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Tropical forcing of Circumpolar Deep Water Inflow and outlet glacier thinning in the Amundsen Sea Embayment, West Antarctica

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ABSTRACT. Outlet glaciers draining the Antarctic Ice Sheet into the Amundsen Sea Embayment (ASE) have accelerated in recent decades, most likely as a result of increased melting of their ice shelf termini by warm Circumpolar Deep Water (CDW). An ocean model forced with climate reanalysis data shows that, beginning in the early 1990s, an increase in westerly wind stress near the continental shelf edge drove an increase in CDW inflow onto the shelf. The change in local wind stress occurred predominantly in fall and early winter, associated with anomalous high sea level pressure (SLP) to the north of the ASE and an increase in sea surface temperature in the central tropical Pacific. The SLP change is associated with geopotential height anomalies in the middle and upper troposphere, characteristic of a stationary Rossby wave response to tropical SST forcing, rather than with changes in the zonally symmetric circulation. Tropical Pacific warming similar to that of the 1990s occurred in the 1940s, and thus is a candidate for initiating the current period of ASE glacier retreat.

INTRODUCTION

Pine Island Glacier and Thwaites Glacier, the two largest of several fast-moving outlet glaciers that drain a large fraction of the West Antarctic Ice Sheet (WAIS) into the Amundsen Sea Embayment, have long been recognized as critical elements of WAIS dynamics (e.g. Lingle and Clarke, 1979). Hughes (1979; 1981) argued that it is these glaciers that make the WAIS most susceptible to large-scale collapse, which almost certainly occurred during some previous interglacial periods (Scherer and others, 1998; Pollard and DeConto, 2009; Naish and others, 2009). The inferred sensitivity of the Amundsen Sea Embayment (ASE) glaciers reflects their bed geometry, which deepens inland (Lythe and others, 2001), the small size of the floating ice shelves at their termini, and the direct exposure of the ice shelves to the influence of warm, Circumpolar Deep Water (CDW) (Jacobs and others, 1996; Nitsche and others, 2007). In the mid-1990s it was discovered that melt rates under the Pine Island Glacier ice shelf are two orders of magnitude greater than under the much larger ice shelves in the Ross and Weddell Seas (Jacobs and others, 1996). Subsequent studies using satellite imagery and interferometry revealed that the grounding lines of Pine Island Glacier as well as Smith and Thwaites glaciers had retreated recently (Rignot, 1998; 2001), that there was significant thinning well inland of the grounding lines (Wingham and others, 1998; Shepherd and others, 2002), and that glacier surface velocities were increasing (Joughin and others, 2003; Rignot, 2008).

Although the thinning of ASE glaciers could in part reflect changes in surface mass balance (i.e. changes in snowfall in the drainage areas), Shepherd and others (2002) showed that the magnitude of thinning is too large to be explained that way, and that the pattern of changes is consistent with the diffusive upstream migration of a force perturbation beginning at the grounding line (Schmelz and others, 2002; Payne and others, 2004; Thomas and others, 2004). This in turn is consistent with a response to the thinning of the ice shelves (Shepherd and others, 2004) resulting from an increase in submarine melt rates due to enhanced delivery of heat from warm CDW, as had been suggested much earlier (Jacobs and others, 1992; 1996). These ideas were validated in 2010 by direct observations made by an autonomous underwater vehicle under the Pine Island Glacier (PIG) ice shelf, which mapped the subglacial topography in great detail, and measured water temperature, salinity, and oxygen content (Jenkins and others, 2010). These observations showed that CDW now floods the cavity below PIG, more than 30 km upstream of areas that were at least partially grounded as recently as the early 1970s. At several degrees above freezing, this CDW carries enough heat to be melting the ice from below at rates in excess of 50 m/year, in good agreement with estimates independently derived from the observed ice velocities and thinning rates (e.g. Rignot and others, 2008).

Thoma and others (2008) used a regional ice-ocean model to show that recent changes in the influx of CDW in the ASE could be attributed to changes in the frequency and strength of westerly winds near the edge of the continental shelf. The timing of modelled increases in CDW influx is similar to the timing of two distinct phases of acceleration on PIG, in 1974-1987 and after 1994, that were separated by a period of quiescence (Joughin and others, 2003; Rignot and others 2008; Scott and others, 2009; Wingham and others, 2009). Jenkins and others (2010) found that as of 1973, the grounding line of PIG had already retreated from the top of a subglacial topographic ridge, suggesting that the retreat throughout the observational record, as well as into the future, was likely to be an inevitable result of the well-known marine ice sheet instability (Hughes, 1981; Schoof, 2007) associated with deepening of the seabed inland of the ridge crest. This finding raised a question about the relative roles of contemporaneous ocean forcing and continuing ice sheet response to an earlier event in controlling the current behaviour of the glacier. However, Joughin and others (2010) suggest that subtle topographic highpoints on the otherwise downwards slope from the ridge crest could have halted grounding line retreat, at least temporarily, and that ocean forcing could have played a role in re-starting the retreat in the 1990s. Furthermore, not only PIG, but also Thwaites and Smith Glaciers thinned in the 1990s (Shepherd and others, 2002), and glaciers are thinning at present nearly everywhere along the Amundsen Sea margin of the WAIS despite varying bed geometry (e.g. Pritchard and others, 2009).

