Harry van Loon Symposium
Studies in Climate II

CLIMATE AND GLOBAL DYNAMICS DIVISION
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Harry van Loon Symposium

Studies in Climate II

Editors R. A. Madden, E. C. Stephens,
J. W. Hurrell, G. N. Kiladis, G. A. Meehl, D. J. Shea

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.
<table>
<thead>
<tr>
<th>Author</th>
<th>Title</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>van Loon</td>
<td>The Annual Range over the Oceans Revisited</td>
<td>1</td>
</tr>
<tr>
<td>J.S. von Storch</td>
<td>On the Ability of a General Circulation Model to Simulate Two Large Scale Variations in the Southern Hemisphere</td>
<td>15</td>
</tr>
<tr>
<td>Trenberth</td>
<td>General Circulation Studies Using NCEP/NCAR Reanalyses</td>
<td>30</td>
</tr>
<tr>
<td>Shea</td>
<td>The Potential for Long-Range Prediction of Precipitation Over India for the Southwest Monsoon Season an Analysis of Variance Approach</td>
<td>44</td>
</tr>
<tr>
<td>Rogers</td>
<td>On the Causes of Mild Winters in Northern Europe</td>
<td>51</td>
</tr>
<tr>
<td>Rasmusson</td>
<td>Is there a Warm Season Relationship Between Precipitation Over the United States and Tropical Pacific Sea Surface Temperature?</td>
<td>68</td>
</tr>
<tr>
<td>Raphael</td>
<td>Quasi-Stationary Waves in the Southern Hemisphere of a GCM with and without an Interactive Ocean</td>
<td>76</td>
</tr>
<tr>
<td>Mo</td>
<td>Planetary Waves in the Southern Hemisphere and Linkages to the Tropics</td>
<td>90</td>
</tr>
<tr>
<td>Milliff</td>
<td>Quasi-Stationary Waves in the Southern Hemisphere: Revisiting a Theory due to Harry van Loon</td>
<td>107</td>
</tr>
<tr>
<td>Meehl</td>
<td>A Modulation of the Mechanism of the Semianual Oscillation in the Southern Hemisphere</td>
<td>122</td>
</tr>
<tr>
<td>Madden</td>
<td>Semi-Diurnal Variations in the Budget of Angular Momentum in a General Circulation Model and in the Real Atmosphere</td>
<td>138</td>
</tr>
<tr>
<td>Leder</td>
<td>The Effect on the Lower Stratosphere of Three Tropical Volcanic Eruptions</td>
<td>153</td>
</tr>
<tr>
<td>Large</td>
<td>Semiannual Oscillation in Models and Observations of the Southern Ocean</td>
<td>166</td>
</tr>
<tr>
<td>Labitzke</td>
<td>The Signal of the 11-year Sunspot Cycle in the Upper Troposphere-Lower Stratosphere</td>
<td>176</td>
</tr>
<tr>
<td>Kiladis</td>
<td>Interannual Variability of the South Pacific Circulation Related to the Southern Oscillation</td>
<td>197</td>
</tr>
</tbody>
</table>
Jenne
The Southern Hemisphere Climatology Project
An Historical Perspective ................................................................. 213

Jaegar
Linkages Between Scientific Expertise and Policy Development in the Climate Change Case ..................................................... 220

Hurrell
Decadal Variations in Climate Associated with the North Atlantic Oscillation ................................................................. 224

Chen
An Observational Study of the Semiannual Oscillation in the Tropics and Northern Hemisphere ............................................. 246

H. von Storch
Storm and Surge Climate in the North Sea Area:
Changes in the Past Century ............................................................. 252

Newton
Extratropical Cyclone Climatology in Northern and Southern Hemisphere Winter, and a Comparison with Northern Winter .......... 269
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

vi
WHAT HAS HARRY BEEN UP TO?

NUMBER OF PUBLICATIONS

NCAR YEARS

Projected
Actual
Publications


---, 1955: Mean air temperature over the southern oceans. *Notos*, 4, 292–308.


