A Reexamination of the Mechanism of the Semiannual Oscillation in the Southern Hemisphere*

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ABSTRACT

The cause of the semiannual oscillation (SAO) at middle and high southern latitudes, proposed by van Loon, is reexamined using observations and general circulation model (GCM) simulations. The model results and the more recent observed data [sea surface temperature (SST), ocean heat storage, temperature profiles in the upper ocean, and atmospheric transient eddy momentum and heat fluxes] support van Loon's original hypothesis that the mechanism involves the different annual cycles of temperature between the Antarctic polar continent and the surrounding midlatitude southern oceans. A strong semiannual oscillation is noted in the observed atmospheric transient eddy momentum flux in southern midlatitudes corresponding with the two times of year when the circumpolar trough around Antarctica is most intense. The products of the dynamical coupling of ocean and atmosphere—the annual cycle of SST near 50°S and associated ocean heat storage—are important to the amplitude and phase of the SAO in the atmosphere. GCM simulations are analyzed to provide insights into the consequences of changing elements of the ocean forcing near 50°S. The GCM simulation with the specified annual cycle of SSTs has the correct phase of the SAO but reduced amplitude. The model with a simple mixed-layer ocean (shallow fixed depth with no dynamics) produces an altered annual cycle of SSTs and ocean heat storage at 50°S and a similarly altered SAO as a consequence. These model results, along with the observed upper-ocean temperature profiles and heat storage values, suggest that changes in the annual cycle of SST and ocean heat storage near 50°S could lead to a modulation of the observed SAO and affect its role in the El Niño–Southern Oscillation and the Indian monsoon.

1. Introduction

The semiannual oscillation (SAO) in the Southern Hemisphere (SH) occurs throughout the depth of the troposphere and is characterized at the surface by an expansion and weakening of the circumpolar trough of low pressure surrounding Antarctica from March to June and September to December and by a contraction and intensification from June to September and December to March. This twice-yearly pulsation of the circumpolar trough is associated with similar fluctuations of tropospheric temperature gradients, heights, pressure, and winds at middle and high latitudes in the SH (e.g., van Loon 1972; van Loon and Rogers 1984). At the surface, the strongest westerly winds occur during March and September south of about 50°S and during June and December north of about 50°S. The SAO explains more than 50% of the observed variance of sea level pressure over vast areas of the SH middle and high latitudes (Xu et al. 1990).

The SAO has been widely known at least since the 1930s. Van Loon (1967) summarizes the early literature on the SAO. Since van Loon (1967), several papers have described further aspects of the observed SAO (e.g., van Loon 1972; Hsu and Wallace 1976; White and Wallace 1978). Van Loon and Rogers (1984) updated the original van Loon (1967) results by noting that the SAO was present in unfiltered data in individual years and by showing some aspects of the interannual variation of the SAO. More recently, the SAO has been documented in ocean currents in the southern extratropics (Large and van Loon 1989) and in ocean wind stress at those latitudes (Trenberth et al. 1990).

The SAO has also been described in general circulation models (GCMs) of the atmosphere (Weickmann and Chervin 1988; Xu et al. 1990; Kitoh et al. 1990). The current generation of spectral GCMs does not simulate well some aspects of the SAO, but a gridpoint GCM described by Kitoh et al. (1990) appears to be somewhat more successful.

Even though numerous papers have been written on the description of the SAO in observations and models, none since the van Loon paper has addressed the mechanism that maintains it or how part of that mechanism may vary to produce interannual variability in

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the SH. The purpose of this paper, then, is to reexamine van Loon's original hypothesis of the cause of the SAO in light of more recent data and to use GCM simulations to begin to address some of the possible external forcing processes: in the present case, the role of SST and ocean heat storage in the circumpolar ocean near 50°S. If the mechanism is better understood, anomalous forcings that could cause the SAO to affect interannual variability could also be monitored.

It was once thought that the SAO was peculiar to the far southern reaches of the SH and, as such, of little interest outside the sub-Antarctic. However, van Loon (1984) and van Loon and Shea (1985, 1987) showed that a critical time in the evolution of an extreme in the Southern Oscillation (SO) (May–June–July, or MJJ) coincided with the time of year in the mean when the SAO was seen in the northward expansion of the sub-Antarctic circumpolar trough and with enhancement of midlatitude wavenumber three. The circumpolar trough itself is a climatological feature characterized by the region of decay of baroclinic disturbances that form much farther north and spiral in toward Antarctica (van Loon 1972). The two main paths these disturbances take originate in the subtropical Pacific and Atlantic oceans (Trenberth 1981a). Therefore, a stronger-than-normal trough is associated with pressure falls northward into the subtropics and manifested by a weakened subtropical high in the Pacific, reduced trade winds, and increased sea surface temperatures (SSTs) in the equatorial Pacific in a warm event. Conversely, the failure of the trough to amplify in the Pacific is associated with strengthening of the subtropical high and trades and decreasing equatorial Pacific SSTs in a cold event. Meehl (1987, his Fig. 16) notes similar relationships in the Pacific and also draws attention to the possible role of the SAO in the southern Indian Ocean during the initiation of a relatively strong or weak Indian monsoon. Other linkages between the tropics and mid- and high-latitude circulations in the context of the El Niño–Southern Oscillation (ENSO) are shown by Meehl (1988, 1990).