Thus, while the complex interaction between bed slope, glacier dynamics and ocean forcing remains to be fully understood, the evidence appears to be firm that changes in CDW inflow to the ASE, driven by changes in local wind forcing, have played a role in influencing the thinning and retreat of glaciers in the ASE. This raises the question of how the observed changes in winds in the ASE are related to larger-scale changes in atmospheric circulation. Thoma and others (2008) found that the correlation between local atmospheric circulation in the ASE and commonly used indices of the large scale circulation, the Southern Annular Mode (SAM) index and the Southern Oscillation Index

(SOI), was low. However, the causes of local wind changes cannot generally be ascribed to a single large-scale index, and the relevant dynamics probably depends on the season in which changes in CDW influx have occurred. In this paper, we explore the relationship between ASE winds, modelled CDW upwelling, and the large-scale atmospheric circulation in more detail.

DATA AND METHODS

Thoma and others (2008) modelled CDW intrusions onto the ASE shelf using a regional ice-ocean model based on a version of the Miami Isopycnic Coordinate Ocean Model of Bleck and others (1992) adapted to include sub-ice-shelf cavities by Holland and Jenkins (2001). They forced the model with sea level pressure and sea surface temperature variations from the NCEP climate reanalysis data (Kalnay and others, 1996). We use their monthly model output of the thickness of CDW layers (isopycnic model layers 7 and 8) on the continental slope and on the inner shelf, near the margin of PIG, for the period 1980 through 2004. The continental shelf edge in the ASE region is oriented approximately east-west, and Thoma and others (2008) argued that increased westerlies would lead to enhanced CDW intrusion across the shelf. As a proxy for the westerlies they used a simple index for the geostrophic wind, based on the sea level pressure (SLP) difference across a box (~68°S to 72°S) straddling the shelf edge. Here, we use the zonal wind stress centred at 70°S and averaged between 100° W to 125°W (Fig. 1) as a more direct measure of atmospheric forcing. Both measures are highly correlated with one another and are in practice directly comparable.

A simple explanation for the relationship between westerly wind stress and CDW inflow is the northward Ekman transport of surface waters that drives an eastward geostrophic current and an associated up-slope, on-shelf flow of CDW within the bottom boundary layer. While this process undoubtedly operates, the bulk of the inflow, which is focussed on the shelf-edge troughs, is driven by more complex interactions between the temporally varying shelf-edge currents and the spatially varying topography, and is probably related to both the overall strength and the variability of the westerly wind stress (Klinck and Dinniman, 2010). Once on the continental shelf, CDW is transported into Pine Island Bay by a combination of the cyclonic circulation on the shelf and the deepening of the seabed along the axes of the main glacial troughs (Nitsche and others, 2007). An important caveat is that the resolution of the Thoma and others (2008) model (as well as that of the atmospheric forcing data), cannot explicitly capture the smaller-scale processes that control the on-shelf flow. While the isopycnic formulation of that model provides a natural framework in which to parameterise the transports associated with sub-grid-scale eddies, the results remain to be corroborated by a model that resolves the eddy processes. These caveats on grid resolution notwithstanding, seasonal mean westerly wind stress is likely to be a physically meaningful measure of the atmospheric dynamics relevant to CDW inflow.

To obtain wind stress data, and for analysis of the large-scale climate fields, we use both NCEP2 (Kanamitsu and others, 2002) and a combination of ERA-40 (1979-2002) and ERA-interim (2003-2009) reanalysis products (Uppala and others, 2005; 2008). For the variables we consider, ERA-40 and ERA-interim are indistinguishable during their overlap period (1989 to 2002). Following other recent work (Ding and others, 2011), in some figures we show ERA-40/ERA-interim data only, but the results are not dependent on which product is used. While problems in the reanalysis data have been noted, particularly at high southern latitudes (e.g. Bromwich and Fogt, 2004), due to changes in the availability of direct observations, we emphasize that the focus here is on the relationship between the reanalysis data and the results from an ocean model that has been forced by those same data; possible errors in the reanalysis data sets will thus have little influence on the results.

RELATIONSHIP BETWEEN LOCAL WIND STRESS AND MODELLED CIRCUMPOLAR DEEP WATER INFLOW

In the climatological mean, westerly wind stress at the ASE shelf edge occurs predominantly in fall through spring, with a maximum in late winter (Fig. 1). In austral summer, the wind stress is weak easterly. The true seasonal variability of CDW intrusions in the ASE is not known, due to the very limited available data. However, the model results of Thoma and others (2008) show the greatest quantity of CDW on the continental slope during spring, about one month after the climatological maximum in westerly wind stress, and a subsequent maximum on the inner shelf one to two months later. The lag between wind stress over the continental slope and modelled inner shelf CDW layer thickness is quite consistent (~2.5 months) on both seasonal and interannual timescales, as clearly seen in a simple lag correlation plot (Fig. 2a). A spectral coherence calculation suggests the same phase lag also extends to decadal timescales (Fig. 2b,c), though this cannot be demonstrated to be statistically significant.

