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TRANSIENTS AND THE AIR-SEA FLUXES IN THE ANTARCTIC SEA ICE ZONE

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1. INTRODUCTION

There are very strong thermal gradients between the Antarctic continent and the sea ice zone, and between that zone and the ocean to the north. As a result of these contrasts the sea ice domain is one of strong cyclogenesis and high cyclone frequency. The presence of these systems and of sea ice dramatically impacts on the fluxes between the atmosphere and surface. The area covered by Antarctic sea ice varies seasonally from a minimum of about 4 x 10^6 km^2 in February to a maximum of 19 x 10^6 km^2 in September (Gloersen et al., 1993); this strong seasonality significantly modulates the nature of the interactions. In particular, marginal ice zones are regions of intense air sea interactions and many studies (e.g., Watkins and Simmonds, 1998) have explored the complex relationships between the variabilities of wind stress, sea-ice concentration, and temperature.

One of the difficulties of exploring the interaction of the atmosphere and the surface over sea ice is associated with obtaining adequate and reliable data for analysis. The high latitude oceans and sea ice zones are obviously challenging places from which to obtain data, and in general they are regions of low population density. Hence even at the periphery of these regions only modest amounts of relevant data are available. These problems are particularly marked in the Southern Hemisphere (Simmonds, 2003). Until the IGY few data were collected at high southern latitudes, and there were very few stations monitoring meteorological conditions on the Antarctic coast (King and Turner, 1997). This state of affairs improved during and after the IGY, and has been further ameliorated by the more recent deployment of automatic weather stations on the Antarctic continent, but it is still true that the coverage afforded by these is far from adequate. The data situation is even worse over the ice zone itself. Much can be obtained from various platforms such as buoys trapped in the sea ice (e.g., Kottmeier and Sellmann, 1996) and observations taken from ships during cruises and voyages to resupply coastal stations. Clearly, however, the latter source gives a somewhat distorted view as such voyages will be undertaken in regions of lower ice concentration and will be subject to a 'fair-weather' bias.

Given these difficulties it is of importance that our understanding of the large scale manifestation of surface-atmosphere interaction be based on representations of the atmosphere and ocean which has been assembled with as much of the available data as possible, as well as being consistent with the fundamental large scale atmospheric physics. As such, the use of the products of meteorological 're-analysis' are very important to this process. These products are assembled by inserting historical data into a global atmospheric model. A consequence of using a numerical model as a vehicle for this process is that the influence of data can be felt remotely and in a manner in accord with the physics of the atmosphere. This means, in principle, one can have a degree of confidence in the quality of the analyses so-derived, even if there are no data in the immediate vicinity. Further, the 'data assimilation' process used in re-analysis means that optimum use can be made of remotely sensed data and, in particular, the vertical temperature profile that many platforms provide can, through the physics incorporated into the model, produce physically consistent wind structures in the lower layers of the atmosphere.

Here we explore aspects of the spatial distribution of Antarctic sea ice and of the low level atmospheric structure over the sea ice zone and, in particular, the synoptics of this highly baroclinic region. Using this perspective we quantify the rate at which momentum and kinetic energy is transferred to the ocean in the broad region centred on the sea ice zone. We address the contribution of synoptic 'turbulence' to these fluxes, and quantify the extent to which these contributions depend on how this turbulence is defined.

2. DATA SETS

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The sea ice concentration data used in this study has been complied from the 25 km resolution combined Scanning Multichannel Microwave Radiometer (SMMR) (years 1978-1987) and Special Sensor Microwave/Imager (SSM/I) (years 1987-2002) Goddard Space Flight Center data set, as described by Cavalieri et al. (1999). These data cover the period from October 1978 to December 2000. The SSMR data are available every second day, while the sampling frequency is daily for the SSM/I set.

The atmospheric data are the NCEP-DOE Reanalysis (Kanamitsu et al., 2002). These global analyses (on a 2½° x 2½° latitude-longitude grid) are available every 6 hours, and we use the period from 1 January 1979 to 31 December 2002.