Other than these documented associations, the cause–effect relationships still remain unclear. Yet, there is a reasonable possibility that the SAO has links to the circulation at much lower latitudes than its commonly perceived domain, and, therefore, understanding the cause of the SAO could clarify its role in ENSO and in interannual variability in the Indian monsoon.

2. Data and model simulations

This study uses the European Centre for Medium-Range Weather Forecasts (ECMWF) observed data analyses from 1979 to 1985 described by Trenberth and Olson (1988). Southern Hemisphere temperature data from the Australian analyses cover the period from May 1972 to May 1984. Southern Hemisphere sea level pressure (SLP) data represent two periods—the first is from the South African analyses for January 1951 to December 1958, and the second is covered by the Australian analyses from May 1972 to May 1984.

Observed SSTs are from three sources: 1) atlas values of monthly means compiled by Alexander and Mobley (1976), 2) monthly mean SSTs from COADS (Comprehensive Ocean Atmosphere Data Set) for the period 1950–79 taken mainly from ships of opportunity, and 3) monthly means from the atlas put together by Levitus (1982). Vertical temperature profiles are from ship soundings and are compiled in Gordon et al. (1986).

The first GCM simulation is forced with the observed specified annual cycle of SSTs. Called SPEC SST and described by Meehl (1989), a global, spectral (R15, about 4.5° latitude by 7.5° longitude) atmospheric GCM is run with the annual cycle of observed monthly mean SSTs and sea ice from Alexander and Mobley (1976). Averages are from the last three years of a 5-year run. These model data were also used in the model intercomparison of the SAO by Xu et al. (1990). The atmospheric model has realistic geography, nine layers in the vertical, and includes calculations for soil moisture, snow cover, cloudiness—radiation, convective adjustment, and seasonal cycle.

A second GCM simulation is run with the same atmospheric model coupled to a single, fixed-depth (50-m), mixed-layer ocean (called MIX1) first described by Washington and Meehl (1984). The SSTs in this model are computed from surface energy balance and simple heat storage in the 50-m deep slab ocean and are not corrected to compensate for the omission of meridional heat transport. Results are averaged over a 3-year period at the end of a 12-year integration. This was preceded by two earlier periods run with an accelerated annual cycle to bring the model more quickly to equilibrium (Washington and Meehl 1984).

Details of the air–sea interaction of both model versions in the tropical Indian and Pacific sectors are given in Meehl (1989). Meehl and Washington (1985) show the systematic SST errors simulated with the simple mixed-layer model coupled to the atmospheric model. Meehl and Washington (1990) have also run this model with a revised snow–sea ice albedo scheme with results for the SAO similar to those described below.

3. The semiannual oscillation in the Southern Hemisphere

The SAO is evident in the four long-term monthly mean maps of observed SLP (Fig. 1). The trough of minimum SLP (stippling in Fig. 1) is farthest south and deepest in March and September (Figs. 1a,c) and farthest north and weakest in June and December (Figs. 1b,d). As noted earlier, the trough of minimum SLP is indicative of the areas where mature cyclones decay near Antarctica. Associated with the seasonal intensity
and latitudinal movement of this band of minimum SLP are large regions between about 25° and 60°S where seasonal rise and fall of SLP is influenced by the collective tracks of individual cyclones. Fluctuations of the circumpolar trough are thus indicative of changes in a much larger system of cyclonic activity. The net effect is pressure changes over extensive areas of the SH and consequential changes of SLP in the circumpolar trough.

For example, in Fig. 2a (reproduced from van Loon and Rogers (1984)) the expansion of the trough from March to June (Figs. 1a,b) is manifested by pressure
falls over the three oceans between about 25° and 60°S and pressure rises poleward of about 60°S. This is indicative of an equatorward expansion of cyclonic activity over the southern oceans. Similar pressure changes over ocean regions occur when the trough expands from September to December. Opposite seasonal changes take place when the trough contracts from December to March and June to September. Expansion and contraction of the mean subantarctic trough, then, are associated with expansion and contraction of associated cyclonic activity over ocean areas of the Southern Hemisphere extending from the high latitudes to the tropics. Over the subtropical continents, the first harmonic dominates with pressure rises from December to June and falls the other half of the year (van Loon and Rogers 1984, Fig. 1).

Figure 2b suggests a possible link of the seasonal cycle of the SAO to interannual variability in the tropics (e.g., van Loon and Shea 1985). The pattern of change of SLP in MJJ during the inception of warm events is similar to the long-term mean changes of SLP as the trough expands during that season. That is, enhanced expansion of the trough in most of the Pacific and Indian ocean sectors is manifested by pressure falls between about 40° and 60°S, with negative pressure anomalies extending into the tropics in the Pacific sector. Accentuation of the seasonal rise of SLP over Australia in Fig. 2a is also evident in Fig. 2b for the warm-event composite anomalies. Possible links of the seasonal evolution of the SAO and interannual events in the tropics will be discussed further in section 8.

As mentioned earlier, the SAO is evident throughout the depth of the troposphere. For example, van Loon (1972, Fig. 5.15) shows that the amplitude of the second harmonic in zonal mean geostrophic wind actually increases with height in the troposphere. Van Loon (1967) noted a twice-yearly intensification of the midtropospheric temperature gradient between 50° and 65°S associated with the SAO in SLP.