In the early 1990s, there is a significant increase in the thickness of modelled CDW in the ASE (Figure 3), at least approximately coincident with the observation of resumed acceleration of PIG after 1994 (Joughin and others, 2003). This appears as a gradual increase beginning in the early 1990s, and a transition to a period of larger and more variable inner shelf layer thickness around 1994. Although the mean monthly westerly wind stress (Figure 1b) does not show such an obvious transition, there is a pronounced increase in austral fall and early winter in the early 1990s (March through June; Figure 4). Indeed, while the seasonal maximum wind stress remains in winter and spring, the westerly wind stress in fall more than doubles between the 1980s and 1990s (Figure 4). Layer thicknesses remain high through the end of the model period (2005), and similarly, the fall wind stress remains anomalous through this period. Importantly, the maximum layer thickness change between the 1980s and 1990s –to early 2000s on the continental slope also occurs in fall and early winter, and the maximum layer thickness change on the inner continental shelf occurs one to three months later (Figure 3), indistinguishable from the average phase lag seen for seasonal and interannual variability. Thus, the significant increase in modelled CDW inflow beginning in the early 1990s appears to be the result of a corresponding shift in atmospheric conditions occurring in fall and early winter.

RELATIONSHIP BETWEEN LOCAL WIND STRESS AND LARGE-SCALE ATMOSPHERIC CIRCULATION

We now turn to the causes of the observed variability and change in westerly wind stress in the ASE. The climatological seasonal variations are well understood, and are associated with the development of a pattern of increased sea level pressure (SLP) immediately to the north of the ASE, a corresponding weakening of the low-pressure trough (Fig. 1) along the ASE coastline, and a westward shift and contraction of the Amundsen Sea Low (e.g. van den Broeke, 2000; Simmonds and King, 2004). This pattern of variability is associated with the 'semiannual oscillation', the amplitude of which is maximized near the ASE (Simmonds, 2003). A similar pattern of variability also occurs on longer timescales, and various mechanisms have been proposed depending on the season involved. Two commonly used indices of atmospheric circulation relevant to Antarctic climate variability are the Southern Annular Mode (SAM) index (e.g. Marshall, 2003), which reflects the strength of the average circumpolar westerlies, and the Southern Oscillation Index (SOI; e.g. Trenberth, 1984), which reflects conditions in the tropical Pacific. Ding and others (2011) have shown that correlation between the SOI and Amundsen Sea climate is linked more to variability in sea surface temperature (SST) in the central tropical Pacific than in the eastern Pacific region. The Niño3.4 region (5°S-5°N, 190°-240°E) is a commonly used measure of central Pacific SST variability.

Table 1 shows the correlation between the zonal wind stress over the ASE shelf edge and various measures of large-scale climate variability as a function of season. Statistically significant correlations in summer (DJF) are found only with the SAM index. In winter (JJA) and spring (SON), statistically significant correlations are found with measures of tropical variability but not with the SAM index. In austral fall (MAM), significant correlations are found with both the SAM index and central tropical Pacific and South Pacific Convergence Zone (SPCZ) SSTs, as well as with the SOI.

The seasonal differences in correlation patterns shown in Table 1 support previous work on the causes of interannual variability in the Amundsen Sea region. Most studies of the SAM have focused on summer, during which a significant trend in the SAM has occurred (e.g. Thompson and Solomon, 2002). There is no significant trend in the winter or spring SAM index in the last thirty years, but there are large changes observed in Amundsen sector sea ice and Antarctic surface temperatures in those seasons (Comiso and Nishio, 2008; Steig and others, 2009) that have been linked with changes in tropical Pacific SSTs (Ding and others, 2011; Schneider and others, 2011). Turner and others (2009) showed that recent trends in sea ice as well as changes in SLP and geopotential height in the Amundsen sector are significant in the fall in the last 30 years, and suggested that those changes – reminiscent of the climatological fall-to-winter change in the Amundsen Sea Low – could be explained by changes in the SAM. While this is consistent with our finding of a significant correlation between the fall SAM index and the ASE westerlies, this result is quite sensitive to the season chosen: in the fall/early winter season (AMJ), there is no correlation with the SAM. In contrast, wind stress in MAM and AMJ is consistently as high or more highly correlated with indices of tropical variability than with the SAM.