3. BACKGROUND CONSIDERATIONS

The ice 'edge' is defined to be the location of the 15% concentration contour (Parkinson, 2004). Figure 1 shows the longitudinal structure of the mean edge in December, March, June and September. There is a significant seasonal cycle, at virtually all longitudes, and hence the nature of the interaction of the atmospheric boundary layer and the surface will change dramatically over the course of the year.

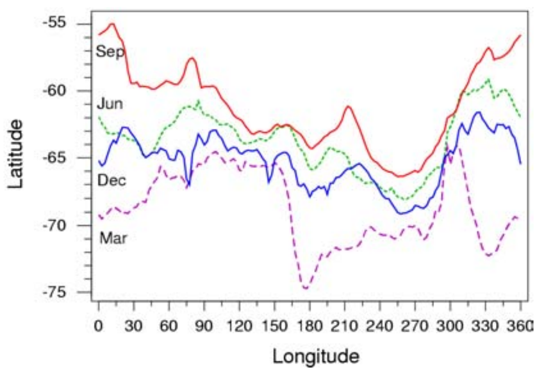

FIG. 1. Mean latitude of sea ice edge as a function of longitude for December, March, June and September (1979-2000).

The sea ice extent reaches its maximum in spring (September to November (SON)). The climatological average mean sea level pressure for this season is presented in Figure 2. Also shown on the plot is the mean vector winds at elevation 10 m above the surface, and the mean location of the sea ice edge. The low-level winds can be seen to be close to geostrophic, with a small component of cross-isobaric flow toward the circumpolar trough. The southern sea ice area is overlain by broad regions of both westerlies and easterlies which induces complex interactions between the ice and the atmospheric boundary layer.

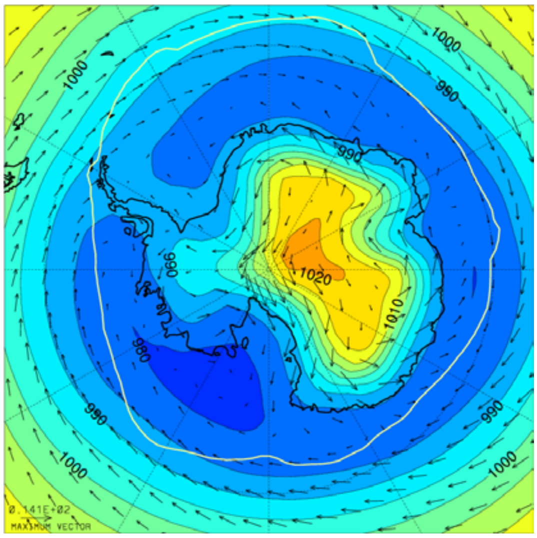

FIG. 2. Climatological average MSLP, 10 m vector winds and northern limit of sea ice for September-November. The contour interval for MSLP is 5 hPa and the length of the arrow in the bottom left corner represents a vector wind magnitude of 14.1 m s^{-1}. The mean ice extent is represented by the light line.

The subantarctic region is host to very energetic and frequent cyclonic systems (e.g., Simmonds and Keay, 2000a Simmonds et al. (2003) and as we shall see these play a very significant role in the interactions of interest.

A comprehensive summary of the characteristics of cyclones over the spring sea ice domain in our 24-year sample is presented in Figure 3. (The objective cyclone tracking software of Simmonds and Murray (1999) is used to obtain these.) Part (a) exhibits the average cyclone 'system density' and shows that the axis of highest density is found to the south of 60°S, and that maxima are positioned in the Indian Ocean. The largest systems (i.e., those with the greatest mean radius (see, e.g., Simmonds et al., 1999)) are also found in this region, as well as just east of the Greenwich meridian (Figure 3b). The 'depth' of systems, which can be interpreted as an integrated measure of the influence of cyclones (Simmonds and Keay, 2000b; Simmonds et al., 2003), (Figure 3c) indicates that the most influential systems are found in the Indian Ocean, which is also the region of greatest density.

