Harry van Loon Symposium
Studies in Climate II


Papers presented at the Symposium organized by the National Center for Atmospheric Research, 21 and 22 October 1996, to honor the contributions that Harry van Loon has made to our science. The National Center for Atmospheric Research is sponsored by the National Science Foundation.
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PREFACE

Harry van Loon has been a scientist at NCAR since 1963. Since that time he has published over 100 papers and inspired many more. His ideas have influenced collaborators from all over the world. Several of them came together at NCAR to join in a symposium dedicated to honor his continuing contributions to our science. The 21 papers presented at the symposium are contained in this volume. Nearly all of the authors have written papers with Harry in the past. Harry organized a similar symposium in 1984 which was titled “Studies in Climate” (NCAR/TN-227), thus the current title, “Studies in Climate II”.

Harry van Loon’s many publications (facing page and following list) represent a comprehensive set of “Studies in Climate” themselves. They provide a thorough study of Southern Hemisphere meteorology and climatology. His work comprises the major contribution in the AMS Monograph, METEOROLOGY OF THE SOUTHERN HEMISPHERE (1972) which continues to be the definitive work on the subject.

He wrote a series of papers on annual and semiannual variations in the atmosphere of both hemispheres, and in ocean currents and stratospheric winds, which have provided considerable insight into the workings of seasonal forcings. This is also true of his careful documentation of the structure of the large-scale spatial waves. He demonstrated the importance of changing advection patterns to show changes in temperature and documented the relation between eddy heat transports and temperature gradients.

He has increased our understanding of large-scale interannual and low-frequency variations in climate including the North Atlantic Oscillation, the Southern Oscillation, the whole ENSO phenomenon, and the QBO. In recent years his careful work with Karin Labitzke documenting 10-12 year oscillations in several atmospheric variables has stimulated many papers and meetings about the role and importance of atmosphere-solar interactions.

His papers are studied and referenced by observationalists, theoreticians and modellers alike. Besides his own publications he has inspired countless others. His insights and enthusiasm and encouragement lie behind many papers that don’t bear his name.

March 1997, The editors
WHAT HAS HARRY BEEN UP TO?

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Studies in Climate II, 21-22 October 1996

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23. Ralph Milliff
24. Marilyn Raphael
25. Dennis Shea
26. Harry van Loon
1. INTRODUCTION

The topic I have chosen for today's talk is the one with which I began my career at the NCAR. It deals with the observation that the annual range of the temperature in the surface layer of the open ocean and in the air above decreases across middle latitudes as one goes toward the pole. I was not the first to observe this phenomenon, although I thought so at the time; it had been noted by the classical, insightful climatologist Wilhelm Meinardus and described by him in two papers (Meinardus, 1929 and 1940). I became aware of it in 1954, described it in Notos (van Loon, 1955), and showed later (van Loon and Taljaard, 1958) that the decreasing annual temperature range with increasing latitude extends as high as the tropopause. A further description of the annual range over the ocean, including its implication for the interseasonal change in the atmospheric circulation, appeared in van Loon (1964). Later, a paper (van Loon, 1966) explained physically the meridional distribution of the range, and finally, the description was widened in several chapters of the AMS monograph "Meteorology of the Southern Hemisphere" (van Loon et al., 1972).

There are no new insights in this talk. I undertook the work out of curiosity to see if the picture of the annual range in the papers above was borne out by the global analyses from the European Centre for Medium Range Forecasting (ECMWF) since 1979.

The range of sea surface temperature on the Southern Hemisphere (SH) between February and August (Fig. 1) shows the decrease from the subtropics to about 50°S. At first thought a poleward decrease of the annual range does not seem to fit the fact that the annual range of insolation at the top of the atmosphere increases from equator to pole; if the tropospheric temperature range followed the pattern of the insolation it should also increase with increasing latitude. It does indeed over the continents, as seen in Fig. 2, but in the open ocean on Northern Hemisphere (Fig. 3) and in the air above (Fig. 2), the annual range follows the pattern of the SH: increasing from near the equator to the subtropical ridge, and decreasing further poleward to a minimum in the subarctic. This behavior of the annual range was explained in van Loon (1966), and can be considered an essential quality of a water hemisphere.

2. COMPARISONS: THEN AND NOW

a. Temperature

The comparisons in the following between the annual range in the analyses of the data before 1960 and those after 1979 are between monthly means in the former and seasonal means in the latter...
period. For instance, the differences in Fig. 4 are between January and July in the old analyses but between January-February-March and July-August-September in the recent ones. There is no reason to believe that the results from two different periods should be exactly the same, climate varies and so do data coverage and analysis methods. Considering these points and the fact that the comparisons are between data from extreme months and from extreme seasons one should expect only fair agreement; as it turns out, the agreement is good.

Figure 4a shows the difference in the zonally averaged temperature for only the SH, whereas Fig. 4b reaches from pole to pole. One should therefore compare Fig. 4a with the right half of Fig. 4b, and in doing so it is immediately apparent that there is little difference between the two periods — when allowance is made for the month versus season data: the zonally averaged temperature difference between summer and winter decreases in the whole troposphere as one goes polewards across middle latitudes on the SH, and the pattern and the values in the two periods are to all practical purposes the same. The annual range in the zonal average in Fig. 4b decreases in the troposphere between 40°N and 55°N too, although markedly less so than on the SH, owing to the comparative narrowness of the northern oceans.

The areal distribution over the SH of the annual range in the surface-air temperature in the 1954 analyses (Fig. 5a) reflects the range in the ocean's surface. It compares favorably with the interseasonal range derived from the European Centre analyses in Fig. 2.

The diminishing annual temperature range across middle south latitudes is conspicuous in the thickness of the 1000–500 mb layer in the early-period map (Fig. 5b) as well as in the thickness range between two extreme months in the International Geophysical Year 1957–1958 (Fig. 6a) and in the ECMWF post-1978 data (Fig. 6b). The three maps agree well, considering that the first one uses an analysis of a heterogeneous set of stations (Fig. 5b), the other is the difference between two months in a single year (Fig. 6a), and the third is a 15-year mean of seasonal differences (Fig. 6b).

On the basis of the four concentric zones of annual range on the SH—low in the tropics, high in the subtropics, low in middle latitudes, and high in the Antarctic—one can make the following deductions about the change from summer to winter in the meridional gradients of temperature in the troposphere, referring to the schematic diagram in Fig. 7:

1. Because the temperature falls more from summer to winter in the subtropics than in the tropics, the gradient between the two regions must be steeper in winter than in summer.

2. The decrease of the temperature from summer to winter is smaller in middle latitudes than in the subtropics and the gradient between the two regions must therefore be steeper in summer than in winter.

3. The temperature falls more from summer to winter in the polar regions than in middle latitudes, so the gradients across high latitudes must be steeper in winter than in summer.
b. Wind

The variation of the zonal component of the geostrophic wind with height is related to the horizontal, meridional gradients of temperature, as expressed in the thermal wind equation. Where the temperature gradient is steeper in summer than in winter the thermal wind is stronger in summer, and, everything else being equal, the west wind should also be stronger in summer.

The zonally averaged westerlies are, indeed stronger in summer than in winter at middle south latitudes, as shown in van Loon (1964 and 1966), and as seen in Fig. 8a (from van Loon et al., 1971). This figure is based on a climate analysis of the data available before the end of 1966 (Taljaard et al., 1969), and it differs little from the same, interseasonal, difference in the 15-year mean of ECMWF analyses in Fig. 8b. There is a limited region near 55°N where, in the zonal mean, the wind is slightly stronger in summer than in winter (Fig. 8b), corresponding to the narrow region of poleward decreasing annual temperature range on the Northern Hemisphere in Fig. 4b.

The similarity between the early analyses and the ECMWF analyses is also evident in Fig. 9, which contains the zonally averaged, vertical profiles of the zonal wind: The stronger wind in winter than in summer in the subtropics of the SH and in the subantarctic, and the stronger wind in summer than in winter at middle south latitudes, are found in the analyses of both periods; and the strength of the wind differs no more than one could expect from data for different periods and from single-month versus seasonal data.

The same figure also demonstrates both that the zonal mean velocities are higher on the Southern than on the Northern Hemisphere in both seasons, and the fact that the zonally averaged wind on the Northern Hemisphere is stronger in winter than in summer in all three zones (van Loon, 1964).

c. Global annual temperature range

Table 1 shows the area-averaged temperature in January and July for the Southern Hemisphere ("BEFORE", from van Loon et al., 1972), and the same in the extreme seasons ("AFTER", from the ECMWF analyses). An annual range based on three-month averages is apt to be lower than one based on two extreme months, and this emerges from the comparison in Table 1. The difference between the mean temperatures in the two data sets is thus no larger than one could expect from the different periods and from the use of extreme months and seasonal values.

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<th>So. Hem.</th>
<th>January</th>
<th>July</th>
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<tr>
<td>Surface Air</td>
<td>16.5</td>
<td>10.6</td>
</tr>
<tr>
<td>500 mb</td>
<td>-13.2</td>
<td>-18.6</td>
</tr>
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4. CONCLUSION

The explanation of the decreasing annual temperature range as one goes polewards across middle latitudes over the open ocean on both hemispheres is given by means of a heat balance study in van Loon (1966). It is the combined effect of (a) the mitigation of radiation influences by clouds and (b) the mixing of heat to greater depths in the ocean in middle latitudes than in the subtropics, due to stronger winds and ocean currents in middle latitudes.

To end this talk on an old-timer's note: it is gratifying to see that the early, so-called subjective analyses of the climate of the SH—and thus the studies which were based on them—can hold their own against the present-day climate analyses based on daily maps with a better data coverage analyzed by computer.

Acknowledgment

I thank James W. Hurrell for preparing those figures in which the ECMWF data were used.

References


Figure 1. Change of mean sea surface temperature (°C) on the Southern Hemisphere, February minus August. From van Loon et al., 1972.
Figure 2. Change of mean sea surface temperature (°C) in the North Atlantic Ocean, August minus February. From van Loon, 1991.
Figure 3. The mean surface air temperature ($^\circ$C) difference, January-February-March minus July-August-September 1985–1993, based on ECMWF analyses. From Hurrell et al., 1997.
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Figure 6. (a) The 1000-500 mb thickness difference (dkm) on the Southern Hemisphere, February minus August 1958. From van Loon (1967). The mean 1000-500 mb thickness difference (dkm) on the Southern Hemisphere, January-February-March minus July-August-September. From ECMWF analyses.

EC T42 1000-500 mb Thickness 1979-1993
JFM-JAS
Figure 7. A schematic diagram of the annual cycle in the air temperature in subtropical, middle, and high latitudes. The vertical arrows show the meridional temperature contrasts between latitude belts in summer and winter. From van Loon and Shea, 1988.
Figure 8. (a) The difference in the zonally averaged mean geostrophic wind (m s\(^{-1}\)) on the Southern Hemisphere, January minus July. From van Loon et al., 1971. (b) The difference in the zonally averaged \(u\)-component of the wind (m s\(^{-1}\)), January-February-March minus July-August-September. From ECMWF analyses.
Figure 9. Vertical profiles of the zonally-averaged geostrophic wind ($m \ s^{-1}$) at 25°S, 45°S, and 60°S. 
On the Ability of a General Circulation Model to Simulate Two Large-Scale Variations in the Southern Hemisphere

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1. INTRODUCTION

There are two large-scale modes which are unique features of the Southern Hemisphere (SH). One is the semi-annual wave and the other is a mode associated with the mass redistribution. The semi-annual wave is extensively described by van Loon (1967, 1972) and van Loon and Rogers (1981). The distributions of the amplitude, phase and the explained variance of the observed semi-annual wave shown in Fig. 1 (derived from South African and Australian analysis for the periods 1951–1958 and 1972–1983, Xu et al., 1990) indicate that the wave is a large-scale feature. The amplitude of the wave (Fig. 1a) peaks over the three oceans at mid-latitudes, reaches a minimum near 60°S, and reaches a second peak at polar latitudes. In the regions where the amplitude is large, the semi-annual wave accounts for over 50% of the total variance (Fig. 1b). The phase (Fig. 1c) is uniform in longitude and changes at 60°S from maxima in the spring and autumn in middle latitudes to maxima in the extreme seasons over the Antarctica. In a recent study concerning the long-term modulation of the Southern Hemispheric annual cycle, Hurrell and van Loon (1994) have shown that the semi-annual wave undergoes an appreciable change after the late 1970s and has lost its dominance in the 1980s. They suggest that these changes are related to the concurrent rise of sea surface temperature at low latitudes. It appears that the semi-annual wave is a large-scale feature which fluctuates on long time scales.

The mass mode was first reported by Rogers and van Loon (1982) using the Australian analysis. The mode in sea level pressure (SLP), as identified by Rogers and van Loon in both summer and winter data, is shown in Fig. 2. For both seasons, pressures poleward of 50°S rise and fall in opposition to those at lower latitudes. Trenberth and Christy (1985) pointed out that these SLP fluctuations are related to fluctuations in the distribution of atmospheric mass.

There have been a series of studies concerning the ability of general circulation models (GCM) of the atmosphere to simulate the Southern Hemispheric features. A review of these studies suggests that the first generation of low resolution GCMs are unable to simulate the large-scale feature of the semi-annual wave (Xu et al., 1990), whereas the mass mode is found in the GFDL (Lau, 1981) and the CCC (Zwiers, 1987) atmospheric GCM and in the coupled atmosphere-ocean ECHAM1/LSG GCM (von Storch, 1994). The ability in producing the mass mode on the one hand and the failure in generating the semi-annual wave on the other hand indicate that the causes of these two large-scale Southern Hemispheric features are of distinctly different nature. In this paper we analyze a 19-year integration
performed with a newly developed GCM, and discuss the simulation results focusing on the model performance in the SH.

The general circulation model used is the vertically extended and modified T21 version (hereafter referred to as ECHAM3.5) of the ECHAM3 spectral GCM (Roeckner et al., 1992). It is considered as one of the newly developed low-resolution models. The vertical extension is specifically designed for the simulation of the stratospheric climate. Because of this model feature, the model data may also indicate the effects of the above mentioned large-scale tropospheric variations on the stratosphere, if these features are captured by the model. The physical parameterizations in common to both ECHAM3.5 and ECHAM3 and the specifics of the modifications to the ECHAM3 can be found in Manzini and Bengtsson (1994). The climate and the interannual variability of ECHAM3.5 model are discussed in Manzini and Bengtsson (1996). The model is forced by a seasonal cycle in solar radiation and the climatological monthly mean sea surface temperature, as computed from the AMIP 1979–1988 data set. The first 19 years of a 30-year integration are considered in this paper. In Section 2, the semi-annual wave and the mass mode simulated by the ECHAM3.5 T21 model are described. The possible mechanisms and the model’s ability in capturing these mechanisms are discussed in Section 3. A summary is given in Section 4.

2. THE SIMULATED SEMI-ANNUAL WAVE AND MASS MODE

Figure 3a, 3b and 3c show, respectively, the amplitude, explained variance, and the phase of the semi-annual wave of the simulated mean SLP. Some local maxima of amplitudes and explained variances are found over the South Atlantic and south of Australia. The simulated semi-annual wave is confined to limited areas, and can only be considered as a local feature. The observed large-scale distribution with wave maxima at mid- and high latitudes, separated by minima along 60°S, are absent in Fig. 3a and 3b. The observed phase distribution with the essentially zonally orientated phase change at 60°S cannot be found in Fig. 3c.

On the other hand, the large-scale feature of the mass mode, as described by the first EOF derived from the monthly anomalies of sea level pressure in Fig. 4, is well captured by the ECHAM3.5 model. The positive (negative) pressure anomalies at mid-latitudes are related to the negative (positive) pressure anomalies at high latitudes. The location of the zero line in Fig. 4 lies in the same area as that in Fig. 2.

Although both Southern Hemispheric phenomena have large zonally symmetric components and are located both at middle and high latitudes, it seems to be extremely difficult to simulate the semi-annual wave, while relatively easy to simulate the mass mode.
3. CAUSES OF THE SEMI-ANNUAL WAVE AND THE MASS MODE

a. Mechanism of the semi-annual wave

The semi-annual wave was first explained by van Loon (1967) in terms of the differences in the seasonal heating and cooling between mid- and high latitudes, combined with the special circumstances of nearly equal annual temperature ranges in the middle troposphere. The situation is illustrated in Fig. 5 (from van Loon, 1967). Two features of tropospheric temperature are demonstrated by Fig. 5. First, the first harmonic dominates the annual cycles of the temperature at both mid- and high latitudes. Secondly, the annual marches of temperature at about 50°S and 65°S, as represented by data from the nearby stations (Fig. 5a), have distinctly different shapes. The curves at 50°S show a faster rate of fall during autumn (March to June) than the rate of rise in spring (September to December). The reverse is true for the curves at 65°S. The temperature maximum occurs around February/March at 50°S, but around December/January at 65°S. The temperature minimum appears in July/August at about 50°S, but is not reached until September at 65°S. The distinctly different annual marches at middle and high latitudes can be found throughout the whole troposphere. Van Loon suggested that these different seasonal temperature behaviors were tentatively related to the different responses of the earth's surface in the two regions to the heat budget.

As long as the annual temperature range at 65°S differs significantly from that at 50°S, as it is the case at the surface (Fig. 5b, the curves marked by b), the first harmonic dominates the annual cycle of the meridional temperature difference between the two latitudinal belts. In the middle troposphere where the annual temperature ranges in the two latitudinal bands are nearly equal, a pronounced semi-annual component in the meridional temperature difference results from the different shapes of the annual cycles at middle and high latitudes, as described in Fig. 5b (the curves in Fig. 5b marked by a). This leads to a semi-annual component in the strength and position of the subpolar trough and the semi-annual wave in SLP. A reexamination of this mechanism in light of more recent data is given by Meehl (1991). Since the different annual marches of the temperature in mid- and high latitudes is the key aspect of the mechanism, the semi-annual wave can be considered as a variation which is thermodynamically driven.

Figure 6 shows the simulated annual marches of the temperature at 50°S (Fig. 6a) and 65°S (Fig. 6b), and the meridional difference of these temperatures (Fig. 6c). The curves marked by A to E represent the values at 1000, 800, 600, 400 and 200 hPa respectively. As in the observations, the first harmonic dominates the annual cycles of temperature. However, the difference between the simulated annual marches at middle and high latitudes are no longer that pronounced. In particular, the annual maximum and minimum at middle latitudes occur at almost the same time as those at high latitudes. Consequently, the semi-annual component in the tropospheric temperature gradients (curve D in Fig. 6) is not generated in the same manner as in the observations. This discrepancy between the model and
observation might be responsible for the distribution of the strength and phase of the semi-annual wave in SLP.

On the other hand, the near surface curves are expected to be strongly affected by the climatological SST derived from the period 1979–1988. Since the annual march of the upper-air temperature is likely related to that of the surface temperature, the failure in simulating the semi-annual wave might be related to the SST forcing. This suggestion is consistent with the result that the observed semi-annual wave has undergone a significant change since the late 1970s (Hurrell and van loon, 1994). However, in order to answer the question of whether the ECHAM3.5 model is able to produced the semi-annual wave observed before the late 1970s, additional experiment using SST from the early period is required.

b. Mechanism of the mass mode

By performing a detailed scale analysis for the linear shallow water system, von Storch (1995) suggests that the mass mode is, on the lowest order, produced by the quasi-geostrophic dynamics. The essence of these dynamics are summarized in the potential vorticity equation which is derived by taking the curl of the momentum equations and then making the use of the continuity equation. Equation 1 is the dimensionless form of the inviscid potential vorticity equation of the linear shallow water system on the sphere.

$$\varepsilon_i \frac{\partial}{\partial t} (\zeta^* - FN \sin \eta^*) + \frac{L_m}{r} \cos \varphi \nabla v^* = 0 \quad (1)$$

The relative vorticity $\zeta^*$, surface elevation $\eta^*$ and meridional velocity $v^*$ are dimensionless variables. $\varepsilon_i$, $F$, $N$ and $\delta$ are dimensionless parameters defined as

$$\varepsilon_i = \frac{1}{2\Omega T} \quad (2)$$

$$F = \frac{L_m^2}{R^2}$$

$$N = \frac{g N_a / L_m}{2\Omega U}$$

$$\delta = \frac{L_m}{L_z} = \frac{V}{U}$$

$L_m$, $L_z$, $U$, $V$, $T$ and $N_a$ are characteristic scales of meridional and zonal lengths and velocities, and characteristic scales of time and surface elevation, respectively. $\varepsilon_i$ is the local Rossby number. It is smaller than $10^{-2}$ for $T = 8$ days. The decorrelation time of the mass mode indicates that its dominant time scale is of the order of one to two months. $F$ is the ratio of the square of the meridional length scale and the square of the Rossby radius of deformation $R$. For a barotropic system, $R$ equals the external
Rossby radius of deformation $R_{\text{ext}} = (gH)^{1/2}/(2\Omega)$ where $H$ characterizes the depth of the fluid. For a barocline system, $R$ should be replaced by the internal Rossby radius of deformation $R_{\text{int}}$, which is a function of the stratification. If the stratification is characterized by the Brunt-Väisälä frequency $N_z$, one has $R_{\text{int}}/R_{\text{ext}} = (N_z^2 H/\beta)^{1/2}$. $N$ in (2) describes the ratio between the meridional pressure gradient and the Coriolis force induced by zonal velocity. At middle and high latitudes, $N$ is of the order of one.

The first term on the left hand side of (1) describes the temporal changes of the relative potential vorticity which is the sum of the time rate of the relative vorticity $\zeta^*$ and the changes of relative potential vorticity through vortex tube stretching induced by variations of the surface elevation $\eta^*$. The second term describes the gain or loss of planetary vorticity through meridional motions. Equation 1 shows that the temporal changes of the relative potential vorticity must be balanced by the gain or loss of the planetary vorticity.

Using the dimensionless potential vorticity (1), one can determine the characteristics of the solutions, without explicitly solving the system of equations.

The first characteristic concerns the values of $\delta$. $\delta$ describes the degree of isotropy. In the conventional scaling analysis which concerns essentially isotropic motions, $\delta$ is one. For strong anisotropic motions, $\delta$ is much less than one. For strictly zonally symmetric motions, $\delta = 0$. The potential vorticity constraint of (1) suggests that for large-scale motions with $L_m$ being comparable to the earth's radius $r$, as long as the time scale of the motions is long enough so that $\varepsilon_{\text{max}}(1, FN \sin \varphi) << 1$, the motions must be anisotropic with $\delta << 1$. Otherwise the time rate of change of relative vorticity which is of the order of $\varepsilon_{\text{max}}(1, FN \sin \varphi)$ will not be balanced by the gain/loss of the planetary vorticity which is of the order of $\delta L_m/r$. In contrast to that, small-scale ($L_m << r$) and low-frequency ($\varepsilon_{\text{max}}(1, FN) << 1$) motions can satisfy (1) with $\delta = O(1)$.

Given the spatial scale of the motion, the degree of isotropy can be estimated. Figure 7a and 7c show the regressions of zonally averaged zonal and meridional wind anomalies with the time series of the EOF pattern shown in Fig. 4. The wind anomalies which are related to the mass mode have a meridional scale of the order of one third to one half of the earth radius $r$. This leads to $L_m/R_{\text{ext}} = O(1)$. In this case, $F = L_m^2 R_{\text{int}}^2 = (L_m R_{\text{ext}}^2)/(R_{\text{int}} R_{\text{ext}}^2) = R_{\text{ext}}^2/R_{\text{int}}^2 = O(5)$. The balance $\varepsilon_{\text{max}}(1, FN \sin \varphi) = \delta L_m/r$ leads then to a $\delta$ which is much smaller than $10^2$, when $\varepsilon_{\text{max}} << 10^2$ and the tropospheric value of $N_z$ with $N_z^2 = 2 \cdot 10^4 \, s^{-2}$ are used. Thus, the strong anisotropic feature of the mass mode results directly from the potential vorticity constraint.

The degree of anisotropy of the simulated mass mode can be estimated from Fig. 7. The largest values of the zonal and meridional wind anomalies in Fig. 7a and 7c are 2 m/s and 0.08 m/s, respectively. The resulting $\delta$ is smaller than $10^1$ which is consistent with the theoretical considerations. The explained variances (Fig. 7b and 7d) suggest that these wind anomalies are significant, even though their amplitudes are small.
The second characteristic concerns the dispersion relation which describes the essence of linear solutions. The equation which determines the dispersion relation describes the essential dynamics of the solutions. The scaling analysis is able to identify the dispersion relation and the equation which determines it. We demonstrate this first in term of the mid-latitudes small-scale ($L_m \ll r$) and isotropic ($\delta = I$ and $L_m = L_c = L$) Rossby waves. For these waves, the dispersion relations are known as part of the analytical solutions. This allows us to compare the analytically derived relations with those obtained from the scaling analysis.

Even though the mid-latitudes Rossby waves are normally studied using the mid-latitude $\beta$-plane approximation of (1), the nature of the dynamics remains the same. The balance described in (1) must also be satisfied by the small-scale isotropic Rossby waves. This balance determines the dispersion relations. Depending on the values of $FN \sin \varphi$, two types of dispersion relations can be derived. At mid-latitudes where pressure gradient and the Coriolis force are of the same order (e.g. $N = I$) and $\sin \varphi = O(1)$, one has $FN \sin \varphi = F$. The values of $FN \sin \varphi$ is controlled by the length scale of the motions. For large-scale motions with $L >> R$ and $F >> I$, the balance described in (1) is characterized by

$$\epsilon = \frac{L}{r}$$

(3)

At mid-latitudes with $O(\sin \varphi) = O(\cos \varphi) = O(1)$, (3) leads to

$$\frac{1}{T} = \frac{\beta gH}{f^2} \frac{1}{L}$$

(4)

where $\beta = 2 \Omega \cos \varphi / r$, $f = 2 \Omega \sin \varphi$.

Besides the sign, (4) is identical to the long wave approximation for the Rossby waves (Gill, 1982)

$$\omega = -\frac{\beta gH}{f^2} k$$

(5)

where $\beta = 2 \Omega \cos \varphi / r$, and $\Omega$, and $k$ are frequency and zonal wave number, respectively.

On the other hand, for small-scale motions with $L << R$ and $F << I$, the balance described in (1) is characterized by

$$\epsilon = \frac{L}{r}$$

(6)

or

$$\frac{1}{T} = \beta L$$

(7)

which corresponds to the short wave approximation for Rossby waves.
\[ \omega = -\frac{\beta k}{k^2 + l^2} \] (8)

The dispersion relation, a relation between the time and length scales, can be derived from any of the equations of the system which contains derivatives of both time and space. In the case of shallow water system, they are the equations of the zonal and meridional momenta and the continuity equation. The fact that there exist different dispersion relations suggests that the equations of the system are not dynamically equivalent. For large-scale motions satisfying \( FN \sin \varphi = F >> 1 \), the solutions which are characterized by the dispersion relation (3) are, to the lowest order, determined by the continuity equation. The equations of momenta reduce, to the lowest order, to balance equations (i.e. the time derivatives which contain the time scale can be neglected in these equations). The situation reverses for small-scale motions with \( FN \sin \varphi = F << 1 \) whose dispersion relation is determined by the momentum equations.

Motions characterized by (3) can be easily found in the ocean where \( R = R_{int} \) is small because of the strong stratification. In the atmosphere, \( R \) is not that small. However, the spatial scale of the mass mode is extremely large, so that it is possible for \( F = L_m^2/R_{int}^2 \) to be large. As discussed earlier, \( F = O(5) \), so that \( FN \sin \varphi > 1 \). The continuity equation becomes the dynamically most relevant equation. The mass mode should be considered as a solution of an equation system whose linear operator is, to the lowest order, determined by the equation of mass.

In order to identify the detailed spatial structure of the mass mode, the boundary condition which affects the dynamically most relevant equation, i.e., the continuity equation, must be considered. This boundary condition results from the attempt of the flow to satisfy the no-slip conditions at the surface, and can be expressed in terms of the Ekman pumping/sucking velocity \( w^* \) in the continuity equation of the interior flow. The resulting potential vorticity equation becomes, to the lowest order,

\[ \epsilon_r \frac{\partial}{\partial t^*} (FN \sin \varphi \eta^*) - \delta \frac{L_m}{r} \cos \varphi v^* = h \nu^* \] (9)

where \( l \) depends on the eddy viscosity coefficient.

Equation 9 suggests that the time rate of change of mass anomalies, as described by \( \eta^* \), is controlled by the internal meridional velocity \( v^* \) and the mass convergence in the Ekman layer which results \( w^* \). For motions with long time scale \( T \), the magnitude of the first term in (9) is small. This is only possible when convergence due to the interior flow balances, to a large extent, the convergence within the Ekman layer. As a consequence, a much smaller meridional velocity is found in the interior than in the Ekman layer, since the Ekman layer is expected to be much thinner than the interior layer. Such a meridional circulation is produced by the model, as indicated by the regression (Fig. 7c) between the zonally averaged meridional velocity and the time series of the first EOF shown in Fig. 4.
Figure 7 suggests also that the mass mode is not strictly confined to the troposphere. The zonal wind anomalies penetrate into the stratosphere (Fig. 7a). A significant amount of variance in zonal wind, up to 20–30%, is explained by the mass-mode-related anomalies in the lower stratosphere (Fig. 7b). A regression between the zonally averaged temperature and the time series of Fig. 4 shows that 5–10% of the total variance in temperature is related to the mass mode. This is a novel result, given that the previous GCMs examined did not include the stratosphere.

SUMMARY

According to van Loon (1967), the SST forcing plays an important role in generating the semi-annual wave. A possible cause of the failure in simulating the semi-annual wave may therefore be that the model has been forced with climatological SST derived from a period when the observed semi-annual wave is weak. However, from the integration considered here, it is not clear whether the model would produce the observed prior-1980 semi-annual wave, if the model is forced by the pre-1980 SST.

The ECHAM3.5 model is able to produce the observed mass mode. A scaling analysis of linear shallow water system suggests that the mass mode is generated by the straightforward potential vorticity dynamics. For large-scale motions such as the mass mode, only the surface elevation which represents the mass anomalies contributes, to the lowest order, to the relative potential vorticity. The potential vorticity dynamics are controlled by the equation of mass. The low-order spatial structure of the mass mode is determined by the convergence of mass caused by the flow in the interior and Ekman layer.

Since the ECHAM3.5 model includes both troposphere and the stratosphere, it can be used to study the effect of the mass mode on the stratospheric variations. This study suggests that 20–30% of the total variance of the zonally averaged zonal wind in the lower stratosphere of the SH is related to the mass mode.

References


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Figure 1. The observed semi-annual wave in SLP: distribution of its amplitude a) and phase c), and the percentage of the annual variance explained by it. (From Xu et al., 1990, spacing: 1 mb, 10%, 1 month).
Figure 2. The first EOF of SLP anomalies in summer (DJF) and winter (JJA), as derived from the Australian Southern Hemisphere analyses. (From Rogers and van Loon, 1982).
Figure 3. The semi-annual wave in SLP as simulated by the ECHAM3.5 model: Distribution of its amplitude a) and phase c), and the percentage of the annual variance explained by it. (Spacing: 1 mb, 10%, 1 month). Note: Fig. 3 is rotated 90 degrees with respect to Fig. 1.
Figure 4. The first EOF of monthly SLP anomalies derived from a 19-year integration with the ECHAM3.5 GCM. Note: Fig. 4 is rotated 90 degrees with respect to Fig. 2.

Figure 5. The annual march of temperature (°C) at Marion Island and Mawson a), and the schematic representation b) of the annual march of temperature at 50°S and 64°S at the surface (denoted by b) and at 500 mb (denoted by a). (From van Loon, 1967).
Figure 6. The annual march of the zonally averaged temperature at 50°S a) and 65°S b), and the meridional temperature difference c), as derived from the 19-year integration with the ECHAM3.5 model.
Figure 7. Regression patterns (upper panel) between the zonally averaged anomalies of zonal and meridional wind and the time series of the EOF pattern shown in Fig. 4. (zonal wind in m/s) a), meridional wind in 0.1 m/s c), and the variance explained by the regression patterns (lower panel, spacing 5%).
1. INTRODUCTION

This symposium is to honor my good friend and colleague Harry van Loon. Harry's most notable works prior to the completion of my Sc.D. at Massachusetts Institute of Technology in 1972 in the field of atmospheric sciences were undoubtedly all of his papers and atlases, and the monograph on Southern Hemisphere meteorology. The monograph, edited by Chester Newton, and featuring several van Loon contributions came out in 1972, so I hope we can say 1972 was a good year for Southern Hemisphere meteorology! At that time I returned to New Zealand, as I had been on a New Zealand Research fellowship, and so I began to deal with all the problems Harry had been wrestling with for some time. Not that I knew Harry or had come across his work at that time.

Of special note in Harry's work are the difficulties of analyzing a very sparse observational database and trying to make sense out of the weather and short-term climate variations. It is amazing, in retrospect, how much information Harry and his colleagues were able to glean. The ingenuity that went into fashioning quality analyses that have stood the test of time is not appreciated, I believe, by those who have never had to do it. Recall that this was largely before the days of satellites, and computers were just entering the scene. The fact that these works are not sufficiently appreciated is evident from the many times that things written about by Harry are often rediscovered, or worse, corrupted by newer poor data analysis. So it is often left to Harry to update his analyses and remind us that there is really little new in the world to discover.

I recall some of my first encounters with Harry's work were when I did some research on interannual variations of the atmospheric circulation around New Zealand and I wrote a paper on the Southern Oscillation that was published in 1976. This made me aware that 1957–58, the International Geophysical Year (IGY) period on which a lot of Harry's work was based, was a period of a strong ENSO warm event. I vaguely recall having written to Harry and I pointed out to him that his wonderful analyses were probably aliased by ENSO. Of course this did not upset Harry at all, instead he subsequently wrote a series of eight papers on the Southern Oscillation, and a number of others besides! I am not sure whether I had anything to do with getting Harry into this line of research or whether my memory is even correct, but it makes a good story.

Harry's synoptic background lead him naturally to a careful scrutiny of the data he was analyzing. Misrepresentation of data was a sin but all too frequent in the early days of numerical objective analysis schemes. What I am going to report on today is a continuing example of this in very recent analyses from the NCEP/NCAR reanalysis project. However, the quantities are not ones dealt with by Harry, but
he will take some pride in the fact that it is still difficult for a machine to beat out a careful manual analyst.

What I deal with is moisture and the hydrological cycle. Moisture plays an important role in the heat budget of planet Earth especially through the greenhouse effect of water vapor and, at the surface, by moderating surface temperature changes as heating goes into evaporating moisture rather than increasing temperature. In addition, the transport of moisture by the atmosphere effectively redistributes the latent heat. Yet these aspects are not well determined in our climate system.

Reasons for the scanty knowledge of both moisture in the atmosphere and precipitation stem from the lack of observations, especially over the oceans, and the nature of the quantities. Rainfall and clouds often occur on quite small time and space scales, so that a single moisture or precipitation observation may not be representative of more than an area with dimensions of a few kilometers across or for more than a small fraction of a day. Experience with atmospheric models indicates that they quickly adjust the moisture fields to be compatible with the model moist physics, of which moist convection is probably the most critical. Therefore, observed information has not proven to be very valuable in numerical weather prediction. Moreover, it follows that moisture fields from 4DDA may not be very good estimates of the real world.

Measurements of accumulated precipitation occur only where humans live in relatively widely-spaced locations, and the buckets used to measure accumulations may not catch it all, especially under snowy and windy conditions (Legates and Willmott, 1990). Satellite data on moisture have been made available to the global analyses from TOVS, although with mixed results (Liu et al., 1992; Wittemeyer and Vonder Haar, 1994). After July 1987 fields of precipitable water and other quantities from the SSM/I have become available, but these are not produced in time for operational purposes. Recently, a special set of global analyses of water vapor has been compiled taking advantage of the radiosonde measurements combined with SSM/I and TOVS data called NVAP (Randel et al., 1996). Also, a number of different data sources are being utilized to put together global monthly mean fields of precipitation through the GPCP for the period after 1987 and these are extended to cover the period after 1979 by Arkin and Xie (1994) and Xie and Arkin (1996). Over land these fields are mainly based on information from rain-gauge observations, while over the ocean they primarily use satellite estimates made with several different algorithms based on outgoing longwave radiation (OLR), and scattering and emission of microwave radiation. Because the latter consist of an integration of spot estimates of precipitation rate, they are subject to considerable sampling uncertainties. In some regions, such as the ITCZ in the Pacific, results from these algorithms do not agree very well, and so uncertainty exists as to the true values (e.g., Chiu et al., 1993). Nevertheless, we will use the NVAP and Xie-Arkin (GPCP) results as one version of "truth". Alternative estimates of these fields come from the assimilating model used in 4DDA, usually from a short model integration, but they need to be evaluated using other sources.

Other quantities related to moisture and its sources and sinks are less well known. Trenberth and Guillemot (1995) evaluated the precipitable water and moisture budgets from the global analyses of
European Centre for Medium Range Weather Forecasts (ECMWF), the U.S. National Meteorological Center (NMC) (now National Centers for Environmental Prediction, NCEP) and NASA/Goddard and examined the differences between evaporation $E$ and precipitation $P$, $E - P$, from 1987 to 1993. The precipitable water from the global analyses was computed and compared with satellite data from the SSM/I. Fluxes of moisture and their divergence were used to estimate the vertically integrated moisture budget and thus $E - P$ as residuals. Here we use a similar approach but with the NCEP reanalyses (Kalnay et al., 1996) and use will be made of the analyses in model (sigma) coordinates at full resolution.

The NCEP system is based on a numerical weather prediction model with T62 spectral resolution and 28 levels in the vertical with five of those levels in the atmospheric boundary layer. The Spectral Statistical Interpolation (SSI) scheme is employed in the analysis with complex quality control. Fields are not initialized. Although products from the reanalyses include estimates of evaporation $E$ and precipitation $P$, they are "C- variables" which are generated entirely by the model used in the 4DDA.

2. COMPUTATIONS AND METHODS

We have carried out a systematic evaluation of the monthly, seasonal and annual fields by computing means, standard deviations and anomalies of NCEP precipitable water, model-based evaporation, model-based precipitation, and implied $E - P$, the vertically integrated flux of moisture (as a vector), the divergence of the latter, the tendency in moisture, $E - P$ from the moisture budget from the previous two quantities, and the differences between the two estimates of $E - P$. In effect, this is a determination of the extent to which the moisture budget balances. The base period for the climatology is 1979 to 1995. Similar statistics are computed from precipitable water from NVAP and $P$ from Xie-Aarkin GPCP, along with differences with the model-based values when available, so that we can thoroughly document the mean bias, the standard deviations of the differences (indicating the typical errors), and the correlation between the anomalies. For NVAP we are restricted to the more limited period from 1988 to 1992. These calculations provide indications of whether the model-estimated anomalies are meaningful even if there are biases present in the total fields.

3. RESULTS

Results are presented here only for the annual means. Complete results are presented in an atlas by Trenberth and Guillemot (1996).

Figure 1 shows the precipitable water and how well the NCEP/NCAR reanalyses replicate the NVAP fields which are believed to be quite an accurate depiction of the truth. Although the shortness of the NVAP record limits the comparison, the results reveal substantial shortcomings in the NCEP reanalyses. Although correlations between the two fields are quite high, moisture is depleted in the tropical convergence zones by 4 to 12 mm, but is too high in the South Pacific high. The tropical structures are less well defined in the NCEP reanalyses and values are generally smaller where they should be high and higher where they should be low, a pattern also present in earlier operational
analyses (Trenberth and Guillemot, 1995). In addition, the NCEP fields reveal much less variability from year to year in analysed $w$ in the tropical Pacific than in the NVAP data (not shown). Dominant variations are found in the tropical Pacific in association with the ENSO phenomenon, but the variance in the NCEP fields are especially deficient in the central and western tropical Pacific in all seasons.

Figure 2 shows the annual precipitation field from NCEP compared with the Xie-Arkin GPCP product. While the latter can not be considered to be fully quantitatively accurate (e.g., see Chiu et al., 1993), the patterns should be reasonable. There appear to be noteworthy biases in the NCEP $P$. The NCEP model $P$ generally reveals a pronounced double ITCZ in the central Pacific and the location of the SPCZ is therefore not well captured. In the ITCZ, rainfalls are weak relative to the Xie-Arkin product. A bias for too much rainfall in the model over the southeastern U.S. is also present in northern summer. The standard deviation maps (not shown) reveal that the variability in the central tropical Pacific of $P$ associated with ENSO is severely underestimated in the NCEP reanalyses, and moreover, is not very well correlated with the Xie-Arkin product. In the correlation between the two $P$ fields (Fig. 3), values exceed 0.8 only around Antarctica, in the Arctic, and in tropical eastern Pacific. In both of the polar oceans, the deficit of real data led to the substitution of NCEP $P$ values in the Xie-Arkin analysis, and so the two products are dependent and should be closely related. This is not so in the tropical eastern Pacific, where a large ENSO-related signal is apparently captured by the NCEP reanalyses. Elsewhere, values are mostly lower than about 0.4 and approach zero in the tropical western Pacific.

In places where $P$ is strong, especially the ITCZs, there is a low bias in model $P$. Thus the low bias in precipitable water in the tropics could be a factor in contributing to lower rainfall rates and, at the same time, lower rainfalls mean lower latent heating and feedback to the divergent flow which transports the moisture into the region, thereby contributing to biases in moisture amounts.

Figure 4 shows the moisture transport, $E - P$ from the residual calculation and evaporation from the model, while Fig. 5 shows the computed differences between the two $E - P$ fields. The locations of the dry subtropics and ITCZ are qualitatively well depicted by the side panels representing zonal averages of monthly mean $E - P$. Maximum zonally-averaged values of $E - P$ of about 3 mm/day are observed in January near 20°N while they reach 4 mm/day near 20°S in July. Excess precipitation of the order of 2 to 2.5 mm/day is noted near the ITCZ every month of the year with a somewhat broader latitudinal extent of the peak during the northern winter. Standard deviations of $E - P$ (not shown) are largest in the tropics but do not appear to be very seasonally dependent.

During the northern winter the dry cold air spilling eastward into the Pacific from Asia gives rise to a very pronounced moisture and evaporation gradients over coastal waters. The same is true off the east coast of North America and, to a lesser extent, in the Indian Ocean as a result of the winter monsoon. On an annual mean basis the largest evaporation of over 6 mm/day is in the subtropical Indian Ocean, as was also found in ECMWF analyses (Trenberth and Solomon, 1994). Smaller areas of similar magnitude occur over the East Australia current and Gulf Stream. Otherwise, the largest values of evaporation generally occur over the warm northward-flowing ocean currents in the western Pacific.
and Atlantic in winter. The more maritime Southern Hemisphere is characterized by a more zonal
distribution of evaporation with the largest values found over subtropical waters.

The computed differences between the two $E - P$ fields (Fig. 5) reveals the extent to which the
moisture budget balances. To some degree the values are smaller than anticipated, apparently
indicating the dominance of the divergence field in shaping both products. However, remarkably, nearly
all island stations show up as bull’s-eyes in this difference field calculated from the model and the
residual technique and there are several other striking systematic differences that appear almost every
month. Many features are strong and very consistent in all months, and their origin is quite puzzling.
Each $E - P$ field by itself seems to be quite coherent and reasonable. Yet when differences are taken,
bull’s-eye features emerge centered almost over island stations throughout the global domain.

Note that positive differences in Fig. 5 imply that the model $E$ is too high, the model $P$ is too low,
and/or that the analyzed moisture divergence is too negative. Extensive efforts have been made to track
down the analysis characteristics that lead to these features. They appear to have multiple causes and
eamples of all three kinds of biases can be found. In retrospect, perhaps these differences are not so
surprising as they highlight sources of systematic biases whether from the analyses or the model
parameterizations.

