TASK 2: COMPARISONS OF MODEL SIMULATIONS OF CLIMATE VARIABILITY WITH DATA

Summary of Work Completed in the Current Funding Period

1. Investigations of Climatic Variability

Significant progress has been made in our investigations aimed at diagnosing low frequency variations of climate in General Circulation Models. We have analyzed three versions of the Oregon State University General Circulation Model (OSU GCM). These are: (1) the Slab Model in which the ocean is treated as a static heat reservoir of fixed depth, (2) the coupled upper ocean-atmosphere model in which the ocean dynamics are calculated in two layers of variable depths representing the mixed layer and the thermocline; this model is referred to OSU2 in the following discussion, and (3) the coupled full ocean-atmosphere model in which the ocean is represented by six layers of variable depth; this model is referred to as OSU6 GCM in the discussion. Multi-year integrations of these models were carried out at Lawrence Livermore National Laboratory (LLNL) by Drs. G. L. Potter and W. L. Gates. Our analyses of the climatic variability in these multi-year simulations are presented in the following publications:


Pages 1 thru 17 intentionally omitted.


The Southern Oscillation

The first three papers contain detailed diagnoses of the Southern Oscillation in the OSU GCM and comparisons with observations. The Southern Oscillation signal was initially found in the OSU2 GCM (Sperber et al., 1987) which incorporates the ocean by calculating two layers of variable depth representing the mixed layer and the thermocline. Analysis of a 23-year simulation of global climate with this coupled model revealed that the interannual oscillation of the trans-Pacific sea-level pressure (SLP) gradient is accompanied by anomalies in surface air temperature and precipitation typical of the Southern Oscillation (see Figures 1, 2, and 3 of Sperber et al., 1987). During the course of the simulated warm and cold events anomalies of SST and mixed layer currents evolve in a fashion similar to observations.

An important question is: Where does the impulse for the Southern Oscillation originate? Is it in the ocean or is it in the atmosphere, or does the ocean-atmosphere interaction trigger this oscillation? In Paper No. 1 we have pursued this query by examining the nature of interannual variations
in a 37-year simulation of the same atmospheric GCM without a coupled ocean. This is the slab version of the OSU GCM. In this model the ocean is represented by a mixed layer of constant depth (50 meters) with no horizontal heat transport and no exchange of heat with the ocean below. Results showing that the slab model generates the signal of the Southern Oscillation in atmospheric pressure during the northern spring season but the signal is not propagated to other seasons. The coupled model manifests the Southern Oscillation in other seasons and in several climatic variables in which it is found in nature.

The comparison of these two models also indicates that the trigger for the Southern Oscillation is contained within the atmosphere and that the ocean provides the inertia required to perpetuate the Southern Oscillation to other seasons and other climatic fields in response to the trans-Pacific spring SLP asymmetry. This is supported by observational evidence (Meehl, 1987; Gordon, 1986; Wright, 1985; and Troup, 1965) that the phase transition of the Southern Oscillation (from warm phase to cold phase and vice versa) occurs during northern spring with the interaction of the ocean maintaining and accentuating the ".... alternate strengthening and weakening of the mean west to east exchange of mass from the Indian to Pacific sectors in the atmosphere" (Meehl, 1987).
In Papers 2 and 3 the signature of the Southern Oscillation in sea-level pressure, sea-surface temperature, thermocline temperature, surface air temperature, precipitation and mixed layer currents in the OSU2 GCM simulation has been analyzed and compared with observations.

Of particular interest is the evolution of SST anomalies during extreme phases of the SO. During the low phase the southeast tradewind system is weaker than normal or may even reverse. Associated with this breakdown of the trans-Pacific pressure gradient are anomalously warm SST anomalies, usually on the order of several °C. These extreme warmings, El Niño events, are typically characterized by westward propagation of SST anomalies from the coast of South America beginning early in the calendar year of onset. Towards the end of the year these anomalies merge with those propagating from the western Pacific (Rasmusson and Carpenter, 1982). Thus, these events usually occur in phase with the seasonal cycle. Atypical events, such as the 82/83 and 86/87 El Niños, evolve out of phase with the seasonal cycle. That is, eastward propagation of SST anomalies from the western Pacific dominates with the warm anomalies in the eastern Pacific not developing until late northern summer and early fall.

Evolution of Trans-Pacific SST Anomalies: GCM simulated composite monthly SST anomaly maps for warm episodes (low Southern Oscillation Index) have been generated. These
provide the time sequence of the spatial evolution of SST anomalies in the region 130°E-70°W, 30°S-30°N. Averages of selected months that capture the temporal evolution are shown in Figures 14a-d. These may be compared with the temporal evolution of SST anomalies presented in Rasmusson and Carpenter (1982) based upon the composite evolution for the 1951, 1953, 1957, 1965, 1969, and 1972 El Niños. Following the notation of Rasmusson and Carpenter (1982) we denote months of the year during which the maximum SST anomaly in the eastern Pacific occurs with (0) and months during the subsequent year with (+1) respectively.

In the early spring [March(0) and April(0)], see Figure 14a, the simulated SST anomalies adjacent to South America reach their maximum values corresponding to the timing found by Rasmusson and Carpenter (1982). Subsequent evolution of the model warm event in the eastern Pacific departs from the typical scenario previously outlined. Rather development mimics the evolution of the atypical event. During early summer, [May(0), June(0), and July(0)], see Figure 14b, the spatial extent of the .2°C isotherm in the western Pacific increases with anomalies ≥ .4°C now located near 150°W. In the monthly anomalies, not shown, the SST anomalies in the eastern Pacific became negative during May(0) and June(0) with out of phase positive anomalies developing during July(0). Rapid growth of anomalous conditions occurs during
late summer and early fall. By August(0) the eastern and western 0.2°C isotherms merge resulting in SST anomalies of > 0.2°C that span the tropical basin from South America to Indonesia. From September(0) through November(0) the strength of the anomalies continue to enhance throughout the basin. The resulting mean August(0), September(0), October(0), and November(0) anomalous distribution is given in Figure 14c. The largest SST anomalies of the warm event, ~.9°C, occur near the equator at 115°W during October(0). Thus, the model warm event reaches its mature phase somewhat earlier than seen in observations which show the greatest extent of the warm anomalies between November(0) and January(+1) after onset (Rasmusson and Carpenter, 1982; Philander, 1983). The location of the simulated maximum is adjacent to the location of the negative node of the annually averaged SLP isocorrelation seen in Sperber et al. (1987) which characterizes the primary atmospheric signature of the SO. The subsequent decay of the positive anomalies in the model also occurs earlier than has been observed. In the December(0), January(+1), and February(+1) composite, Figure 14d, the model warm event has decayed with anomalies of ~.2°C spanning most of the tropical and subtropical Pacific.

