Observational studies of low-frequency oscillations in the Southern Hemisphere

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Observational studies of low-frequency oscillations in the Southern Hemisphere

Graves, Charles Edward, Ph.D.
Iowa State University, 1988
Observational studies of low-frequency oscillations in the Southern Hemisphere

by

Charles Edward Graves

A Dissertation Submitted to the Graduate Faculty in Partial Fulfillment of the Requirements for the Degree of DOCTOR OF PHILOSOPHY

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For the Graduate College

Iowa State University
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1988
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### APPENDIX A: OVERVIEW OF ATMOSPHERIC DYNAMICS

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GENERAL INTRODUCTION

Atmospheric phenomena are often classified and studied in spatial and temporal regimes. For example, sound waves have spatial scales of a few meters and temporal scales less than one second. In contrast, planetary Rossby waves have spatial scales of several thousand kilometers and temporal scales of several days. The present study examines atmospheric phenomena with temporal scales having periods near 40-50 days and which are currently the subject of intense investigation by researchers worldwide.

Observational Studies

This phenomenon was first discovered by Madden and Julian (1971) in ten years of rawinsonde data at Canton Island in the western Pacific (3°S, 172°W). (Possible unfamiliar terms like "rawinsonde" are defined in the Glossary.) Their study revealed a peak centered around 40-50 days in the spectra of surface pressure, zonal wind, and temperature. The zonal wind near the surface was found to be out of phase with the zonal wind near the tropopause (at a height of about 15 km). A similar phase shift was also observed for the temperature. They referred to this phenomena as the 40-50 day oscillation.

In a follow-up study, Madden and Julian (1972) found similar results at a variety of stations around the tropical Indian and Pacific Oceans. A schematic adapted from Madden and Julian (1972) depicting their essential results is shown in an equatorial cross-section in Fig.
1. The circulation that they observed was characteristic of a Walker circulation with upward motion associated with low-level convergence and downward motion associated with low-level divergence. The convergence and divergence created an approximate zonal wavenumber one pattern (i.e., one wavelength along a latitude circle). The observed circulation pattern moved eastward at about 5 m/s until it reached the eastern Pacific where it died out. Typical zonal wind values of 5 m/s were observed in the upper branch of the circulation. Their study suggested that the meridional wind was not associated with the oscillation.

Madden and Julian anticipated that cloud development would be enhanced in regions of low-level convergence and suppressed in regions of low-level divergence. They also suggested that nonlinear effects might be important since typical zonal wind values were of the same order as the overall tropical zonal wind.

Others studying this phenomena have generally found results similar to Madden and Julian. Lau and Chan (1985) found a strong dipole pattern in satellite measurements of Outgoing Longwave Radiation (OLR) over the tropical Indian and Pacific Oceans. One pole of the dipole pattern containing high OLR values represented a region of suppressed convection, while the other pole represented a region of enhanced convection. Their study revealed that each pole of the pattern covered 60 degrees of longitude and 20 degrees of latitude and was centered about the equator. Also, they observed that the dipole
Figure 1. A schematic adapted from Madden and Julian (1972) depicting the temporal (top to bottom panel) and spatial (in a zonal plane) circulation associated with the 40-50 day oscillation.
pattern moved eastward at about 5 m/s, dissipating as it reached the eastern Pacific Ocean.

In the OLR data, Lau and Chan observed the anticipated cloud fluctuations suggested by Madden and Julian. Also, the OLR data were derived from polar-orbiting satellites, hence giving a more complete spatial picture of the oscillation than the scattered station data used by Madden and Julian. Lau and Chan thereby confirmed the role of convection and tropical heating as part of the 40-50 day phenomena.

Several studies have examined the relationship between the 40-50 day oscillation and the Asian monsoon cycles (e.g., Yasunari, 1979; Krishnamurti and Subrahmanyan, 1982; Lorenc, 1984; Murakami et al., 1984a b; Chen and Yen, 1986). In general, they all observed that 40-60 day low-level westerly winds in the Indian Ocean were associated with periods of active monsoon rainfall. Likewise, the 40-50 day low-level easterly winds were associated with breaks in the monsoon rainfall. Murakami et al. (1984a, b) observed a separate component of the 40-50 day oscillation over the Indian Ocean, which moved northward until it dissipated at the foot of the Tibetan Plateau. They suggested that this northward propagating component provided the low-level convergence and divergence that resulted in the cycles of rainfall during the monsoon season.

The connection with the Asian monsoon might suggest that the oscillation occurs only during the Northern Hemisphere summer. However, Anderson et al. (1984), as well as others, observed the oscillation
in the tropics throughout the entire year. Still, the general consensus seems to indicate that the strongest amplitudes occur in the Indian and Pacific Oceans during the Asian Monsoon season (May - September).

Most of the studies mentioned above examined the 40-50 day oscillation only over a limited region. However, other studies have investigated the oscillation over much of the globe. Weickmann et al. (1985), using 250 mb zonal wind data from ten Northern winters, presented evidence suggesting that the oscillation was more of a global phenomena, propagating completely around the equator. Thus, the 40-50 day time scale arose as the time needed for the wave to propagate around the globe. The weakest link to the global propagation was over eastern Pacific Ocean where very little signal was found.

Madden and Julian (1972) did not observe global propagation possibly because the few stations available in the Atlantic and equatorial Africa do not have records comparable to those in the tropical Indian and Pacific Oceans. Also, Lau and Chan (1985) did not observe global propagation possibly because the colder waters of the eastern Pacific tend to suppress cloud development — the latter being a necessary element for OLR detection of the 40-50 day oscillation.

Weickmann et al. (1985) also found significant coherency between tropical signals and the Northern Hemisphere 40-50 day fluctuations. The relationship between the tropics and Northern extratropics has also been investigated by Liebmann and Hartmann (1984), Lau and
Phillips (1986), and Chen (1987). Overall, these studies find significant correlations between tropical and extratropical phenomena on the 40-50 day time scale. The results of Lau and Chan, as well as Liebmann and Hartmann, suggest extratropical signals behave as quasi-stationary wavetrains propagating into or out of the tropics. However, the tropical regions that correlate best with wavetrains propagating out of the tropics are not located where the wavetrains are seemingly forced. The region of forcing is located by extrapolating the wavetrain back into the tropics. Likewise, wavetrains propagating into the tropics appear to correlate with signals that are downstream from the location where the wavetrains seemingly enter the tropics. Thus, connections between the tropics and Northern extratropics appear more complicated than simple wavetrain propagation.

Knutson et al. (1986) did a similar study to that of Weickmann et al. (1985), but for nine Northern Hemisphere summers. The tropical results were much the same but the extratropical relationships were only found in the Southern Hemisphere. The relationship observed in the coherency indicated tropical signals were out of phase with signals near 30°S, while they were in phase with signals at 50°S.

Gao and Stanford (1987, 1988) examined the 40-50 day oscillation over the entire globe for four years using satellite-derived microwave infrared brightness temperatures. They observed in Microwave Sounding Unit Channel 4 data (representative of 30-150 mb mean temperatures) the tropical dipole which has become quite common in most studies. They also observed other signals outside the tropics, most
notably a region near Easter Island (27°S, 109°W). Section I of the present study confirms the existence of that signal. Gao and Stanford also observed, in one-point correlation maps, patterns of alternating positive and negative correlations reminiscent of a wavetrain. The wavetrain path began in the eastern half of the dipole over the Pacific Ocean, curving southeastward to a turning point in the Southern Atlantic, and then returning northeastward to the western half of the dipole over the Indian Ocean. This path was suggested as a possible feedback mechanism of the tropical oscillation. Using infrared data (stratospheric sounding unit channels 1, 2, and 3), Gao and Stanford also found 40-50 day signals in the middle and upper stratosphere which had not been previously observed.

Krishnamurti and Gadgil (1985) looked at the variance in the 40-50 day band in the zonal wind, meridional wind and temperature during 1979. They found local areas of large variance in both the tropics and extratropics. They also observed that the vertical structure of zonal wind was more barotropic in the extratropics (no vertical phase change or tilt with height) while the tropical vertical structure was more baroclinic (vertical tilt with height). The baroclinic structure is consistent with the schematic in Fig. 1. They did not explain the reason for large regions of extratropical variance but described them as low-frequency storms.
Theoretical Studies

Besides the observational studies, there have been several modeling studies of the 40-50 day oscillation. With the belief that the meridional wind was not part of the oscillation, most theories proposed the generation of Kelvin waves from tropical heating. (Kelvin waves do not have a meridional wind component.) Yamagata and Hayashi (1984) used a primitive shallow water beta-plane model with a prescribed localized heating which pulsated with a 40 day cycle. The response to the heating consisted of standing and traveling wave components. The traveling component contained both westward modes (Rossby waves) and eastward modes (Kelvin waves). The results were somewhat similar to those of Gill (1980) where the prescribed heating was steady. The extent of the propagation, both eastward and westward, was determined by the level of dissipation in the model.

The circulation resulting from the eastward components (Kelvin waves) generally agreed with the observational results, while the westward components (Rossby waves) were considered to be a discrepancy between the model and observations. However, some of the studies mentioned above reported extratropical signals which cannot be associated with tropical Kelvin waves. (Tropical Kelvin waves decay away from the equator.) These extratropical signals might be related to Rossby wave components. In addition, Madden (1987) observed that the tropical meridional wind (indicative of a Rossby wave and not a Kelvin wave) contained 40-50 day signals which were in phase with the zonal wind
from June to August but out of phase from December to February. Hence, previous studies examining the relationship between zonal and meridional wind over an entire year found them unrelated.

Therefore, the Rossby wave component observed in the model of Yamagata and Hayashi (1984) may be correct, though possibly too strong. A major weakness of this model is that it did not explain the physical basis for a 40 day heating cycle in the tropics.