The bull’s-eyes arise in both pressure and model coordinate calculations. Moreover they were
clearly features of the moisture divergence, not the tendency term. Attempts to trace the features to
individual levels or layers were not very successful because of the fact that they are systematic with
height. At low levels, although identifiable, the features are quite ordinary and tend to be submerged in
the overall field noise and interannual variability. The contributions between about 300 and 400 mb
emerge somewhat more clearly from the background noise but, as moisture amounts fall off substantially
with height, their total contributions are not that large. So it is a systematic pattern with height from 1000
to 300 mb that separates these features from the noise and which tends to cancel elsewhere as
integrations are performed both in height and in time. After much investigation it was determined that
the primary term contributing to the bull’s eyes is the eastward advection of moisture. Thus it emerges
that the dominant contribution to the bull’s-eye features comes from the eastward gradients in $\bar{q}$, arising
from an almost imperceptible decrease in $\bar{q}$ at the station ($\frac{\partial \bar{q}}{\partial x}$ negative). Surprisingly, for the advection
term, there is not a strong or systematic feature with reverse sign (i.e., a dipole structure). The
implication seems to be that at isolated island stations throughout the southern oceans, the model first
guess in the vicinity is systematically slightly moister than the observed value, and the observed
information is advected downstream affecting the analysed values in that area. Consequently, the only
feature emerging in the analyses is a bull’s-eye slightly upstream from the station location.

Evidence for a dry bias in the observations from certain rawinsondes, which include many of
those identified in Fig. 5 with positive bull’s-eyes, comes from the Soden and Lanzante (1996) study
which uses upper tropospheric moisture from satellite-based water vapor channels. They find a negative
bias in relative humidity as measured by the radiosondes with capacitive or carbon hygristor sensors of 10–20% in the upper troposphere.

A bull's-eye of reverse sign appears over Apia (Pago Pago; 14°S 170°W) which is an area where model precipitation is too high. Frequently in northern winter other bull's-eyes occur nearby, with a positive center near Penrhyn (9°S 153°W) and a negative center near Tahiti (17°S 150°W), and presumably these also relate to the characteristics of the different rawinsondes used, the biases in the analyzed moisture fields, and the model biases in precipitation and evaporation.

Other systematic features in Fig. 5 appear to have different origins that stem more from the model-generated fields of $E$ and $P$. In particular, there are very sharp gradients in $E$ across coastlines (Fig. 4), so that the feature south of Japan over the East China Sea appears to originate primarily from very strong $E$ in the model although there may also be insufficient model precipitation there. West of California, on the other hand, the positive feature stems from a deficiency in model $P$, as verified by the GPCP values. Over the southeastern part of the United States in northern summer, the bias originates from a systematic overestimate of $P$ by the model, again as verified by GPCP. These biases seem to extend to the Caribbean and Gulf of Mexico and also rainfall regions in the western tropical Pacific in the northern summer.

Many other features around the coast lines, such as the negative centers near 30°E 30°N and over Aden (42°E 12°N) and the positive center over Saudi Arabia, may be traceable to the presence of negative $E$ in the model surrounding the Red Sea where values are large and positive, yet with a maximum in $E$ over Saudi Arabia. In some places $P$ is also negative, and these negative values of $E$ and $P$ presumably originate from ringing effects at finite spectral model resolution. The evaporation over Saudi Arabia appears to be excessive.

Over southern Africa, the tendency for positive values in Fig. 5 arises from errors in the moisture divergence. In June 1995, for instance, model $E$ and $P$ estimates are very small (< 1 mm/day) and $E - P = 0$, whereas the residual $E - P = -2$ mm/day implying excessive precipitation which does not verify from GPCP estimates.

Consequently, there are some places (such as North America and other areas where the precipitation is clearly wrong, or Saudi Arabia and the East China Sea where the evaporation seems to be at fault) where it appears that the residual method produces better answers, but in other places (such as southern Africa) the residual method estimates are clearly inferior to those from the model parameterizations. Both sets of estimates are affected by biases in moisture, as analyzed, while the moisture divergence depends critically on the velocity divergence field. The model estimates also depend upon the parameterizations of subgrid scale processes, such as convection, that influence $E$ and $P$.

Therefore, while the comparison of $E - P$ from the moisture budget with the model result reveals similarities, there are also strong and systematic differences. In particular, the remarkable island station
bull's-eyes are identified as originating from the moisture budget calculation through rather subtle effects arising from small but systematic differences in vertical moisture profiles from those in the surrounding oceans. In part this may reflect differences between radiosonde moisture amounts with either the model first guess or TOVS soundings. It indicates that the influence radius of rawinsonde moisture observations in the analyses, while perhaps appropriate for an individual sounding, is probably too small in the analyses of these data on average. These analysis problems are somewhat reminiscent of numerical analysis problems manifested in heat transports found by van Loon (1980).

4. MOISTURE BUDGET SOURCES OF ERRORS

A continuing major source of errors in the tropical moisture budget is the divergence field. The negative bias in precipitation in the NCEP reanalyses is perhaps an indication that the divergent circulation is too weak. It may be that improvements in this area will have to wait for global satellite-based wind measurements, although it seems that substantial progress should be possible if scatterometer winds at the surface were fully utilized in the analyses.

With 6-hourly analyses, the diurnal cycle errors are not a source of concern. However, mass imbalances are quite large in the NCEP reanalyses (Trenberth, 1996) and can distort other budgets unless corrected for, although the impact is fairly small on the moisture budget. Vertical resolution can be an issue for analyses on pressure coordinates, but it is not an issue when use is made of all the levels in model coordinates, as done here. Horizontal resolution can be an issue where sharp gradients occur as there is evidence of spectral ringing and physically impossible values of negative precipitation and physically unlikely values of negative evaporation. Horizontal resolution is also believed to be important in the vicinity of steep orography and associated low level diurnal jets (Helfand and Schubert, 1995).

We have shown that substantial biases continue to exist in the moisture fields in the NCEP reanalyses. Information available from SSM/I is not utilized and nor are the water vapor channels of TOVS. At ECMWF and operationally at NCEP, brightness temperatures of the TOVS channels are directly assimilated in place of retrievals and this apparently provides a substantial improvement in the depiction of moisture over the oceans. Nevertheless, it is apparent from the bull's-eyes in the $E - P$ difference fields, that there is difficulty in assimilating moisture into models and the information content inherent in these fields is not being adequately utilized. The moisture errors probably feed back and influence the divergent circulation through negative biases in latent heating arising from precipitation. Therefore this is one area where it seems possible to do a much better job and where it is extremely important to do so.

It is apparent that there are errors arising in the precipitation field from the model physics. In the southeastern part of the United States in summer, the model is not able to sustain the observed high humidities giving rise to spurious precipitation. Errors in assigning soil moisture values or in the land surface moisture budget appear to be responsible for some errors in evaporation. We have demonstrated the discrepancy between the model $E - P$ and that from the residual technique, so that the
moisture budget is not balanced in the analyses, a point made also by Mo and Higgins (1996). Improvements in parameterization of the moist physics in the model is an obvious need, and should pay off in improved forecasts as well.

5. CONCLUDING REMARKS

It is apparent that there are substantial problems with the NCEP moisture-related fields and the moisture budget is not close to being balanced. Some problems are identified with the assimilating model physics, but several stem from the quality of the analyses. The latter certainly depend also on the assimilating model, but many go well beyond that. We have shown that there are systematic differences between radiosonde moisture amounts with either the model first guess or TOVS soundings that show up as bull's eyes in the $E - P$ difference fields. The implication is that the influence radius of rawinsonde moisture observations, which perhaps is appropriate for an individual sounding, is nevertheless too small in the analyses of these data on average. This is the sort of thing that was found in the late 1970s global analyses over the southern oceans by Harry van Loon, and it seems likely that Harry's expert manual analyses could have produced a better result.

Substantial shortcomings mean that the analyses should only be used with great caution in climate and hydrological studies. The reanalyses are a great step forward for climate studies and are much better than operational analyses in general in a number of ways. The strategy in reanalyses is that they should be done again and indeed it is crucial that they be done again taking advantage of lessons learned.

Harry, it is a pleasure to have worked with you, and you are a joy to have around. I only hope that I am as healthy and active when I rise to your esteemed status.

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References


Figure 1. Annual mean precipitable water for the period 1979–1995 (NCEP) and 1988–1992 (NVAP), and their differences. All quantities correspond to vertical integrals and have been truncated to T31 for presentation.
Figure 2. Annual mean precipitation for the period 1979-1995 from the model, GOCPP and their differences. GPCP refers to the Xie-Arkin precipitation data set which uses a slightly different algorithm for the period 1979-1995 than was used for the original GPCP data. All quantities correspond to vertical integrals and have been truncated to T31 for presentation.
Figure 3. Correlation between the seasonal anomalies in precipitation in NCEP and Xie-Arkin GPCP analyses over the period 1979 to 1995 (67 seasons). Values exceeding 0.8 are stippled.
Figure 4. The moisture transport, $E - P$ derived as a residual from the indirect moisture budget computation, and $E$ for the period 1979–1995. All quantities correspond to vertical integrals and have been truncated to T31 for presentation.
Figure 5. $\Delta (E - P)$ is the difference between the residual computed from model E and P, and $E - P$ computed from the indirect method for the period 1979–1995. All quantities correspond to vertical integrals and have been truncated to T31 for presentation. Some island station locations have been identified.
1. INTRODUCTION

The potential for long-range prediction of monthly time averages as defined by Madden (1976), Madden and Shea (1978; hereafter MS) and Shea and Madden (1990; hereafter SM) is estimated by the ratio of the observed interannual variability of monthly means ($\sigma^2_A$) to the variability due to "climate noise" ($\sigma^2_T$). The square root of the climate noise is synonymous with the standard error of estimating monthly means. It is the error to be expected strictly from statistical sampling fluctuations. It is a result of daily variability introduced by the internal dynamics occurring within the atmosphere and as such cannot be predicted at long lead times. Thus it represents an unpredictable component in the climate system. The actual variance, $\sigma^2_A$ is assumed to consist the climate noise and variance introduced by slowly changing external conditions (i.e., climate signals). Assuming $\sigma^2_T$ is independent of $\sigma^2_A$ then the amount by which the observed interannual variability exceeds the climate noise is a measure of the potential long-range predictability. For example, F-ratios $\sigma^2_A/\sigma^2_T$ of 1.5 and 2.0 would indicate that it would potentially be possible to explain 33% and 50% of the variance of monthly means, respectively. It is important to note that large estimates of $\sigma^2_T$ do not necessarily imply low potential for prediction.

The estimates of climate noise were evaluated within the conceptual framework outlined by Leith (1975, 1978). Briefly, this may be described as follows. Under the influence of constant external conditions individual realizations from an ensemble of realizations occurring within a climate allowing a unique set of statistics are sampled. The resulting time means from each realization will exhibit fluctuations about the ensemble mean. This variability is the climate noise. In fact, under these ideal conditions the climate noise and the interannual variability would be nearly identical since all variability would be due to internal dynamics alone. In this ideal case, there would be no potential for long-range prediction since the F-ratios would be approximately one everywhere.

Unfortunately, climate noise in the 'real world' can not be estimated under these ideal conditions. Neglecting possible atmospheric intransitivity (Lorenz, 1968, 1976), the continuously changing external conditions (e.g., sea-surface temperature, snow-ice cover, etc.) within the ensemble of realizations (approximated by daily data within, say, a season) and even within an individual realization affect daily weather and, thus, the estimates of the climate noise. Nevertheless, it is important to assess the magnitude of the climate noise and to estimate the potential for long-range prediction. In the following, we briefly describe the method used to estimate $\sigma^2_T$ for precipitation.
2. DATA

The daily rainfall data used in this study were from the India Meteorological Department (IMD-Pune) and from the Climate Prediction Center (CPC), Washington, D.C. The data set contained data from 1596 Indian stations and 61 stations from countries bordering India. These stations were selected based upon length of record and the desire for representative spatial coverage. Generally, stations within India spanned more than 60–65 years of data. Thirty-seven stations spanned 1901–4/93 and 100+ stations spanned 1901–84 with the period 1971–74 missing. The data for surrounding countries spanned, at most, 14 years. Results from these areas should be viewed as tentative. For this reason, the emphasis in this study will be on India. Shea and Sontakke (1995; hereafter SS) provide more details on this data set. SS also presents various statistics computed from these data.

3. ESTIMATING CLIMATE NOISE

The climate noise associated with monthly and seasonal precipitation totals was estimated by assuming that daily precipitation could be modeled by a two-state first-order Markov process.

\[ \sigma^2_T = T \left[ q p \sigma^2 + p(1 - p) \frac{1 + d}{1 - d} \mu^2 \right] \] (1)

where \( T \) is the length of the noise process (e.g., 30 or 31 for monthly totals and, say, 92 for a 3-month seasonal total); \( q \) is a term which includes the effects of autocorrelation within sequences of wet days (\( \geq 2.5 \text{mm} \)); \( p \) is the unconditional probability of precipitation; \( \sigma^2 \) and \( \mu^2 \) are the variance and squared mean of precipitation amounts utilizing wet days only; and, \( d \) is a 'persistence parameter'. This model is a more general version of the model first proposed by Katz (1983). Equation (1) includes the occurrence and intensity of the precipitation process and the autocorrelation of sequences of wet days. Because it is more difficult to unambiguously lessen the effects of slowly varying external conditions on precipitation (as opposed to temperature; see MS and SM), it is possible that the estimates of climate noise are inflated. This would cause the F-ratios presented in Section 4 to be conservative.

4. RESULTS

Estimates of the noise and the F-ratios for different time periods within the monsoon season (June through September; J–S) were calculated. These provided some insight into the robustness of the model parameters used and the noise estimates obtained using (1). Qualitatively, all periods exhibited the same pattern. The largest noise estimates coincided with the areas of largest total precipitation (see SS). For example, the values of \( \sigma_T \) for combined July–August (JA; \( T=62 \)) are presented in Figure 1. These values of \( \sigma_T \) may be interpreted to mean that 62-day JA precipitation totals would have a standard deviation of about 142 mm near Bhopal in central India strictly as a result of fluctuations due to daily weather variability.
Figures 2 and 3 present F-ratios for the entire monsoon season (J–S) and for the peak monsoon months (JA). Both figures show that throughout India there exists variance above that attributable to climate noise. The largest F-ratios (≥2.5) are along the west coast where the southwesterly winds associated with the monsoon impinge on the Western Ghat mountain range which runs parallel to the coast. Stronger (weaker) than normal southwesterly monsoon winds will produce above (less than) normal rainfall in these areas. The central and northwestern portions of India show ratios of 1.4 to 2.0. These regions are affected by low pressure systems which originate in the northern region of the Bay of Bengal. These systems travel west–northwestwards along the monsoon trough which generally located over the Ganges Valley. Periods when the monsoon trough move northwards to the foot of the Himalaya mountains are characterized by short period drought over central India and copious rainfall near the Himalaya. The smallest F-ratios (~1.25 or less) are located over southeastern India. During JJAS this region is less affected by the southwest monsoon. This region generally gets the majority of its rainfall during the northeast monsoon which occurs from October through December.

5. SUMMARY

A Markov model has been used to estimate the variance in monthly and seasonal precipitation totals over during the southwest monsoon season (June through September). F-ratios calculated using the observed interannual variability of monthly and seasonal precipitation totals in the numerator ($\sigma_A$) and the variance associated with weather fluctuations ($\sigma_T$) in the denominator indicate that it may be possible to predict fluctuations of rainfall totals over India with long lead times.

References


Figure 1. Estimates (mm) of $\sigma_T$ for the combined July–August precipitation totals.
Figure 2. Estimates of F-ratios for the entire southwest monsoon season (June through September).
Figure 3. Estimates of F-ratios for the peak southwest monsoon months (July and August).
On the Causes of Mild Winters in Northern Europe

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1. INTRODUCTION

The seesaw in winter air temperatures between Greenland and northern Europe (van Loon and Rogers, 1978; hereafter vLR) is generally thought to be linked to interannual variability of the North Atlantic Oscillation (NAO). Figure 1 shows that this presumption is generally correct although the relation between the two is not always significant with 99% confidence. This paper will show that despite the generally high correlations in Fig. 1, the NAO is, at best, only indirectly the cause of mild winters in northern Europe during Greenland Below (GB) events. The evidence leading to this conclusion is part of an analysis about the characteristics and variability of the North Atlantic storm track (Rogers, 1997), described here in Section 3. The Greenland side of the temperature seesaw in GB cases appears to be very strongly dependent on the NAO (Rogers, 1997) and, along with the Greenland Above (GA) seesaw mode, is only briefly mentioned in this paper.

2. DATA

Northern Hemisphere gridded daily and monthly mean sea level pressure (SLP) data are used. The data are available at every 5° of latitude and longitude from 20–85°N for the period November 1899 through March 1992. Gridded maps are available once daily for either 1300Z (from 1899–1939) or 1200Z and are available twice daily (00Z and 1200Z) from 1955–56 through 1959–1960 and for all winters starting with 1962–63. Daily maps are missing from December 1944 through December 1945 but monthly charts are available for calendar year 1945. Other than this 13 month period, missing daily data were replaced by pressure averages of the day prior and the day after. Monthly mean surface air temperature data on a 5°×10° latitude/longitude grid (Jones et al., 1991) for 1899–1990 are used. Monthly surface air temperatures are also used for Oslo (59.9°N, 10.7°E), Norway and Jakobshavn (69.2°N, 51.0°W) and Egedesminde (68.7°N, 52.8°W) on western Greenland. Temperature departures at the latter two stations are combined to make a complete western Greenland record, necessitated by a lack of data at Jakobshavn after 1970.

3. THE NORTH ATLANTIC STORM TRACK

Once-daily gridded SLP data, spanning the period November 27 through March 4 from 1899–1900 through 1991–1992, were high-pass filtered using a binomial filter with weights -0.0625, -0.25, +0.625, -0.25, -0.0625, the weights associated with the low-pass binomial filter of n = 4 (1–4–6–4–1). The filter has maximum response in the 2–8 day periodicity range, typically associated with passage of
The scores of the first rotated principal component are a time series of the primary mode of Atlantic monthly rms fields of high-pass filtered SLPs. Monthly RPCA scores are standardized and have randomly distributed numerical values ranging from -2.1 to +3.3 (Fig. 2). The 32 highest positive and 27 lowest negative monthly scores are identified and used to illustrate spatial changes associated with extremes of the RPCA storm track pattern. Differences in monthly mean rms values (Fig. 3a), obtained by subtracting the mean rms distribution during negative cases from those of positive cases, have a spatial pattern very similar to that displayed in the rotated principal component loadings. The largest mean rms variations form a dipole (Fig. 3a) with centers in the extreme northeastern Atlantic and Norwegian Sea, where the net mean rms differences exceed 4 mb, and over the eastern Atlantic around Portugal. The mean rms differences between the two data sets are statistically significant at the 95% confidence interval, based on a two-tail t-test, over large areas of the northern Atlantic from 50°W to 50°E and from 55°N to 80°N, as well as over southern Siberia at 55–60°N.

The composite mean rms for the 32 largest positive cases (Fig. 3b) exceeds 7 mb over Newfoundland and Labrador, and in the area around Iceland, with values over 5 mb from the East Greenland Sea northeastward to Novaya Zemlya. The composite rms for the 27 negative cases (Fig. 3c) exhibits a maximum near Newfoundland, extending from Maine to nearly the southern tip of Greenland, and comparatively low values in the northeastern Atlantic, only reaching 3–4 mb. The main axis of the rms maximum is oriented toward the Bay of Biscay and the Mediterranean basin. The net mean rms differences around the southern dipole center west of Portugal (Fig. 3a) are statistically significant with 95% confidence from 0 to 30°W and 30–45°N, due to small rms variability at these latitudes in the cases comprising the composites.

Composite means of raw monthly SLPs were obtained after stratifying the RPCA scores into five groups separated at points corresponding to numerical score values of -1.0, 0.0, +1.0 and +2.0 (Fig. 4). This data stratification illustrates changes occurring in mean intensity and spatial locations of Atlantic centers of action as score values change. The composite mean for the set of months with scores between 0.0 and +1.0 (Fig. 4c) resembles closely the long-term mean Atlantic SLP field with a minimum of 996 mb over the Denmark Strait and a subtropical maximum near the Azores at 30°N, 30°W with a central pressure of just over 1024 mb.

In months with scores lower than -1.0 (Fig. 4a), the subtropical high and Icelandic low are weaker than normal and shifted to the south and west of their mean positions (Fig. 4c). The mean low (1001 mb) extends over a large area southeast of Greenland and relatively high pressure between 1010–1016 mb occurs across the northeastern Atlantic and the Barents and Kara Seas to Novaya Zemlya. The Azores high is relatively weak (1021 mb) and lies over the south-central North Atlantic near 25°N, 45°W (Fig.
4a). A trough of comparatively low pressure extends toward the Bay of Biscay and across southern Europe and the Mediterranean Sea. This case corresponds primarily to that of Fig. 3c with comparatively higher than normal mean rms values over the east-central Atlantic, suggesting an active storm track toward Portugal and the Mediterranean and a weak subtropical high.

In months with the highest positive RPCA scores (Figs. 4d and 4e), both the mean subpolar low and the subtropical high extend farther northeast of normal. The mean subpolar low is 994 mb at 70°N 10°E in the Norwegian/Barents Sea area (Fig. 4e), with pressure under 996 mb as far east as 50°E near Novaya Zemlya. The highest rms values of high-pass filtered pressures occur (Fig. 3b) over Iceland and farther northeast suggesting that cyclone activity proceeds northeastward into the Norwegian and Barents Seas and even farther east in these cases. The subtropical high extends northeast of normal, well over the Mediterranean Basin, with a maximum pressure of 1028 mb. The mean SLP is 1008–1012 mb around the Bay of Biscay in Fig. 4a but it is about 1026 mb in Fig. 4e. Hatching in Fig. 4a shows that SLPs are significantly different between Figs. 4a and 4e in areas centered over southern Europe and northern Europe and the northeastern Atlantic.

It is apparent from Fig. 4 that with increasingly positive RPCA scores (1) the mean subpolar low intensifies as it shifts to the northeast, (2) the subtropical high intensifies and migrates northeastward of its mean position, (3) the pressure gradient between the centers of action intensifies as they shift northeastward, and (4) the storm track shifts from a northwest-southeast orientation (for low negative scores) to a southwest-northeast orientation, extending deep into the high Arctic. The eastward shift in the subpolar and subtropical SLP fields is apparent in the eastward shifts in areas of statistically significant pressure differences in Figs. 4b–4d.

4. THE STORM TRACK AND SEESAW IN WINTER SURFACE AIR TEMPERATURES

Winter means of the storm track scores (Fig. 2) were correlated to gridded hemispheric winter seasonal surface air temperatures spanning 1900–1990 (Jones et al., 1991). Statistically significant coefficients of correlation (Fig. 5) have maximum positive values over Ireland, the United Kingdom and southern Scandinavia extending eastward into north-central Asia between 55°–75°N, 70°–100°E. Positive correlations imply higher than normal winter surface air temperatures over Europe and Eurasia (Rogers and Mosley-Thompson, 1995) during months with highest scores, while below normal air temperatures concurrently occur over western Greenland, Baffin Island and over the Mediterranean Basin and northern Africa. Conversely, Africa and the Mediterranean have above normal temperatures when the storm track scores are negative, occurring as storms migrate toward the Mediterranean basin and during which northern Europe and Eurasia have unusually cold winters.

The winter air temperature seesaw between western Greenland and northern Europe can be identified in Fig. 5. Further analysis of the seesaw is based on vLR’s rule for identification of cases: monthly seesaw extreme GB and GA events only occur if the sign of the western Greenland temperature
anomaly is opposite that at Oslo, with an absolute temperature anomaly difference between them (Greenland minus Oslo) exceeding 4°C.

The storm track scores are negative in 55 of 67 GA winter months since 1899 (Fig. 6a), with scores most frequently falling between -0.5 and -1.5. Cyclone activity is concentrated near southern Greenland in these cases while much higher mean pressures occur over northern Europe (Figs. 4a and 4b). On the other hand, the storm track index values have a wider distribution across the 63 GB events with 24 negative cases and only 39 positive (Fig. 6b). The two distributions differ significantly from each other with 95% confidence. Histograms such as these were constructed individually for temperature anomalies greater than absolute 4°C at Greenland and at Oslo, ignoring the seesaw criteria, and were then stratified by storm track scores. The tendency for a broader distribution of storm track scores across positive temperature anomalies is very predominant at Oslo (not shown), more so than is shown in Fig. 6b. The results suggest that, like GA cases, about 40% of GB cases have large synoptic activity along the southern-dipole storm track.

Mean SLP composites are created for the 30 GB cases when the storm track scores are greater than +0.5 and for 24 cases when the scores were negative, ignoring nine "overlap" cases with scores between 0.0 and +0.5. The positive score months (Fig. 7a) are characterized by broad subpolar low pressure with centers west of Iceland and over the Norwegian Sea. Pressures under 1000 mb extend to Novaya Zemlya and the isobars around the double low extend zonally into Europe. The Atlantic subtropical high, as measured by the 1024 mb isobar, extends farther northeast than usual, and strong maritime westerly flow extends well into Europe and Asia. The entire pattern is typical of high RPCA scores in Figs. 4d and 4e.

A deep low also occurs near Iceland in GB cases with negative scores (Fig. 7b), but the isobars to the east generally lie parallel to the Scandinavian coast. Comparatively high pressure covers the Barents and Kara Seas. The Siberian anticyclone, as measured by its 1020 mb isobar, spreads much farther north and west in Fig. 7b than it did in Fig. 7a, while the Atlantic subtropical anticyclone is displaced west, over the mid-ocean basin.

Mean SLP differences between these modes (Fig. 7c) consist of a dipole with centers over eastern Europe and southwest of Ireland with a strong pressure gradient between the two centers lying across much of Scandinavia and northern Europe. The eastern European dipole center is an area of anomalous high pressure in the GB/negative score months (Fig. 7b and 7c) and the flow around this anticyclonic anomaly produces an anomalous southeasterly flow over northern Europe (Fig. 7c) in conjunction with the above normal surface air temperatures. This mild southeasterly return flow on the time-averaged charts, such as Fig. 7b, occurs during periods of westward extension of the Siberian anticyclone. The westward extension of the Siberian anticyclone is a well-known synoptic feature among meteorologists in southern Europe. Makrogiannis et al. (1981) obtained mean SLPs, 500 mb heights and 1000–500 mb thicknesses for 20 winter cases when the Siberian high was displaced to the west and found: (1) the westward extension develops due to negative vorticity advection aloft (it is not entirely be
due to radiational cooling), (2) cyclogenesis in the Bay of Biscay and Mediterranean basin often accompanies synoptic development leading to a westward extended Siberian high, and (3) much of western Europe, and particularly northwestern Europe undergoes significant warm air advection during the westward extension of the anticyclone. Note that (2) is consistent with the synoptic development found here for negative score cases while (3) is consistent with above normal temperatures in northern Europe. The northern European dipole center in Fig. 7c is in the same location where the positive correlation coefficients of Fig. 5 are somewhat lower than at other points between Ireland and eastern Siberia, suggesting that in this area another mechanism beside strong zonal flow (and positive RPCA scores) is linked to higher than normal winter air temperatures.

The dipole centers in Fig. 7c are not the standard centers of action of the NAO. Pressure differences over Iceland and the Denmark Strait are not even statistically significant. In a climatic context, the results suggest that two separate synoptic settings and time-mean flow patterns are linked to mild winter months in the northern European segment of the winter temperature seesaw. The maritime flow producing mild conditions at Oslo in Fig. 7a is conditional on the extension of the low pressure into the Norwegian and Barents Seas and extending into northern Europe. In Fig. 7b the flow is more southeasterly because of the westward extension of the Siberian high: the impact on abnormally high temperatures in Europe may primarily be due to the Siberian anticyclone extension. The Icelandic low, in the sense of the subpolar low over the Denmark Strait, seems to play little direct role in above normal winter air temperatures over northern Europe.

Composite mean pressures are also obtained during the 55 GA winter months, when the storm track scores are negative (Fig. 7d). The mean Icelandic Low is weak and displaced south of Greenland with a trough of low pressure extending northwestward over the Davis Strait. The 1008, 1012 and 1016 mb isobars imply southeasterly flow and a trough over western Greenland, a situation often accompanying GA west-coastal above normal winter air temperatures (Rogers, 1985). A trough over the Norwegian Sea is very weak and high pressure extends westward over much of Europe.

The pressure differences obtained by subtracting GB (Fig. 7b) from GA (Fig. 7d), when the RPCA scores are negative, is shown in Fig. 7e. This pattern now appears similar to the NAO, with centers near Iceland and the Azores, and areas of statistical significance over the Denmark Strait and central Atlantic. The pattern correlation between Fig. 7e and that of the winter NAO (Fig. 2a in Rogers, 1990) is $r = 0.865$ across 76 grid points common to both figures. Pressure differences of 18 mb occur in the Denmark Strait and 8 mb near 35°N, 25°W. The elongated maximum of 6 to 8 mb extending northeastward of the Black Sea is the net result of the westward extended Siberian anticyclone in Fig. 7b and its absence in Fig. 7d. Fig. 7e suggests that the traditional NAO, with centers near Iceland and the Azores, is embedded in its entirety in the realm of negative and weakly positive ($< +1.0$) scores in the storm track index. The highest positive scores ($> +1.0$; Figs. 4d, 4e and 7a) are instead cases when the storm track intensifies to the north and shifts farther east, and arguably is not linked to the NAO. The implied geostrophic flow variations around the Denmark Strait center indicate a stronger northerly (southerly) flow in GB (GA) over Greenland. The SLP anomalies illustrated in Fig. 7e represent the GA
cases and cold flow over Europe would originate in the northeasterly flow across the Barents Sea and into Scandinavia.

5. HOW CLOSELY IS THE NAO LINKED TO THE SEESAW?

a. Storm Track and Mean Circulation Considerations

This study links high RMS scores (>+1) with (1) high rms variability in the extreme northeastern Atlantic (Fig. 3b), (2) deep mean low pressure over the Norwegian and Barents Seas at the expense of a separate low over the Denmark Strait (Figs. 4d and 4e) and (3) strong zonal flow over Europe (Fig. 7a) linked to above normal surface air temperatures as far east as Siberia (Fig. 5; see also Rogers and Mosley-Thompson, 1995).

This study also shows that the GB mode of the winter air temperature seesaw can be explained by two separate sea level circulation patterns. Europe has mild winters (1) when the storm track and the mean subpolar low extend into the Norwegian and Barents Seas bringing strong maritime zonal flow far into northern Europe (Fig. 7a), and (2) when the storm track does not extend beyond Iceland, and the mean low lies over the Denmark Strait with isobars parallel to the Scandinavian coast (Fig. 7b). The latter case is the less common but in these winters the NAO centers of action are near their normal ocean basin positions (Fig. 7b) and a westward-extended Siberian anticyclone assists in producing a strong southeasterly flow into northern Europe. The first case, with strong European zonal circulation (Fig. 7a), has long been considered the NAO-based cause of GB events. This is indeed the more frequent mechanism for GB events (Fig. 6), but it is brought about by the strong northeastward-extended storm track with a deep trough in the Norwegian Sea occurring in conjunction with northeastward movement of the subtropical high. This case of maritime flow is arguably linked here to a non-NAO eastward extension of Atlantic cyclone activity.

The argument of whether the NAO has an atmospheric circulation and climatic imprint extending far into Europe hinges on whether the Atlantic subpolar low is really "Icelandic" when there is either (1) a Norwegian or Barents Sea pressure minima occurs or (2) a deep trough occurs, extending northeastward from Iceland. On the long-term climatological charts, the wintertime Icelandic low almost always appears over the Denmark Strait with a trough extending to the northeast, much as in Figs. 4b and 4c. For case (2) the strength of the trough is at issue, reflecting the amount of eddy activity occurring in the extreme northeastern Atlantic. Some examples of unusual winter pressure distributions include that of Fig. 8a in which an isolated mean low pressure center is found over the Norwegian Sea (case #1 above). Figure 8b illustrates the mean pressure in a set of months when a deep mean low is located over the Barents Sea and a very weak secondary low lies near Labrador/Newfoundland. Figure 8c illustrates the mean pressure field for cases with a split low, with two deep centers south of Greenland and in the Barents Sea. Each of these instances show substantial deviation from the long-term mean pressure distribution, which exhibits a single mean "Icelandic Low" in the Denmark Strait, and each is associated with strong zonal flow extending far into Europe.
b. Methodological Considerations in Defining the NAO

This study and two others suggest the possibility that sea level Atlantic regional teleconnections other than the NAO may exist. Hsu and Wallace (1985) for example identify the sea level NAO and another dipole pattern over western Europe and North Africa which they did not illustrate. Rogers (1990) identifies an additional three Atlantic patterns beside the NAO. Most notable among them is the “SENA” pattern which is a dipole, like the NAO, with centers shifted about 50° to the east of those in the NAO and located over Southern Europe and the Northeastern Atlantic (Rogers, 1990; his Fig. 4). The SENA index is significantly correlated, as the NAO, to the winter air temperature seesaw (Rogers, 1990). Methodologically, Hsu and Wallace (1985) and Rogers (1990) identified a relatively large number of principal component patterns over the Northern Hemisphere and the NAO centers of action over the ocean basin were oriented such that the Icelantic low has isobars roughly parallel to the Scandinavian coast with little suggestion of strong zonal flow into Europe (similar to Fig. 7b).

A majority of earlier EOF studies were designed to only identify two or three hemispheric teleconnection patterns, one of which was typically the NAO. In these cases, the NAO seems to "acquire" the variability of surrounding areas and the Atlantic centers of action are placed somewhat east of their mean position with zonal flow between them extending far into Europe. This is observed in SLP-based analyses of Kutzbach (1970; his Fig. 1a), Rogers (1981; his Fig. 1a), Trenberth and Paolino (1981; their Fig. 4a), and Wallace and Gutzler (1981; their Fig. 4) as well as in studies in which the NAO is identified using a pressure index (Hurrell, 1995; his Fig. 1b) or one-point pressure correlations (Wallace and Gutzler, 1981; their Fig. 8). It is also noted in the index-based analysis of Rogers (1984; Figs 4 and 5) who points out that the strong-zonal pattern is less apparent in earlier decades of this century. Studies identifying only 2–3 EOFs, or which use raw pressure data indices or one-point correlations, generally identify an expansive NAO with zonal flow across Europe, while EOF studies identifying many patterns isolate the NAO to the Atlantic basin.

c. Comparison to Storm Tracks in Rogers (1990)

Finally, comparison is made between the results of this study and the cyclone trajectories obtained in the extremes of other SLP low-frequency teleconnections (Rogers, 1990; Fig. 9). The cyclone tracks in the extremes of the NAO (Rogers, 1990; Figs. 9a and 9b) are characterized by large latitudinal differences over the central Atlantic. The NAO positive mode has maximum cyclone frequency near the Denmark Strait with few cyclones occurring east of Iceland, and the cyclones have a path toward the Bay of Biscay in the NAO negative mode but they do not enter the Mediterranean basin. Each of the other three Atlantic sector sea level teleconnections (Rogers, 1990) have one polarity mode characterized by a pronounced cyclone frequency maximum to the east or northeast of Iceland (along with a northeastward extended mean subpolar low), while in the other phase there is a tendency for cyclones to penetrate into the Mediterranean. The results of this and the earlier study suggest that while the NAO is linked closely to the structure of the storm track, in terms of its latitudinal variability, the easternmost non-NAO sea level teleconnections may be linked to the intensification and extreme
eastward extensions of the storm tracks into either the northeastern Atlantic (positive mode) or the Mediterranean basin (negative mode).

6. CONCLUSIONS

There are two circulation-related causes of mild winters in northern Europe. The primary mechanism (about 60% of cases) is due to strong zonal flow into northern Europe. The zonal flow is associated with an Atlantic storm track extended far northeastward of Iceland and toward the high Arctic, where it influences air temperatures across Siberia (Rogers and Mosley-Thompson, 1995). It is associated, on the monthly mean sea level pressure charts, with an Icelandic low having either (1) a deep trough displaced northeastward of Iceland or (2) a northeastward displaced subpolar low. In both cases zonally oriented isobars cross the northern Atlantic and extend into Europe (Fig. 7a). The second cause of mild winters in northern Europe is one in which the storm track is displaced southward toward the Mediterranean basin. The Icelandic low maintains a normal position (in the Denmark Strait) and may even be deeper than normal (Fig. 7b). The isobars in the mean field form only a weak northeastward trough and they are oriented parallel to the Scandinavian coast. The NAO plays little role in this situation, with insignificant pressure variations over the Denmark Strait, and the cause of the mild winter is southeasterly return flow around a westward displaced Siberian anticyclone. The southward displaced storm track brings anticyclones across south-central Europe that help build the westward displacement of the mean Siberian high on the monthly charts (Makrogiannis, 1981).

The question arises as to whether the first case, with a northeastward displaced storm track and eastward displaced mean zonal flow, should also be considered the NAO. As discussed in Section 5a, the displacement of the mean subpolar low far northeastward of Iceland is an anomalous circumstance, arguably no longer making it the "Icelandic" low (see examples in Fig. 8). The dipole in Fig. 7c is observational evidence that the NAO is not a statistically significant factor in these cases. Section 5b points out that spatial characteristics of the NAO, and the subpolar low, vary in RPCA-based studies depending upon how many rotated patterns are retained in the analysis. The northern subpolar NAO center is confined to the Denmark Strait when more than 5–6 hemispheric teleconnection patterns are identified. Isobars about the low are then confined to the ocean basin and lie parallel to the Scandinavian coast. Other EOF-based studies that are limited to identifying only 2–3 eigenvectors, typically identify a broader eastward extended NAO with zonal flow into Europe. The same is typically true of studies examining the pressure field after using raw data to form NAO indices or using raw data with correlation analysis. It is clear however, that the storm track frequently extends far beyond Iceland and that the mean subpolar low can appear elsewhere beside its mean position in the Denmark Strait (Fig. 8). The argument that the sea level NAO is the only regional teleconnection pattern would make it unique in light of the multiple patterns identified in upper air studies.
This study has shown, if nothing else, that analysis of high frequency pressure data can be useful in identification of subtleties in climatic variability that might otherwise be difficult to distinguish in monthly or seasonally averaged fields.

Acknowledgments

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References


Figure 1. Coefficients of correlation between the winter index of the North Atlantic Oscillation and the index of the seesaw in winter air temperatures between Greenland and Northern Europe. The coefficients are for running twenty year periods dated on the final year. The dashed line \((r = +0.60)\) corresponds to the 99% confidence interval.

Figure 2. Time series of scores associated with the first rotated principal component of the monthly rms fields of high-pass filtered sea level pressures. Monthly (thin solid line) and seasonal (thicker solid line) values are shown, including zero values for the missing data from December 1944 through 1945.
Figure 3. Composites of monthly root-mean-squares (in mb) for sets of months with extreme opposite modes of the first principal component of Atlantic area root-mean-square fields of high-pass filtered sea level pressures, 1900–1992. The three diagrams include the (a) net mean rms differences (mb) between the (b) composite positive mode cases and (c) the composite negative cases.
Figure 4. Composites of Atlantic mean sea level pressures (mb) for subset groups when the monthly scores (Fig. 2) of the first principal component monthly root-mean-squares of high-pass-filtered sea level pressures are (a) lower than -1.0; (b) between -1.0 and 0.0; (c) between 0.0 and +1.0; (d) between +1.0 and +2.0; and (e) for cases higher than +2.0. Lighter and darker hatching represent areas where the differences in pressure are statistically significant with 95 and 99% confidence between different combinations of maps. Hatching on (b) through (e) represents significant differences with the preceding map while hatching on (a) represents significant differences between (a) and (e).
Figure 5. Spatial distribution of coefficients of correlation between RPCA scores of the Atlantic storm track eigenvector and gridded winter mean air temperatures for land areas of the Northern Hemisphere, 1900–1990 (from Jones et al., 1991). Correlation coefficients of $r = \pm 0.32$, $r = \pm 0.44$ and $r = \pm 0.55$ are significant at the 95%, 99% and 99.9% confidence levels. Positive coefficients are represented by dashed lines.

Figure 6. Frequencies of RPCA storm track scores, at increments of 0.5, during individual winter months 1900–1992 when the GA and GB modes occur of the seesaw in winter air temperatures between Greenland and northern Europe.
Figure 7. Mean sea level pressures (mb) when the GB seesaw mode occurs and the RPCA storm track scores are (a) positive and (b) negative, and (c) the net pressure differences, (b) minus (a), between those sets of cases. Mean sea level pressures are also shown (d) for the GA seesaw cases that occur with negative RPCA storm track scores and for (e) the net pressure differences for the sets of cases, (d) minus (b). Lighter and darker hatching on (c) and (e) represent areas where the differences in pressure are statistically significant with 95 and 99% confidence.
Figure 8. Examples of mean sea level pressure (mb) in the northern Atlantic during winter months (a) when a deep mean low occurs over the Norwegian Sea, (b) when a deep mean low occurs over the Barents and Kara Seas, and (c) when a double low occurs south of Greenland and in the Barents Sea.
Figure 9. Isopleths of the frequency of wave cyclones, per 5° latitude x 5° longitude tesserae, for nine winter months, 1957–1986, when the rotated principal component scores for the Atlantic–European sector teleconnections are highest (Figs. a, c, e, and g) and for nine months when they are lowest (b, d, f, and h).
Is there a Warm Season Relationship Between Precipitation Over the United States and Tropical Pacific Sea Surface Temperature?

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1. INTRODUCTION

The influence of the El Niño-Southern Oscillation (ENSO) cycle on U.S. warm season precipitation has been examined by a number of investigators over the past decade. Ropelewski and Halpert (1986) analyzed the ENSO signal in surface temperature and precipitation over the U.S. by characterizing the ENSO events in terms of a two-year cycle and then fitting a two-year harmonic dial representation to the two year cycle. Adjacent stations were then aggregated over areas which exhibited similar phasing in an attempt to identify regional ENSO responses in precipitation. Their analysis brought out the well documented ENSO cold season precipitation anomalies over northern Mexico and the southeastern U.S., but results for the warm season were somewhat ambiguous. This could be due to the lack of a consistent signal or to deficiencies in the broad-brush method used to isolate the ENSO signal.

Trenberth et al. (1988), in an analysis and model simulation of the spring and early summer drought of 1988 over the upper Mississippi Basin, concluded that for this particular event “large-scale atmospheric circulation perturbations associated with natural variations in the coupled atmosphere-ocean system in the tropical Pacific were most likely the primary cause of the drought”. Since this was a single case study, and since the drought episode occurred primarily during the spring and early summer seasons (April–June), the relationship of these results to typical summer season linkages is unclear.

More recently, O’Brien and Sittel (1995) found little evidence of a consistent ENSO cycle warm season modulation of precipitation over the U.S. Taken together, these mixed results left the issue of an ENSO modulation of U.S. warm season precipitation unresolved. On the other hand, more recent studies, some of which have not yet been published, seem to clarify this issue, in that they all tend to point to a rather consistent relationship between tropical Pacific sea surface temperature (SST) and warm season precipitation east of the Rockies. These newer results are summarized in Section 3 of this review. First, however, the broad scale features of the North American summertime circulation and precipitation regime described in Section 2 provide a climatological setting for the discussion of non-seasonal variability. This discussion is largely a synthesis based on a number of studies during the past decade that together reveal the seasonal-evolving elements of a North American warm season monsoon system. Finally, a summary and further interpretation of the more recent findings is contained in Section 4.
2. NORTH AMERICAN SUMMER CLIMATE REGIME

The mean seasonal evolution and interannual variability of the North American warm season monsoon system provides a useful framework for describing, understanding and modeling the warm season precipitation regime over the U.S. and Mexico. Monsoon circulation systems develop over low latitude continental regions in response to the seasonal change in the differential heating between the continent and adjacent ocean regions and the seasonal change in the distribution of continental heating associated with terrain and land surface conditions. The broadscale features of the North American summer monsoon system can be described in terms of development, mature and decay phases.

a. Development Phase (May–June)

This is the period of transition from the cold season continental-scale circulation regime to the warm season regime. There is a decrease in mid-latitude synoptic-scale transient activity over the U.S. and northern Mexico as the extratropical storm track weakens and migrates poleward to its mean summertime position near the Canadian border. By late June the increased role of continental forcing is apparent. The onset of the Mexican Monsoon (Douglas et al., 1993), a major component of the continental monsoon system, is marked by the outbreak of heavy rainfall over southern Mexico which quickly spreads northward along the western slopes of the Sierra Madre Occidental into Arizona and New Mexico by early July.

