A Study of the Southern Oscillation and Walker Circulation Phenomenon

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ABSTRACT

A survey of the literature dating back to the early 1920's along with some appropriate statistical studies delineate an atmospheric-oceanic phenomenon of considerable interest. The Southern Oscillation—an oscillatory exchange of atmospheric mass between the eastern south Pacific and Indonesia—and the Walker Circulation—its counterpart in wind circulation—have a time-scale of years and are manifestations of a near-global variation in circulation, clouds and precipitation, centered in the equatorial eastern Pacific. Ocean surface temperatures in this region are intimately involved; in their warmest phase these variations are known as El Niño events. Some evidence that the strength of the Northern Hemisphere subtropical jet stream varies in conjunction with this phenomenon is given. Since a fully coupled atmosphere-ocean model is presently impractical, a set of general circulation model experiments using altered ocean boundary temperatures has been performed with the NCAR 5° global atmospheric model. Simulation of the phenomenon was successful in that many of the observed atmospheric variations are reproduced. A relative, thermally direct circulation is produced which is driven principally by the latent heat of condensation. The interaction between the dynamics and thermodynamics is underscored since atmospheric response to increased ocean surface temperatures occurs westward of the largest temperature increases, and the subtropical jet stream moves off the prescribed change is significantly affected. While the results of this modeling study do not completely illuminate the problem of atmosphere-ocean interaction as a closed-loop feedback process on these annual time scales, they do suggest where the more important areas for future research must lie.

1. Introduction

Over a period of some 55 years meteorologists and oceanographers have published results of research on various phases of what recently has become evident as a coherent ocean-atmosphere phenomenon of considerable interest. This phenomenon in its parts has become known by a variety of names, although the total phenomenon remains unnamed. Since the early 1960's the work of J. Bjerknes (1961, 1966, 1969, 1970) has demonstrated forcefully that these parts are pieces of a coherent phenomenon which is impressive for a variety of reasons: the atmosphere-ocean phenomenon has a spatial scale which extends more than halfway around the globe, it has a time-scale of years, and the evidence is persuasive that it encompasses both the tropics and at least some parts of middle latitudes. The phenomenon is manifest in all significant meteorological variables, namely, pressure, temperature, wind, moisture, cloudiness and precipitation, and many oceanographic variables, namely, surface temperature and currents, thermocline depth and possibly undercurrents. We intend in this study to outline a historical perspective of the phenomenon and its parts and to show that general circulation models (GCM's) of the atmosphere can contribute to its study. The historical survey will be supplemented by some new statistical studies designed to demonstrate the coherence of the parts of the phenomenon.

2. The phenomenon described

a. The Southern Oscillation

The Southern Oscillation (SO) was originally described in the 1920's by Walker (1923, 1924, 1928) and has been the subject of the work of various investigators since that time. A monograph by Berlage (1966) sums up the empirical evidence for the SO with some curious omissions, however. Perhaps the best summary is in Walker's (1924) own words "By the southern oscillation is implied the tendency of (surface) pressure at stations in the Pacific (San Francisco, Tokyo, Honolulu, Samoa, and South America), and of rainfall in India and Java ... to increase, while pressure in the region of the Indian Ocean (Cairo, N.W. India, Port Darwin, Mauritius, S.E. Australia and the Cape) decreases ..." and in Walker (1928) "We can perhaps best sum up the situation by saying that there is a swaying of pressure on a big
scale backwards and forwards between the Pacific and Indian Oceans ...". A figure adapted from Berlage (1966) (Fig. 1) outlines regions of the globe encompassed by the SO. In this figure lines of equal correlation of station pressure with Djakarta, Indonesia (Dj) are shown. Various surface pressure record lengths enter into this figure; the statistical significance of the SO has been established by other means. We are unaware of any published spectral studies of the SO other than the one by Trenberth (1976) who performed cross-spectral analyses on Port Darwin, Easter Island and Tahiti surface pressure over a 32-year interval. We have computed the coherence-square statistic between station pressure at Port Darwin and Santiago (Fig. 2). Monthly averages for every other month from January 1882 through November 1973 were used so that the Nyquist frequency was 1/4 per month, and the series were standardized by subtracting the sample monthly mean and dividing by the sample standard deviation for each month. The bandwidth of the analysis is shown and it is equivalent to 17 degrees of freedom. We note that the 0.95 prior confidence limit for the coherence square for uncorrelated series is thus 0.33. Two features of the maximum in the coherence square need emphasis. First, the phase angle of the cross spectrum (not shown) is almost exactly ±π indicating that the pressures at Port Darwin and Santiago in the SO are out of phase. Further, the bandwidth of the phenomena is rather large—estimates of the width at the "maximum divided by two" values give a period range of 87 to 27 months (7.2 to 2.2 years). The maximum coherence square falls over a period range of about 42 to 33 months (3.5 to 2.8 years). This fact emphasizes what Berlage and others have pointed out—that the SO is not periodic in the conventional sense of the word—but it is certainly oscillatory.

Various combinations of stations have been used to compute an index for the SO from differ-
ences in station pressure. Most commonly the stations used have been Djakarta (D), Port Darwin (D), Santiago (S), Apia, Samoa (A), and Easter Island (E), whose locations are shown in Fig. 1. As will be obvious in a later section, which particular combination is used is not important. From Fig. 1 it is obvious that the “sloshing” back and forth of pressure which characterizes the SO influences a very large area of the globe and the “centers of action”, namely, Indonesia and the eastern Pacific, are apparently rather large.

b. Relationships between the Southern Oscillation and precipitation in the equatorial Pacific

Even before Walker and co-workers were calculating the hundreds of correlation coefficients which led to the definition of the SO, C.E.P. Brooks and co-worker H. Braby published a paper (1921) which was concerned with newly acquired meteorological observations in the central equatorial Pacific. Working with surface observations at Ocean Island (O), eastward to the Line Islands (P,F,X,M) they noted a distinct inverse relationship between the strength of the trades at these equatorial islands and the amount of precipitation received. Furthermore, they noted strong episodic behavior of the monthly precipitation and the trades and strong coherence in the precipitation anomalies at Ocean, and at Fanning and Palmyra in the Line Islands over a distance of 30° longitude. During anomalously wet periods, October 1911–February 1912, and January–May 1915, they noted southwesterly winds rather than the trade easterlies at Ocean Island and weak and irregular winds in the Line Islands. During the notable dry periods, November 1909–May 1910 and January–April 1916, the entire group of islands reported easterly winds. They postulated mean sea level pressure maps which placed an equatorial low pressure area west of the date line in the dry periods, but approximately on the date line during the wet periods.