—, and —, 1989: Recent work correlating the 11-year solar cycle with atmospheric elements grouped according to the phase of the quasi-biennial oscillation. *Space Science Reviews*, 49, 239–258.


—, 1991: A review of the surface climate of the Southern Hemisphere and some comparisons with the


Meteorologische Zeitschrift, Neue Folge, 5, 166-169.


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The Annual Range over the Oceans Revisited

Harry van Loon
National Center for Atmospheric Research

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 sub tropics 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 sub-tropics, 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 sub-tropics 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 sub-tropics 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.

<table>
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<th>Table 1.</th>
<th>BEFORE</th>
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<td>So. Hem.</td>
<td>January</td>
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<tr>
<td>Surface Air</td>
<td>16.5</td>
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<td>500 mb</td>
<td>-13.2</td>
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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 (°C) difference, January-February-March minus July-August-September 1985–1993, based on ECMWF analyses. From Hurrell et al., 1997.
Figure 4. (a) The zonally averaged, mean temperature difference on the Southern Hemisphere (°C), January minus July. Based on data from Taljaard et al., 1969. Shading indicates regions where the annual range decreases as one goes poleward. (b) The zonally averaged, mean temperature difference (°C) January-February-March minus July-August-September, based on ECMWF analyses.
Figure 5. (a) The mean surface air temperature (°C) difference on the Southern Hemisphere, January minus July. From van Loon, 1964. (b) The mean 1000–500mb thickness difference (dkm) on the Southern Hemisphere, January minus July. From van Loon and Taljaard, 1958.
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.
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 ($m \text{s}^{-1}$) at 25$^\circ$S, 45$^\circ$S, and 60$^\circ$S. Left: based on Taljaard et al., 1969. From van Loon, 1990. Right: based on ECMWF analyses. From Hurrell et al., 1997.
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

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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 produce 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.

\[
e_r \frac{\partial}{\partial t^*} \left( \zeta^* - FN \sin \varphi \eta^* \right) + \delta \frac{L_m}{r} \cos \varphi \nu^* = 0
\]  

The relative vorticity \( \zeta^* \), surface elevation \( \eta^* \) and meridional velocity \( \nu^* \) are dimensionless variables. \( e_r, F, N \) and \( \delta \) are dimensionless parameters defined as

\[
e_r = \frac{1}{2\Omega T} \\
F = \frac{L_m^2}{R^2} \\
N = \frac{gN_o}{2\Omega U} \\
\delta = \frac{L_m}{L_x} = \frac{V}{U}
\]

\( L_m, L_x, U, V, T \) and \( N_o \) are characteristic scales of meridional and zonal lengths and velocities, and characteristic scales of time and surface elevation, respectively. \( e_r \) 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 \( N_s \), one has \( R_{\text{int}}/R_{\text{ext}} = (N_s^2 H/g)^{1/2} \). \( N \) 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 \( 10^{-1} \), when \( \varepsilon_{\text{max}} << 10^{-2} \) and the tropospheric value of \( N_s \) with \( N_s^2 = 2 \cdot 10^4 \text{ s}^{-2} \) 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 terms of the mid-latitudes small-scale ($L_m << r$) and isotropic ($\delta = l$ and $L_m = L = L$) 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. $N = l$) 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 $F >> l$, the balance described in (1) is characterized by

$$\varepsilon r F = \frac{L}{r} \tag{3}$$

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

$$\frac{1}{T} = \frac{\beta}{f^2} \frac{gH}{1} \frac{1}{L} \tag{4}$$

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

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

$$\omega = -\frac{\beta}{f^2} \frac{gH}{k} \tag{5}$$

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 $F << l$, the balance described in (1) is characterized by