The upper troposphere stationary wave pattern undergoes a significant evolution from June to July. The general height increase in the middle latitudes associated with the seasonal heating of the troposphere is not zonally uniform. Over the North American sector, the largest increases in height occur over the western and central U.S., which likely results from a combination of enhanced atmospheric heating over the elevated terrain of the western U.S. and Mexico, and increased latent heating associated with the development of the Mexican Monsoon. An upper troposphere "monsoon high", analogous to the Tibetan High over South Asia and the warm season Bolivian High over South America migrates northward over Mexico during June, becomes established near the head of the Gulf of California by early July, and remains in this general vicinity during the next two months.

b. Mature Phase (July–August)

The North American Monsoon System is fully developed by early July, and changes during July and August are relatively small. The upper tropospheric circulation and associated changes in the divergence field (mean vertical motion) from June to July can be related to the seasonal evolution of the continental precipitation regime. A region of enhanced upper troposphere divergence in the vicinity and south of the upper troposphere monsoon high coincides with enhanced upper troposphere easterlies or weaker westerlies and enhanced Mexican Monsoon rainfall. In contrast, the upper tropospheric flow is more convergent and rainfall diminishes from June to July in the increasingly anticyclonic westerly flow to the north and east of the monsoon high (Tang and Reiter, 1984; Mock, 1996). There is also some
indication of increased divergence and precipitation in the vicinity of an "induced" downstream "troughing" over the eastern U.S.

c. Decay Phase (September–October)

The evolution of the monsoon system during its decay phase can be broadly characterized as the reverse of the changes during the development phase, although the process tends to proceed at a slower pace. The western U.S. ridge weakens, as the monsoon high and Mexican Monsoon precipitation retreat southward into the deep tropics. Other features associated with continental forcing, such as diurnal variability diminish.

The evolutionary nature of the monsoon system during the warm season has important implications for the analysis of non-seasonal variability. More specifically, non-seasonal anomalies generally represent a small perturbation on the climatological base. The evolutionary nature of the monsoon circulation and precipitation regimes throughout much of the warm season leads climatological month-to-month changes in the base state which are as large or larger than the non-seasonal anomalies, and this raises a serious question regarding the typical approach of most studies, including those discussed in this review, which examine only multi-month seasonal averaged conditions.

3. TROPICAL PACIFIC SST – U.S. PRECIPITATION RELATIONSHIPS

A number of recent analyses point to a warm season relationship between tropical Pacific SST and precipitation over the central and eastern U.S. First of all, the summer (JJA) warm and cold phase ENSO cycle composites which Higgins et al. (1996) included in their recent atlas of hourly precipitation statistics (1963–1993) show a distinct warm phase increase (cold phase decrease) in precipitation over the upper Midwest and a tendency for anomalies of opposite sign over the southeast and mid-Atlantic states (Fig. 1).

The upper Midwest anomalies which appear on these composites are consistent with the results of Bunkers et al. (1996). They composited precipitation data from stations within and immediately adjacent to the Dakotas for ENSO cycle warm and cold episodes during the period 1880–1990. They identified a highly significant warm episode signal (increased precipitation) during April–October, and a highly significant cold episode signal (decreased precipitation) during May–August.

X. Wang (personal communication, 1996) has examined the relationship between variations in mean seasonal SST in the Niño3 index area and U.S. climate division precipitation (1895–1995). His composites for seasonal SST departures from the 100-year mean exceeding one standard deviation bring out the dipole pattern over the central and eastern U.S. even more clearly than the Higgins et al. composites.

Further evidence for a tropical Pacific SST-U.S. summertime precipitation link is provided by the results of a study by Ting and Wang (1996). Specifically, the leading mode of an SVD analysis of the
relationship between Pacific SST and Great Plains rainfall showed a spatial and temporal pattern in the SST field that is broadly characteristic of the ENSO cycle, and a relationship between the SST variations and Great Plains consistent with the results of the other studies reviewed in this section.

4. DISCUSSION AND CONCLUSIONS

At least three phenomenological features of the climatological warm season circulation could link the North American Monsoon System to tropical Pacific variability: (1) changes in the position and intensity of the eastern Pacific ITCZ, which clearly affect the warm season precipitation regime over at least central America and southern Mexico (Cavazos and Hastenrath, 1990; Ropelewski and Halpert, 1987; 1989), (2) changes in the upper tropospheric westerly flow around the low-latitude Mid-Pacific Trough, which extends to the southwestern U.S., and (3) ENSO-related changes in the mid-latitude stationary waves and associated storm tracks.

Focusing on the last of these, the North American summer monsoon circulation system is bounded on the north by the mid-latitude storm track, as discussed in Section 2. The interaction between the mid-latitude stationary waves and associated storm tracks and the monsoon system is quite pronounced. Over the past three decades, a variety of studies of summer drought and wet regimes over the Midwest have revealed their relationship to a characteristic subtropical/mid-latitude geopotential anomaly pattern in the middle/upper troposphere. This continental-scale anomaly pattern reflects changes in the intensity and/or configuration of the upper troposphere monsoon ridge and its neighboring troughs over the eastern Pacific and eastern U.S. (see, for example, Fig. 11 from Mo et al. (1995)). Characteristic anomalies in the lower troposphere include a weakening/strengthening of the Great Plains low level jet, which is in turn associated with a diminished/enhanced northward inflow of moisture into the central U.S. (Mo et al., 1995).

The characteristic upper troposphere geopotential anomaly pattern was more or less in evidence during the 1988 drought (Trenberth et al., 1988). Furthermore, Ting and Wang (1996) showed that this upper troposphere anomaly pattern to be associated with their SVD 1 mode. Unfortunately, geopotential anomaly fields do not adequately reflect circulation anomalies in the equatorial belt, and so cannot show whether there is indeed a direct connection between the upper troposphere circulation anomalies in the subtropics and middle latitudes and the characteristic ENSO cycle circulation anomalies in the equatorial belt i.e. the anticyclonic/cyclonic couplet. To establish this linkage between middle latitude and equatorial circulation anomalies, one must analyze the stream function anomaly field.

This has been done by Y. Dai as part of her PhD thesis research at the University of Maryland (Y. Dai, 1996, personal communication). Figure 2, derived from ECMWF analysis for the period 1985–1995, shows the leading mode of a rotated principal component analysis of combined interannual variability of the 200 and 850 mb streamfunction, SST and the upper- and lower-level diabatic heating anomalies during the 1985–1995 summer months (May–August). The relationship between tropical Pacific SST and the classical wet/dry circulation anomaly pattern over the U.S. is apparent, but in this
case, the analysis clearly shows the equatorial cyclonic/anticyclonic couplet characteristic of the SST anomaly pattern, thus forging the link between the ENSO cycle circulation anomalies and the wet/dry circulation anomaly pattern over the U.S. Recognizing that this a relatively short time series, Ms. Dai also performed a similar analysis of the Oort data set (1964–1989) with similar results.

Since the Mexican Monsoon is a major component of the North American summertime Monsoon system, and since it is also associated with the mean seasonal development of the upper troposphere monsoon ridge, whose modulation plays a central role in the interannual variability of precipitation over the upper Midwest, it is important to establish whether the Mexican Monsoon is also related to tropical Pacific SST. As previously noted, empirical studies have clearly established a statistical linkage between the ENSO cycle and precipitation variability over central America and the southern Mexico, but the correlations weaken to the north, where the relationship, if any, is unclear. There is some evidence of a negative correlation between Mexican Monsoon precipitation and precipitation over the upper Mississippi Basin (PACS Steering committee, 1994). It is intriguing that this pattern of non-seasonal variability is somewhat similar to the mean seasonal changes in the precipitation pattern associated with the development of the monsoon ridge from June to July.

This anomaly pattern, which was observed during the summer of 1993, suggests a continental-scale mode of variability that is similar in character to the seasonal changes associated with the development and decay of the continental monsoon system. However, a cautionary note is in order at this time, since analyses by Mathew Barlow at the University of Maryland (M. Barlow, personal communication, 1996), have identified a correlation pattern similar to that shown in the PACS document, but the correlations he is finding are too weak to be of much practical significance.

References


Figure 1. Composite difference from normal (1963-1993) June-August seasonal precipitation (mm day$^{-1}$) for ENSO warm (left) and cold (right) events. Solid (dotted) lines denote positive (negative) anomalies. Light (dark) shaded regions indicate where anomalies are ≥ 25% (50%) of the mean. From Higgins et al., 1996
Figure 2. The leading mode of a rotated principal component analysis of combined interannual variability of the 200 and 850 mb streamfunction (a,b), SST (c) and the upper-level (200–500 mb) and lower-level (700–200 mb) diabatic heating anomalies (d,e) during the 1985–1995 summer months (May–August). The contour interval is indicated on the top of each panel. The common coefficient time series is shown in the bottom right panel (f). This mode explains 12% of the combined domain variance. Courtesy Y. Dai.
Quasi-stationary Waves in the Southern Hemisphere of a GCM, with and without an Interactive Ocean

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1. INTRODUCTION

In this study, some aspects of the Southern Hemisphere (SH) atmospheric circulation as simulated by the Climate System Model (CSM) and the National Center for Atmospheric Research (NCAR) Climate Community Model, Version 3 (CCM3) are examined and some initial findings are presented. The motivation for this research is twofold: first, to evaluate how well these models emulate the atmospheric circulation (in particular the SH) and second, if they prove reliable, as appears to be the case, to then use them as mechanistic tools to examine the generation and maintenance of the quasi-stationary waves in the SH and the cause of decadal variation in the SH. In this paper the temperature distributions and the principal quasi-stationary waves in the geopotential height field simulated by the CSM and the CCM3, are compared with similar variables from the NCAR/National Center for Environmental Prediction (NCAR/NCEP) reanalyses.

2. DESCRIPTION OF MODELS AND DATA

The CCM3 is a three-dimensional global atmospheric general circulation model (GCM). It descends directly from the CCM2 and has improved physical representation of specific climate processes. These include substantial improvements to the top of the atmosphere and surface energy budgets, a reduction in the magnitude of the hydrological cycle, and a more realistic distribution of tropical precipitation. This results in a better simulation of the extra-tropical stationary wave pattern. The CSM is NCAR’s new comprehensive climate system model which includes the CCM3, and component models for a dynamic, interactive ocean, sea-ice and land surface processes (Acker, et al 1996). The stand-alone CCM and the CCM3 in the CSM both have 18 levels in the vertical and realistic orography. Their resolution is a triangular truncation at wavenumber 42 (T42) on a Gaussian grid. The fine vertical and horizontal resolution of the models allow the smaller scale climatic processes to be better represented.

CCM3 results from an integration performed with specified thermal forcing using observed, monthly averages of sea surface temperatures (SST) for the period 1979–1993 (from the Shea, Trenberth and Reynolds (1988); hereafter, STR) climatology, NCAR/NCEP reanalyses for the concurrent period and CSM simulations for 15 years are compared. In the CSM, climatological SSTs from the STR climatology are used as an initial condition and the surface temperature that is used by the model atmosphere is obtained from the interaction between the model ocean and the atmosphere. The model
creates its own surface temperatures by merging these SSTs with those from the land surface and ocean ice models. The CCM3 is run without an ocean. Instead, daily SST values are interpolated from the monthly averages and the model merges these with the temperatures produced by the land surface submodel to create the surface temperature used by the model atmosphere. The CCM3 and the CSM are similar in most aspects, the chief difference being the part in thermal forcing played by the interactive ocean in the CSM. This makes their comparison also a comment on the role of an interactive ocean in climate.

The NCEP Reanalyses are an improvement on the old National Meteorological Center data. They represent an international effort in which historic data (1957–1996) were reanalyzed, incorporating data and methods that were previously unavailable. NCEP reanalyses lack the inhomogeneities that have been introduced into the operational analyses due to changes in the NCEP model’s data assimilation system over many years. The new analyses are computed at a much finer resolution (T62) and do not contain model-induced climatic shifts. In order to make them directly comparable to the model data, the NCEP data are spectrally truncated to T42 before processing.

The variables compared here are the geopotential height and temperature for the southern winter (July to September, JAS), a period when the circulation of the atmosphere is more vigorous. Evaluation of the geopotential height variations simulated by the models is effectively an evaluation of the ability of the models to simulate the atmospheric circulation. Fourier analysis is used to decompose the geopotential heights into the individual waves and the first three waves are extracted. In the observed data, the first three waves explain most (>90%) of the variation in the geopotential height field (van Loon and Jenne, 1972; Trenberth, 1980). For each simulation, the amplitude, phase and percent variation explained by these waves are compared to the observed data (NCAR/NCEP reanalyses).

The temperature distributions generated by the models for the 1000 mb level is examined. The waves in the geopotential height field in the SH are thought to be forced by orography and by surface temperature variations. This is particularly true for Wave 1, the principal fluctuation in the southern circulation. Therefore, it is necessary to examine the temperature distributions that the models use in their simulation of the atmospheric circulation. Although the focus here is on the SH, some references to the simulation of the Northern Hemisphere’s atmospheric circulation will be made.

3. RESULTS

a. Temperature

In the SH, the 1000 mb temperature differences (Fig. 1) are rather small except along the Antarctic continental margin where the large differences are associated with variations along the ice margins. Poleward of 60°S, the surface temperature of CSM is greater than that of CCM3 except near the Antarctic peninsula. The warmer CSM surface temperatures are attributed in part to the fact that a grid point is considered completely covered in ice if any ice exists in the CCM3 whereas the CSM is
more conservative, allowing partial coverage. Therefore the temperature differences in the Antarctic are largely an artifact of ice assignments. Equatorward of 60°S, the CSM produces colder temperatures than the CCM3.

The zonal deviations of the 1000 mb surface temperatures produced by the models and the observed (Fig. 2) exhibit a close resemblance to each other. Given that the temperature distributions are similar, this is not surprising. Continental regions in the subtropics are cooler than the zonal average. This is the SH winter so such a pattern is expected from the influence of continentality. Over Antarctica there is a clear Wave 1 apparent in the zonal deviations; the eastern hemisphere is cooler than the western.

b. Geopotential Height Variations

Wave 1. In the vertical cross-section of the observed geopotential height Wave 1, there are two well-defined peaks in amplitude in the upper troposphere; one between 30–40°S (at 300 mb) and the other which propagates into the stratosphere between 55–60°S (Fig. 3a). In the subpolar region, the location and shape of the simulated Wave 1 match that of the observed very well. Differences lie in the amplitudes of each structure; the CCM3 has a stronger amplitude than the observed by 20 gpm while the CSM is weaker by 10 gpm. In the subtropics, the CSM simulates the morphology of Wave 1 better but its amplitude is higher than the observed by 30 gpm. In the CCM3, Wave 1 in the subtropics propagates into the stratosphere and has a weaker amplitude than the observed, by 10 gpm. Some 80–90% of the variance in the observed data is explained by Wave 1. Similar levels of explanation is offered by the CCM3, but for the CSM, this value is lower, 50–90%.

Geographically, the ridge of Wave 1 is located over the eastern Atlantic in the subtropics and over the central Pacific at subpolar latitudes. It has a phase reversal at 40°S where the wave nearly vanishes (Fig. 4). Both models place the ridge and trough of Wave 1 in the correct locations as dictated by the observed. The amplitudes of the CCM3 simulated Wave 1 are larger than the observed but that of the CSM are closer in magnitude. Note also that CCM3 does not exhibit a clear Wave 1 over Antarctica but the CSM does, albeit a weak version. The ridges of this Wave 1 are not accurately placed since they are a bit east of the observed. Also, the ridge of the CCM3 Wave 1 is further east than the observed in the subtropics. Wave 1 in the SH is thought likely to be forced thermally, since it follows closely the zonal pattern of temperature anomalies (Fig. 2). Both models simulate the distribution of these zonal temperature anomalies well and this may underlie the excellence of their simulated Wave 1.

At 300 mb model simulation of the phase of the wave (not shown) is also good. This indicates that the phase of the simulated wave does not change with height. Unchanging phase with height is an important feature of the observed SH circulation and is the reason why there is little net poleward transport of heat by the stationary waves. In the SH, the pressure and temperature waves are in phase so there is no phase tilt with height. This is also true of the simulated values. Again, the amplitudes of the modeled waves are larger than the observed except in the subtropics where the CSM is very close to the observed.
In the NH, the models do not simulate Wave 1 as well as they do in the SH. In the observed (Fig. 2) there are two distinct features, one centered upon 30°N peaking close to the surface and between 100–200 mb and the other at 70°N peaking between 200–300 mb. The first feature is due to the Asian summer monsoon, while the second is due to the weakened subpolar jet. Wave 1 is better developed in the CCM3 than in the CSM, in particular the subpolar feature. The difference in simulation may be due to problems in the ice distribution experienced within the ice submodel in the CSM.

Wave 2: The vertical cross-section of Wave 2 in the observed data in the SH has a peak in amplitude in the upper troposphere at 35°N and 70–80°S (Fig. 5). In CCM3 the subtropical peak in the amplitude of Wave 2 lies between 25°S and is weaker than the observed while the subpolar peak coincides with the observed but is almost twice as large. In the CSM, there are two well-defined peaks in Wave 2, both of which are greater than the observed. The subtropical peak is in the correct position, approximately, but the subpolar peak is poleward of the observed. The latter closely resembles the structure of Wave 1 so it might be due to some leakage of the power from Wave 1. Some 20–30% of the variance is explained by this feature. In the subtropics, 15% of the variance is explained while over Antarctica approximately 50% is explained.

Wave 2 lies over Antarctica with ridges near 90°E and 90°W (van Loon and Jenne, 1973; Fig. 6). Therefore the phase of Wave 2 appears associated with the presence of Antarctica. This wave is well simulated by CCM3 but not so well by the CSM. In both models the phase of the wave appears associated with Antarctica. The CCM3 simulates Wave 2 in the subtropics and subpolar regions with the approximate amplitudes of the observed but the spatial distinction between both features is not well-preserved. The CSM exhibits a strong Wave 2 close to 60°S but no Wave 2 in the subtropics.

Wave 3: The vertical cross-section of Wave 3 in the observed data in the SH shows a peak in amplitude at about 55°S and between 200–300 mb (Fig. 7). Here it explains no more than 10% of the variance. In the CSM, Wave 3 is much like the observed but weaker. It also explains less than 10% of the variance. The CCM3 exhibits no Wave 3.

In the observed, Wave 3 lies between 25°S and 60°S with ridges in the vicinity of the three lower latitude continents (Fig. 8). This indicates that the phases of this Wave have an association with the land masses. While the CCM3 does not simulate Wave 3 in the subtropics, it is present in the CSM. In the latter, Wave 3 is weaker and its ridges are located east of the observed. The fact that the CCM3 does not simulate Wave 3 may be because the land/ocean contrast is not well represented in that model.

4. SUMMARY AND CONCLUSIONS

Despite the fact that one model is forced by specified SSTs while the other generates its own from an interactive ocean, Wave 1, the most important variation in the SH circulation wave is well represented by both models in both the subtropical and subpolar latitudes. That the CCM3 can produce Wave 1 well in the SH supports the argument that this wave is thermally forced by ocean surface
temperatures. Moreover, given that the zonal anomalies of temperature match the observed, then this suggests that the thermal forcing is key to the generation and maintenance of Wave 1. It emphasizes that Wave 1 is a stable and necessary fluctuation in the general circulation of the southern atmosphere. In this respect, it also shows that both the CSM and CCM3 models perform well and that monthly averages are good enough for simulating the largescale features of the general circulation in the SH.

Wave 3 is not simulated by the CCM3 while the CSM produces a credible facsimile of the observed. The difference between the two models is the interactive ocean. Wave 3 may be amplified by the contrast between land and ocean which is not present in the CCM3 boundary conditions but is created in the CSM. Wave 3 may also be affected by differences in convection over land and this will be further explored by examining the difference in convection between the two models.

The major consequence of this study is the determination that both the CSM and CCM3 simulate the important features of the circulation very well (Wave 3 accounts for very little of the variation in the data). This has been achieved by verification with the NCAR/NCEP reanalyses and by validating the models against each other. This is a significant achievement since previous GCMs have been unable to simulate well all the major features of the SH general circulation. (For example Xu et al., 1990). Some studies have been able to simulate the primary Wave well (e.g. Quintanar and Mechoso, 1995) but the models under discussion perform better in many more aspects. The possibilities for explanation that arise from these models' potential as mechanistic tools are enormous. Also, the results of this study show that an atmospheric model alone can be used to study the mean circulation of the SH.

The distribution of SSTs, the asymmetry (around the South Pole) of Antarctica, the orography of Antarctica have all been suggested as generation mechanisms for the fluctuations in the circulation. Some have suggested forcing from lower latitude transient activity (Quintanar and Mechoso, 1995). However, it is not yet firmly proven what the relative roles of these elements are in forcing the general circulation. In the NH, the roles of thermal forcing and topography in the generation of the stationary waves in winter is clear. However, in the SH the same is not true. One reason is that the arrangement of land in the SH is different and there is less land in the SH. Also, the physical processes underlying the climate are better understood and specified in work that focused on the NH. Now that the CCM3 and the CSM have proven to be reliable, we will conduct a number of experiments designed to alter the symmetry, orography and SST distribution in the CCM3, in order to isolate their role in forcing the primary stationary waves in the SH.

Acknowledgments

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References


Figure 1: 1000 mb temperature differences CSM minus CCM3. Contour interval is 2 K.
Figure 2: Zonal deviations in surface temperature at 1000 mb for (a) NCAR/NCEP Reanalyses; (b) CCM3; and (c) CSM. Contour interval is 2 K.
Figure 3: Amplitude of quasi-stationary wave 1 for (a) NCAR/NCEP Reanalyses; (b) CCM3; and (c) CSM. Units are meters.
Figure 4: Amplitude of quasi-stationary wave 1 at 500mb for (a) NCAR/NCEP Reanalyses; (b) CCM3; and (c) CSM. Units are meters.
Figure 5: Amplitude of quasi-stationary wave 2 for (a) NCAR/NCEP Reanalyses; (b) CCM3; and (c) CSM. Units are meters.
Figure 6: Amplitude of quasi-stationary wave 2 at 500mb for (a) NCAR/NCEP Reanalyses; (b) CCM3; and (c) CSM. Units are meters.
Figure 7: Amplitude of quasi-stationary wave 3 for (a) NCAR/NCEP Reanalyses; (b) CCM3 and (c) CSM. Units are meters.
Figure 8: Amplitude of quasi-stationary wave 3 at 500mb for (a) NCAR/NCEP Reanalyses; (b) CCM3; and (c) CSM. Units are meters.
Planetary Waves in the Southern Hemisphere and Linkages to the Tropics

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ABSTRACT

Decadal trends associated with the half yearly wave and the mean planetary waves in the Southern Hemisphere (SH) are examined using the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) Reanalysis from 1973 to the present. While the climatology of the waves observed by van Loon and Jenne (1972) is generally reproduced, there is large interannual variability. From the 1950’s to the present, the half yearly wave weakened with the largest changes in the Pacific Ocean. Pressure changes reported by van Loon et al. (1993) using the Australian analysis are also found in the NCEP Reanalysis. The largest impact is on wave number 3 at 50°–60°S. Wave number 3 intensified during the 1980s and continues to strengthen through the 1990s, though the seasonal variation continues to diminish.

Wave numbers 1 and 3 dominate at low frequencies. The first mode is the global, zonally symmetric mode associated with El Niño-Southern Oscillation (ENSO) and the tropical intraseasonal oscillation. The next two modes are the Pacific South American (PSA) modes with wave number 3 in midlatitudes. These two modes together represent intraseasonal oscillations in the SH, which interact episodically with the tropics.

1. INTRODUCTION

van Loon and Jenne (1972) and van Loon et al. (1973) studied the zonal harmonic standing waves in the SH using monthly mean maps from the South African Weather Bureau, especially for the International Geophysical Year. They stated that the dominant wave in the extratropics was wave number 1, with a peak in the subtropics near 30°S and a peak at higher latitudes near 55°S. They also found that wave number 2 was relatively weak but that wave number 3 had significant standing components. Wave number 3 is strongest during June–August (JJA) with a maximum at 50°S. The amplitude of wave number 3 during summer is about 30% that of wave number 1. Their results were later confirmed by Trenberth (1980), who revisited this topic using the Australian analyses. While the climatology of planetary waves is stable, there is large interannual variability. van Loon et al. (1993) noted that a major circulation regime change had occurred during the 1970s. While the sea level pressure (SLP) fell at high latitudes, it rose at low latitudes and wave number 3 intensified at both high and mid-latitudes. Wave number 3 is particularly important because of its association with persistent anomalies in the SH. (Trenberth and Mo, 1983; Mo and Higgins, 1996).
Wave numbers 1 and 3 also dominate low frequency variability. Rogers and van Loon (1982) examined the spatial variability of daily 500 hPa heights and SLP using eigenvector analysis. Their results were obtained using the Australian analyses from 1972–1979, but were verified against station data. The first EOF is largely zonal and indicates an out of phase relationship between anomalies at mid-latitudes and high latitudes. The second EOF has a wave number 3 structure. Later, Kidson (1988; 1991) reproduced both patterns using the ECMWF analysis from the 1980s and found that both patterns exist at both interannual and intraseasonal time scales.

Rogers and van Loon (1982) noticed that the phase reversal in EOF 1 between high and low latitudes is a signature of the ENSO composite (van Loon and Madden, 1981). Later, Lau et al. (1994) and Mo and Kousky (1994) used a rotated EOF analysis on 200 hPa streamfunction anomalies to relate the low frequency variability in the extratropics to the tropics. The zonally symmetric pattern appears as EOF 1 and it can be related to enhanced convection in the central Pacific. The next two patterns in the SH are two wave number 3 patterns in quadrature with each other with large amplitudes in the PSA sector. They are referred to as the Pacific South American patterns (Karoly, 1989) analogous to the Pacific North American pattern (Wallace and Gutzler, 1981) in the Northern Hemisphere. The PSA 1 mode occurs frequently during warm ENSO events, so it is primarily forced by tropical convection (Karoly, 1989). On the intraseasonal time scale, both patterns are associated with persistent anomalies in the SH (Mo, 1986; Mo and Higgins, 1996). These two PSA patterns represent intraseasonal oscillations in the SH but no significant correlations between the time series associated with the PSA patterns and outgoing long wave radiation (OLR) anomalies in the tropics can be found.

Recently, NCEP/NCAR have completed a global reanalysis for the period from 1 January 1973 to 31 August 1996. Documentation of the model and the assimilation system used for reanalysis can be found in Kalnay et al. (1996). Although there are changes in the input data base, the reanalysis data are produced with a fixed assimilation system, they are homogeneous. They are better suited for examining decadal changes than operational analyses. In this paper, we revisit the planetary wave regimes in the SH and their linkages to the tropics.

2. DATA

The data used in this study are daily mean global gridded data from the NCEP/NCAR Reanalysis for the period 1973 to the present. Data are on a 2.5° × 2.5° latitude-longitude grid. Daily averages of the NOAA satellite OLR from 1979 to the present are used to represent tropical convection (Leibmann and Smith, 1996). The seasonal cycle at each grid point is defined as the grand mean and the first and second harmonics with periods of 12 and 6 months, respectively. Anomalies are defined as the difference between the full field and the seasonal cycle.
3. DECADAL VARIATION OF PLANETARY WAVES

Mo and van Loon (1984) examined the seasonal changes associated with the half yearly wave, and the mean planetary waves in the monthly mean sea level pressure (SLP) using two data sets. One was obtained from the South African Weather Bureau (Taljaad and van Loon, 1964) for the period 1951 to 1958 and the other was obtained from the Australian Bureau of Meteorology for the period 1972 to 1980. They found large differences between the two data sets over the southern oceans with the largest changes in the Pacific Ocean. These changes have an impact on planetary waves, especially at wave number 3. We updated these changes using the NCEP/NCAR Reanalysis in Figs. 1 and 2 and also reproduced the results from the African analysis from Mo and van Loon (1984) for comparison.

The half yearly wave can be represented by the SLP difference between March and June or the difference between June and September. In each case, the SLP rises in the polar region while it drops in midlatitudes over the three oceans from March to June (Fig. 1). The situation reverses from June to September (Fig. 2). While the basic pattern does not change, the amplitudes and locations of the maxima and minima vary significantly. The largest differences can be found over the Pacific Ocean. The pressure difference in the Pacific from March to June changed from 11.8 mb in the 1950s (Fig. 1a) to 4.5 mb in the first half of the 1990s (Fig. 1d). The pressure rise from June to September also changed by about the same amount from the 1950s to the 1990s. Over the Atlantic and Indian Oceans, the largest differences can be found between the 1950s and the 1970s. The pressure difference between June and March from 1951–1958 was positive in the Atlantic (Fig. 1a), but the pressure difference in the same region was negative for the other three decades (Fig. 1c–1d). The African analysis shows a clear maximum of about 4.5 mb in the Indian Ocean from September to June. This maximum decreased in the later years and there was no longer a well defined maximum in the 1990s. These changes in the half yearly waves between the 1970s and the 1980s were also observed in the Australian analyses as well as in station data (van Loon et al., 1993). Overall, there is good agreement between results from the NCEP Reanalysis and the Australian analyses.

Decadal changes also occur in the planetary waves. Figure 3 shows the annual march of the amplitude of the mean harmonic waves for the 1970s, the 1980s and the first half of the 1990s. A comparison with the planetary waves in the 1950s (Mo and van Loon, 1984) shows that the largest changes are in wave number 3. The amplitude of wave number 1 shows two maxima. The one in the subtropics appears in winter and has an amplitude of about 4 mb. The one at higher latitudes increased slightly from 6 mb in the 1950s to 8 mb in the 1990s. The amplitude of wave number 2 has only one maximum at 55°S which appears in late winter or early spring and has a magnitude of about 4 mb. The wave number 3 maximum stays at 50°S, but magnitudes continue to increase. The wave number 3 in the 1950s showed a well defined seasonal cycle with a maximum of about 3 mb in July. Wave number 3 strengthened somewhat during the 1970s but the marked seasonal variation was still visible. While the wave number 3 amplitudes continued to increase in mid-latitudes in the 1980s and the 1990s, the seasonal variation decreased. There was very little seasonal variation in the first half of the 1990s.
annual cycle consists of the 12 months and 6 months harmonics. Fig. 5a plots the annual mean of the wave number 3 amplitude and the amplitude of the annual cycle is given in Fig. 5b. The wave 3 amplitude was weak for the period from 1973 to 1979 (Fig. 4a). After 1980, the maximum stayed at about 50 m between 50°–55°S while the amplitude of the annual cycle decreased. Thus, the changes in the half yearly wave and planetary waves demonstrated by van Loon et al. (1993) using the Australian analyses are also found in the NCEP/NCAR Reanalysis and these changes continue to this day.

4. LOW FREQUENCY MODES

Figures 5a–5c show the first three EOFs of fluctuations in 500 hPa height anomalies over the SH with periods longer than 10 days using 17 years (1979–1995) of the NCEP/NCAR Reanalysis. The height anomalies are not normalized but a latitudinal cosine weighting factor was used to compute the covariance matrix. Because of weak seasonal variation of wave number 3 during this period, all seasons are pooled. EOFs 1 and 3 are similar to the first two EOFs obtained by Rogers and van Loon (1982). These are also the first three EOFs reported by Kidson (1988; 1991) using ECMWF 500 hPa height anomalies.

The leading EOF (Fig.5a), which explains 15% of the total variance, shows strong zonal symmetry and an out of phase relationship between high latitudes and subtropics. Apart from zonal symmetry, a zonal wave number 3 is also evident with three centers of action located in the three oceans. Rogers and van Loon (1984) noticed that this mode implies the weakening of westerlies equatorward of 35°–40°S and the strengthening of westerlies poleward of 35°–40°S and an increase in tropical easterlies. They also pointed out that this is the signature of ENSO. Figure 5d shows the first EOF obtained using the 200 hPa streamfunction anomalies for JJA. The winter season was selected to avoid dominance of the NH waves. The correlation between the two monthly mean time series of PCs associated with Figs. 5a and 5d is 0.95 , which is significant. This indicates that Fig. 5a is the SH part of a global pattern which shows strong zonal symmetry in both hemispheres with a dipole straddling the equator in the central Pacific. For positive PC 1, it indicates strong easterlies in the tropics and enhanced westerlies in mid-latitudes and a weakening of westerlies at polar latitudes.

The next two EOF patterns (Figs. 5b and 5c) explain roughly the same amount of the total variance (8.83% and 8.39%). They depict wave number 3 patterns in quadrature with each other with large amplitudes in the PSA sector. The degeneracy and the quadrature relationship suggest that they represent oscillation in the SH. In this paper, we label them PSA 1 (EOF 3) and PSA 2 (EOF 2) modes. They are the first two rotated EOFs in the 200 hPa streamfunction anomalies (Mo and Higgins, 1996) after removing the zonal mean (Fig. 5e and 5f). Weak loadings in the Northern Hemisphere suggest that the two PSA modes are regional modes.

The two PSA modes represent the intraseasonal oscillation and their evolution determined from the lag and lead correlations, reveals this path:
We present the 200 hPa zonal wind composites for positive and negative large PSA days in Fig. 6. These are days when the amplitudes of the PC associated with any given PSA mode are larger than 1.5 standard deviations computed from the daily time series for the period from 1979 to 1995. The annual mean 200 hPa zonal wind averaged over 1979 to 1995 shows that in the SH the subtropical jet is located at 30°S across Australia with the jet core at 160°E and with a second jet located in the Indian Ocean (not shown). The positive PSA 1 composite (Fig. 6a) shows that the subtropical jet starts to extend eastward into the South Pacific with the jet core located east of the dateline and the South American jet strengthens. Westerlies diminish near the location of large negative streamfunction anomalies near 120°W, 60°S (Fig. 5e) similar to a blocking situation. The positive PSA 2 case shows further extension of the subtropical jet. The negative phase of PSA 1 shows that the SH subtropical jet retracts into the western Pacific while the jet in the Indian ocean strengthens. The negative PSA 2 composite shows that the jet core is located north of New Zealand and the South American jet weakens.

5. TROPICAL CONNECTION

a. Zonally Symmetric Pattern

We correlated the monthly mean PC time series associated with streamfunction EOFs (Figs. 5d–5f) with OLR anomalies (OLRA). Again, we pooled all seasons. The correlation should be larger than 0.23 to be significant at the 95% level. For positive PC 1, there are negative OLRA (enhanced convection) in the central Pacific and positive OLRA (less than normal convection) in the western Pacific which is a typical ENSO signal. Lau et al. (1994) and Mo and Kousky (1984) show that this pattern also exists in the intraseasonal band and that it can be excited during strong tropical intraseasonal oscillations.

b. PSA Modes

There is no statistically significant correlation pattern between OLRA in the tropics and monthly mean or 10 day mean time series of PCS associated with PSA 1 or PSA 2. However, Mo and Higgins (1996) suggested that the PSA modes can interact with the tropics episodically. A composite analysis was performed to study the tropical linkages of the PSA modes. Lagged composites for 200 hPa winds, velocity potential and OLRA for large positive and negative PSA 1 and PSA 2 days were produced for the pentads from 20 days before onset to 20 days after onset. The onset of each event is defined as the time when PC 1 (PC 2) first crosses a threshold magnitude which is defined as 1.5 standard deviations. There is no requirement for persistence. We interpret this as a moving average of the canonical flow pattern for large PCS. There are on average about 415 maps in each composite. Composites were also computed with a requirement for persistence of 5 days or more; the major conclusions do not change. Generally, the anomaly patterns for positive and negative events are similar but with a sign reversal. We present the composite difference between positive and negative events.
Figure 8 shows the composite difference between positive and negative large PSA days averaged for day 0 to 8. The positive PSA 1 mode is associated with enhanced convection in the central Pacific and suppressed convection in the western Pacific. The OLRA associated with PSA 1 and PSA 2 are in quadrature with each other. The OLRA just south of the equator show a familiar dipole which is the signature of the tropical intraseasonal oscillation (Lau and Chen, 1985; Knutson and Weickmann, 1987). In the central Pacific, the OLRA associated with REOF 1 have strong stationary components. The response of the divergent flow indicated by the velocity potential anomalies is consistent with the OLRA.

The evolution of the tropical heating anomalies is given in Fig. 9. It shows eastward propagation of the OLRA dipole in the tropics. The shift of convection is accompanied by the dynamical signal, depicted by 200 hPa meridional wind anomalies. Two pentads before the onset of large PSA 1 days, enhanced tropical convection is already established in the central Pacific. The enhanced tropical convection is accompanied by an enhanced local Hadley circulation as indicated by positive OLRA in the subtropics near 25°S and 25°N in both hemispheres with a stronger center in the SH. In the central Pacific, a meridional wind dipole appears at 20°–25°S near the dateline south of the tropical heating (Fig. 9a). Meanwhile, a wave number 3 structure establishes itself in midlatitudes. A pentad later, the convergence at 25°S deepens as the local Hadley circulation continues to strengthen. The South Pacific Convergence Zone (SPCZ) strengthens. The dipole in the subtropics and wave number 3 in midlatitudes propagate eastward together. After onset, the anomalies diminish upstream, but anomalies downstream amplify. As the wavetrain like pattern reaches South America, the South Atlantic Convergence Zone (SACZ) strengthens (Fig. 9c). Similar tropical extratropical relationships hold for PSA 2. The subtropical meridional wind dipole forms one pentad before onset when convection in the central Pacific approaches the dateline (Figs. 9e). There is a phase reversal between anomalies in the subtropics and in midlatitudes. Wind anomalies associated with PSA 2 are in quadrature with those associated with PSA 1. Similar to the PSA 1 case, the subtropical dipole and wave number 3 propagate eastward together.

Blade and Hartman (1995) used a simple model to examine the dynamical response in the extratropics to tropical heating. They imposed eastward propagating dipole heating similar to the OLRA during the tropical intraseasonal oscillation. The period of the tropical intraseasonal oscillation is 40 days. When the nonlinear interaction is allowed, a dipole appears in the area of tropical heating with a wavetrain downstream at day 5. As the heating propagates eastward, the wavetrain in the extratropics and the dipole in the subtropics propagate together with the heating dipole. These responses are very similar to the meridional wind composites based on REOFs shown in Fig. 9. The principal difference is that for their case, the responses in the two hemispheres are symmetric. Since composites here are keyed to the PCS associated with the PSA modes which represent circulations in the SH extratropics, the responses are stronger in the SH.
6. CONCLUSIONS

The zonal harmonic waves in the SH were revisited using the NCEP Reanalysis data from 1973 to the present. The NCEP Reanalysis was produced with a fixed assimilation system so the reanalysis data are well suited for studies of decadal trends. While the climatology of waves observed by van Loon and Jenne (1972) is generally reproduced, there is large interannual variability. From the 1950s to the present, the half yearly wave weakened with the largest changes in the Pacific Ocean. The pressure changes reported by van Loon et al. (1993) using the Australian analyses are also found in the NCEP Reanalysis. The largest impact is on wave number 3 at 50°-60°S. Wave number 3 intensified during the 1980s and continued to strengthen during the 1990s, while the seasonal variation continued to diminish.

These changes also appear in the low frequency modes. The EOF patterns observed by Rogers and van Loon (1984) are well reproduced using the NCEP Reanalysis from 1979 to the present. The first mode is the zonally symmetric pattern which is part of a global mode. Correlation with OLRA indicates that the positive phase of this mode is associated with enhanced convection in the central Pacific and suppressed convection in the western Pacific. Therefore, this mode can be excited during ENSO as suggested by van Loon and Rogers (1984). On the intraseasonal time scales, this mode is also related to the tropical intraseasonal oscillation (Mo and Kousky, 1994; Lau et al., 1994.)

The next two modes represent the intraseasonal oscillation in the SH. They are SH modes and both have wave 3 in midlatitudes with large amplitudes in the PSA sector. They interact with tropical convection episodically (Mo and Higgins, 1996).

References


Figure 1. Average sea level pressure difference (June-March) from (a) 1951-1958 reproduced from Mo and van Loon (1984), (b) 1973-1980, (c) 1981-1989 and (d) 1990-1996 to the present from the NCEP Reanalysis. Contour interval is 2 mb.
Figure 2. Same as Fig.1, but for the average changes from June to September.
Figure 4. The (a) annual mean, and (b) the seasonal cycle of the monthly mean wave number 3 amplitude for 500 hPa heights. Contour intervals are (a) 20 m and (b) 8 m.
Figure 5. (a) EOF 1, (b) EOF 2 and (c) EOF 3 for 10 day mean 500 hPa height anomalies from the NCEP Reanalysis for 1979 to 1995. Negative values are shaded. (d) Same as (a) but for 200 hPa streamfunction, (e) the PSA 1 mode which is the first EOF from the 10 day mean zonally asymmetric 200 hPa streamfunction anomalies, and (f) same as (e) but for the PSA 2 mode which is the second EOF.
Figure 6. 200 hPa zonal wind composite averaged over all large (a) positive PSA 1 days, (b) positive PSA 2, (c) negative PSA 2 and (d) negative PSA 1 days. Contour interval is 8 m s$^{-1}$. 
Figure 7. Correlation between the PC monthly mean time series associated with the zonally symmetric mode (Fig. 5d) and OLRA from 1979 to the present. Contour interval is 0.1. 0.25 and -0.25 contours are added.
Figure 8. OLRA (shaded) and velocity potential anomaly (contour) composite between positive and negative large (a) PSA 1 days and (b) PSA 2 days. The unit for OLRA is $5 \text{ W m}^{-2}$. Contour interval for the velocity potential is $1 \times 10^6 \text{ m}^2 \text{ s}^{-1}$.
Figure 9. Map sequences of 200 hPa meridional wind and OLRA represented as the composite difference between positive and negative (a) PSA 1 days for the pentad centered at day -8, (b) PSA 1 days centered at day -3, (c) PSA 1 days centered at day 2, (d) PSA 2 days centered at day -6, (e) PSA 2 days centered at day -3 and (f) PSA 2 days centered at day 2. Meridional wind anomalies are contoured every 3 m s$^{-1}$. Zero contours are omitted. OLRA are shaded.
Quasi-Stationary Waves in the Southern Hemisphere:
Revisiting a Theory due to Harry van Loon

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1. INTRODUCTION

We are revisiting a theory for the positions and maintenance of the quasi-stationary waves (QSW) of the Southern Hemisphere (SH). This note will describe our work in progress to diagnose the QSW signature in a new surface wind data set, and to begin to understand the dynamics of the QSW as well as other large-scale features of the surface winds of the SH. We briefly review the QSW theory due to van Loon and co-workers (van Loon and Jenne, 1972; van Loon et al., 1973). Surface wind observations derived from satellite scatterometer measurements are introduced, and we compare the long-wave patterns in the scatterometer winds with the QSW theory. Finally, we examine surface pressure fields from the recent National Centers for Environmental Prediction (NCEP) Reanalysis data set (Kalnay et al., 1996) and contrast the geostrophic winds they imply with the scatterometer winds. In analyzing the modern surface wind data, we will focus on meridional and seasonal averages in the latitude band 55°S to 60°S, for January through March (JFM) of the years 1992–1995.

The QSW of the geopotential height fields of the SH were described by van Loon and Jenne (1972) using long-term average geopotential height fields due to Taljaard et al., 1969. In 1973, van Loon and Jenne joined with Labitzke to expand their study to contrast the zonal harmonic QSW of the Northern Hemisphere (NH) and SH (van Loon et al., 1973). Taken together, these papers demonstrate that the QSW of the SH explain more than 95% of the total variance in the 500 hPa geopotential height field; the lion’s share (>90%) due to zonal-wave 1 (ZW1), and then zonal-wave 3 (ZW3). In Figure 2 of van Loon and Jenne (1972; not shown here), the positions of QSW ridges in 500 hPa and 200 hPa geopotential height fields were identified. The ridge associated with ZW1 at 500 hPa occurs in the Pacific Ocean around 120°W for the latitude range of interest in this note. The weaker ZW3 contribution is associated with ridges in 500 hPa geopotential height fields aligned with Africa, Australia, and South America (i.e. equatorward from the circumpolar region toward the nearest continental land masses). Zonal-wave 2 is confined to the Antarctic continent and is not a major contributor to the variance explained by the QSW in the troposphere of the circumpolar region above the ocean.