That the composite model event decays somewhat earlier than the typical El Niño as well as the atypical El Niños is probably related to several factors including the nature of
the mean climate state at the outset of the event and the strength of the anomalies that develop in the atmospheric and oceanic circulation. The interplay of these and other factors in determining the duration of a warm event in the GCM may be investigated in further diagnostic studies. The simulated atmospheric and oceanic anomalies are smaller than found in nature and suggest the possibility that the shorter duration in comparison with observations may be related to the weaker strength of the signal.

**Mixed Layer Current Anomalies:** Concomitant with the variation of SLP between the eastern Pacific and Indonesia is a mixed layer current variation due to the coupling of the ocean currents to the variation of the atmospheric forcing by windstress. During years of low Southern Oscillation Index the SLP gradient across the equatorial Pacific decreases resulting in diminished Southeast trade winds. Firing et al. (1983), Wyrtki (1975, 1977) and Enfield (1981) have deduced from observations that a weakening of the South Equatorial Current occurs during El Niño resulting in an eastward transfer of warm water that is responsible for the anomalous increase in SST in the central and eastern Pacific. Rasmussen and Carpenter (1982) have noted the development of anomalous equatorward flow in both hemispheres in the Pacific during a warm episode. The simulated anomalous ocean current occur not only in the tropical Pacific basin but also in
extra-tropical regions. The current anomalies are shown using velocity vectors of three different lengths: the smallest arrow is for speeds $< 2\text{cm s}^{-1}$, a medium arrow is for speeds $> 2\text{cm s}^{-1}$ but $< 5\text{cm s}^{-1}$, and the largest arrow is for speeds $\geq 5 \text{cm s}^{-1}$. It is interesting to note that the anomalies in equatorial flow converge near $115^\circ W, 2^\circ N$, the grid-point noted earlier as having the strongest anti-correlation in SLP with the Indonesian region. The South Equatorial Current has been observed to reverse in direction, particularly in the western Pacific during El Niño years. The simulated anomalous atmospheric forcing of the currents is weaker than observed, recalling that the amplitude of the model Southern Oscillation Index is $1/3$ to $1/2$ that of the observed Southern Oscillation Index (Sperber et al. 1987). Hence the currents do not reverse; however the anomalies are in the observed direction. These eastward anomalies are supported by tropical flow converging to the equator from both hemispheres, as noted in the empirical observations of Rasmusson and Carpenter (1982).

The temporal evolution of the current anomalies during years of low Southern Oscillation Index can be evaluated by examining successive seasonal anomalies. It will be seen that the Pacific basin SST anomalies, given in Figure 14a-d evolve roughly in parallel with the development of the current anomalies.
During spring, Figure 15a, eastward anomalies of 3-7.8 cm s\(^{-1}\) are located in the equatorial region west of the dateline. The corresponding distribution of SST anomalies, given in Figure 14a, shows the highest SST anomalies in the same region. By summer, Figure 15b, the eastward equatorial Pacific anomalies west of 120°W and east of the dateline have increased in strength to 3-6 cm s\(^{-1}\). Concomitant with the expansion of equatorial eastward current anomalies is the spread of positive SST anomalies into the central/eastern Pacific Ocean. The distribution of SST anomalies in Figure 14b shows the central Pacific as the region of greatest warming. Northeastward flow from 8°S and 4°S in the longitude range 160°W to 130°W converges with the eastward equatorial current anomalies in the area of the largest positive SST anomalies. From the coast of South America to about 120°W westward anomalies of 2-6 cm s\(^{-1}\) generally reinforce the climatological flow. These current anomalies are associated with westward expansion of eastern Pacific positive SST anomalies (clearly seen in the individual monthly composites). In the Fall, Figure 2c, the eastward equatorial anomalies in the Pacific are highly developed with velocities of 5-11.7 cm s\(^{-1}\) extending to nearly 120°W. The SST anomalies, Figure 1c, span the Pacific basin in conjunction with the well developed current anomalies. The equatorward flow in both hemispheres has increased relative to summer together
with greater eastward current anomalies at the equator. In winter, Figure 15d, the current anomalies are highly organized. The eastward anomalies in the equatorial Pacific now range from 6-26 cm s\(^{-1}\). The strength of the equatorward current anomalies in both hemispheres has increased. Equatorward flow in the Pacific now comes from as far north as 16\(^\circ\)N and as far south as 8\(^\circ\)S, particularly west of the dateline. The waning of the positive SST anomalies precedes the decrease of eastward current anomalies in the model. Here we have discussed the anomalous behavior in the Pacific basin. In Paper 2 we show that the simulated ocean currents throughout the globe depart from the mean state in a systematic fashion. The anomalous flow during low Southern Oscillation Index may be characterized as follows in general terms: Flow in the tropical and southern regions of the Pacific, Indian, and Atlantic Oceans opposes the normal circulation. In the regions north of the tropics in the Pacific and Atlantic Oceans the current anomalies tend to augment the usual flow.

It should be noted that high phase cold event simulated SO SST anomalies evolve in the same fashion but with the sign of the anomalies reversed, a known characteristic of the empirical data (Meehl, 1987; Kiladis and van Loon, 1988). During this phase the trans-Pacific pressure gradient is enhanced resulting in mixed layer current anomalies in the
tropical Pacific Ocean which augment the usual flow. Thus the cold event evolution is antithetical to that of the warm event.