In 1933, Leighly published a report dealing with surface observations in the Marquesas Islands (9–10°S, 139–140°W). Remarking on the extreme variation of rainfall he noted dry spells there from 1907–11 and again in 1916 and wet periods in 1912–15, 1927 and 1931. In particular, he noted an exceptional relation between rainfall at Malden Island in the Line Islands and pressure at Port Darwin and Apia, Samoa. He said, “The correlation with dry and rainy years at Malden is distinct and unequivocal: dry years at Malden are years of steep gradient [in surface pressure] westward, rainy years of weak gradient.” He noted both the works of Brooks and Braby and of Walker and concluded that the episodic nature of precipitation in the equatorial central Pacific was related to the SO.

Modern interest in this set of relationships was reawakened in the early 1960’s. Bjerknes (1961) became interested in the occurrence of anomalously warm ocean surface temperatures along the western coast of South America and Ichiyi and Petersen (1963) published a paper analyzing precipitation and sea level pressure patterns in the equatorial Pacific in 1957–58. In particular they pointed out the anomalously large precipitation amounts at the islands in the central Pacific and noted the associated changes in the strength of the trades and the anomalous positive ocean surface temperatures (OST) recorded at these island stations. They proposed a coupled sea-air mechanism which was similar to that proposed by Bjerknes.

A final paper will be discussed in this section. Doberitz (1968) placed the relation of the ocean surface temperature, precipitation and surface wind anomalies on a solid quantitative foundation by computing cross spectra between combinations of time series of monthly precipitation at various Pacific Islands and ocean surface temperatures at Puerto Chicama, Peru (PC). Fig. 3 (top), adapted from his work, presents the coherence-square statistic between monthly precipitation for selected combinations of islands in the central Pacific and Fig. 3 (bottom) coherence square for the precipitation at the islands and the ocean surface temperature anomalies at Puerto Chicama.

Confidence limits of 95% on the basis of zero population coherence square have been entered on the ordinate. There is no doubt that significant coherence exists over large distances in the central equatorial Pacific for all resolved frequencies less than ½ per year for precipitation and also for ocean surface temperature-precipitation. Doberitz calculates that between 1935 and 1965 five years selected from the lowest Puerto Chicama ocean surface temperature gave precipitation anomalies of 50% of normal from Nauru and Ocean Islands eastward to the South American coast, while the five highest OST years gave precipitation anomalies of 150% over the same geographical region. He concludes, “Significant teleconnections of rainfall and sea temperature in the tropical Pacific are therefore limited, in practice, to the arid equatorial region and its immediate neighborhood. However, this area between 4N and 10S has a zonal extension of more than 13,000 km.” Surely this spatial relationship, called by Doberitz and others a teleconnection, together with the SO, are among the most remarkable known in meteorology.

c. Ocean surface temperatures and currents

The behavior of ocean surface temperature in the eastern equatorial Pacific has been the subject of
many investigations, most of them concerned with ocean-atmosphere interaction. A comprehensive bibliography on the recurrent appearance of anomalously warm surface water off the South American coast, the El Niño, and its effects is available (Inter-American Tropical Tuna Commission, 1975). In the context of the present paper we note simply that Bjerknes was concerned with the ocean-atmosphere interaction accompanying these warm waters in the early 1960's. Since that time numerous articles have appeared on the general subject of El Niño and the atmospheric phenomena summarized here in earlier sections.

Reflecting our concern with the terminology involved, as expressed in earlier sections, we would like to comment here that the term El Niño, while of ethnic and descriptive value and concise (few letters), describes a phenomenon of considerable interest in a qualitative and somewhat uncertain manner. In the past there has been a tendency for people, from the Peruvian anchovy fishermen who coined the term to dynamic oceanographers who are concerned with understanding it, to use the term in an either/or fashion. That is, a year is either an "El Niño year" or it is not. Thus, workers do not always agree that above average ocean surface temperatures along the Peruvian coast represent an El Niño or not. Some examples of work which arbitrarily define the phenomenon are Quinn and Burt (1972) and Ramage (1975). It seems to us that the fluctuations in ocean-surface temperatures on the time scale of years over the equatorial eastern Pacific very likely encompass a spectrum of amplitudes and areas. Rather than attempting to classify each swing in the temperature oscillation as either an El Niño occurrence or not, the use of quantitative measures of the ocean surface temperature in relation to time series of meteorological and oceanographical variables (e.g., Kidson, 1975 and Trenberth, 1976) is to be preferred.

The interrelationships between the ocean surface temperatures near the equator and the currents in the tropical Pacific have been given by Wyrski (1974). Using the dynamic topography established by deviations of sea level he has established the fluctuations of the major ocean currents over approximately a 20-year interval. Most important are the fluctuations on the time scale of a year to
years in the South Equatorial Current (SEC), a
westward flowing current between about 5°N and
15°S; its counterpart in the Northern Hemisphere,
the North Equatorial Current (NEC) (10–20°N);
and the countercurrent (CC), a narrow eastward
flowing current centered at about 7°N. The NEC and
CC vary in phase and are almost exactly out of
phase with the SEC (on all time scales) with
frequencies of less than 1 per year predominating.
He presents evidence that the annual variability of
these current systems can be accounted for by varia-
tions in wind stress produced by the trade winds.
However, an analysis of the connection between
the trade winds and these currents on time scales
greater than one year, which is a critical point in
the discussion to follow, has only recently been
attempted by Wyrtki (1975) and Reiter (1978a,b).