The QSW are quasi-barotropic throughout the troposphere. The maximum amplitudes, and largest percent variance explained for ZW1 and ZW3 occur in the mid to upper troposphere, but the signal at the surface is still appreciable. ZW1 explains more than 50% of the variance in 1000 hPa geopotential height in a long-term average for JFM (Dennis Shea, personal communication, 1996).

van Loon and Jenne (1972) credit Anderssen (1965) with the identification of the QSW in the SH.
van Loon and his co-workers put forward the following theory for the maintenance of the QSW in their papers. Consider the schematic diagram of sea-level pressure (SLP) distribution in the SH shown in Fig. 1. The equator is off the top of the page and the South Pole is off the bottom of the page. The surface is the circumpolar ocean. A ZW1, trough and ridge system is depicted here by a pressure isoline, or "geostrophic streamline" that marks the wind direction near the surface. From Ekman theory we know that the surface ocean will be advected at an angle of about $45^\circ$ to the left of the surface wind direction in the SH. This advection creates net surface divergence under the atmospheric low pressure lobe and net surface convergence under the high pressure center. There is perhaps, a secondary SST affect due to the net equatorward advection of colder water in the region of the trough. Thus the oceanic impacts upon the SLP distribution serve to reinforce the ZW1 structure. The coincidence of low pressure with lower temperature, and high pressure with higher temperature is the reason for the barotropic structure of the QSW. We have begun to examine this interplay between SLP and surface winds in the QSW using better data sets than were available to van Loon and co-workers in the early 1970's. This note introduces applications of scatterometer winds and SLP from the NCEP Reanalysis.

2. WHAT IS SCATTEROMETRY?

Wind shear at the air-sea interface creates capillary waves or "cat's paws" on the ocean surface. A scatterometer instrument measures the cm-wavelength capillary wave field by bouncing a radar pulse off the ocean surface and detecting the backscatter signal. The capillary wave field in the cm-wavelength range has been related quantitatively to surface shear stress on the ocean, and then by a drag law, to the surface wind at 10 m height. The process whereby a radar backscatter signal is related to a 10 m wind is an active area of current research. Present-day methods can produce surface wind speeds and directions accurate to about a meter per second, and to within a few degrees of direction.

A space-borne scatterometer has been operating aboard European Space Agency satellites, ERS-1 and ERS-2 since July 1991. We will examine surface winds in the SH from the ERS-1 data set for JFM of 1992 through 1995. Our work is in preparation for data from a NASA scatterometer (NSCAT) that was launched in August 1996 from Japan, and is expected to be in operational use by early 1997. Both NSCAT and ERS-1 are polar-orbiting platforms; each complete orbit lasting about 100 minutes. The surface winds from ERS-1 are reported in a 5° wide swath that is scanned, normal to the satellite ground track, off one side of the orbiting platform (NSCAT is a 2-sided instrument, sweeping two 5° swaths off either side of the platform, with a 3° nadir gap between). The wind vector cell resolution within the swath is 50 km$^2$. The sampling pattern precesses westward such that more than 90% of the surface wind field can be measured in about 35 days (2 days for NSCAT).

Figure 2a depicts the ERS-1 scatterometer coverage of the SH for the 24-hour period of 1 January 1993. The domain spans the globe from 45°S to 65°S (ERS-1 winds are reported to 60°S). South America and the Antarctic Peninsula bound the Drake Passage; the South Island of New Zealand is the other substantial land mass in the domain. A gray-scale value (according to speed, scale not
shown) is assigned to each wind vector cell location from which at least one wind speed has been returned during the sampling period. Note that there are missing data within swaths, and missing swaths in some locations. Figure 2b depicts the 3-day coverage for the domain, starting with 1 January 1993. Up to the limit imposed by missing data, we are approaching synoptic coverage. The NSCAT sampling scheme will more than double this coverage given its two-sided configuration. By 17 days (Fig. 2c), every location in the domain has been sampled by the ERS-1 scatterometer at least once. However, the coverage is a function of the orbit configuration. For most of 1993, ERS-1 was in a so-called 35-day repeat orbit (Figs. 2a-c). In 1994, this orbit was changed to a 3-day repeat configuration. Figure 2d is the 17-day coverage starting from 3 February 1994, after the ERS-1 orbit had been altered. The unsampled diamond patterns are never covered in this orbital configuration.

Obviously, the scatterometer data represent several orders of magnitude more information about the surface winds in the SH than was previously available. The scatterometer winds can give good estimates of synoptic, and certainly monthly variability. As with any new data set, there are biases and sources of error to be considered.


The 90-day average meridional winds for JFM'93 are depicted in Fig. 3. The ERS-1 winds have been collected in 1° longitude by 5° latitude bins (55°S–60°S), and the 90-day average meridional component for each bin is drawn as a single vector. The vector length is scaled such that 1° latitude represents 1 ms\(^{-1}\). The bottom panel measures the number of ERS-1 wind observations per bin. As we have seen from the coverage maps (Fig. 2), the bin-weights are nearly a uniform function of longitude with around 2000 observations per bin for JFM'93.

The ZW1 and ZW3 structures are apparent in the average meridional winds. Moreover, there is the signature of a large-scale surface high-pressure system in the Pacific sector. We were a bit concerned at first to note a bias toward northerly winds in these seasonal averages. However, the concern abated when we rediscovered a paper by Vowinckel and van Loon (1958), based on whaling ship reports, that demonstrates a preference for northerly components in the meridional part of the total wind vector in the latitude band of interest. We have yet to determine where in the SH troposphere the implied local mass imbalance is made up.

Figure 4 presents a power spectrum (Fourier coefficients squared vs. wavenumber) for the meridional winds in Fig. 3. We have segmented the ERS-1 winds into five non-overlapping sequences of 17 days each (2 day separation between sequences). Power spectra have been computed for each segment, and averaged by wavenumber bin. This yields about 10 degrees of freedom for each amplitude-wavenumber pair, given the optimistic assumption that each 17 d segment is independent. No windowing has been performed to control spectral leakage. The peaks at ZW1 and ZW3 are significant and dominant. The largest peak at ZW3 merges with a peak at ZW4, possibly due to leakage.
To begin to develop a sense of the interannual variability in the QSW, and to begin to look at the
effects of orbit configuration, Fig. 5 shows the 90-day average meridional winds for JFM'94. The bin-
weight profile demonstrates the effect of the unsampled diamond-shaped regions characteristic of the 3-
day repeat orbit (Fig. 2d). In this year and season, the meridional components of the surface winds are
even more predominantly northerly. The large-scale signature of the QSW is interpretable as additions
to, and subtractions from, the amplitudes of these northerly components. The power spectrum for the
JFM'94 meridional winds (Fig. 6a) again picks out ZW1 and ZW3. A power spectrum for the bin weight
profile (Fig. 6b) shows clearly the wavenumber band of the sampling artifact; far from ZW1 or ZW3.

We will depict the meridional average winds for JFM'92 and JFM'95 after introducing the (SLP)
data to set up a comparison with geostrophic wind components for the circumpolar region.

4. NCEP REANALYSIS SLP 92–95

Figure 7 is the 4-year average global SLP field from the NCEP reanalysis product available 4
times daily on a regular 2.5° grid. The ZW1 and ZW3 structures in the SLP average for the latitude
band 55°S to 60°S are evident. We have used these data to compute the meridional component of the
geostrophic velocity \( v_{\text{geos}} \) in each of the summer periods JFM'92–JFM'95, from the zonal gradients of
the averaged SLP. The pressure gradients were computed by an overlapping central difference scheme
that employs a 5° span between SLP data points.

Figure 8 depicts a superposition of the four \( v_{\text{geos}} \) profiles as well as the 4-year average for JFM.
Again we can see interannual variability and the large-scale QSW signature that is a combination of
ZW1 and ZW3. The meridional components of the surface geostrophic winds also show a northerly bias.
There is visual evidence for interannual variability on length scales of 20°–60° that is as yet unexplained.
The interannual variability on shorter spatial scales is largely cancelled in the mean profile (Fig. 8).

In Fig. 9a-d, the longitudinal profiles of the average meridional component of the surface
geostrophic velocities have been rescaled to compare with ERS-1 meridional velocity vectors averaged
over the same 90 days (JFM) for each year (1992–5), using 5° latitude by 2.5° longitude bins.
Differences between vector amplitudes and the amplitude of the longitudinal profiles in Fig. 9 are too
large to be explained by errors in the ERS-1 data set and/or errors in the NCEP SLP data set. More
likely, the amplitude differences represent departures from geostrophy in the time and space averaged
surface winds. Such departures can be due to surface boundary-layer effects not accounted for in the
genostrophic model, and/or a mean meridional flow (northerly) for the period, and in the region, over
which averages have been computed. We will explore each of these possibilities as our research
program evolves.
5. SUMMARY

The surface positions of the QSW of the SH between 55°S and 60°S have been determined from: a) seasonal (JFM) and spatial averages of the meridional wind vectors measured by the ERS-1 scatterometer; and b) the meridional components of surface geostrophic winds implied by similar averages of SLP from the NCEP reanalyses. The ZW1 and ZW3 structures proposed by van Loon and co-workers have been confirmed. In addition our cursory examination of these data sets demonstrates evidence of: variability on spatial scales shorter than ZW3; interannual variability; and significant departures from geostrophy.

Obviously, we have presented a work-in-progress. We are eager to examine, in greater detail than has previously been possible, the variability on several timescales of the QSW at the surface in the circumpolar region of the SH. These preliminary analyses suggest two truths that we may have already known: 1) early workers were correct in their description of the QSW; and 2) increases in resolution, be they temporal, spatial, or in radiometry, always reveal new structure and phenomena in the geophysical fluid processes under scrutiny.

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References


In-Phase T and P in SH Quasi-Stationary Waves
(van Loon et al., 1973)

EQUATOR

TROUGH

RIDGE

Ekman flux
surface ocean

Poleward advection
(warming)

Pressure isoline/
"geostrophic streamline"

Westerly Belt
(Mid- to Sub-Antarctic
Latitudes)

Equatorward advection
(cooling)

Low
net divergence
\(\Rightarrow\) colder surface

High
net convergence
\(\Rightarrow\) warmer surface

"ATLANTIC/WEST INDIAN"

"PACIFIC"

SOUTH POLE
Figure 2. ERS-1 coverage of the SH circumpolar region. a) 1-day coverage for 1, January 1993; b) 3-day coverage beginning 1, January 1993; and, c) 17-day coverage beginning 1, January 1993. Panels a-c are representative of the 35-day repeat orbit configuration. Panel d) is the 17-day coverage beginning 1, February 1994 during a 3-day repeat orbital configuration phase of the ERS-1 mission.
Figure 3. Average JFM 1993 surface wind meridional component from ERS-1. Scatterometer winds have been averaged over 90 days beginning 1 January 1993, in 1.00000 degrees longitude by 5.00000 degrees latitude bins. Wind vectors (top panel) represent the average meridional component for each bin, where 1.0 of latitude represents 1 m/s. The total number of ERS-1 observations per bin is depicted in the bottom panel.
Figure 4. Power spectrum for meridional wind profile from Figure 3. Amplitude ($a^2 = b^2$) vs. wavenumber (deg$^{-1}$) from an average of 5 spectra; one for each non-overlapping 17-day coverage segment during the period JFM 1993.
Figure 5. As in Figure 3, but for the period JFM 1994 during the 3-day repeat orbit phase.
Figure 6. Power spectra for: a) average surface meridional winds from ERS-1 as in Fig. 4 for the period JFM 1994; and b) power spectra for bin-weight totals (see bottom panel Fig. 5) for the period JFM 1994.
**JFM Average $P_{\text{m}}$ (hPa) from NCEP Re-Analysis 1992–95**

Figure 7. Average global SLP from NCEP Reanalysis. SLP averaged over JFM for 1992–5 from the 4 times daily, 2.5° product of the NCEP reanalysis.

**Average Meridional Geostrophic Velocity JFM (55S–60S) NCEP 1992–95**

Figure 8. Superposition of longitudinal profiles of $v_{\text{geo}}$ for the periods JFM 1992–5 (one profile for each year). $v_{\text{geo}}$ is computed from overlapping 5° centered differences of SLP. The 4-year mean JFM profile is depicted by the dashed line.
Figure 9. Comparisons of $v_{geo}$ and the average surface meridional wind components from ERS-1 for a) JFM 1992; b) JFM 1993; c) JFM 1994; and d) JFM 1995.
A Modulation of the Mechanism of the Semiannual Oscillation in the Southern Hemisphere

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ABSTRACT

The twice-annual contraction and intensification of the circumpolar trough of low pressure around Antarctica, termed the semiannual oscillation (SAO), was the dominant signal in the annual cycle at mid- and high southern latitudes prior to 1979. The mechanism, as postulated by van Loon (1967), arises from the polar continent surrounded by ocean and the different seasonal evolution of surface temperatures associated with the characteristics of the surface between roughly the latitudes of 50°S and 65°S. It has subsequently been shown that since 1979 the SAO has weakened considerably.

Circumstantial evidence is presented here to suggest that the basic mechanism as originally proposed by van Loon is still valid, but that secular warming since 1979 has not been evenly distributed through the seasonal cycle at each latitude. Thus an anomalous change in the temperature gradient between 50°S and 65°S, with peaks in roughly May and November, has interfered with or modulated the mechanism that produces the SAO, with its peaks in March and September. Consequently, the magnitude of the SAO has decreased in the more recent period.

1. INTRODUCTION

The 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 (van Loon, 1967; 1972). 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. Prior to 1979, the SAO explained 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).

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. 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; Meehl, 1991).

After 1979 the amplitude of the SAO decreased dramatically (van Loon et al., 1993; Hurrell and van Loon, 1994). We will first describe the behavior of the SAO prior to 1979 and the mechanism proposed by van Loon (1967) to explain it. Then we will present circumstantial evidence from limited observational data to suggest that a modulation of the original mechanism has likely been responsible for the reduction in magnitude of the SAO in the more recent period.

2. THE SEMIANNUAL OSCILLATION IN THE SOUTHERN HEMISPHERE

The SAO is evident in long-term monthly mean maps of observed sea level pressure (SLP, Meehl, 1991). The trough of minimum SLP is farthest south and deepest in March and September and farthest north and weakest in June and December. Associated with the seasonal intensity and latitudinal movement of this band of minimum SLP are large regions between about 25°S and 60°S where the 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.

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°S and 65°S associated with the SAO in SLP. Thus, a useful index of the SAO, first used by van Loon (1967), is the difference of the zonal mean 500-mb temperature between 50°S 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 (1967), 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. 1, 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)]. In contrast, stations at comparable latitudes in the NH show sharper midwinter temperature minima.

3. OBSERVATIONS AND MODEL SIMULATIONS OF THE SAO

As noted above, a useful index of the SAO is the 500-mb temperature difference, 50°S minus 65°S. Figure 2 from Meehl (1991) shows this index for observations and two model simulations. In the observations (Fig. 2a, data from the Australian analyses 1972–84, with 8 years of strong SAO, 1972–79, and 5 years of weak SAO, 1980–84) the second harmonic is still dominant (59% of the variance versus 19% for the first harmonic), with maxima, indicating an intensification of the gradient, occurring in the transitional seasons of March and September, the times of year when the circumpolar trough is deepest.

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.

The two model simulations from Meehl (1991, see that paper for model details) shown in Fig. 2 are similar in that both have a weaker-than-observed zonal mean 500-mb temperature gradient between 50°S and 65°S. Yet, the SPEC SST simulation (an atmospheric GCM run with observed SSTs compiled prior to 1979, Fig. 2b) shows a dominant, though small, second harmonic (46% explained variance compared to 38% for the first harmonic). The MIX1 simulation in Fig. 2c (the same atmospheric GCM coupled to a non-dynamic 50 meter deep mixed layer or "slab" ocean) shows a more dominant first harmonic compared to the second harmonic (30% for the second harmonic versus 60% for the first harmonic), with the maxima of the second harmonic occurring about half a month later compared to the SPEC SST case.

Figure 3 from Meehl (1991) 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. 3b), the SPEC SST again follows reasonably well the observed annual curve of 500-mb temperature. In contrast to 50°S, the SPEC SST simulation at 65°S is within 1°C to 2°C of the observed values throughout the year. The MIX1 simulation has larger quantitative errors from January to July (about 3°C to 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 mid-latitudes 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. 3 show that the shapes of the curves of the annual cycle are mostly correct at 50°S and 65°S. The main contributor to the reduced amplitude of the gradient associated with the weaker SAO shown in Fig. 2b is the colder-than-observed 500 mb temperature at 50°S in Fig. 3a. 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.

Meehl (1991) shows that the seasonal cycle at 500 mb is similar to that at the surface. The SPEC SST model reproduces the phase of the SAO but underestimates the amplitude. 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.

Meehl (1991) goes on to point out that 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 simulate 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 case mainly because of the different simulations of SSTs and ocean heat storage at 50°S. This implies that an alteration of the seasonal cycle of SSTs at 50°S in the real system could alter the manifestation of the SAO.

Thus, the main conclusions from the Meehl (1991) study are:

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 midtroposphere. Ocean heat storage and surface energy balance combine to produce a rapid decrease of SSTs in the southern autumn and a slow increase to the annual maximum in March. This is true in the observations and in the SPEC SST case at the surface and the midtroposphere. 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 at 50°S.

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 long wave radiation while the sun is low and below the horizon), coupled with the amplification of the trough and enhanced meridional flow in winter, would produced the coreless winter. Transient eddy heat flux from observations and model experiments indicates a large winter heat flux convergence south of about 50°S. 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) Of most relevance for the observed changes in the SAO since 1979, it was postulated that, since a critical aspect 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 manifestation of the SAO.

4. MODULATION OF THE MECHANISM OF THE SAO SINCE 1979

As noted earlier, a change in the SAO was noted to have occurred around 1979 (van Loon et al., 1993; Hurrell and van Loon, 1994). This change in the seasonal cycle at mid and high southern latitudes is illustrated in Fig. 4 from Hurrell and van Loon (1994). Zonal mean SLP at 50°S and 70°S are shown for the 1970s and the 1980s at top, with the difference, 50°S minus 70°S, shown at the bottom for the two periods. The difference of SLP at the two latitudes is a measure of the intensity of cyclonic activity at those latitudes, with large positive differences indicative of strong cyclonic activity. For the 1970s (the solid line at the bottom of Fig. 4), the two well-documented peaks of activity in the circumpolar trough in March and September are clearly seen, with large positive values of the SLP difference at the two times of year when the trough contracts and intensifies as noted by van Loon (1967). However, for the 1980s (the dashed line), there is a pronounced change in the seasonal cycle with greater values of the difference in May-June-July, decreased values in August-September-October, and greater values again in November-December. If this index is taken as indicative of baroclinicity between these two latitudes (and thus indicative of associated cyclonic activity), these results suggest that there was anomalous increased baroclinicity near the middle and end of the year in the 1980s compared to the 1970s, flattening the annual cycle of cyclonic activity and consequently decreasing the amplitude of the SAO in the later period compared to the earlier one.

Hurrell and van Loon (1994) showed an association between the magnitude of the second harmonic of SLP at Chatham Island at 44°S and SST anomalies in the tropical Pacific with warm SST anomalies associated with a decreased magnitude of the SAO at Chatham at decadal timescales. They suggested a possible modulation of the SAO from tropical convective heating anomalies implied by the
tropical Pacific SST anomalies. However, as shown by Bottomly et al. (1990), the low frequency variability of SST anomalies in the tropical Pacific occurs with very similar character nearly globally. Thus there is the possibility that the correspondence of tropical Pacific SST anomalies with the SAO amplitude could be coincidental, with SST variations at other latitudes or locations having a similar low frequency signature that could affect the SAO.

This brings us back to the original mechanism for the SAO proposed by van Loon (1967), and the result from analysis of model experiments by Meehl (1991). That is, a change in the seasonal cycle of SSTs at the two key latitudes for the mechanism, the ocean-dominated latitude of roughly 50°S and the polar continental latitude near 65°S, could result in a modulation of the mechanism that produces the SAO. Thus, if there was an annual mean warming of SSTs near 50°S for the 1980s compared to the previous period, and if this annual mean warming was not evenly distributed throughout the year (as suggested by Thomson, 1995, and Mann and Park, 1996), there could be a modulation of the van Loon mechanism that could reduce the amplitude of the SAO as in the MIX1 model simulation of Meehl (1991) illustrated in Figs. 2 and 3.

To explore this hypothesis, we show surface temperature anomalies from a combined land (Jones, 1994) and SST (Parker et al., 1995) data set that has been used extensively to document climate trends (e.g. IPCC, 1996). Though there are limitations to this data set in the Southern Hemisphere (see discussion in IPCC, 1996), these data could provide circumstantial evidence for a possible change in the seasonal cycle at mid- and high latitudes of the SH. The seasonal cycle near the two key latitudes for the SAO mechanism are shown in Fig. 5 (52.5°S at top, and 67.5°S at bottom) for the pre-1979 era (solid line, 1951–1979), and the post-1979 era (dashed line, 1980–1995). Reflecting the low-frequency aspect of surface temperatures documented by Bottomly et al. (1990) and noted for the tropical Pacific by Hurrell and van Loon (1994), there is annual mean warming for the later period compared to the earlier period at both latitudes (+0.26°C at 52.5°S, and +0.22°C at 67.5°S). However, this annual mean warming is not distributed evenly throughout the seasonal cycle. At both latitudes there is somewhat greater warming in the first half of the year, while at 67.5°S there is actually relative cooling in October-November-December in the more recent period.

Since there could be problems with data sampling in the surface data, it is useful to examine station data to see if a similar kind of change in the seasonal cycle at these two latitudes took place in the more recent period. As pointed out by van Loon (1967) and Meehl (1991), the 500 mb temperatures are a good indicator of the midtropospheric temperature gradient associated with the SAO (e.g. see Fig. 1). The 500 mb temperatures for the two periods are plotted in Fig. 6 for a pair of stations situated in positions representative of the two key latitudes, Marion Island (47°S, 38°E) and Davis (69°S, 78°E). As in the surface temperatures in Fig. 5, the 500 mb temperatures for these two stations reflect similar secular changes, with annual mean warming for the post-1979 period (1980–93) compared to the earlier period (1951–1979) of +0.61°C at Marion with slight annual mean cooling (-0.05°C) at Davis. However, as was shown for the surface temperature data, there is greater warming in the first half of the year at
both stations for the more recent period compared to the earlier period, with relative cooling at Davis in
the latter part of the year.

The consequence of such alterations of the seasonal cycle of monthly mean temperatures is
illustrated in Fig. 7 which shows a plot of the difference in the temperature gradient between the two
latitudes defined as

\[ T_{\text{index}} = (\Delta T_{50S}) - (\Delta T_{65S}) \]  

(1)

where \( \Delta \) is the the difference between the post-1979 minus pre-1979 temperatures. The dashed lines in
Fig. 7 are the latitude pair differences from 52.5°S minus 67.5°S as depicted in Fig. 5, and to check for
consistency the nearby latitudes 47.5°S minus 62.5°S are also shown. For both the surface
temperatures and the station pair at 500 mb (solid line), there is an anomalous increase of the
temperature gradient for the post-1979 period compared to the pre-1979 period near the middle and end
of the year. These are the times of year noted in Fig. 4 to have a suggestion of increased cyclonic
activity in the post-1979 period compared to the pre-1979 period. The evidence from this limited data
suggests that alterations of the seasonal cycle of monthly mean surface temperatures, present through
the depth of the troposphere, changed the second harmonic of the temperature gradient between 50°
and 65°S. That is, the second harmonic of the change in temperature gradient between these two
 latitudes with May-November maxima interferes with or modulates the climatological second harmonic
with March-September maxima (Fig. 2a). The result is a flattening of the seasonal cycle of baroclinicity
in the later period and a consequent reduction of SAO amplitude. This is similar to what occurred in the
MIX1 model simulation of Meehl (1991) discussed in Figs. 2 and 3. Thus secular changes in the
amplitude of the SAO can be traced to a modulation of the seasonal cycle of surface temperatures near
50°S and 65°S that are an integral part of the SAO mechanism first proposed by van Loon (1967).

5. CONCLUSIONS

1. The SAO is characterized by expansion and contraction of cyclonic activity in the mid and
high southern latitudes twice a year, and was the dominant signal in the annual cycle prior to 1979.

2. The mechanism, as first proposed by van Loon (1967), arises from a polar continent south of
about 65°S surrounded by ocean around 50°S, and the different seasonal cycle of surface temperatures
seen throughout the depth of the troposphere at those two latitudes.

3. A GCM with a weak SAO of the correct phase, when coupled to a simple non-dynamic slab
ocean with an altered seasonal cycle of SSTs at 50°S and surface temperatures near 65°S, has an SAO
of reduced amplitude as a consequence, suggesting that an alteration of the seasonal cycle of surface

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temperatures at the key latitudes of 50°S and 65°S could change the manifestation of the SAO (Meehl, 1991).

4. Analysis of observations of surface temperatures, along with 500 mb temperatures from a station pair, provides circumstantial evidence that, in addition to annual mean surface warming since 1979 at both latitudes, the warming has not been uniformly distributed throughout the year. There has been greater warming in the first half of the year, and slight cooling during the later part of the year near 65°S. Thus, the second harmonic of the change in temperature gradient in the post-1979 period with May–November maxima interferes with or modulates the climatological second harmonic with March–September maxima in the pre-1979 period. The consequence is a reduction in amplitude of the SAO since 1979.

References


Figure 1. Annual cycle of monthly mean values of 500 mb temperature (C) at 50 and 65°S, smoothed long term mean observed values from Australian analyses (1972–84) superimposed to show the two times of year in the mean when the gradient between 50°S and 65°S intensifies (indicated by arrows), thus providing a manifestation of the SAO at middle and high southern latitudes. Solid line is 50°S (temperature scale at left); dashed line is 65°S (temperature scale at right) (after Meehl, 1991).
Figure 2. Annual cycle of global zonal mean 500 mb temperature differences (C), 50°S minus 65°S, for (a) long-term mean observations (1972–84), (b) SPEC SST model case (atmospheric GCM run with pre-1979 climatological SSTs), and (c) MIX1 model case (same atmospheric GCM coupled to a non-dynamic 50 m deep slab ocean). Dots connected by thin line indicate monthly mean values; dashed line is first harmonic; solid line is second harmonic. Percentages indicate amount of total variance of monthly mean values accounted for by first and second harmonics (after Meehl, 1991).
Figure 3. Annual cycle of monthly mean values of 500 mb temperature (°C) at 50°S and 65°S for (a) values at 50°S for unsmoothed observations (solid line), SPEC SST (dot-dash line), and for MIX1 (dashed line). (b) same as (a) except for 65°S (after Meehl, 1991).
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Semi-Diurnal Variations in the Budget of Angular Momentum in a General Circulation Model and in the Real Atmosphere

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1. INTRODUCTION

In a study of the atmospheric angular momentum ($M$) budget in data from every time step of a simulation by the NCAR Community Climate Model (CCM2) we noticed a striking, mostly semi-diurnal, variation. Here we look more closely at the diurnal and semi-diurnal variations in the budget of $M$ in both the model and in available observations of the real atmosphere. Special attention is paid to a semi-diurnal variation in mountain torques.

That there are diurnal and semi-diurnal variations in $M$ budget in the simulation is not surprising since they have been documented in the wind and pressure fields in studies of other General Circulation Model (GCM) output (e.g. Zwiers and Hamilton, 1986), and in that of the CCM2 itself (Lieberman et al., 1994). That they exist in the real atmosphere has also been known for a long time. Some references that describe relevant, observed pressure and wind oscillations are Haurwitz and Cowley (1973), Wallace and Tadd (1974), and Whiteman and Bian (1996).

It is important to document the diurnal changes in $M$ in our quest for a fuller understanding of the atmosphere. In addition, they have important geodetic implications. Better estimates of the rotation rate of the earth, or length-of-day (LOD), have allowed its close connection to $M$ changes to be clearly demonstrated (see review by Rosen, 1993). To date, most related studies have considered variations in $M$ and LOD on time-scales of a few days or longer. Now, time series of LOD are becoming available with time resolution of three hours and better (Freedman et al., 1994). As the diurnal variation of the LOD is better defined it is important to better define the $M$ budget of the atmosphere as well as that of the ocean. Herring and Dong (1994) concluded that subdaily changes in $M$ were on the order of 1 μs ($10^{-6}$ second) while those of LOD were about an order of magnitude larger suggesting that ocean tides were the biggest player in exchanging momentum with the solid earth on these short time scales. Here, we look again at the $M$ changes, in both a GCM (CCM2) and in the real atmosphere using NCEP/NCAR Reanalysis data.

In Section 2 to follow, we briefly describe the simulated and observed variables that are studied. The diurnal and semi-diurnal variations found in the CCM2 simulation are presented in Section 3. Section 4 contains parallel findings from the real atmosphere as depicted in the NCEP/NCAR Reanalysis.

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2. VARIABLES

The $M$ is often considered as made up of two components. One is omega momentum ($M_\Omega$) which is angular momentum that a parcel of air that is at rest with respect to the earth would have by virtue of the rotation rate of the earth. It is determined by:

$$M_\Omega = \frac{\Omega a^4}{g} \int_0^{2\pi} \int_{\frac{\pi}{2}}^{\frac{\pi}{2}} p_s \cos^3\phi \, d\phi \, d\lambda$$

where $a$ is the earth radius, $\Omega$ is the earth angular velocity, $p_s$ the surface pressure, $\lambda$ the longitude and $\phi$ the latitude. The second is relative angular momentum ($M_r$) and is that part of the total $M$ contributed by air motion relative to the earth. It is determined by:

$$M_r = \frac{a^3}{g} \int_0^{2\pi} \int_{\frac{\pi}{2}}^{\frac{\pi}{2}} u \cos^2\phi \, d\phi \, d\lambda \, dp$$

and thus:

$$M = M_\Omega + M_r$$

The rate of change of $M$ is forced by pressure and friction torques. Pressure torques arise as a consequence of unequal pressure on the eastern and western sides of mountains and small-scale topographic features, and they are often referred to as mountain torques ($T_m$). Lower pressure to the east of mountains speeds up the rotation rate of the earth, shortens the LOD, and decreases $M$. $T_m$ is estimated by:

$$T_m = -a^3 \int_0^{2\pi} \int_{\frac{\pi}{2}}^{\frac{\pi}{2}} p_s \cos^2\phi \frac{\partial H}{\partial \chi} \, d\phi \, d\lambda$$

where $H$ is the height of the earth’s surface.

Friction torques ($T_f$) always act to slow down the winds no matter what their direction, so they contribute to increasing ($T_f > 0$) $M$ when acting on easterlies and decreasing ($T_f < 0$) $M$ when acting on westerlies. $T_f$ is given by

$$T_f = a^3 \int_{\lambda=0}^{2\pi} \int_{\phi=-\frac{\pi}{2}}^{\frac{\pi}{2}} \tau_\lambda \cos^2\phi \, d\phi \, d\lambda$$

where $\tau_\lambda$ is the east-west wind stress which is negative for westerlies.
Early climate simulations tended to have too large values of $M$. It was recognized that small-scale features that were not included in the models could be the cause. In an attempt to include some of their effects, so called gravity wave stress ($\tau_g$) was introduced. It is a stress representing the interaction of the surface with the atmosphere caused by subgrid scale vertically propagating gravity waves. Gravity wave torque ($T_g$) between the surface and some upper level is given by:

$$T_g = a^3 \int_{p_f}^{p_t} \int_0^{2\pi} \int_{\frac{\pi}{2}}^{\pi} \tau_g \cos^2 \phi \, d\phi \, d\lambda \, dp$$

(6)

where $p_t$ is the pressure of the top level.

\textbf{a. Simulation}

The simulation is the same as that studied in Lejenäs et al. (1997). The model has a horizontal spectral resolution of T42 with 18 levels in the vertical. The top level of the model is at 4.809 hPa and there is a rigid lid at 2.917 hPa. Time step intervals are 20 minutes. The model was discussed fully by Hack et al. (1993). The simulated data considered here consist of 2448 time steps or 34 model days starting at 30 December of year 2 of the 20-year simulation. All the variables necessary to solve the above equations were available with one exception. The exception is $\tau_g$ at $p = 2.917$ hPa, which was not saved. As a result we use only the surface value of $\tau_g$ to estimate the gravity wave drag torque, as opposed to the difference between the gravity wave stress at the top and bottom of the model atmosphere.

\textbf{b. Observations}

The observational data used here are the NCEP/NCAR Reanalysis products for the 14-year period 1982–1995. We worked with the zonal winds and surface pressure interpolated to the same horizontal, T42, scale as that of the simulation. The zonal winds were at 14 levels in the vertical from 1000 to 10 hPa. We used pressure intervals in approximating the integral of (2) so that the lowest pressure was zero as opposed to the 2.917 hPa for the simulation data. Data were available at 00, 06, 12, and 18 UTC. Only January data were included. We estimate $M_\Omega, M_r, T_m,$ and $T_f$ from available data but made no attempt to account for unresolved effects such as done with the model data and (6).

\textbf{3. RESULTS FROM CCM2}

Figure 1 shows values of $T_f, T_m,$ and $T_x$ along with $M_r$ and $M_\Omega$ for the 34-day time period extending from late December of a model year through January of the next. These data are plotted for every time step. The fast variations in $T_m, M_r,$ and $M_\Omega$ are predominantly semidiurnal variations. Those of $T_f$ and $T_x$ are diurnal although there are indications of a semi diurnal variation as well.

We can isolate the diurnal and semidiurnal oscillations by averaging together the 34 values that occur at the same time step each day. The resulting values for each of the 72 time steps are presented in Fig. 2. The 72 respective time step average has been subtracted from the values in each panel of Fig.
The semidiurnal variation in $T_m$ is clear. Peak-to-trough variations of the semidiurnal variation are about 15 Hadleys as compared to variations over a few day periods of about 50 Hadleys (from Fig. 1). The predominate diurnal variation in $T_f$ has a range near 20 Hadleys while that at longer time scales evident in Fig. 1 are also about 50. In the case of $T_g$, diurnal variations are about the same size as ones on the longer time scales of Fig. 1. Interestingly, phases of the large diurnal variations in $T_f$ and in $T_g$ are such that, taken together, they tend to reinforce the semidiurnal variations. Asterisks in Fig. 2 are observed values based on NCEP/NCAR Reanalysis data, and they will be discussed in the following section.

The diurnal variation in the balance between the rate of change of total atmospheric angular momentum and total torque is shown in Fig. 3. The total torque is $T_f + T_m + T_g$. Values in Fig. 3 are not anomalies so that the average over 24 hours (72 time steps) need not be zero. The angular momentum can change during this model January. Of course, the average rate of change of $M$ should equal the average total torque though. The slight bias (averages of 3.2 versus 1.4 Hadleys) is due, we think, to the way we estimated $T_g$. There is a well marked semidiurnal variation in torque and in the resulting rate of change in $M$. There are maxima near 0300 and 1500 UTC.

Figures 4, 5 and 6 show the contribution from sectors of the globe to the diurnal variation of anomalies $T_f$, $T_m$, and $T_g$, respectively. The sum of the four traces in each of these figures gives the respective total torques of Fig. 2. The Western Hemisphere is the chief contributor to the sharp minimum in $T_f$ that occurs after 18 UTC.

The semi-diurnal variation in $T_m$ (Fig. 5) is very clear in the Eastern Hemisphere, although it is roughly one quarter of a cycle out of phase between the north and south. On the other hand, north of the equator, the semi-diurnal variation in the Western Hemisphere is very nearly in phase with that in the Eastern Hemisphere. These semi-diurnal variations are driven by a semi-diurnal surface pressure tide, whose zonal wave length is wave number two. Lieberman et al. (1994) studied the atmospheric tides in CCM2, and the semi-diurnal pressure tide looks very much like the one observed in the real atmosphere (Haurwitz, 1956; also reproduced in Chapman and Lindzen, 1970). The fact that the Rockies and Himalayas are approximately separated by 180 degrees of longitude results in the nearly in phase variation in $T_m$ in the two hemispheres. The diurnal variation in $T_g$ (Fig. 6) results almost entirely from the Eurasian region.

4. RESULTS FROM NCEP REANALYSIS

We used 14 Januaries (1982–1995) of NCEP/NCAR Reanalysis data to compare the model results to. Wind data were interpolated to 14 pressure levels in the vertical from 1000 to 10 hPa. We also used surface pressure data for $M_{\Omega}$ and winds from the first sigma level (0.995, about 5 hPa or 40 m above the surface) and a simple bulk formula to estimate $T_f$. Data were separated according to synoptic time, and resulting anomaly values of $T_f$, $T_m$, $M$, and $M_{\Omega}$ at 00, 06, 12,18 UTC are indicated in Fig. 2 by the asterisks. The anomalies are simply the difference between the value at the indicated hour and the value averaged over all four hours.

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A semidiurnal variation which is reasonably well simulated by the CCM2 is evident in the observed $M_r$ and $M_\Omega$. The four observation times occur near the inflection points of the expected semidiurnal variation in $T_f$ and $T_m$, and, accordingly, semidiurnal variations are not observed in these variables.

We can attempt to learn a little more about the approximate behavior of $T_m$ at more than the four synoptic times. We could, for example, take the amplitude of the semidiurnal pressure wave from Haurwitz (1956) and simulate its passage around the world in 24 hours and compute $T_m$ at any time resolution we like. Rather than that, we use the semidiurnal pressure wave determined from the reanalysis data. Fig. 7 is the anomaly surface pressure at the four synoptic times. The wave-2 pattern is evident as is a relative maximum in amplitude in the tropics as shown by Haurwitz (1956). There are also some small-scale irregularities, as for example near the Andes Mountains. The panels at the right of Fig. 7 show positive (negative) anomalies in zonal mean pressure at 06(00) and 18(12) UTC in the equatorial region consistent with the positive (negative) anomalies in $M_\Omega$ at those times.

The observed semidiurnal variation in $T_m$ was approximated at 64 time steps during a 24-hour period (22.5 minute intervals) by shifting the 00 UTC anomaly pressure pattern of Fig. 7 two grid point westward at a time and computing $T_m$. Sixty-four time steps result because two grid points are a little more than 5.6° of longitude apart in our T42 data. Of course the small scale features that may be regularly associated with the semidiurnal pressure wave, such as those near the Andes at 00 UTC, are definitely not realistic when they are shifted over the ocean. Nevertheless, the resulting semidiurnal variations (Fig. 8) are realistic and those from the CCM2 taken from Fig. 2 are consistent with them. Besides values determined from the 00 UTC pattern of Fig. 7, those determined from the patterns of the three other synoptic times are also included in Fig. 8.

Figure 9 is like Fig. 5 with areal contributions to $T_m$ determined from the CCM2 data, and here the average of those determined from the 00, 06, 12, and 18 UTC anomaly pressure patterns of Fig. 7. The agreement between CCM2 and observed values is very good. These four areal contributions can be directly related to mountain ranges and to the pressure anomalies of Fig. 7. For example, the lower left panel (Andes mountains) indicates a maximum in $T_m$ about 12 UTC. From Fig. 7 we see that the anomaly pressure gradient is directed eastward over the Andes as it should at that time.

5. CONCLUSIONS

The CCM2 simulation shows diurnal and semidiurnal variations in the $M$ budget. Data from the NCEP/NCAR Reanalysis are generally consistent with these simulated results. The semidiurnal variation in $T_m$ is due to the westward propagating semidiurnal solar tide. The semidiurnal changes in $M$ correspond to about 10 $\mu$s changes in LOD.
References
Figure 1. Frictional ($T_f$), mountain ($T_m$), gravity wave drag ($T_g$) torques, and relative ($M_r$), omega ($M_\Omega$), and effective total atmospheric angular momentum ($M_r + 0.7M_\Omega$; see Rosen, 1993) for each 20-minute time step during a 34-day period from late December of a CCM2 model year through January of the next model year. Torque units are $10^8 \text{ kg m}^2/\text{s}^2$ (Hadleys), and those of angular momentum are $10^{25} \text{ kg m}^2/\text{s}$. 
Figure 2. Like Fig. 1, but with data composited according to the time of day showing the diurnal variations. Results from the CCM2 are depicted by the smooth line which is determined by 72 time steps averaged over the 34 days. Asterisks denote observed values calculated from NCEP reanalysis data for 14 January at the synoptic times of 00, 06, 12, and 18 UTC. The vertical bar at the left-hand side of the panel for effective total $M$ represents an equivalent change in the length of day (assuming an $M$ change lasted for 24 hours) of 10 microseconds. (For example see Rosen, 1993)
Figure 3. The rate of change of total atmospheric angular momentum determined from the sum of $M_r + M_\alpha$ from Fig. 2 (dashed) and the sum of the three torques from Fig. 2 (solid) for 72 time steps (one day from 00 to 23 hours 40 minutes UTC).
Figure 4. Average areal contributions to anomaly $T_f$ for 72 time steps (one day from 00 to 23 hours 40 minutes UTC) for CCM2 data. Bottom panel is the sum of the four areal contributions and equal to $T_f$ values of Figure 2.
Figure 5. Same as Fig. 4 but for $T_m$. 

Mountain Torque – Hadleys
Figure 6. Same as Fig. 4 but for $T_g$. 
Figure 7. Observed surface pressure anomaly at four synoptic times determined from averages over 14 Januarys of NCEP/NCAR Reanalysis data. The anomalies are determined by subtracting the average of all four hours.
Figure 8. The diurnal variation of $T_m$. The line is for CCM2 data and comes from Fig. 2. Other symbols represent the observations approximated by moving the four synoptic hour pressure maps around the earth past the mountains. There are 64 points, or simulated time steps, for each of the observed pressure anomaly maps. The 'o' at 00UTC is the same as the asterisk at 06 UTDC in Fig. 2, etc.
Figure 9. Same as Fig. 5 but with observed areal contributions simulated for 64 time-steps during the day in a manner similar to that of Fig. 8 depicted by the asterisks. Rather than plotting values determined from each of the four anomaly pressure maps, the asterisks are the average of the four for clarity.
The Effect on the Lower Stratosphere of Three Tropical Volcanic Eruptions

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ABSTRACT

The eruptions of three low-latitude volcanoes (Mt. Agung, El Chichón, and Pinatubo) have in the recent past disturbed conditions in the stratosphere. In particular, the temperature structure in the tropics and the dynamics of the extratropical circulation in winter were affected. We describe some of the probable effects of the volcanic eruptions as they appeared in the stratosphere on the Northern Hemisphere, and stress those differences between the effects of the three eruptions that are associated with the state of the Quasi-Biennial Oscillation during and after each of the eruptions.

1. INTRODUCTION

Many studies have tried to show that volcanic eruptions cause a drop in the earth's surface temperature (see, e.g., the summaries in Lamb, 1972; Kondratyev, 1983); but evidence was equivocal until the arrival of satellite technology. The Microwave Sounding Unit carried by NOAA satellites showed beyond doubt that the eruptions of both El Chichón (April 1982) and Pinatubo (June 1991) were followed by a drop in the tropospheric temperature. Pinatubo's eruption had the greater effect. The direct solar radiation decreased by as much as 30% and the temperature of the Northern Hemisphere fell 0.7°C in the lower troposphere during the 14 months after the eruption (Dutton and Christy, 1992).

Newell (1970) showed, with the comparatively meager material available in and before 1963, that Mt. Agung's eruption in March 1963 raised the temperature of the lower stratosphere. His paper was met with scepticism, but Labitzke and Naujokat (1983) confirmed his conclusions. In the present paper we compare the effects of the three volcanoes on the stratosphere in terms of 1) the temperature changes after the eruptions, 2) the importance of the state of the Quasi-Biennial Oscillation (QBO), and 3) the response of the cyclonic vortex in winter. The analyses made in the Stratospheric Research Group, Freie Universität Berlin, provide the data for this investigation. They can be obtained from the University and from the data archives at the National Center for Atmospheric Research, Boulder, Colorado. The series of geopotential heights begins in 1958, and the temperature series in July 1964.
2. THE EFFECT OF AN ERUPTION ON THE LOWER STRATOSPHERE

a. The Temperature Rise

Newell (1970) provided, as mentioned, the first evidence of a warming of the tropical lower stratosphere after a low-latitude volcanic eruption; and several papers appeared after El Chichón’s eruption that outlined the temperature rises which then took place (Fujita, 1985; Labitzke et al., 1983; and Parker and Branscombe, 1983, among others). Briefly outlined: after the arrival of the volcanic plume in the stratosphere the temperature in the tropical stratosphere begins to rise, and the warming spreads across the tropical and parts of the subtropical regions.