**The North Atlantic Oscillation**

An examination of the pressure distribution in the North Atlantic Ocean suggests that there is a tendency for it to be low near Iceland and high near the Azores and southwest Europe, in winter (Walker and Bliss, 1932). The interannual variation of this meridional pressure gradient has been identified as the North Atlantic Oscillation (NAO) and its state is of interest since it is known to influence climates in Europe, eastern North America, and north Africa. The state of the NAO influences the speed and direction of the midlatitude westerlies across the North Atlantic Ocean, which in turn, affect the tracks of low pressure storm systems in the European-sector.

Paper No. 4 presents evidence that the basic characteristics of the NAO are simulated in the OSU2 GCM. The NAO analysis in the model was begun by examining the winter sea-level pressure correlations in the North Atlantic. Figure 16 displays areas showing significant anticorrelation with a region comprising of 9 gridpoints centered around 22°N, 40°W (say region B, seen in the figure as a box lying west of North Africa). The regions significant at 95% level are
shown along with those displaying significance at the 99% level (darker shade). We see that there are three regions—a large region in Canada and the United States, a region north of Iceland, and one in central Europe—that show significance at the 95% confidence level. Each of these has a subregion significant at the 99% level.

It is noteworthy that the 9 stations on the periphery of the North Atlantic Ocean selected by Walker and Bliss (1932) for their definition of the NAO index (shown as solid circles in Figure 16) are located in the proximity of the regions of significant anticorrelation seen in the figure, i.e., instead of just a bimodal nature, the NAO simulation exhibits a multimodal character as perceived by Walker and Bliss. However, the NAO index definition in recent literature has been narrowed to the use of pressure anti-correlation between two centers (Rogers, 1984; Lamb and Peppler, 1987). In order to make comparisons with the observed teleconnections described in these studies, we define the OSU2 model NAO index (NAOI) as the difference between the winter pressure anomalies for two centers, between which the negative correlation is significant at the 99% level. These centers, both shown in Figure 16, were chosen to be the region B defined earlier, and a region A containing the two gridpoints at 74°N, 10°W and 74°N, 15°W. Centers A and B are somewhat displaced from their counterparts in observed data, viz. the Azores and
Iceland, and belong to larger areas correlating negatively with each other. The time series of the index is shown in Figure 17. We see from the figure that the index undergoes significant year to year fluctuations, sometimes as large as 10 mb. This helps in categorizing some years as 'high' NAOI years (such as years 1, 3, 6, 11, 17, 20, 21, 22) and some others as 'low' NAOI years (such as years 4, 5, 7, 14, 15, 18, 19, 23).

Researchers have documented teleconnections involving the North Atlantic Oscillation, and have found it to affect winter precipitation in the northwest African region of Morocco (Lamb and Peppler, 1987), and temperature distribution in southeastern United States (Rogers, 1984) and in the Greenland--northern Europe region (van Loon and Rogers, 1978; Meehl and van Loon, 1979). The OSU2 was tested for simulations resembling the observations in the above mentioned works. It was found that the model simulated NAO influences precipitation in northwest Africa and temperatures in Greenland and Northern Europe, in a way qualitatively to those observed in station data. Furthermore, the model's North Atlantic Oscillation extremes affect surface-level winter air temperatures off the eastern coast of the United States, a teleconnection that occurs in observations as well, though in the latter case, the southeastern region of the country itself is seen to be influenced. An analysis of the simula-
tion of these teleconnections in the GCM will be presented in a separate paper.

The North Pacific Oscillation

Walker and Bliss (1932) state that pressure variations in Hawaii are opposed to those over Alaska and Alberta, this being a manifestation of the North Pacific Oscillation (NPO). The authors state that there is a marked association between the low temperature in the Aleutian Islands and high temperature in south-west Canada. They define an index for the NPO by incorporating the pressures and temperatures at eight locations scattered in Hawaii, Alaska, and western Canada.

Figure 6 of Walker and Bliss (1932) suggests that there is an opposition in winter temperature anomalies between western Alaska and western Canada. Rogers (1981) has used this as an indication of the strength of the NPO and has chosen Edmonton, Canada to represent western Canada-eastern Alaska, and Dutch Harbor and St. Paul to represent the other center of the temperature "seesaw", viz. western Alaska-eastern Siberia. The author has grouped the NPO modes as either Aleutian Above or Aleutian Below, these being Januaries when the differences in departure from normal between the two regions have an absolute value of 4°C. The relationship between the NPO and hemispheric patterns of sea level pressure, air temperature, precipitation, sea surface temp-
Temperature and sea ice, and the temporal variability of the NPO were illustrated by Rogers (1981). In his study, Rogers (1981) has used Januaries and winters (D-J-F) between 1906 and 1978.

The purpose of Paper No. 5 is to present comparable results obtained from the OSU2 GCM that suggest that the modeled physics captures the interannual variations of regional climate represented by the NPO.

In the OSU2, a study of the distribution of the winter sea level pressure (SLP) in the North Pacific Ocean shows air oscillation in two cells, one bounded by 26-38°N, 175-165°W (Hawaii region), and the other by 58-66°N, 165-150°W (Aleutian region), shown in Figure 18. The correlation coefficient between the two centers in the figure is -0.72, significant at the > 99.9% level. The north-south pressure anticorrelation was noted by Walker and Bliss (1932) and in their formula for the NPO, they used the pressures from stations close to these centers. The authors also note that there is an association between low pressure in the northern region of the Pacific with low temperature in the Aleutian Islands and high temperature in southwest Canada. Consequently, temperature data from stations in these regions also contribute to their quantification of the NPO. We note that this pressure anticorrelation associated with the NPO is much stronger than the other pressure oscillations studied in this
GCM. The anticorrelation between the equatorial eastern and western Pacific ocean atmospheric pressure associated with the Southern Oscillation was -0.61 (Sperber et al., 1987) and the anticorrelation between the Azores and Iceland regions associated with the North Atlantic Oscillation was -0.57.