Fig. 4 presents a cross-spectrum analysis between
every other month of the SO index (defined in
the figure legend) and the standardized Puerto
Chicama ocean surface temperatures, 1925–73. The
coherence-square statistic shows a very significant
feature centered at the frequency of 0.03 per month
(period of 2.8 years), agreeing with the feature
(Fig. 2) of the SO itself. And as with the SO
itself, the significant coherence extends over a rela-
tively broad frequency band. The phase spectrum
indicates that the SO index (as defined) and the
ocean surface temperatures along the Peruvian
coast are out of phase. This result is the strongest,
and from a statistical point of view the most
significant, air-ocean relationship known to us and
forms the basis for linking the oceanic behavior of
the El Niño with the atmospheric SO and its
associated behavior.

d. The Walker Circulation

Because of the prolification of nomenclature
applied to the interrelated ocean-atmosphere phe-
nomena described in this paper, we open this sec-
tion by defining the "Walker Circulation". The term
was originated by Bjerknes (1969) with the follow-
ing statement: "The Walker Circulation . . . must
be part of the mechanism of the still larger 'Southern
Oscillation' statistically defined by Sir Gilbert
Walker [1923 (et. seq.)]. . . whereas the Walker
Circulation maintains east-west exchange of air
covering a little over an earth quadrant of the
equatorial belt from South America to the west
Pacific, the concept of the Southern Oscillation
refers to the barometrically recorded exchange of
mass along the complete circumference of the
globe in tropical latitudes. What distinguishes the
Walker Circulation from other tropical east-west
exchanges of air is that it operates a large tapping
of potential energy by combining the large-scale
rise of warm-moist and descent of colder dry air."
In this same paper and a later one, Bjerknes
(1970) describes this thermally direct circulation
oriented in a zonal plane by reference to mean
monthly wind soundings at opposing "swings" of
the SO and ocean surface temperature anomaly
patterns.

This depiction of the circulation associated with
the redistribution of mass described by the SO
was presaged by Troup (1961). He drew attention
to the marked anomalies in the 200 mb zonal
wind in December 1957–January 1958 over the
equatorial Pacific and Indian Oceans. Positive
anomalies (westerlies) of 5 m s⁻¹ (10 kt) or greater
were observed over the eastern Indian Ocean and Indonesia and corresponding negative anomalies (easterlies) at Canton, Majuro and Christmas Islands in the central Pacific. In a later paper Troup (1967) presents correlation coefficients between the 50 000 ft (150 mb) zonal winds at Canton Island and Singapore with an index of the SO (the difference between monthly mean pressure at Darwin and Papeete, Tahiti). From about 12 years of monthly data he finds a positive correlation (depending on season) of 0.51 to 0.85 with Canton’s zonal 150 mb wind and the SO index; and a negative correlation of -0.31 to -0.77 for Singapore’s 50 000 ft zonal wind and the SO index. While no attempt was made to assess the statistical significance of these coefficients, Troup nevertheless drew attention to the presence of interannual changes in the upper troposphere flow over the tropics and indicated that the anomalies in the flow covered a large range of longitude.

Bjerknes’ (1966) paper did not discuss the particular circulation he later termed the Walker Circulation, but did postulate an ocean-atmosphere interaction in which anomalously warm equatorial surface water could influence the Hadley Circulation, which influence could be felt in middle latitudes. In his 1969 and 1970 papers he clearly sets out the zonally oriented nature of this thermally direct circulation which distinguishes it in essence from the classical Hadley circulation. By comparing zonal wind profiles for November 1964 and November 1965 at a number of rawinsonde stations along the equator, he compared the circulation of a month with typical, cool surface water along the equator in the eastern Pacific with a month with anomalously warm water there. The resulting direct circulation was consistent with a movement eastward of the low-pressure center and as described in the references in Section 2.

Krueger and Gray (1969) examined the same time period and confirmed Bjerknes’ description. In addition, they pointed out that variations in rainfall and cloudiness were consistent with previous work and, importantly, examined the changes in atmospheric stability and equivalent potential temperature \( \theta_e \) at Canton Island. Particularly noteworthy were changes of about 16°C in surface \( \theta_e \) between the “dry” and “wet” periods (about one year apart). Satellite cloudiness measurements were examined and suggested that although the equatorial eastern Pacific remained nearly cloud-free in the “warm water” period, zones of cloudiness developed north of the equator (5–10°N) and along a span of longitude 140–150°W. Moreover, in agreement with the papers described in earlier sections, the major equatorial zone of clouds located (normally) at about 10–20°S and 160–180°E was located during the anomalous period northward and eastward of that position extending along the equator to 175°W.

In a recent paper, Krueger and Winston (1975) examine the circulation at 700 and 200 mb during the 1971–72 period using objectively analyzed upper air maps prepared operationally at the National Meteorological Center, NOAA. In 1972 large-scale positive ocean surface temperature anomalies again appeared in the eastern equatorial Pacific and the changes in the circulation observed were in accord with the Walker Circulation. Important were the reversal from 1971 to 1972 of the zonal wind anomalies at 200 mb over the entire Pacific Ocean and a weakening of the 700 mb easterlies over a region nearly as large.

We have attempted to extend the work mentioned above by aggregating wind data for a number of upper air stations along the equator for all of the various periods considered by the various workers. To present a composite Walker Circulation, we have first identified a number of 3-month intervals (over the last 20 years) at both maximum and minimum of the SO index. That index is defined here as the difference between Santiago and Darwin’s monthly mean standardized station pressure and is the same index as was used in Figs. 2 and 4. A plot of this index appears in Fig. 6 from 1950–73 and the intervals chosen at maximum and minima are shown by the horizontal
bars under the curve. A total of five maxima and five minima were chosen subjectively by referring to both SO indices shown in Fig. 6. For all the months for the maximum and minimum separately, the deviations of the 200 and 850 mb wind components from the respective monthly means were averaged. The results are shown in Fig. 5. Four upper air stations, Cocos Island (C), Port Darwin (D), Canton Island (Ca) and Lima, Peru (L) were used. In the top portion of the figure the situation during cool surface water periods and high SO index is shown. Following Bjerknes, we assume much smaller meridional deviations (no consistent pattern was evident in the data examined) to depict the streamlines. Subsidence is noted over the relative high pressure east of Canton. In the lower portion of the figure the opposite phase of the oscillation is shown. Low pressure is shown east of Canton and the sense of the circulation cell overlying Canton is reversed. A total change of 6.6 m s\(^{-1}\) occurred at 200 mb and 3.3 m s\(^{-1}\) at 850 mb at Canton and a change in the opposite sense but lesser magnitude at Cocos Island. In addition, the time series of monthly zonal wind anomalies are 850 and 200 mb are plotted as the lower curve in Fig. 6. The distinct out-of-phase relationship can be clearly seen, as well as the association with the SO index. In closing this section we emphasize, following Bjerknes, that the circulation depicted in Fig. 5 should not be interpreted literally. The winds shown are deviations from the climatologically expected winds, the exact mechanism by which the vertical mass flux occurs in the circulation is not known, and the ascending and descending branches of the circulation may not exist on the scale suggested by this schematic figure. However, the circulation apparently does exist in the same statistical–physical sense as the classical Hadley Circulation.

