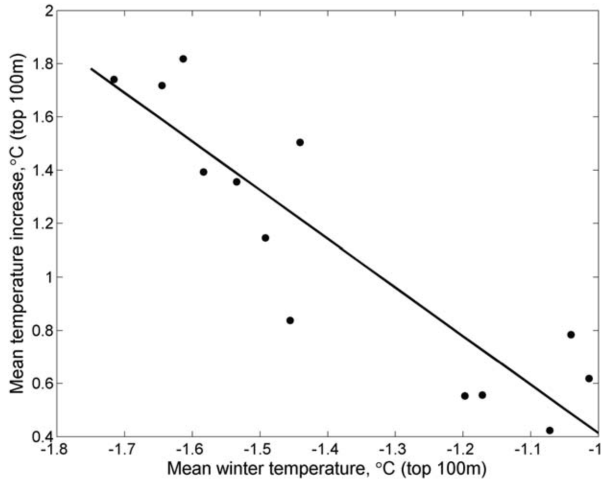

Figure 10. Top 100 m average temperature summer increase against winter ($r^2 = 0.81$).

Assuming a sea ice salinity of 6 [*Gough et al.*, 2012], the salinity changes can be converted to a depth of net ice production, assuming no net effect from advection. These estimates are 2.9 m for SMHS (high ice) years and 1.9 m for DMLS (low ice) years. Using a sea ice salinity higher or lower by 3 leads to an approximately 10% increase or decrease, respectively, in the ice production estimate. Converting again to latent heat release leads to heat fluxes of 82 W m^{-2} and 53 W m^{-2} on average over SMHS and DMLS years, respectively (mean over calendar days 140–280). For DMLS years the rate of salinity increase is approximately even in this period, but in SMHS years it occurs mostly in the first 45 days, during which the associated latent heat released is 190 W m^{-2}.

Figure 11 shows the heat fluxes (changes in ocean sensible heat content) between casts, again split between SMHS and DMLS years and smoothed with a 30 day running mean. Changes in ocean heat content could, in principle, be due to horizontal advection through the sampling site of different water masses, though without a net source or sink of heat nearby, the anomalies thus generated would be likely to average out. The cycles shown are therefore likely to approximate well the heat fluxes between ocean and atmosphere in the region of the time series site [*Smith and Klinck*, 2002]. Positive values represent a gain of heat by the ocean.

Heat loss to the atmosphere is greater in winter in DMLS years (February to August mean 33 W m^{-2}) than SMHS years (mean 24 W m^{-2}), linked to both the greater seasonality in heat content above the temperature minimum and the release of CDW-stored heat from very deep mixing events. The release of heat from shallow depths is gradual through the autumn and early winter but releases from depths greater than 100 m are more episodic and only occur during winter when there are deep mixing events. There is relatively little heat released from below 200 m: Figure 2b shows that reductions in temperature below this depth are much reduced compared to shallower depths. The heat uptake is higher in low ice years throughout the summer, extending well into the period when sea ice is absent even in high ice seasons. This suggests that the cooling effects of ice, through albedo and latent heat, do not dominate the difference in heat uptake between high and low ice years.

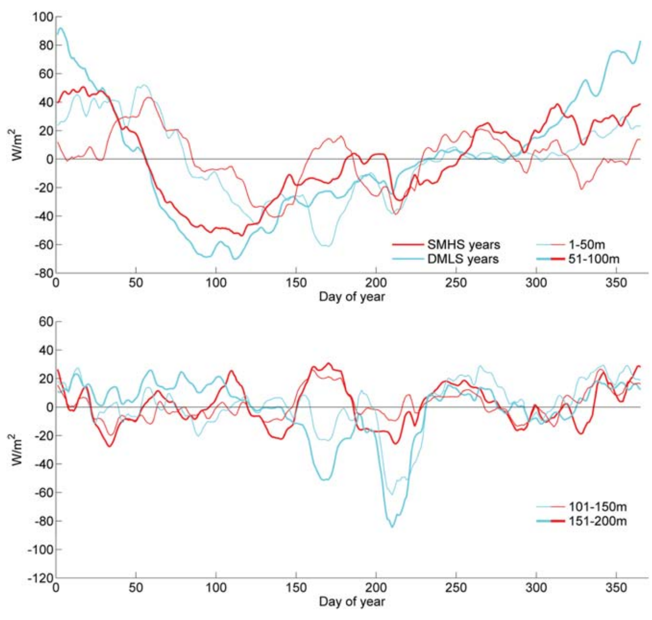

Figure 11. Change in heat content between successive casts, split between heavy ice (shallow winter mixing) years, and low winter ice (deep winter mixing) years.

Overall, the extra heat released from shallower than 100 m, which is replenished each year, is approximately the same as the heat released from CDW below 100 m, though the deep release is more concentrated in short-lived deep mixing events, which are stronger in some DMLS years than others (Figure 2b). If these subannual events were to become more frequent, the multiyear timescales of temperature recovery that follow them would result in less heat being available for release. The reduced latent heat release in low ice DMLS years is about a quarter of the increased sensible heat released from the ocean.

4. Discussion

The processes and changes described here for Ryder Bay give additional insight into the feedbacks between ice cover and ocean heat content, both in the surface layer and in deeper layers. This can allow greater understanding of polar areas that are less heavily sampled, especially in winter; however, it is noted that magnitude of the feedbacks and relative importance of this process will vary in different locations. The conditions in Ryder Bay are shown to be similar to those found in Marguerite Bay, including areas distant from glacial inputs of freshwater; hence, they cannot be determined dominantly by local/coastal processes. This wider area of Marguerite Bay is exposed to much greater fetches so could potentially experience different levels of wind-driven mixing.