A useful index of the SAO, first used by van Loon (1967), is the difference of the zonal mean 500-mb temperature between 50° and 65°S. He postulated that this index, indicative of the state of the SAO, was associated with the forcing of the phenomenon.

The idea is that the twice-yearly intensification of the midtropospheric temperature gradient between the ocean-dominated latitude of 50°S and the polar continental latitude of 65°S is associated with a twice-yearly increase of baroclinicity and storm activity due to changes in intensity of the circumpolar trough.

The mechanism, as posed by van Loon, arises from the different slope of the annual curves of temperature over the midlatitude ocean and the polar continent and their nearly equal amplitudes in the midtroposphere. As seen in Fig. 3, this circumstance produces an intensification of the temperature gradient twice a year in the midtroposphere and can be linked to the phase of the annual cycles at the surface being reflected higher in the troposphere. Heat storage in the ocean delays the summer temperature maximum and winter minimum at the ocean latitudes near 50°S, while the "coreless winter" over Antarctica is associated with no well-defined midwinter temperature minimum there. Instead, the temperatures drop rapidly in autumn, then gradually continue to decrease in the mean throughout the winter (with a slight increase at some locations in midwinter), and culminate with an early spring minimum and a rapid rise in early summer [shown by van Loon (1967) and discussed further in Fig. 6]. In contrast, stations at comparable latitudes in the Northern Hemisphere (NH) show sharp midwinter temperature minima.

Van Loon used widely dispersed stations in the SH to describe the character of the SAO. As noted in subsequent studies, because of its dominant and geographically widespread influence on the circulation of the SH, even a few stations can describe the SAO. But van Loon had virtually no additional observed data in hand in 1967 to prove or disprove his proposed mechanism. Concerning the first element of the mechanism, little was known about the annual cycle of SST or ocean heat storage near 50°S at that time. Instead, he showed surface energy balance atlas data to infer the role of ocean heat storage. Northern Hemisphere weather-ship data were used as analogues for the qualitative character of the SH SST. The other part of the mechanism, the "coreless winter" over Antarctica, was also inferred (from station data) to be partly caused by midwinter incursions of relatively warm maritime air transported into high latitudes by the amplification of the circumpolar trough into the Indian, Pacific, and Atlantic oceans in early winter. The increased longitudinal pressure contrast between the oceans and the lower-latitude continents would amplify the meridional flow southward from middle to high latitudes often enough to delay a winter temperature minimum until early

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**Fig. 2.** (a) Difference in SLP, long-term mean, June minus March; the sign of the difference signifies the seasonal change in SLP as the circumpolar trough system expands into the three Southern Ocean regions from the February–March–April (FMA) season to the May–June–July (MJJ) season (from van Loon and Rogers 1984). (b) Anomalies of SLP for the MJJ season for warm events minus the long-term mean during the inception of composite warm events. Arrows indicate the direction of the geostrophic wind anomaly. This is the season when the circumpolar trough system is expanded, and the similarity of anomalies with part (a) in the Pacific and Indian sectors suggests that the expansion of the circumpolar trough system is connected with the inception of warm events as an enhancement of the seasonal cycle of the SAO at this time of year (from van Loon and Shea 1987).
4. Observations and model simulations of the SAO

As noted by van Loon (1967), a useful index of the SAO is the 500-mb temperature difference, 50°S minus 65°S. Figure 4 shows this index for observations and the two model simulations. Note in the observations (Fig. 4a) the dominant second harmonic (59% of the variance versus 19% for the first harmonic), with intensification of the gradient occurring in the transitional seasons of March and September, the times of year when the circumpolar trough is deepest (Fig. 1). Significance testing of this phenomenon is difficult with the abbreviated length of the observed record, but van Loon and Rogers (1984) show the SAO in individual years and note that there is a manifestation of it in every year that they examined (1957, 1958, and 1972–1979).

Van Loon and Rogers (1984) have shown that the geographical distribution of the observed SAO in the SH is zonally uniform in phase, with maxima in the three ocean sectors. In the zonal means, the SAO is enhanced because the first harmonic tends to cancel itself in the averaging due to the heterogeneous phase (van Loon and Rogers 1984).

Weickmann and Chervin (1988) and Xu et al. (1990) have shown that the National Center for Atmospheric Research (NCAR) model-simulated SAO is much less uniform in space. The Pacific sector in the model has the most realistic simulation of the SAO in SLP. In particular, the latitudinal position and phase of the SAO in the 500-mb temperature gradient are well simulated in that region (Xu et al. 1990). In the present paper, zonal means calculated over all longitudes will be used to identify the operative mechanisms and to indicate the processes taking place in the model integrations and the observed system. The purpose of including the model simulations is to provide insight into the role of the various elements of the SAO deduced from observations.

The two model simulations are similar in that both have a weaker-than-observed zonal mean 500-mb temperature gradient between 50° and 65°S. Yet, the SPEC SST simulation (Fig. 4b) shows a dominant, though small, second harmonic (46% explained variance compared to 38% for the first harmonic). The MIX1 simulation in Fig. 4c shows a more dominant first harmonic compared to the second harmonic (30% for the second harmonic versus 60% for the first harmonic).
monic), with the maxima of the second harmonic occurring about half a month later compared to the SPEC SST case.