The 23 year time series of SLP for each of the four seasons at the two centers (shown in Figure 18) is shown in Figure 19. We see from the figure that not only has the anticorrelation been graphically captured, but also that the fluctuations are strong in three of the four seasons (summer being the exception). The correlation coefficients for winter, spring summer and autumn are -0.72, -0.63, -0.36 and -0.55, respectively, and are significant at > 90% level. With the exception of summer, they are significant at > 99% level too, suggesting that the study of the NPO in station data in spring and autumn may yield useful teleconnections in this region.

Rogers (1981) has used the seesaw of surface air temperatures in the North Pacific region to distinguish composites of sea level pressure, surface air temperatures and precipitation in the northern hemisphere associated with opposite phases of the NPO. To make an objective comparison with his results, surface air temperatures in the OSU2 were used as well in the analysis of the NPO, and just as in Rogers (1981), January data were utilized.
The region enclosed by 180 and 90°W, and 30 and 90°N, stretching across the North Pacific, when searched for centers of negative correlation in January surface air temperatures, showed an anticorrelation between two 'cells', comparable to those used by Rogers (1981). These were the regions extending from 130 to 120°W and 38 to 74°N, and 170 to 160°W and 30 to 58°N (see Figure 19). Here the former will be referred to as region B, while the latter will be referred to as region A. The correlation between A and B was -0.76, significant at > 99.9% level. Time series of the temperatures in the two model regions is shown in Figure 20 where the anticorrelation between them is apparent. Those years when the departure from the normal between the two regions exceeded 3°C were used to define 'Aleutian Below' (AB) and 'Aleutian Above' (AA) years (Rogers (1981) has used a difference of 4°C to achieve the same with the temperature data of two Alaskan stations and Edmonton, Canada). In observations, an AB mode of the NPO is said to have occurred when both representative Alaskan stations have a negative temperature anomaly in January, and the temperature anomaly is simultaneously positive in Edmonton. When the opposite conditions prevail, the NPO mode is considered to be of the AA type. In the OSU2, we have used a similar criterion in designating the model Januaries. By applying the AB and AA definitions to regions A and B explained above, two cate-
gories of OSU2 NPO modes were obtained. The $3^\circ$C condition in the OSU2 classified years 4, 6, 11, 15 and 20 to be AB years while years 10, 12, 16, 21 and 23 fell into the group of AA years. The relationships between the NPO and hemispheric distributions of sea level pressure, air temperature, precipitation and sea surface are presented in Paper No. 5.

**Study of the Seasonal Cycle in Precipitation using Harmonic Analysis**

Traditionally, comparisons between precipitation simulated in a general circulation model and observations are made in terms of the geographical distributions of total precipitation for a month or a season (Washington and Parkinson, 1986) or in terms of zonal mean values (Potter and Gates, 1984). The general conclusion from such comparisons is that most GCMs provide a reasonably satisfactory simulation of the large scale features of the major rain belts, but do poorly when comparisons are made on regional and local scales. Because the annual precipitation curve, aside from displaying monthly values, only reveals the basic seasonal variation, it is useful to evaluate the data in a more analytical manner. The method of harmonic analysis is useful in that it reduces the curve to six independent terms which describe the tendencies for annually, semiannually, etc. occurring features clearly. Harmonic analysis of precipi-
itation over Africa is presented in Paper No. 6; the results show that a greater amount of detail of the regional precipitation produced by the GCM becomes evident when the seasonal variations are analyzed by the method of harmonic analysis. Previously we have presented a similar analysis for continental United States (Kirkyla and Hameed, 1989).

The usefulness of the method in comparing regional precipitation patterns between GCM simulations and observations can be illustrated by considering the amplitude and phase of the first harmonic.

Figures 21 and 22 present the first harmonic amplitudes (mm/day) for recorded and model data, respectively. The season when the first harmonic reaches its maximum is displayed for the two data sets in Figs. 23 and 24.

The first harmonic amplitudes in Fig. 21 clearly delineate the extent of the Sahara desert. Exceptionally low values, ranging from 0.01-0.04 mm/day, are observed in this region. Although the amplitudes of the simulated precipitation over the Sahara region are the lowest in Africa (Fig. 22), they are in fact erroneously large—approximately ten times greater than the amplitudes obtained from measurements.

The northern portion of Africa may be decomposed into zones of varying precipitation patterns that fluctuate northward and southward during the course of the year. During the summer months the zones are positioned as follows. The
first, located poleward of the ITCZ, rests within the drying influence of the anticyclone located over the Sahara desert, hence precipitation totals are smaller indicated by the smaller first harmonic amplitudes. Southward several degrees latitude, precipitation increases with a summer maximum (Figs. 23 and 24) as surface winds out of the west or southwest provide some maritime influence. Yet farther south, at approximately 10N, the maritime surface current is deeper and precipitation totals are further enhanced. Finally, along the coastal regions of the Gulf of Guinea the stronger maritime influence manifests itself in higher precipitation totals, and accordingly larger first harmonic amplitudes. Trewartha (1981) has discussed this progression of zones for northwestern Africa, but as is seen in Fig. 21 a north to south progression of increasing amplitudes if observed across the entire longitudinal extend of Africa between the Sahara and the equator. This distribution of precipitation in northern Africa, also noted by Griffith and Soliman (1972) seems to be represented by the GCM.

Portions of the northwestern coast within 20N-33N are subjected to the aridifying effects of the cold Canaries current. Western Sahara and neighboring Atlantic have low first harmonic amplitudes in both the model and recorded data (Figs. 21 and 22). The model, however, produces an April maximum in the first harmonic in this region when an autumn
maximum is actually apparent (Fig. 23). The GCM simulates a region with maximum rainfall in autumn a few degrees south of where it is observed. This rainfall represents influx of moist air masses from the Atlantic which is still warm in autumn.

In Jaeger's data the winter maximum in northern Africa (Fig. 23), representing the influence of the polar front, is replaced by a summer maximum near 20°N. In the GCM simulation a large belt with maximum precipitation in Spring is indicated in north Africa. Based on station data Griffith and Soliman (1972, p. 98) find May to be the wettest month of the year over a large region stretching over northern Niger, Chad and Sudan. They attribute the rainfall to infrequent cold fronts moving inland from the Indian Ocean.

