Role of daily surface forcing in setting the temperature and mixed layer structure of the Southern Ocean

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1. Introduction

The role of high-frequency surface forcing in setting the temperature and mixed layer structure in the Southern Ocean is investigated in an ocean general circulation model. The analysis suggests that daily fluctuations in the momentum flux, and the enhanced vertical mixing they produce, have the biggest effect on the time-mean temperature and mixed layer structure. In particular, summertime (December–February) surface temperatures are cooler, subsurface temperatures are warmer, and the mixed layer is thicker in the experiment with daily forcing than in the one without. The wintertime mixed layer in the Pacific sector is, however, thinner in the presence of daily forcing owing to the enhanced density contrast between the mixed layer and the ocean below. Daily fluctuations in the surface momentum flux are the leading cause of the high-frequency temperature variability at the high latitudes of the Southern Ocean, whereas the daily variability in the surface heat flux is more important at the midlatitudes. Since the air and sea-surface temperatures do not co-vary on short timescales in this study, daily fluctuations in the air temperature/humidity cause substantial variability in the air-sea turbulent exchanges of heat. The high-frequency part of the wave number–frequency spectrum of the daily sea surface temperatures is interpreted using a conceptual model of the response of the Southern Ocean to passing atmospheric storms. For high-to-medium frequencies, the preferred wavelengths in the conceptual model are determined by fast atmospheric advection; preferred low frequencies, in contrast, are determined by slow oceanic advection.


1. Introduction

[2] Several properties make the Southern Ocean a unique component of the climate system. The Southern Ocean encompasses the globe and, unlike the Atlantic, Pacific, and Indian oceans, does not have meridional boundaries and associated western boundary currents. The Southern Ocean connects the Atlantic, Pacific, and Indian oceans, and is a formation site of several important water masses. In particular, wintertime convection in the northern part of the Southern Ocean forms a deep mixed layer known as the Subantarctic Mode Water (SAMW), which is a precursor for the formation of the Antarctic Intermediate Water [McCartney, 1977].

[3] The Southern Ocean is a region with persistent strong winds and powerful storms, in which the air-sea exchanges of heat and momentum and their variability strongly influence stratification and water mass properties. Thus Rintoul and England [2002] demonstrate that properties of SAMW are controlled by fluctuations in the wind stress. A wide range of timescales characterizes climate variability over the Southern Ocean. The focus of this study is on the atmospheric fluctuations on the timescales of days and on their effects on the upper ocean. These high-frequency fluctuations have a form of propagating powerful storms with wind speeds exceeding 15 m s$^{-1}$ [Yuan, 2004].

[4] Atmospheric storms are capable of having a pronounced effect on the oceanic temperature and the mixed layer structure. Several studies of the upper-ocean response to hurricanes in the Northern Hemisphere report significant cooling of SSTs [Price, 1981; Large et al., 1986; Price et al., 1994; Large and Crawford, 1995; Doney, 1996], which is mainly caused by intensified mixing and deepening of the mixed layer [Large et al., 1986; Large and Crawford, 1995]. Large et al. [1986] hypothesize that vertical mixing draws its energy from shear instability when the wind forcing and inertial currents are in phase. The magnitude of the cooling strongly depends on the direction of wind stress vector [Price, 1981; Price et al., 1994; Large and Crawford, 1995]: The response is amplified when the vector is aligned with inertially rotating mixed layer current. Doney [1996] describes significant variability in the mixed layer depth at the synoptic timescale in the vicinity of Bermuda, and reports success of a one-dimensional model of the synoptic oceanic variability. Three-dimensional models of the tropical Pacific confirm that the daily anomalies in the wind stress enhance vertical mixing and cool the surface [Chen et al., 1999; Sui et al., 2003]. Sui et al. [2003] also demonstrate that daily winds act to warm the subsurface.
ocean and to deepen the mixed layer; these changes are compensated by intensified surface heating; changes in horizontal advection also play a role in the eastern tropical Pacific.

[5] Temporal variability in the atmospheric forcing over the Southern Ocean can be considered stochastic for sufficiently long timescales. For example, Connolly [1997] demonstrated that interannual variability in mean sea level pressure has a white spectrum consistent with random forcing by storm events with large-scale spatial patterns. The ocean, however, can integrate a white spectrum of atmospheric forcing into a red response signal [Hasselmann, 1976], and the preferred interannual modes of oceanic variability can be excited in the midlatitude oceans [Frankignoul et al., 1997; Junge et al., 2000]. For example, in their model of the Southern Ocean forced by random spatially correlated monthly anomalies of atmospheric fluxes of heat, moisture, and momentum, Weiss et al. [1999] reported propagating patterns similar to the Antarctic Circumpolar Wave [White and Peterson, 1996]. The timescale of these anomalies is set by oceanic advection as the time required by a water particle to travel over a given length scale. For timescales shorter than a few months, however, atmospheric advection can affect spectral properties of the oceanic response, as is demonstrated by a conceptual model of the oceanic response to passing atmospheric storms in the present study.