Figure 5 shows maps of the interannual correlation between ASE zonal wind stress and SLP, the upper troposphere streamfunction, and SST for austral fall (MAM). The most prominent feature in SLP is a significant correlation in the Amundsen Sea sector of the Southern Ocean. The SLP anomalies are the surface expression of a deep coherent tropospheric circulation, with corresponding geopotential height anomalies in the middle and upper troposphere. The dynamical connection between ASE wind stress and the tropical Pacific is apparent in the correlation with the streamfunction at the 200 hPa level (Fig. 5b), which shows a sequence of positive and negative correlation centres extending from the central equatorial Pacific to the far south Pacific. These patterns, along with the nearly equivalent barotropic structure of the high latitude geopotential height anomalies, are characteristic of a stationary Rossby wave response to tropical SST forcing (e.g. Gill, 1980; Hoskins and Karoly, 1981; Mo and Higgins, 1998). Correspondingly, the correlation between ASE westerly wind stress and SST features a positive SST anomaly in the central tropics, shown in Fig 5c.

The dynamics that are responsible for the teleconnection between the tropical Pacific and the south Pacific are well established (e.g. Sardeshmuhk and Hoskins 1988; Lachlan-Cope and Connolley, 2006): anomalously high SSTs in the central Pacific force an increase in tropical convection in regions of strong potential vorticity gradients (associated with the subtropical jet east of Australia), which creates a strong Rossby wave response along a great-circle path towards the Amundsen Sea. Indeed, Ding and others (2011) showed that modest positive tropical SST anomalies in the central Pacific, very similar in pattern to that shown in Fig. 5c, force atmospheric circulation anomalies consistent in pattern and amplitude with those shown in Fig. 5(a,b), including in the Amundsen Sea region.

The same physics that relates tropical SSTs with interannual anomalies in ASE zonal wind stress in MAM also appears to be responsible for the decadal changes in wind stress that account for the modelled changes in CDW layer thickness. Figure 6 shows the change in SLP, upper level stream function and SST between the 1980s and the 1990s that accompany the CDW layer thickness and wind stress changes shown in Fig. 3 and 4. The patterns of surface and upper level circulation and SST changes associated with the decadal changes in the ASE zonal winds are strikingly similar to the pattern of interannual correlations. This is not surprising, because decade-to-decade changes in the tropical Pacific are in part a reflection of differences in the character of El Niño-Southern Oscillation (ENSO) variability. For example, an important difference between the large El Niño event in 1982/1983 and that in 1997/1998 is that SST anomalies associated with the former were restricted to the eastern Pacific, whereas the latter involved significant warming in the central Pacific, the more important region for exciting a poleward propagating Rossby-wave response. Trenberth and Hoar (1997) noted that another difference between the 1980s and the 1990s is that the La Niña events of the latter decade were relatively weak. Hence, it appears that tropical SST forcing is responsible for a significant fraction of both interannual variability and decade-to-decade change in ASE zonal wind stress. Tropical SST forcing thus plays an important role in influencing the amount of warm CDW that flows across the continental shelf to bathe the PIG ice shelf and the floating termini of other outlet glaciers in the Amundsen Sea Embayment.

DISCUSSION

The results above show that both interannual variability and longer term changes in westerly wind stress in the Amundsen Sea Embayment, relevant to forcing CDW inflow to the continental shelf, are significantly influenced by conditions in the tropics, particularly during the non-summer seasons. In the annual mean, as much of 30% or more of the variance in zonal ASE wind stress can be attributed to tropical forcing, depending on which measure of tropical variability is used. Regional, high latitude atmospheric processes are also important and must dominate the unforced variability. We cannot rule out an independent role for the SAM, but only in the summer is there evidence that the SAM plays a role that is greater than that of tropical forcing. The largest changes in wind stress occur in fall and early winter, when tropical forcing is clearly at least comparable, and probably dominant.

The distinction between low latitude versus high latitude forcing of the ASE wind stress has significant implications for our understanding of both recent and future changes in CDW inflow, and therefore to the past and future evolution of PIG and other outlet glaciers in the ASE. Recent changes in the SAM in summer have widely been attributed to radiative forcing resulting from the decline in stratospheric ozone (Thompson and Solomon, 2002; Gillett and Thompson, 2003; Son and others, 2008). If changes in the ASE winds in fall were similarly attributed to ozone-related SAM changes, this would imply that as the ozone hole recovers over the next few decades, the current, apparently anomalous wind field pattern will change, implying a return to pre-1990s CDW inflow strength.

The attribution of fall SAM changes to ozone forcing is somewhat problematic because this would require a three to six-month lagged response to spring ozone depletion (see e.g. Keeley and others 2007, Sigmond and Fyfe, 2010). Furthermore, the SAM index itself is not independent of tropical conditions: observations show, for example, that a positive SAM index is more likely to occur during strong La Niña years, and weak SAM index during strong El Niño years (Fogt and Bromwich, 2006; Fogt and others, 2011). L’Heureux and Thompson (2006) concluded that about 25% of the interannual variability in the SAM can be attributed to El Niño-Southern Oscillation variability in austral summer, while Grassi and others (2005) show that the observed zonal SAM pattern change at 500 hPa geopotential heights between the 1980s and the 1990s can be simulated with an atmospheric general circulation model using observed tropical SST forcing alone. Thus, correlation between the SAM index and ASE wind stress does not in itself provide evidence of independent high latitude forcing. In contrast, the role of tropical forcing is unequivocal.