During the midsummer months, the southwesterlies penetrate the continent; moist and unstable in character, they bring precipitation to the Guinean coastlands and northeastward into central Africa. Their influence is felt as far eastward as the western edges of the Ethiopian highland (Lydolph, 1985) with a corresponding July-August maximum in rainfall. The summer maximum is observed in both Figs. 23 and 24, although along the northern coast of the Gulf of Guinea, the model produces a maximum occurring later in summer, whereas in nature the maximum along this littoral region occurs in early summer shifting to a midsummer maximum
in the interior. A broad thunderstorm belt stretches from 10N in western Africa, southeastward to Lake Victoria (Lydolph, 1985) with maximum thunderstorm activity occurring during the summer months. These factors, in addition to the incidence of precipitation during periods of high sun, bring more precipitation to the region twenty degrees north of the equator than its southern counterpart, and correspondingly first harmonic amplitudes are larger (Figs. 21 and 22) as is the predominant annual variation in this region.

South of the equator to approximately 20S, the region experiences high sun (southern hemisphere summer) precipitation (Fig. 23), the model reproducing this characteristic accurately (Fig. 24). South of 20S, first harmonic amplitudes distinguish the drier west coast and the moist eastern portion, as the region experiences diminishing effects of tropical controls and falls more under the influence of subtropical anticyclones (Trewartha, 1981) producing a drier western coast. The model reproduces this feature realistically (Fig. 24), as first harmonic amplitudes increase from the western coast eastward.

Aridity along southwestern coastal Africa is enhanced by the combined effects of the stable subsidence of the south Atlantic anticyclone and by the upwelling of sea water in the cold Benguela current that travels northward from the Cape of Good Hope along the coast of Africa, and bends westward into
the Atlantic ocean just south of the equator. Similar to the western coast of South America, the cold waters close to shore strip moisture from the land creating a dry and arid climate. Accordingly the western coasts of Southwest Africa and South Africa reveal low first harmonic amplitudes in both the model and recorded data (Figs. 21 and 22), although the model produces slightly larger amplitudes along coastal South Africa.

During the winter months of the southern hemisphere, the subtropical high pressure cells over the south Atlantic and Indian oceans migrate equatorward exposing the southern tip of Africa to the stormy westerlies and pronounced winter precipitation. Jaeger's data show a June-July maximum along the southwestern coast of South Africa. Results from the OSU GCM also indicate a winter maximum in this region.

**Chandler Wobble as an Atmospheric Fluctuation**

A fixed point on the earth's surface wobbles about the rotation axis at two discrete periods of 12 months and ~14 months, resulting in a variation in latitude with respect to the fixed stars. The annual term has been recognized as a forced wobble caused largely by seasonal shifts in air mass. The 14-month variation in latitude was discovered in astronomical data by S. C. Chandler in 1891 and this periodicity has subsequently been found worldwide in observations of sea
level (Miller and Wunsch, 1973; Currie, 1975) where it is termed the "pole tide", and also in air pressure (Maksimov, 1960).

Most of the problems associated with the wobble, and discussed by Chandler and his contemporaries, are still with us today. Only the lengthening of the period to about 434 days (from the 305 days predicted for a rigid earth by Euler), has been quantitatively explained. Questions not yet resolved are: (1) What excites the wobble and maintains it? (2) Where is the rotational energy dissipated? And (3), is the period constant or variable, or perhaps multiplet structured? Questions 1 and 2 are treated by Lambeck (1980), while Carter (1981), 1982) has reopened discussion of the third. Another topic that has been neglected, except for work by Maksimov (1960), Bryson and Starr (1977, 1978), and Starr (1983) is the possible influence of the wobble on climate.

In Paper No. 7 we present evidence that a signal of mean period 14.7 months in surface air pressure has been produced by the modeled physics of OSU6 GCM. At high northern latitudes the mean amplitude is 42% as large as the annual term indicating that it is an important contributor to climate variability at these latitudes. Evidence is also found for a signal of period 6.7 months, i.e., the sum frequency for the wobble and the annual cycle, and a signal of period near 10
Table 1. Comparison of Chandler wobble amplitudes in mb given by Maksimov (1960) with the OSU6 model, and of the OSU6 wobble with the annual cycle.

<table>
<thead>
<tr>
<th>Latitude</th>
<th>North</th>
<th>South</th>
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<tbody>
<tr>
<td></td>
<td>Maksimov wobble&lt;sup&gt;a&lt;/sup&gt;</td>
<td>OSU6 wobble&lt;sup&gt;b&lt;/sup&gt;</td>
</tr>
<tr>
<td>80-90°</td>
<td>0.70</td>
<td>0.93</td>
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<td>70-80°</td>
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<td>60-70°</td>
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<td>50-60°</td>
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<td>20-30°</td>
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<td>0-10°</td>
<td>0.13</td>
<td>0.11</td>
</tr>
<tr>
<td>Average</td>
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<td>0.49</td>
</tr>
</tbody>
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<sup>a</sup>Average amplitude in mb of the Chandler wobble in surface air pressure as a function of latitude obtained by Maksimov (1960) from 194 stations during the 1903/16 period, and from 182 stations during 1922/35.

<sup>b</sup>Average amplitude of the Chandler wobble in surface air pressure in mb as a function of latitude obtained from the Oregon State University General Circulation Model OSU6 (see Fig. 2).

<sup>c</sup>Average amplitude of the annual cycle in mb as a function of latitude in the GCM.

months, i.e., at the frequency difference between the six month period and the wobble. Signals at sum and difference frequencies are expected in a nonlinear physical system.

Table 1 shows the comparison of Chandler wobble amplitudes generated by the GCM with those found in station data by Maksimov (1960). For both hemispheres upto 60° the calculated wobble amplitudes are from 7 to 15% as large as the annual term. From 60° to 90° the fractional contribution
of the wobble increases dramatically, suggesting that it is a significant factor in climate at high latitudes, as suggested by Maksimov (1960).