Even though the SPEC SST model is qualitatively simulating the SAO, it is reasonable to question whether any model has effectively demonstrated the link between the intensity of the 500-mb temperature gradient between 50° and 65°S, the position and intensity of the circumpolar trough, and baroclinic eddy activity. This linkage has been shown with a version of the SPEC SST model by Meehl and Albrecht (1988). A large intensification of the 500-mb temperature gradient (their Fig. 1b) in the SH midlatitudes (forced by an alteration of the convective parameterization) produced a dramatic deepening and southward movement of the circumpolar trough of low pressure (their Figs. 2a,b) and an increase in baroclinic eddy activity in that model (their Fig. 8). That study suggests that the correct mechanisms are in place in the model to reproduce the SAO if the amplitude of the temperature gradient is correct.

Xu et al. (1990) note that the semiannual wave of 500-mb temperature gradient in the model best resembles the observations in the Pacific sector with one maximum at 55°S (March–September phase) and one at 35°S (June–December phase). This pattern is indicative of the expansion and weakening of the trough in June–December and the contraction and intensification in March–September, as noted in Fig. 1. That the model displays these features best in the Pacific sector and not in the Indian or Atlantic is partly manifested by the small-amplitude zonal mean values in Fig. 4b.

Figure 5 depicts the annual curves of zonal mean 500-mb temperature for the observations and the model simulations. At 50°S, the SPEC SST follows the observations very well, with both showing maxima in February and minima in August. But the SPEC SST model is colder than the observations by about 3°C throughout the year. The MIX1 model (with the simple slab-ocean mixed layer and computed SSTs) shows quite a different annual cycle of 500-mb temperature at 50°S with a minimum in September and a maximum in March.

At 65°S (Fig. 5b), the SPEC SST again follows reasonably well the observed annual cycle of 500-mb temperature. In contrast to 50°S, the SPEC SST simulation at 65°S is within 1° to 2°C of the observed values throughout the year. The MIX1 simulation has larger quantitative errors from January to July (about 3°–5°C too warm). Yet, the general character of the seasonal cycle is qualitatively reproduced with a maximum in January and a minimum in early spring.

It was noted earlier from the Meehl and Albrecht (1988) results that the amplitude of the 500-mb temperature gradient in the southern midlatitudes is very important in the intensification and poleward movement of the circumpolar trough and in the increase of baroclinic eddy activity. The SPEC SST model 500-mb temperatures in Fig. 5 show that the shape of the curves of the annual cycle is mostly correct at 50° and 65°S. The main contributor to the reduced amplitude of the gradient associated with the weaker SAO shown in Fig. 4b is the colder-than-observed 500-mb temperature at 50°S in Fig. 5a. An improved tropospheric temperature structure, such as that associated with a revised convective scheme demonstrated by Meehl and Albrecht (1988), could contribute to an improved simulation of the amplitude of the SAO in the model.

5. Elements of the mechanism near 50°S

One part of van Loon's proposed mechanism involves the annual cycle of SST and heat storage at the ocean-dominated latitude of 50°S. Figure 6a shows SST at 50°S from 1) observed atlas values of Alexander and Mobley (1976) used in the SPEC SST model, 2) observed monthly mean SSTs from COADS for the period 1950–79, 3) observed monthly mean SSTs from
from the insolation maximum in December indicative of heat storage in the 50-m slab ocean (see Meehl and Washington 1985), but the minimum occurs in October. The amplitude of the annual cycle is too large, which suggests that the 50-m mixed layer is too shallow. Similar characteristics of this annual cycle of SST are also portrayed at 500 mb in Fig. 5a for the MIX1 model. The reflection of the shape of the annual curve of SST at 50°S in the 500-mb temperatures for the observations and both model simulations (one from a model simulation with quite a different annual cycle of SSTs) agrees with van Loon’s idea about the workings of the SAO—that the principal forcing of the SAO is from the surface.

Another element of van Loon’s mechanism has to do with the role of surface energy balance and ocean heat storage in contributing to the annual cycle of SSTs at 50°S. Surface energy balance can be defined:

\[ [I_\downarrow - I_\uparrow] + [(Q + q)(1 - \alpha)] - H - LE - G = 0, \]

where \( I_\downarrow \) is incoming infrared radiation, \( I_\uparrow \) is upward infrared from the surface, \( Q + q \) is the sum of direct and diffuse solar radiation, \( \alpha \) is albedo, \( H \) is sensible heat, \( LE \) is latent heat, and \( G \) is net heat flow through the surface.

In the ocean, \( G \) has two major components:

\[ G = Q_v + Q_s, \]

where \( Q_v \) is the rate of change of local heat content, usually referred to as “heat storage,” and is the dominant term in the upper ocean heat budget (Large et al. 1986), and \( Q_s \) is the sum of vertical advection, horizontal advection, and vertical diffusion (Large et al. 1986).

The quantity \( Q_v \), heat storage, is defined as

\[ Q_v(h, t) = \frac{\partial}{\partial t} \int_0^h c_v \rho T(z) dz, \]

where \( Q_v \) is the heat storage in a column of depth \( h \) and unit surface area, \( c_v \) is the specific heat of seawater, \( \rho \) is the water density, and \( T(z) \) is the temperature profile.