CONCLUSIONS

Flow of warm CDW onto the continental shelf has played a critical role in the high melt rates and recent thinning and retreat of glaciers in the Amundsen Sea Embayment region of West Antarctica. Variability in CDW inflow is strongly influenced by the westerly wind stress over the continental slope, and tropical SST forcing has played an important, if not dominant, role in recent changes in the zonal wind regime in the ASE. Continued changes in tropical SSTs can be expected in the future, due to increased global radiative forcing from anthropogenic greenhouse gases, and warming in the central tropics is particularly pronounced in most AR4 model runs (Ding and others, 2011), suggesting that the current wind stress regime in the ASE is likely to persist.

We caution that the link from wind forcing to CDW inflow changes to glacier retreat is not a simple linear process, and that once the PIG retreated past a subglacial ridge some decades ago, continued glacier thinning and retreat was probably inevitable even without the recent changes in wind forcing (Jenkins and others, 2010; Jacobs and others, 2011). In this context, it is interesting to note that significant warming in the central tropical Pacific last occurred in the 1940s, and ice core evidence indicates that the impact on climate in the Amundsen Sea sector of Antarctica was comparable with what has been observed recently (Schneider and Steig, 2008). This suggests that tropical SST forcing during the 1940s is a viable candidate for the initiation of the current period of change in the Amundsen Sea ice shelves, which clearly was underway at least by the 1970s (Jenkins and others, 2010). Photographic evidence shows that in 1947, the PIG ice shelf was only slightly more advanced than in the early 1970s, but that a large area of icebergs and sea ice extended seaward of the ice front, which may be evidence of significant calving over the preceding decade (Rignot, 2002). There is also independent evidence from sediment cores that a larger ice shelf may have occupied the ASE at some time prior to this, possibly during the 20th century (Kellogg and Kellogg, 1987). We speculate that a more extensive ice shelf may have partially collapsed following the very large El Niño event of 1939-1942.

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	Season					
	DJF	MAM	AMJ	JJA	SON	Annual
SAM Index	***0.34***	**0.35**	0.11	0.07	0.09	0.05
Southern Oscillation Index	-0.20	**-0.36**	**-0.36**	-0.31	**-0.61**	**-0.60**
Eastern Tropical Pacific SST	0.21	0.19	0.20	0.13	**0.35**	**0.36**
Central Tropical Pacific SST	0.19	**0.36**	**0.43**	***0.34***	**0.47**	**0.53**
Nino3.4 SST	0.20	0.33	**0.40**	**0.37**	**0.45**	**0.55**
SPCZ SST	0.14	**0.49**	0.33	0.14	0.26	**0.40**

Table 1. Correlations between zonal wind stress (ERA-40/ERA-interim) near the shelf edge of the Amundsen Sea Embayment, and the SAM index, Southern Oscillation Index, and sea surface temperatures (SST) in the tropical and subtropical Pacific (ERSST3), for the period 1979-2009, for different 3-month seasons and the annual mean. Latitude and longitude ranges for the SSTs are as follows: Eastern: 6°S-6°N, 240°-280°E; Central: 6°S-6°N, 160°-240°E; Nino3.4: 6°S-6°N, 190°-240°E; SPCZ: 20°S-8°S, 180°-240°E. Bold numbers indicate significant correlation above the 95% level, italics at the >90% confidence level. Autocorrelation in the wind stress data is <0.1, so that correction of significance levels for autocorrelation is negligible.