3. Summary of the phenomenon

Fig. 6 presents a summary of the course of various oceanic and atmospheric variables dis-
cussed above from 1950 through 1973. The topmost curve is the strength of the South Equatorial Current taken from Wyrtki (1974). The next curve (reading downward) is the ocean surface temperature average over the equatorial eastern Pacific after Allison et al. (1972), and the Puerto Chicama monthly ocean surface temperature anomalies are also given. The solid and dashed curves next in order are the SO indices: the dashed line [after Quinn (1974)] are 12-month running averages of the Easter Island-Darwin differences in sea level pressure; the solid line is the smoothed Santiago-Darwin station pressure differences. Finally, the last curves are of upper wind data. The solid and dashed lines are the 850 and 200 mb zonal wind anomalies for Canton Island. (The Canton aerological data lasted only from 1954 to 1967.) Most of these curves have been low-pass filtered—the exact details for each curve are given in the figure legend or in the cited references.

The statistical significance of the association between Santiago and Darwin station pressure on the time scales of interest (the SO index) has already been established as well as the association between the Puerto Chicama ocean surface temperature differences and the SO Index. The significance of the association of the upper atmospheric variables with the SO Index and the oceanic variables remains to be established. Qualitative examination of the zonal wind anomalies in the lower two curves and the remaining curves is certainly suggestive, particularly in view of the correspondence in respective maxima and minima.

Recently, Reiter (1978b) has contributed two additional time series of relevant variables which could well be added to those in Fig. 6. He has calculated the mean strength of the meridional component of the surface trade winds over a very broad area of the equatorial north and south Pacific. The observations are the same as those of Wyrtki (1974), gathered from ships of opportunity. The average trade wind \( \nu \) component anomalies (from monthly means), low-pass filtered, show a striking in-phase relation to precipitation in the Line Islands and to the El Niño events.

The mutual correlation and particular phase association demonstrated by this collection of oceanic and atmospheric variables is, in our opinion, most impressive. They indicate an atmosphere-ocean coupling with a time scale of years and a spatial scale of tens of thousands of kilometers involving the tropics as well, at the least, parts of the sub tropics.

J. Bjerknes, who first clearly recognized the coherence (in the nonstatistical sense) of the phenomenon, also put forth a physical explanation of the atmosphere-ocean coupling. Indeed his work was not at all statistical but was always very close to the meteorology and oceanography involved. Briefly, his hypothesis is as follows: during the cool equatorial surface water phase of the oscillation the strength of the South Equatorial Current is strong as are the surface trade winds that drive it. Ekman drift results in net surface water divergence along the equator producing cool surface waters by upwelling. The cool water inhibits convection, clouds and precipitation and the so-called dry zone extends along the equator and the tropics into the west central Pacific. The resulting reduced vertical mass and momentum energy flux reduces the intensity of the Hadley Circulation over most of the western Pacific and there is reduced transfer of zonal angular momentum into the subtropics. (We do not subscribe to the use of the term Hadley Circulation to refer to a mean circulation oriented in vertical, meridional planes for restricted ranges of longitude. We are here following Bjerknes’ usage.) Such a reduced Hadley Circulation must have the effect of reducing the strength of the surface trades and when the trades drop to some critical level, the upwelling ceases and warm water appears in the eastern Pacific. The resulting increased air temperature and moisture produce greater instability leading to increased cloudiness and precipitation along the “dry zone”. This convection produces a significant mass and momentum flux in the local Hadley Circulation and increased zonal momentum transport into the subtropics at least in the eastern Pacific. The westerlies there increase and the surface trades which have been weak begin to increase again and the “cycle” begins again.

A key element in the Bjerknes hypothesis is that the regional atmospheric circulation in the equatorial Pacific during the increased OST’s of El Niño event manages to export increased angular momentum to subtropical latitudes, although the hypothesis is not specific on the exact transfer mechanism. We examined this element of the Bjerknes atmosphere-ocean interaction by using the monthly mean 200 mb zonal wind data for Lihue, Hawaii. This station’s upper tropospheric winds are representative of the intensity of the Northern Hemisphere subtropical jet in the longitude range of interest. A calculation of the correlation coefficient between low-pass filtered Lihue, 200 mb zonal wind anomalies and the SO Index for the years 1951–73 gives \(-0.45\). Although there are 276 months in this sample, the normalization and smoothing on both series reduces the appropriate degrees of freedom. The particular filter used on the SO and upper air data uses binomial weights of length 17 (months) and reduces the number of independent observations in the sample by about 86%. Adjusting for the normalization and for the digital filtering the approximate number of degrees of freedom is 33. Thus this correlation is just
significant at the 0.99 confidence level. This result, if characteristic of other regions of the eastern Pacific subtropics, appears to confirm Bjerknes' description of events: namely, that increased eastern Pacific OST's are associated with an increased strength of the subtropical westerlies.

Now there are some difficulties with this reasoning chain which, we believe, Bjerknes himself recognized. Conventional wisdom suggests the time scale of atmospheric circulation changes is short compared to the ocean's circulation changes and the fluctuating Walker Circulation should very quickly respond to the OST change. Thus, the increase in the strength of the trades should occur very quickly (on the time scale of months) after the appearance of the warm surface waters. Such a conclusion indicates that the trades should be strongest at the time of the appearance of the warm surface waters or shortly thereafter. The critical point in the reasoning chain then becomes what is the time scale for the Ekman-drift-produced upwelling to reestablish the cool surface waters? Recently, Wyrski has put forth a very important hypothesis and produced an important missing piece of the observational picture by showing that increased zonal surface wind stress in the region 10°-20°S, 70°-80°W occurs shortly before or during El Niño. Thus, the argument that the appearance of warm surface water is (directly) caused by a weakening of the trades, in the eastern Pacific at least, must apparently be abandoned. Wyrski's proposal for the El Niño mechanism involves the accumulation of water in the western Pacific produced by the strong SEC and strong trades. When a weakening of the trades occurs the east-west sea level gradient cannot be maintained and the water moves quickly eastward modifying not only the surface water but also the thermocline structure. This proposal, however, requires that the atmospheric trades weaken after a gradual buildup; this weakening, to be consistent with Bjerknes' hypothesis, would occur at the time of decreasing Hadley Circulation which as indicated above should be at about the time the cool surface water was reestablishing itself. Thus, this critical piece of evidence (and the resulting hypothesis) indicates that the air-ocean interaction is not in the correct phase to be consistent with a "fast" atmospheric response.