Figure 7 shows observed ocean heat storage \( Q_v \) from Levitus (1987) at 50°S compared to heat storage computed from the MIX1 model. Levitus (1987) integrated temperature soundings to a depth of 275 m to compute heat storage. Levitus notes that June to September is the most poorly observed period (compared to the rest of the months that are somewhat less poorly observed) in the Southern Ocean south of about 40°S. Levitus (1984) also points out that some regions of the Southern Ocean have contributions to heat storage from depths in excess of 275 m.
The observed heat storage values in Fig. 7 show a transition from heat loss to heat gain in late August and a switch from heat gain to heat loss in March (the two zero-line crossings). Note in Fig. 6 that the observed SST minimum is in July (Alexander and Mobley), August (COADS), or August–September (Levitus) and that the maximum is in March.

For late summer (March), the SST and heat storage data are consistent with the switch from heat gain to heat loss associated with the maximum of SST. In late winter, the heat-storage transition also occurs around the time of the SST minimum. The disagreement of the three observed SST sources as to the exact month of the SST minimum at 50°S (all in late winter but either in July, August, or September) is probably a reflection of the paucity of data during that time of year at that latitude. This uncertainty would also be manifested in the heat storage calculations by Levitus (Fig. 7).

A consistency check of the relationship between SST and heat storage values can be made by comparing calculations at latitudes that are better sampled. For example, Meehl (1984) uses data from weather ship Victor (34°N, 164°E) for calculations of ocean heat storage, taking into account a full integration of the annual cycle of monthly mean vertical temperature profiles versus a calculation that depends only on SST. Meehl shows that the calculation of ocean heat storage dependent only on SST is seriously in error at that location, particularly in autumn. Victor, located in a regime with a shallow summer mixed layer and a nearly isothermal winter temperature profile, is representative of many regions of the northern midlatitude oceans. This calculation is most inaccurate in autumn because heat in the shallow summer mixed layer is distributed down into the thermocline at that time of year during episodic cooling associated with passing storms. As noted by Large et al. (1986), entrainment and vertical diffusion can result in a decrease of SST with heat content remaining the same or even increasing at that time of year in such a regime. Therefore, the maximum and minimum of SST do not correspond with the zero-line crossings of heat storage since the column is not well mixed at certain times of the year.

If, on the other hand, the extremes of the annual cycle of SST occur nearly coincident with the zero crossings of heat storage, the ocean column must be very well mixed to considerable depth throughout the year. The ocean column, therefore, does not undergo the shallowing of the summer mixed layer with the transition to a nearly isothermal profile in winter (i.e., no episodic cooling in autumn), and the temperature profiles must warm and cool nearly uniformly to considerable depth year-round. Several investigators have suggested that this is the case near 50°S, using limited data at that latitude (e.g., U.S. Naval Oceanographic Office 1957; van Loon 1966). Colborn (1975) shows vertical temperature profiles for the southern Indian Ocean (reproduced in Fig. 8a). The regime of shallow summer mixed layer and isothermal winter temperature profile for the latitudes near 35°–40°S is reminiscent of the NH midlatitude oceans discussed above and used in the calculations of Meehl (1984). These regions are characterized by cooling episodes in autumn [Large et al. (1986) for the NH; Large (1987) for the SH] that often result in a decrease of SST with heat content in the column mostly unaffected.

However, the region near 50°S, just north of the polar front in the ocean, is encompassed by the water mass defined by Sverdrup et al. (1942) as the “subantarctic upper water.” As described by Colborn (1975), “...the general lack of significant thermal gradient formation during the warming period produces only a weak density gradient in the upper layers that is insufficient to prevent deep vertical convections at the initiation of winter cooling during May–June. Surface water cools and sinks to relatively great depths....”

Figure 8b compares the summer and winter temperature profiles from Colborn (1975) for the region near 50°S in the southern Indian Ocean with the profiles from latitudes just to the north in Fig. 8a. As implied by the relationship between SST and heat storage discussed above, the column is very well mixed and the entire temperature profile warms and cools to at least 500 m (the limit of the vertical extent shown by Colborn) during the seasonal cycle at 50°S. Subsurface temperature profiles also have been compiled by Levitus (1982). With these data, Levitus (1984) shows that the region near 50°S is very well mixed year-round, especially compared to similar latitudes in the NH. For first-guess fields for the analyses, however, the Lev-
Fig. 8. (a) Temperature profiles (°C) for winter (dashed line) and summer (solid line) for area in the southern Indian Ocean near 40°S from Colborn (1975). (b) Same as (a) except for area near 50°S. (c) Temperature profiles (°C) from the southern Indian Ocean taken from Gordon et al. (1986); dashed line is the winter profile (July 1972, 49°S, 85°E), and solid line is the summer profile (December 1969, 50°S, 132°E). (d) Temperature profiles (°C) from the southern Pacific Ocean taken from Gordon et al. (1986). Dashed line is the winter profile (August 1964, 50°S, 160°W), and solid line is the summer profile (December 1971, 55°S, 170°E).

- Itus monthly means use either seasonal means or, where data are sparse (certainly during winter in the southern oceans), the annual means of existing observations. This could introduce a bias toward the annual mean profile at poorly sampled latitudes or depths. A more accurate description of the seasonal characteristics of temperature profiles at 50°S is still probably best obtained by examination of individual ship soundings.

Gordon et al. (1986) have compiled a number of these into representative longitudinal sections. Figures 8c and 8d show winter and summer profiles near 50°S for the Indian and Pacific oceans, respectively. Since these
soundings and sections are from different ships, different years, and somewhat different longitudes, this comparison must remain largely qualitative. But these profiles confirm that, at locations near 50°S, the ocean column is very well mixed to depths of about 1000 m year-round. This is consistent with the extremes of SST coinciding with the zero-line crossings of heat storage noted in Figs. 6 and 7.