4. SURFACE - AIR FLUXES

The scalar wind stress acting on the surface can be parameterized as

$$\tau = \rho C_D u^2 \qquad (1)$$

where ρ is the atmospheric density, u the wind speed at some (near surface) reference level, and C_D the drag coefficient with respect to that reference level. The density and the exchange coefficient vary with space and time, and it is known that around the Antarctic periphery there are significant covariances between these and the basic meteorological parameters which, in turn, impact on the surface-air fluxes in that region (Simmonds and Dix, 1989). Our prime focus here is on the effect of these covariances and hence, for simplicity, we regard the density and drag coefficient as constant and hence the wind stress (or momentum flux) is proportional to u^2. Over the sea ice zone and surrounding ocean we are also interested in the rate at which mechanical (i.e., kinetic) energy is deposited at the surface. Again, taking these terms to be constant, this rate is proportional to the cube of the wind speed (u^3). This flux is of direct relevance to wave setup and the development of the fully-developed sea.

To explore the role that the high 'turbulence' intensity (Simmonds et al., 2005) (an outcome associated with the frequent passage of intense cyclones) plays in these fluxes we perform the traditional decomposition of the wind speed into a mean and turbulent part

$$u = \overline{u} + u' \qquad (2)$$

where the prime denotes the deviation from the mean. We may then express the time mean of u^2 (S) (proportional to the mean magnitude of the surface stress) as

$$\overline{u^2} = \overline{u}^2 + \overline{u'^2} \qquad (3)$$

A similar expansion for the mean of u^3 (C) leads to

$$\overline{u^3} = \overline{u}^3 + 3\overline{u}\,\overline{u'^2} + \overline{u'^3} \qquad (4)$$

(Simmonds and Keay, 2002).

Simmonds et al. (2005) have considered in detail the structure of each of the covariance terms, and here we only consider the contribution form the first terms on the right hand sides, denoted respectively as $S1$ and $C1$.

Note that the decomposition performed in equations (2) – (4) does not fully reflect the contributions of the synoptic eddies. For example, suppose that a cyclonic system passed near a given point and that the vector wind there rotated through a full 360 degrees but maintained its magnitude. As defined above, the wind speed would always equal the mean wind speed and hence the deviations would be zero. This clearly does not convey a true reflection of the variability. An alternative approach in which the signature of the transients is perhaps better represented is to decompose the wind speed with respect to $|\overline{\boldsymbol{u}}|$, the magnitude of the time mean wind vector. That is, represent

$$u = \tilde{u} + u'' \qquad (5)$$

where

$$\tilde{u} \equiv |\overline{\boldsymbol{u}}| \qquad (6)$$

(Then in the special case presented above, *all* of the wind speed is tied up in the turbulent component.) Note, defined in this manner, the time mean of the perturbations is not identically zero, and indeed it is easily shown that $\overline{u''} \geq 0$. Using this method of partition, the mean of the products are

$$\overline{u^2} = \tilde{u}^2 + 2\tilde{u}\,\overline{u''} + \overline{u''^2} \qquad (7)$$

and

$$\overline{u^3} = \tilde{u}^3 + 3\tilde{u}^2\,\overline{u''} + 3\tilde{u}\,\overline{u''^2} + \overline{u''^3} \qquad (8)$$

Analogous to before, we refer to the first terms on the right hand sides of these equations as $SV1$ and $CV1$.