[6] This paper considers the oceanic response to daily fluctuations in the surface fluxes of heat and momentum in the Southern Ocean. The approach is to force an ocean general circulation model with surface fluxes with and without daily fluctuations, and to analyze the differences in the oceanic temperature and mixed layer structure. The analysis is restricted to the Southern Ocean south of 40°S, the region unique in its geometry and dynamics. The model and data sets are described in section 2. The analysis of the response of the time-mean temperature and mixed layer depth to daily fluctuations in surface forcing is presented in section 3.1. Effects of the daily forcing on the high-frequency variability in the upper ocean are analyzed in section 3.2. Finally, analysis of the spectral properties of a conceptual model of the response of the Southern Ocean to passing atmospheric storms is presented in section 3.2. Summary and discussion are given in section 4.

2. Model Description and Experiment Design

[7] The model is based on the GFDL MOM3 code [Pacanowski and Griffies, 1999]. The horizontal resolution is 2° in longitude and latitude, and the model domain is global and extends from 78°S to 84°N. There are 25 levels in the vertical, with resolution increasing from 17 m at the surface layer to 510 m at the bottom. The bathymetry of the model is derived from the Scripps Topography with the surface layer to 510 m at the bottom. The bathymetry of the model is proportional to the difference between the observed and model-simulated ice cover. The observed values for years 1979–2001 are taken from the National Snow and Ice Data Center data set.

3. Results From Numerical Experiments

[8] Following initial spinup of the model with climatological forcing, four 23-year simulations of years 1979–2001 are performed. In the experiment SMOOTH, daily fluctuations in the wind stress, wind speed, and air temperature and humidity are removed by a running 30-day average. Although in a strict sense this smoothing technique does not preserve a monthly mean, the differences between monthly means of the smoothed and unsmoothed data are verified to be negligible in this study. Daily fluctuations
henceforth are defined as the difference between the unsmoothed and smoothed data. The variability in the daily forcing generally intensifies toward the pole; Figure 1 shows the standard deviation for the daily anomalies in the zonal component of the wind stress.

The Southern Ocean is simulated reasonably well given the coarse spatial resolution in the model. Deviations from observations (World Ocean Atlas 2001) in the simulated annual-mean temperatures and salinities do not exceed 3°C and 0.4 psu at any location in the upper 100 m; zonally averaged annual mean temperatures and salinities are within 0.4°C and 0.1 psu of the observations. The total volume transport through the Drake Passage is approximately 115 Sv, which is smaller than the modern estimates of 134 ± 11 Sv [Cunningham et al., 2003].

The mixed layer depth is defined in this study as the depth at which potential density exceeds its surface values by 30 kg m⁻³. The simulated annual mean mixed layer depth is within 5–10 m of the values calculated from the World Ocean Atlas temperatures and salinities. The simulated wintertime mixed layer at several locations is, however, noticeably deeper than the mixed layer computed from the observations. The largest disagreements are found in the ice-covered regions, which can be explained by crude representation of convection in the model, and by large uncertainties in the observed values of temperature and salinity.

In three other experiments, daily fluctuations are consequently added to components of surface forcing (Table 1). Comparison of experiments SMOOTH and DAILY_STRESS thus serves as an explicit evaluation of the role of daily fluctuations in the wind stress, comparison of experiments DAILY_STRESS and DAILY_WINDS serves as an explicit evaluation of the role of daily wind speeds, and comparison of experiments DAILY_WINDS and DAILY serves as an explicit evaluation of the role of daily air temperature/humidity. The analysis in this study is restricted to the upper-ocean temperatures and the mixed layer depth. As demonstrated by the results of this study, the stratification below 100 m is generally unaffected by daily anomalies in surface forcing. The response of the upper-ocean salinity is not considered in detail here, owing to a simplified form of the freshwater forcing, which includes restoring. The interannual variability in the model is also very similar in all four experiments, and is therefore not discussed here.

3.1. Time-Mean Temperature and Mixed Layer Depth

The time-mean temperatures (Figures 2 and 3) and mixed layer depth (Figure 4) are noticeably different between experiments SMOOTH and DAILY_STRESS. These differences are caused by the addition of daily fluctuations to the wind stress in DAILY_STRESS. In contrast, the time-mean temperatures and mixed layer depth are very similar in DAILY, DAILY_WIND and DAILY_STRESS, and therefore sensitivity of these fields to the presence of daily fluctuations in the wind speed and air temperature/humidity is not significant. Section 3.1 thus mostly analyzes the effects of daily fluctuations in the wind stress and describes differences between SMOOTH and DAILY_STRESS. Effects of daily fluctuations in the wind speed and air temperature/humidity on the time-mean stratification are only briefly discussed at the end of this section.