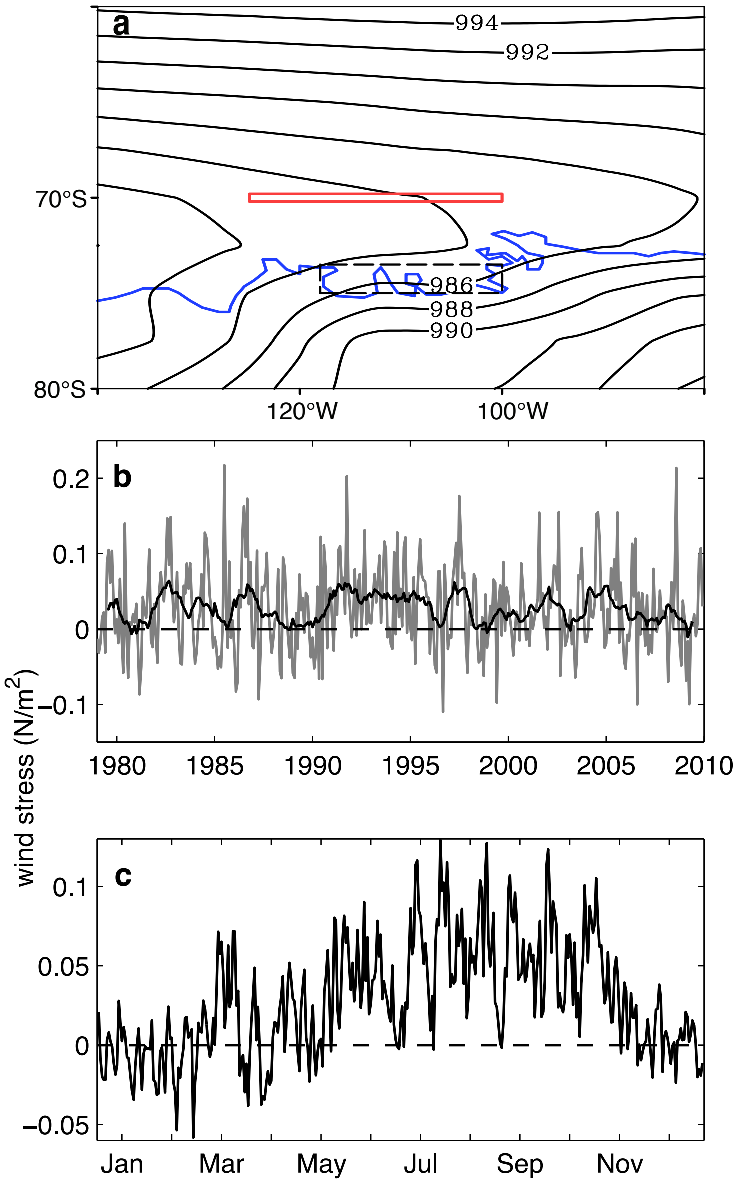

Figure 1. a) Climatological sea level pressure (hPa) for June-July-August over the Amundsen and Bellingshausen Seas, with outline of the continent in blue. Red box shows the location (70°S, 100° to 125° W) centered on the continental slope used for the zonal wind stress. Dashed box denotes the Amundsen Sea Embayment. **b**) Monthly zonal wind stress along with a 13-month running mean (bold) and **c)** daily climatological wind stress from ERA40/ERA-interim 1979-2009. Positive values are westerly.

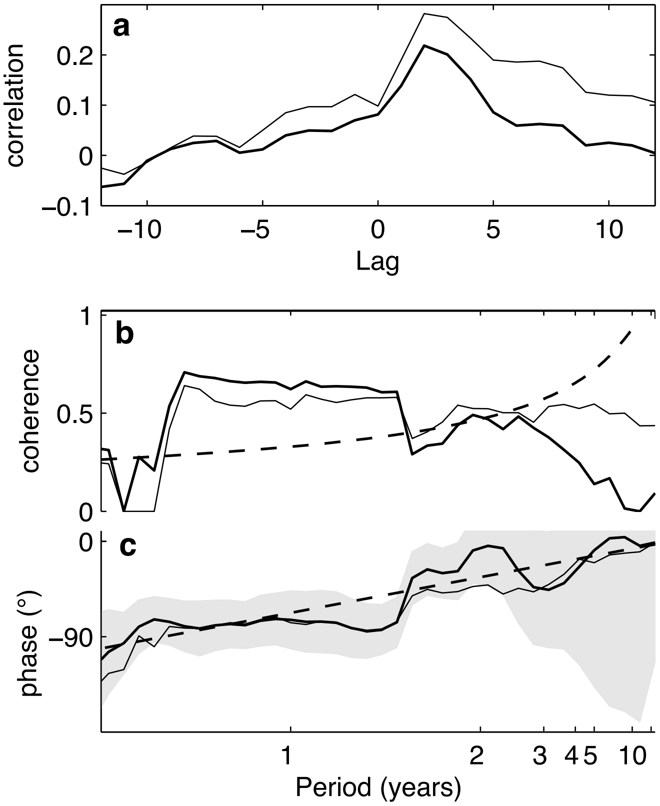

Figure 2. Relationship between westerly wind stress near the continental slope and CDW layer thickness on the inner shelf in the Amundsen Sea Embayment. **a**) Correlation of monthly anomalies (mean seasonal cycle removed) of zonal wind stress with CDW layer thickness (layers 7 and 8 from Thoma and others, 2008), for December 1979 through November 2004. **b**) Spectral coherence between wind stress and CDW inner shelf layer thickness as a function of period. Dashed line shows 95% confidence limit. **c**) Phase of the coherence estimates, with 95% uncertainties (shading). Dashed line shows phase for a constant 2.5-month lead of wind stress over CDW changes. Spectral coherence and phase calculated using the Thomson (1982) multi-taper method with a bandwidth of ~0.6/yr. In each figure, thin lines are NCEP2 data, thick lines are ERA-40/ERA-interim