The spring distribution of S (Figure 4a) (calculated from the 10 m winds) shows a region of values over 200 m^2 s^{-2} in the Indian Ocean midlatitudes. Values are somewhat smaller over the sea ice zone, but are still close to 100 m^2 s^{-2} there. It will be observed that these mean values increase significantly in the coastal sea ice zone, a reflection of the cooperative interaction of synoptic systems and katabatic flow inducing very strong coastal winds in certain sectors (e.g., Murphy and Simmonds, 1993). We present in Figure 4b the percentage of S that is contributed by mean of the wind speed ($S1$). In the mid latitude regions identified earlier more than 85% is contributed by this term. Further south and over the coastal regions this decreases to about 70%, meaning that there the temporal deviations contribute about 30% to *S*.

When the alternative view of the role of transients presented by equation (7) is taken we obtain a rather different perspective. Figure 4c shows that *SV*1, (i.e., $\widetilde{u}^2$) contributes only about 50% to *S* in the mid latitudes, and drops to about 10% at a distance of about 600km from the coast (and hence the turbulence contributes about 50% and 90% respectively in these regions). This decrease in value is expected, occurring over regions where the magnitude of the mean vector wind drops to very small values (Figure 2). Further south and over the coastal region the winds are strong and have great directional constancy (Simmonds and King, 2004), and hence the contribution of *S*1 starts to increase and assumes values of about 60%.

The two methods of breaking down the rate at which mechanical energy is deposited over the ocean and the sea ice zone also reveals the central role played by the strong transient systems over this domain. The mean SON distribution of *C*, shown in Figure 5a, reveals values in excess of 3000 $m^3\ s^{-3}$ in the Indian Ocean near 45^{o}S. Into the sea ice zone this rate of transfer decreases, but in a number of sectors of the East Antarctic coast assumes values comparable with those mentioned above. Figure 5b indicates that in midlatitudes *C*1 (mean term) contributes about 70% of *C*, and approximately 50% further south. When the alternative decomposition (equation (8)) is used the cube of the magnitude of the mean vector wind (*CV*1) makes only a very modest contribution to *C*, with values dropping as low as 10% over much of the sea ice zone (Figure 5c).