We point out, however, that the interannual variations in both the u and v components of the surface trades throughout the central and eastern Pacific presented by Wyrski (1974) and Reiter (1978a,b) leave in doubt a description of the forcing of the wind stress on the ocean and that role in the El Niño phenomenon. For example, in the area 0°-20°S and 120°-180°W Wyrski's data do show a minimum in zonal wind stress during the warm surface water periods (such as 1957-58 and 1965) and maxima during the cold water episodes (particularly in 1955-56, 1966-67 and 1970-71). Reiter (1978a) approximates the curl of the wind stress in the Northern Hemisphere tropics and his data show that El Niño events generally but not always follow periods of decreasing curl. Certainly, the data set used both by Wyrski (1974) and Reiter (1978a) does not provide a definitive descriptive relation between the variations in wind stress and tropical Pacific OST's in our estimation.

4. Model simulation of the phenomenon

Although testing the hypothesis of Bjerknes, of a strongly interacting atmosphere and tropical ocean, is beyond present capabilities of numerical modeling, general circulation models can still be used to examine complete intra-atmospheric processes. Rowntree (1972) has investigated the effect of altering the ocean surface boundary temperature on one version of the GFDL (Geophysical Fluid Dynamics Laboratory) model. Ocean surface temperature increases in the eastern equatorial Pacific were specified and the model integrations started with two different initial conditions based on observed data. Results were made by comparing days 20-30 after start between the control and anomaly integrations.

The particular version of the GFDL model used was hemispheric and contained a "wall" at the equator. Nevertheless, some impressive changes in the atmospheric flow in response to this altered ocean surface boundary were noted. In particular, a direct thermal circulation relative to the increased ocean temperatures was seen, with increases in precipitation occurring west of the anomaly, and a strengthening of the subtropical jet stream in the eastern Pacific.

Rowntree's work stimulated us to attempt a similar experiment with a version of the NCAR general circulation model. Such an experiment would allow us to remove the artificial constraint of an equatorial "wall" and to compare the differences in the model integrations, control and altered OST's, with the model climatological statistics.

The experimental design used here is similar to that used by Chervin et al. (1976) and Kutzbach et al. (1977). The model control case evolved from an isothermal atmosphere at rest (Day 0) and continued for 60 days. The control case was run with constant January external forcing; i.e., with fixed solar declination and OST distribution appropriate for mean January conditions but with a diurnal cycle of solar heating. Two prescribed change experiments were run to test the sensitivity of the NCAR GCM to tropical Pacific Ocean surface temperature anomalies. The first involved the OST
anomaly distribution in the eastern Pacific shown in Fig. 7a. This pattern was constructed to smooth completely the prescribed OST gradient from the International Date Line to the continent and thereby modify the hypothesized drive for the Walker Circulation. The second used the OST anomaly distribution in the equatorial Pacific shown in Fig. 7b. This pattern is a composite of several real data anomalies and closely resembles the anomaly of Rowntree (1972). Both prescribed change experiments were run with perpetual January conditions to Day 60 in a parallel fashion to the control case. The simulated climate in all cases was analyzed in terms of 30-day mean fields which were determined by averaging over the final 30 days of the simulations to allow the model to reach a quasi-equilibrium state.

The prescribed change response is presented in this paper by subtracting simulated climatic means in the control case from each of those in the prescribed change experiments. For statistical significance testing purposes, a normalized response is constructed using the procedures and estimates of the model's inherent variability from Chervin and Schneider (1976).

5. Description of model

The version of the NCAR GCM used in this study is global in domain and has six layers with a 5° latitude-longitude grid. The model is identical to that described in Kasahara and Washington (1971) with the exception that an adjustment has been added to maintain a constant global average sea level pressure of 1013.25 mb. A detailed comparison of the model's simulated January climate with the observed mean climate may also be found in Kasahara and Washington (1971). The physical processes included in the model are solar and infrared radiation, cloudiness, and boundary and subgrid-scale diffusion of momentum, sensible heat and moisture. The dynamic effects of orography and a hydrological cycle are taken into account. The unmodified ocean surface temperature distribution is specified from Washington and Thiel (1970) and held constant in time at January climatological values. Over land or snow and ice regions the surface temperature is calculated from an energy balance equation having a diurnal variation. An extensive description of the model details can be found in Oliger et al. (1970). Those physical processes which appear to be most relevant to this study (and hence warrant further discussion) are the vertical fluxes of sensible heat \( h \) and water vapor \( w \) in the surface boundary layer, convection and cloud formation.

In the model these surface fluxes are determined from the simple bulk aerodynamic formulations

\[
h = -c_D \rho S (T_S - T_G) V_S, \tag{1}
\]

\[
w = -c_D W \rho S (q_S - q_G) V_S, \tag{2}
\]

where the subscript \( S \) denotes the anemometer level or top of the constant flux surface layer, \( T_G \) is the temperature of the earth's surface (prescribed for ocean grid points but calculated otherwise from a surface energy balance equation), \( q_G \) is the saturation specific humidity at the earth's surface, \( c_D \) is the drag coefficient (set equal to 0.003 for all grid points and wind conditions), \( c_D W = 0.7 \ c_D \) (an empirical factor to reduce the model's evaporation rate), \( \rho_s \) is the density at the anemometer height, and \( V_S \) is the maximum of the magnitude of the model's derived surface wind and 5 m s\(^{-1}\).

Convection is taken into account by means of a convective adjustment which adjusts the temperatures in adjacent layers to mimic the processes.
which maintain the dry adiabatic lapse rate in unsaturated conditions or the moist adiabatic lapse rate in saturated conditions. Temperatures are made to coincide with a moist adiabat for upward motion and with a dry adiabat for downward motion. The adjustment is subject to the constraint that the internal energy remains constant between adjacent layers.