The temperature anomalies at the 30 hPa level from the month of the eruption till October of the same year are shown in Fig. 1 for El Chichón (Mexico) and Fig. 2 for Pinatubo (the Philippines), respectively at 17°N and 15°N. In the month when the eruptions took place the 30 hPa temperature anomalies were negative in the equatorial belt in both instances, and the vertical wind shear was easterly in the QBO (Fig. 3a). The easterly shear indicates that the vertical motion was upward which is consistent with the negative temperature anomalies on the equator—when the vertical wind shear is westerly the air sinks and warms (e.g., Reed, 1965, Fig. 9).

In the month following the eruption (May 1982 for El Chichón, Fig. 1b; and July 1991 for Pinatubo, Fig. 2b) there were positive anomalies at 20°N–25°N in 1982 and at 15°N–30°N in 1991; that is, in the areas over which satellites showed that the volcanic debris had drifted. The positive anomalies in the arctic in 1982 were present before the eruption and were gone in June. The positive anomalies in the east Pacific Ocean in 1982 were four standard deviations above the mean, and in the east Atlantic Ocean in 1991 they were three standard deviations above. On the equator the vertical wind shear was still easterly and the temperature anomalies were therefore negative in 1991 and negative or weak positive in 1982. The volcanic aerosol had in both instances spread from the latitude of the eruption to the equator (Bluth et al., 1992), but its warming effect was strongest in the outer tropics whereas the cooling associated with the rising motion in the QBO prevailed on the equator.

Two months after the eruption (Figs. 1c and 2c) the wind shear was almost zero at 30 hPa in 1982, but in 1991 it was still easterly. The anomalies were then positive all over the tropics; in 1982 the largest positive deviations were on the equator, but in 1991 they were north of the equator along more than half of the circumference. The effect of the aerosol had thus overcome or diminished the effect of the vertical motion in the QBO on the equator, but more so in 1982 when the wind shear at 30 hPa was weak easterly below 50 hPa and westerly above, than in 1991 when it was easterly as high as 20 hPa.

Three months after the eruption the vertical wind shear was westerly in 1982 and the positive temperature anomalies were as large as 10°C on the equator (Fig. 1d), which equals five standard deviations. At the same stage in 1991 (Fig. 2d) the positive deviations on the equator were only 2°–4°C, a fact which we ascribe to the strong easterly vertical shear with associated rising motion which counteracted the warming effect of the volcanic aerosols.
In order to compare the magnitude of the warming due to the volcanic aerosols with the "normal" temperature anomalies connected with the respective phase of the QBO, we selected "undisturbed" periods comparable to both cases, Figs. 3b–e. For the El Chichón case one should compare Fig. 3b (April 1977) with Fig. 1a (April 1982), i.e. in both cases an undisturbed east phase situation with weak negative anomalies near the equator; and Fig. 3c (July 1977) with Fig. 1d (July 1982), i.e. the "undisturbed" positive anomalies of +4°C against positive anomalies up to +10°C following the El Chichón eruption. And for the Pinatubo case: Fig. 3d (June 1986) compares with Fig. 2a (June 1991), i.e. in both cases again an "undisturbed" east phase situation with weak negative anomalies near the equator; and Fig. 3e (September 1986) with Fig. 2d (September 1991), i.e. continuously negative anomalies in the "undisturbed" case against positive anomalies up to +4°C in the subtropics following Pinatubo's eruption.

The difference in the size of the positive low-latitude temperature deviations continued in the following month (Figs. 1e and 2e), presumably because the westerly vertical shear increased in 1982 into 1983, whereas the easterly vertical shear strengthened in 1991 and continued into 1992.

The historical analyses of the stratospheric temperatures do not cover the period after Mount Agung on Bali erupted in March of 1963. However, monthly mean maps of July 1962 and 1963 were analyzed as a special project in the Stratospheric Research Group after the eruption of El Chichón in order to compare the influence of the two volcanoes (Labitzke and Naujokat, 1983). The map of July 1963 is reproduced in Fig. 4 and presents the temperature anomalies four months after the eruption, and it thus corresponds to Figs. 1e and 2e. The size of the deviations is the same as of those observed at the same stage after the eruption of El Chichón. The warming of the lower tropical stratosphere was therefore appreciably larger at this stage after Agung's than after Pinatubo's eruption, and it is likely that this difference is due to the fact that the equatorial vertical wind shear at 30 hPa in July 1963 was westerly (Fig. 3a), that is, the same as it was at the same time after El Chichón's eruption.

Figure 5 shows the difference—for each of the three tropical volcanic eruptions—between a month after the eruption and the same month in a year without an eruption. The phase of the QBO in Figs. 5a,b was westerly and in Fig. 5c easterly in both the year of the eruption and the one without. The figure demonstrates that when the effect of the QBO is subtracted, the volcanic effect is the same in all three cases.

b. The Stratospheric Cyclonic Vortex in Fall and Winter

In a study of the influence of the Southern Oscillation on the stratosphere (van Loon and Labitzke, 1987) we found a tendency for the polar vortex to be comparatively weak in the winter of the mature phase of a warm event, as defined by Rasmusson and Carpenter, 1982 (as an example see Fig. 6a, from Pawson et al., 1993). Two winters in our study did not fit into this pattern: the winters of 1963/1964 and 1982/1983. In both of these winters the stratospheric cyclonic vortex remained strong. In another paper (Labitzke and van Loon, 1989) we pointed out that both of these winters followed eruptions of tropical volcanoes (Mt. Agung and El Chichón), and it turned out that the polar stratospheric cyclone in
the northern winter after Pinatubo's eruption—which was in the mature phase of a warm event too—also failed to weaken. Figure 6b shows the average 30 hPa height anomalies in February for the three winters after volcanic eruptions and in the mature phase of the Southern Oscillation.

In the following we suggest a reason why the vortex in the winters after the eruptions failed to weaken, although they were all in the mature phase of a warm event of the Southern Oscillation. In the fall after the eruptions of 1982 and 1991 the temperature anomalies were positive from the equator into middle latitudes as a result of the eruptions, and negative or negligible farther north. The meridional temperature contrasts (thermal wind) between low and high latitudes were thus stronger than normal in the season when the cyclonic vortex forms (Figs. 1f and 2e).

At the beginning of winter (December, Fig. 7) negative anomalies of the geopotential heights over the arctic were surrounded by the volcanically induced positive anomalies in all three instances; that is, the vortex was above normal strength in the early winter after each of the eruptions. The winter as a whole (Fig. 8a) is in each instance characterized by positive anomalies in lower and negative anomalies in higher latitudes, with associated anomalously strong westerlies north of about 40°N and weaker westerlies to the deviations below normal at higher latitudes (Fig. 8b).

### 3. CONCLUSION

In the eruptions of three low-latitude volcanoes (Mt. Agung, 1963; El Chichón, 1982; and Pinatubo, 1991) the volcanic plume reached into the stratosphere where it first spread along the circumference of the earth in the tropics and later into high latitudes. The presence of volcanic products in the tropical stratosphere caused a warming there which persisted for many months after the eruption. The vertical wind shear in the equatorial belt associated with the Quasi-Biennial Oscillation was easterly at the time of each eruption, but in 1963 and 1982 it soon changed to westerly, whereas in 1991 the easterly shear persisted. Easterly vertical shear is associated with lower and westerly shear with higher temperatures, and therefore the warming of the equatorial stratosphere was more noticeable in 1963 and 1982 than in 1991. If one corrects for the influence of the QBO, the effect of the volcanic eruption is the same in all three instances. The fact that the tropical stratosphere stayed abnormally warm into the following fall when the cooling began at high latitudes resulted in steep meridional temperature gradients (thermal wind) across middle latitudes at the time when the stratospheric cyclonic vortex forms. In all three instances the volcanic eruption took place in a year when a warm event developed in the Southern Oscillation. The signal of a warm event in the extratropical stratosphere in winter is a weaker than normal vortex, but because of the lower latitude warming in the stratosphere associated with the eruptions and the resulting strong meridional gradients, the winter vortex stayed strong in the winter after each eruption.
References


Figure 1. Deviations of the 30 hPa temperature from the mean of 1965–1974, after the eruption of El Chichón in 1982. In °C.

Figure 1. Deviations of the 30 hPa temperature from the mean of 1965–1974, after the eruption of El Chichón in 1982. In °C.
Figure 2. The same as Fig. 1, but after the eruption of Pinatubo in 1991.
Figure 3. (a) The wind in the Quasi-Biennial Oscillation during the period surrounding each of the three volcanic eruptions. The heavy lines are drawn through the strongest easterly winds and separate vertical easterly from vertical westerly shear. In m s⁻¹.
Figure 3. (b–e) The same as Figs. 1 and 2, but for "undisturbed" periods.

Figure 4. The 30hPa temperature anomalies in July 1963, after the eruption of Mt. Agung. In °C.
Figure 5. The difference between a month in years with volcanic eruptions and a month in a year without - both months being in the same phase of the QBO.
Figure 6. (a) The 30 hPa anomalies of geopotential height (in dekameters) in February during warm events in the Southern Oscillation without volcanic disturbance. (b) The same deviations in three warm events after volcanic eruptions. From Pawson et al., 1993.
Figure 7. The 30 hPa geopotential height deviations from the mean of 1964–1973 (in dekameters) in December 1963, 1982, and 1991.
1. INTRODUCTION

The Southern Hemisphere westerly winds exert a primarily zonal eastward wind stress over the Southern Ocean between 30° and 70°S (Trenberth et al., 1990). The absence of any continental boundary at the latitudes of Drake Passage (55° to 65°S) means this stress cannot be balanced by a zonal pressure gradient (Sverdrup balance) as it is at other latitudes. Instead a predominantly zonal Antarctic Circumpolar Current (ACC) is generated, that transports more than 100 Sv (1 Sv = 10^6 m^3/s) of water through Drake Passage (Nowlin and Klink, 1986). The dynamical forces balancing the wind stress forcing of this fast moving current system are still the subject of debate (Warren et al., 1996) with some of the candidates being continental boundaries, bottom topography, stratification and dynamical instabilities (Nowlin and Klink, 1986). If the dominant drag mechanisms do not adjust instantaneously, then the current speed and transport would be expected to respond to changes in the zonal wind stress.

Large and van Loon (1989) presented evidence of ACC response to semiannual forcing during the First GARP Global Experiment (FGGE) year of 1979. This work is reviewed in Section 2. However, the semiannual oscillation equatorward of about 50°S was very different throughout much of the 1980s (Hurrell and van Loon, 1994). The associated change in forcing and in ACC response is discussed in Section 3. Finally recent ocean observations suggesting a return to pre-1980 conditions are presented in Section 4.

2. PRIOR TO THE 1980s

Studies of Southern Hemisphere near-surface meteorology prior to 1980 reveal that the half-year wave in sea level pressure (SLP) has circumpolar amplitude peaks over Antarctica and at about 50°S (van Loon and Rogers, 1984). There is an amplitude minimum around 60°S, where there is a 180° phase shift. This pressure pattern leads to a semiannual wave in the zonal surface wind stress that has a minimum amplitude at about 50°S, where the phase changes from equinoctial maxima at higher latitudes to solstitial maxima equatorward (van Loon and Rogers, 1984).

This latitudinal distribution of the amplitude and phase of the semiannual harmonic of the 1979 zonal geostrophic wind is shown in Fig. 1 (dashed traces) for the Atlantic Ocean sector. At 35°S there is a nearly 7 ms⁻¹ increase in this wind component from March to June and again from September to December. A comparable increase is seen at 60°S, but instead from December to March and again from June to September. These cycles contribute as much as 74% of the zonal wind variance of monthly means. At 45°S the amplitude of the semiannual harmonic is a minimum of less than 1 ms⁻¹. A similar distribution is evident in the other ocean basins (van Loon and Rogers, 1984). In the central Pacific
Large and van Loon (1989) used the observed displacements of the FGGE drifting buoy array as a proxy for surface currents, and discuss the implications of using these imperfect current followers. They found highly significant, large scale, low frequency variability in the zonal drift of the buoys. This variability appeared to be wind forced, because of similar meridional and zonal patterns in the amplitude and phase of the semiannual harmonic. The results from the Atlantic sector are shown in Fig. 1 (solid traces) for easy comparison to the wind. The buoy drift also has relative maxima in second harmonic amplitude about 60°S and around 40°S, with solstitial maxima at the latter and equinoctial maxima at the former. At 45°S, the 16 cm/s March to June increase in speed is a considerable fraction of the 24 cm/s mean buoy drift and the second harmonic accounts for about 60% of the variance in the 12 monthly means of 1979.

Similar amplitude and phase patterns are found in the Pacific and Indian Ocean sectors, but the correspondence with the wind results is not always exact. In general, the drift patterns tend to be even larger scale than those of the wind, with even more low frequency contributions. Therefore, Large and van Loon (1989) concluded that the drift is a low-pass filter response to the wind that is advected away from the local forcing regions and that the advection extended over at least the scale of an ocean basin.

There are several details of the Fig. 1 comparison that also suggest that the buoys are not being directly forced by windage. First, south of 50°S the buoy drift displays a two week to one month phase lag relative to the local winds. Second, the solstitial and equinoctial amplitude maxima are similar in the winds, but the former are significantly greater in the drift. Finally, there are drift and wind amplitude maxima at 60°S, but the former accounts for much less a fraction (11%) of the variance than does the latter.

Three months of Seasat altimeter data (July–October 1978) have been used by Fu and Chelton (1984, 1985) as an indicator of large scale variability of the ACC. Their time series of north-south differences in sea level across the ACC are measures of eastward geostrophic flow poleward of 50°S. At each section there is a general increase, which can be interpreted as a rise from a June minimum to the second equinoctial maximum of a semiannual wave in response to the wind forcing. These results indicate that the amplitude of the semiannual harmonic is a minimum in the vicinity of Drake Passage, with maxima of about 2 cm/s in the middle of the three ocean basins. A small 0.5 cm/s amplitude wave with an equinoctial maximum is also shown by measurements of the pressure difference across Drake Passage over 4 years 1977–1981 (Large and van Loon, 1989). Other studies have concluded that Drake Passage transport and ACC as a whole respond to the circumpolar averaged wind stress from 43°S to 65°S (Wearm and Baker, 1980). Therefore, the equinoctial maxima in Drake Passage current and transport is expected.
3. THE DECADE OF THE 1980s

Hurrell and van Loon (1994) document changes in the semiannual harmonic of the middle and high latitude Southern Hemisphere that began around 1980. These sea level pressure changes would be expected to have the following effects on the surface wind stress. South of 50°S the semiannual amplitude should be somewhat weaker, but still significant with equinoctial maxima. Equatorward of 50°S the semiannual amplitude should be much reduced compared to earlier years, and perhaps insignificant compared to a much more dominant annual harmonic.

In order to show these effects, wind stress from the surface wind products of the NCEP/NCAR Reanalysis (Kalnay et al., 1994) have been averaged into a mean monthly climatology using the four years 1985–1988. The amplitude and phase of the semiannual harmonic of zonal wind stress are shown in Fig. 2. Large amplitudes are found poleward of 50°S throughout most of the Indian and Pacific basins, but not the Atlantic. First maxima are generally equinoctial, occurring between 1 February and the end of March. The abrupt phase shift to solstitial maxima at about 45°S is evident only in the Atlantic and Indian Ocean sectors.

The above NCEP/NCAR climatology was used to force a global ocean general circulation model to equilibrium. Model details and further analysis are presented in Large et al. (1997). Of interest here are the amplitude and phase of the semiannual harmonic of zonal surface velocity shown in Fig. 3. The distinct amplitude and phase patterns seen in Fig. 1 are weakly evident in the western Indian Ocean, but not in the Atlantic. Only small isolated regions have amplitudes greater than 1 cm s⁻¹ and these tend to be found near the edges of the ocean basins. The companion Drake Passage transport from the equilibrium model solution is shown in Fig. 4. Although as might be expected equinoctial peaks are evident at 1 April and 1 September, the variance is dominated by an annual period cycle. Whereas the 1977–1981 pressure observations indicate a 1 August minimum in the transport (Large and van Loon, 1989), Fig. 4 shows the flow to be near its maximum at this time of year.

Chelton et al. (1990) study the variability of Southern Ocean surface circulation with the Geosat altimeter data from November 1986 through December 1988. The first empirical orthogonal function (EOF) accounts for 15.2% of the variance and has an annual cycle, in accord with the above results. A semiannual harmonic is associated with the second EOF (9.7% of the variance), but its amplitude is only 1 cm s⁻¹ compared to the 2 cm s⁻¹ suggested by Seasat in 1978. Chelton et al. (1990) conclude that the inferred patterns of semiannual circulation from Geosat altimetry bear some resemblance to those inferred by Large and van Loon (1989). However, unlike Seasat, the 1980s Geosat altimetry shows only a disconnected transition zone between regions of equinoctial and solstitial maxima. At Drake Passage the first EOF shows a large amplitude (8 cm) annual cycle with an early September maximum, that is consistent with Fig. 4. Unlike Seasat, the semiannual harmonic in this region is incoherent, changing phase across the channel. Chelton et al. (1990) discuss several possible explanations for differences between the semiannual variability in 1978–1979 and the Geosat period, including measurement
uncertainty and sampling differences. Another contributor that should be considered is the very different semiannual wind forcing between these two periods.

4. THE DECADE OF THE 1990s

The decade of the 1990s promises a very rich and varied observational dataset of Southern Ocean meteorology and oceanography. The World Ocean Circulation Experiment (WOCE) expended a considerable fraction of its effort in the region. In addition to repeating Drake Passage observations, other areas such as the ACC south of Australia were the focus of unprecedented attention. The surface wind forcing has been very well observed since the launch of the ERS-1 satellite in 1991 with its C-band scatterometer measurements of the surface wind vector. The measurement time series is being continued with the NSCAT Ku-band scatterometer launched on the Japanese ADEOS satellite in August 1996. In addition, a long time series of satellite altimetry began with ERS-1 and is continuing with the TOPEX satellite.

Very few of these observations have yet been analyzed and synthesized. However, surface currents from ERS-1 altimetry passes across Drake Passage have been computed by Challenor et al. (1996). Despite considerable spatial variability across the channel and temporal variability over the eight months, the mean current is predominantly eastward with a mean speed of about 20 cm s⁻¹, which is not inconsistent with the 16 cm s⁻¹ annual average buoy drift found by Large and van Loon (1989).

Meredith et al. (1996) show North-South bottom pressure differences across Drake Passage from most of the 5 year period 1990 through 1994. For the four most complete years these data have been subsampled 24 times per year and replotted in Fig. 5. In 1990 and to a lesser extent in 1991, there is a dominant annual harmonic (dashed curves in upper two panels of Fig. 5), which explains 60% and 34% of the variance, respectively. These high values and the similarity of these two years to Fig. 4 are indicative of a persistence of the conditions of the 1980s to at least 1991. This percentage continues to drop to 21% in 1993 and only 0.4% in 1994, suggesting another transition has occurred. In both these years more variance is explained by the semiannual harmonic (dashed curves in lower two panels of Fig. 5); 35% and 51%, respectively. Both semiannual harmonics have equinoctial (March and September) maxima, and are therefore, reminiscent of the 1977–1981 mean annual cycle shown in Large and van Loon (1989). However, this average has a 4 mb September–October peak, a -4 mb July–August minimum and only a 2 mb March peak. In contrast, the 3 mb March 1994 peak is larger than the September peak and the winter minimum is only -3 mb (Fig. 5, bottom panel). Nonetheless, there is a strong indication of a return to conditions similar to those of the 1970s.

Has the semiannual harmonic surface wind forcing after 1992 returned to a state similar to that described by van Loon and Rogers (1984)?

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Figure 1. Amplitude and phase of the 1979 semiannual harmonic of zonal geostrophic wind (bottom scale; dashed traces) and zonal buoy drift speed (top scale; solid traces) as a function of southern latitude, averaged over 0° to 40°W in the South Atlantic.
Figure 2. Amplitude and phase of the semiannual harmonic of zonal wind stress from a 1985 to 1988 mean monthly climatology based on NCEP reanalysis.
Figure 3. Amplitude and phase of the semiannual harmonic of zonal surface current from an equilibrium ocean general circulation model solution, forced with the winds that produced Fig. 2.
Figure 4. Annual cycle of Drake Passage transport from an equilibrium ocean general circulation model solution, forced with the winds that produced Fig. 2.
Figure 5. Annual cycles (1990, 1991, 1993, and 1994) of North-South bottom pressure difference across Drake Passage (solid curves) from the data of Meredith et al. (1996). The dashed curves trace the harmonic component with the largest amplitude (annual in 1990 and 1991, but semiannual in 1993 and 1994).
The Signal of the 11-year Sunspot Cycle in the Upper Troposphere-
Lower Stratosphere

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1. INTRODUCTION

The concern and uncertainty surrounding the question about a possible human influence on the earth’s climate have in recent years led to renewed interest in the sun’s role in climate change and variability. The research activity in this field is still limited in comparison with the attention paid to, for instance, the effect of greenhouse gases and the change in ozone amounts; and the research is to a large extent carried out by solar and upper-atmosphere physicists rather than meteorologists because the latter tend to approach the subject with habitual scepticism. Their scepticism is rooted in our present inability to explain how changes in solar irradiance can affect the troposphere; and with respect to the 11-year sunspot cycle (SSC) the scepticism was seemingly justified when it turned out that the amplitude of the solar irradiance over an 11-year cycle, as measured by satellite, was only one tenth of a percent. If the solar effect were purely radiative those who doubt that the observed small irradiance change could cause a signal that would rise above the noise level might be right. But the radiance changes are not equally distributed across the solar spectrum, for the changes in the ultraviolet from maximum to minimum in the SSC amount to several percent; and besides, it would be reducing the problem to a false simplicity to think that the effect on the atmosphere is through direct radiation rather than through changes in the general circulation.

Although our purpose is to report on the detection of an appreciable signal of the SSC in the tropical-subtropical lower stratosphere and upper troposphere, we should like to mention some recent investigations which successfully have related long term trends in the surface air temperature statistically to the sun. Friis-Christensen and Lassen (1991) and Lassen and Friis-Christensen (1995), for instance, have found that the variability of the length of the SSC during the past several centuries is well correlated with global temperatures; Reid (1991) found good agreement between 11-year running sunspot numbers and sea surface temperature anomalies since the mid-19th century; and Hameed and Gong (1994) showed that the spring temperature in the Yangtze River valley during the last three centuries closely followed changes in the length of the SSC. Our work is limited to the period after 1950 and to the influence of the SSC, but the areal distribution of the solar influence which it reveals suggests some mechanisms for transmitting the solar variability to the earth’s atmosphere which may be applicable to the association between the SSC and temperature trends which is dealt with in the studies above.
2. THE SOLAR SIGNAL IN THE ANNUAL MEAN

The data which we have used in our analyses are derived from the daily stratospheric maps produced in the Stratospheric Research Group at the Freie Universität, Berlin, supplemented with observations made by radiosonde at selected stations and with 700 hPa grid point data from the U.S. National Meteorological Center. There were 38 years of daily maps of the stratosphere available at the time of writing, 1958–1995, and several radiosonde stations had data as far back as the mid-1950s. We correlated the 38-year time series of annual mean 30 hPa heights (about 24 km above sea level) at all points north of 10°N with the 10.7 cm solar radio flux—used as measure of the SSC—and Fig. 1 shows lines of equal correlation between the two quantities. The significance level of 0.1% is just above 0.5, and the areas where the local significance equals or exceeds this level are shown in the figure by shading. The annual mean heights in the arctic stratosphere at this level are apparently not influenced by the sun, but in the latitudes south of 45°N the correlations are large, with their highest values along 30°N. During this period the solar variation accounted for as much as 50% of the interannual variance of the annual mean 30 hPa heights over the central and eastern Pacific Ocean. The correlations decrease across the tropics along the whole periphery, and the pattern is dominated by a strong zonal harmonic wave 1.

We began our investigation of the association between the SSC and the stratosphere in 1987 and published the first correlation map of the annual mean 30 hPa heights with the SSC in 1990 (van Loon and Labitzke, 1990). This map is reproduced in Fig. 2; the similarity with Fig. 1 is striking and proves that the additional eight years since 1987 have not only not changed the pattern but have enhanced its statistical significance. The curves in Fig. 3 further illuminate this point: They are from a location, 30°N 150°W, where the correlations in Fig. 1 exceed 0.7 and the number of years of 30 hPa heights at the location is the same as that used in the map in Fig. 1. The thin solid line consists of the annual 30 hPa heights and the heavy solid line is made up of the three-year running means of annual 30 hPa heights. It is obvious that the time series of the heights follows the curve of the SSC in the same diagram, and also that in the last solar cycle the fit between solar and stratospheric variation is as good as in the previous cycles. Following the argument in von Storch (1995), the last eight years may be considered an independent sample and their eminent fit to the solar curve can be regarded as a confirmation of the reality of the association between the stratospheric and solar variation. We have previously, out of caution, labelled the oscillation in Fig. 3 the Ten-to-Twelve-Year Oscillation (TTO), but in view of the continued clear association with the SSC during the past cycle, we no longer use the name TTO.

The height of a constant pressure surface, such as the 30 hPa level, depends on the pressure at sea level and the cumulative effect of the temperature in the column between sea level and the pressure surface. The former influence is unimportant with respect to the solar signal (Labitzke and van Loon, 1992: Figs. 3 and 5; see also Fig. 7 in the present paper), whereas the interannual temperature changes in the column below 30 hPa do have a decadal component in-phase with the SSC (Labitzke and van
Loon, 1992; van Loon and Labitzke 1994). Figure 4 shows the temperature difference between maxima and minima in the SSC for four stations in the Pacific Ocean near the largest correlations in Fig. 1. The difference is between the means of two years in solar maxima and two in solar minima in each solar cycle for three to four cycles. At each station the troposphere is warmer and the tropopause levels colder in the solar maxima. This characteristic distribution of the temperature difference will be discussed further below; it suffices here to say that the large correlations between the 30 hPa height and the SSC in Fig. 1 in these instances are overwhelmingly associated with the tropospheric temperature.

3. THE SOLAR SIGNAL THROUGH THE YEAR

The pattern of correlation between two-monthly mean 30 hPa heights and the SSC (Fig. 5) is the same through the year. The correlation values are smallest in winter, grow to a peak in summer, and wane in autumn. We shall discuss the low correlations in winter in a later section and concentrate below on the high correlations in summer.

We assume that the increase in correlation from winter to summer and the decrease through autumn are associated with the north- and southward movement of the sun, and that this in itself is an indication of interaction between the lower stratosphere and the SSC. The fact that the largest correlations are in the latitudes near 30°N suggest an association with the sinking branch of the Hadley circulation in the troposphere. The position of the Hadley circulation can be gauged from Fig. 6, which shows the long term mean, zonally averaged relative humidity (Hurrell et al., 1997): where the humidity is low near 30° lat. in the summer hemisphere the air sinks in the Hadley circulation; and it rises in the equatorial latitudes with high humidity.

We present some circumstantial evidence of a connection between the SSC and the lower stratosphere-upper troposphere in the following: Figure 7a shows the average temperature difference in summer between SSC maximum and minimum in four maxima and four minima of the cycle, using two years in each extreme, from the Caribbean to the western Pacific Ocean. At all the stations the temperature in the troposphere is higher in the solar maxima. At the tropopause levels the difference between SSC extremes is near zero, and in the stratosphere the difference is again positive. The height difference (Fig. 7b) is zero at the surface, but becomes positive and grows with increasing height in response to the cumulative effect of the temperature, as mentioned in Section 2 about the temperature differences in the annual means in Fig. 4. We ascribe the positive difference at the low latitude station, Truk at 7.5°N 152°E, to a stronger convection in the Intertropical Convergence Zone (that is, the rising branch of the Hadley circulation) in solar maximum than minimum, and the positive difference at the stations at the northern edge of the tropics, for instance Lihue at 22°N 159°W, to stronger subsidence in the sinking branch of the Hadley circulation at maximum of the SSC. This idea is discussed in more detail and with more data in Labitzke and van Loon (1995).
The 30 hPa constant pressure level is the highest level for which we have long series of data in the lower stratosphere in summer. It is therefore not possible to say how high the strong correlation pattern observed at 100 hPa to 30 hPa extends into the stratosphere. There exists a series of 10 hPa (about 30 km above sea level) analyses for the winter months from the Stratospheric Research Group which can be used to judge the upward extension of the pattern. We show the correlations at 10 hPa for October–November in Fig. 8; it is evident that in these months the correlation pattern at 10 hPa is appreciably weaker than 6 km farther down, and statistically insignificant. It is thus possible that the strongest correlation between heights of constant pressure levels in the stratosphere and the SSC culminates at levels between 20 km and 25 km. Considering the strong interdependence between the temperature in the troposphere-lower stratosphere and 30 hPa heights—and the weak correlations at 10 hPa—it is therefore possible that the solar influence observed in the lower stratosphere has its origin in the atmosphere below 30 hPa.

We present Fig. 9 as further indication of a relationship between the signal of the SSC in the height of the constant pressure surfaces and the tropospheric temperature. The figure shows the 10.7 cm solar flux together with three-year running means of the annual mean, zonally-averaged 30 hPa height and 700 hPa temperature in the area between 20°N and 40°N. The results discussed in the following change little if one uses the entire area between 20°N and the North Pole, because the solar signal is overwhelmingly strong in the subtropics (see Fig. 1).

The connection between the solar curve and the 700 hPa temperature and 30 hPa height is clear. The correlation coefficients—computed with the single years and not with the three-year running means in Fig. 9—are large, 0.50 and 0.68 respectively, and quite significant (Table 1); and the same is true for the 50 hPa level which is not shown in Fig. 9. We have previously published a similar curve for the surface-air temperature (Labitzke and van Loon, 1994a,b), and the correlation coefficient with the solar flux was 0.52 for the zonally-averaged, area-weighted surface air temperatures between 10°N and the North Pole.

Apparently, the solar cycle component in the temperatures and heights in Fig. 9 is superposed on a long-term trend similar to the long-term trend for the same period which was attributed to solar changes by Friis-Christensen and Lassen (1991) and Lassen and Friis-Christensen (1995).

Table 1. Correlations between the single year annual means of 700 hPa temperature (T700), Solar flux, 30 hPa heights (H30), and 50 hPa heights (H50). The percentages are the levels above which the correlations are significant.

<table>
<thead>
<tr>
<th>Correlation</th>
<th>p-value</th>
</tr>
</thead>
<tbody>
<tr>
<td>r(H30, Sun)</td>
<td>0.68 (0.1%)</td>
</tr>
<tr>
<td>r(H50, Sun)</td>
<td>0.59 (0.1%)</td>
</tr>
<tr>
<td>r(T700, Sun)</td>
<td>0.50 (1.0%)</td>
</tr>
<tr>
<td>r(T700, H30)</td>
<td>0.49 (1.0%)</td>
</tr>
<tr>
<td>r(T700, H50)</td>
<td>0.67 (0.1%)</td>
</tr>
</tbody>
</table>

In addition to the solar signal in the tropical-subtropical temperatures in the troposphere described above, it has been brought to our attention (White et al., 1996, private communication) that a
strong solar signal has been found in the mixed layer of the global tropical oceans. This corroborates the idea of a modulation of the Hadley circulation on the time scale of the SSC.

4. THE SOLAR SIGNAL IN WINTER

The correlation between the 30 hPa heights and the SSC are small in winter (Fig. 5)—especially in late winter (February)—although the pattern is the same as in the rest of the year. However, Labitzke (1987) found that by dividing the time series at the North Pole according to the phase of the Quasi-Biennial Oscillation (QBO) in the winds of the equatorial lower stratosphere, the two series were well correlated with the SSC, albeit the correlations were of opposite sign: Positive in the west years of the QBO and negative in the east years. Further work showed that the correlation map as a whole in the east years looked like the maps of the rest of the year with large correlations in the subtropics on the Atlantic-Pacific side of the hemisphere of opposite sign to those at the Pole (Fig. 10a). This distribution reflects the dominant teleconnection pattern in the stratosphere (Shea et al. 1992): when heights or temperatures are above normal in the lower latitudes they tend to be below normal at higher latitudes, and conversely; but this does not seem to function in the west years. There is little correlation with the SSC in the west years outside the polar region (Fig. 10b).

The reason for the positive correlation with the SSC at the North Pole in the west phase is, in the first approximation, that midwinter warmings in high latitudes, with accompanying breakdown and reversal of the cyclonic polar vortex, in the west years so far have happened only in maxima of the SSC, whereas the vortex tends to be cold and stable in the solar minima in these years. We illustrate this by means of Fig. 11 which shows the 30 hPa heights at the North Pole in February plotted against the solar flux in (a) the east and (b) the west years of the QBO. In the east years the breakdowns occurred primarily at low solar flux, which is what theory and model experiments predict (Holton and Tan, 1980; Dameris and Ebel, 1990, among others). The opposing signs of correlation in the polar region in Figs. 10 and 11 are characteristic of the lower stratosphere, for instance of the standard pressure levels of 100 hPa, 50 hPa, and 30 hPa; but at the next higher standard pressure level, 10 hPa, the correlations in the west phase are weaker and insignificant and their pattern is different from the lower levels, whereas the pattern in the east phase is the same as at the standard pressure levels below 10 hPa, and although it is weaker it is still significant. We surmise that the weakening of the correlations with height indicates that the solar connection also in winter is associated with the troposphere.

The fact that large correlations with the SSC in winter in the lower polar stratosphere are contingent on the division of the data into the two extremes of the QBO has encouraged critics to suggest that the effect is an artefact of aliasing (e.g., Salby and Shea, 1991) and therefore has a mathematical rather than a physical cause. We have suggested (Labitzke and van Loon, 1993; van Loon and Labitzke, 1994) that the cause should be sought in the different conditions for vertical and horizontal wave propagation in the two phases of the QBO, since in early winter the critical line is near the equator in the west phase and well into the subtropics in the east phase of the QBO. Model
experiments by Balachandran and Rind (1995) and Rind and Balachandran (1995) apparently support
this concept. In addition, model experiments by Haigh (1996) have confirmed that a modulation of the
Hadley circulation is a likely link in the chain of cause and effect of solar forcing. These experiments
take into account the effects of solar ultraviolet variability (see also Kodera et al., 1990), but it has been
suggested that the sun might also influence the atmosphere through energetic particle precipitation (e.g.,
Tinsley and Heelis, 1993).

5. THE SOLAR SIGNAL IN OZONE

The correlations between total column ozone, observed by the Total Ozone Mapping
Spectrometer (TOMS) between 1979 and 1993, are weak in the annual mean and in the extreme
seasons in the equatorial belt, where ozone is produced, and in subpolar latitudes where it is most
plentiful (Fig. 12a). The correlations are strong in the tropics outside the equatorial regions, with the
highest correlations in the summer hemisphere (Labitzke and van Loon, 1996).

There are some marked differences between the two solstice seasons: in the months about the
boreal solstice, Fig. 12b, the correlation maxima are symmetrical with respect to the equator and are
only 20°–25° lat. apart; but in the austral solstice season the two maxima are 45° lat. apart (Fig. 12c).
The correlations in the equatorial belt are smaller by far in the southern than in the northern summer.
We assume that these differences are related to the fact that the equatorial temperature in the ozone
layer has a comparatively large annual range with the lowest temperature in the southern summer (see,
e.g., van Loon and Jenne 1970; van Loon et al., 1973).

As regards the question why the SSC and ozone are well correlated in the outer tropics-
subtropics and nowhere else (Fig. 12), we suggest that the higher pressure in the subtropics at maxima
in the SSC, which is evident in Figs. 1–3, restricts the poleward flow through the subtropics—causing
convergence—and thus creates an abundance of ozone in the tropics outside the equatorial belt relative
to the solar minima.

6. CONCLUSION

This review deals with an oscillation in the lower stratosphere-upper troposphere which the
records allow us to follow during four of its periods. During that time it has been in-phase with the 11-
year solar cycle. The basic pattern of correlation between the solar cycle and the stratospheric heights at
constant pressure levels is one in which the highest correlations are found south of 45°N, with a zonal
asymmetry such that the western, ocean-dominated parts of the Northern Hemisphere contains the
largest correlations. The association with the sunspot cycle appears to culminate in the lower
stratosphere at levels below 10 hPa (30 km).
The basic pattern of correlation is found in all months but is strongest in summer. It is associated with corresponding changes in the temperature of the middle and upper troposphere in the tropics and subtropics, and the temperature variations in these regions suggest that the interannual variation of the Hadley circulation contains a component on the time scale of the solar cycle.

The large zonal asymmetry in the basic correlation pattern (Fig. 1) is owing partly to the fact that the size of the correlations is more variable over the subtropical parts of Central Asia than elsewhere in the same latitudes; and partly to the fact that whereas the position of the highest correlations with the solar cycle is comparatively stable through the available period over the oceans, the highest correlations move meridionally during the period over the continents. This movement is especially marked in the western Pacific-China area where it covers as much as 15° lat. (Fig. 13).

The correlations with the 11-year sunspot cycle are smallest in January-February. The smallness is due to the fact that the Quasi-Biennial Oscillation modulates the solar signal.

The pattern of correlation between total column ozone and the sunspot cycle is such that the correlations reach their highest values between 5° lat. and 30° lat. on either hemisphere, and in these regions the highest correlations are found in the summer hemisphere. The position of the best correlation between the gas and the solar cycle in the outer tropics-subtropics suggests that the solar influence is not entirely radiative but that it also affects the poleward transport of ozone.

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Figure 1. Correlations between the annual mean 30 hPa height and the 10.7 cm solar flux, 1958–1995. In the shaded areas the statistical significance is better than 0.1%. From van Loon and Labitzke (1990), updated.
Figure 2. As Fig. 1, but for 1958–1987. From van Loon and Labitzke (1990).
Figure 3. Time series of the 10.7 cm solar flux (dotted line), the annual mean 30 hPa height at 30°N 150°W, in gpdam, (thin solid line), and of the three-year running means of the latter curve (heavy solid line). From Labitzke and van Loon (1995) updated.
Figure 4. The annual mean temperature difference (°C) between maxima and minima in the 11-year sunspot cycle at San Diego (33°N 117°W), Lihue (22°N 159°W), Guam (14°N 145°E), and Truk (8°N 152°E). From Labitzke and van Loon (1995), modified.
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Figure 10. Correlation between the 10.7 cm solar flux and the 30 hPa height in February in (a) the east years and (b) west years of the QBO shown in Fig. 11, using the average wind in January-February at the 40–50 hPa levels on the Equator to define the QBO.

Figure 11. The 30-hPa heights at the North Pole in February plotted against the 10.7 cm solar flux in (a) the east years and (b) the west years of the QBO. The asterisks denote winters with major warmings and reversals of the polar cyclonic vortex.
Figure 12. Correlations between 15 years (1979–1993) of TOMS total ozone measurements and the 10.7 cm solar flux: (a) annual values, (b) the mean of May, June, July, August, and (c) the mean of November, December, January, February. From Labitzke and van Loon (1996). In the shaded areas the statistical significance is better than 5%.
Figure 13. (a) Correlations between the 10.7 cm solar flux and annual mean 30-hPa heights, 1958–1977. (b) as (a) but for 1976–1995. The heavy lines connect the points of highest correlations, and the line on (a) is shown on (b) too. In the shaded areas the statistical significance is better than 0.1%.
1. INTRODUCTION

Variability of the tropospheric circulation over the South Pacific Ocean is known to be closely linked to fluctuations in sea surface temperature (SST) over the tropical Pacific. In general, when SST is above normal during a warm event in the Southern Oscillation (SO), the Pacific trade wind field tends to be weak, but the mean upper level South Pacific trough and associated subtropical jet is stronger than normal (van Loon and Shea, 1985; 1987). Strong trades and a weak jet are characteristic of cold events. These circulation patterns are also linked to hemispheric-wide anomalies, which in some seasons appear as wave trains originating over the subtropical Pacific (e.g. van Loon and Madden, 1981; Karoly, 1989). The vertical structure of interannual circulation variability over the extratropics of both hemispheres is generally equivalent barotropic, such that anomalies of one sign persist through the depth of the troposphere (Kidson, 1988) and even into the stratosphere (van Loon and Labitzke, 1987; Labitzke and van Loon, 1989).

One signal which is particularly well-developed during warm events in the SO is the so-called Pacific South America (PSA) pattern (Mo and Ghil, 1987). This pattern has alternating signs of pressure or height anomalies stretching from the central South Pacific eastward over southern South America and on into the South Atlantic. The PSA pattern is also a frequently seen intraseasonal circulation mode, thus is not necessarily excited by tropical forcing, although it may have some preferred relationship to the Madden-Julian Oscillation (Mo and Higgins, 1997, this volume).

In this paper we examine the structure of the previously documented warm event signals over the Southern Hemisphere (SH) using a longer data set from the National Centers for Environmental Prediction and National Center for Atmospheric Research (NCEP/NCAR) Reanalysis project. We also compare the warm event signals to those observed during the cold events, or opposite phase of the SO. An attempt is made to account for at least the Pacific portion of the observed patterns dynamically, by a comparison of rudimentary vorticity budgets of the 1982/83 warm event and 1988/89 cold event.

2. DATA AND METHODOLOGY

For this study, NCEP/NCAR Reanalysis data are used to represent the large-scale circulation. The reanalysis data provide a consistently derived set that is not subject to operational constraints or changes in the assimilation system (Kalnay et al., 1996). Four times daily grids were averaged to produce a daily reanalysis set for this study, utilizing winds and temperature on 14 pressure levels from
1000 mb to 50 mb. Also used were Outgoing Longwave Radiation (OLR) data as a proxy for deep convection during the case studies examined. Both of these data sets are on a 2.5 degree latitude/longitude grid. The 24 year study period spans from January, 1973 to February, 1996.

Since we are dealing with interannual variability and seasonal dependence, the data have been averaged into standard seasonal means (e.g. December-February (DJF), March-May (MAM), etc.) for the purposes of compositing. Composite anomalies were calculated for JJA and SON of the years during which the events started (year 0), through DJF and MAM of the following year. The warm events year 0's used in this compositing are 1972, 1976, 1982, 1986, and 1991, and the cold events are 1973, 1975, 1983, 1988, and 1995. The June through December anomalies during 1972 were not available for these composites, but the January through May fields were used for the DJF and MAM year+1 calculations.

3. WARM AND COLD EVENT COMPOSITES

a. Sea Level Pressure and 500 mb Heights

To illustrate the anomalous circulation pattern over the SH during warm events, we reproduce in Figs. 1 and 2 the composite 500-mb height and sea level pressure anomalies. Although only 5 events are used, the patterns are robust in that they agree well with those for individual events in the earlier historical record (not shown), and with those obtained by Karoly (1989) using a smaller sample.

Sea level pressure and 500-mb height anomaly patterns are similar over the extratropics, although a slight westward tilt with height is seen in many regions. The vertical structure is generally consistent with the equivalent barotropic nature typical of extratropical interannual circulation anomalies. At low latitudes, however, the sea level pressure anomalies tend to retain the same sign as those in the subtropics, whereas there is most often a sign reversal between the subtropics and equator at 500 mb in the Pacific. This reflects the tendency for first baroclinic mode vertical structures at low latitudes associated with diabatic heating anomalies.

In JJA (Figs. 1a and 2a), a region of anomalously low pressure extends along 40°S over the Pacific from the Tasman Sea eastward to 90°W. Poleward of this trough a wave train of anomalies stretches around Antarctica, starting with a strong positive anomaly over the Bellingshausen Sea and extending across the southernmost Atlantic into the Indian Ocean. This pattern has a strong projection onto the PSA mode described in this volume by Mo and Higgins, as a preferred mode of interannual variability. These features were also discussed by Karoly (1989).