Oscillations similar to the observed have also been found in General Circulation Models (GCM), although the GCM oscillations usually have shorter periods than observations, about 20–40 days (Lau and Lau, 1986; Hayashi and Sumi, 1986; Swinbank et al., 1988). Hayashi and Sumi (1986) observed that a wave-CISK process in an ocean covered GCM generates an eastward propagating anomaly that circles the equator in about 30–40 days (phase speed of about 15 m/s). If moisture effects (evaporation and condensation) are removed from the model, then the generated anomalies are Rossby and Kelvin waves similar to Yamagata and Hayashi (1984) but with much faster phase speeds (about 30 m/s) resulting in a 15–20 day oscillation.

The possibility that moisture plays a significant role in this tropical oscillation has led to the theoretical development of mobile wave-CISK theories. Essentially, mobile wave-CISK or moist Kelvin wave theories are based on the movement of tropical convection along with the eastward moving Kelvin waves. Thus, the Kelvin wave "carries" its forcing mechanism along, as it propagates eastward. The Kelvin wave
continues to exist as long as there is enough positive feedback between the forcing and the wave. The eastward phase velocity is controlled by moisture variables (i.e., evaporation, condensation, and convection) and is slower than a dry Kelvin wave. Also, the eastward movement of the forcing causes the successively generated Rossby waves to almost cancel, reducing the Rossby wave response. The key to achieving a 40-50 day oscillation with this theory, is to generate a Kelvin wave with the needed moisture effects which takes 40-50 days to propagate around the globe.

A variety of studies (Lau and Peng, 1987; Hendon, 1988; Swinbank et al., 1988) have used versions of this theory with reasonable success. They generally observe a circulation pattern reminiscent of Fig. 1, but phase speeds are somewhat too fast and are probably too sensitive to the vertical distribution of heating. In addition, the limited knowledge of the tropical hydrological cycle may also result in poor parameterizations for the CISK mechanism in both the simple and complex models.

Finally, in a different approach, Salby and Garcia (1987) and Garcia and Salby (1987) examined the linearized primitive equations as a dynamical system subject to random forcing. They looked for normal modes of the system that might be expected from a red noise type of forcing. They reasoned that tropical heating, as observed in convective systems, appears to the atmosphere as a forcing with a red noise character in both space and time.
Garcia and Salby's results revealed a variety of solutions on several time scales; at the lowest frequencies they found two separate responses, a barotropic component (Rossby wave-like) and a baroclinic component (Kelvin wave-like). The Kelvin wave is confined to the tropics and propagates eastward and upward, while the Rossby-like response propagates into the extratropical regions. This model is one of the few that actually considers extratropical responses. Their results indicate that extratropical responses will interact with the mean flow and may then propagate vertically into the stratosphere. In certain respects, their results are similar to Yamagata and Hayashi (1984).

Summarizing, attempts to model the observed characteristics of the 40-50 day oscillation have been moderately successful. The main difficulty lies in obtaining the observed frequency and phase speed; the modeled results are usually too fast. Because the extratropical responses are still being observationally identified, modelers have mainly concentrated on reproducing the observed tropical circulation.

Present Study

The extratropical component of the tropical 40-50 day oscillation has yet to be clearly resolved. The present studies aim at improving our understanding of 40-50 day extratropical fluctuations, particularly, in the Southern Hemisphere. Our first paper (Section I) details the 40-50 day response in temperature and zonal wind observed at Easter Island (27°S, 109°W). The motivation for this work was to corroborate
signals observed by Gao and Stanford (1987). The results also confirm the barotropic vertical structure observed by Krishnamurti and Gadgil (1985). This work also served as my introduction into this field.

The second paper (Section II) details a more complete structure of the fluctuations on the 40-50 day time scale in the extratropical Southern Hemisphere. These fluctuations appear to behave as quasi-stationary wavetrains which may interact with the tropics over the Indian Ocean. These wavetrains gain energy from the mean zonal flow by baroclinic energy conversions and lose energy to the mean zonal flow by barotropic energy conversions.

The value of these studies is: (1) that they establish the existence of 40-50 day oscillations in the extratropical Southern Hemisphere without reference to the tropical phenomena and (2) that the tropical 40-50 day oscillation may be influenced by midlatitude phenomena.

Details concerning aspects of atmospheric dynamics, omitted in Sections I and II, are given in Appendix A. A glossary of possible unfamiliar terms is found in Appendix B and a list of symbols are given in Appendix C.

Explanation of Dissertation Format

The alternate dissertation format is followed, in which two papers have been or will be submitted to scholarly journals.
The candidate had primary responsibility for both papers. All data analysis was performed by the candidate. Section I has been published in the 1987 January issue of the *Journal of the Atmospheric Sciences*. Section II is to be submitted to the same journal. The style is as required for the *Journal of the Atmospheric Sciences*. 
LOW-FREQUENCY ATMOSPHERIC
OSCILLATIONS OVER THE SOUTHEASTERN PACIFIC

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SECTION I. LOW-FREQUENCY ATMOSPHERIC OSCILLATIONS OVER THE SOUTHEASTERN PACIFIC

Abstract

A 40-50 day atmospheric oscillation has recently been reported in the southeastern Pacific based on analyses of satellite-derived microwave brightness temperature data. Prior to this, such oscillations have not been generally recognized in this region. The purpose of this Note is to provide corroboration of the microwave analyses from an independent data set, four years of rawinsonde data from Easter Island (27°S, 109°W). Significant spectral peaks with period 45-53 days were found in the zonal wind near the jet core (300-200 mb) and in the temperature on the underside of the jet. The signal is also present in the 100 mb temperature as shown by strong coherence between 100 and 400 mb temperature perturbations. Direct comparison of the 100 mb temperature perturbations and the microwave data show good agreement.

There is also an indication of a lower frequency signal, about 65-84 day period, in the meridional wind at 1000 mb. This signal appears to be vertically evanescent with maximum amplitude at the surface.

Introduction

Madden and Julian (1971, 1972), hereafter MJ, first reported a 40-50 day oscillation in 10 years of rawinsonde data; subsequently, many others have investigated these low-frequency oscillations (LFO), with much of the work being concentrated over the tropical western
Pacific and Indian Oceans, for example, Lau and Chan (1985) and Mura­
kami et al. (1984). Recently, others such as Weickmann et al. (1985),
have examined the LFO on a global scale. In all previous studies,
there was little or no LFO specifically reported in the southeastern
(SE) Pacific. However, Gao and Stanford (1986), hereafter GS, using
satellite-derived microwave brightness temperatures have recently re­
ported a strong low frequency signal in the SE Pacific. GS attributed
the lack of previous evidence of the low frequency signal in the SE
Pacific to (1) the lack of analyses of rawinsonde data from this area,
and (2) the fact that many of the observational studies have been based
on the outgoing longwave radiation data sets which depend implicitly
on the occurrence of high clouds. The latter tend to be conspicuously
absent in satellite photos of the SE Pacific, a region of general atmos­
pheric subsidence. The motivation behind the present work is to provide
corroborative evidence for the signal found by GS, based on an entire­
ly independent data set.

Describing the frequency band of the LFO as 40-50 days is more
generic than exact; the feature is generally considered to have a broad
frequency range somewhere in the neighborhood of 30-60 days. The fre­
quency bands examined in this work are relatively narrow, with periods
near 40-50 days, chosen to compare specifically with the results of MJ
and GS. In addition, the present study also detects a lower frequency
signal, with 65-84 day period, in the meridional wind over the SE Pacific
with maximum amplitude near the surface.
Data and Analysis

This study used four years of rawinsonde data from Easter Island (109°W, 27°S), from December 1, 1979 to November 30, 1983. The rawinsonde data consisted of once a day measurements of temperature and wind fields at the standard pressure levels from 1000 to 100 mb. For computational efficiency, three-day nonoverlapping means were used to reduce the length of the data set to 488 points without significantly affecting the low frequency signal. (The data set was actually only 487 points, so the last point was repeated to increase the efficiency of the analysis.) Any missing three-day means were replaced by linearly interpolated values. As expected with a rawinsonde data set there were more missing data at the higher levels, with number of the interpolations reaching a maximum of 4% for the temperature and 20% for the wind, both at 100 mb. At 200 mb, the percentages were 3% and 16%, respectively.

After removing the mean from each time series and at each level, the resulting time series were Fourier analyzed to obtain raw spectral estimates. Leakage from the annual cycle and its harmonics was minimized by choosing the data set to be exactly 4 years long, so the annual cycle was an exact harmonic of the data set. Spectral estimates with periods longer than 183 days or shorter than 28 days are not considered here. The raw estimates were then band averaged over five spectral points, reducing the frequency resolution from 0.00068 to 0.0034 day\(^{-1}\), but giving more reliable spectral estimates. The sensitivity of the spectrum to the analysis technique was checked by comparing the spectrum
with a hanned version. In neither case did the zonal wind or temperature show significant changes. The meridional wind contained no significant peaks within the typical LFO frequency range.

Results

Power

Figure 1 shows the spectra of zonal wind and temperature at 250 mb and 500 mb, respectively. At these levels, the 45-53 day peak is the strongest. The dashed lines represent the background estimate (lower line) and the 95% confidence level (upper line). The null hypothesis (true background) is not known but an objective estimate is given for the background in the Appendix. Because this feature has been previously reported, a priori statistics or a 95% confidence level is considered sufficient to establish significance. In both cases, the 45-53 day peak is significant at the 95% level. (See the Appendix for significance calculations.) This same peak is also significant at 300 and 200 mb in the zonal wind and at 400, 300, and 250 mb in the temperature spectra. In altitude, these perturbations in the zonal wind are strongest near the jet stream core (300 - 200 mb) while the temperature perturbations are maximized on the underside of the jet.