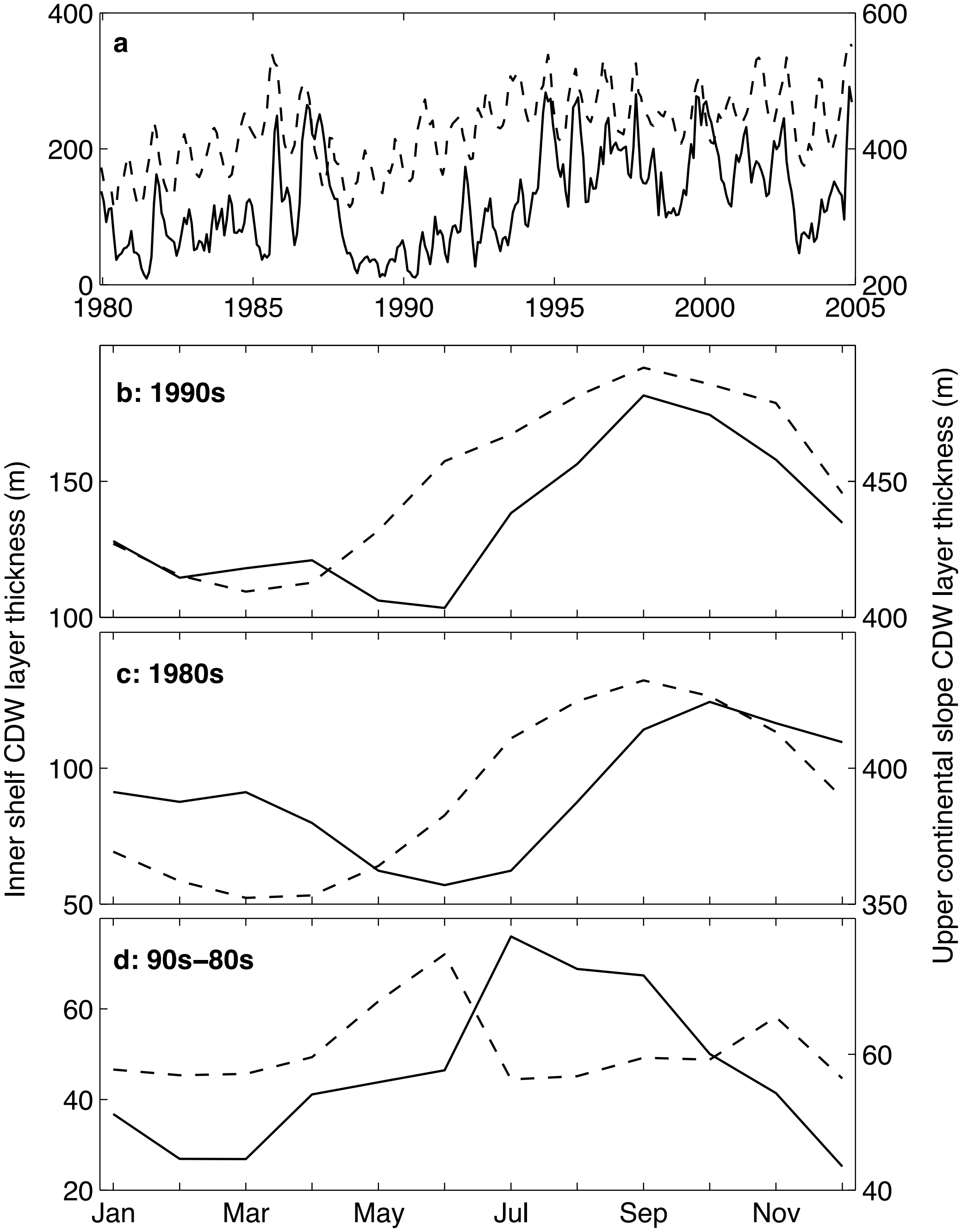

Figure 3. Variations in thickness of CDW layers (layers 7 and 8) on the continental slope (dashed) and inner continental shelf (solid) from Thoma and others (2008). **a**) Monthly averages from 1980 to 2004. **b**) Seasonal climatology for the period 1990-1999. **c**) Seasonal climatology for the period 1980-1989. **d**) The difference between b) and c), significant at the 95% confidence level for the fall (MAM) season, based on a two-tailed *t*-test.

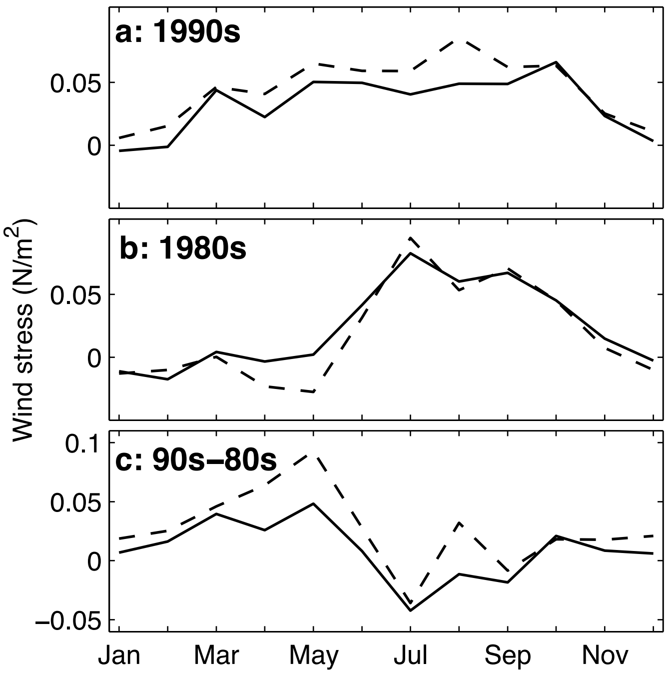

Figure 4. Seasonal wind stress climatology from NCEP2 (dashed) and ERA-40 (solid) for **a)** 1990-1999 **b**) 1980-1989, and **c)** their difference. Although the mean wind stress maximum occurs in the same seasons in both periods (winter and spring), the seasonal mean westerly wind stress more than doubled in the fall between the 1980s and the 1990s.