The Pacific trough anomaly strengthens significantly into southern spring (Figs. 1b and 2b), and although the Bellingshausen Sea anomaly retains its strength, the downstream portion of the PSA mode has broken down and the development of a strong zonally symmetric component is apparent. This strengthening of the anomalous Pacific trough results in a suppression of the usual seasonal cycle over the Pacific, since in the mean pressure normally rises over the Pacific into the southern spring (van Loon, 1984).
The zonal symmetry of the anomalies strengthens into summer (Figs. 1c and 2c), with negative anomalies centered on 50°S ringing above normal heights over and surrounding Antarctica. The Pacific trough is still strong, as are troughs over the South Atlantic and South Indian Ocean, giving a significant wave number three component to the flow as well. This is also evident in the high heights over Antarctica which protrude equatorward between the oceanic troughs. By the following MAM (Figs. 1d and 2d), the pattern begins to break down, although pressure remains high over Antarctica and zonal asymmetries are still pronounced.

During the cold event phase of the SO, the pressure anomalies are in many regions opposite to those during warm events (Figs. 3 and 4). The most robust feature is the weak Pacific trough in all seasons. However, the "inverse PSA" is not very evident during JJA in Figs. 3a and 4a, with a negative anomaly located southeast of New Zealand, well west of its positive counterpart in Figs. 1a and 2a. An inverse pattern between warm and cold events is much more evident in SON (Figs. 3b and 4b), when the pattern over the Pacific sector is the opposite of that in Figs. 1b and 2b, and positive perturbations cover Antarctica. As in warm events, the zonally symmetric portion of the signal is maximized during DJF (Figs. 3c and 4c). Again there is a wavenumber three component to this pattern, although it is much weaker and has a different longitudinal phasing than the patterns in Figs. 1c and 2c. This signal persists more strongly into MAM (Figs. 3d and 4d) than it does during warm events (compare Figs. 1d and 2d). The lack of a perfect reversal in the pressure patterns between warm and cold events, though partly due to sampling, may also be due to a real non-linearity in the atmospheric response to the presence versus the absence of large-scale convection over the tropical Pacific.

b. 200 mb Zonal Wind

It is well-established that enhanced rainfall over the central and eastern equatorial Pacific accompanies warm events, as manifested by an equatorward shift in the ITCZ and SPCZ (e.g. Kiladis and van Loon, 1988). Opposite anomalies are observed during cold events. In his pioneering work on the SO, Bjerknes (1969) noted a persistently strong Aleutian low during warm events, resulting in stronger than normal westerlies over the North Pacific, especially during winter. He reasoned that an enhanced Hadley circulation can maintain a strong subtropical jet stream by the large scale meridional flux of westerly momentum.

Figure 5 shows zonal wind anomalies at 200 mb during SON and DJF for warm and cold events, calculated from the same sample of years as used in Figs. 1 and 2. Anomalies are most pronounced over the Pacific sector, consistent with the region of largest forcing by diabatic heating displacements, as envisioned by Bjerknes. The patterns are remarkably symmetric about the equator over the Pacific. In this sector the warm event patterns during both SON and DJF have easterly anomalies along the equator and along 40°N and 40°S latitude (Figs. 5a and 5b). Enhanced westerlies are observed in the subtropics of both hemispheres, and also along the coast of Antarctica in phase with the implied 500 mb and sea level pressure geostrophic winds of Figs. 1b, 1c, 2b, and 2c. Over the Pacific, at least, the cold event anomalies (Figs. 5c and 5d) are virtually opposite to those observed during warm events.
These zonal wind signals essentially represent an equatorward movement of the storm track and strengthening of the Pacific jets during warm events, and poleward storm track migration during cold events. Over other longitudinal sectors the signals are weaker, although the symmetry about the equator is also evident over the Atlantic, and the anomalies there are generally of opposite sign for a given latitude to those over the Pacific. As was seen in Figs. 1 through 4, the Indian Ocean region shows less consistency, although there is a tendency for strong (weak) westerlies centered on 15°S during warm (cold) events.

c. Zonally Symmetric Signals

Karoly (1989) analyzed the zonal mean signals in wind, temperature and pressure over the SH associated with warm events from the surface up to 100 mb, and found primarily equivalent barotropic structures from polar regions well into the subtropics. Figure 6 shows a similar analysis from 80°N to 80°S for DJF zonal mean zonal wind during warm events (Fig. 6a), cold events (Fig. 6b), and for temperature during warm events (Fig. 6c). Comparison of Figs. 6a and 6b with Figs. 5b and 5d reveals that the zonal mean signatures are dominated by the large anomalies over the Pacific sector. Poleward of about 15°N and 15°S, the anomalies are of the same sign from the surface all the way up to 50 mb and evidently beyond, with little poleward tilt with height apparent. During warm events (Fig. 6a) strengthened westerly flow is found centered on 30°N and 40°S, extending to within 10° of the equator and out to 45° latitude in both hemispheres. These westerlies peak at jet level, just above 200 mb. Poleward of the jets deep easterly anomalies dominate the entire column from the surface into the stratosphere, while westerlies are again found below 150 mb over Antarctica. These signals peak at a lower level than the westerly anomalies in the subtropics, but continue to increase in strength with height to at least 50 mb at around 60°N. This weaker than normal stratospheric polar vortex in the Northern Hemisphere winter during warm events was documented in detail by Quiroz (1983) and van Loon and Labitzke (1987).

Zonal mean zonal winds during cold events are not simply the inverse of the warm event signals. The most notable deviation from this anti-symmetry concerns the fact that easterly anomalies during cold events are centered on 15°N and 20°S in Fig. 6b, rather than at the mean latitudinal position of the jet cores. This latitudinal shift is also noticeable by comparing Figs. 5b and 5d, where anomalies in the Pacific sector appear to account for the bulk of the difference in zonal mean signals. Beyond the subtropics, however, the cold event patterns are more like those during warm events but opposite in sign. The strengthened stratospheric polar vortex in the Northern Hemisphere and stronger westerlies over the Southern Ocean at around 60°S are robust, appearing to some degree in each individual cold event (not shown). This is also the case for the warm event signals.

The zonal mean temperature signal during DJF warm events is shown in Fig. 6c. The warming of the tropical troposphere out to about 20° latitude is a well-documented feature of warm events (e.g. Newell and Weare, 1976). This pattern can be related to the warming of SST and concomitant increase in sensible and latent heat release over the central and eastern tropical Pacific, along with increases in
shortwave radiation over the drier than normal Eastern Hemisphere during warm events (e.g. Kiladis and Diaz, 1989). The observed warming of the Northern Hemisphere stratosphere and cooling of the tropical stratosphere is discussed in detail by van Loon and Labitzke (1983). The amplitude of these stratospheric signals is quite large compared to those in the troposphere. Since they are far removed from the surface boundary forcing, they must have a dynamical origin, although this is yet to be investigated. The negative temperature anomalies in the mid- to lower troposphere of the mid latitudes reflect the upward extension of surface zonal mean temperature anomalies at these latitudes. At least part of this signal may be related to westerly wind stress forcing of the ocean surface, which would produce lower than normal SST due to latent and sensible heat fluxes (e.g. Lau and Nath, 1996). These storm track anomalies exhibit a slight poleward tilt with height, and are of the right sense to produce a thermal wind balance of the zonal winds in Fig. 6a away from the equator.

4. MASS CIRCULATIONS DURING WARM AND COLD EVENTS

The zonal wind features described above were observed during the 1987–1989 ENSO cycle (Vincent et al., 1995), and were related to anomalies in the vorticity budget by Rasmusson and Mo (1993). Sardeshmukh and Hoskins (1985) also showed that the large anomalies in tropospheric zonal wind during the 1982/83 warm event could be accounted for by changes in the forcing of the rotational flow by the divergent circulation, as originally implied by Bjerknes (1969). Here we examine a portion of the vorticity budget, namely the forcing due to the divergent wind advection, during 1982/83 and 1988/89, to see if these are consistent with the idea that anomalous convection over the tropical Pacific leads to changes in the local Hadley circulation which are directly linked to anomalies in flow at jet level.

A vorticity budget at 200 mb for DJF was calculated for the 1982/83 warm event and 1988/89 cold event, using the daily averaged NCEP/NCAR Reanalysis data. The budget residual, which also includes the effects due to vertical advection and the twisting and tilting terms (which were not calculated) was found to be small relative to the leading terms in the balance over the tropical Pacific Ocean (not shown). To illustrate the contrasting stationary forcing during warm and cold events, we present here the forcing due to the advection of absolute vorticity by the mean divergent wind, \(-\mathbf{v} \cdot \nabla \zeta\), where \(\mathbf{v}\) is equal to the divergent wind and \(\zeta\) is the absolute vorticity.

Figure 7a shows contours of the advection of absolute vorticity by the divergent wind at 200 mb for the DJF 1982/83 mean state. Also shown is the mean divergent wind, with OLR less than 240 Wm\(^{-2}\) outlined by the shaded region. Divergent outflow from the region of convection is evident over the central and eastern Pacific, leading to the generation of anticyclonic (negative in the Northern Hemisphere and positive in the Southern Hemisphere) vorticity over large regions of the Pacific subtropics. This forcing would be associated with an easterly acceleration along the equator and westerly accelerations poleward of the anticyclonic forcing, in precisely the regions where these zonal wind anomalies were observed during that season.
The fact that the SH has more substantial summer anomalies than in Northern Hemisphere summer is likely related to the fact that the forcing of the Rossby wave sources in the subtropics by the divergent wind (see Sardeshmukh and Hoskins, 1988) is stronger in the former, since the westerlies and thus the associated absolute vorticity gradient is much closer to the upper level divergent wind forced by the tropical convection. However, it is still necessary to take into account the effect of feedback by the transients, especially within the storm track (e.g. Trenberth, 1984; Karoly, 1990), in order to completely account for the strengthened subtropical jets (Held et al., 1989; Hoerling and Ting, 1994).

As a contrast, Fig. 7b shows the same calculations for the DJF 1988/89 cold event. In this case the most active convection was located farther west over Indonesia and in the SPCZ, and there was a notable lack of low OLR values over the central and eastern equatorial Pacific. Accordingly, the divergent outflow along with the stationary forcing of anticyclonic vorticity was absent over the subtropical Pacific, leading to weaker than normal jet flow over the central part of the basin.

5. SUMMARY AND CONCLUSIONS

Substantial circulation variability is observed over both hemispheres during extremes in the SO. These anomalies reach their largest amplitude over the Pacific sector, closest to the region of greatest forcing by SST and associated convective anomalies. In general, during warm events enhanced convection over the central and eastern equatorial Pacific drives a stronger than normal divergent circulation which is associated with anticyclonic forcing of the subtropical circulation. This leads to easterly anomalies along the equator and stronger jet flow in the upper troposphere over that sector. This is in turn associated with a reorganization of the storm track and associated transient forcing, amplifying the response.

The Pacific anomalies are tied to global scale teleconnections in the circulation. Over the Southern Hemisphere during winter a wave train originating in the Pacific and extending across the southern oceans is observed, the so-called PSA pattern. During summer, however, the anomalies are much more zonally-symmetric, although some projection onto zonal wavenumber three is evident. The seasonal dependence of these perturbations is likely related to the response to forcing in different basic state circulations between summer and winter, although this has not yet been thoroughly investigated for the SH. The fact that the SH has more substantial summer anomalies than the Northern Hemisphere is likely related to the fact that the westerlies and associated absolute vorticity gradient is closer to the upper level divergent wind forced by tropical convection in the SH.

The circulation perturbations observed during cold events are in many regions opposite in sign to those during warm events. The lack of a perfect reversal in the pressure patterns between warm and cold events, though partly due to sampling, may also be due to a real non-linearity in the atmospheric response to the presence versus the absence of large-scale convection over the tropical Pacific.

The vertical structure of the circulation anomalies during extremes in the SO is nearly equivalent
barotropic, having in most regions the same sign from the surface well into the stratosphere. In the equatorial Pacific, however, there is a sign reversal in the pressure perturbations between 500 mb and the surface, reflecting the first baroclinic mode structure of circulations associated with diabatic heating. Thus while upper level easterly (westerly) anomalies are observed along the equator during warm (cold) events, the surface flow is characterized by weaker (stronger) than normal easterly trade wind flow.

Zonal mean circulations during extremes of the SO are dominated by the signals over the Pacific. Patterns over the Atlantic generally oppose the zonal mean at a given latitude, which weakens the zonal mean signal somewhat. Zonal wind anomalies are consistent with the thermal wind associated with the zonal mean temperature perturbations.

In summary, the convective reorganization during extremes in the SO appears to force a modified basic state which the transients appear to amplify within the storm track regions. This leads to the characteristic zonal wind anomalies observed over the Pacific basin. Rossby wave propagation is one likely mechanism which can explain the observed global teleconnections associated with the SO, although radiative forcing in the tropics and driving of oceanic latent and sensible heat fluxes and associated feedbacks may also need to be taken into account to explain some of the far-field signals. This is an area in which observational work in conjunction with modelling studies could prove to be quite fruitful.

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Figure 1. Composite warm event 500-mb height anomalies for a) June-August. b) September-November. c) December-February. d) March-May. Contour interval is 1 dam, with negative contours dashed.
Figure 2. Composite warm event sea level pressure anomalies for a) June-August. b) September-November. c) December-February. d) March-May. Contour interval is 1 mb, with negative contours dashed.
Figure 3. Composite cold event 500-mb height anomalies for a) June-August. b) September-November. c) December-February. d) March-May. Contour interval is 1 dam, with negative contours dashed.
Figure 4. Composite cold event sea level pressure anomalies for a) June-August. b) September-November. c) December-February. d) March-May. Contour interval is 1 mb, with negative contours dashed.
Figure 5. Composite 200-mb zonal wind anomalies for a) September-November warm events. b) December-February warm events. c) September-November cold events. d) December-February cold events. Contour interval is 1 m s$^{-1}$, with negative contours dashed. Positive anomalies greater than 3 m s$^{-1}$ are shaded, negative anomalies less than -3 m s$^{-1}$ are hatched.
Figure 6. Latitude-pressure diagrams of composite zonal mean a) December through February zonal wind during warm events. Contour interval is 0.4 m s$^{-1}$. b) December through February zonal wind during cold events. Contour interval is 0.4 m s$^{-1}$. c) December through February temperature during warm events. Contour interval is 0.2°K. Negative contours are dashed.
Figure 7. Forcing due to the advection of absolute vorticity by the mean divergent wind at 200 mb, \(-\vec{v}_x \cdot \nabla \vec{z}\), for (a) December through February, 1982/83. (b) December through February, 1988/89. Contour interval is \(1 \times 10^{-11} \, \text{s}^{-1}\), with negative contours dashed. Shading denotes regions of OLR less than 240 W m\(^{-2}\). Also shown is the mean divergent wind at 200 mb, largest vectors are about 5 m s\(^{-1}\).
The Southern Hemisphere Climatology Project An Historical Perspective

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The project to prepare a surface and upper air climatology of the Southern Hemisphere was a large project that was mainly accomplished during 1967–1971. The analysis project used surface data (land & ships) and rawinsonde data. The data coverage was not very good in the Southern Hemisphere, but we could mostly compensate for this by using very careful analysis procedures. We will focus on this project, but will first trace some earlier events in the career of Harry van Loon.

Harry van Loon grew up in Denmark and was there until 1951. He was married in 1949, and then was at MIT from July 1951–April 1954. While at MIT he worked on a Southern Hemisphere analysis project that was started by Hurd Willett. Some people from South Africa visited MIT and asked Harry if he might be interested in working on Southern Hemisphere analyses in South Africa.

1. WORK FOR NINE YEARS IN SOUTH AFRICA

Harry lived in South Africa from July 1954 to June 1963. He worked on some climatology projects and also had forecast duties. South Africa started drawing daily maps of Southern Hemisphere sea level pressure from 1950-on. Harry drew these maps starting with data from about December 1954. The maps were completed in delayed time so that past and future continuity could be used to obtain better analyses in areas with poor data coverage (which was most of the hemisphere). The International Geophysical Year (IGY) was a period when more upper air stations started in the Southern Hemisphere. Daily Southern Hemisphere maps at the surface and 500 mb were drawn for this period (SLP: June 57–Dec 58, 19 months; and 500 mb: July 57–Dec 58, 18 months). These maps also exist in digital form at NCAR. van Loon and Taljaard drew the first nine months of the IGY daily maps before Harry left to work at NCAR in July 1963 (Taljaard and van Loon, 1964). The next nine months of maps were completed by Taljaard and another person.

2. SOME EARLY CLIMATE WORK IN SOUTH AFRICA

While in South Africa, Harry made an analysis (in about 1960) of 500 mb heights and sea level pressure for the mid-season months for the Southern Hemisphere and these were later published. This was an early effort toward a climatology, and would later help to kindle the desire for the larger project.

When he came to NCAR in July 1963, Harry worked on projects relating to the nine months of IGY maps that he and Taljaard had already analyzed. About mid-1966, Harry got the idea of starting a larger Southern Hemisphere climatology project. This would be a large task, even to analyze the maps,
so he soon asked Taljaard if he wanted to join the effort. Taljaard was very interested so things started rolling. He visited NCAR for 18 months starting early 1967.

Roy Jenne was brought into the project to help with data and computing. The team at NCAR found that Harold Crutcher at NCDC (Asheville) had collected a lot of monthly rawinsonde data and was planning to carry out an analysis project. The two projects were joined.

3. UPPER AIR OBSERVATIONS FOR THE PROJECT

The project had monthly rawinsonde data (in digital form) available through October 1966, a total of 52,868 monthly reports. But only 11,261 of these were for stations south of 5°N latitude. Since data for several stations were still completely missing from the record, we obtained daily raob data from the USAF (39,288 raobs). Stations included were Ascension Island; four stations in Brazil (Belem, Sao Paulo, Fernando Noronha, Porto Alegre); and Salinas, Ecuador. The daily reports were checked and monthly statistics were calculated. Finally, all of the monthly reports were used to calculate long term statistics for each level and month. These were plotted on the maps. A few stations were still missing but published sources could be found. Wind statistics were not available on some of the early monthly reports, so some published sources were used to complete the record. This is documented in Volume 1 of the atlas (Taljaard et al., 1969).

4. THE ANALYSIS OF THE SURFACE DATA

The analysis of surface pressure and temperature required special procedure because some regions had only a few historical observations. Procedures were first used to insure time and space continuity of surface temperature and sea level pressure over the ocean areas. Grid point data were read from the U.S. Navy Marine Climatic Atlas series for selected latitudes in order to construct curves as shown in Fig. 1. Data for Orcadas provided guidance in drawing the curves showing the annual march of temperature at the other grid points at this longitude. After the curves were drawn, points were read from them to use in drawing the monthly maps.

5. THE LEVELS THAT WERE ANALYZED

The atlas has data for seven levels which are the surface, 850, 700, 500, 300, 200, and 100 mb. The basic analyses were for the seven levels for the four mid-seasonal months of January, April, July, and October. For the other eight months, analyses were only made at sea level, 500 mb, and 200 mb (see Table 1). Later, the other eight 100 mb height maps were also hand-drawn by Crutcher at Asheville. The computer derivation of the remaining intermediate month grids and of the various derived grids such as geostrophic winds was discussed in Jenne et al., 1974. The archive tapes have a complete set of
data for every month, and they include geostrophic winds that were calculated using the geopotential heights.

<table>
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<th>Table 1: Analyses in the Climate Atlas</th>
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<tr>
<td>Sea level Pressure</td>
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<td>Surf Temp, Dew Pt. (DP)</td>
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<td>850 mb Height, Temp, DP</td>
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<td>700 mb Height, Temp, DP</td>
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<td>100 mb Height, Temp</td>
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6. METHODS FOR PREPARING THE ANALYSES

The analysis procedures started with maps of sea level pressure and surface virtual temperature. To obtain a thickness for 1000/850 mb, 1000/700 mb, or 1000/500 mb, it was noted that the stability between 35°S and 60°S is nearly constant (approximately 0.62°C per 100 geopotential meters) and that the errors at 500 mb would be acceptable (within ± 20 gpm of observations). From the stability assumptions, a thickness field was calculated that could be added to the 1000 mb chart (derived from sea level pressure). This would provide a good guess for the 500 mb heights (or similar for 850 or 700 mb). Then the reported data for 500 mb heights at stations was used to modify the 500 mb heights based only on thickness. Similar procedures were used to prepare the analyses at 300, 200, and 100 mb.

Upper air temperature analyses that were consistent with the heights were desired. Therefore, first guess virtual temperatures were calculated from the heights. These were modified by using station data.

7. DATA FLOW AND COMPUTING

The maps were drawn by van Loon and Taljaard at NCAR. They were sent to Crutcher at NCDC, where grid points were read each 5° Lat-Lon. The numerical data was sent to Jenne at NCAR. Programs were used to detect possible errors in the grid points. A light smoother was used to filter out the small scale noise that is always present in human estimates of grid points. Please recall that data for a number of levels were only analyzed for four months in a year (not twelve). Computer methods were used to derive a complete set of data. Many products were calculated such as zonal means, geostrophic winds, temperature-dew point spread, precipitable water, cross sections, and motion pictures. A lot of output was generated on microfilm. The computer methods and the available output was described in Jenne, et al., 1974.
Table 2: Four Atlases of the Climate of the Southern Hemisphere

Four atlases were prepared by Taljaard, van Loon, Crutcher and Jenne and were published during 1969-1971. The title of each starts with “Climate of the Upper Air (Part 1, Southern Hemisphere).” The content of each atlas is summarized below.

- **Volume I** Temperatures, dew points, and heights at selected pressure levels (Sep 1969)
- **Volume II** Zonal geostrophic winds (1971)
- **Volume III** Vector mean geostrophic winds. Isogon and Isotach Analyses (May 1971)
- **Volume IV** Selected Meridional Cross Sections of temperature, dew point, and height (June 1971)

8. THE AMS MONOGRAPH: METEOROLOGY OF THE SOUTHERN HEMISPHERE (1972)

The atlas project resulted in a data base of maps and digital data that could be used for many other research projects. One collection of this research was in the AMS monograph of the Southern Hemisphere (Newton, ed., 1972). Some of the contents of this monograph are outlined in Table 3. Taljaard visited NCAR for nine months in about 1970-71 to help work on the monograph.

Table 3: Overview of the Contents of the AMS Monograph (Southern Hemisphere)

1. **Physical features of the Southern Hemisphere:** Topography, pack ice, SST (Taljaard)
2. **Radiation budget of the Southern Hemisphere:** Radiation model, clouds, water vapor, radiation budget (Sasamori, London, Hoyt)
3. **Temperature, pressure and winds in the Southern Hemisphere:** Annual and half yearly waves, cross sections, variability (van Loon)
4. **Cloudiness and precipitation in Southern Hemisphere:** Clouds, humidity, precipitation (van Loon)
5. **The Stratosphere in the Southern Hemisphere:** Means for seasons, mid-winter warmings, QBO, yearly and half-yearly waves, (Labitzke and van Loon)
6. **Synoptic meteorology of the Southern Hemisphere:** Historical survey, observation networks and analysis, air masses, fronts, cyclones, tropical weather systems (Taljaard)
7. **Southern Hemisphere General Circulation (and global energy and momentum):** Angular momentum, eddy transfer, water balance, energy balance (Newton)

9. A SPECIAL ISSUE OF NOTOS

The annual issue of NOTOS (Vol. 17, 1968) was printed about Nov 1969. Volume 17 of NOTOS contains four papers (pages 23–140) about the climate of the Southern Hemisphere that were mainly a
result of the Southern Hemisphere Climatology Project. The subjects were: zonal means, frontal zones, yearly and half-yearly waves (temperature and wind), seasonal range and anomalies for selected levels (by Taljaard, van Loon, Jenne, Crutcher).

10. A MOTION PICTURE OF THE SOUTHERN HEMISPHERE CIRCULATION

We produced a motion picture that shows how different hemispheric maps and cross sections evolve through the year. The film includes some similar charts for the Northern Hemisphere so that comparisons can be made. It even includes 1.7 minutes for each hemisphere to show changes in daily 500 maps for a 395 day period. There is a text that describes the preparation of the film, and describes some of the features to observe (van Loon and Jenne, 1970).

11. USE OF THE ANALYSES

Over 1000 copies of Volume I of the Atlas were printed. This was not enough. It soon became impossible for people to obtain a copy. Many people obtained copies of the digital output. Larry Gates assembled a collection of climatological data that modelers could use. He took surface and upper air data from this climatology (plus our similar data for the Northern Hemisphere based on Crutcher's work) and interpolated it from a 5° x 5° grid to a 5° x 4° grid.

The data still looks very good in modern comparisons. van Loon is preparing comparisons with means of ECMWF daily operational analyses for the 1980's. The two sets of data are very similar. Even the comparison of zonal mean winds in the west wind belt look good (and these winds were derived from the heights). We are very happy that this climatology has stood the test of time, and that it still looks very good!

References


Figure 1. Dashed lines show the annual march of temperatures from the U.S. Navy Marine Atlas for different latitudes at 40°W. Data for the station Orcadas is plotted as a dotted line. The solid lines show an estimate of the annual march of temperatures at each latitude.
Linkages Between Scientific Expertise and Policy Development in the Climate Change Case

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1. INTRODUCTION

The material upon which this paper is based is derived from the results of a multi-year, international project on “Social Learning in the Management of Global Environmental Risks”. The project has looked at the responses of various social actors (scientists, policy-makers, business and industry, the media and non-governmental organizations) to the risks of acid deposition, stratospheric ozone depletion and climatic change. The study looked in detail at the responses in a variety of arenas: the USA, Canada, Mexico, Japan, the Netherlands, the UK, Germany, Hungary and the former Soviet Union, as well as for the European Union as a whole and selected international institutions (e.g. the United Nations Environment Program and the World Meteorological Organization).

By taking a longer term perspective, considering the time period from before the time of the Stockholm Conference in 1972 to recent years and looking at three global environmental issues that have been on national and international policy agendas during that time, it is possible to trace how the relationship between knowledge (monitoring and assessment of risks and response options) and action (goal and strategy formulation, implementation and evaluation) changes systematically with issue evolution.

2. RISK MANAGEMENT FUNCTIONS

Monitoring plays an important, if often neglected, role in dealing with global environmental risks. There are numerous motivations to begin monitoring of emissions, concentrations, and other environmental variables. These motivations are not necessarily related to the management of global environmental risk. Sometimes a monitoring program exists and becomes drawn into the risk management process. In other cases, monitoring begins or is intensified as a result of assessments of global environmental risks. There is, however, a concentration on the natural science components with much less monitoring of human needs, much less monitoring of social responses and this is, of course, mirrored by the fact that risk assessments are also focusing on the center of “the causal chain” rather than on the outside elements. Monitoring has played a major role in reframing the debate and stimulating risk assessments, goal and strategy formulation and implementation.

1 Further Information on the project can be obtained from William C. Clark, Center for Science and International Affairs, Harvard University (clark@ksgsch.harvard.edu). The results of the study will be published by MIT Press in a book: “Learning to Manage Global Environmental Risks: A Comparative History of Social Responses to Climate Change, Ozone Depletion, and Acid Rain”.

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Risk assessment is concerned with understanding the nature, causes, consequences, likelihood and timing of the risk in question. The empirical evidence shows that over the time period studied here for each of the global environmental risks—ozone depletion, climate change, and acid rain—there is a trend over time toward multi-agency design of the assessment. There are, however, in each case, for each of the issues, difficulties and tensions in trying to look at all of the elements of “the causal chain”. There is generally a concentration on the risk assessment with respect to emissions and concentrations, with some attention to environmental impacts. However, very little attention is focused upon the driving factors, the human needs and wants and choices that are causing these emissions or changes in concentrations and much less attention to risk assessments of the socio-economic and social responses that are a result of these changes in concentrations and changes in the environment.

Risk assessment of global environmental risks also demonstrates “stickiness”. For instance, in the climate change debate, in 1979 the estimate was made that if the concentration of carbon dioxide in the atmosphere doubled, the temperature rise would be 3°C ± 1.5°C (i.e. a range of 1.5–4.5°C). In most of the major risk assessments on climate change that have been made since 1979, although climate models have changed since then and although other aspects of climate science have been evolving, the assessments have concluded that there was no reason to alter the previously agreed range of 1.5–4.5°C.

Response assessment is concerned with finding possible options for responding to global environmental risks. In many cases examined in this study, risk assessment and response assessment are closely linked and difficult to separate. Indeed single assessment efforts can include both risk and response option assessments. In the ozone depletion case, as another example, very early in the 1970s an estimate was made that if the amount of ozone in the stratosphere is decreased by 1 percent with a resulting 2 percent increase in the amount of ultraviolet radiation coming to the earth’s surface, there will be a 2 percent increase in the incidence of skin cancer. This number, the 2 percent increase in skin cancer is used until the beginning of the 1990s although there were advances in scientific knowledge, especially through public health studies. There are clearly resistances to changing a number that has become accepted in previous risk assessments. For each of the global environmental risks, there is a bias towards emission reduction technologies, with a suppression of options to change the environment or adapt to environmental changes. This is not only in the climate case where the response assessments have been very strongly biased toward discussing CO₂ emissions reductions.

Goal and strategy formulation is an important part of the process of managing global environmental risks. Goals are statements of objectives or of conditions that an actor wishes to bring about. Strategies are plans for how goals will be achieved. For each of the global environmental risks, the empirical data show that goals to reduce the emissions of pollutants dominate. The data also show that during the early phases of management of the risk there is a domination of goals to build capacity to understand or manage the risk, for example to set up a research program or set up a monitoring program. For broad targets, over time, and for each of the global environmental risks studied here, the empirical data show that the goals are converging across countries. In a relatively short period of time the same goal is found in most arenas, whether it’s a certain percentage reduction of emissions or other goals in terms of capacity-building.
The empirical data on implementation show, not surprisingly given the data on goal and strategy formulation, that in the regulatory action there is a clustering on actions that reduce emissions and build capacity. However, over the time period studied here, there is an increasing number of flexible regulatory measures introduced in response to the global environmental risks. The implementation process is opening over time from the beginning of the 1970s to 1992 with increased Non-Governmental Organisation involvement in particular. At the same time, the debate about implementation has evolved from being very technical to being more political in terms of what is going to be implemented and by whom.

Evaluation is a self-conscious effort to assess the performance of oneself and/or others. The study shows that evaluation practices in the management of global environmental risks are not universal or homogeneous. Societal culture in the different arenas studied here plays a very important role in shaping how evaluation is done and indeed what evaluation is done. The types of evaluation that were sought included evaluation of research programs and of integrated efforts at assessments of policy. The evidence suggests that there are very few explicit examples of evaluation in the management of global environmental risks.

3. LINKAGES BETWEEN KNOWLEDGE AND ACTION

The study shows that as the issues of global environmental risks emerge, knowledge and action are largely disconnected. They are scientific issues that receive some acknowledgment at certain levels in the policy-making sphere but are not on the political agenda. As the issues move higher on the political agenda, knowledge begins to drive action. In particular, assessments stimulate goal statements (in the cases mentioned here, most often with respect to emissions reductions). Subsequently, there is an emergence of policy-driven assessment and monitoring with some evidence of self-conscious evaluation of progress in responding to the issue at hand. These observed changes in the relationship between knowledge and action are generally not recognized by the actors involved (scientists or policy-makers). It is likely that different kinds of assessments are most useful at different stages of issue evolution. A fact that appears to have been largely ignored to date in the management of global environmental risks.

The evolution can be illustrated by looking at the issue of global climate change. Until the mid-1980s, scientific capacity was building up with a series of assessments and monitoring activities by individuals, national and international groups. The Study of Man's Impact on Climate at the beginning of the 1970s belongs in this category, as do the contributions to the First World Climate Conference in 1979. Climate monitoring and assessment of the data to identify the nature and causes of climatic changes were ongoing and provided an important basis for later assessments.

This was followed by the emergence of the issue onto the policy agenda. At the international level this occurred at the Villach Conference of 1985, where scientists reached a consensus about the threat of global warming and concluded that it was time to begin a dialogue with policy-makers. This was followed by international meetings in Villach, Bellagio, Toronto, Nordwijk etc. On each of these occasions, the assessment of the risk and response options stimulated a goal and/or strategy statement.
Most significant of these was the so-called "Toronto goal", which called for CO₂ emissions reductions of 20% by the year 2005. This goal stimulated both assessment and policy action at the national and international level and even at the municipal level.

After the establishment of the IPCC in 1988, the interactions between knowledge (monitoring and assessment) and action (policy development and implementation) intensified and were in both directions. For example, implementation included the establishment of the Intergovernmental Negotiating Committee for the Framework Convention on Climate, after the first report of the IPCC in 1990. The goal of returning CO₂ emissions to 1990 levels by the year 2000 led to assessments of options for emissions reduction. In this policy-driven phase, however, there has so far been very little self-evaluation (i.e. asking the question "How are we doing at managing the risk of global climatic change?"). Reorganisation of the IPCC after the first assessment is an indication of evaluation as are the country reports under the Framework Convention on Climate Change. But major feedback from such evaluations into the assessment or policy processes has not yet occurred.

4. CONCLUDING REMARKS

This paper reports on a small part of a much larger study. It serves to illustrate, however, that there is much to be learned by looking at how social actors have been managing global environmental risks over the past twenty years or more. The results shown here regarding the linkages between knowledge and action especially in the climate change case could be useful in the future in the design of assessments and in developing effective policy responses to global environmental problems.
Decadal Variations in Climate Associated with the North Atlantic Oscillation

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1. INTRODUCTION

At the core of the debate of anthropogenic climate change is the observation that global mean surface temperatures have risen between 0.3° and 0.6°C over the last century, including a warming of 0.2° to 0.3°C over the past four decades when the data are considered to be most reliable (IPCC 1996). The average warming has not been uniform, however, with differing decadal trends and marked seasonal and regional variations. Over the Northern Hemisphere (NH), average temperatures decreased from the late 1930s to the mid-1960s. Since the mid-1970s, NH temperatures have increased to record levels during the past decade. The recent warming has been largest during the winter and spring seasons over the continents between 40°N and 70°N, while weak cooling has occurred over the northern oceans (Fig. 1a).

It is the purpose of this paper to show that the cooling over the northwest Atlantic and the warming across Europe and downstream over Eurasia since the early 1980s are directly related to decadal changes in the North Atlantic Oscillation (NAO), while the temperature anomalies over the Pacific basin and North America result in part from tropical forcing associated with the El Niño–Southern Oscillation (ENSO) phenomenon but with important feedbacks in the extratropics. The changes in circulation over the Atlantic are also linked to coherent large-scale anomalies in precipitation since the early 1980s including dry conditions over southern Europe and the Mediterranean and wetter-than-normal conditions over northern Europe and parts of Scandinavia. The results summarize and update the findings of Hurrell (1995a, 1996) and Hurrell and van Loon (1997) and emphasize the point that the NAO, in addition to the Southern Oscillation (SO), is a major source of interannual variability of weather and climate around the world.

2. CHANGES IN THE PACIFIC

Decadal variations in the climate over the North Pacific with associated teleconnections downstream across North America have long been of interest and have been particularly highlighted by the work of Namias (1959, 1963, 1969); see also Dickson and Namias (1976), Douglas et al. (1982) and Namias et al. (1988). Recently, a large amount of evidence has emerged of a substantial change in the wintertime atmospheric circulation over the North Pacific that began in the mid-1970s and lasted throughout the 1980s. The changes involved the Pacific-North American (PNA) teleconnection pattern and corresponded to a deeper and eastward shifted Aleutian low pressure system (Fig. 1b) which advected warmer and moister air along the west coast of North America and cooler and drier air over the central North Pacific (Nitta and Yamada, 1989; Trenberth, 1990; Trenberth and Hurrell, 1994).
Consequently, there were increases in temperatures and sea surface temperatures (SSTs) along the west coast of North America and Alaska but decreases in SSTs over the central North Pacific (Fig. 1a). Changes in coastal rainfall and streamflow have also been noted (Cayan and Peterson, 1989), as well as decreases in sea ice in the Bering Sea (Manak and Mysak, 1987). Changes in the mean flow were accompanied by a southward shift in the storm tracks and associated synoptic eddy activity (Trenberth and Hurrell, 1994) and in surface ocean sensible and latent heat fluxes (Cayan 1992). On decadal and longer time scales, changes in the ocean must also become a factor in maintaining the extratropical circulation anomalies (Latif and Barnett, 1994).

A simple index used to measure the variations over the North Pacific (NP) is the area-weighted mean sea level pressure (SLP) over the region 30° to 65°N, 160°E to 140°W (Trenberth and Hurrell, 1994). The NP index (for North Pacific), averaged over the winter months from December through March, is shown in Fig. 2 since 1925 (the given year corresponds to the January of the winter season). Pressures from 1977 to 1988 were lower by 2.2 mb relative to the 70-winter NP-area mean. The only previous period that comparable values occurred was for a much shorter interval in the early 1940s. The pattern of temperature change associated with the NP index (Fig. 3) shows that the recent North Pacific basin anomalies (Fig. 1a) are consistent with the longer record: below normal NP values are associated with below-normal temperatures over the North Pacific and above normal surface temperatures along the west coast of North America extending into Alaska and across much of Canada. The departure pattern during winter also reveals below-normal temperatures over the southeast United States, which illustrates the PNA teleconnection and occurs in opposition to the temperature changes associated with NAO (see Fig. 5).

The decadal changes over the North Pacific have been linked to variations in the tropics (Trenberth and Hurrell, 1994; Kawamura, 1994), and several modeling studies have confirmed that North Pacific atmospheric variability is controlled in part by anomalous tropical Pacific SST forcing (Kitoh, 1991; Chen et al., 1992; Lau and Nath, 1994; Graham et al., 1994; Miller et al., 1994; Kumar et al., 1994). Fluctuations in tropical SSTs are related to changes in ENSO, and the observed warming of the tropical waters since the mid-1970s (e.g., Trenberth and Hoar, 1996) has been linked to increased tropospheric temperatures and water vapor in the western Pacific (Hense et al., 1988; Gaffen et al., 1991; Gutzler, 1997) and a more active hydrological cycle (Nitta and Yamada, 1989; Graham, 1995). The variability of the SO is evident in the NP index (Fig. 2), but feedback effects in the extratropics may serve to emphasize the decadal over interannual time scales relative to the tropics (Trenberth and Hurrell, 1994).

3. CHANGES IN THE ATLANTIC

a. The NAO Index

Over much of the past decade, anomalous coldness during winter has prevailed near Greenland and the eastern Mediterranean, while very warm conditions have been notable over Scandinavia,
northern Europe, the former Soviet Union and much of central Asia (Fig. 1a). Interannual and longer time-scale changes in the atmospheric circulation and lower tropospheric temperatures during winter over the North Atlantic and adjacent land areas do not appear to be strongly influenced by tropical SST variability (e.g., Barnett, 1985; Kumar et al., 1994; Graham et al., 1994; Lau and Nath, 1994). Rather, the anomalies are more strongly linked to the recent behavior of the NAO.

The NAO, which is associated with changes in the surface westerlies across the Atlantic onto Europe, refers to a meridional oscillation in atmospheric mass with centers of action near the Icelandic low and the Azores high (e.g., van Loon and Rogers, 1978). Although it is evident throughout the year, it is most pronounced during winter and accounts for more than one-third of the total variance of the SLP field over the North Atlantic (Wallace and Gutzler, 1981; Barnston and Livezey, 1987). Because the signature of the NAO is strongly regional, a simple index of the NAO can be defined as the difference between the normalized mean winter (December–March) SLP anomalies at Lisbon, Portugal and Stykkisholmur, Iceland (Hurrell, 1995a). The winter SLP anomalies at each station were normalized by dividing each seasonal pressure by the long-term mean (1964–1985) standard deviation. The variability of the NAO index since 1864 is shown in Fig. 4, where the heavy solid line represents the low pass filtered meridional pressure gradient. Positive values of the index indicate stronger-than-average westerlies over the middle latitudes associated with low pressure anomalies over the region of the Icelandic low and anomalously high pressures across the subtropical Atlantic.

In addition to a large amount of interannual variability, there have been several periods when the NAO index persisted in one phase over many winters (van Loon and Rogers, 1978; Barnett, 1985). Over the region of the Icelandic low, seasonal pressures were anomalously low during winter from the turn of the century until about 1930 (with the exception of the 1916–1919 winters), while pressures were higher than average at lower latitudes. Consequently, the wind onto Europe had a strong westerly component, and the moderating influence of the ocean contributed to higher-than-normal temperatures over much of Europe (e.g., Parker and Folland, 1988). From the early 1940s until the early 1970s, the NAO index exhibited a downward trend, and this period was marked by European wintertime temperatures that were frequently lower-than-normal (van Loon and Williams, 1976; Moses et al., 1987). A sharp reversal has occurred over the past 25 years and, since 1980, the NAO has remained in a highly-positive phase with SLP anomalies of more than 3 mb in magnitude over both the subpolar and subtropical Atlantic (Fig. 1b, see also Walsh et al., 1995). The 1983, 1989 and 1990 winters were marked by the highest positive values of the NAO index recorded since 1864 (Fig. 4). Beniston et al. (1994) note that blocking highs over Switzerland were more frequent during the 1980s than at any other time this century, with a decadal frequency 2–3 times greater than during previous decades. Nearly 25% of the total observed blocking highs over Switzerland since 1900 occurred during the 1980s.

b. Relationships to Temperature

The changes in local surface temperatures and SSTs based on linear regression with the NAO index are shown in Fig. 5. The surface temperature data are the same as those used in Fig. 1a and
consist of land surface temperatures blended with SST data (Jones and Briffa, 1992; Parker et al., 1994). Changes of more than 1°C associated with a one standard deviation change in the NAO index occur over the northwest Atlantic and extend from northern Europe across much of Eurasia. Changes in temperatures over northern Africa and the southeast United States are also notable. The similarity between the departure pattern of temperature (Fig. 5) and the decadal anomalies over the North Atlantic and surrounding landmasses (Fig. 1a) is striking and suggests that the recent temperature anomalies over these regions are strongly related to the persistent and exceptionally strong positive phase of the NAO index since the early 1980s.

The effect of circulation changes on temperature can be quantified through multivariate linear regression (Palecki and Leathers, 1993). Previously, a common application of linear regression has been to remove the influence of the SO from hemispheric and global temperature time series (e.g., Jones, 1994; Christy and McNider, 1994). The NAO and NP indices, in addition to an index of the SO defined by the normalized Tahiti minus Darwin SLP anomalies (Trenberth, 1984), were regressed upon the NH extratropical (20°N to 90°N) temperature anomalies for each winter since 1935. The results can be extended farther back in time, but only with the tradeoff of less reliable NP and SO indices (Trenberth and Hurrell, 1994) and less reliable estimates of surface temperature and SST (Trenberth et al., 1992).

The regression model using all three indices explains 49% of the NH extratropical surface temperature variance (Table 1), and the leading regression coefficient is associated with the NAO index. The strong relation between the NP and the SO indices is readily apparent: the two are correlated at 0.51 (the 1% significance level is $r = 0.33$), and this contributes to the relatively high standard errors of the regression coefficients associated with the two indices. The partial correlation coefficients in Table 1 show the actual variance explained by a circulation index when the effects of the other indices are eliminated. For instance, while the correlation of the SO index with NH temperature anomalies is $-0.39$, the partial correlation coefficient is $-0.27$ when the NP index is held fixed.