Vertical structure

Rather than obtaining mean phases by cross-spectral analysis, we emphasize the episodic nature of these signals by examining the vertical structure of the LFO reconstructed from (band-pass) Fourier components
Figure 1. Band-averaged power spectra for (a) zonal wind at 250 mb and (b) temperature at 500 mb. The period in days is given at the top and the frequency is given in cycles per 1464 days on the bottom of the figure. The units of power are $m^2 \, s^{-2}/(0.026 \, \text{day})$ for (a) and $K^2/(0.026 \, \text{day})$ for (b), with the area under the curve being proportional to the variance. The bandwidth is 0.0034 day$^{-1}$, indicated by the plotted widths.
PERIOD IN DAYS

(a) 250 MB U WIND

(b) 500 MB TEMPERATURE

95% C.L.

BACKGROUND
within the 39–53 day band. Figure 2a is the reconstructed zonal wind time series at all the pressure levels, 1000 to 100 mb. Although only three of the pressure levels were shown to have statistically significant peaks (250, 300, and 400 mb), a number of wind anomalies can be traced over a considerable vertical depth. One of the characteristics of the LFO described by MJ was a nearly 180 degree phase shift between the upper and lower tropospheric zonal wind anomalies. Figure 2 reveals that the LFO at Easter Island exhibits very little change of phase in the vertical. The nearly vertical phase structure was also found for other frequency bands (not shown). This barotrophic structure is consistent with other observational LFO investigations outside the tropics, such as Krishnamurti and Gadgil (1985) and is also predicted by recent modeling studies (Garcia and Salby, 1987).

The reconstructed time series in Fig. 2a also appears to have a seasonal pattern with the strongest amplitude during the Southern Hemisphere winter (June, July, and August). The enhanced activity during the third year of the time series coincides with the strong 1982–1983 El Niño event (beginning about day 830 in Fig. 2a).

The reconstructed temperature time series (Fig. 2b) does not show the same simple vertical structure as the zonal wind, even though the temperature spectra were statistically significant at four pressure levels (250, 300, 400, and 500 mb). While these four levels appear to be in phase, the 100 mb level is most often out of phase with these lower levels, although this can only be determined when amplitudes are
Figure 2a. The reconstructed zonal wind from the Fourier components within the 39-53 day band from 1000 to 100 mb. The distance between consecutive levels (or vertical tic marks) is 10 m/s.
Figure 2b. The same as for Figure 2a but with the reconstructed temperature. The distance between consecutive levels (or vertical tic marks) is 4 degrees Kelvin.
reasonably large. Moreover, during the enhanced activity in the zonal wind, the amplitude of the temperature perturbations at 250 and 300 mb is near zero while the amplitudes of the zonal wind perturbations were maximum. The change of phase between 100 mb and mid-tropospheric levels is exactly that expected if the geopotential perturbations maximize in the core of the jet stream and the temperature perturbations are proportional to the vertical derivative of the geopotential. This out-of-phase characteristic was substantiated by cross-spectral analysis which revealed a nearly 180 phase shift between 100 mb and 400 mb with coherency squared exceeding 0.90 (a value of 0.53 is needed to achieve the 95% confidence level with 10 degrees of freedom). This strong coherency suggests that the LFO also exists in the 100 mb temperature data, although the noise level of the rawinsonde data inhibited detection directly in the temperature power spectrum.

**Comparison with MSU data**

This study was stimulated by the results of GS, who analyzed Microwave Sounding Unit (MSU) channel 4 brightness temperatures. These represent weighted vertical averages of temperature from about 150 to 30 mb. Figure 3 is a comparison of the Easter Island rawinsonde 40-50 day reconstructed temperature at 100 mb and 40-50 day reconstructed MSU channel 4 brightness temperatures at several locations surrounding Easter Island. The comparison is considered good even though the amplitudes of the rawinsonde temperature fluctuations are larger than those in the MSU data. This is consistent with the facts that (1) the MSU
Figure 3. The reconstructed Easter Island temperature at 100 mb and the MSU brightness temperature from the 40-50 day band. The geographical locations of the MSU brightness temperatures are labelled to the right of the corresponding line. Easter Island is located at 27°S, 109°W. Vertical tic marks are separated by 4 degrees Kelvin.
data are a vertical and horizontal average and (2) the method of analysis by GS eliminated zonal wavenumbers larger than 10; both points tend to reduce the amount of signal amplitude. The overall phase and envelope agreement indicates that the MSU instrument has detected the same feature found in this study. The four-year data set used by GS did not completely overlap that used in the present study. The GS data set began April 1, 1980, whereas ours began December 1, 1979.

**Lower frequency signal**

Figure 4 shows the spectrum of the meridional wind at 1000 mb. Similar to the results of MJ, there is no significant signal in the LFO frequency band at this or at any pressure level in the meridional wind. However, note the strong peak at very low frequencies, 65-84 day periods. Because this peak was unexpected, stronger (a posteriori) statistical requirements must be met before it can be established as significant. This is especially true in light of the tendency of atmospheric spectra to exhibit red noise characteristics at low frequencies. For our data and analysis parameters, a posteriori statistics require a spectral peak to reach the 99.9% confidence level for significance (see the Appendix of GS whose parameters are identical to those employed here). The dashed lines represent different confidence levels, and the dotted line is the subjectively estimated background. The peak meets the 99.9% confidence level required to establish significance. The peak is also observed at 850 and 700 mb, although the amplitude of the peak decays with increasing altitude and is difficult to discern above 700 mb.
Figure 4. Same as Figure 1a but for the meridional wind at 1000 mb.
Evidence of the peak is also seen in the temperature spectra (see Fig. 1). The relationship between this 65-84 day feature and the broad band of LFO described by other authors is yet unclear. Further studies are in progress.

Conclusion

Significant power was found in the SE Pacific in the zonal wind and temperature spectra within the frequency band of the LFO, revealing a barotropic structure consistent with other reports of LFO outside the tropics. The signal was also found in the temperature at 100 mb, out of phase with the lower levels. This 100 mb temperature signal is comparable to the 40-50 day feature found by GS using satellite-derived microwave data. These results, together with those of GS, are important because they validate the use of Microwave Sounder Unit data for LFO investigations. In particular, they reveal the first clear evidence of annually occurring LFO in the SE Pacific, a region whose general subsidence and relative lack of cloudiness (except in El Niño years) evidently generally precludes observations of LFO by outgoing longwave radiation data sets. It may be that these results reflect a link between the Southern Hemisphere subtropical jet and tropical convection on this time scale. Finally, an unexpected lower frequency signal (65-84 day periods) was also found near the surface in the meridional wind.
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Appendix

Background estimation procedure

The background of a given spectral peak is calculated in a two-fold manner. First, a biased background is calculated by a least square fit to a second-order polynomial using all the spectral estimates within the band of periods 30-98 days. (It makes little difference whether the raw or band averaged estimates are used.) Second, individual raw spectral estimates above the 95% confidence level from the biased background were removed from the spectrum and replaced with the average of adjacent spectral values. The final estimated background is found via another second-order polynomial least square fit using the spectral estimates without the significant peaks. These backgrounds are shown by the longer dashed lines in Fig. 1a, b.

The background for the lower frequency peak in Fig. 4 was drawn subjectively.

Significance determination

For the band-averaged spectral estimates (5-point nonoverlapping means), there are approximately ten degrees of freedom in our spectra,
two for each raw spectral estimate. Therefore, the 95% confidence level is achieved when the ratio of the spectral peak to the background is greater than 1.8 (i.e., the ratio of the chi-squared distribution at the 95% confidence level for 10 degrees of freedom, to the number of degrees of freedom, 10). Further details may be found in the Appendix of GS and more generally in Mitchell (1966).

References


MIDDLE AND HIGH LATITUDE SOUTHERN HEMISPHERE
OSCILLATIONS ON THE 35-60 DAY TIME SCALE

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SECTION II. MIDDLE AND HIGH LATITUDE SOUTHERN HEEMISPHERE OSCILLATIONS ON THE 35–60 DAY TIME SCALE

Abstract

Geopotential height fluctuations are examined in the extratropical Southern Hemisphere on the 35–60 day time scale. These fluctuations, in 200 mb one-point correlation maps and anomaly maps, are shown to behave like quasi-stationary wavetrains with eastward energy propagation. The wavetrains follow an elliptically shaped circuit identifiable from low Southern latitudes in the Indian Ocean to the coast of Antarctic near 70°W. There is some suggestion of reflection from a critical layer or else stimulated emission of a new wavetrain north of Australia. Correlations found in the tropics in regions of easterlies may result from the evanescent tail of the wavetrain.

Eliassen-Palm flux diagrams and energetics calculations suggest that energy of the 35–60 day features is extracted from the mean zonal flow through baroclinic processes and lost to the mean zonal flow through barotropic processes.

These observational results constitute several tests for models, particularly, the existence and amplification of the midlatitude fluctuations and their quasi-stationary nature.
Introduction

Since the pioneering work of Madden and Julian (1971, 1972), most investigations of 40-50 day atmospheric signals have concentrated on the tropics. Recently, studies have noted relationships on the same time scale between this tropical phenomena and fluctuations in the midlatitudes of the Southern Hemisphere (e.g., Knutson et al., 1986; Gao and Stanford, 1988). The present study investigates these extratropical fluctuations in geopotential height one-point correlation maps, anomaly maps, and by examining their interactions with the mean zonal flow.

The investigations of Knutson et al. (1986) found that significant coherency existed at 250 mb between the zonal wind in the tropics and the zonal wind in Southern midlatitudes. They also found similar relationships between tropical outgoing longwave radiation and the zonal wind in the Southern midlatitudes. Krishnamurti and Gadgil (1985) found regions of large variance in the zonal wind and temperature on the 40-50 day time scale in the Southern and Northern Hemisphere for 1979. They found that the midlatitude fluctuations show a more barotropic structure than their tropical counterparts. Gao and Stanford (1988) found a wave-like pattern in Microwave Sounding Unit channel 4 (MSU4) brightness temperatures. The wave-like pattern followed a path into the Southern Hemisphere from the tropical western Pacific to a turning point over the southern Atlantic and then returning to the tropical Indian Ocean. Gao and Stanford suggested that this route may
constitute a feedback mechanism for the 40-50 day oscillation.