$$\varepsilon r = \frac{L}{r} \tag{6}$$

or

$$\frac{1}{T} = \beta L \tag{7}$$

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 \( FN \sin \varphi = F \gg l \), 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 \( FN \sin \varphi = F \ll l \) 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 \( F = L_{int}^2 / R_{int}^2 \) to be large. As discussed earlier, \( F = O(5) \), so that \( FN \sin \varphi > l \). 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 \( w_{e}^* \) in the continuity equation of the interior flow. The resulting potential vorticity equation becomes, to the lowest order,

\[ \epsilon, \frac{\partial}{\partial t^*} (FN \sin \varphi \eta^*) - \delta \frac{L_{int}}{r} \cos \varphi v^* = \omega_{e}^* \]  

(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 \( w_{e}^* \). 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


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), 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; Wittermeyer 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-Aarkin (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-Arkin 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
examples 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 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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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, GOCP 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.
The Potential for Long–Range Prediction of Precipitation Over India for the Southwest Monsoon Season An Analysis of Variance Approach

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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 (σ^2_A) to the variability due to “climate noise” (σ^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, σ^2_A is assumed to consist the climate noise and variance introduced by slowly changing external conditions (i.e., climate signals). Assuming σ^2_T is independent of σ^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 σ^2_A/σ^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 σ^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’ cannot be estimated under these ideal conditions. Neglecting possible atmospheric intranessibility (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 σ^2_T for precipitation.

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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.

$$\sigma^2_T = T \left[ q \sigma^2 + p(1-p) \frac{1+d}{1-d} \mu^2 \right]$$  \hspace{1cm} (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 ($\geq 2.5$ 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 in 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 (σ_A) and the variance associated with weather fluctuations (σ_I) 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
synoptic systems. The rms of the high-pass filtered data are then obtained for each winter month. A rotated principal component analysis (RPCA) is performed on Atlantic monthly rms values extending from 80°W–20°E and from 30°N to 70°N, following procedures outlined in Rogers (1990). The first, or primary, component of the RPCA represents variability of the main Atlantic storm track.

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.

52
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 Icelandic 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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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%, 98% 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 × 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 nonseasonal 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 Hassenrath, 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/cycloic 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 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.

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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 \(\geq\) 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. Dal.
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 large-scale 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 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, topography 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 (Taljaard 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. The
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 5W m$^{-2}$. Contour interval for the velocity potential is 1 x 10$^8$ m$^2$ 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).

1 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° 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 km². 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 m s⁻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 ($v_{geos}$) 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 $v_{geos}$ 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 geostrophic 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

Ekman flux
surface ocean

Poleward advection
(warming)

Low
net divergence
→ colder surface

Westerly Belt
(Mid- to Sub-Antarctic
Latitudes)

~45°
equatorward advection
(cooling)

High
net convergence
→ warmer surface

~45°
Pressure isoline/
"geostrophic streamline"

"ATLANTIC/WEST INDIAN"

~45°

"PACIFIC"

~45°

SOUTH POLE

Figure 1: Schematic diagram of ZW1 QSW at the surface. The in-phase relation between SLP and SST proposed by van Loon and co-workers as a reason for the baroclinity of the QSW.
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 components from ERS-1. Scatterometer winds have been averaged over 90 days beginning 1 January 1993, in 1° longitude by 5° latitude bins. Wind vectors (top panel) represent the average meridional component for each bin, where 1° of latitude represents 1 ms⁻¹. 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.
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.

Figure 8. Superposition of longitudinal profiles of $v_{geo}$ for the periods JFM 1992–5 (one profile for each year). $v_{geo}$ 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.

121
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).
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 mid-tropospheric 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, 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 produce 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

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

where \( \Delta \) is 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 M10 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
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).
Figure 4. Zonally averaged sea-level pressure (mb) for 50°S and 70°S (top) and their difference (bottom), averaged from 1972–79 (solid) and 1980–89 (dashed). Amplitude (A) in mb, percentage variance explained (V), and phase (P) in months are given for the first and second harmonics (H) (after Hurrell and van Loon, 1994).