Such a well-mixed water column at 50°S has implications for interannual variability associated with SST anomalies. For example, near 40°S (Fig. 6a), a 0.5°C increase of SST (distributed proportionately through the column to the level of no seasonal change during summer at about 400 m) results in an increase in total heat content of the column of only about 2%. But for the same increase of SST at 50°S (distributed proportionately through the column to the level of no seasonal change at around 1000 m during summer as shown, for example, in Fig. 8c), the total heat content in the column increases 5%, more than twice the increase of heat content for the same SST anomaly at 40°S. This means that a small SST anomaly near 50°S may be associated with a relatively large change in heat content in the ocean column because the ocean is well mixed in those regions. Such an anomaly would presumably be long-lived since the heat is mixed to considerable depth.

Another important consideration in the determination of SST near 50°S is the net surface heat balance [the sum of terms 1–4 in Eq. (1)]. Van Loon (1967) used surface heat balance data from the Budyko atlas (Budyko 1963) to infer changes in SST near 50°S. Those data show seasonal extreme values of around −110 W m⁻² occurring in late June–early July and about +90 W m⁻² in late December. More recent compilations of observed net surface heat flux (Esbensen and Kushnir 1981; Oberhuber 1988) show roughly similar magnitude and timing of the seasonal cycle, taking into account the presumed uncertainties of such calculations made with limited data. For Eq. (1) to balance, G (heat flow into the ocean that is largely heat storage at 50°S, as discussed above) must compensate for this net energy input at the surface (terms 1–4). Figure 7 shows that, indeed, maximum heat loss is in June and maximum heat gain is in December with extremes of around ±100 W m⁻². These values coincide roughly (within the margins of error of the observations) with the net surface heat flux values. More exact calculations of such balances must await more extensive and accurate data.

Van Loon (1967) noted that two characteristics of the annual cycle of SST at 50°S contributed to the mechanism of the SAO. One is the delay of the seasonal extremes of SST about two to three months from the seasonal extremes of net surface heat flux. The other is the faster rate of fall of SSTs after the seasonal maximum than the rate of rise after the seasonal minimum. Within the constraints of the limited data, it is readily apparent that ocean heat storage plays critical roles in both these processes. Also, because the ocean is well mixed at these latitudes, a very small SST anomaly would be associated with a very large relative change in heat content of the ocean column.

The net surface heat flux has similar seasonal timing in the MIX1 model case (Meehl and Washington 1985, their Fig. 9). The shallow 50-m mixed layer is about 3.5°C too warm in late summer (March), but it begins to cool rapidly in autumn, as does the observed SST. The MIX1 SSTs cool rapidly because the 50-m mixed layer is too shallow and cools well into October. In the observed ocean, the rapid cooling in autumn to great depths is accomplished because weak density gradients in the upper layers allow deep vertical convection as surface water cools and sinks to considerable depths (Colborn 1975, as discussed previously). In late winter, the SST begins to rise as a direct consequence of the seasonal march of heat storage (Figs. 6 and 7). The 50-m slab also begins to warm consistent with the heat storage, but this happens somewhat later than in the observations because of the excessive cooling that takes place in the shallow 50-m layer during winter (Figs. 6 and 7).

The asymmetry of the seasonal march of SST, then, is provided by a combination of surface heat flux noted by van Loon in his Fig. 7 ("... less radiative energy is available during the period of cooling to offset the energy loss through evaporation and sensible heat transfer than is available during the warming period...") and the dynamics of the well-mixed upper 1000 m of the ocean at 50°S. This progression of the seasonal march of SSTs at 50°S and the dynamical coupling of ocean and atmosphere at that latitude are reflected in the troposphere (also very well mixed at that latitude) and contribute an essential element to the mechanism of the SAO.

6. Elements of the mechanism near 65°S

The second major part of the mechanism of the SAO postulated by van Loon involves the “coreless winter” over Antarctica. Figure 6b shows the annual cycle of surface air temperature from the station Syowa at 68°S (representative of temperature curves at that latitude) and temperatures from the lowest model layer from the SPEC SST and MIX1 model integrations. As seen for the 500-mb temperature results in Fig. 5, the SPEC SST model follows the observed values throughout the year with a maximum in January and a minimum in August, one month earlier than the observations. Computed temperatures are slightly higher than observed during the Antarctic winter, with the zonal mean model temperatures not reflecting the slight temperature rise at Syowa in June.

The MIX1 model also follows the annual cycle of observed temperature qualitatively, but remains too cold by about 3°–5°C throughout the year. This is as-
associated with overextensive sea ice in this model, as discussed by Washington and Meehl (1984) and Meehl and Washington (1990). Yet, the agreement with the shape of the annual curve of observed surface air temperatures is much better at this latitude in the MIX1 model than at 50°S in Fig. 6a.

The SPEC SST model reproduces the phase of the SAO but underestimates the amplitude (Figs. 4, 5, and 6). The forcing at the surface provides the correct phase, with the annual cycle of observed SSTs reflected in the 500-mb temperatures at 50°S. The coreless winter in the model at 65°S also gives about the right phase. The 500-mb temperatures are consistently about 3°C too low at 50°S in the SPEC SST model. In the Antarctic, they tend to be somewhat too high most of the year but are within about one degree of the observed values. This contributes a great deal to the reduced baroclinicity and, thus, to too small an amplitude of the SAO even though the mechanism appears to be working in the correct sense in the model with observed SSTs.