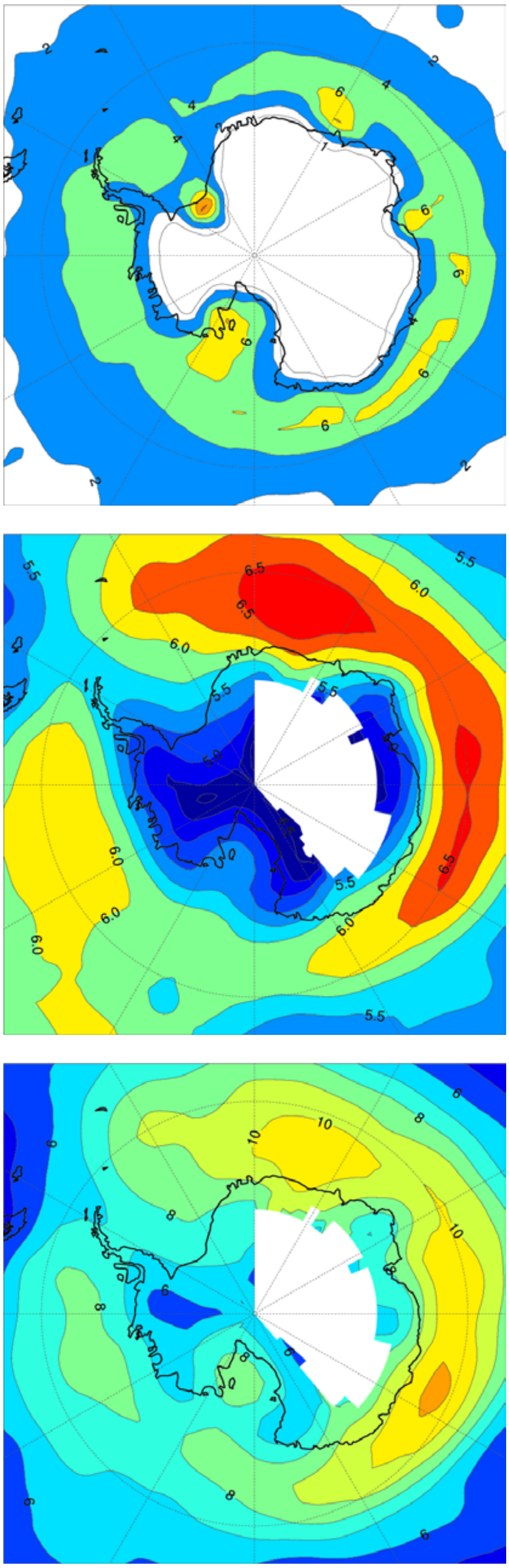

FIG. 3. Mean spring (a, top) system density, (b, middle) radius, and (c, bottom) depth of cyclones in the Antarctic sea ice region. The contour intervals are 2 cyclones per analysis per 1000 (deg. lat.)^2, 0.25 deg., and 1 hPa.

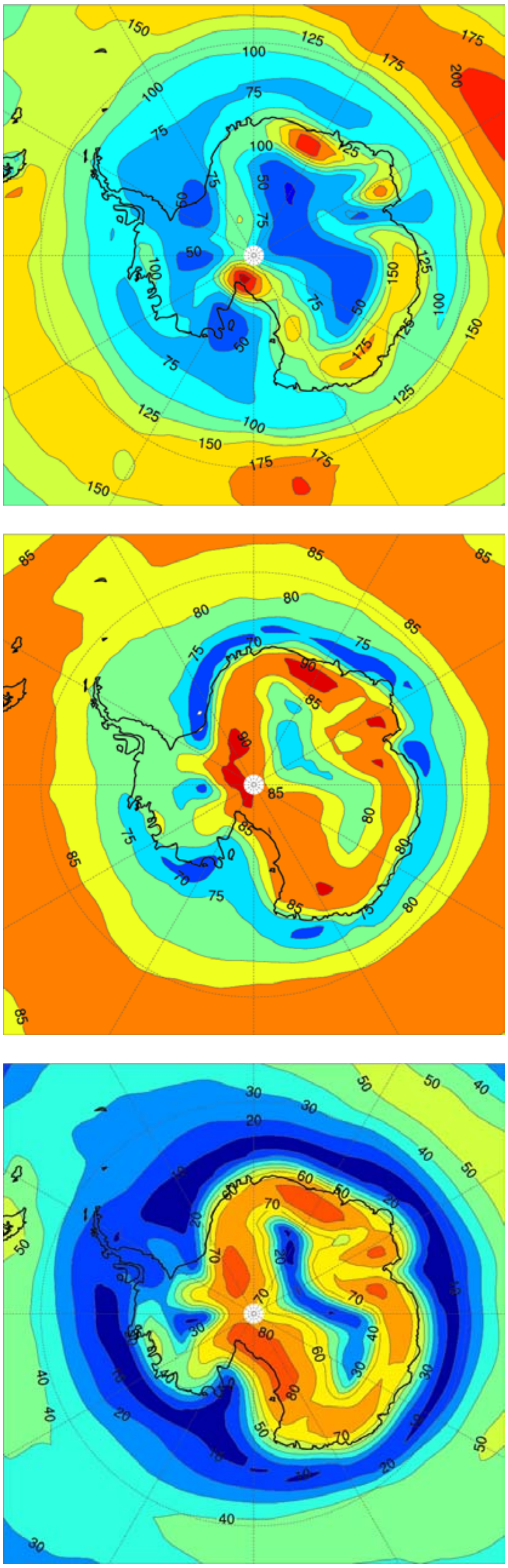

FIG. 4. Spring distribution of (a, top) *S* and the percentage of *S* that is contributed by (b, middle) *S*1, and (c, bottom) *SV*1. The contour intervals are 25 $m^2\ s^{-2}$, 5%, and 10%.