Atmospheric Signals at High Northern Latitudes

Our studies of air pressure variations simulated in the OSU GCM led to reports of several atmospheric oscillations (Currie and Hameed, 1988; Hameed and Currie, 1989). These were identified as the quasi-biennial oscillation (or QBO) with period ca. 26 months, and the atmospheric Chandler wobble of period ca. 14.7 months. We also found spectral peaks near 10 and 6.7 months which were explained as the interaction of the Chandler signal with the annual term and its 6 month harmonic; it was further pointed out that another term near 66 months should also eventually be detected.

Such terms (the 66, 10, and 6.7 month terms) are called "combination tones" in nonlinear mechanics, and they are generated when the damping term in the equations of motion varies quadratically.

The OSU GCM calculates atmospheric pressure and other climatic variables at 3312 locations around the globe. This large number of time-series of equal lengths greatly facilitates the identification of statistically significant oscillations in spectra. In Paper No. 8 we present the spectrum for high northern latitudes; this region is of particular
interest because it is known to be characterized by large climatic variability. The results presented show that half of this variability is contained in band-limited signals of known physical origin.

In addition to the signals mentioned above (with periods of ca. 26, and 14.7 months) the high latitude region shows another in the sub-annual time scale with period ca. 40 months which we propose represents one signature of ENSO and denote as the quasi-triennial oscillation, QTO. For monthly sampled data, the QTO, the QBO, and the Chandler signal would interact with the annual cycle and its harmonics at 6, 4, 3, 2.4, and 2 months to theoretically produce 33 difference and summation tones. We report evidence that the GCM has simulated 29 of these 33 tones at 216 locations between 74° and 82° latitude north. In terms of total variance, 51.1% is accounted for by the signals and 49.9% represents pink noise.

Since the spectrum of atmospheric pressure on these time scales is not continuous, but consists of a large number of discrete and physically explained lines, the climate system may not be characterized as chaotic on these time scales. Attempts to describe atmospheric variations as a chaos process (see Pool, 1989) originated from simulations of atmospheric circulation in highly simplified models (Lorenz, 1963). In these simulations the temporal variation of solar radiation was neglected, and thus the physical phenomena
Table 2. Percentages of variance contained in three classes of signals and noise in three latitude zones.

<table>
<thead>
<tr>
<th>Latitudes</th>
<th>Sub-annual signals $T_1$, $T_2$, $T_3$</th>
<th>Annual + harmonics</th>
<th>Combination tones</th>
<th>Sum of signals</th>
<th>Noise</th>
</tr>
</thead>
<tbody>
<tr>
<td>86-90</td>
<td>5.4%</td>
<td>11.4%</td>
<td>35.1%</td>
<td>51.9%</td>
<td>48.1%</td>
</tr>
<tr>
<td>74-82</td>
<td>4.4</td>
<td>27.6</td>
<td>19.1</td>
<td>51.1</td>
<td>49.9</td>
</tr>
<tr>
<td>58-70</td>
<td>1.7</td>
<td>58.0</td>
<td>5.0</td>
<td>64.7</td>
<td>35.3</td>
</tr>
</tbody>
</table>

listed in columns 2-4 of Table 2 could not be produced; only the noise listed in column 6 was simulated and identified as the dominant characteristic of weather and climate.

The analysis of climate simulated in the more realistic General Circulation Model presented here demonstrates that inclusion of the correct temporal and spatial variation of solar radiation input into the atmosphere-ocean system is essential to a physically proper description of the system. The periodic solar variations impose order on the system as described in this paper.

2. Analysis of Station Data

Our work on instrumental meteorological data has focused on questions of reliability of data. Investigations to evaluate data quality of sunshine and precipitation measurements were presented in the following papers:


The Issue of Change of Sunshine Instruments

Duration of sunshine measurements have been made in the United States since 1891, thus generating a useful record for correlations with solar radiation and daytime cloudiness, and for examining climatic fluctuations. Hoyt (1977) has demonstrated that sunshine-derived values of cloud cover are in good agreement with results drawn from satellite measurements, and strongly suggest that for radiation budget or climate modeling studies, either of these values is closer to the "true" cloud cover than conventional ground-based observations of cloud cover by meteorological observers. Sunshine duration is a useful parameter in understanding climate and regional solar energy availability, and a major influential factor in the agricultural economy. Spatial and temporal changes of United States sunshine duration have been studies (Changnon 1981; Angell et al. 1984). Some analyses have been
restricted to periods not encompassing the entire period of observation (Angell et al. 1984) because of a possible contamination in the sunshine record, due to a instrument change at all stations in the United States in 1953.

The change of instrument issue was addressed by Powell (1983), who by comparing mean sunshine duration values in Phoenix, Arizona, for the periods 1896-1953 and 1954-81, established statistically significant differences between these lengthy periods of data monitored by different instruments, and thereby ignored the possibility of any long-term secular trends that could have influenced his results. In Paper No. 10 we have restricted our analysis to data for short periods immediately before and after the change of instrument in the United States. This way, we have attempted to eliminate the interference of secular trends in our results. By subjecting the short record to the Student's t-test for differences between two means and the Mann-Whitney U-test, statistically insignificant differences between data recorded by the two different instruments were found for a large majority of stations in the United States, thereby suggesting that any contamination of the record due to instrument change is small in comparison with natural fluctuations in the data, and may be ignored in discussions of climatic change.

The sensitivities of the thermometric recorder (0.37 cal cm\(^{-2}\) min\(^{-1}\)) and the photoelectric switch (0.12 cal cm\(^{-2}\)
min⁻¹) indicate that the latter may be expected to record greater sunshine durations. The results of the Student's t-test and the Mann-Whitney U-test suggest that any such increase in the registered sunshine is small compared with natural fluctuations in the data. Thus, it seems that the change of instrument did not introduce an inhomogeneity so large that it would exclude the use of these data in examining secular changes in the data series bridging the time of instrument change. We have used this argument to study the variation of United States sunshine durations from 1908-84 in seven regions of the country in Paper No. 9.