In the same atmospheric model coupled to the simple ocean mixed layer (the MIX1 case), the annual cycle of SSTs at 50°S is not well reproduced, in part due to lack of a deep enough mixed layer with ocean dynamics to accurately reproduce ocean heat storage. This is also reflected at 500 mb in the MIX1 model. Nevertheless, the coreless winter over Antarctica is reasonably well simulated in a qualitative sense in the MIX1 model. Therefore, the SAO in the MIX1 case is different from the observed and SPEC SST mainly because of the different simulation of SSTs and ocean heat storage at 50°S. This has implications for the observed interannual variability that will be discussed in section 8.

7. Eddy flux mechanisms

With the advent of global data assimilation and analysis schemes, it is now possible to examine transient eddy flux mechanisms in the SH that may play a role in the SAO. The magnitude of analyzed transient fluxes may still be called into question by the paucity of stations available (e.g., van Loon 1980; Karoly and Oort 1987), but the annual cycle should be well described qualitatively in the analyses. Transient eddy momentum flux [\(\bar{u}'\bar{v}'\)] and eddy heat flux [\(\bar{v}'T'\)] are computed to produce long-term monthly means. The product of the primed quantities (primes denote deviation from the time mean at each grid point) is computed and averaged for each month (overbar indicates monthly mean). Then those quantities are zonally averaged (brackets indicate zonal mean) and each set of monthly values is plotted as a time--latitude plot in Figs. 9 and 10.

The annual cycle of zonal mean transient eddy momentum flux [\(\bar{u}'\bar{v}'\)] from the ECMWF analyses (Fig. 9a) can be compared to the same quantity from the SPEC SST and MIX1 models (Figs. 9b and 9c). Values are shown for the upper troposphere where this quantity is largest (Trenberth 1987; Trenberth and Olson 1988; Newton 1972 also has shown geostrophic eddy momentum flux for 1957 and 1958). Seven-year averages from the ECMWF analyses show large maxima of poleward transient eddy momentum flux and eddy momentum flux convergence in the midlatitudes—stippling in the figure—during the times when the circumpolar trough is deepest and the 500-mb temperature gradient between 50° and 65°S is strongest in the transitional seasons. The small-amplitude SAO described earlier in the SPEC SST model is manifested in the transient eddy momentum flux with two poleward maxima around February and September. These two months are also characterized by an intense circumpolar trough (Xu et al. 1990).

The MIX1 model shows an altered SAO pattern in transient eddy momentum flux compared to the ob-
the troposphere (van Loon 1972). In the annual cycle, the SAO is manifested by an intensification of the circumpolar trough indicative of deeper lows and stronger westerlies in the midlatitudes. This is associated with the maxima of observed transient eddy momentum flux convergence at those latitudes around March and September (Fig. 9a).

Values for zonal-mean transient eddy heat flux \( [\nu T''] \) are shown in Fig. 10 for the lower troposphere where this quantity is observed to be largest (van Loon 1980; Trenberth and Olson 1988). Convergence of transient eddy heat flux is indicated by stippling. The observed analyses show a broad poleward maximum spanning southern winter for nearly nine months from March almost to November near 50° to 55°S. This indicates a second harmonic component to the midlatitude eddy heat flux with maxima in about March and September and a first harmonic with a winter maximum. The two model simulations also show large wintertime values of poleward transient eddy heat flux. This is associated with a credible simulation of the coreless winter in both model simulations, in spite of a quite different simulation of surface air temperature and 500-mb temperature variation at 50°S with the different ocean surfaces (Fig. 6).

Van Loon (1967) suggests that the coreless winter is associated with enhanced meridional flow from middle to high latitudes during the winter. This relatively warmer advection air, along with the wintertime radiational cooling, prevents a sharply defined midwinter minimum of temperature and a delay of the annual minimum until early spring.

The more recent analyses in Fig. 10 show that a wintertime maximum of transient eddy heat flux is present in the observations and the model simulations. This could be interpreted as increased meridional flow associated with transient eddies advecting warmer air southward (and colder air northward). Since the observations and the model simulations are reproducing the annual cycle of eddy heat flux at middle and high latitudes and all are characterized by a coreless winter at those latitudes, it is likely that transient eddy heat flux contributes to the coreless winter. This would be consistent with van Loon’s suggestion that enhanced meridional flow is part of the mechanism of the SAO.

8. Links to the tropics

Associations were noted earlier between the meridional expansion of the midlatitude trough in the South Pacific in MJJ and the development of a warm event in the SO (Fig. 2). This is characterized by a weakening of the subtropical high in the South Pacific, a reduction of the trade winds, and a warming of the waters in the equatorial Pacific. The opposite set of circumstances occurs in a cold event. In the Indian Ocean, the SAO may play a role in the development of relatively strong
or weak Indian monsoons. If, indeed, the SAO is playing a role in the evolution of extremes in the SO or in Indian monsoon variability, it would be of interest to monitor the mechanisms that seem to control it.