In the austral summer (December–February), daily fluctuations in the wind stress cause cooling of the time-mean SSTs by 0.2–0.5°C (Figure 2a), which is significant for a time-mean response. Note, however, that the episodic cooling caused by an isolated storm can be of much greater amplitude; see section 1 for a summary of the studies in the Northern Hemisphere and see section 3.2. The largest storm-induced time-mean cooling is found in the western

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Pacific and Atlantic sectors in this study. Below the surface, the sensitivity of the time-mean temperature to daily fluctuations in the wind stress is stronger (Figure 3a); the ocean at the 40-m depth is warmer by $0.5$ to $1^\circ C$ in DAILY_STRESS than in SMOOTH everywhere in the Southern Ocean except in the region south of Australia. The described summertime changes in the temperature therefore weaken the stratification in the region. Changes in salinity (not shown) are also consistent with reduced stratification: Surface salinity increases whereas subsurface waters become fresher. It is noteworthy that the addition of daily wind stress anomalies improves simulation of the 40-m temperatures in the Southern Ocean, but increases errors in the simulated SSTs in several regions.

It is noteworthy, that the time-mean SSTs are less sensitive to the presence of daily forcing than are the subsurface temperatures in this ocean-only model. Seasonal means of the daily fluctuations in the wind stress/speed and air temperatures/humidities are almost zero by definition. If the upper-ocean temperatures depended only on the surface fluxes, and this dependence were linear, the seasonal-mean values of temperature would be identical in all runs. The reported differences therefore indicate the importance of other processes, such as advection and mixing. These processes play a larger role for the subsurface temperatures than for SST, making the response to daily forcing of the temperatures at the 40-m depth less linear than at the surface.

With the onset of autumn (Figure 2b), seasonal cooling of the surface is somewhat delayed in DAILY_STRESS compared to SMOOTH, and, as a result, the surface appears warmer in the Pacific and Atlantic sectors. The differences between the 40-m temperatures in the two experiments decrease toward the winter with deepening of the mixed layer (Figure 3b). In the austral winter and spring, the effects of daily fluctuations on SST (Figures 2c and 2d) and the 40-m temperatures (Figures 3c and 3d) appear negligible almost everywhere; this is explained by the very deep mixed layer [Price, 1981].

The oceanic mixed layer is sensitive to changes in the stratification caused by daily momentum fluxes.
Weaker stratification in the upper 50–70 m in the presence of daily winds manifests itself in a deeper mixed layer over most of the Southern Ocean in the austral summer (Figure 4a); the heat content (not shown) is also larger in DAILY STRESS than in SMOOTH. However, this storm-induced deepening of the mixed layer becomes spatially non-uniform in the following autumn (Figure 4b). In particular, the regions south of Australia and west of the tip of South America exhibit a shallower mixed layer in DAILY STRESS than in SMOOTH. The mixed layer in these regions is particularly deep; these are also the formation sites for the Antarctic Intermediate Water [McCartney, 1982]. In the austral winter, the mixed layer is noticeably thinner in DAILY STRESS than in SMOOTH in most of the Pacific sector and to the south of Australia (Figure 4c). The response of the very deep wintertime mixed layer to daily forcing in these regions is thus different from that in summer. In the spring, the mixed layer shoals; it is considerably deeper in DAILY STRESS than in SMOOTH over most of the domain (Figure 4d).

To understand the described response of the Pacific sector mixed layer to daily winds, we next consider seasonal changes in the stratification (Figure 5). The depth of the vertically homogeneous boundary layer, which is shown in these figures, is a penetration depth of the turbulent mixing by boundary layer eddies in the KPP model [Large et al., 1994]. This boundary layer in DAILY STRESS remains deeper than in SMOOTH during most of the year; this is a result of more vigorous stirring by daily winds in the former case. In the austral summer and early autumn, the mixed layer is shallow; the turbulent mixing penetrates almost to its bottom, and thus strongly affects the mixed layer depth. As Figures 5a and 5b show, the mixed and boundary layers are similar in depth, and the mixed layer reaches greater depths in DAILY STRESS. In the autumn, the mixed layer deepens faster in DAILY STRESS, and the seasonal surface cooling is delayed compared to SMOOTH.

Figure 3. Difference in seasonal-mean temperature at 40 m depth between experiments DAILY STRESS and SMOOTH, averaged over the length of the experiments for: (a) summer (December–February); (b) autumn (March–May); (c) winter (June–August); and (d) spring (September–November). Units are degrees; the contour interval is 0.2°C. Values under the ice are not shown.
By the end of autumn, the influence of wind stirring is limited, since it does not reach the base of the mixed layer (Figures 5b and 5c). Further deepening of the mixed layer is instead driven by convection. The density of the mixed layer is, however, lower in DAILY_STRESS than in SMOOTH, and the convective deepening in the former case is, therefore, impeded by stronger stratification at the base of the mixed layer. As a result, the SAMW is noticeably thinner in most of the Pacific sector in the presence of daily winds. By the end of winter, however, the depth of the mixed layer and the properties of the SAMW layer are very similar between the experiments with and without daily forcing. (SAMW layer thickness changes significantly in the zonal direction in the model with maximum values of 600–700 m in the southeast Pacific. Zonal-mean values shown in Figures 5 and 6 are smaller.)