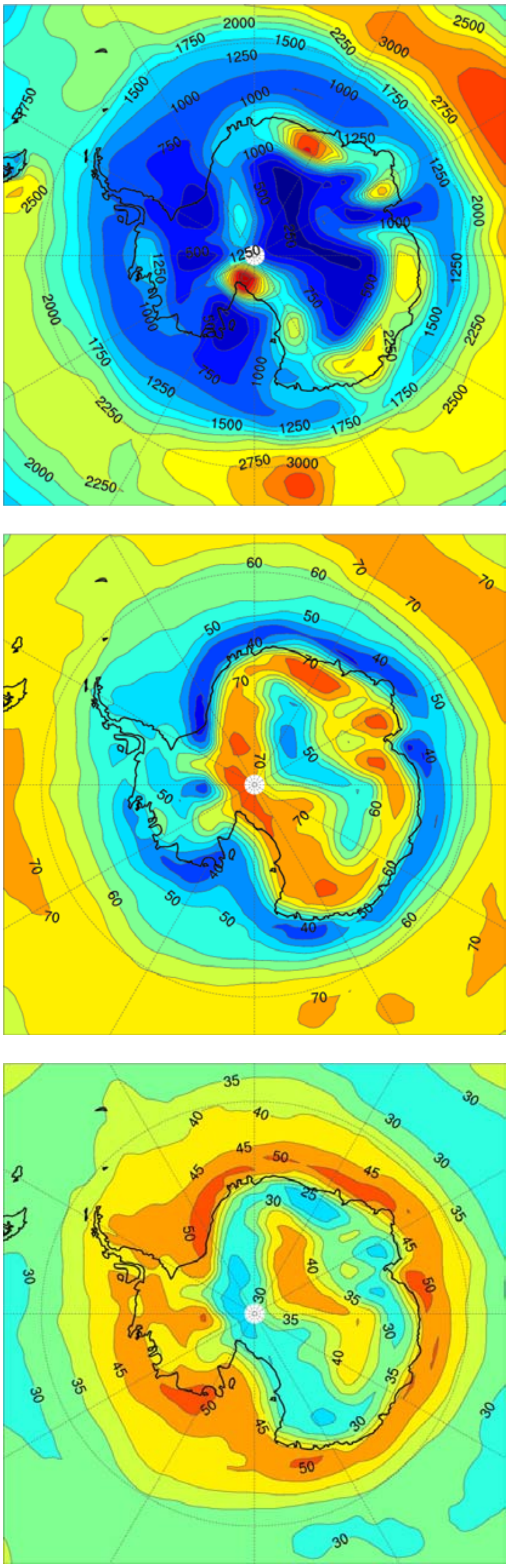

FIG. 5. Spring distribution of (a, top) *C* and the percentage of *C* that is contributed by (b, middle) *C*1, and (c, bottom) *CV*1. The contour intervals are 250 $m^3\ s^{-3}$, 5%, and 5%.

5. CLOSING COMMENTS

We have explored aspects of the spatial distribution of Antarctic sea ice and of the low level atmospheric structure over the sea ice zone and, in particular, the synoptics of this highly baroclinic region making use of modern reanalysis products. Cyclonic systems have been shown to play a very important role in influencing boundary layer exchanges of momentum and kinetic energy over the Antarctic sea ice zone. The Indian Ocean sector in particular is one of very vigorous and very frequent synoptic systems and high levels of ‘turbulence intensity’.

The paper has documented the mean square and the mean cube of the low-level wind speed (respectively proportional to the mean magnitude of the surface stress and the rate at which kinetic energy is deposited in the ocean) and confirmed that these assume very large values in the Indian Ocean. To quantify the contribution made to these terms by the transient synoptic systems we have decomposed the wind speed in two different ways. Our results show that the cyclonic systems make a significant contribution to these fluxes over the sea ice domain. It is very important to note that over the sea ice zone the magnitude of this contribution depends on precisely how the transients are defined, and the method suggested here (which perhaps reflects better the presence of transients than do traditional methods) shows the role of these to be central.

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