In the model, clouds are permitted at heights of 3 km ($c_{3km}$) and 9 km ($c_{9km}$). They are assumed to be blackbodies for radiation calculations and have thickness small compared to the 3 km vertical mesh size. Cloud fractions are computed from empirically derived formulas assuming a linear dependence on relative humidity (RH), i.e.,

$$c_{3km} = 2.4 \text{ RH} - 1.6,$$

$$c_{9km} = 0.24 \text{ RH} - 0.16,$$

subject to the restriction that $0.2 \leq c_{3km} \leq 0.8$ and $0 \leq c_{9km} \leq 0.08$. The relative humidity (in hundredths) is computed from linearly interpolated mixing ratios and temperatures in the two adjacent vertical layers. The minimum value of cloudiness is also imposed whenever the vertical motion at the particular cloud level is less than $-2 \text{ cm s}^{-1}$ or the relative humidity is less than 75%.

6. Discussion of results

a. Description of circulation differences

In general, the prescribed change responses utilizing the eastern Pacific anomaly and the equatorial Pacific anomaly were qualitatively similar. Except when explicitly mentioned, therefore, the results discussed will be concerned with the eastern Pacific anomaly alone. For each grid point of the GCM computational mesh a normalized response

$$r = |\Delta_{30}|/\sigma_{30}$$

is calculated, where $\Delta_{30}$ is the difference of the 30-day averages, anomaly minus control, for any variable (i.e., the prescribed change response) and $\sigma_{30}$ is the estimate of the standard deviation of 30-day averages of the variable for the GCM version used for the control integration. [The details of this procedure for the NCAR GCM have been given by Chervin and Schneider (1976).] This ratio, proportional to a t-statistic, allows a potential assessment of the significance of the change in model behavior attributable, in this case, to a modified ocean boundary condition. Our significance criterion, $r \geq 4$, resulted from applying the two-sided t-test at a single arbitrary grid point to determine the normalized response necessary to reject the hypothesis (at the 5% significance level) that the prescribed change response is due to random fluctuations and not the imposed prescribed change itself. That is, a normalized response greater than or equal to 4 implies that the probability of such a response arising from mere chance sampling is 5% or less. Although it is difficult to assign a quantitative significance level to the pattern of these r values because of the non-independence of the grid points and the posteriori nature of their selection we have chosen a value of $r \geq 4$ as representing significant change in the behavior of the model.

Fig. 8 presents the prescribed change response for sea level pressure averaged from Days 31 to 60. The stippled areas delineate significant change in the sea level pressure according to the criterion mentioned above. We note that, with the warmer ocean boundary, sea level pressures are lower over the eastern Pacific and most of South America and higher over Indonesia, Australia and the Indian Ocean. The large response south of 40°S is not judged significant and is comparable to the model’s inherent variability. However, the three regions of significant response located in the latitude band between 10° and 40°S seem physically plausible since the maximum prescribed change in SST is also located in this band. The intriguing significant “downstream teleconnection” halfway around the globe in the Indian Ocean just west of Australia will appear consistently in the response charts for other variables. The stippled areas in the Northern Hemisphere appear to be isolated grid-point events without any obvious physical connection to the prescribed change. Consistent with the observations of Bjerknes (1966) for the single winter of 1957–58, when above normal surface temperatures existed in the tropical Pacific from the American coast to the date line, the model response shows a deepening of the Gulf of Alaska low and a weakening of the Icelandic low. However, these changes are not significant at the 5% level and hence cannot be attributed to the OST anomaly with much confidence. Thus, although the total redistribution of monthly-mean sea level pressure for the tropical Pacific does not achieve the level of significance selected, it is qualitatively patterned after that of the Southern Oscillation (see Fig. 1) with a major exception being the different longitudinal locations for the zero correlation and zero pressure response lines. Furthermore, unlike the GCM integrations using ocean temperature modifications at higher latitudes (Chervin et al., 1976), these results show distinctly significant changes in sea level pressure as a result of the modified tropical ocean surface temperatures.

The changes in the wind field are rather more marked than in sea level pressure. For the eastern Pacific anomaly comparison Figs. 9 and 10 present differences in the surface zonal and meridional wind components, respectively. Again (and subse-
Fig. 8. The distribution of 30-day average sea level pressure difference, Eastern Pacific anomaly (Fig. 7a) minus control, for the GCM simulations. Areas in which the observed difference divided by the standard deviation of pressure for the GCM climatology equals or exceeds 4 are stippled.

Consequently, those areas stippled represent $r$ values which are greater than or equal to 4. Very striking changes in the surface trades flowing into the region of the increased ocean temperatures can be noted: the $r$ values reach maxima of 6–7 for both components. The strength of the model trade winds in the eastern Pacific increases significantly in the prescribed change case; indeed, the increase in the zonal component of the surface wind extends entirely across the Pacific. Most importantly,

Fig. 9. The distribution of 30-day average surface zonal wind velocity, Eastern Pacific anomaly minus control. Areas in which the observed difference divided by the standard deviation of surface zonal wind from the model climatology equals or exceeds 4 are stippled.
the changes in the meridional component indicate intensified convergence into the region of the OST increase in excellent agreement with the observed behavior of the trades (Reiter, 1978b).

In the model upper troposphere, significant changes are also found. Figs. 11 and 12 give the component differences for the 10.5 km wind field. Very evident is the strong anomalous easterly flow over the ocean temperature anomaly; in this particular case the influence of the anomaly was sufficient to almost change the weak equatorial westerlies in the control case to easterlies. From the four figures (Figs. 9–12), we may assess the GCM's ability to simulate the Walker Circulation as shown in Fig. 5. In this assessment we recognize that the OST distribution in the control inte-
gration did not exactly correspond to the "cold water" OST distributions representative of the cases composite in the top panel of Fig. 5. (In the control integration long-term mean January OST's were used.) We will assume that the control integration represents the long-term mean and that the prescribed change experiment represents the El Niño "warm water" case, or the bottom panel of Fig. 5. From Fig. 11 we see that the large longitudinal extent of relative easterly flow in the upper troposphere to the west of the OST anomaly corresponds to the easterly anomaly shown at Canton Island. This easterly relative flow, however, extends also over and to the east of the OST anomaly and thus does not correspond with the westerly anomaly shown for Lima. However, relative westerly flow is shown over the Indian Ocean and northern Indonesia which corresponds with the observed anomalies over Cocos Island and Port Darwin, but is displaced northward.

At the surface, the model Walker Circulation fails to exhibit the westerly anomalies in the central Pacific (e.g., Canton Island) but mimics the observed strengthening of the trades over the Indian Ocean and along the South American west coast.