To further study causes of the differences in the mixed layer structure between experiments with and without daily forcing, we consider a mixed layer heat budget. The equation for the evolution of the mixed layer heat content has the following form:

\[
c_{p} \rho \frac{\partial H}{\partial t} = -c_{p} \rho \int_{-H}^{0} \nabla \cdot (uT) dz + c_{p} \rho \omega T(-H)
\]

\[
- c_{p} \rho \frac{\partial H}{\partial t} T(-H) + \int_{-H}^{0} D_{xy} dz
\]

(1)
where $H(x, y, t)$ is the depth of the mixed layer, $T(x, y, z, t)$ is the temperature, $c_p$ is the heat capacity of seawater, and $\rho$ is the density of seawater. The first two terms on the right-hand side of (1), denoted by “ADV,” represent convergence of the horizontal advective heat flux and the vertical advective heat flux through the base of the mixed layer. The third term is the vertical heat flux due to the entrainment (“ENT’’), $F_s$ is the surface heat flux (“SURF”), $D_z$ is the heat flux through the base of the mixed layer due to vertical mixing and convection (“VMIX”), and $D_{xy}$ is the convergence of horizontal diffusive fluxes (“HMIX”).

[21] In Figure 6, zonal averages of the main terms in the mixed layer heat budget (equation (1)) are compared between experiments with (DAILY_STRESS) and without daily forcing (SMOOTH). In the austral summer (Figure 6a), zonal-mean surface heating is stronger by approximately 5 W m$^{-2}$ in DAILY_STRESS than in SMOOTH. The enhanced warming in DAILY_STRESS is explained by cooler SSTs (Figure 2a), since the daily fluctuations in the wind speed and air temperature/humidity are removed in this experiment. Moreover, the heat flux into the ocean is similar among all the experiments with daily forcing (DAILY, DAILY_WIND and DAILY_STRESS), and therefore addition of daily fluctuations in the wind speed and air temperature/humidity does not considerably affect the time-mean heat flux. Increased surface heating is balanced by intensified vertical mixing at the base of the mixed layer in DAILY_STRESS as compared to SMOOTH (Figure 6a); the heat uptake by the ocean below the mixed layer is also intensified by daily winds. The heat budget in summer is therefore essentially one-dimensional. In the austral autumn, the mixed layer begins to deepen, and the surface starts to
deviations of daily anomalies in SST decrease toward the pole (Figure 7a) (the standard deviation is computed from the anomalies which are defined as a difference between the unsmoothed and smoothed data); the daily standard deviations normalized by the SST (not shown), however, increase with latitude, which is more consistent with the increasing variability in the wind stress (Figure 1). Daily variability in both SST and 40-m temperature (Figures 7a and 8) is not uniformly distributed, and exhibits local maxima in the western part of the Indian sector and in the regions east of New Zealand and South America. It is noteworthy that the temperature variance is higher at the surface than below (Figures 7a and 8), whereas the sensitivity of the time-mean SSTs to daily forcing is considerably weaker than that of the 40-m temperatures (section 3.1). This fact confirms a conclusion of section 3.1, that the response of the temperature fluctuations in surface forcing is more linear at the surface than at depth.

[24] The daily fluctuations in temperature in experiment SMOOTH are the smallest among all the experiments; the standard deviations of daily temperatures in SMOOTH are on average less than 40% of those in DAILY (Figure 7b). The remaining fluctuations in SMOOTH can be partly attributed to intrinsic short-term variability in the upper-ocean temperature due to fluctuating currents. In particular, areas in the eastern Indian sector, west of the Drake Passage in the southeastern Pacific, show similar variability of temperature in experiments SMOOTH and DAILY (Figure 7b); the absolute values are, however, very small (Figures 7a). In addition, the smoothing procedure applied to all forcing fields in SMOOTH does not completely remove high-frequency fluctuations in the forcing.

[25] Addition of daily fluctuations to the wind stress leads to significant enhancement of the daily variability in SST at high latitudes (Figure 7c). Close to the ice edge, more than 80% of the standard deviation in DAILY is recovered in the experiment with daily momentum fluxes (DAILY_STRESS). Toward the north, the wind stress variance (Figure 1) and the importance of the wind stress anomalies in exciting SST variability both decrease; the standard deviation of SSTs at 45°S in DAILY_STRESS is approximately half of that in DAILY. The addition of daily fluctuations in the wind speed intensifies the SST variability (experiment DAILY_WIND), but the standard deviations north of 45°S remain less than 70% of those in DAILY (Figure 7d). Daily fluctuations in the air temperature and humidity are, therefore, important for exciting high-frequency temperature variability at midlatitudes. The importance of daily fluctuations in all forcing fields in exciting high-frequency variability in the 40-m temperatures is very similar to that in the SST.