4.1. Feedbacks on Ice Cover

Years that exhibit a reduction in winter fast ice are subject to deeper winter mixing and a reduction in stratification the following summer. This lower stratification leads to increased uptake of heat in the top 100 m in summer, with this increase being greater than the original heat loss caused by the deep winter mixing. The reduced ice cover and associated oceanic heat loss in winter, due to the mixing of salinity persisting into the following summer, is therefore a positive feedback effect on ice loss the following sea ice season due to the increased ocean heat content in autumn. As additional heat is lost from the upper 100 m during periods of low sea ice, the process is also a positive feedback toward reduced ice within each winter as well as for the subsequent winter.

There are plausible physical reasons why there should be increased warming after the increased winter cooling, but there is no a priori reason why the magnitude of this warming should necessarily exceed the cooling. In general, this process may either be a positive feedback or lessen the negative feedback of winter cooling, and in other locations its impact may differ from that observed in Marguerite Bay. Overall, this process leads to increased seasonality in ocean temperatures, but also to increased heat loss to the atmosphere in winter and uptake in summer, and thus will contribute to reduced seasonality in atmospheric temperatures with increased air temperatures in the winter, as observed along the Antarctic Peninsula [*Turner et al*., 2013].

As with all positive feedbacks, it is likely that there are other processes and natural variability that can break the cycle, as conditions would otherwise progress rapidly toward a different equilibrium. In this case the cycle can be broken by one or more of: calm winters, likely leading to less mixing and prolonged sea ice cover; calm summers, when there is less energy to mix heat downward; cold summers, when there is less heat to be pumped down; increased summer freshwater inputs, leading to a greater increase in stratification in spring and summer and likely less heat mixed downward.

4.2. Implications for Observing Temperature Trends

The increased seasonality in the top 150 m temperature means that subsampled data, especially if collected only during summer and/or near the surface, may lead to misleading or incomplete interpretations of temperature changes associated with changes in ice cover. There is net annual heating to 40 m (summer increases outweigh winter decreases). At 40 m summer-only sampling would suggest that reduced ice cover leads to a temperature increase of 0.4°C and winter-only sampling would indicate a cooling of 0.3°C, whereas full-year sampling shows no net change, but increased seasonality. Between 40 and 90 m the summer/autumn peak is warmer in DMLS years but there is net cooling on an annual average basis.

There is an observed trend toward less extensive sea ice cover along the Antarctic Peninsula [*Stammerjohn et al*., 2012]. This could potentially lead to the processes observed in this time series being increasingly replicated across a wider area. Observing and interpreting near-surface temperature trends in the ocean as a consequence of this trend in ice cover therefore needs to be done with caution, depending on the seasonality and depth of the available subsurface temperature data. Increased winter mixing will, in general, lead to reduced stratification but the relative effects of this will depend on local features such as the amount of buoyancy added during spring and summer and the amount of wind and tide energy available for mixing in all seasons. It could therefore logically be expected that increased warming would occur, but the magnitude of this relative to winter cooling will vary with local conditions. It is therefore not possible to make quantitative predictions for other areas, but the mechanism itself cannot be excluded from consideration when seeking to identify the causes and nature of ocean sea/ice changes.

Below 100 m there is cooling throughout the year (peaking at 0.8°C at 140 m as an annual average, or 1.3°C in winter). Below about 150 m, the temperatures cycles last more than a year after very deep mixing in some years of very sparse sea ice. This means that the timing of sampling relative to deep mixing events is important on interannual as well as seasonal timescales.

4.3. CDW Ventilation

The increased mixing significantly increases the link between CDW and the surface. This releases heat and carbon [*Montes-Hugo et al.*, 2010] to the atmosphere. The deep mixing events are episodic and are then followed by a period of recovery toward previous temperatures. This recovery is gradual (up to 2 years) and can be stopped by another deep mixing event, so the time mean temperatures are reduced at all depths.

The release of heat is due to reduced sea ice and hence is a positive feedback on local sea ice conditions within each winter. The multiyear recovery in temperatures means that the strength of this process may reduce during periods of persistently low winter sea ice. It may also be a negative feedback on downstream locations where this heat would otherwise be released by mixing or melting ice [*Padman et al.*, 2012]. A net increase in ventilation of the old water mass associated with a reduction in ice cover will also increase the release of carbon to the atmosphere from the CDW layer.

5. Summary

We have identified a previously unrecognized feedback mechanism that operates on sea ice in the nearshore environment of Ryder Bay at the western Antarctic Peninsula. Here reduced ice cover leads to increased mixing and heat loss in winter, with the reduction in stratification created persisting until the following summer. This preconditions the water column to a greater vertical extent of surface-generated mixing, with the result that more heat is mixed down in summer than was lost the preceding winter. Consequently, it is found that increased winter cooling actually results in warmer ocean conditions the following summer and autumn, and hence is a positive feedback on reducing sea ice coverage. Locally, the trend toward reduced winter ice cover has led to stronger deep mixing events, with implications for the venting of heat and carbon from the deep CDW layer. Whilst the time series used to generate these findings is situated in a comparatively sheltered environment that is subject to significant glacial freshwater inputs, we have demonstrated the representativeness of the results across the broader region of northern and central Marguerite Bay using data from autonomous ocean gliders.

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