It has been shown that a critical element in the mechanism of the SAO is the annual evolution of SSTs (and thus ocean heat storage) at around 50°S. By changing the shape of the annual curve of SSTs at that latitude in the MIX1 model simulation, the SAO is simulated differently compared to the model simulation with observed SSTs. The implication for the observed system is that a change of SSTs at 50°S could affect the behavior of the SAO.

For example, higher SSTs at 50°S in MJJ could enhance the 500-mb temperature gradient and be conducive to a stronger trough and associated expanded cyclonic activity as far north as 25°S during that time of year (Fig. 2). This more intense trough could be associated with a greater number of low-pressure systems between about 25° and 60°S that could weaken the subtropical high in the South Pacific. This weakened high, then, could be associated with reduced trade winds, less vigorous equatorial upwelling, an increase of SSTs, and enhanced convection there. At the same time in the Indian Ocean, a similarly strong midlatitude trough could be associated with a less intense Mascarene high, less vigorous cross-equatorial flow, a weak Indian monsoon, and decreased convection, as discussed by Meehl (1987). An expanded trough in both Indian and Pacific sectors (with larger-than-normal pressure falls between 25° and 60°S as an intensification of the seasonal cycle of that pattern during MJJ) can be associated with weak convection in the Indian monsoon and strong convection in the tropical Pacific. Therefore, zonally uniform anomalies in the southern midlatitude Indian and Pacific sectors can occur in conjunction with nonzonal uniform convective anomalies in the tropics directly to the north. These associations are now being studied in greater detail and will be the subject of a subsequent paper.

9. Conclusions

The mechanism of the SAO of pressure and winds at middle and high southern latitudes (not addressed since van Loon’s study of 1967) is described and reviewed. Originally proposed and supported by van Loon with heat balance calculations, station data records, and synoptic weather analyses, the mechanism is reexamined with more recent data and with two GCM simulations. One model is run with the observed annual cycle of SSTs and sea ice (SPEC SST), and the other is run with a 50-m thick fixed-depth mixed layer (MIX1). The following conclusions are reached.

1) The annual cycle of SSTs at the ocean-dominated latitude of around 50°S and the shape of the annual cycle of SSTs there are the products of the dynamical coupling between ocean and atmosphere and are reflected in the annual cycle of temperature in the mid troposphere. Ocean heat storage and surface energy balance combine to produce a rapid decrease of SSTs in the southern autumn and a slow increase in the annual maximum in March. This is true in the observations and in the SPEC SST case at the surface and the mid troposphere. The model simulations provide insights into the consequences of changing elements of the ocean forcing near 50°S. The MIX1 case produces an altered annual cycle of SSTs due to the exclusion of ocean dynamics, the lack of a deep enough variable-depth mixed layer, and the resulting inadequate simulation of the annual cycle of ocean heat storage. The altered annual cycle of computed SSTs is evident in a similarly altered annual cycle of 500-mb temperatures in the MIX1 simulation.

2) The “coreless winter” over Antarctica is characterized by the lack of a well-defined midwinter temperature minimum and a slow mean decrease of temperatures (and sometimes a slight increase in midwinter at some stations) until the annual minimum in early spring. Van Loon postulated that the radiational forcing (the continuous outgoing longwave radiation while the sun is low or below the horizon), coupled with the amplification of the trough and enhanced meridional flow in winter, would produce the coreless winter. Transient eddy heat flux from observations and both model runs indicates a large winter heat flux convergence south of about 50°S (Fig. 10). This, combined with the radiational forcing common to both model simulations and the observed system, is associated with the slow decrease of Antarctic winter temperatures with a minimum in September.

3) The twice-annual maxima in the transient eddy momentum flux convergence south of about 50°S is associated with the strengthening of the circumpolar trough twice a year around March and September. This signal in the transient eddy activity occurs in conjunction with the twice-annual intensification of the mid tropospheric temperature gradient that is associated with the two main aspects of the SAO mechanism—the SSTs at 50°S and the coreless winter over Antarctica.

4) The phase of the SAO is simulated correctly in the SPEC SST case, but its reduced amplitude is mostly caused by lower-than-observed 500-mb temperatures at 50°S even though the basic mechanism seems to be working in the model.

5) It is postulated that, since a critical element of the SAO is the annual cycle of SSTs and ocean heat storage near 50°S, a change of those elements at that latitude could alter the behavior of the SAO.

This last point could be relevant to the development of an extreme in the SO (a warm or cold event) or to the relative strength of the Indian monsoon. Previous
studies have shown a connection between the inception of an ENSO event and the phase of the SAO in MJJ when the trough and associated cyclonic activity between about 25° and 60°S are expanding northward into the South Pacific. A more intense and expanded trough at that time of year is associated with a weakening of the subtropical high, a reduction of the trade winds, and a warming of the SSTs in the equatorial Pacific. Opposite conditions have been noted for cold events. In the Indian Ocean, a more intense trough in MJJ would weaken the Mascarene high and be associated with a weaker monsoon in India. Thus, a uniformly expanded and intensified trough in both southern midlatitude Indian and Pacific sectors (as part of an enhanced seasonal cycle in MJJ) could be associated with greater convection and rainfall in the tropical Pacific and less convection and rainfall in the Indian monsoon. If, indeed, such processes in the tropics are linked to the SAO, it would be worthwhile to monitor the mechanism of the SAO and document the various possible connections to determine cause and effect. Such a study will be the subject of a subsequent paper.

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