The present study analyzes the Southern Hemisphere extratropical fluctuations by themselves and then considers possible connections with the tropics. In doing so, it becomes advantageous to conceptually separate the tropical phenomena -- which we denote the Madden and Julian Oscillation (MJO) -- from the extratropical fluctuations. As will be shown, we can successfully interpret our results in terms of propagating wavetrains. The energy of the wavetrains propagates along a route that suggests a possible connection with the MJO over the Indian Ocean. We also find evidence consistent with a low latitude reflection of the wavetrain or stimulated wavetrain emission over the Indian Ocean. In addition, we find that these fluctuations extract energy from the mean zonal wind through baroclinic processes and lose energy to the mean zonal wind through barotropic processes.

We investigate the low frequency fluctuations in the Southern Hemisphere height fields mainly through one-point correlation maps. Before discussing the results from these maps, we motivate our choice of reference points and the frequency limits of our filter. Using the one-point correlations, we then examine the spatial and temporal structure of filtered geopotential heights. Finally, we investigate the manner in which these geopotential height fluctuations interact with the mean zonal flow.
Data

This study uses daily geopotential heights and Microwave Sounding Unit channel 4 (MSU4) brightness temperatures on a five by five degree grid obtained from the British Meteorological Office. The data sets cover five years, from 1 April 1980 to 31 March 1985. The height fields are from NMC analyses up to 100 mb; above that the heights are determined by integrating satellite temperature soundings. The pressure levels used are at 850, 500, 300, 200, 100, 50, 20, 10, 5, 2, and 1 mb. The height fields are analyzed from 20°S to 80°S and the MSU4 fields are analyzed from 10°N to 10°S. The MSU4 data are representative of 30 to 150 mb mean temperatures.

Missing data were linearly interpolated along a latitude circle if fewer than 24 (of 72) longitudes were missing. For those latitude circles missing more than 24 points, the entire latitude circle was linearly interpolated in time. In general, less than 15% of the data were obtained by interpolation for any single level or latitude. There was only one string of 28 days (from 2 July 1983 to 29 July 1983) in the geopotential heights where the gap of missing data was larger than 15 days. The number of elements in each time series was compressed by using three-day non-overlapping means.

Frequency Analysis

The data were zonally and temporally decomposed into zonal wave-numbers \( k \) and frequencies \( \omega_n \) \((=2\pi n/1824 \text{ days}^{-1}, n=0 \text{ to } 304)\) so that
for a particular latitude and pressure level the geopotential height, $Z(x,t)$, was given by

$$Z(x,t) = \sum_{n} A_n(x) \cos(\omega_n t) + B_n(x) \sin(\omega_n t).$$

(1)

Here,

$$A_n(x) = \sum_{k} a_{kn} \cos(kx) + b_{kn} \sin(kx),$$

and

$$B_n(x) = \sum_{k} c_{kn} \cos(kx) + d_{kn} \sin(kx),$$

where $x$ is longitude, and $t$ is time. The data were spatially filtered by restricting the sum over $k$ to 1 through 10.

The variance at a given location (i.e., $(A_n(x)^2 + B_n(x)^2)/2.0$) was averaged over $n=32$ to $53$ (periods of about 58–34 days) and compared to a red noise background to infer where statistically significant fluctuations exist in the Southern Hemisphere. Red noise estimates were made following Mitchell (1966) and were determined using the variance for periods from 5 years to about 18 days. Before estimating the red noise background, the variance at the annual and semi-annual harmonics ($n=5$ and 10) was replaced with the average variance from the two neighboring harmonics. Figure 1 shows the ratio between the 58–34 day variance and the red noise background at 200 mb. The contours represent the 70%, 80%, 90%, 95% and 97.5% confidence levels, respectively (see Appendix for details). Revealed in Fig. 1 are three
Figure 1. The ratio of variance from 58-34 days to a red noise background estimate for the geopotential heights in the Southern Hemisphere at 200 mb. The contour levels represent the 70%, 80%, 90%, 95%, and 97.5% confidence levels, respectively. The outer, middle and inner latitude circles are 20°, 50°, and 80°S.
distinct regions where fluctuations on the 40-50 day time scale are highly significant in the Southern Hemisphere. These regions are centered near (1) 45°S, 10°W; (2) 70°S, 60°W; and (3) 20°S, 130°E. The majority of the results presented here are based on correlations with the reference point at 45°S, 10°W. This location was chosen for two reasons: 1) it is a region of statistical significance, and 2) this region has previously been recognized as one containing fluctuations on the 40-50 day time scale (Gao and Stanford, 1988; Knutson et al., 1986).

The power spectrum and the estimated background at this location (45°S, 10°W and 200 mb) are shown in Fig. 2. The spectrum is estimated by applying a 7 point running mean to the periodogram. (Again, the annual and semi-annual harmonics were replaced by the average of their respective two neighboring harmonics.) The spectrum is well above the background estimate over a frequency range centered at 40-50 days. Also present in this spectrum (and more prominently in others examined) is a peak above the background estimate in the spectrum between 25-30 days. We chose to explicitly exclude this peak in the present analysis because it is on the fringe of the generally accepted frequency range of the MJO and it is near frequencies related to lunar or solar rotational periodicities. To exclude this peak, a filter was constructed which effectively discriminates against periods of less than one month as well as periods of more than two months. The filter response is shown in Fig. 3. This 41-point filter has half amplitude points at
Figure 2. The spectrum and red noise background estimate for 45°S, 10°W, and 200 mb. The spectrum is estimated from a 7 point running mean over the periodogram. The power is in units of $10^4 \times \text{gpm}^2$/unit frequency. The red noise estimate follows Mitchell (1966)
Figure 3. The response of the 41-point filter applied to the geopotential height and MSU4 data. Half amplitude points are at approximately 35 and 61 days.
approximately 35 and 61 days. The filter was applied by appropriately weighting the Fourier coefficients $A_n(x)$ and $B_n(x)$ from Eq. (1) and then performing an inverse Fourier transform. In the resulting time series, the first and last 20 points were removed because of the 41 points in the filter. The data are also spatially filtered since only zonal wavenumbers $k=1-10$ were used. These filtered time series are used for all correlations and anomaly studies described in later sections.

The MSU4 data were filtered in exactly the same manner as the geopotential heights. The MSU4 data have been shown to be capable of detecting the MJO (Gao and Stanford, 1987; 1988) and are used here to examine possible connections between the MJO and the 35-60 day fluctuations in the Southern Hemisphere geopotential heights.

One-Point Correlations

The correlations presented in this study are simple linear correlations described in textbooks on time series analysis (e.g., Chatfield, 1980). As shown in a later section, the filtered fluctuations are generally strongest in the Southern Hemisphere winter, thus we calculated the correlations from May through October for each year (1980 to 1984) and then averaged them together. The correlations are determined at all pressure levels and for lags of -18 to +18 days with a step in lag of 3 days, though for brevity only selected results are presented here. Correlations are calculated with respect to the
reference point at 45°S, 10°W, and 200 mb, unless otherwise noted. A positive lag indicates that the reference point leads the rest of the height field. That is, a correlation at a lag of +6 days between the reference point and longitude x, implies that when an event occurs at the reference point, a corresponding event occurs 6 days later at x.

Local statistical significance is achieved in our analysis by correlations stronger than ±0.4. In the correlation maps presented here, the contours are ±0.3, ±0.4, ... ±1.0, so that the area within the second contour is considered significant. All maps presented here are considered globally significant unless there is an asterisk on the upper right corner of the graph. Both local and global statistical assessments are discussed in the Appendix.

**Horizontal cross-sections**

One-point correlation maps at 200 mb for lags of -6, -3, 0, +3, +6, and +9 days are shown in Fig. 4. To aid in the discussion of these correlation maps, distinct regions of significant correlation have been assigned the letters A through I. In Fig. 4a (lag of -6 days), regions of alternating positive and negative correlations (labelled as A, B, C, and D) form a wave-like pattern. From a lag of -6 to +6 days (Figs. 4a-4e), the wave-like pattern observed in Fig. 4a evolves in a systematic manner. Regions B, C, and D strengthen at first and then weaken at later lags. Region A weakens and disappears while E and then F appear and develop downstream of the existing wave-
Figure 4. One-point correlation maps in the Southern Hemisphere at 200 mb with the reference point at 45°S, 10°W and 200 mb (region D). Figures (a) through (f) show results at lags of -6, -3, 0, +3, +6, and +9 days, respectively. The outer, middle and inner latitude circles are 20°, 50°, and 80°S, respectively. The contours are correlations of ±0.3, ±0.4, ..., ±1.0 and dashed contours represent negative correlations. The letters A through I denote those regions of correlation discussed in the text. Local significance is achieved with correlations stronger than ±0.4 and the asterisk in the upper right corner of a graph indicates this correlation map did not satisfy global significance requirements (see Appendix for details).
like pattern. Finally, at a lag of +9 days (Fig. 4f), most of the regions have vanished except for regions F and C which have diminished in size and strength.

Overall, the systematic development of regions E and F, together with the subsequent weakening of regions A and B is indicative of a wavetrain pulse propagating along the path from A to F, as depicted in Fig. 5. Hence, we adopt a wavetrain philosophy, describing the results of the correlations in terms of a wavetrain with the regions of positive (negative) correlations representing crests (troughs) and the overall strength of the correlation representative of the amplitude envelope of the pulse.