In view of the importance of the sunshine data in a number of applications related to sunshine availability, it would be useful to more accurately establish the difference made by the change of instrument. We suggest that a direct procedure for this purpose would be to reestablish the thermometric recorder at several stations so that sunshine duration may be recorded simultaneously by both instruments for several years.

Trends in United States Precipitation

There has been considerable interest in estimating secular trends in precipitation data in various regions of the world. It is therefore important to ascertain the manner in which errors of observation affect estimated trends. In
Paper No. 11 we have compared trends at 1219 stations in the contiguous United States for two data sets: (a) original observations, also called "raw" observations, and (b) the observations adjusted to compensate for suspected errors. The adjustments were made at the National Climate Data Center, Asheville (Quinlan et al., 1987; Karl and Williams, 1987). In order to focus on the effects of observational errors we attempted to avoid the effects of filling of missing data by limiting the analysis to the period 1940-1984 for which the number of missing values is much smaller than earlier periods. A least-square linear regression was performed on the raw and adjusted data for each station and the slopes of the fitted lines were compared. The comparison was made for monthly, seasonal and annual precipitation values.

The results for annual precipitation showed that 23 percent of the stations have trends of opposite signs in the raw and adjusted data. The trends were identical in annual data at only 11 percent of the stations. When monthly data are combined to form seasonal and annual averages the magnitude of the difference between the slopes of the adjusted and the raw observations generally increases, indicating that the errors in the individual monthly observations are correlated. When the station data were averaged to obtain state-wide averages, the effects of the errors became less pronounced in most of the states.
These results indicate that trends obtained from station data may be misleading, unless a careful analysis has shown the errors of observation to be small.

If the errors are estimated to be relatively large then the adjusted data set also may not yield the true trend. There are several reasons for this skepticism. Procedures for adjusting data are generally based on correlations with a number of neighboring stations. The choice of the stations to be included in the correlation is nonunique and depends on the judgment of the investigator. Moreover, the calculated correlations between a given station and its neighbors vary with time; these may not be applicable outside the period for which they were calculated. Even within this period, an adjustment procedure can usually be justified only on a time averaged basis; it has unspecified accuracy for individual measurements.

3. Studies of Climatic Variations in China in Collaboration with the Institute of Geography, Chinese Academy of Sciences

China has a rich legacy of documents describing climatic and agricultural conditions in historical times. Examples of such documents include official histories compiled by county, provincial and federal authorities as well as personal diaries, travelogues and other literature. Professor Gaofa Gong of the Institute of Geography of Academia Sinica is a
visiting scholar at Stony Brook sponsored by U.S. DOE since November 1988. The results from our collaboration with him have been presented in the following papers:


History of Moisture Condition Variations in China for the Last Two Thousand Years

Professor Gaofa Gong has collected documentary reports on floods and droughts in different parts of China. Although the oldest records found are 3000 years old, the data from that period is chronologically discontinuous. In Paper No. 12 we present an analysis of time series of drought and floods found in the historical records for the past two thousand years.

The method followed in collecting the data is as follows. The number of floods and droughts reported in the official record of each "Fu" (or "Xian"), the administrative unit roughly equivalent to a county, were arranged as separate tables of floods and droughts for each year. A flood report refers to any description of extraordinary wetness in
the records studied. The reports usually mention undesirable events associated with drought or flood occurrence.

The provinces for which data are discussed are Gansu, southern sections of Inner Mongolia, Shaanxi, Shanxi, Hebei, Henan, Shandong, Anhui, Jiangsu and Zhejiang. This area can be divided into three climatic subregions, as shown in Figure 25, according to the classification given in the atlas published by the State Meteorological Administration of China (1966). The semi-arid area to the north-west ($I_1$) receives about 300-500 mm of rain per year at present. To the south and east of it is the semi-wet region ($I_2$) with the current annual rainfall of nearly 500-700 mm. The region ($I_3$) to the south, designated as the wet region receives abundant rainfall amounting to nearly 700-1000 mm per year.

Previously, drought-flood series for China in historical times have been presented by Chu (1926) and Yao (1943, 1944). These investigators covered almost the whole extent of eastern China, from Hebei province in the north to Guangdong province in the south. There are relatively few records for regions south of Changjiang (Yangtze) river before the 15th century. Our investigation was limited to the part of eastern China shown in Figure 25 to maintain a homogeneous data base. Chu's and Yao's data were based on the following sources: for the period 206 B.C. to 1643 A.D. both used "Cu jin tu shu ji cheng" (The anthology of ancient and modern
books), prepared by the Qing Imperial government in 1726 A.D., in which a volume is dedicated to astronomical and meteorological observations. For the Qing period (A.D. 1644-1911) the source of data for Chu (1926) was "Dong Hua Lu" while Yao (1943, 1944) used the information given in "Qing Shi Gao", these being different compilations of the history of the Qing Dynasty. In addition to anthologies such as these we have utilized the information given in more than five thousand local manuscripts. Addition of the historical sources in each county greatly increased the spatial and temporal density of the data. The total number of reports in each of the previous studies was less than fifteen percent of the present investigation. Time resolution of Chu's and Yao's series was 100 years. With the data available to us continuous series could be constructed with the time resolution of 5 years. The time series developed for the semi-wet area covers the years 1 to 1950 A.D., while for the semi-arid and wet regions the results are presented for the period 201 to 1950 A.D.

Because of the large number of data reports multiple references to the same event are usually found. The redundancy in data facilitated quality control of the analysis because in the vast majority of events the various reports are mutually consistent. Where disagreements occur, common sense rules were used to choose among the conflicting re-
ports. Comparisons were made with neighboring counties to seek consistency. In many instances the disagreeing reports about an event originate from chroniclers of different periods. In such cases the earlier reporters were assumed to be more credible than later ones. We defined the 'moisture index' \( I = \frac{2F}{F+D} \) where \( F \) and \( D \) are the numbers of reported flood and drought events during a 5 years interval.