Table 1. Summary statistics from the multivariate regression using the NAO, the NP and the SO indices as the independent variables and NH (20°N to 90°N) temperature anomalies as the dependent variable. Also shown is the correlation matrix, and the partial correlation coefficients. The index that is held constant is indicated by the subscript.

| Regression Coefficient | Estimate | Standard Error | t-Statistic | Probability of Larger $|t|$ |
|------------------------|----------|----------------|------------|-------------------------|
| NAO                    | 0.57     | 0.09           | 5.5        | 0.000                   |
| NP                     | -0.22    | 0.11           | -1.9       | 0.069                   |
| SO                     | -0.24    | 0.11           | -2.3       | 0.026                   |
| $R^2 = 0.49$           |          |                |            |                         |

<table>
<thead>
<tr>
<th>Correlation Matrix</th>
<th>Partial Correlation Coefficients</th>
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<tbody>
<tr>
<td>NAO 1.00</td>
<td>(NAO,T)$_{NP}$ = 0.61</td>
</tr>
<tr>
<td>NP 0.03 1.00</td>
<td>(NAO,T)$_{SO}$ = 0.60</td>
</tr>
<tr>
<td>SO -0.07 0.51 1.00</td>
<td>(NP,T)$_{NAO}$ = -0.44</td>
</tr>
<tr>
<td>T 0.58 -0.34 -0.39 1.00</td>
<td>(NP,T)$_{SO}$ = -0.18</td>
</tr>
<tr>
<td>NAO NP SO T</td>
<td>(SO,T)$_{NAO}$ = -0.43</td>
</tr>
<tr>
<td></td>
<td>(SO,T)$_{NP}$ = -0.27</td>
</tr>
</tbody>
</table>
When only the NAO and the SO indices are used in the regression more confidence can be placed in the estimated coefficients, and 46% of the variance of the extratropical NH temperatures is explained (Table 2). Shown in Fig. 6 are the temperature changes since 1935 associated with the NAO, the SO and their sum. Also shown are the mean NH temperature anomalies relative to the 61-winter mean and the residual time series after the removal of the temperature contributions by the two circulation indices. Over the entire period, variations of temperature associated with the NAO account for 34% of the hemispheric interannual variance, while the SO accounts for 15% of the interannual variance (Table 1). Both the NAO and the SO indices can account linearly for much, but not all, of the hemispheric warming over the past two decades. Over the period 1981 to 1995, the NH extratropical temperature anomaly relative to the 61-winter mean was 0.31°C, while the anomaly of the residual series was 0.10°C. Of the 0.21°C difference, the NAO-related warming was 0.16°C. Similarly, the NAO remained in a negative phase during much of the 1960s (Fig. 4), a period during which wintertime blocking events were frequent over the Atlantic basin. From 1961 to 1971, the NH extratropical temperature anomaly was −0.11°C and the NAO-related anomaly was −0.16°C. The 10-winter average of the residual temperature anomaly was 0.07°C. While the selection periods are arbitrary, the above examples serve to illustrate the importance of the NAO, in addition to the role of circulation changes associated with ENSO, in determining the hemispheric temperature anomalies.

Table 2. Same as for Table 1, but for only the NAO and SO indices.

| Regression Coefficient | Estimate | Standard Error | t-Statistic | Probability of Larger | | |
|------------------------|----------|----------------|-------------|----------------------|---|
| NAO                    | 0.56     | 0.09           | 5.3         | 0.000                | |
| SO                     | −0.35    | 0.09           | −3.75       | 0.001                | |

\[ R^2 = 0.46 \]

Regression was also performed locally using the NAO and the SO indices as the independent variables. The anomalies of the residual temperatures averaged over the winters since 1981, recomputed relative to a 1951–1980 base period, reveal that the distinctive regional features evident in Fig. 1a are linearly related to variations in these two indices (Fig. 7). Most notable is the reduction of the warm anomalies over the NH extratropical landmasses and the elimination of the strong cooling over the northwest Atlantic. The residual coolness over the North Pacific, the warmth over Alaska and the coolness over the southeast U.S. is partially related to the exclusion of the NP index from the regression. Note that the residual coolness over the eastern U.S. and Europe resembles the signal of aerosol forcing. It should also be recognized that the anomalies in Figs. 6 and 7 reflect the exclusion of other teleconnection indices that affect patterns of interannual temperature variability in the NH, and that linear regression techniques do not account for feedback (nonlinear) effects. The claim is not that the NAO and the SO indices provide the best fit to NH extratropical anomalies; rather, the indices were selected because they relate to well-understood circulation anomalies that have persisted for much of the past two decades.
c. Relationships to Precipitation

Changes in the mean circulation patterns over the North Atlantic are accompanied by pronounced shifts in the storm tracks and associated synoptic eddy activity (Rogers, 1990; Hurrell, 1995b). Changes in the mean and eddy components of the flow affect the transport and convergence of moisture and, therefore, can be directly tied to changes in regional precipitation. Since the early 1980s conditions have been anomalously dry over southern Europe and the Mediterranean and wetter-than-normal over northern Europe and parts of Scandinavia (Fig. 8). Over the Alps, for instance, snow depth and duration over the past several winters have been among the lowest recorded this century, causing economic hardships on those industries dependent on winter snowfall (Beniston and Rebetez, 1996).

The atmospheric moisture budget cannot be adequately computed prior to 1979 because of the lack of high quality global analyses. To provide an analogy to the low-frequency changes evident in the NAO index (Fig. 4), use is made of recent interannual variations from which low or near-normal NAO winters are compared with very-high NAO winters. The data used are the global analyses produced by the European Centre for Medium Range Weather Forecasts (ECMWF), which appear to be of sufficient quality to adequately evaluate the large-scale moisture budget (Trenberth and Guillemot 1995). The low or normal NAO winter composite is the average December through March ECMWF analyses for the winters 1979, 1985, 1986, 1987, and 1988. Because only three of these winters have negative index values, the composited index value is \(-0.6 \pm 0.8\). The high NAO composited index is \(3.5 \pm 0.9\); the average of the 1983, 1989, 1990, 1992 and 1993 winter indices.

Vector plots of the vertically integrated total moisture transport for both the high and the normal/low NAO composites (Fig. 9) show that, during times of a high NAO index, the axis of maximum moisture transport shifts to a more southwest-to-northeast orientation across the Atlantic and extends much farther to the north and east onto northern Europe and Scandinavia. A significant reduction of the total atmospheric moisture transport occurs over parts of southern Europe, the Mediterranean and north Africa. It is the divergence of the moisture transport, however, that determines the excess of precipitation over evaporation.

Evaporation \((E)\) exceeds precipitation \((P)\) over much of Greenland (Fig. 10) during high NAO winters, especially in the south where changes are on the order of 1 mm day\(^{-1}\). When viewed with the low-frequency changes evident in the NAO index (Fig. 4), these results are consistent with observational and modeling evidence of a declining precipitation rate over much of the Greenland Ice Sheet over the past two decades (Bromwich et al., 1993). Drier conditions during high NAO index winters are implied over much of central and southern Europe and the Mediterranean, while enhanced moisture flux convergence is indicated from Iceland through Scandinavia.

Long-term station data from the NCAR archives lend further support to the \(E - P\) pattern. Listed in Table 3 are 39 stations that contain December–March records of precipitation for at least 40 winters. Shown are the correlation coefficients with the NAO index and the number of winters that were included in the correlations. Many of the correlations are highly significant and show that the recent precipitation
anomalies over Europe (Fig. 8) can be directly linked to decadal anomalies in the NAO. Also indicated in Table 3 is the mean winter precipitation rate for each station, computed over the total number of years, as well as the difference in precipitation rates for winters in which the NAO index exceeded 1.0 minus winters in which the index was below –1.0. The station data provide further evidence that the recent precipitation anomalies are related to the persistent positive phase of the NAO since about 1980.
Table 3. Stations (latitude, longitude) that contain records of December–March precipitation for at least 40 winters. The correlation coefficients $r(\text{NAO}, P)$ with the NAO index (Figure 4) and the total number of winters ($n$) that were included in the correlations are indicated. Also indicated is the mean precipitation rate $\bar{P}$ over the total number of winters ($n$), and the difference in precipitation rate between winters with a NAO index > 1.0 and those with an index < -1.0. One asterisk indicates statistical significance at the 5% level and two indicate significance at the 1% level.

<table>
<thead>
<tr>
<th>Station</th>
<th>$r(\text{NAO}, P)$</th>
<th>Years</th>
<th>$\bar{P}$ (mm day$^{-1}$)</th>
<th>$\Delta P$ (mm day$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bergen (60.4°N, 5.3°E)</td>
<td>0.77**</td>
<td>72</td>
<td>5.8</td>
<td>3.6**</td>
</tr>
<tr>
<td>Stornoway (58.2°N, 6.3°W)</td>
<td>0.75**</td>
<td>63</td>
<td>3.5</td>
<td>1.4**</td>
</tr>
<tr>
<td>Tiree (56.6°N, 6.9°W)</td>
<td>0.67**</td>
<td>63</td>
<td>3.4</td>
<td>1.2**</td>
</tr>
<tr>
<td>Stavanger (58.9°N, 5.6°E)</td>
<td>0.66**</td>
<td>43</td>
<td>2.7</td>
<td>1.4**</td>
</tr>
<tr>
<td>Thorshavn (62.0°N, 6.8°W)</td>
<td>0.53**</td>
<td>116</td>
<td>4.8</td>
<td>1.1**</td>
</tr>
<tr>
<td>Lerwick (60.1°N, 1.2°W)</td>
<td>0.49**</td>
<td>63</td>
<td>3.6</td>
<td>0.8**</td>
</tr>
<tr>
<td>Reykjavik (64.1°N, 21.9°W)</td>
<td>0.48**</td>
<td>73</td>
<td>2.7</td>
<td>0.9**</td>
</tr>
<tr>
<td>Akureyri (65.7°N, 18.1°W)</td>
<td>0.43**</td>
<td>63</td>
<td>1.6</td>
<td>0.4*</td>
</tr>
<tr>
<td>Stykkisholmur (65.1°N, 22.7°W)</td>
<td>0.40**</td>
<td>117</td>
<td>2.2</td>
<td>0.7**</td>
</tr>
<tr>
<td>Haparanda (65.8°N, 24.2°E)</td>
<td>0.37**</td>
<td>130</td>
<td>1.2</td>
<td>0.4**</td>
</tr>
<tr>
<td>Karesuando (68.5°N, 22.5°E)</td>
<td>0.21*</td>
<td>115</td>
<td>0.6</td>
<td>0.1</td>
</tr>
<tr>
<td>Oslo (60.2°N, 11.1°E)</td>
<td>0.21*</td>
<td>125</td>
<td>1.3</td>
<td>0.2</td>
</tr>
<tr>
<td>Helsinki (60.3°N, 25.0°E)</td>
<td>0.18*</td>
<td>130</td>
<td>1.5</td>
<td>0.1</td>
</tr>
<tr>
<td>Edinburg (56.0°N, 3.4°W)</td>
<td>0.14</td>
<td>130</td>
<td>1.7</td>
<td>0.2</td>
</tr>
<tr>
<td>Stockholm (59.4°N, 18.0°E)</td>
<td>0.14</td>
<td>126</td>
<td>1.1</td>
<td>0.1</td>
</tr>
<tr>
<td>Copenhagen (55.7°N, 12.6°E)</td>
<td>0.14</td>
<td>129</td>
<td>1.3</td>
<td>0.1</td>
</tr>
<tr>
<td>Valienta (51.9°N, 10.3°W)</td>
<td>0.09</td>
<td>123</td>
<td>4.5</td>
<td>0.1</td>
</tr>
<tr>
<td>DeBilt (52.1°N, 5.2°E)</td>
<td>0.08</td>
<td>130</td>
<td>1.9</td>
<td>0.1</td>
</tr>
<tr>
<td>Belfast (54.7°N, 6.2°W)</td>
<td>0.01</td>
<td>63</td>
<td>2.3</td>
<td>0.0</td>
</tr>
<tr>
<td>London (51.2°N, 0.2°W)</td>
<td>-0.02</td>
<td>129</td>
<td>1.7</td>
<td>-0.0</td>
</tr>
<tr>
<td>Angmagssalik (65.6°N, 37.6°W)</td>
<td>-0.02</td>
<td>90</td>
<td>2.6</td>
<td>-0.0</td>
</tr>
<tr>
<td>Athens (38.0°N, 23.7°E)</td>
<td>-0.11</td>
<td>98</td>
<td>1.7</td>
<td>-0.1</td>
</tr>
<tr>
<td>Egedesminde (68.7°N, 52.8°W)</td>
<td>-0.13</td>
<td>38</td>
<td>0.6</td>
<td>-0.1</td>
</tr>
<tr>
<td>Paris (49.0°N, 2.5°E)</td>
<td>-0.19*</td>
<td>119</td>
<td>1.5</td>
<td>-0.3**</td>
</tr>
<tr>
<td>Frankfurt (50.1°N, 8.7°W)</td>
<td>-0.19*</td>
<td>130</td>
<td>1.4</td>
<td>-0.2*</td>
</tr>
<tr>
<td>Godthåb (64.2°N, 51.8°W)</td>
<td>-0.20*</td>
<td>100</td>
<td>1.1</td>
<td>-0.4</td>
</tr>
<tr>
<td>Jakobshavn (69.2°N, 51.1°W)</td>
<td>-0.21*</td>
<td>95</td>
<td>0.4</td>
<td>-0.1</td>
</tr>
<tr>
<td>Ivigtut (61.2°N, 48.2°W)</td>
<td>-0.31*</td>
<td>89</td>
<td>2.8</td>
<td>-0.9*</td>
</tr>
<tr>
<td>Marseille (43.5°N, 5.2°E)</td>
<td>-0.32**</td>
<td>120</td>
<td>1.5</td>
<td>-0.5**</td>
</tr>
<tr>
<td>Milan (45.4°N, 9.3°E)</td>
<td>-0.35**</td>
<td>130</td>
<td>2.2</td>
<td>-0.8**</td>
</tr>
<tr>
<td>Istanbul (41.0°N, 29.1°E)</td>
<td>-0.36**</td>
<td>64</td>
<td>2.7</td>
<td>-0.7**</td>
</tr>
<tr>
<td>Lyon (45.7°N, 5.0°E)</td>
<td>-0.37**</td>
<td>129</td>
<td>1.6</td>
<td>-0.5**</td>
</tr>
<tr>
<td>Rome (41.8°N, 12.2°E)</td>
<td>-0.37**</td>
<td>119</td>
<td>2.6</td>
<td>-0.8**</td>
</tr>
<tr>
<td>Ajaccio (41.9°N, 8.8°E)</td>
<td>-0.48**</td>
<td>42</td>
<td>2.2</td>
<td>-1.1**</td>
</tr>
<tr>
<td>Ponta Delgado (37.8°N, 25.7°W)</td>
<td>-0.49**</td>
<td>98</td>
<td>3.2</td>
<td>-1.1**</td>
</tr>
<tr>
<td>Belgrade (44.8°N, 20.5°E)</td>
<td>-0.50**</td>
<td>94</td>
<td>1.4</td>
<td>-0.6**</td>
</tr>
<tr>
<td>Casablanca (33.6°N, 7.7°W)</td>
<td>-0.61**</td>
<td>82</td>
<td>2.0</td>
<td>-1.1**</td>
</tr>
<tr>
<td>Lisbon (38.8°N, 9.1°W)</td>
<td>-0.64**</td>
<td>130</td>
<td>3.0</td>
<td>-1.8**</td>
</tr>
<tr>
<td>Madrid (40.4°N, 3.7°W)</td>
<td>-0.69**</td>
<td>129</td>
<td>1.3</td>
<td>-1.0**</td>
</tr>
</tbody>
</table>
4. SUMMARY

The pattern of temperature change over the past two decades has been one of warming over NH continents and cooling over the oceans (Fig. 1a), and this pattern is strongly tied to decadal changes in the atmospheric circulation (see also Wallace et al. 1995). Nearly all of the observed cooling in the northwest Atlantic and the warming across Europe and downstream over Eurasia since about 1980 results from changes in the NAO, and the NAO accounts for 34% of the hemispheric interannual variance of surface temperature over the past 61 winters. In addition, the NAO can be linked to decadal anomalies in precipitation including wet conditions over northern Europe and Scandinavia since 1980, and dry conditions over southern Europe, the Mediterranean, parts of North Africa, and western Greenland. It is interesting to note that the NAO changed phase during the winter of 1996 (Fig. 4) when SLP anomalies over the North Atlantic were nearly opposite to those in Fig. 1b and temperatures were colder-than-normal throughout Europe. The mechanisms responsible for such large interannual variations of the NAO, in addition to those associated with decadal variations, are not well understood and are the subject of ongoing research.

References


Figure 1. Fifteen winter (1981–1995) average (a) surface temperature and SST anomalies and (b) SLP anomalies expressed as departures from the 1951–1980 means. Temperature anomalies >0.25°C are indicated by dark shading, and those <–0.25°C are indicated by light shading. The same shading convention is used for SLP anomalies greater than 1 mb in magnitude. Regions of insufficient data coverage are not contoured in (a).
Figure 2. Time series of normalized mean North Pacific SLP averaged over 30° to 65°N, 160°E to 140°W (the NP index) for the months December–March beginning in 1925 and smoothed with a low-pass filter with seven weights to remove fluctuations with periods less than 4 years.
Figure 3. Changes in temperatures ($\times 10^{-1}^\circ$C) corresponding to a unit deviation of the NP index, multiplied by minus one, computed over the winters from 1935 through 1995. The contour increment is $0.2^\circ$C. Temperature departures $>0.2^\circ$C are indicated by dark shading, and those $<-0.2^\circ$C are indicated by light shading. Regions of insufficient data are not contoured.
Figure 4. Time series of the winter (December–March) index of the NAO (as defined in the text) from 1864–1996. The heavy solid line represents the meridional pressure gradient smoothed with a low pass filter to remove fluctuations with periods less than 4 years.
Figure 5. As in Fig. 3, but for changes in temperatures ($x 10^{-1}^\circ$C) corresponding to a unit deviation of the NAO index computed over the winters from 1935 through 1995.
Figure 6. Temperature changes associated with the NAO, the SO and their sum (20°N–90°N) relative to the 1935–1995 mean. Also shown are the mean NH temperature anomalies and the residuals after removing the linear effects of the NAO and SO.
Figure 7. Average (1981–1995) winter surface temperature and SST anomalies as departures from the 1951–1980 mean after the linear effects of the NAO and the SO have been removed. Residual anomalies >0.25°C are indicated by dark shading, and those <−0.25°C are indicated by light shading. Regions of insufficient data coverage are not contoured.
Figure 8 Fifteen winter (1981-1995) average precipitation anomalies expressed as departures from the 1951-1980 mean. The contour increment is 0.2 mm day$^{-1}$, except the ± 0.1 mm day$^{-1}$ contours are included. Anomalies >0.1 mm day$^{-1}$ are indicated by dark shading, and those <−0.1 mm day$^{-1}$ are indicated by light shading. The dataset was kindly provided by Dr. Mike Hulme and is an extension of the dataset described by Eischeid et al. (1991).
Figure 9. Vectors of the vertically integrated moisture transports for (a) high NAO index winters and (b) normal or low NAO index winters. The contour interval of the magnitudes is 50 kg m$^{-1}$s$^{-1}$.
Figure 10. Evaporation ($E$) minus precipitation ($P$) computed from the atmospheric moisture budget for high minus normal or low NAO index winters. The contour interval is 0.5 mm day$^{-1}$, and the stippling indicates values significantly different from zero at the 5% level using a $t$-test.
An Observational Study of the Semiannual Oscillation in the Tropics and Northern Hemisphere

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Iowa State University
* National Center for Atmospheric Research, Boulder, CO 80307

1. THE SEMIANNUAL OSCILLATION IN THE TROPICS

A pronounced asymmetry in the semiannual oscillation (SAO) of the middle and upper-tropospheric zonal wind and temperature between the eastern and western hemispheres was found by van Loon and Jenne (1969). Later, focusing on the monsoon longitudes, van Loon and Jenne (1970) made a more comprehensive analysis of the semiannual cycle. Based upon results obtained in these two studies, they suggested that the semiannual variation in the Southern Hemisphere (SH) tropics may be associated with the interseasonal migration of the upward branch of the local Hadley circulation over the monsoon regions in the eastern hemisphere, in contrast to the western hemisphere where it mainly stays in the Northern Hemisphere (NH) through the year. Our interest in van Loon and Jenne's mechanism was stimulated by two previous studies:

(a) The global divergent circulation undergoes an east-west seesaw SAO between the Asian-Australian (AA) monsoon (60°E–120°W) and the extra-AA monsoon (120°W–60°E) hemispheres in agreement with that of tropical cumulus convection (Chen and Wu 1992).

(b) The streamfunction budget equation provides a way to illustrate the interaction between the global divergent circulation and rotational flow through the vorticity source (Chen and Chen 1990).

The large-scale circulation in the tropics is well depicted by the streamfunction. In order to explore the asymmetry in the SAO of the upper-level zonal wind, an effort was made in the present study to use the National Meteorological Center (NMC) data of 1986-1992 to examine (a) the three-dimensional structure of the tropical stationary eddies and (b) the role played by the global divergent circulation in the SAO of the tropical stationary eddies. The major findings related to the tropical SAO are:

(a) The stationary eddy streamfunction (\(\tilde{\psi}_E\)) undergoes a semiannual seesaw between the eastern and western hemispheres (Fig. 1). This SAO is particularly clear in the Southern Hemisphere tropics. The semiannual component of the eddy streamfunction exhibits a vertical phase reversal at the 400-500 mb layer and at the equator in the meridional direction (not shown). Evidence of this phase reversal can be seen upon examination and contrast between the stationary eddy streamfunction at 200 mb (Fig. 1a) and 850 mb (Fig. 1b) in the Northern Hemisphere tropics, the semiannual component of eddy streamfunction does not show a vertical phase reversal over the western Pacific and the Atlantic.
The note that eddy streamfunction and velocity potential (χ) are linked through the streamfunction
budget equation, which is the inverse Laplace transform of the vorticity equation (Chen and Chen 1990),

\[ \frac{\partial \tilde{\Psi}_E}{\partial t} = \nabla^2 \left[ -V_\psi \cdot \nabla (\zeta + f) \right]_E + \nabla^2 \left[ -\nabla \cdot \left( V_\psi (\zeta + f) \right) \right]_E + \nabla^2 \left( \tilde{R} \right)_E. \]

We find that two physical processes, the streamfunction tendencies induced by the horizontal
advection of vorticity (\( \tilde{\Psi}_{AVG} \)) and by the vorticity source (\( \tilde{\Psi}_{XE} \)) maintain the SAO of tropical stationary
eddies, counterbalance each other, and are spatially and temporally in quadrature with \( \tilde{\Psi}_E \) (the
semiannual component of the eddy streamfunction) (Fig. 2).

It is inferred from the relation \( \tilde{X}_E \rightarrow \tilde{\Psi}_{XE} \rightarrow \tilde{\Psi}_{AVG} \) that the east-west seesaw in the SAO of
the tropical stationary eddies between the eastern and western hemisphere is caused by that of the
global divergent circulation between the AA-monsoon and the extra-AA hemisphere in response to the
east-west seesaw in the SAO of tropical cumulus convection. This mechanism is particularly clear in the
Southern Hemisphere tropics, but less so in the Northern Hemisphere tropics.

2. THE SEMIANNUAL OSCILLATION IN THE NORTHERN HEMISPHERE

So far, there is no uniformly accepted explanation of the hemispheric-scale dynamic mechanism
for the SAO in the NH. It was suggested by Weickmann and Chervin (1988) that the NH high latitude
semiannual cycle may be due to east-west land-sea contrasts. In order to examine the NH extratropical
SAO, 5° latitude/longitude gridded monthly geopotential heights were obtained from the National Center
for Atmospheric Research (NCAR). We extracted data from 1961 to 1992.

The semiannual harmonic amplitude of the NH extratropical 500-hPa height field is shown in Fig.
3a. This figure includes five primary centers of semiannual cycle amplitude. In addition, the area north
of about 60°–65° shows a strong SAO amplitude. Perhaps the most interesting feature of this spatial
variation is the east-west asymmetric structure in the midlatitudes and the more north-south structure in
the high latitudes, with larger maxima in the hemisphere from 40°E eastward through 140°W. The SAO
phase distribution is shown in Fig. 3b. It is clear that the areas with maximum SAO amplitudes generally
have midwinter and midsummer peaks. The areas with smaller SAO amplitudes tend to have spring-fall
peaks. The pronounced zonally asymmetric distribution in the SAO amplitude field suggests that it would
be useful to look at the stationary eddy pattern.
We hypothesize that the SAO in the NH midlatitudes is related to the asymmetric response of the atmospheric circulation to the annual variation of solar heating. This relationship is evident in the asymmetric spatial response of the stationary eddy pattern between winter and summer. In order to see how the asymmetric response of the stationary eddies leads to a semiannual cycle in the midlatitudes, consider location A (40°N, 110°W) as shown in Fig. 3a. Figure 4a shows the longitude-time plot of the monthly average 500-hPa stationary eddy deviations (ZE) at 40°N. Following this longitude from January through December, it is clear that the secondary positive peak in the summer is related to the westward extension of the eddy anomaly, located to the east of location A. Figure 4b shows that the secondary positive eddy anomaly in the summer is also influenced by the northward extension of the positive anomaly centered at 35°N. The apparent westward and northward shift of the eddy (Fig. 4) is simply a result of the asymmetric response to seasonal heating in the northeastern North American summertime positive anomaly.

We consider the semiannual cycle maxima shown in Fig. 3 to be located in the subpolar region. Although the eddy asymmetries contribute to the SAO at these locations, it is suggested by the differences between the SAO and ZE time series (not shown) that the asymmetries in ZE are not sufficient to explain the SAO in the subpolar region. Thus, it seems possible that the semiannual cycle of the zonal mean component may also contribute to the observed SAO in the polar region.

The vertical structure of the SAO was explored in terms of its spatial distribution of amplitude and phase for the 300-, 500-, and 700-hPa height field, as well as the sea level pressure field. Although the SAO amplitude increases with height, the general structure of the amplitude and phase fields (not shown) is similar throughout all levels. The general structural similarity in SAO phase and the locations of SAO amplitude centers in the vertical (particularly between the levels from 700 to 300 hPa) suggests an equivalent barotropic structure in the NH SAO.

Acknowledgments

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References


Figure 1. Amplitude and phase of the semiannual component of eddy streamfunction $\tilde{\psi}_E$ at (a) 200 mb and (b) 850 mb. The amplitude and phase of $\tilde{\psi}_E$ are represented by the length and direction of a straight line. A vertical line indicates maxima on January 1/July 1 and a horizontal line indicates maxima on April 1/October 1. The direction of a line is measured counterclockwise as the conventional measurement of an angle. The amplitudes of both $\tilde{\psi}_E(200 \text{ mb})$ and $\tilde{\psi}_E(850 \text{ mb})$ are contoured with intervals of $10^6 \text{ m}^2 \text{s}^{-1}$ and $5 \times 10^5 \text{ m}^2 \text{s}^{-1}$, respectively. The phases of $\tilde{\psi}_E$ in two quadrants (April 1–July 1 and October 1–January 1) are stippled.

Figure 2. Same as Fig. 1, except for (a) $\tilde{\psi}_{AVE}(200 \text{ mb})$ and (b) $\tilde{\psi}_{XE}(200 \text{ mb})$. The amplitudes of $\tilde{\psi}_{AVE}(200 \text{ mb})$ and $\tilde{\psi}_{XE}(200 \text{ mb})$ are contoured with an interval of $15 \text{ m}^2 \text{s}^{-2}$. 

250
Figure 3. (a) Semiannual harmonic amplitude for the 500-hPa height field. The 5-m contour interval with values greater than 20 m are lightly shaded and values greater than 40 m are heavily shaded. (b) Semiannual harmonic phase line segments for the 500-hPa height field. Phase is indicated by the orientation of the line segments, with a segment oriented north-south corresponding to maxima on 1 January and 1 July, and an east-west orientation corresponding to maxima on 1 April and 1 October.

Figure 4. (a) Longitudinal-time (x - t) plot of Z\textsubscript{E} at 40°N. (b) Latitude-time (y - t) plot of Z\textsubscript{E} at 110°W in (a) and the horizontal line at 40°N in (b) correspond to the Z\textsubscript{E} time series in Fig. 3a. The contour interval is 20 m with solid lines for positive anomalies, dashed lines for negative anomalies, and a dotted line for the 0-meter line. Positive contours greater than 20 m are shaded.
Storm and Surge Climate in the North Sea Area: Changes in the Past Century

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1. BACKGROUND

In the public debate concerning climate change due to increasing emissions of radiatively active gases into the atmosphere many people are concerned about the possibility of an intensification of extratropical storms. Even though the IPCC took a cautious stand in this matter because of lack of evidence, a mixture of indirect evidence (van Hoff, 1993; Hogben, 1994) and misleading scientific statements (Schinke, 1992) created a substantial uneasiness in the public. The offshore oil industry in the North Sea was confronted with reports about extreme waves higher than ever observed. The insurance industry organized meetings with scientists because of greatly increased storm-related damages. The Northern European newspapers were full of speculations about the enhanced threat of extratropical storms in the early part of 1993.

As a response to this situation the Norwegian Weather Service organized two workshops “Climate Trends and Future Offshore Design and Operation Criteria”, in Reykjavik and Bergen, and brought together people from the oil industry, certifying agencies and scientists to discuss the reality of a worsening of the wave and storm climate. The workshops did not create definite statements but the general impression that hard evidence for a worsening of the storm and wave climate was not available (for a summary see von Storch et al., 1994). A group of participants then established the “WASA project”. A major part of WASA (Waves and Storms in the North Atlantic) deals with the assessment of the wave climate—first results will be published in near future by WASA Wave Group (1997).

Parallel to the WASA project, the question of whether the storm surge climate in the North Sea has changed, and may possibly change in future, was pursued with funding by the German Ministry of Science and Technology (BMBF).

In the present paper conclusions from both projects are summarized. First, the results concerning the storm climate, deduced from various atmospheric parameters in WASA, and results deduced from changes in the high-frequency statistics of storm surges at the North Sea (BMBF), are presented—these data unequivocally indicate no change in storminess (Section 2). Then, in Section 3 a statistical model is designed which relates intramonthly quantiles of surge statistics at one location to the monthly mean air-pressure field over the Northeast Atlantic and Europe. This statistical model is fitted to 1970–1988 data of high water levels in Cuxhaven (a harbour in the German Bight at the mouth of the river Elbe; for more details, refer to von Storch and Reichardt, 1996) and tested with independent data since 1900. The model is used for deriving a scenario of future surge statistics in Cuxhaven at the expected time of doubled CO$_2$ concentrations around 2035.

1 Preliminary summaries of the WASA project are offered by WASA (1994,1995).
2. ANALYSIS OF THE OBSERVATIONAL RECORD OF STORMINESS

When assessing the temporal evolution of the storm climate, in principle two different types of data may be considered. One source of information could be the archive of weather maps, which covers more than hundred years. Indeed, several attempts have been made to count the number of storms, stratified after the minimum core pressure, in the course of time (Schinke, 1992; Stein and Hense, 1994). These studies are useful in describing the year-to-year fluctuations in the past, say, 10 years. However, for the longer perspective this approach is rendered inconclusive simply because the quality of the weather maps has steadily improved. Thus any creeping worsening of the storm climate apparent in the weather maps (as reported by Schinke, 1992) might reflect a real signal or be a result of the ever increasing quality of the operational analyses due to more and better observations, more powerful diagnostic tools and other improvements in the monitoring of the state of the troposphere. A more detailed mapping of the pressure distribution, however, automatically yields deeper lows.²

The inhomogeneity problem is illustrated by Fig. 1 in which the ratio of high-pass filtered standard deviations of air-pressure variations in winter in the decade 1984–93 and in the decade 1964–’73, as derived from the DNMI analyses³, is plotted. Obviously the variability is greatly enhanced since the 1960’s in areas where little or no in-situ observations are routinely available; this increase is likely spurious. In the area marked in Fig. 1, between 70°N and 50°N and east of 20°W the bias seems less severe. For this area slightly more storms were found in the 1984–93 decade than in the previous decades (348 as opposed to 339, 336 and 330). We do not know to what extent changes in the analysis scheme are responsible for the changing storm numbers in that area, therefore the result of this storm count should be taken as an upper bound of an increase of storm frequency and intensity.

Any analysis of changes of the storm climate should be supported by an analysis of local observations which are unaffected by improvements in the process of mapping the weather.⁴ A good parameter would be wind-speed, since it relates directly to damages and impact of waves and surges. However wind observations—either determined instrumentally or estimated—are usually of limited value due to inhomogeneities such as the change of scale, change of observer, change of surroundings etc. (cf. Peterson and Hasse, 1987).

Therefore one must look for other and more homogeneous proxies for storminess. An obvious choice is to base these on station air pressure, the time series of which are considered to be rather homogeneous because more or less the same instrument (mercury barometer) and procedures have been used throughout the entire observation period.

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² This problem is severe for (daily) weather maps; when dealing with monthly mean maps, the inhomogeneity becomes less significant.
³ The Norwegian Weather Service DNMI has prepared 6-hourly high resolution air pressure analyses from 1955 until today. Until 1982 the analyses were done manually with a resolution of 150 km; after that time an objective technique was used and the product became available on a 75 km grid. Commonly, these analyses are considered as the best product for the region.
⁴ According to "van Loon's Rule", this advice is almost universal—results obtained from analysed products should be checked with the help of local data.
From air-pressure two proxies for storminess may be formed, namely the annual (seasonal, monthly) distribution of the geostrophic wind speed derived from three stations in a triangle (Schmidt and von Storch, 1993; see Section 2.1) or the annual (seasonal, monthly) distribution of the pressure tendency, possibly after suppressing the non-synoptic variations by means of a digital filter (Schmith, 1995) (see Section 2.2). Another homogenous proxy data time series is provided by high-frequency sea level variations at a tide gauge. The variance of such variations is controlled by the variance of the synoptic atmospheric disturbances (see Section 2.3).

These proxy data, geostrophic wind, high-frequency pressure tendency and sea level variations, can not be used to reliably estimate actual wind speeds; however, changes of the annual (monthly) distributions are connected with similar changes in the distributions of the wind speed (cf. WASA (1995)).

2.1. Geostrophic Wind Analyses

For 15 stations, situated in Northwestern Europe and the Northeast Atlantic, an uninterrupted pressure record for about the last 100 years of three or four daily observations were identified in the WASA project, which could be homogenized (Alexandersson, 1986). With these stations, triangles were set up and daily geostrophic winds were derived. Here, the results for two such triangles are shown: one over Denmark and one over southern Sweden.

For the Danish triangle, annual distributions of geostrophic winds speeds were derived, and from each annual distribution the annual 50%, 90% and 99% quantiles were determined. The resulting time series of the annual quantiles (Fig. 2) show no significant upward or downward trend but some interdecadal variability.

A similar result is obtained for the Swedish triangle Goteborg-Visby-Lund. This time, the result is presented by plotting the annual number of cases with geostrophic wind speeds exceeding 25 m/s per year. Figure 3 indicates that recently, there may have been a weak reduction of the number of such cases in the past.

Also the analysis of other triangles offers no evidence of an ongoing worsening of the European storm climate.

2.2 Pressure Tendency Analysis

Large air pressure tendencies are indicative for major baroclinic developments so that large wind speeds are likely to occur somewhere in the neighborhood. Therefore Kaas et al. (1995) suggested the use of the absolute value of the 24-hourly pressure tendency as another possible proxy for storminess. They investigated the record of two stations, namely Fanø in Denmark and Torshavn on the Faroe Islands in the North Atlantic. In both cases no systematic increase of the 50%, 90% and 99% percentiles of the annual distributions of the pressure tendencies were found, as is exemplified in Fig. 4 for Fanø. In

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5 The 90% quantile of a distribution \( X \) is that number \( X_{90\%} \) so that the probability to observe any realization of \( X < X_{90\%} \) is 90%. In case of distributions formed from 366 daily geostrophic wind speeds during one year, the actual geostrophic wind is at 36 days equal or larger than the 90% percentile. Quantiles are often called fractiles or percentiles.
the later years, since about 1970, an increase is present for all three tide gauges, but this increase does not appear as alarming when compared to earlier time-limited trends. The recent trend may well be linked with the intensification of the North Atlantic Oscillation in the past decades (Hurrell, 1995), and it is certainly important to closely monitor the future development.

2.3 High Frequency Sea Level Variations

The idea to use high-frequency variations of sea level as a proxy for storm activity was suggested by de Ronde (cf. von Storch et al., 1994), who analysed data from Hoek van Holland. To do so, the annual mean water level is subtracted from the data, because changes in the mean water level are thought to reflect processes unrelated to the storm activity, such as local anthropogenic activity (e.g., harbour dredging), mean sea level rise or land sinking. After subtraction of the annual mean, intraannual distributions of the water level variations are formed, as in the case of geostrophic winds discussed above, and intraannual quantiles are determined.

The time series of annual mean sea level\(^6\) as well as the time series of intraannual quantiles at Cuxhaven (German Bight, at the mouth of the river Elbe) are displayed in Fig. 5. The mean water level has risen by about 30 cm/100 years, but the storm related intraannual quantiles have remained almost constant.\(^7\)

By now, the observational record at several tide gauges around the southern and eastern North Sea has been examined. As for Cuxhaven, at the other locations no increase of the storm-related intraannual quantiles were found. As examples, we present the time series for Den Helder (The Netherlands) and Esbjerg (Denmark) in Fig 6.

3. THE EMPIRICAL MODEL LINKING INTRAMONTHLY PERCENTILES AND MEAN AIR PRESSURE DISTRIBUTIONS

The empirical model is based on a Canonical Correlation Analysis which links two sets of random vectors \(\mathbf{S} \) and \(\mathbf{Q} \) (Barnett and Preisendorfer, 1987; von Storch, 1995). Both vectors are assumed to be centered, i.e., their time means are subtracted prior to the analysis. Also, a data compression with the help of EOFs is done prior to the analysis in order to avoid artificially enhanced correlations.

In the cases considered below, one vector time series, \(\mathbf{S} \), represents the winter (DJF) monthly mean air-pressure distributions. The other vector time series, \(\mathbf{Q} \), is formed by a few quantiles of the

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\(^6\) Here given as the height of the high tide, which takes place about twice daily.

\(^7\) Note that the effect of unchanged storm-related variations and a rise of the mean water level leads to a significant increase of the height of storm surges!
intramonthly distributions of our local parameter of interest. In case of storm-related water level

distributions, the 50%, 80% and 90% quantiles are considered so that 8.
q5ob
_50%

1

(1)
(1)

Q. = q80%
q90o%,

The result of a CCA are pairs of vectors ( ;k, pq;k) time coefficients a , (t) and a

;

(t) so that

S= A a sj(t)p;k
k

(2)

=Q,=aq;t(t)"q'
k

The k = I coefficients share a maximum correlation p', the k = 2 coefficients another maximum
correlation P2 obtainable under the constraints of being uncorrelated with the k = 1 coefficients and so

forth.
The patterns are eigenvectors of a matrix essentially featuring the cross-covariance matrix

between S and Q. The CCA coefficients a

are determined as a least squares fit by minimizing

Il t-X

K

.I S;k

I

(3)

The coefficients are normalized to one
VAR(a;k) = VAR(a,) = 1

so that the three components of pk may be interpreted as anomalies which occur typically together
with the "field distribution" p;k
The downscaling model which relates the large-scale air-pressure information to the local-scale
storm-related water level information is a regression model aq;k = pkas;k for the CCA-coefficients a,;k

and a q;k with a reconstruction in the three-dimensional space using (2):
q 50 %

(4)

K

Qt = 180%

Pka.s(t)P
k=l

,q 90% ,
The regression model (4) may e applied to anomalies of observed or simulated air pressure fields
S=E

8

as.;k

Note that the quantiles are also centered, so that the vector

Q. is composed of anomalies.
256


The success of the reconstruction of observed intramonthly water level percentiles is quantified by two measures of skill, namely the correlation skill score $\rho_\kappa$ and the percentage of represented variance $\varepsilon_\kappa$ for $\kappa = 50\%, 80\%$ and $90\%$ (Livezey, 1995):

$$\rho_\kappa = \frac{\text{COV}(\hat{q}_{\kappa,t}, q_{\kappa,t})}{\sqrt{\text{VAR}(\hat{q}_{\kappa,t})\text{VAR}(q_{\kappa,t})}}$$

$$\varepsilon_\kappa = 1 - \frac{\text{Var}(\hat{q}_{\kappa,t} - q_{\kappa,t})}{\text{VAR}(q_{\kappa,t})}$$

where $\hat{q}_{\kappa,t}$ is the estimated $\kappa$ percentile in the month $t$.

In the following Section 4 we show and discuss the resulting patterns $\bar{p}^{q;k}$ and $\bar{p}^{s;k}$ and their relationship for the case $\bar{S} = \text{monthly mean air pressure distribution}$ and $\bar{Q} = \text{intramonthly quantiles of observed high-pass filtered high water levels in Cuxhaven}$.

### 4. THE DOWNSCALING MODEL FOR WATER LEVEL QUANTILES

Before calculating for each winter month the intramonthly quantiles, first the annual means are computed and subtracted from the data. Then, the CCA is done with the 18-year subset of data December 1970 to February 1988. The remaining data, prior to 1970, are kept as independent data for verifying the statistical model.

<table>
<thead>
<tr>
<th>$\kappa$</th>
<th>50% [cm]</th>
<th>80% [cm]</th>
<th>90% [cm]</th>
<th>$\varepsilon_\kappa$ [%]</th>
<th>$\rho_\kappa$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>-21</td>
<td>-16</td>
<td>-18</td>
<td>83</td>
<td>0.89</td>
</tr>
<tr>
<td>2</td>
<td>-10</td>
<td>1</td>
<td>10</td>
<td>13</td>
<td>0.32</td>
</tr>
</tbody>
</table>

The two largest correlations amount to 0.89 and 0.32 (see Table 1) The first two CCA patterns of air pressure distributions are shown in Fig. 7 and the associated anomalies of storm related intramonthly water level percentiles are given in Table 1.

The first air pressure anomaly pattern, $\bar{p}^{s;1}$, describes a southeasterly flow across the North Sea, which is connected with a almost uniform decrease of all three considered percentiles of $-20$ cm. This CCA pair describes the dominant atmospheric control of water level variations - as much as $83\%$ of the...
variance month-to-month variability of intramonthly percentiles is represented by this first pair of patterns. From the distributions shown in Fig. 7 and an analysis of the characteristic pattern of high-pass filtered air pressure variance (see von Storch and Reichardt, 1996) we conclude that the first CCA pattern encompasses two factors affecting water level variations; first, there is a weakened mean northwesterly flow; secondly, this pattern reduces the formation of synoptic disturbances which travel in a southeasterly direction into the North Sea, where they pile up water in the German Bight.

The second air pressure pattern, $\overline{p}^{x,2}$, is less important for the variations of Cuxhaven water levels, since it represents in the fitting interval no more than 13% of the variance of the combined vector of percentile anomalies. Its link to water level variations is rather different from that of the first CCA pattern: The 50% percentile is lowered by 10 cm, the 80% percentile is almost unchanged and the 90% percentile is lifted by 10 cm. Thus, the water level distribution becomes markedly broader if this air pressure distribution prevails; if the sign of air pressure anomaly is reversed then the water level distribution tends to be narrower than on average. Also the second pair of CCA pairs is physically plausible. The anomalous air-pressure distribution of $\overline{p}^{x,2}$ in Fig. 7 does not cause an additional accumulation of water in the German Bight. Indeed the mean air flow across the North Sea is southeasterly and, consistently, the 50% percentile is reduced. However, this pattern steers occasionally energetic synoptic disturbances into the area of the North Sea (see von Storch and Reichardt, 1996) so that the higher percentiles are enhanced.

Table 2. The correlation skill $p_k$ and the percentage $\varepsilon_k$ of the month-to-month variability of the intramonthly percentile $q_k$ represented by the regression model ((4)), determined from independent data (1899–69).

<table>
<thead>
<tr>
<th>$K$</th>
<th>$\kappa = 50%$</th>
<th>$80%$</th>
<th>$90%$</th>
<th>$\varepsilon_k = 50%$</th>
<th>$80%$</th>
<th>$90%$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>.75</td>
<td>.73</td>
<td>.69</td>
<td>53%</td>
<td>50%</td>
<td>45%</td>
</tr>
<tr>
<td>2</td>
<td>.79</td>
<td>.72</td>
<td>.63</td>
<td>62%</td>
<td>50%</td>
<td>40%</td>
</tr>
</tbody>
</table>

The regression model (4) has been used for estimating intramonthly percentiles for Cuxhaven for the winters 1899 to 1988. The skill of the model (4) is summarized in Table 2. The inclusion of the second canonical pair improves the skill for the 50% percentile but deteriorates the skill for the 90% percentile. We include it in the regression model to have more degrees of freedom when designing the scenario below.