As observed in Fig. 4, the location of each region (A to F) remains about the same, suggesting little phase propagation. For example, region D (the region containing the reference point) is found to be at nearly the same location for the lags from -6 to +6 days. To determine an approximate group velocity, we estimate that the center of the amplitude envelope is between regions B and C (about 100°W, 65°S) at a lag of -6 days (Fig. 4a) and centered on region E (about 40°E, 40°S) at a lag of 6 days. These estimates result in an eastward group velocity of about 10 m/s and a northward group velocity of about 3 m/s. The southward group velocity occurring when the maximum of the amplitude envelope passes from A to C is estimated to be about the same as the northward group velocity. Since this zonal group velocity is estimated to be less than the zonal wind speed (about 25 m/s), the
Figure 5. The same as Figure 4c but the suggested propagation path along A through F is depicted.
Intrinsic group velocity is westward.

The location of strong negative correlation at C (about 70°S, 60°W) is very close to the second region of strong signal-to-background ratio observed in Fig. 1. Thus, choosing 70°S, 60°W as a reference point produces one-point correlation maps (not shown) nearly identical to those in Fig. 4. The only noticeable difference is a 10 degree shift to the west in the location of region F.

**Vertical structure**

The wavetrain observed in Fig. 4 is found throughout the vertical column in the troposphere. Figure 6 shows the correlations at 850 and 500 mb at a lag of 0 days. At both pressure levels, a wavetrain similar to that at 200 mb (Fig. 4c) is observed. The comparison of regions C, D, E, and F between Figs. 6a and 6b reveals a small westward tilt with height of about 5 to 10 degrees from 850 to 500 mb. From 500 to 200 mb there is very little tilt with height except in region F which exhibits a small eastward and southward tilt with height.

In the stratosphere, the wavetrain is observed at 100 mb, but at 50 mb the complete wavetrain is not observed; only regions C, D, and E are occasionally observed. At higher altitudes, correlations are found only at a lag of -6 days and to a lesser extent at -3 days. The correlations at a lag of -6 days are shown in Fig. 7 at 50 and 2 mb. At 200 mb and -6 day lag (Fig. 4a), the wavetrain extends from regions A to D, while at 50 mb (Fig. 7a) only regions D and E are found. At 2 mb (Fig. 7b), region F is found together with a negative
Figure 6. Similar to Figure 4 but for one-point correlation maps at (a) 850 mb and (b) 500 mb at a lag of 0 days both with the reference point at 45°S, 10°W and 200 mb. The outer, middle and inner latitude circles are 20°, 50°, and 80°S, respectively. The contours are correlations of ±0.3, ±0.4, ..., ±1.0 and dashed contours represent negative correlations.
Figure 7. Similar to Figure 4 but for one-point correlation maps at (a) 50 mb and (b) 2 mb at a lag of -6 days both with the reference point at 45°S, 10°W and 200 mb. The outer, middle and inner latitude circles are 20°, 50°, and 80°S, respectively. The contours are correlations of ±0.3, ±0.4, ..., ±1.0 and dashed contours represent negative correlations.
(a) 50MB LAG=-6

(b) 2MB LAG=-6
correlation near 25°S, 70°W. This three-dimensional correlation pattern is reminiscent of a spiral staircase, with regions of correlation found eastward and upward from the previous lower regions. The spiral pattern suggests possible stratospheric connections with tropospheric 35-60 day fluctuations. The regions of correlations observed in Fig. 7 are near locations of stratospheric 40-50 day signals found by Gao and Stanford (1987). The other region of negative correlation observed at 2 mb (Fig. 7b) indicates that the 35-60 day signal in the stratosphere is concentrated strongly in zonal wavenumber one.

Overall, the correlation pattern found at other levels in the troposphere are very similar to that at 200 mb. Therefore, the results obtained at 200 mb are representative of the Southern Hemispheric troposphere. Also, the small vertical tilt suggests a barotropic structure to the fluctuations, similar to the results of Krishnamurti and Gadgil (1985); however, because of the approximate zonal wavenumber four structure, even 5 to 10 degree longitude tilts can imply non-negligible baroclinic effects, as will be substantiated later in the energetics calculations.

Consequences of Observed Propagation

Tropical connections

If the wavetrain continues to propagate in a manner suggested by Fig. 5, then it might be possible for these fluctuations to propagate into the tropics. To test this hypothesis, we correlated the reference
point (45°S, 10°W and 200 mb) with the MSU4 brightness temperatures from 10°N to 10°S. Based on Fig. 4, one might anticipate correlations between the MSU4 data and the reference point to occur somewhat east of 90°E and at lags of 9 days or later. The correlations with the MSU4 data for lags of 9, 12 and 15 days are shown in Fig. 8. As anticipated, correlations are observed at about 100°E for latitudes of 5°N to 5°S. The area of correlation increases from a lag of 9 to 12 days which is consistent with the hypothesis of propagation from the higher latitudes into the tropics. Thus, midlatitude fluctuations could have an effect on the MJO. However, the correlations observed in the MSU4 data are much weaker than observed in the geopotential heights and might be the result of the evanescent tail of the wavetrain as it enters the negative index of refraction region of easterly flow in the tropics. The region near 100° to 120°W achieves the necessary a priori statistical significance, based on the hypothesis of propagation from higher latitudes along the path denoted in Fig. 5. The other regions of correlation in Fig. 8 are not discussed because they were not expected a priori and hence require stricter statistical considerations.

Reflection or regeneration?

Easterly flow is expected in the tropics over the Indian Ocean north of 10° or 15°S (Newell et al., 1972; Stadler, 1975). Thus, if the wavetrain in Fig. 5 continues to propagate northward from region F, it would be expected to encounter a critical layer (i.e., region
Figure 8. Correlations of MSU4 data from 10°N to 10°S, with the geopotential reference point at 45°S, 10°W and 200 mb. Lags of 9, 12 and 15 days are shown in (a) through (c), respectively. The contours are correlations of ±0.3, ±0.4, ..., ±1.0 and dashed contours represent negative correlations. Correlations near 100°-120°W achieve a priori statistical significance based on the hypothesis of energy propagating into the tropics as suggested in Figure 5.
(a) MSU4 LAG=9

(b) MSU4 LAG=12

(c) MSU4 LAG=15
where the phase speed equals the background wind speed, in this case, the zero wind line). Under certain approximations, the wavetrain is expected to reflect, especially when nonlinear effects become important (Killworth and McIntyre, 1985). While little evidence suggesting a reflection is observed in the correlations in Fig. 4 (i.e., with the reference point at 45°S, 10°W, and 200 mb), a third region of strong signal-to-background is shown in Fig. 1 where reflection might be expected (low latitudes along 100° to 140°E).

Correlations based on this reference point (20°S, 130°E) are shown in Fig. 9. These correlations have regions similar to A, B, and F (and are marked as such) along with the region (marked as G) associated with the new reference point. At lags of -6 to 0 days (Figs. 9a-9c), regions F and G increase in size and G increases in strength. From a lag of 3 to 9 days (Figs. 9d-9f), regions F and G weaken while A and B begin to develop. The direction of energy propagation changes from northeastward (along D-E-F in Figs. 4c-4f) to southeastward (along F-A-B in Fig. 9). The energy pulse then appears to merge smoothly into the wavetrain observed in Fig. 4a. But why is this apparent reflection not observed in the lag correlations of Fig. 4 (i.e., with the reference point at 45°S, 10°W)? We speculate that if the reflection is due to a critical layer, the nonlinear behavior of critical layer dynamics may destroy linear relationships. Therefore, the correlation between regions D and G will be weak. Also, the strong signal-to-background ratios observed at low latitudes along 100° to 140°E in Fig. 1 suggest
Figure 9. Same as Figure 4 but with the reference point at 20°S, 130°E and 200 mb (located in region G)
this region may represent an accumulation region caused by the critical layer dynamics.

It may also be possible that the incoming wavetrain (D-E-F) stimulates tropical forcing which, in turn, generates another wavetrain (G-A-B). These two possibilities would not be differentiable in our results.

Summarizing, the weak correlations with tropical temperature fluctuations (Fig. 8) and those observed in Fig. 9 suggest that the wavetrain enters the tropics where it may be reflected or may stimulate the generation of a new wavetrain.

Geopotential Anomaly Maps

The results presented so far are based on seasonal correlations averaged over five years. However, most of the results can be observed in time sequences of the filtered geopotential anomalies themselves. Figure 10 is such a sequence from 19 August 1981 to 3 September 1981 at 200 mb. The filtered geopotential height anomalies have been divided by the sine of latitude and thus are proportional to stream-function anomalies.

The anomalies C through G on 19 August (Fig. 10a) trace out a wavetrain very similar to those observed in the correlation maps (Figs. 4 and 9). From 19 to 25 August anomalies C, D, and E weaken while F and G increase, in much the same manner as the correlations in Fig. 4. From the 25 August to 3 September, F slides along south of G even-
Figure 10. Geopotential height anomalies divided by sine of latitude for 19 August 1981 to 3 September 1981. The outer, middle and inner latitude circles are 20°, 50°, and 80°S, respectively. The contours are in intervals of ±40.0 gpm and the zero contour has been omitted. Dashed contours represent negative anomalies. The letters A through J denote local anomalies discussed in the text.
tually becoming A, much like the reflection process observed in Fig. 9.

There are some differences between the anomaly maps and the correlation maps. First, the correlation maps suggested little phase propagation while the anomalies exhibit more noticeable phase progression (although the latter mainly occurs when the anomalies are at their minimum). Also, anomalies H and I seem to be a further extension of the wavetrain C through G. Region I is statistically significant in the correlation maps (Fig. 4). The anomaly J appears to propagate southward, possibly from the tropics, and coalesces into the main wavetrain replacing D one-half cycle later. In noting these differences, it should be remembered that the geopotential anomaly maps (Fig. 10) are a case study of a single event, whereas the correlation maps (Figs. 4 and 9) represent 5 year averages.