From the definition we see that the index takes the values \( 0 \leq I \leq 2 \), the larger values reflecting wetter conditions. Four indices were developed, \( I_1 \) for the semi-arid area, \( I_2 \) for the semi-wet area, \( I_3 \) for the wet area and \( I_w \) for the whole area.

A continuous record of monthly averages of instrumentally measured precipitation is available for Beijing from 1881 onwards. Beijing is situated in the semi-wet area shown in Figure 25. Five-year averages of summer precipitation were calculated and compared with the moisture index \( I \) in Beijing derived from the traditional data base of locally recorded number of drought and flood events. The comparison between the instrumental data and the moisture index indicates that the index gives a reasonable representation of monsoon precipitation. The correlation coefficient between the index and instrumental precipitation is 0.91 when June to September totals are taken and 0.85 for June to August totals.

In Paper No. 12 we focused on climatic changes on the time scale of centuries. In order to examine such long term
trends we have calculated 15-point (i.e., 75 year) moving averages of the indices $I_1$, $I_2$, $I_3$, and $I_W$. These are shown in Figure 26.

The mean index values were 0.67 (semi-arid region), 1.21 (semi-wet region), 1.29 (wet region) and 1.19 (whole region). These long term mean values may be used to distinguish dry and wet epochs for each of the regions. Cramer's test was used to identify periods in which the moisture condition differed from the mean in a statistically significant fashion. In Fig. 26 intervals marked by three, two and one dots differ from the mean index at greater than 99%, 95% and 90% confidence, respectively.

We can see that the fluctuations of moisture condition were different in the three climatic regions. The variation in the semi-arid area ($I_1$) was characterized by two waves, each lasting for a millennium. The minimum and maximum of the first wave occurred around the years 400 A.D. and 850 A.D. The trough of the second wave was marked by two minima, around 1200 A.D. and 1450 A.D., while the peak occurred in the 19th century.

The fluctuation of moisture condition in the semi-wet area ($I_2$) was very similar to the semi-arid area until the 11th century. Between the 13th and 18th centuries these regions had opposite phase of the moisture condition.

The variation of moisture condition in the wet area ($I_3$)
was quite different from the semi-wet and semi-arid areas. There were seven waves of alternating conditions during the last 18 centuries marked by peaks in moisture around the years A.D. 270, 450, 700, 1000, 1300, 1600 and 1900.

We note that the frequency of alteration between dry and wet conditions increases from the semi-arid region to the semi-wet region to the wet-region, i.e., regions that are wetter on the average experience greater variability in wetness.

The composite curve for the whole area ($I_w$) also shows seven prolonged epochs of alternating dryness and wetness. Among these, the dry epoch from 326 to 630 A.D. was the longest. Extremely dry conditions prevailed for several centuries over the semi-arid and semi-wet regions during this period.

The variations of moisture in the whole area were compared with reported changes in the level of the Caspian Sea. Broad agreement between the two data sets suggested that variations in moisture in distant sectors of mid-latitude Asia are similar on the time scale of centuries.

**Comparative Reliability of Dynastic and Local Records in Observations of Sunspots**

Because of widespread interest in the history of the sun and suggestions that solar variations may impact climate of
the earth, attempts have been made by several investigators to extend the record of sunspot observations to the pre-telescopic era. These efforts are based on information found in ancient astronomical documents from the Orient. Several catalogues of naked-eye sunspot sightings have been published, and lists of these may be found in Yau (1988) and Wittman and Xu (1988). The most complete record to date has been published by Yau and Stephenson (1988). It consists of 235 reports of sunspot sightings from 165 B.C. to 1918 A.D. 194 of the reports are from China, 37 from Korea, 3 from Vietnam and 1 from Japan. Each record consists of a description of the sighting together with the data of the observation.

Major dust storms were recorded by local officials in China during historical times. In severe events the thickness of dust in air is usually such that its precipitation on the ground builds up to measurable depths over short periods of time. This phenomenon is known as Yutu or 'Dust-Rain' in the historical documents. The available record of the number of dust-rain events for the last 2000 years was presented by D. Zhang (1982).

In Paper No. 13 we have compared the decadal frequencies of naked-eye sunspot observations with those of dust-rain events, C-14 and the telescopic sunspot observations. The C-14 deviation values were taken from Figure 1 of Stuiver and
Braziunas (1988). The non-telescopic sunspot record is divided into pre-1620 and post 1620 data. The separation was made at 1620 because there are only 4 court reported sightings in subsequent years. We note that:

1) Naked-eye observations prior to 1620 are significantly anti-correlated with the C-14 data while those after 1620 are not.

2) The pre-1620 naked-eye observations are not significantly correlated with dust events. The correlation between post-1620 sunspot reports and dust rain reports has statistical significance at >99% level.

3) The post-1620 naked-eye reports are poorly correlated with their contemporary telescopic observations. The latter are highly anti-correlated with the C-14 values as expected. The small correlation between telescopic sunspot data and dust-rain events shows that solar activity does not influence atmospheric turbidity.

The results obtained above show that the pre-1620 naked eye sunspot observations are a more authentic record than the later data. This is explained by noting the difference in their sources. All reports in the post-Ming period are based on provincial documents. Upto the Ming dynasty the reports are, with very few exceptions, official records maintained by the court astronomers. Needham (1959) has
underscored the distinction between astronomy and other sciences in ancient China by noting that astronomical observations were considered especially important by the emperor who depended on them not only for omens but also for forecast of droughts and floods. We may expect, therefore, that in general astronomical observations were made and recorded with care as is illustrated by the voluminous record of eclipses. It is likely that the sun was looked at on a frequent basis and those sunspots that were observed were recorded. The small number of records can be explained by noting that the spots could not be observed if smaller than a certain size.

The sightings in the local documents are in a different category because regular observations were not made outside the royal court. The local observations were probably made by interested individuals who observed the sun when possible. The occurrence of haze in the atmosphere therefore played a much greater role in these sightings.

Proposed Research

1. Global Climatic Oscillations and Their Teleconnections

Our work so far in this project has shown that GCM's simulate important aspects of interannual variability and can therefore be potentially useful tools in understanding the variability of global and regional climates.
END

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