As in most cases with statistical models, a marked percentage of variance is not represented by the model. This "failure" matters if the goal of the model is to reproduce the details of a development. In the present case, however, these details do not matter; instead, all that is needed are the statistics of storm-related water level variations. The achievement of this goal is demonstrated by Fig. 8 which displays the time series of the 80% quantiles as derived from the detrended in-situ observations and as
reconstructed by the regression model (4). The differences between the in-situ data and the indirectly derived data appear as mostly irregular, while the low-frequency variations are captured remarkably well.

In order to determine a consistent scenario of expected future storm-related water levels in Cuxhaven, the mean difference field of air pressure in a paired "2 × CO₂"/"control" time slice experiment with a T106 atmospheric GCM is determined and plugged into the regression model (4). In these time slice experiments, the atmospheric GCM simulates the equilibrium response to present day sea surface temperature and sea ice distribution and present carbon dioxide concentrations. For the "2 × CO₂" experiment, SST and sea ice distributions from a simulation with a coupled low-resolution atmosphere-ocean GCM with gradually increasing carbon dioxide concentrations are determined from the time of doubled carbon dioxide concentrations at about the year 2035 (Cubasch et al., 1992). These SST and sea ice distributions are then used as specified, time constant lower boundary conditions for the T106 atmospheric GCM. Additionally, the carbon dioxide concentrations are doubled. For more details, refer to Bengtsson et al. (1995; 1995) or Cubasch et al. (1996). Beersma et al. (1997) found the wind climate in the doubled carbon dioxide world insignificantly changed when compared to the control run.

The resulting change in the air pressure distribution is plotted in Fig. 9. When fed into the regression models for anomalies of intramonthly quantiles of high tide levels in Cuxhaven a slight increase of about 7 cm for all three quantiles is found.

5. CONCLUSIONS

The storm and surge climate along the North Sea coasts has not roughened in the past hundred years. This result is consistent with findings for other European coasts and other analyses. For instance, Jónsson (1981) studied the number of "storm days" on Iceland, as defined by local observations, and found no systematic changes (cf. von Storch et al., 1994). The Koninklijk Nederlands Meteorologisch Instituut published an official assessment on the state of climate and its change for the territory of the Netherlands (KNMI, 1993). According to that report the maximum wind speeds observed during severe storms have not been increased between 1910 and today.

Thus, results derived from analysed products (weather maps) that allude to a roughening of the storm climate, seem to be misleading, since a significant part of the identified signal is due to changes in the observational, reporting and analyzing procedure.

Our study has a number of caveats. The analysis of geostrophic winds, pressure tendencies and high-frequency sea level variations covers only the near-coastal areas of Northern Europe, and no robust analysis is available for the open ocean regions. Also, one may speculate whether the link between these proxy data and the wind speeds holds for extreme wind speeds.

Our scenario for the expected time of doubled carbon dioxide concentrations points to moderate increases of surges in the German Bight. This scenario is consistent with, and within the range of, previously observed variations. As such, it is plausible. However, it depends crucially on the validity of
the driving GCM experiments; if these GCM simulations turn out to be inadequate then also our numbers are inadequate. Thus, not too much informational weight should be given to this scenario.

Also, no error bars are given for the scenarios. Such error bars are often demanded as indispensable by physicists, but can not be given with any degree of confidence. The expected error is composed of many factors, ranging from natural variability in the system to proper descriptions of the various parameterizations in the climate model used. Since the climate models are tuned to the observational record, and only one such record exists, rigorous statements about errors can not be made.

In the present analysis, the aspect of wave heights, and the reports about increasing wave heights in the past decades (Neu, 1984; Carter and Draper, 1988; Bacon and Carter, 1991; van Hoff, 1993; Hogben, 1994; Buows et al., 1996) have not been considered since a detailed publication about these matters is to be expected in near future (WASA Wave Group, 1997; see also WASA 1994 and 1995).

Acknowledgements

I thank my WASA and BMBF partners for a good and fruitful cooperation: specifically I am indebted to Viacheslav Kharin, Marek Stawarz, Hinrich Reichardt and Arnt Pfizenmayer. The T106 scenario was supplied by Lennart Bengtsson (MPI für Meteorologie). Thanks to Eigil Kaas, Torben Schmith and Hans Alexandersson for permission to use their diagrams.

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Figure 1. Ratio of synoptic scale standard deviation of air pressure variations in winter (DJF) as derived from DNMI analyses in the decade 1984–93 and in the decade 1964–73. The analyses in the marked area south of 70°N and east of 20°W seem to be relatively homogeneous. Courtesy: Viasheslav Kharin.
Figure 2. Time series of percentiles of geostrophic wind speed over Denmark. Units: m/s. Courtesy: Torben Schmith.

Figure 3. Time series of number of daily geostrophic wind speeds exceeding 25 m/s, derived from the triangle Göteborg-Visby-Lund in Southern Sweden. The solid line represents a low-pass filter. Courtesy: Hans Alexandersson.
Figure 4. Time series of percentiles of the absolute value of the 24-hour pressure tendencies over Denmark. Units: 0.1 hPa/3 hrs. Courtesy: Torben Schmith.
Figure 5. Time series of the annual mean of the sea level reported by the Cuxhaven (German Bight) tide gauge (left), and the time series of various annual percentiles (10%, 50%, 90%, 99% from bottom to top) of sea level relative to the annual mean (right). Units: cm

Figure 6. Time series of the intraannual quantiles of storm-related water level variations (defined as deviation from the annual mean) at the gauges in Den Helder and Esbjerg. The den-Helder time series has been corrected for systematic changes induced by the dam construction closing the Ijsselmer in the 1930s. Units: cm
Figure 7. Two characteristic patterns $\bar{p}_1$ (left) and $\bar{p}_2$ (right) of monthly mean air pressure anomalies over the Northeast Atlantic. The coefficients of these CCA vectors share a maximum correlation with the coefficients of Cuxhaven water level percentile patterns given in Table 1. Units: hPa

Figure 8. Time series of 80%-percentiles of intramonthly storm-related water level variations in Cuxhaven as derived from in-situ observations (solid) and estimated from the monthly mean air-pressure field through Equation 4 (dashed). Units: cm
Figure 9: Simulated atmospheric response to doubled carbon dioxide concentrations, as derived from a T106 time slice experiment. The variable shown is air pressure at sea level, given in hPa.
Extratropical Cyclone Climatology in Southern Hemisphere Winter, 
and a Comparison with Northern Winter

Chester W. Newton
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When I arrived at the National Center for Atmospheric Research in summer 1963, I found Harry, who had recently come from the South African Weather Bureau. Having been appointed head of the embryonic Synoptic Meteorology group, and feeling the obligation to display some leadership, I was immediately and shockingly confronted by the question: “What shall I do with Harry?” As it soon turned out, the question was needless; Harry knew what to do with himself. I watched as he invented the climate of the Southern Hemisphere, or so it seemed to me.

A few years and more than a dozen papers later (for Harry), the question came up in a different way, although Harry did not put it into words: “What should I do with Chester?” The answer was to put me to work on a contribution to the monograph Meteorology of the Southern Hemisphere (van Loon et al., 1972), of which Harry’s and Jan Taljaard’s contributions form the backbone. In this way, as well as through earlier exposure to his enthusiasms, Harry impelled me to acknowledge that there is another half of the global atmosphere with its own fascinations. And that was only part of my continuing enlightenment from Harry.

1. INTRODUCTION

The publication of daily surface and 500-mb maps for the International Geophysical Year (IGY) (South African Weather Bureau 1962–66) enabled the first comprehensive census of cyclones and anticyclones over the entire Southern Hemisphere (van Loon, 1965; Taljaard, 1967). This supplemented Petterssen’s (1950) statistics for the Northern Hemisphere, to give a complete global view. Prior to the IGY, analyses had been blank in high latitudes of the SH, so that the establishment of numerous surface and aerological stations around Antarctica, with several in the interior, made it possible for the first time to fill this void.

Outside the four continents and the New Zealand-west Pacific region, two-thirds of the hemisphere was still devoid of observations except for a few island stations. Procedures for coordinated surface and upper-air analyses, making fullest use of the available data, are described by Taljaard and van Loon (1964). These included the preparation of time sections of 6-hourly observations and ships’ observations at all hours, aerological diagrams at upper-air stations; frontal analysis as “an indispensable framework for obtaining plausible sequences of systems over the sparsely covered southern oceans”; and the development of a systematic scheme for thickness analysis to construct charts at 500 mb consistent with synoptic systems at the surface.
Over the decades since the studies mentioned above, numerous investigations have been carried out on both hemispheric and regional scales, based on manual and numerical analyses and on satellite imagery of cloud systems. The purpose of this review is to draw together some of the salient results of published studies. To this end, I have selected figures (for winter only) from much richer arrays in the original papers. Secondly, although physical connections are often mentioned, it is frequently necessary to turn to other sources to confirm these relationships. Accordingly, I have modified some of the original figures by superimposing a mean jet-stream axis and the locus of maximum sea surface temperature (SST) gradient. The July mean 500-mb jet axis from the climatological atlas of van Loon et al. (1971) is reproduced on all maps, so as to serve as a common frame for intercomparison. It is acknowledged that cyclone development is frequently connected with anomalous upper-air patterns, so that seasonal or monthly means only reflect the most habitual relationships.

"Cyclone frequencies" have been presented in two basically different ways that provide different information: cyclone density representing counts of events in grid boxes at synoptic times, and track density representing counts of cyclone tracks that cross a grid box. Various studies are for different periods of years, lengths of seasons, box area sizes and shapes, and sources of synoptic analyses, with counts of once to four times daily. In some cases, counts are given for the entire data sample over a number of years. For these reasons, direct comparisons of numerical values in different studies are not possible. In the figures appropriated, I have superimposed converted values in terms of a "standardized" cyclone density unit (CDU). This is defined as one cyclone (counted once daily) per month per $10^6$ km$^2$ (the area of a 9 x 9 degree latitude square or, closely, a circle with radius 5 degrees latitude). Finally, the criteria for identifying a cyclone vary, in some cases (following Petterssen) being the presence of at least one closed isobar that persists for 24h, and, in others, a specified Laplacian of pressure or geostrophic vorticity without regard for the presence of a pressure minimum. While objective, the results are sensitive (especially for the forming or dying stages of cyclones) to the criteria chosen and to the manual or numerical model analyses used.

2. CYCLONES IN NORTHERN HEMISPHERE WINTER

The general distributions of cyclogenesis and cyclone density over the Northern Hemisphere (NH) in winter are shown in Fig. 1. Petterssen (1956, 266–277) categorized the principal features of cyclogenesis as a zonal pattern over the oceans, generation in the lee of mountain ranges, and the formation of lows over bays and inland water bodies. Over both oceans, maximum cyclogenesis (Fig. 1a) is closely associated with the axis of maximum mean wind at 500 mb. Most prominently over the eastern oceans, greatest cyclone density (Fig. 1b) is observed well poleward, and maxima of track density (dashed lines) are intermediate.

The most striking orographic effect is the abrupt decrease of Pacific cyclones off the North American west coast, with regeneration in the lee of the Rocky Mountains. Petterssen remarks that "Neither of these mountain maxima [Fig. 1b] show much extension downwind, indicating that many of the
centers that form in this way remain local or are of short life, although some of them develop and travel across the North American continent." Chung et al. (1976; see also Newton 1996) found that, among lows that formed in the lee of mountain ranges both in North America and East Asia, about half developed into moderate to intense cyclones that had long trajectories. All of these were associated with the advance of pre-existing well-developed upper troughs, and developed rapidly. Of the rest, which either remained weak or did not leave the lee slope, none "was directly associated with an advancing upper trough at any time during its life span." In contrast to the well-defined association between the latitudes of peak cyclogenesis and of strong 500-mb wind over the Asian and North American east coasts, this relationship is ill-defined over the meridionally extensive Rockies where the mean winds are weak (Olson et al., 1983). They note that "the lack of a mean wind maximum does not preclude the existence of occasional [polar] jet maxima propagating through this region and aiding cyclone development."

The patterns of cycloysis, cyclogenesis, and track density derived by Zishka and Smith (1980) clearly show that cyclones forming in the lee of the Rocky Mountains dominate cyclone activity over all of North America except near the east coast. They point out that the prominent maximum of cyclones over the Great Lakes region (Fig. 1b) is not due to cyclogenesis there, but results from the confluence of tracks of cyclones that originate in the lee of the Canadian and "Colorado" Rockies.

3. THE SETTING FOR SOUTHERN HEMISPHERE CYCLONES

3.1. The Earth's Surface in Winter

Figure 2a shows the orography of SH continents and the mean SST in August (Taljaard, 1972a). It will be convenient to allude to two main longitudinal sectors wherein the topographical controls are distinctly different. These will be called the Atlantic–Indian Ocean (AIO) sector from South America eastward to eastern Australia, in which are situated the continents including the bulk of Antarctica; and the Pacific Ocean (PAC) sector in which the extent and elevation of Antarctica are comparatively small. These features control the marked asymmetry of the mean atmospheric structure and, in turn, the distribution and character of extratropical cyclones.

The effective eccentricity of Antarctica is augmented by the greater extent of pack ice around the AIO sector, especially in the mid-Atlantic as a result of discharge of cold water and ice by the Weddell Sea gyre and the east-northeastward deflection of the Antarctic Circumpolar Current through the Drake Passage. As a combined effect of the low latitude of the pack-ice boundary and the southward and westward intrusion of the warm Agulhas Stream, the SST gradient (Fig. 2b) is most concentrated in a band from south of Africa to the central Indian Ocean. Around the PAC sector, the SST gradient at a corresponding latitude is weak, and the separate concentrations in subantarctic and subtropical latitudes are significant for the cyclone distribution.
3.2. The Mean Structure of the Atmosphere

The remarkable response of the mean atmospheric structure, including its annual and semiannual variations, to the distribution of sea-surface temperature over the SH, has been emphasized by van Loon (1965; 1972a; 1972b). Isotachs at 500 mb (Fig. 3a) show two discrete maxima, both highly asymmetrical. The subtropical jet stream (STJ), which is present at 200 mb all around the hemisphere and is intense downstream of the southward upper-level branch of the Australasian monsoon circulation, is prominent at 500 mb only from the central Indian Ocean to the mid-Pacific. The midlatitude polar jet (PJ) is distinctly fastest over the SW Indian Ocean where the SST gradient is most concentrated. Around the PAC sector where both the STJ and PJ are present, zonal winds are notably weak in midlatitudes, especially in the Australia–New Zealand region.

As a measure of upper-tropospheric wave activity, Physick (1981) demonstrated for the winter of 1979 that the variance of the v-component of the 300-mb wind (band-pass filtered to admit synoptic time scales) is very much greater around the AIO than the PAC sector. A similar result (Fig. 3b) was derived by Trenberth (1991) from a multi-year series of ECMWF analyses. Around the AIO sector, eddy activity is vigorous and the core of the "storm track" closely coincides with the mean jet axis. This feature contrasts sharply with the western PAC sector where eddy activity is feeble, especially in midlatitudes.

Vertical sections in Fig. 4 (van Loon, 1972c) emphasize the asymmetry of the hemispheric wind system in which, in midlatitudes, the vertical shear in the longitude of New Zealand (Fig. 4a) is about half that over the SW Indian Ocean (Fig. 4b). This has consequences for the types as well as the occurrences of cyclones in the two regions. In summary, it is evident that all of the "environmental" features illustrated in Figs. 2–4 are far more favorable to baroclinic cyclone development over the AIO than the PAC sector. The belt of enhanced SST gradient (Fig. 2b) is not only most concentrated over the SW Indian Ocean but also lies at lowest latitude. The combination of great baroclinity, conditional instability and accessibility to moist tropical air thus favor baroclinic instability. The concurrence of maximum eddy activity in the storm track (Fig. 3b), over the same region, reflects the frequent passage of upper troughs whose vorticity advection can initiate disturbance development.

4. CYCLONE AND FRONT FREQUENCIES DURING THE IGY

 Frequencies of surface fronts and cyclones during winter of the IGY are shown in Fig. 5. Although in summer a simple belt of frontal frequency maximum (FFM) encircles the hemisphere in midlatitudes, in winter (Fig. 5a) there are two zones (van Loon, 1965). "One has the shape of a spiral winding from southernmost Brazil, across the South Atlantic and Indian Oceans to high latitudes in the Pacific. The other begins in low latitudes over the South Pacific Ocean and ends in the South Atlantic." For the most part, the FFM "coincides with the strongest temperature gradient in the ocean surface...." The high frontal frequency at low latitudes over the central Pacific is "likely a consequence of the deformation field created between the frequent anticyclones in middle latitudes in the vicinity of New
Zealand [Fig. 5c] and the subtropical highs which are commonly situated in the eastern part of the ocean. Part of the reason...may lie in the tendency for fronts to become stationary in this area...

Taljaard (1967; 1972b) presented analyses of cyclone and anticyclone densities over the SH, and profiles of their genesis by latitude, for winter, summer and the transitional seasons (spring and autumn combined). The distribution of cyclones in winter is shown in Fig. 5b. Taljaard cautioned that the daily map sample (7 months) should not be construed as representative of long-term climatology, and that “over large parts of the southern oceans the reliability of these maps is medium to poor.” Notwithstanding, the major features derived have well stood the test of time.

Anticyclones are arranged in a “well defined belt of high frequency which encircles the hemisphere” just south of the mean subtropical ridge (SR), with occurrences “near zero over much of the southern oceans between about 45°S and the Antarctic coast.” The “zone with most frequent cyclogenesis falls midway between the zones with highest anticyclone and cyclone frequencies”, the latter being located near the mean circumpolar trough (CT). “Spiral arms” of enhanced cyclone frequency extend from South America and the Tasman Sea toward the “graveyards” around Antarctica; “the Antarctic Ocean is a region with many mature and occluded cyclones rather than the cradle of new systems.”

In general, there is a consistent relationship between the frontal frequency maxima and the “spiral arms” in the cyclone density distribution. The significant exception is the west Pacific where, over the Tasman Sea, there is a prominent maximum of cyclones with a minimum of fronts. As discussed by Taljaard (1972b), many of the lows over the eastern Australia–New Zealand region originate not as waves on fronts, but rather are induced by upper-tropospheric “cut-off” cyclones that develop equatorward from the polar jet. Being cold-cored, the upper-level lows outnumber surface cyclones. They occur in blocking situations, which are most common in these longitudes, and the cut-off lows are paired with (situated equatorward of) surface anticyclones that move along the southern track in Fig. 5c (cf. Taljaard, 1972b, Fig. 8.39).

5. SATELLITE-OBSERVED CLOUD VORTICES: LIFE CYCLE STAGES

As remarked by Troup and Streten (1972), “The advent of satellite observations has effected a considerable improvement in synoptic analyses for the [SH oceans], enabling systems to be tracked that would otherwise have gone unobserved.” Making use of digitized visual mosaics, they introduced a classification scheme for the evolution of cloud vortices and, from conventional observations coincident with particular stages, derived the associated mean patterns of sea-level pressure and 500- and 300-mb geopotential. These were expressed in terms of local anomalies from the monthly climatological atlas data of Taljaard et al. (1969).

West-east profiles of these quantities, and their sequential values in relation to the forms and ages of cloud vortices (for summer plus transition seasons), are given by Streten and Troup (1973).
their classification scheme (Fig. 6a), two initial forms (W and A) are recognized, both of which may evolve into the later forms B to D. The "wave development" form (W) is seen as "localized thickening of a cloud band" and the "comma cloud" form (A) as early vortex development either "merged with a major cloud band or in isolation." These initial forms correspond generally to the distinctive modes of cyclogenesis described by Petterssen and Smeybe (1971): baroclinic instability on a frontal zone without a pronounced upper trough initially; or development of a low-level circulation as a result of strong upper divergence downwind of an initially well-developed trough, with or without a front initially but with increasing baroclinity as the storm intensifies. For both modes of cyclogenesis, "The end result of the development is a thermal structure that resembles the classical occlusion."

Making use of infrared images, Carleton (1979) derived the climatology of satellite vortices for winters (JJAS) of 1973–77. Latitudinal distributions according to type are shown in Fig. 6b. As found by Streten and Troup for the other seasons, there is a poleward progression of peak frequency with development stage. The two modes of initial formation are concentrated at very different latitudes, with peak occurrences of "wave" vortices (W) near 40°S and of "comma" clouds over the belt 45–60°S.

The distribution of (W,A,B) vortices, stages leading up to the "mature" phase, is shown in Fig. 7a, along with locations and densities of most frequent "dissipating" (D) vortices. For the most part, these locations correspond well with the maximum cyclone densities in Fig. 5b, corroborating Taljaard's (1967) analysis and his interpretation of these sites as cyclone "graveyards". Carleton notes that although there was strong year-to-year variability of vortex frequencies in the subantarctic centers, they "show a tendency to persistence with little shift in location from one winter to another." The distribution of "cut-off" (F/G) vortices (Fig. 7b) shows that these are common over the Australia-New Zealand region and that they diminish eastward, with frontal and comma vortices dominating near the same latitude over the eastern Pacific. As shown by all studies, the frontless vortices are numerous only in winter.

6. CYCLONE CLIMATOLOGY FROM NUMERICAL ANALYSES

6.1. General Remarks on Methods

Synoptic analyses by numerical models enable the construction of cyclone tracks by totally automated procedures. They present the great advantage that, for very large data sets, cyclone and anticyclone locations at individual times, along with characteristics such as central pressure and vorticity, can be recorded and retrieved for any desired subset. While objective, the derived statistics are sensitive to the criteria specified for counting a perturbation as a cyclone, namely a threshold intensity and a minimum lifetime.

Using versions of a cyclone finding and tracking scheme devised by Murray and Simmonds (1991), but with different tests for identifying cyclones, Jones and Simmonds (1993) and Sinclair (1994, 1995) derived multi-year climatologies from numerical analyses by the Australian Bureau of Meteorology and by ECMWF, respectively. The Jones and Simmonds (JS) threshold criterion is expressed in terms
of $\nabla^2 p$, and Sinclair's (SI) as a minimum magnitude of geostrophic vorticity. Because of the latitude variation of coriolis parameter, these onset criteria result in greatly different cyclone distributions with, in JS, an apparent rejection of weak systems at lower latitudes and concentration of cyclogenesis and cycolysis around Antarctica. On the other hand, as discussed by SI, his low threshold ($|\mathcal{\zeta}| \geq 0.2 \times 10^{-4}$ s$^{-1}$) admits an excessively large population of vorticity maxima. Vorticity calculations are sensitive to analysis errors, of this magnitude, and the methods (as shown by example) "may have a tendency to include elongated geostrophic shear or curvature zones not generally thought of as cyclones."

6.2 Statistics of "Mobile Cyclones"

Sinclair (1994) derived seasonal cyclone statistics from ECMWF analyses for seven years (1980–1986). Figure 8a shows the density of all vorticity centers in winter (JJA) with $|\mathcal{\zeta}|$ exceeding the threshold magnitude $0.2 \times 10^{-4}$ s$^{-1}$. In this count, individual centers were not required to be formed into tracks, and the average number per analysis time was 45. "The maxima over the three midlatitude SH land-masses dominate the statistics during both seasons...the fact that these maxima straddle elevated terrain suggest that lee troughs within alternating easterlies and westerlies may contribute to these maxima."

Sinclair demonstrated that the pattern of cyclone density around Antarctica closely resembles the Laplacian of terrain height and of pressure reduced to sea level, leading to the conclusion that "many of the cyclone centers analyzed are likely spurious." Also most of the lows in the vicinity of South America and Africa are evanescent and/or spurious, tending to be nearly stationary. On these grounds, he applied a "mobility test" that excluded all centers that moved a distance less than 10° latitude from further consideration (more than half the population in Fig. 8a). According to Sinclair, the procedure does not exclude slow-moving cyclones that "have formed and deepened in midlatitudes...or lee cyclones that eventually moved offshore...." Note, however, that the centers of high concentration around Antarctica in Fig. 8a, eliminated in the process (possibly due to rejection when cyclone movements become slow or irregular), are identical with the high-density "graveyards" in Fig. 5b and concentrations of dissipating cloud vortices in Fig. 6b.

After eliminating nonmobile centers, the "mobile cyclone" density (Fig. 8b) shows an arc around the AIO sector in latitudes 40–60°S. The pattern over the PAC sector is distinctly different, with "two bands—one near 40°S and the other near 65°S—separated by a zone of minimum cyclone occurrence southeast of New Zealand." The band near 40°S is clearly defined from Australia to the central Pacific where cut-off lows are frequent in winter, as noted in connection with Figs. 5b and 7b.

While this band of cyclones is clearly situated in a region of weak mean winds aloft, the circumhemispheric ring at middle latitudes over the AIO sector and continuing poleward of 60°S over the PAC sector, appears to be loosely related to the mean polar jet. This relationship becomes more definitive when the population in Fig. 8b is stratified into open cyclones (those with vorticity maxima but no closed isobars, in the numerical analyses) and closed cyclones. Open cyclones (Fig. 9a) are fairly
well related to the PJ and maximum SST gradient, whereas closed “mobile cyclones” (Fig. 9b) are concentrated in a ring well to the south.

The locus of maximum average speeds of cyclone movement (not shown) is congruent with the polar jet axis at 500 mb, with the fastest movements (> 15 m s⁻¹) over the south Indian Ocean. As a product of cyclone speed and density (Fig. 8b), the track density (Fig. 10a) is a measure of the frequency of passage of cyclone centers within 5° latitude of a given point. Sinclair notes the “excellent agreement with baroclinic storm tracks obtained by Trenberth” (cf. Fig. 3b). In this light the close similarity of Fig. 10a, with the distribution of open centers in Fig. 9a, illustrates the association of upper-level eddy activity with cyclones in their earlier stages with little contribution from those in their senescent stages (Fig. 9b). For a broader population of cool-season cyclones (MJIASO) “genesis locations” are shown in Fig. 10b, defined (Sinclair 1995) as “all first track points weaker than [\(\bar{\omega} = 0.5 \times 10^{-4} \text{s}^{-1}\)] for cyclones attaining that intensity.” For this subset, the “average life span was about 4 days.” Sinclair observes that cyclogenesis (in the meaning of intensification), as well as genesis, “appears to be correlated with the strongest SST gradients. This may be related to the transfer of oceanic baroclinity to the atmosphere. In addition, strong sensible and latent heat input to the atmosphere occurs where cold air moves rapidly across a strong SST gradient. The resulting low-static stability can further augment cyclogenesis. The locations of both cyclone formation and of cyclogenesis...are also correlated with the climatological positions of the upper-tropospheric jet streams.” He suggests that “the genesis maximum adjacent to the Argentina coast near 45°S occurs year round...is more likely associated with frequent lee cyclogenesis in persistent westerly flow...analogous to the high frequency of newly formed cyclones observed year round east of the [North American] Rockies...”

The distribution of cyclone “genesis” in Fig. 10b is concentrated in four discrete regions, each of which appears to have its own special characteristics:

a) **West Atlantic.** Cyclones that form to the east of the southern Andes cordillera on the approach of upper-level troughs on the jet stream, and develop further (as frontal cyclones) over the region of strong SST gradient. It seems likely that many of the “genesis” events can (as in the North American Rockies) be considered as regenerations of systems that have dissipated off the west coast (cf. fig. 8b) but whose upper-level troughs and surface fronts have continued eastward. This is the site of most frequent genesis in summer as well as winter.

b) **South Indian Ocean.** Generation by baroclinic instability induced by upper troughs, in the sector where both the SST gradient and upper-tropospheric winds are strongest (cf. Figs. 2b, 3a). As shown by track densities at the ensuing stages of intensification, maturity and cyclolysis (Sinclair, 1975, Figs. 7–9), this source contributes to the large population of cyclones SW of Australia (Fig. 8b) and also to the cyclone density at high latitudes over the South Pacific.
c) SE Australia-Tasman Sea. Predominantly frontless cyclones that form beneath cut-off upper lows that develop from the polar jet, situated in the latitudes of weak baroclinity between the mean PJ and STJ but close to the latter (virtually absent in summer).

d) Central-Eastern Pacific. Wave cyclones that form in subtropical latitudes where the frontal frequency (cf. Fig. 5a) and SST gradient are maximum. As with Indian Ocean cyclones, genesis takes place over warm waters (~12–16°C) where the mean moist static stability is nearly neutral. Sinclair’s maps at later stages indicate that although this population of cyclones migrates 10–15° poleward, it remains separate from the high-latitude belt with “mature” cyclones at 40–50°S over the easternmost Pacific.

Trenberth (1991) remarks that “activity spawned in the South Pacific Convergence Zone...coincides with the enhancement in storm track activity...” over the east Pacific (cf. Fig. 3b). This correspondence, evident in Fig. 10, is consistent with the different nature of group D cyclones (baroclinic and eddy-generating) as compared with those (C) over the western Pacific.

7. OVERALL CYCLONE ACTIVITY AROUND THE HEMISPHERES

It is of some interest to summarize the cyclone activity of the hemispheres in terms of variations, with longitude, of total numbers of cyclone passages summed over the whole of extratropical latitudes (called cyclone flux below) in relation to cyclogenesis and cyclolysis. For reasons mentioned in the Introduction, numerical values (and detailed patterns) differ among various studies, but at the same time they offer mutual confirmation as to the broad features.

7.1 Northern Hemisphere

Curve (a) in Fig. 11, derived from tabulations in 5° lat.-long. boxes (Whittaker and Horn, 1982), shows (right-hand scale) the total counts of cyclone tracks between latitudes 20°–85°N in January. Curve (b), also from their tables, gives the monthly count, per 10° longitude interval, of origin points of cyclone paths (first closed isobar). Curve (c) is derived as a residual, by subtracting the latter from the difference of track count across the longitude interval. In map (d), the main cyclogetic regions and principal cyclone tracks (maximum track density), along with (dotted lines) the axes of maximum cyclogenesis over the oceans in Fig. 1, are shown against the background of the January SST distribution derived by Shea et al. (1990).

Taylor (1986) demonstrated that, where cyclone paths deviate from a zonal direction, counts of tracks crossing latitude-longitude grid boxes may considerably exceed the actual track density because the effective cross section is greater. From samples in regions of high track density, he found a typical overcount by 14%. In certain regions such as the Gulf of Alaska where many cyclones turn northward and some westward, the inflation can be much greater. Also, in a census of the cold season over the Pacific, Gyakum et al. (1989) found that 37% of cyclone centers analyzed lasted less than 24-h. As a
guess (based partly on the consideration that the origin count in Fig. 1b considerably exceeds the
cyclogenesis frequencies found by other investigators for the regions indicated by bars), the counts on
the right-hand scales have been adjusted downward by 25% to give the values on the left side.

The main features of Fig. 11a are steep eastward increases of cyclone flux over East Asia and
the western Pacific and over western North America, with a minor increase over western Europe due to
Mediterranean cyclogenesis. East of the maxima, steady declines are observed across the oceans and
across the Eurasian continent, with an overall threefold decrease from the West Atlantic to central Asia
at the longitude of the Siberian High. The eastward increases are associated with elevated frequencies
of cyclogenesis in the longitudes of eastern Asia—western Pacific and western North America—western
Atlantic, but a principal factor accounting for the steep rises is the small degree of cycloysis in these
longitudes. The majority of cyclones there are “survivors” that, having developed to considerable
intensity, have long lives. For example, the virtual absence of cycloysis over North America between
85°W and 120°W is shown by Zishka and Smith (1980, Fig. 2c). On the other hand, the coincidence of
peak cyclogenesis with a maximum of cycloysis near 75°W (marked A) is accounted for by overlapping
cyclone tracks, with generation almost entirely along the east coast and dissipation over the continent
from New England northward as shown by their maps. (The surviving cyclones dissipate to the west of
Greenland, accounting for the cycloysis maximum at 55°W.) A similar feature appears off the east
coast of Asia.

Across the oceans to the east of the regions of most active generation, cyclogenesis continues at
moderate levels but is apparently exceed by cycloysis. As confirmed by Gyakum et al. (1989) for the
cold season over the Pacific, the curves in Fig. 11b, c mainly reflect cyclogenesis around 30°–40°N and
cycloysis in the belt 50°–60°N, of storms that intensify over relatively warm waters (especially the
Kuroshio Current) in latitudes of frequent jet streams and migrate over cold waters (Fig. 11d). For
comparison with January, the cyclone flux over the Asiatic region for April is shown in Fig. 11a. The
marked westward shift is related to great increases of both orographic and coastal cyclogenesis,
especially over the lower Yangzi Valley and East China Sea SW of Japan (Chen et al., 1991). The
combined effect (Fig. 11b), along with a decrease over the open ocean, is to shift the maximum of
cyclogenetic activity westward.

Broad-scale upper-level features (at 500 mb) related to surface cyclone activity are summarized
schematically in Fig. 11d, for activity associated with the main belt of extratropical westerlies. (The
Mediterranean region is omitted, where, as for group “C” of SH cyclones in Fig. 10b, cyclones are
commonly associated with cut-off lows that evolve from deepening cold troughs, although the surface
cyclogenesis processes are different. “TB” designates the longitudes of most frequent trough “births”
during cold seasons (Sanders, 1988). These “lie over and downwind from the two major mountain
masses” and “may actually represent reappearances of features temporarily hidden” when
“superimposed on a larger quasi-stationary ridge...The great majority of crossings of mountain barriers by
mobile troughs did not result in a break in their continuity, despite weakening in transit.”
Referring to Fig. 1e, Rocky Mountain cyclogenesis occurs at locations dependent on the latitude of the jet stream at the time. Cyclogenesis over East Asia, by contrast, is highly concentrated in the lee of the Altai-Sayan ranges which extend only about 1000 km north-south; and is ordinarily induced by troughs connected with dissipating cyclones that have traversed the continent along the track near 60°N (Chen et al., 1991). As interpreted by Sanders (1988), coastal cyclogenesis is generally induced by upper troughs that emanate from (or pass through, in a weakened state) the TB regions, growing in the larger scale northwesterly flow as they approach the longitudes of the mean troughs (MT). “UCG” and “UCL” indicate the longitudes in which upper-level cyclogenesis and cycloysis are observed in winter, according to a climatology of 500-mb closed lows by Bell and Bosart (1989), the heavier bars corresponding to greatest frequencies. Initiations of these appear a short distance east of the lee cyclone regions and continue across the oceans, terminating over their eastern parts where trough “deaths” are also most concentrated according to Sanders (1988).

For the hemisphere as a whole during one cold season, Sanders found that 8–15 upper-level troughs were present on a given day. They had a median lifetime of 12 days (much longer than that of a surface cyclone), an average zonal phase speed of 13 m s⁻¹, and a mean zonal wavelength of about 2500 km. This corresponds to 13.5 wave passages across a meridian per month, roughly commensurate with the mean frequency of cyclone passages in Fig. 1a although a one-to-one relationship is not to be expected.

7.2 Southern Hemisphere

the cyclone flux in Fig. 12a was obtained by summation, over all latitudes, of the track densities in Fig. 10a. The flux is quite uniform around the hemisphere, except for minima in the longitudes of Australia and South America. Corresponding breaks in cyclone density, between the AIO and PAC sectors, are seen in Fig. 8b. Figure 12b shows, as proxy for surface cyclogenesis, frequencies of initial points of satellite-observed wave vortices (W) and of the total including comma vortices (A). This was transformed, with smoothing, from Fig. 3 of Carleton (1979) using the numbers of vortices given in Fig. 6b. The three genesis peaks are well associated with eastward increases of cyclone flux; over the Indian Ocean, leading to the maxima of cyclone density and flux SW of Australia (Figs. 8b, 10a); and east of Australia and South America restoring the declines of cyclones across the eastern Indian Ocean and Pacific Ocean.

Longitudes of the major genesis regions in numerical analyses (Fig. 10b) are outlined in Fig. 12d and indicated by the top bars of Fig. 12c. Succeeding bars correspond to maximum track densities for each of these cyclone clusters (Sinclair 1995, Figs 7b-9b) at the stages of intensification (24-h vorticity increase exceeding a specified threshold), maturity (maximum vorticity), and dissipation (decrease exceeding a threshold). Solid portions of these bars indicate longitudes in which cyclones go through their life cycles mostly in middle latitudes, and the dashed continuances correspond to cyclones that have migrated to subantarctic latitudes (most notably south of Australia).
Evidently groups A and B, originating over the AIO sector, contribute most to the high-latitude ring of cyclones surrounding Antarctica. Group C cut-off cyclones run their course in midlatitudes. The behavior of group D cyclones is more complex, depending on the character of the upper-level flow patterns across the Pacific in individual episodes; crossing South America near 50°S when the upper flow is relatively zonal and, when deep long-wave troughs are present over the Pacific, taking southeastward paths that join the main cyclone track through the drake Passage (to its west). Encircled crosses in Fig. 12d indicate locations where satellite cloud vortices are most frequent (Fig. 7a) or the "graveyards" in Fig. 5b, where cyclones (having diagonally crossed the cyclone track in Fig. 12d into the subantarctic region of weak zonal wind) come to rest.

The general level of cyclone flux over the SH (Fig. 12a) is remarkably lower than that over the NH, corresponding more nearly to the minima over the eastern oceans in Fig. 11a. It seems plausible to attribute the greater zonal mean flux over the NH to the augmentations by mountain-lee and east coast cyclogenesis, with persistence of much of the resulting cyclone activity eastward from these sources.

8. CONCLUDING REMARKS

Taken together, the studies summarized above support the finding of van Loon and of Taljaard, which stimulated an era of increased understanding of the general circulation of the Southern Hemisphere. While they fully credited the works of their contemporaries and scientific ancestors and built upon them, their own prominence is assured by their having produced a coherent description of the working parts and their interrelations over the full hemisphere. This is a tribute to their meticulous and imaginative combination of skillful analysis and sound interpretation of its meaning. Although later analyses have greatly elaborated the observed features, these have confirmed the main results.

For both hemispheres, the principal features of cyclone activity are now well established. The detailed patterns, however, differ strikingly among some investigations. This problem arises from several causes, primarily extensive observational voids, analysis procedures for synoptic data, and criteria for identifying cyclones and cyclogenesis.

Sinclair (1994, Fig. 16) illustrates a striking increase of "cyclone" numbers during the years he analyzed, owing to changes of the ECMWF model. As regards the uncertainties of identifying surface cyclones, he remarks that "A similar derivation of cyclone statistics for a higher level (say 700 or 500 hPa) would eliminate the problems posed by shallow topographic perturbations and meaningless pressure reductions to MSL." Also, comparison with tracks of "vorticity centers observed in satellite imagery...would provide an independent measure of analysis skill and avoid the 'incestuous' practice of evaluating model prognoses against model analyses."

One of the agonies encountered in comparing the various studies has been the specification of cyclone frequencies in different ways. Also, difficulties in interpretation arise from the question of what the population includes, such as the relative dominance of weak or strong and persistent systems in the
regional distributions. Although this question has not been entirely neglected, to paraphrase Orwell in Animal Farm, "All cyclones are equal, but some cyclones are more equal than others."

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References


Figure 1. (a) Cyclogenesis density and (b) cyclone density in winter (DJF) of years 1899–1939 (Petterssen 1950). For conversion to “cyclone density units” (CDU = cyclones per month per $10^6$ km$^2$), multiply by 3.3. Solid arrows, added, axes of mean maximum wind at 500 mb in January (Olson et al., 1983); dashes in (b), maximum cyclone track density based on isopleths by Whittaker and Horn (1984).

Figure 2. (a) Mean sea-surface temperature in August, and smoothed topography of continents: light stippling > 500 m, medium > 1500 m, dark >3000 m. (Composite of figures in Taljaard 1972a). (b) Winter mean SST gradient, °C per 1000 km (Sinclair 1995).
Figure 3. (a) Zonal component of geostrophic wind (m s\(^{-1}\)) at 500 mb in July (van Loon 1972c). (b) Standard deviation (m s\(^{-1}\)) of 2–8 day band-pass filtered meridional wind at 300 mb in July (Trenberth, 1991). Where added to these and ensuing figures, heavy dots show locus of maximum SST gradient (from Fig. 2b); and bold arrows show 500-mb geostrophic wind axes in July, solid where mean wind speed > 20 m s\(^{-1}\) (from Fig. 3a).

Figure 4. Zonal geostrophic wind (m s\(^{-1}\)) in July along meridians (a) 170°E and (b) 50°E, indicated by dashed lines B and A in Fig. 3a (van Loon et al., 1971).
Figure 5. Frequencies of surface fronts, cyclones, and anticyclones during winter of the IGY. (a) Number of times in 100 days when part of a front was situated within a unit area of 400,000 km$^2$(van Loon, 1965). (b) and (c) Cyclone and anticyclone density. Heavy lines SR and CT are the mean subtropical ridge and circumpolar trough (Taljaard, 1967). Units are given by the inset, with values in CDU to left.

Figure 6. (a) Primary classification of "extratropical vortex evolutionary patterns" in satellite cloud images (Streten and Troup, 1973). (b) Percentage distribution of each type for five winters 1973-77 (Carleton 1979). Total counts for the five years (twice daily) are given in parentheses. "F/G" distribution is for frontless cut-off cyclones, with satellite patterns like $D_x$ and $D_y$ without the cloud band.
Figure 7. Densities (relabeled in CDU) of (a) combined initial-stage (W, A) and "late formative" (B) cloud vortices; and (b) frontless cut-off (F/G) vortices (Carleton 1979). Locations and densities of most frequent "dissipating" vortices are shown in (a).

Figure 8. Densities of (a) all vorticity centers passing threshold $|\zeta| > 0.2 \times 10^4$ s$^{-1}$ and (b) the total subset of "mobile cyclones" with track lengths exceeding 10 deg. lat. Derived from ECMWF analyses for winters (JJA) of 1980–86 (Sinclair, 1994). Information on number of tracks and cyclones selected has been deleted and the figure inverted to the same orientation as other figures. Original units have been left on; CDU magnitudes in whole digits are shown in a few places.
Figure 9. Cyclone densities for (a) open and (b) closed cyclones (Sinclair 1994), with a few CDU magnitudes.

Figure 10. (a) Track density for all "mobile cyclones" in Fig. 8b passing within 5 deg. lat. of a given location (Sinclair, 1994). Multiplication by 0.012 gives track density in numbers of tracks per 10° latitude per month, as indicated in places. (b) Density of "genesis" locations for a subset of cool-season (MJNASO) cyclones (specified as first track points having $|\mathbf{v}| > 0.5 \times 10^{-4}$ $\text{s}^{-1}$, for cyclones that exceed this intensity at some time in their lives). (Sinclair, 1995).
Figure 11. Frequencies of (a) cyclone flux across meridians (between 20°-85°N) in January, and (b) cyclogenesis per 10° longitude sector, derived from 5° lat.-long. grid tabulations by Whittaker and Horn (1982). A 1-2-1 smoother has been applied. (c) Cyclolysis frequency. Dashed curves show frequencies over East Asia for April (orographic and coastal contributions, dots and crosses). (d) January SST climatology (Shea et al., 1990), with 3-km envelope of Tibetan Plateau (TPL). Main cyclogenetic regions (thin outlines) and simplified cyclone tracks are based on Whittaker and Horn (1984), Zishka and Smith (1980) and Chen et al. (1991). Dotted lines (CG) indicate belts of oceanic cyclogenesis, and dashed arrows suggest the dispersal of cyclones toward the concentrated tracks.
Figure 12. (a) Cyclone flux across meridians (20°–90°S) in SH winter, summed from track densities in Fig. 10a. (b) Genesis frequency (events per 10° longitude per month) of satellite observed cloud vortices. Based on Figs. 3 and 4 of Carleton (1979). (c) Longitudes occupied by cyclones of groups A–D in Fig. 10b at stages of their life cycles. The bars correspond to the shaded regions in Fig. 10b and ensuing figures by Sinclair (1995). Solid bars indicate cyclones in the midlatitude belt, and their dashed continuances, offshoots into the subantarctic ring of cyclones. (d) July mean SST isotherms and (hatched) pack ice (Shea et al., 1990). Heavy arrows, track density maxima from Fig. 10a (solid where core densities are greatest); thin lines; cyclogenesis regions in Fig. 10b; crossed circles, maxima of “dissipating” cloud vortices (Carleton, 1979), which appear in nearly the same longitudes as the “graveyards” of Fig. 5b.