The model results exhibit one very strong feature which was not found in the composite wind field depicted in Fig. 5. The meridional wind component changes (Figs. 10 and 12) are significant, while no consistent meridional wind anomalies were noted in preparing Fig. 5. This difference is most likely caused by the few and the particular aerological stations used for Fig. 5. The data from these stations would not be able to detect actual changes in the mean meridional surface wind component if they do, indeed, resemble those in the model.

In summary, it is quite obvious that the OST anomaly has resulted in a relative, thermally direct circulation in the model which, excepting the central Pacific, mimics rather well the phase of the Walker Circulation shown on the bottom of Fig. 5. Increased inflow and stronger trades are evident at the surface. Since the sea level pressure actually is less in the anomaly case over the region of strongest temperature increase, the outflow in the upper model troposphere must be relatively large. Examination of the figures indicates that the (relative) direct circulation is not symmetric about the temperature anomaly nor is the spatial pattern of divergence/convergence a simple one.

In regions away from the temperature anomaly, the model comparison suggests significant changes occur as a result of the change in boundary conditions. A particularly important feature is the regional increase in the strength of the subtropical jet stream. This can be seen by reference to Fig. 11. Here, over Mexico and the Baja peninsula, and in the Southern Hemisphere off the coast of Chile, the increase in the westerlies can be noted.

Figs. 8–12 pertain to the eastern Pacific anomaly as noted. The results for the equatorial Pacific anomaly are quite similar but less intense, except for the 10.5 km zonal component. Fig. 13 gives the comparable chart to Fig. 11 for this case. Notable in this figure is the larger area of significant
change in the latitudes of the subtropical jet (from the equatorial Pacific anomaly) in the eastern Pacific of the Northern Hemisphere. This difference is due to the different location for the maximum OST anomaly in each case and the nonuniform distribution of the model's inherent variability. In spite of the increase in Northern Hemisphere subtropical westerlies, which corresponds reasonably well with what happens in the real atmosphere, we note that over Australia another significant region of altered 10.5 km zonal flow occurs. This region is characterized by a decrease in the model subtropical westerlies and is exactly opposite to some empirical evidence for the month of October for this region given by Nicholls (1977). This discrepancy is likely due to the failure of the GCM to generate distinctly separate subtropical and polar jet streams and the fact that there is a very strong annual variation in the latitude of the observed wind maximum over Australia.

In the sense of a relative “eddy” circulation, then, the model simulation has mimicked the Bjerknes’ Walker Circulation in the following features: 1) increased surface inflow into the temperature anomaly region manifest as a stronger trade circulation in the eastern Pacific; 2) sufficient vertical mass flux that the upper tropospheric circulation in the region of the anomaly has a more easterly component; and 3) an increase in the strength of the subtropical (westerly) circulation immediately to the north and south of the anomaly.

Comparison of Figs. 1–6 with Figs. 8–13 reveals that the following features of the Walker Circulation are not simulated: 1) the development of lower surface pressure in the central equatorial Pacific; 2) the resultant weakening or reversal of the normal trades in that region; and 3) realistic changes in the strength of the subtropical westerlies at longitudes far removed from the OST anomaly.

b. Description of some model parameterization results

The various physical processes that contribute to the relative, thermally direct circulation set up by the OST anomaly in the model are, of course, nearly all coupled through the parameterizations used (see Section 5). However, it is instructive to examine the differences in sensible and latent heat flux, and convection and cloudiness, to obtain an idea as to their spatial variations and magnitudes.

The relative, thermally direct circulation set up as a result of the OST anomaly is mostly produced by the latent heat of condensation. Fig. 14 shows the change in the vertical flux of latent heat across the surface. The maximum change of 140 W m⁻² occurs directly over the maximum OST anomaly, but a secondary maximum of 80–90 W m⁻² occurs west of the main latent flux maximum. The sensible heat flux pattern (not shown) is similar to Fig. 14 in shape but the fluxes are reduced in magnitude by factors of 4 or 5, e.g., the contour in the vicinity of maximum OST increase has a value of 30 W m⁻². Although there are large regional changes in the surface latent heat flux, the global average change is only 0.78 W m⁻². When compared to the estimated standard deviation of this statistic for the
model climate (0.81 W m$^{-2}$), this global average change is definitely not significant at the 5% level.

Actual variations in these quantities in the equatorial Pacific are not known. Various estimates of components of the surface energy budget appear in the literature, of which the most cited is the atlas by Budyko (1963). In the cases with the increased ocean surface temperatures, however, the sensible and latent heat fluxes are larger than those shown by Budyko, as is to be expected.

Further examination of the model results were made to examine the features of the relative, thermally direct circulation produced by the altered ocean surface temperature. East-west cross sections of various quantities averaged over the latitude band 0$^\circ$ (equator) to 10$^\circ$S were prepared. From these figures, the westward tilt with height of quantities such as specific humidity ($q$) change and equivalent potential temperature change can be seen. For example, the maximum increase in surface specific humidity occurs in the longitude interval 80$^\circ$–100$^\circ$W (coincident with the largest increase in OST), while the largest increase in $q$ at 4.5 km (600 mb) occurs at 120$^\circ$W. The magnitude of the increase varies up to 6.5 g kg$^{-1}$ at the latter location—an amount too large by 60% when compared with actual increases observed for appropriate wet and dry months at Canton Island. In terms of $\theta_e$, the observed data at Canton (Krueger and Gray, 1969) show a maximum difference of 11$^\circ$C at 800 mb, while the model produces increases of 21$^\circ$C at about 600 mb (4.5 km).

Fig. 15 shows the difference in fractional cloudiness at 3 km, eastern Pacific anomaly case minus the control case. The increase in fractional cloudiness of 0.25 occurs near 120$^\circ$W, while the strongest oceanic temperature anomaly occurs (Fig. 7a) at 80–90$^\circ$W. The precipitation change pattern closely resembles the cloudiness change pattern and shows a maximum change of 8 mm day$^{-1}$ at 135$^\circ$W. The significant decrease in the vicinity of 25$^\circ$S, 130$^\circ$W is located at the southeastern end of the simulated cloud band in the basic control case which extends eastward and southward from Indonesia (see Fig. 6, Kasahara and Washington, 1971). This decrease coupled with the increase to the north (and east) mimics the shift in cloudiness and precipitation observed (Trenberth, 1976; Rao and Theon, 1977) except that the model response is displaced too far to the east.