[26] High-frequency variability in turbulent surface heat flux (latent and sensible) is very weak in DAILY_STRESS and SMOOTH (Figure 9), and standard deviations of the heat fluxes amount to less than 10% of those in DAILY. Variability in the turbulent surface heating increases when daily fluctuations in wind speed are included in DAILY_WIND, but remains well below the analogous variability level in DAILY (less than 50%). Clearly, most of the variability in the turbulent air-sea heat exchanges is attributable to the high-frequency variability in the air temperature and humidity. The increase in the heat flux variance
Figure 7. (a) Standard deviation (STD) of daily anomalies in SST for DAILY (units are °C). (b) Ratio of standard deviation of SST between SMOOTH and DAILY. (c) Ratio of standard deviation of SST between DAILY_STRESS and DAILY. (d) Ratio of standard deviation of SST between DAILY_WIND and DAILY.

Figure 8. Standard deviation (STD) of daily anomalies in temperature at the 40-m depth (t40m) for DAILY (units are °C).
in DAILY with respect to DAILY_WINDS has to be explained by stronger variability in the SST/air-temperature contrast. Since the SST variance is similar in both experiments (see above), the difference between daily SSTs and air temperatures can only increase when these two fields do not co-vary. SST anomalies are mainly generated by fluctuations in the winds (Figures 7b and 7c); it is therefore not surprising that they are rather weakly correlated with the air temperature anomalies. The corresponding correlation coefficients (not shown) are less than 0.35.

[27] Figure 10 shows a high-frequency part of the wave number-frequency spectrum of SST variability at 55ºS in experiment DAILY. The SST values are high-pass filtered with the cut-off period of 25 days, which approximately corresponds to the time it takes an air particle to travel around the latitude circle. The spectrum exhibits multiple maxima, most notably at the length scale of half the latitude circle (11,500 km) and the timescales of 15–20 days. These length scale and time periods correspond to a phase speed of approximately 7–9 m s⁻¹, which is similar to the speed of the atmospheric zonal flow in the model (6–8 m s⁻¹). The conceptual model presented in the next section is intended to interpret the spectrum in Figure 10 and to demonstrate how the atmospheric advection can affect spectral properties of the oceanic response to passing atmospheric storms.

3.3. A Conceptual Model of the Response of the Southern Ocean to Passing Atmospheric Storms

[28] We hypothesize that the main four factors that affect spectral properties of the response of the Southern Ocean to atmospheric storms are: oceanic advection of temperature anomalies, atmospheric advection of storms, duration and a zonal extent of each storm, and the timescale with which temperature anomalies are dissipated in a coupled atmosphere-ocean system. To incorporate all these ingredients, I developed a conceptual model of the upper Southern Ocean. Consider a zonal channel of constant depth, with a current of speed \( u_a \) that is independent of space and time. Evolution of a temperature anomaly \( \theta \), subject to anomalous atmospheric forcing \( f(x, t) \) and linear damping with a timescale \( \lambda \), is then described by the following equation:

\[
\frac{\partial \theta}{\partial t} + u_a \frac{\partial \theta}{\partial x} = -\lambda \theta + f(x, t).
\]  

(2)

[28] The model (2) is similar to stochastic climate models with linear feedbacks and advection; \( f(x, t) \) is often taken to be white noise forcing in time [e.g., Saravan and McWilliams, 1998; Weisse et al., 1999]. In the present study, the atmospheric forcing \( f(x, t) \) takes a form of a series of storms advected by the atmospheric zonal flow with a constant velocity \( U_{ia} \),

\[
f(x, t) = \sum_n C_n \exp \left( -a_n^2 \left( t - t_n^0 \right)^2 \right.
\]  

\[
- \gamma_n^2 \left( x - x_n^0 \right) - U_a \left( t - t_n^0 \right) \right)^2 \}.
\]  

(3)

[30] Each \( n \)th storm reaches its maximum strength at the time \( t_n^0 \) and location \( x_n^0 \), and decays in time and space on scales given by the parameters \( 1/\alpha_n \) and \( 1/\gamma_n \). We can anticipate that the structure of each storm affects the solution only on sufficiently short timescales, whereas the oceanic response on longer timescales is largely independent of the shape of each particular storm.

[31] Taking a Fourier transform \( \phi(k, \omega) = \int \int e^{i k x - i \omega t} \phi(x, t) dt dx \) of (2) leads to (see Appendix A)

\[
\theta(k, \omega) = \pi \sum_n C_n \exp \left( i (2k U_a - \omega) t_n^0 - ik x_n^0 \right) \exp \left( -k^2 \gamma_n^2 - \frac{(k U_a - \omega)^2}{4 \alpha_n^2} \right) \frac{i (k U_a - \omega) \lambda}{i (k U_a - \omega) + \lambda}.
\]  

(4)

[32] Two distinct advective timescales emerge from equation (4) for a given wavelength \( 2 \pi/k \). The first timescale, \( 2 \pi k U_a / \alpha_n \) is associated with the slow oceanic advection \( u_a \) which is on the order of 0.05 m s⁻¹. For the zonal scales between 500 km (characteristic for an atmospheric storm) and 11,500 km (half the latitude circle at 55ºS), this timescale is between 4 months and 7.3 years. A much shorter timescale \( 2 \pi k U_a / \gamma_n \) is given by the fast atmospheric advection \( U_a \) which is on the order of 7 m s⁻¹; this timescale is between approximately 20 hours and 20 days. These two regimes correspond to very different spectral properties of (4).