**Eliassen-Palm Flux Diagrams**

Using Eliassen-Palm (EP) flux diagrams (Edmon et al., 1980), we examined how these 40-50 day fluctuations interacted with the mean flow. We calculated the quasi-geostrophic EP flux

$$F = (-u'v', \frac{fv'T'}{N^2})$$

and the EP flux divergence

$$DF = \nabla \cdot F/\rho r \cos(\theta)$$
from the filtered geopotential heights using the linear wind estimate of Robinson (1986) and Randel (1987). The notation is standard and follows Randel and Stanford (1985). The EP flux vectors indicate the relative importance of momentum flux, $u'v'$, to temperature flux, $v'T'$, in addition they provide information on the direction of Rossby wave group velocity in a zonally averaged sense. The EP flux divergence indicates the net effect of the eddy forcing on the zonal mean flow.

The seasonally averaged EP flux diagrams for each of the five years are shown in Fig. 11. The plotting conventions follow Edmon et al. (1980). Each of the five years reveal the same overall pattern. The EP flux vectors are mostly horizontal above 300 mb and, north of 50°S, generally point toward the equator. Above 300 mb, we found that $u'v'$ was more than an order of magnitude larger than $v'T'$, indicating mainly a meridional group velocity. The EP flux divergence is mostly negative; in fact, no positive values greater than 0.1 m/s day$^{-1}$ were found in any year. The EP flux in the troposphere is negative along a broad range of latitudes and becomes more negative at higher pressures. Typical values for the divergence of $-0.3$ m/s day$^{-1}$ imply a reduction of the zonal mean wind by the eddies of 15 m/s over one cycle of these fluctuations (about 50 days). In the upper stratosphere in 1982 and 1983, the EP flux divergence has another region of relatively strong negative values, although these regions may be part of the erroneous "dipole" in the momentum flux discussed by Randel (1987).

The propagation of energy toward the equator as suggested by the
Figure 11. Seasonally averaged Eliassen-Palm flux for 35-60 day fluctuations for 1980 to 1984, (a) to (e), respectively. All diagrams have the same arrow scaling and the reference arrow above (a) represents $\rho_{oo}4\pi r^2$ of $F_y$ and $0.01 \times \rho_{oo}4\pi r^2$ of $F_z$. Arrows above 100 mb have been increased by a factor of 10. The contours are in intervals of $-0.1$ m/s day$^{-1}$. The solid line is the zero contour; all positive values are less than 0.1 m/s day$^{-1}$.
EP flux north of 50°S is consistent with the results observed in correlation maps in Fig. 4. The EP flux divergence indicates that the eddies are acting to slow down the mean zonal wind, i.e., the eddies are growing at the expense of the zonal mean wind. The consistency between the EP diagnostics and the correlations suggests some reliability in our results.

Energetics

The energy transfer into the eddies indicated by the EP flux diagrams can be confirmed using the Transformed Eulerian Mean energy formalism of Plumb (1983). In Fig. 12, the eddy energy

$$E' = \int \rho_s (u'^2 + v'^2 + \phi_z'^2/N^2)/2 \, dy \, dp/g,$$

the conversion of energy into the eddies by baroclinic processes

$$BC = - \int \bar{u}\bar{T}/H \frac{\partial}{\partial z} (\rho_s \bar{T}/N^2) \, dy \, dp/g,$$

and by barotropic processes

$$BT = \int \frac{\bar{u}}{\cos \theta} \frac{\partial}{\partial \theta} (\bar{u}' \bar{v}' \cos^2 \theta) \, dy \, dp/g,$$

are displayed over each of the 5 years. Each of the quantities in Fig. 12 represents a channel average from 25°S to 75°S, a vertical integration from 500 to 50 mb and an average over one cycle (running mean of 50 days). The rate of change of $E'$ is given as:

$$\frac{dE'}{dt} = BC + BT + FB.$$
Here, $FB$ represents the flux of energy out of the domain and was found to be small in comparison to $BT$ or $BC$.

As observed in Fig. 12, the eddy energy normally grows at the start of the Southern Hemisphere winter and decays in the spring. Throughout most of this period, the eddies are generally extracting energy from the mean flow by baroclinic processes and losing energy to the mean flow by barotropic processes. The baroclinic growth and barotropic decay is similar to the classic baroclinic life-cycle modelled by Simmons and Hoskins (1978), though their time scales are an order of magnitude faster. Corresponding to this order of magnitude difference is an order of magnitude slower growth rate observed here. Similar slow growth rates for large-scale disturbances extracting energy baroclinically from the mean flow was observed in a model by MacVean (1985). Note, however, that generally large energy exchange only occurs when the eddy energy is large which is not typical for instabilities.

The calculated growth rate from $BC$ and $BT$ is larger than the observed growth rate of the eddy energy $E'$. The added dissipation needed to account for the observed growth rate is about 50% of $BT$. Energy flux into the tropics was anticipated as this source of dissipation, but the calculated flux was found to be more than an order of magnitude too small. We speculate that wave-wave interactions not considered in this energy scheme, in particular, energy transfer from the 40-50 day fluctuations to synoptic scale disturbances may constitute the missing dissipation.
Figure 12. Total eddy energy and energy conversion into the eddies from the mean flow as a function of time for 1980 to 1984, (a) to (e), respectively. $E'$ is the total energy of the filtered data, $BG$ is the baroclinic conversion and $BT$ is the barotropic conversion of the zonal mean energy into the eddies. Each quantity represents an average from 25 to 75°S, a vertical integration from 500 to 50 mb, and a time average over 50 days. $E'$ is in units of $10^2 \text{ J/m}^2$, while $BG$ and $BT$ are in units of $10^2 \text{ J/m}^2 \text{ day}^{-1}$. Positive values for $BT$ or $BG$ indicate eddy growth at the expense of the mean zonal flow.
Conclusions

The results of this study indicate that fluctuations in the Southern Hemisphere geopotential heights on the 35-60 day time scale exhibit wavetrain-like characteristics. In filtered one-point 200 mb geopotential height correlation maps, we found that these wavetrains follow an elliptically shaped circuit passing from low Southern latitudes in the Indian Ocean to off the coast of Antarctica near 70°W (Fig. 5). The wavetrains are found throughout a deep layer from 850 to 100 mb; only in the lowest levels is there a noticeable tilt with height. Correlations between a midlatitude reference point and microwave temperature data in the tropics suggest a relationship between the midlatitude fluctuations and the MJO.

The one-point correlations also suggest that wavetrains from high latitudes impinging on the tropics may be partially reflected back into the Southern midlatitudes rather than completely entering the tropics. The reflection may be the result of a critical layer which is expected just north of the region of the observed reflection, while the weaker tropical correlations may result from the evanescent tail of the wavetrain. Also, consistent with the observational results is the emission of a new wavetrain from tropical forcing which was stimulated by the impinging wavetrain.

The most of the results from the one-point correlation maps are also observed in the geopotential anomalies maps from a case study in 1981. The anomaly maps exhibited similar wavetrain propagation and
apparent reflection or stimulated emission from the tropics.

The EP flux diagrams indicate that the 35-60 day fluctuations interact with the mean zonal flow at altitudes mainly below 300 mb. The EP flux also reveals that above 300 mb the momentum flux \((u'v')\) dominates the temperature flux \((v'T')\) and, in addition, suggests a northward group velocity north of 50°S. The EP flux results are generally consistent with the results observed in the 5 year mean correlation maps.

The energetics calculations indicate that these fluctuations grow through baroclinic processes and decay through barotropic processes. To balance the observed growth of the eddy energy with the calculated baroclinic and barotropic energy exchange, an added source of dissipation must be postulated. Since the flux of energy into the tropics was found to be small, wave-wave interactions is suggested as the needed dissipation mechanism. The minimal flux of energy into the tropics may also result from our inability to estimate the divergent wind field from the geopotential heights. Models such as Gill (1980) suggest divergent winds are important for the coupling between tropical forcing and extratropical responses.

Finally, the feedback route discussed by Gao and Stanford (1988) is somewhat different from that observed here. Two reasons may account for the differences. First, Gao and Stanford's route was derived with a tropical reference point while our route was found independently of any tropical phenomena. Second, their route is observed in micro-
wave data representative of 30 to 150 mb mean temperatures which lie in the stratosphere in middle and high latitudes, whereas our geopotential analyses have concentrated in the troposphere. When we use the Gao and Stanford reference point, we obtain a similar propagation path.

Acknowledgments

We would like to thank Francis Crum, Xin-Hai Gao, Gloria Manney, and William Randel for many helpful discussions. This material is based upon work supported jointly by the National Science Foundation and the National Aeronautics and Space Administration under grant ATM-8603943 and on National Science Foundation grant ATM-8722703.

Appendix A: Statistical Assessment of Power Spectra

The confidence levels for the signal-to-background ratios in Fig. 1 are estimated by assuming that each estimate in the periodogram contains 2 degrees of freedom, while the filter (Fig. 3) contains 22 degrees of freedom. Thus, the total degrees of freedom is approximately 44. The ratio of signal-to-background then varies as a chi-squared distribution divided by 44. The ratios for the 70%, 80%, 90%, 95%, and 97.5% are approximately 1.1, 1.2, 1.3, 1.4, and 1.5, respectively.
Appendix B: Local and Global Statistical Assessment of Correlations

A correlation of ±0.4 or larger is considered to be significant at the 95% confidence level. This estimate is calculated from a two-sided student t test with 24 degrees of freedom. The degrees of freedom are estimated as 1/2 of the total number of degrees of freedom in our filter (Fig. 3) since only about half of the five years were used in calculating the correlations (correlations were determined from May to October).

All geopotential one-point correlations maps were tested for global significance. In this study, global significance was determined by a Monte Carlo method described by Livezey and Chen (1983). With the calculation of three-dimensional correlations, the percentage of total volume, rather than the percentage of total area, would be more appropriate to establish global significance. Because of the increased cost in computing (an order of magnitude more), we did not use this approach. Instead, we applied the Monte Carlo method to a single horizontal cross-section at 200 mb. We ran 200 sets of correlations with the reference point replaced by a random time series. The random time series was generated by constructing a red noise time series with the same autocorrelations as the estimated background spectrum in Fig. 2. This time series was then subjected to the same filtering procedure as the geopotential heights to obtain the random time series used in the correlations. This method captured the effects of our filter. From
the 200 cases, 9% of the total area was determined as the critical percentage to establish global significance. For all maps presented here, only those with an asterisk in the upper right corner did not achieve global significance.