The significant reduction in cloudiness over the Indian Ocean just west of Australia and coincident increase in surface latent heat flux is due to the reduced vertical velocity in that region. The globally averaged cloudiness change is $-0.004$, which is not significant when compared to the estimated standard deviation of this quantity (0.003). Thus, as was the case with the latent heat flux, the model compensates for large regional changes in cloudiness to produce essentially zero net change in the global average. These results are similar to those obtained by Schneider et al. (1978) who performed prescribed change experiments in which positive OST anomalies were
placed in the latitudes corresponding to the descending branch of the Hadley Circulation.

All of these processes result in a dynamic-thermodynamic coupling which produces a relative, thermally direct circulation in response to the warm ocean temperatures. The interaction of the dynamics and the thermodynamics results in a circulation in which the resulting clouds and precipitation are formed west of the maximum temperature anomaly and enhanced surface fluxes. The results here are, again, in excellent agreement with Rowntree's results. However, in the atmosphere the increased cloudiness and precipitation are observed to be principally in the deep tropics from roughly 180°–150°W (Flohn and Fleer, 1975; Doberitz, 1968) farther to the west than the model results indicate. The results of Rao and Theon (1977) using satellite-derived estimates of rainfall from the Nimbus microwave sounder do show, however, more rainfall in the ITCZ in the eastern Pacific in an El Niño year than in a cold water year.

7. Critique of GCM simulation

Two major criticisms may be leveled at the NCAR GCM simulation with fixed ocean temperature boundary conditions. The first, of course, is that the ocean temperatures are fixed thus providing no opportunity to model the air-ocean interaction that is of primary interest. We believe, however, that the experiment has been productive because it permits us to isolate the interactive dynamic and thermodynamic processes in the atmosphere.

There is reason to believe that the incommensurate time scales of atmosphere and ocean motion that hold in the higher latitudes may, in fact, be more nearly incommensurate in the deep tropics. For example, McCreary (1976) has published an ocean model study directed toward investigating the mechanism of El Niño. The principal wave modes that could be responsible for the ocean surface temperature changes of the El Niño are a complex of Rossby waves and equatorially trapped Kelvin waves. McCreary's model gives a phase speed of 200 cm s⁻¹ for these waves which would produce a characteristic time scale for the 10 000 km of the Pacific of about 50 days. The time scale of the oscillatory air-ocean phenomenon of interest is, of course, years. Since McCreary's model provides the possibility that the time scale for the mechanism of the ocean-surface temperature increases is orders of magnitude less than the time-scale of the occurrences of the increases, and provides a number reasonably close to seasonal atmospheric time scales, we may inquire as to the characteristic GCM response times. Such an inquiry is justified because the commensurate ocean and atmosphere time scales suggest that a coupled tropical ocean-atmosphere general circulation model may well be practicable.

For the phenomenon of interest, an additional experiment was performed to impose instantaneously (at model Day 60 of the control case when the model is in a quasi-equilibrium state) the increased ocean surface temperatures (as in Fig. 7a)
and to continue the integration for another 60 days. Fig. 16 shows an example of the rate at which the atmospheric GCM adjusts to the new ocean boundary conditions. We have chosen to show the upper troposphere zonal wind component in the equatorial Pacific, since this particular variable responds in a significant fashion to the altered OST boundary condition. A rough estimate of the response or adjustment time is 10 days to two weeks. Examination of other variables such as surface wind and sea level pressure confirm this estimate. We conclude from this study that the atmospheric adjustment to the onset of the El Niño OST's embodies an increase in the strength of the trades with a time scale less than known time scale for events in the equatorial ocean, but of the same order of magnitude.

However, the point to be made here is the need to couple an ocean and an atmosphere model to investigate the interaction between them and since the response time scales are roughly commensurate and within practical limits for a GCM integration, such an experiment could be reasonably attempted. McCreary's model results underscore this need, also, since the response of his ocean critically depended on the morphology of the wind stress pattern applied.

The second major criticism of the experiment is that the bulk surface transport parameterization and convective adjustment parameterization used in the GCM result in too large a release of potential energy resulting from the increased ocean surface temperature. While this criticism has merit, a number of mitigating comments may be made. The first is that it is impossible to ignore the fact that an increase in OST of an El Niño situation must increase the evaporation and latent heat flux in the eastern and central equatorial Pacific. (However, it may be the case that because of compensatory cooling in other areas, the net, global, latent heat flux does not increase at all.) This increase results in an increase in the potential convective instability of the atmosphere which then is realized, as Bjerknes suggested, in the central and western Pacific. Cornejo-Garrido and Stone (1977) have recently emphasized the importance of the convergence of moisture into the central equatorial Pacific and condensation and heating there as a primary driving force for the Walker Circulation. The particular convective parameterization scheme used by a GCM should thus model this situation in a qualitatively correct fashion and it is perhaps unclear that any scheme that succeeds in doing so, need be correct in detail. Our model results certainly indicate that the latent heat of condensation is realized west of the primary moisture source, as noted above, even though we readily admit that the magnitude of that energy flux may be too large and not displaced far enough westward.

8. Conclusions

The empirical studies outlined in the first portion of this paper demonstrate that a global, large timescale atmosphere-ocean phenomenon of impressive character exists. Investigation of this phenomenon by general circulation modeling seems reasonable because of the space and time scales involved and because of the necessity to use models coupling the dynamics and thermodynamics of atmospheric motion. Rowntree's (1972) initial effort was open to criticism on the basis that he was using a hemispheric model to simulate a global phenomenon involving the interaction of the tropics and higher latitudes. The modeling effort reported here was intended to remove this deficiency, to test the response of a GCM with different model characteristics and physics, and to estimate the atmospheric response time to a rapid change in ocean boundary conditions.

In spite of the acknowledged limitations of a modeling effort in which the ocean and atmosphere are not dynamically coupled, we believe the experiment has been productive. It has isolated specific processes in the atmosphere which could respond to ocean surface temperature changes. The model integrations have been successful in mimicking the broad-scale characteristics of the Walker/Bjerknes phenomenon, although unsuccessful in some of the
details. The modeling suggests that it is possible that the general circulation of the atmosphere is rather more sensitive to changes in tropical oceanic surface temperatures than to changes in midlatitude oceanic temperatures, at least in terms of easily identified relationships between atmospheric and oceanic variables. We, of course, acknowledge that GCM deficiencies in reproducing the apparent intensity of atmospheric baroclinic instability might well affect the previous statement.

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