3.3.1. High-Frequency Regime: \( \omega \sim k U_a \)

[31] Assume for simplicity that all storms have the same temporal and spatial decay scales, \( 1/\alpha_n \) and \( 1/\gamma_n \). Then multiplication of (4) by its complex conjugate (denoted by an asterisk) leads to the following expression for the wave number-frequency spectrum \( F_{\theta}(k, \omega) \):

\[
F_{\theta} = \theta(k, \omega) \theta^*(k, \omega) = \left( \frac{2 \pi}{1/\gamma} \right)^2 \frac{2 \exp \left( -k^2 \gamma_n^2 - \frac{(k U_a - \omega)^2}{4 \alpha_n^2} \right)}{\omega^2 + \lambda^2} \cdot \sum_n C_n \gamma_n \cos \left( 2k U_a - \omega \left( t_n^0 - t_m^0 \right) - k \left( x_n^0 - x_m^0 \right) \right).
\]  

(5)
Figure 10. Wave number–frequency spectrum of the daily SST anomalies at 55°S in DAILY. The daily data are 25-day high-pass filtered. Spectrum power density is normalized by its maximum.

[35] To make further progress, assume that all storms are statistically independent of each other. For a sufficiently large number of storms, the sum of cosines on the right-hand side of (5) then approaches a value $A$ that is independent of $k$ and $\omega$:

$$F_{00} = A \left( \frac{\pi}{\alpha k} \right)^2 \exp \left\{ -\frac{k^2\gamma^2}{2\sigma^2} - \frac{(kU_\omega - \omega^2/2\alpha^2)^2}{\omega^2 + \lambda^2} \right\}. \tag{6}$$

It is noteworthy, that the response of the ocean to a single storm takes the same form.

[35] For each frequency $\omega$, spectrum (6) reaches maximum at

$$k = \frac{\omega}{U_\omega} \left\{ 1 + \frac{\alpha^2}{U_\omega^2 \gamma^2} \right\}^{-1}. \tag{7}$$

Equation (7) implies that for each time period $2\pi/\omega$ the spectrum (6) reaches maximum at the phase speed $\omega/k$ similar to the speed of atmospheric advection; for example, for $1/\alpha \approx 2$ days, $1/\gamma \approx 500$ km and $U_\omega \sim 7$ m s$^{-1}$, equation (7) gives $\omega/k \approx 1.16U_\omega$. Wavelengths predicted by (7) are generally consistent with those corresponding to the local maxima of the discrete wave number-frequency spectrum for the GCM-simulated SST variability (“GCM spectrum” henceforth). In particular, the GCM spectrum (Figure 10) displays several maxima for the time periods of 15–20 days and a wavelength of 11,500 km corresponding to the wave number two. For these time periods, Equation (7) predicts multiple maxima at the wavelengths of approximately 10,500–14,000 km. The maxima in the GCM spectrum shown in Figure 11 (top) for several time periods (4, 8, 12, and 16 days) are also in reasonably good agreement with equation (7). On the other hand, the GCM spectrum exhibits a distinct maximum at the wavelength of 11,500 km and time period of approximately 9 days, which the conceptual model does not capture.

[36] For a given $k$, spectrum (6) has a maximum only in a rather narrow range of wavelengths: between approximately 1600 km and 2700 km for $U_\omega \sim 7$ m s$^{-1}$ and $\alpha \sim 0.5$ day$^{-1}$ (see Appendix B). For the wavelengths in this range, the GCM spectrum shows multiple maxima between 10 and 20 days (Figure 11, bottom). Although the conceptual model predicts a similar range of time periods (9–26.5 days), its agreement with the GCM spectrum is hard to verify.