References


SUMMARY

The results presented in these studies establish the existence of fluctuations in the extratropical Southern Hemisphere on the 40-50 day time scale. The fluctuations were observed in rawinsonde data at Easter Island (27°S, 109°W) and in global geopotential height fields. The fluctuations are quasi-stationary and exhibit wavetrain-like characteristics that propagate energy along an elliptical path. The vertical structure of these fluctuations is almost barotropic, although energetics calculations indicate otherwise. In particular, the eddies gain energy through baroclinic conversions and lose energy through barotropic conversions. Evidence is found that these fluctuations possibly interact with the tropical 40-50 day oscillation, while retaining a separate identity.

A more complete description of these interactions is suggested as a topic for further investigations. Another question that should be addressed in further studies is the explanation of the quasi-stationary nature of the fluctuations. Also, details concerning interannual variability and connections to the stratosphere are suggested as topics for further investigations.

Finally, the significance of these results is:

(1) They represent an increase in the fundamental understanding of the global scale 40-50 day oscillation;

(2) Models whose domain is limited to the tropics may suffer in certain fundamental aspects: the extratropical-
tropical 40-50 day relationships described observationally here conceivably constitute a feedback mechanism missing from present models but of potential importance in reproducing the observed circulation;

(3) Related to the above is the need for these low-frequency oscillations to be adequately obtained in global circulation forecast models to improve long range weather forecasts. The observational results presented here provide several stringent tests for modelers: the middle and high latitude 40-50 day signals and their quasi-stationary nature.
APPENDIX A: OVERVIEW OF ATMOSPHERIC DYNAMICS

Quasi-Geostrophic Theory

The normal set of equations used in dynamic meteorology consist of the momentum equation,

\[ \frac{d\mathbf{v}}{dt} = -\frac{1}{\rho} \nabla P - 2\Omega \times \mathbf{v} + g + f, \]  
(A1)

the thermodynamic equation,

\[ \frac{c_p}{\rho} \frac{dT}{dt} + \frac{1}{\rho} \frac{dP}{dt} = Q, \]  
(A2)

the mass continuity equation,

\[ \frac{1}{\rho} \frac{d\rho}{dt} + \nabla \cdot \mathbf{v} = 0, \]  
(A3)

and the ideal gas law,

\[ P = \rho RT. \]  
(A4)

A list of symbols are given in Appendix C. If the heating \( Q \) is specified, then these equations represent a system of 5 equations in 5 unknowns. However, the total time derivative is actually the derivative following the flow and is given by

\[ \frac{df}{dt} = \frac{\partial f}{\partial t} + \mathbf{v} \cdot \nabla f. \]  
(A5)
The second term on the right-hand side of Eq. A5 introduces a quadratic nonlinearity into the equations of motion. This term is called advection and can increase (or decrease) the local value of \( f \). For example, if warmer air is located south of Ames, Iowa with a wind from the south, then the wind will advect the warm air northward, increasing the temperature in Ames.

These equations need to be simplified to uncover the essential aspects of a particular type atmospheric phenomenon. A systematic and energetically consistent simplification for atmospheric motions in the midlatitudes with horizontal scales of several thousand kilometers is known as the quasi-geostrophic approximation. This approximation is based on the smallness of the Rossby number (see glossary).

The order-one equations are the hydrostatic and geostrophic approximations. The geostrophic approximation is a balance between the horizontal pressure gradient and the Coriolis force, given as,

\[
\begin{align*}
\frac{f_u}{g} &= \frac{-1}{\rho} \frac{\partial P}{\partial y} \quad \text{and} \quad \frac{f_v}{g} &= \frac{1}{\rho} \frac{\partial P}{\partial x}.
\end{align*}
\]

(A6)

The winds evaluated by this approximation are known as the geostrophic winds and flow parallel to lines of constant pressure. The hydrostatic approximation,

\[
\frac{1}{\rho} \frac{\partial P}{\partial z} = -g,
\]

(A7)

is a balance between the vertical pressure gradient and gravity. This approximation is quite good for large-scale atmospheric motions and al-
lows an easy conversion to pressure as a vertical coordinate (known as the isobaric coordinates). The pressure gradient term in the momentum equations (A1) in isobaric coordinates becomes $\nabla \phi$, where $\phi$ is the geopotential (see glossary).

The measurement of atmospheric variables are normally taken at specified pressure levels (or over a pressure interval as with passive satellite instruments), thus allowing easy comparison with equations. Also, isobaric coordinates simplify the equations by removing the density factor. However, we chose to use a log pressure coordinate,

$$z^* = -H \log \left( \frac{P}{P_0} \right),$$

where $H$ is the scale height ($= R T_o / g$), which retains all the advantages of the pressure coordinate. This vertical coordinate $z^*$ is also approximately equal to the geometric height. This vertical coordinate is used throughout this study ($z^*$ will be written as $z$ from now on).

Both the hydrostatic and geostrophic approximations are diagnostic; they have no predictive capabilities. To obtain prognostic equations (which allow the prediction of future of the various variables), we must go to the next order in the Rossby number expansion of Eq. A1. This yields the quasi-geostrophic approximation

$$\frac{D u}{D t} - f(v_g + v_a) = -\frac{\partial \phi}{\partial x}, \quad \text{(A8)}$$

$$\frac{D v}{D t} + f(u_g + u_a) = -\frac{\partial \phi}{\partial y}, \quad \text{(A9)}$$
The equations are "quasi" geostrophic since small departures from geostrophy are needed to induce changes in the flow. With the change to log pressure coordinates, the hydrostatic approximation becomes

\[ \frac{\partial q}{\partial z} = \frac{RT}{H}. \]  

(A10)

Therefore, the thermodynamic equation (A2), using a similar Rossby number expansion, is

\[ \frac{D}{Dt} \left( \frac{\partial \phi}{\partial z} \right) + wN^2 = \frac{\kappa Q}{H}, \]  

(A11)

where \( w \) is the log pressure vertical velocity, \( N \) is the Brunt-Väisälä frequency and \( \kappa = R/c_p \). This set of equations have become the cornerstone of large-scale midlatitude meteorology.

**Rossby Waves**

Equations A8 through All, together with the continuity equation, can be combined to form one equation with one unknown. Assuming \( Q = 0 \), we obtain

\[ \frac{Dq}{Dt} = 0 \]  

(A12)

where
and is known as the quasi-geostrophic potential vorticity. This quantity is conserved following the geostrophic flow.

Now, consider the flow to be composed of a mean zonal component and a deviation from that zonal mean. For example, the quasi-geostrophic potential vorticity, \( q \), would become \( \bar{q} + q' \) where \( \bar{q} \) is the zonal average of \( q \) (i.e., \( \bar{q} \) is an average of \( q \) around a constant latitude circle). If we assume \( q' \) is a small perturbation to \( q \) then the perturbation equation of A13 is

\[
\left( \frac{\partial}{\partial t} + \bar{u} \frac{\partial}{\partial x} \right) q' + v' \frac{\partial \bar{q}}{\partial y} = 0 ,
\]

where terms containing products of the perturbation quantities have been neglected. Assuming a wave-like solution,

\[
\phi = \exp \left( \frac{z}{2H} \right) \exp(ikx + ly + mz - \omega t) ,
\]

we obtain the dispersion relation for the Rossby wave,

\[
\omega = \bar{u}k - \frac{k \frac{\partial \bar{q}}{\partial y}}{K^2 + \Gamma} ,
\]

where \( K^2 = k^2 + \ell^2 + \left( \frac{f}{N} \right)^2 \) and \( \Gamma = \frac{f^2}{4H^2N^2} \).

Note that if \( \bar{u} = 0 \), the phase speed is westward. From this dispersion relation the east, north and vertical group velocities become
These expressions were derived by neglecting diabatic effects (heating), friction and nonlinear interactions which are at work in the real atmosphere. However, we will use them as a guide for interpreting our results. A more formal development of these results is given in Pedlosky (1979).

Eliassen-Palm Flux

Though the separation into a zonal mean and deviation from the zonal mean proved quite useful, it can also be misleading. For example, consider the zonal mean of the quasi-geostrophic zonal momentum equation (A8):

$$\frac{\partial \bar{u}}{\partial t} = f \bar{v} - \frac{\partial \bar{u} \bar{v}'}{\partial y}.$$  

(A19)

The tendency is to interpret $-\frac{\partial \bar{u} \bar{v}'}{\partial y}$ of Eq. A19 as the only forcing of $\bar{u}$ by the waves, while $f \bar{v}$ represents the forcing by the basic state. However, waves can also generate $\bar{v}$, thus $f \bar{v}$ includes eddy effects (Pedlosky, 1979). Following Edmon et al. (1980), we can introduce the Transformed Eulerian Mean (TEM) equations which explicitly separate the
mean and eddy effects. These equations are:

\[
\frac{\partial \bar{u}}{\partial t} = f v^* - \nabla \cdot F \tag{A20}
\]

\[
\frac{\partial T}{\partial t} + w^* N^2 = 0, \tag{A21}
\]

where

\[
F = (-\tilde{u}'v', \frac{FR}{H} \rho_s \tilde{v}'T')
\]

and \(F\) is known as the Eliassen-Palm flux (EP) flux vector. Here, \(v^*\) and \(w^*\) represent residual velocities that have the component due to the eddies (waves) subtracted out. They are given as:

\[
v^* = \tilde{v} - \frac{R}{\rho_s H} \frac{\partial}{\partial z} \left( \frac{\rho_s}{N^2} \tilde{v}'T' \right), \tag{A22}
\]

\[
w^* = \tilde{w} - \frac{R}{HN^2} \frac{\partial}{\partial y} (\tilde{v}'T') \tag{A23}
\]

Thus, the divergence of \(F\) (DF) represents the sole forcing of \(\bar{u}\) by the eddies. The EP flux vector can also be shown to be proportional to the meridional and vertical group velocities given in Eqs. A17 and A18. Thus, latitude-height diagrams showing EP vectors and contours of DF (EP flux diagrams) can provide information on the group velocity and wave-mean flow interactions.