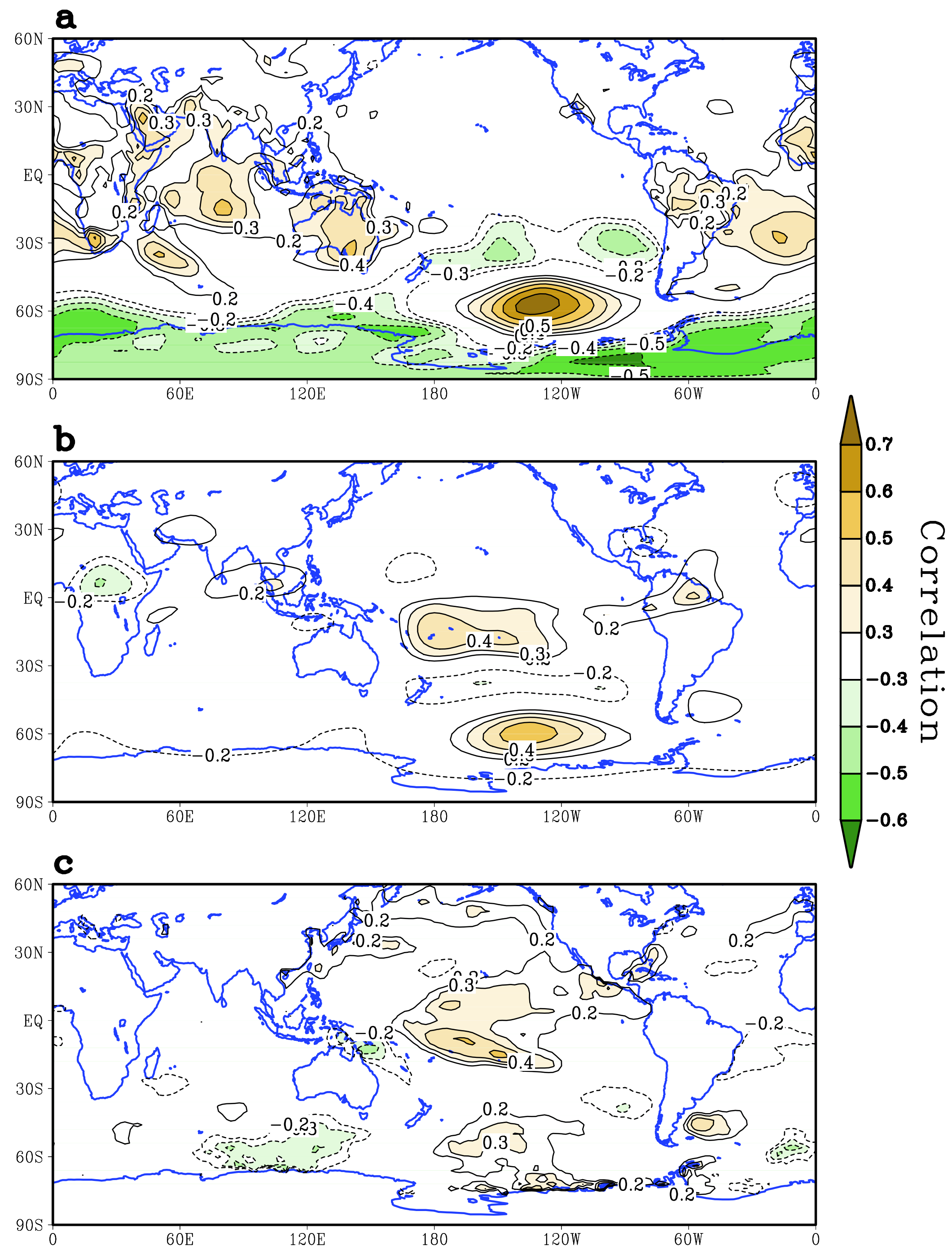

Figure 5. Correlation between zonal wind stress in the Amundsen Sea Embayment region (red box in Figure 1) in austral fall (MAM) **a)** sea level pressure, **b)** upper troposphere (200 hPa) stream function and **c)** sea surface temperature. Data are from ERA-40/interim and ERSST3 (Smith and others, 2008) for the period 1979-2009. Areas of statistically significant correlation are shaded (~0.35 corresponds to 95% confidence level).