### 3.3.2. Low-Frequency Regime: $\omega \sim ku_\omega$

[37] Since the atmospheric advection $U_\omega$ is much faster than the oceanic speed $u_\omega$, in this regime we get that $\omega \ll ku_\omega$. In other words, the timescale $2\pi/\omega$ under consideration is much longer than the time required for a single storm to pass over a characteristic wavelength $2\pi/k$. The numerator in (4) then becomes independent of $\omega$. The wave number–frequency spectrum $F_{00}(k, \omega)$ is then given by

$$F_{00} = \frac{\Phi(k)}{(ku_\omega - \omega^2/2\alpha^2)^2 + \lambda^2}, \tag{8}$$

where $\Phi(k)$ is a function of $k$ only. From (8), it follows that for each wave number $k$ the wave number-frequency response has a maximum at $\omega = ku_\omega$. Weisse et al. [1999] arrived at the same conclusion when they considered response of the same ocean model (2) to white-noise atmospheric forcing. Note that (8) has been derived for the oceanic response with the phase speed $\omega/k$ much slower than the speed of the atmospheric advection. In this regime, low-frequency oceanic modes are excited by localized storms that quickly pass over the oceanic wavelength. The behavior of any individual storm in time, the number of storms, and cross-correlation of the storm events in time do not influence the time period that maximizes the spectrum at each wavelength. The preferred time period is instead determined by the oceanic advection. To summarize, we can identify two regimes depending on a characteristic timescale of oceanic response: a high-frequency regime in which the atmospheric advection sets the preferred time period and wavelength and a low-frequency regime with the preferred time period for a given wavelength determined by the oceanic advection.

### 4. Summary and Discussion

[38] In this study, we consider the effects of daily fluctuations in surface momentum and heat fluxes on the stratification in the Southern Ocean. Roles of daily anomalies in wind stress, in wind speed, and in air temperature and humidity are considered separately. Enhanced vertical mixing by the daily winds is the main cause of the differences in the time-mean upper-ocean stratification between the
experiments with and without daily forcing. In particular, the wind-generated turbulent mixing is stronger and penetrates deeper in the presence of daily momentum fluxes. Consistent with other studies of the oceanic response to atmospheric storms, the enhanced mixing in the austral summer and early autumn leads to cooler surface temperatures, warmer subsurface temperatures, and the deeper mixed layer. This study also demonstrates that the subsurface warming also enhances stratification immediately below the base of the mixed layer, which acts to slow wintertime convection, and results in the shallower mixed layer in the Pacific sector of the Southern Ocean.

[39] High-frequency variability in the upper-ocean temperature is the largest in the experiment with daily anomalies present in all four fields: wind stress, wind speed, and air temperature and humidity. Daily fluctuations in the surface wind stress and wind speed are the primary cause of short-term temperature variability in the high latitudes of the Southern Ocean. The importance of daily winds decreases toward the midlatitudes, where most of the temperature variability is attributable to daily fluctuations in the air temperature and humidity. The daily fluctuations in air temperatures and humidities lead to significant variability in the air-sea turbulent exchanges of heat at all latitudes, since SST and air temperature anomalies do not co-vary on short timescales in this study. However, the role of air temperature and humidity anomalies in exciting variability in SST and surface heat fluxes may be overestimated in an ocean-only model, which does not permit feedbacks of SST on the atmosphere.

[40] The high-frequency oceanic response to passing atmospheric storms is dominated by modes whose phase speed is determined by the speed of atmospheric advection, as suggested by a conceptual model presented in this study. The wavelengths and time periods that maximize the spectrum in the conceptual model are in good agreement with

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**Figure 11.** Wave number–frequency spectrum of the GCM-simulated daily SST anomalies is shown for (top) chosen values of time periods and (bottom) length scales. The dashed lines in the top row show the preferred wavelengths predicted by the conceptual model for the chosen time periods (equation (8)). The dashed lines in the bottom row show preferred time periods predicted by the conceptual model for the chosen wavelengths (equation (B2) in Appendix B).
those observed in the discrete wave number–frequency spectrum of the GCM-simulated SST variability. However, the conceptual model is unable to predict a full richness of the GCM spectrum. These disagreements are explained by the assumptions of the conceptual model, which is formulated in a zonal channel with constant atmospheric and oceanic velocities and does not take into account changes in the mixed layer depth.

[41] In particular, the GCM spectrum exhibits a preferred wavelength of half the latitude circle (approximately 11,500 km at 55°S). It is plausible that this preferred spatial scale of the variability in the Southern Ocean is predetermined by its geometry, perhaps by the width of the Pacific sector. In addition, the discrete GCM spectrum does not provide a good resolution at the long-wave part of the spectrum, which complicates the comparison. At the low-frequency part of the spectrum, the oceanic response is dominated by timescales determined by the slow oceanic advection. This result is independent of a number of the storms and of their shape. It is noteworthy that the same preferred timescale emerges from the spectrum of the oceanic response to stochastic atmospheric variability [Saravan and McWilliams, 1998; Weisies et al., 1999].

[42] This study describes a forced response of the ocean to atmospheric daily forcing. While not allowing a feedback of the SST on the air temperature, humidity, and neutral winds, this approach guarantees that the atmospheric variability remains close to the observed. The NCEP re-analysis data used in this study have a tendency to underestimate high-wind events, when compared with satellite data [Yuan, 2004]. The model described here, therefore, could underestimate the sensitivity of the temperature to high-frequency fluctuations in winds. Estimates of the mixed layer depth can also be quantitatively affected by the coarse vertical resolution. The lack of a dynamic component in the sea-ice model restricted the analysis to the ice-free areas, thus excluding effects of the high-frequency forcing on convection and formation of bottom water at the Antarctic coast. Finally, diurnal fluctuations in atmospheric fluxes are not considered in this study. The results emphasize a need for accurate presentation of the high-frequency variability in SST, winds, and air temperature and humidity in climate models.