When \(DF = 0\), the TEM formalism (Eq. A20) indicates that the mean zonal wind is not being changed by the eddies. Thus, if \(\partial u'v'/\partial y\) is nonzero, the eddies are generating a mean meridional velocity that precisely cancels the tendency for eddy momentum flux (\(u'v'\)) to ac-
celerate the mean zonal wind. This special case occurs for steady, conservative, wave-like disturbances and is known as the non-acceleration theorem first recognized by Charney and Drazin (1961).

Energetics

The TEM equations also lend themselves to energetics studies. The total zonal mean energy (kinetic plus available potential energy — see glossary) is

\[
\overline{E} = \int \frac{\rho_s}{2} \left( \overline{u^2} + \frac{\partial \overline{\phi}}{\partial z} \right) \, dy \, dz \quad (A24)
\]

and the total eddy energy is

\[
E' = \int \frac{\rho_s}{2} \left( \overline{u'^2} + \overline{v'^2} + \frac{1}{N^2} \frac{\partial \overline{\phi'}}{\partial z} \right) \, dy \, dz. \quad (A25)
\]

The conversion of energy from the mean zonal energy into the eddy energy is given as (Plumb, 1983),

\[
C(\overline{E} + E') = \int \rho_s \overline{u} \, \overline{DF} \, dy \, dz. \quad (A26)
\]

The conversion of energy is related to DF and the strength of the mean zonal wind. (This result can be derived by multiplying Eq. A20 by \( \overline{u} \).)

Equation A20 has two components,

\[
BT = \int \rho_s \overline{u} \frac{\partial \overline{v'}}{\partial y} \, dy \, dz, \quad (A27)
\]
BT is known as the barotropic conversion, since it depends on horizontal gradients. This conversion is due to convergence momentum flux \((u'v')\). BT is known as the baroclinic conversion since it depends on vertical gradients. This conversion is due to northward heat flux that varies in altitude. Typical midlatitude storm systems develop by baroclinic energy and decay by barotropic energy conversion (Simmons and Hoskins, 1978; Randel and Stanford, 1985).
APPENDIX B: GLOSSARY

- **Baroclinic**: A state of fluid in which the surfaces of constant density and pressure do not coincide.

- **Barotropic**: A state of fluid in which the surfaces of constant density and pressure coincide.

- **Beta plane**: An approximation of the Coriolis parameter $f$, that retains the first two terms in the Taylor series expansion, $f = f_0 + \beta y$.

- **CISK**: Conditional Instability of the Second Kind. An instability mechanism that produces a feedback between the large-scale flow and the small-scale convection. Typically, large-scale flow provides the low level convergence and flux of moisture that causes the convection to develop. The convection then adds energy to the large-scale flow which is able to continue providing convergence and momentum flux. Originally developed to explain hurricanes. Wave-CISK is a CISK instability in which the large-scale motion is wave-like.

- **Energy, kinetic**: One half the density times the horizontal velocity squared. The vertical velocity is so small for large-scale motions that its contribution to kinetic energy is negligible.

- **Energy, potential**: The sum of the gravitational potential energy and the internal potential energy. Using the hydrostatic approximation, the sum of the two terms becomes, $c^T_p$.

- **Energy, available potential (APE)**: A measure of the net potential energy that can be extracted. To extract the total potential energy from a parcel of air it must be brought to the surface and lowered to zero temperature. Thus, the total potential energy is not "available". A measure of how much potential energy is available is the deviation of the potential energy from some average value. The zonal mean APE is the deviation from a global average, while the eddy APE is the deviation from the zonal mean.
<table>
<thead>
<tr>
<th>Term</th>
<th>Description</th>
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<tbody>
<tr>
<td>GCM</td>
<td>Global Circulation Model. A complex numerical forecast model which possibly includes radiation, moisture, and orography.</td>
</tr>
<tr>
<td>Geopotential</td>
<td>The amount of work needed to raise a parcel of air from the surface to an altitude $z$, $\int g , dz$.</td>
</tr>
<tr>
<td>Kelvin wave</td>
<td>A non-dispersive eastward traveling wave. In the present context, a Kelvin wave refers to an equatorial wave in which the meridional wind is identically zero. The typical vertical structure consists of a 180 degree phase shift (although vertical structures are also possible). The Kelvin wave decays away from the equator.</td>
</tr>
<tr>
<td>Monsoon</td>
<td>Strictly speaking, a monsoon is a seasonal wind. In the context used here, it is the season of heavy rainfall over India and surrounding regions that normally occurs from late May to early September.</td>
</tr>
<tr>
<td>MSU4</td>
<td>Microwave Sounding Unit Channel 4. A satellite instrument measuring radiance in the microwave band. It &quot;measures&quot; the mean temperature from 150-30 mb.</td>
</tr>
<tr>
<td>OLR</td>
<td>Outgoing Longwave Radiation. Radiation emitted from the earth in the infrared usually measured by satellites. In the tropics, it is used as a proxy for convective heating. The tropical OLR signal is &quot;cold&quot; when strong convective clouds are present, otherwise it is &quot;warm&quot;.</td>
</tr>
<tr>
<td>Rawinsonde</td>
<td>A method of upper air observation using balloons that transmits temperature, pressure, and relative humidity while being tracked by radar to measure wind speed and direction.</td>
</tr>
<tr>
<td>Rossby number</td>
<td>A non-dimensional parameter used in the formal expansion of the equations of motion (Al). The parameter is given as $U/(fL)$ where $U$ and $L$ represent the velocity and length scales, respectively, $f$ is the Coriolis parameter.</td>
</tr>
<tr>
<td>Rossby wave</td>
<td>A dispersive westward travelling wave that tries to conserve absolute vorticity (actually potential vorticity). The latitudinal change of the Coriolis parameter $f$ is crucial for the Rossby wave. See Appendix A.</td>
</tr>
<tr>
<td>Term</td>
<td>Definition</td>
</tr>
<tr>
<td>--------------------------</td>
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</tr>
<tr>
<td>Scale height</td>
<td>The vertical distance required for pressure to drop by 1/e in an isothermal (constant temperature) atmosphere.</td>
</tr>
<tr>
<td>SSU</td>
<td>Stratospheric Sounding Unit. A satellite instrument measuring radiance in the infrared. It contains three channels that &quot;measure&quot; mean temperatures within about a 16 km slab of the atmosphere centered at 10, 5 and 1.5 mb, respectively.</td>
</tr>
<tr>
<td>Walker circulation</td>
<td>An east-west circulation in the longitude-height plane observed between the eastern and western Pacific Ocean.</td>
</tr>
<tr>
<td>Zonal wavenumber</td>
<td>The number of wavelengths that can fit around a latitude circle.</td>
</tr>
</tbody>
</table>
APPENDIX C: LIST OF SYMBOLS

\( C_{gx} \) Zonal group velocity of Rossby wave, see A16
\( C_{gy} \) Meridional group velocity of Rossby wave, see A17
\( C_{gz} \) Vertical group velocity of Rossby wave, see A18
\( c_p \) Specific heat of dry air at constant pressure
\( DF \) Divergence of Eliassen-Palm flux, \((= \nabla \cdot F)\)
\( E \) Total mean energy, see A24
\( E' \) Total wave energy, see A25
\( f \) Coriolis parameter, \((= 2\Omega \sin\theta)\)
\( f_f \) Frictional force
\( F \) Eliassen-Palm (EP) flux, see A20
\( g \) Gravity
\( g_0 \) Mean sea-level value for \( g \), \((= 9.80665 \text{ m s}^2)\)
\( H \) Scale height, \((= RT_0/g)\)
\( k \) Zonal wavenumber
\( \lambda \) Meridional wavenumber
\( m \) Vertical wavenumber
\( N \) Brunt Viësälä frequency, \((N^2 = \frac{R}{H} \left( \frac{\partial T}{\partial z} + \frac{KT}{H} \right))\)
\( P \) Pressure
\( P_o \) Standard pressure, \((= 1000 \text{ mb})\)
\( q \) Quasi-geostrophic potential vorticity
\( Q \) Heating
\( r \) Radius of the earth
R  Gas constant

t  Time

T  Temperature

\( T_o \)  Mean global temperature, (chosen as about 273°K)

u  Zonal wind, (positive when flowing from west to east)

v  Meridional wind, (positive when flowing north to south)

v\*  Residual meridional velocity, see A22

V  Three-dimensional wind, (u,v,w)

w  Vertical wind

w\*  Residual vertical velocity, see A23

x  East, (longitude)

y  North, (latitude)

z  Geometric height in Eqs. A1 and A6
   Otherwise log pressure coordinate

z\*  Log pressure coordinate, (= -H log(P/P_o))

Z  Geopotential height, (= \( \phi/g_o \))

\( \phi \)  Geopotential

\( \theta \)  Latitude

\( \rho \)  Density

\( \rho_s \)  Standard density, (= \( \rho_{oo} \ exp(-z/H) \))

\( \rho_{oo} \)  Mean air density at sea level

\( \omega \)  Frequency of Rossby wave, see A15

\( \Omega \)  Rotational frequency of the earth, (= 2\pi rad/day)

\('\)  Deviation from a zonal mean
\[ \hat{\theta} \] Zonal mean value

\[ \hat{\gamma} \] Deviation from a global average
LITERATURE CITED


Chen, T. C. 1987. 30-50 day oscillation of 200-mb temperature and 850-mb height during the 1979 Northern summer.


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Finally, to my parents, to whom this work is dedicated, I thank them for their quiet belief that I could accomplish anything I set my mind to.