Appendix A

[43] Taking the double Fourier transform \( \tilde{\phi}(k, \omega) = \int \int e^{i k x} e^{-i \omega t} \phi(x, t) dx dt \) of equation (2) leads to

\[
\hat{\phi}(k, \omega) = \left( -i \omega + i k u_0 + \lambda \right) \tilde{\phi}(k, \omega) = \tilde{f}(k, \omega). \tag{A1}
\]

[44] The right-hand side of (A1) can be expressed as

\[
\tilde{f}(k, \omega) = \int \int e^{i k x} e^{-i \omega t} \sum_n \int d t \int C_n \exp \left\{- \omega^2 (t - t_0)^2 - \frac{\gamma^2}{U_a} \left( t - t_0 - x + x_0 \right)^2 \right\}
\]

\[
\times \exp \left\{ - \frac{\omega^2}{4 \gamma^2 U_a^2} \left( t - t_0 - x + x_0 \right)^2 \right\}
\]

\[
= \sum_n \int d x \int d \omega \int h_n(t) h_n' \left( t - \frac{x}{U_a} \right) dt
\]

\[
= \sum_n C_n U_a \hat{h}_n(k U_a) g_n(k U_a), \tag{A2}
\]

where now \( \hat{\phi}(\eta) = \int \phi(y) e^{i \eta y} dy \). In derivation of (A2), we made use of the convolution theorem. The functions \( h(y) \) and \( g(y) \) are

\[
h_n(y) = \exp \left\{ - \frac{\omega^2}{4 \gamma^2 U_a^2} \left( y - t_0 - x_0 + x \right)^2 \right\}
\]

\[
g_n(y) = \exp \left\{ - \frac{\gamma^2 U_a^2}{4 \omega^2} \left( y - t_0 + x_0 - x \right)^2 \right\}. \tag{A3}
\]

[45] The Fourier transforms of these functions take the following form:

\[
\hat{h}_n(\eta) = \frac{\sqrt{\pi}}{\omega} \exp \left\{ - \frac{(\eta - \omega)^2}{4 \omega^2} + \frac{i \eta \omega}{\omega} \right\}
\]

\[
\hat{g}_n(\eta) = \frac{\sqrt{\pi}}{\gamma U_a} \exp \left\{ - \frac{\eta^2}{4 \gamma^2 U_a^2} + \frac{\eta \omega}{\gamma U_a} \right\} \left( \frac{\omega}{\gamma U_a} - \frac{\omega_0}{U_a} \right). \tag{A4}
\]

Appendix B

[46] For a given \( k \), frequency \( \omega \) that maximizes spectrum (7) is given by the following cubic equation:

\[
\omega^3 - \omega^2 k U_a + \omega(\lambda^2 + 2 \alpha^2) - k U_a \lambda^2 = 0. \tag{B1}
\]

[47] Equation (B1) can be further simplified after noting that the timescale \( 1/\lambda \) due to damping of SST anomalies to atmospheric feedbacks is typically longer than the timescale associated with the atmospheric advection, \( 1/k U_a \). In particular, for the wavelength of 5000 km, this advective timescale is approximately 6 days. The Haney-type boundary conditions [Haney, 1971] lead to a damping scale of about 60 days for a 50-m-thick surface layer; radiative cooling corresponds to a significantly longer damping scale. Oceanic vertical turbulent diffusion for a 50-m-thick surface layer and turbulent diffusivity of \( 10^{-3} \) m² s⁻¹ corresponds to damping with a timescale of approximately 30 days. Therefore it is instructive to consider oceanic responses with timescales longer than those associated with the linear damping: \( \omega \gg \lambda \). Spectrum (7) then reaches maximum at

\[
\omega = k U_a \left( \frac{1}{2} - \frac{1}{2} \sqrt{1 - \frac{8 \alpha^2}{k^2 U_a^2}} \right). \tag{B2}
\]

which has a physically meaningful solution only for the wavelengths \( 2\pi/k \) that (1) make the atmospheric advective timescale, \( 1/k U_a \), sufficiently short: \( 1/k U_a < \frac{1}{2} / \sqrt{\lambda} \) and (2) keep \( \omega \) of the same order as \( k U_a \). For \( U_a \sim 7 \) m s⁻¹ and \( \alpha \sim 0.5 \) day⁻¹, the corresponding wavelengths \( 2\pi/k \) lie between 1600 km (which makes \( \omega > 0.1 \) k U_a) and 2700 km; the time periods \( 2\pi/\omega \) given by (B2) are approximately 26.5 and 9 days.

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References