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A note on Bjerknes' hypothesis for North Atlantic variability

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ABSTRACT

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On decadal time-scales the historical surface temperature record over land in the Northern Hemisphere is dominated by polar amplified variations. These variations are coherent with *SST* anomalies concentrated in the Northwest Atlantic, but extending with lesser amplitude into the North Pacific as well. Bjerknes suggested that multi-year *SST* anomalies in the subpolar North Atlantic were due to irregular changes in the intensity of the thermohaline circulation. In support of the Bjerknes hypothesis there is evidence that winter overturning in the Labrador Sea was suppressed for a brief period from 1967–1969 by a cap of relative fresh water at the surface. Cause and effect are unclear, but this event was associated with a marked cooling of the entire Northern Hemisphere.

The difference in *SST* averaged over the Northern Hemisphere oceans and *SST* averaged over the Southern Hemisphere oceans from the equator to 40°S is coherent with Sahel summer rainfall on decadal time scales. Empirical evidence is supported by numerical experiments with the British Meteorological Office atmospheric climate model which simulate augmented monsoonal rainfall in the Sahel region of Africa in response to realistic warm *SST* anomalies in the Northwest Atlantic. A coupled ocean–atmosphere global model exhibits two equilibrium climate states. One has an active thermohaline circulation in the North Atlantic and the other does not. The two climate states provide an extreme example which illustrates the type of large scale air sea interaction Bjerknes visualized as a mechanism for North Atlantic climate variability on decadal time-scales.

Introduction

A formidable barrier to empirical identification of greenhouse climate warming is the fact that the atmosphere has large natural climate variations, which at the present time are very poorly understood. Of particular interest is the surface warming of the Northern Hemisphere during the 1930's decade and the subsequent cooling during the 1960's. The Northern Hemisphere historic record of surface temperature appears to be dominated by two basic types of large-scale, low frequency variability. On a three to four year time-scale the El Niño phenomenon is most important. Its greatest amplitude is in the the equatorial region, but it also extends very high into the atmosphere. Pan and Oort (1983) show that for the period 1960–1975 the pressure and area averaged temperature of the entire globe is highly correlated with sea surface temperature (*SST*) in the central equatorial Pacific. Figure 1 shows zonally averaged air surface temperature over land for the entire period 1880–1980 from a compilation by Hansen and Lebedeff (1987). Note that the variability at higher latitudes is much larger than that at lower latitudes. A polar amplified climate variability is indicated. The time-scale is very long, over ten times that of the typical El Niño.

The existence of melting sea ice tends to moderate interannual variations of surface air temperature above the polar ocean in summer (Manabe and Stouffer, 1980). A more detailed examination

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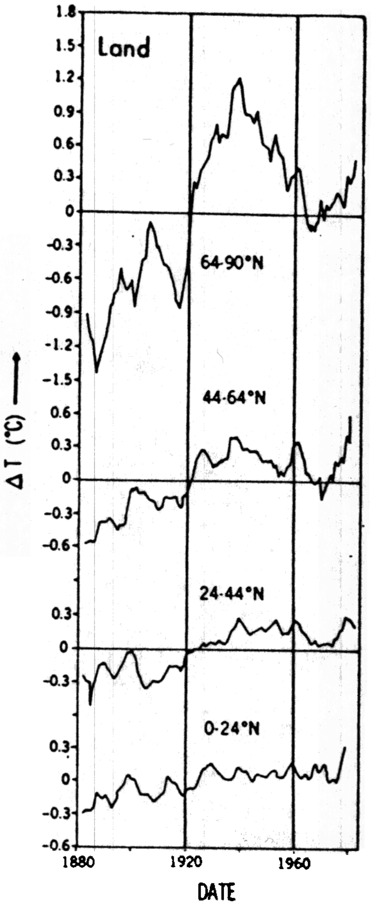

Fig. 1. Zonally averaged surface air temperatures over land in the Northern Hemisphere for different zonal bands (Hansen and Lebedeff, 1987). The curves are based on five year running means. Note the pronounced polar amplification.

of the temperature time series for the Arctic (Kelly et al., 1982) showed that the maximum contribution to this polar trapped climate variability shown in Fig. 1 is due to temperature variability during the autumn, winter and spring. The Hansen and Lebedeff data is taken over land only. A statistical analysis of the record of global *SST* is shown in Fig. 2. Folland et al. (1986a) have computed the EOF's (empirical orthogonal functions) from the British Meteorological Office historical *SST* data set from 1901–1980. The EOF's are based on the covariance matrix and the contribution of 10 degrees of latitude by 10 degrees of longitude squares are weighted evenly, regardless of area. All seasons are lumped together. The three EOF's that account for the largest share of the variance are shown in Fig. 2. The first EOF is of one sign almost everywhere. Its amplitude (not shown) has a nearly monotonic increase during the entire period. The details of this trend may be somewhat suspect because methods of measuring *SST* from ships have changed over the years. A more detailed discussion of this point is contained in a recent paper by Jones et al. (1986) where corrections have been worked out to eliminate some of the possible instrumental bias.

The second EOF is concentrated in the Pacific Ocean and has the familiar signature of the El Niño phenomenon. Note that the highest amplitude is on the equator with two lobes of opposite sign in the subtropics. The amplitude oscillates with a period of three to four years. The third EOF is concentrated in the Atlantic with the largest amplitude in the subpolar region of each hemisphere and almost no amplitude in lower latitudes. Note that the anomaly has the opposite sign in each hemisphere.

The first EOF of Folland et al. (1986a) represents a nearly global upward trend in *SST* since the beginning of the temperature record. The second EOF is the signature of the El Niño phenomenon and the third EOF is associated with a polar amplified climate variation with a period of many decades. The work of Folland et al. (1986a) is an elegant simplification in the description of the historic *SST* data, but several caveats must be kept in mind. Due to changing methods of *SST* measurement, the trend represented by the first EOF is suspect in spite of very careful corrections that have been applied to this data set. The amount of variance accounted for the third EOF is only about 6% (C.K. Folland, pers. commun.). Such a small share of the total variance implies a certain lack of robustness in the statistics which needs further investigation. The amplitude of the third EOF is shown Fig. 3a. A positive amplitude implies that the high latitude oceans of the Northern Hemisphere are warm relative to the Southern Hemisphere. It is negative at the beginning of the century, rising to a maximum positive value in the middle of the century. This behavior is entirely consistent with a slow monotonic temperature rise in the Southern Hemisphere and the low frequency,

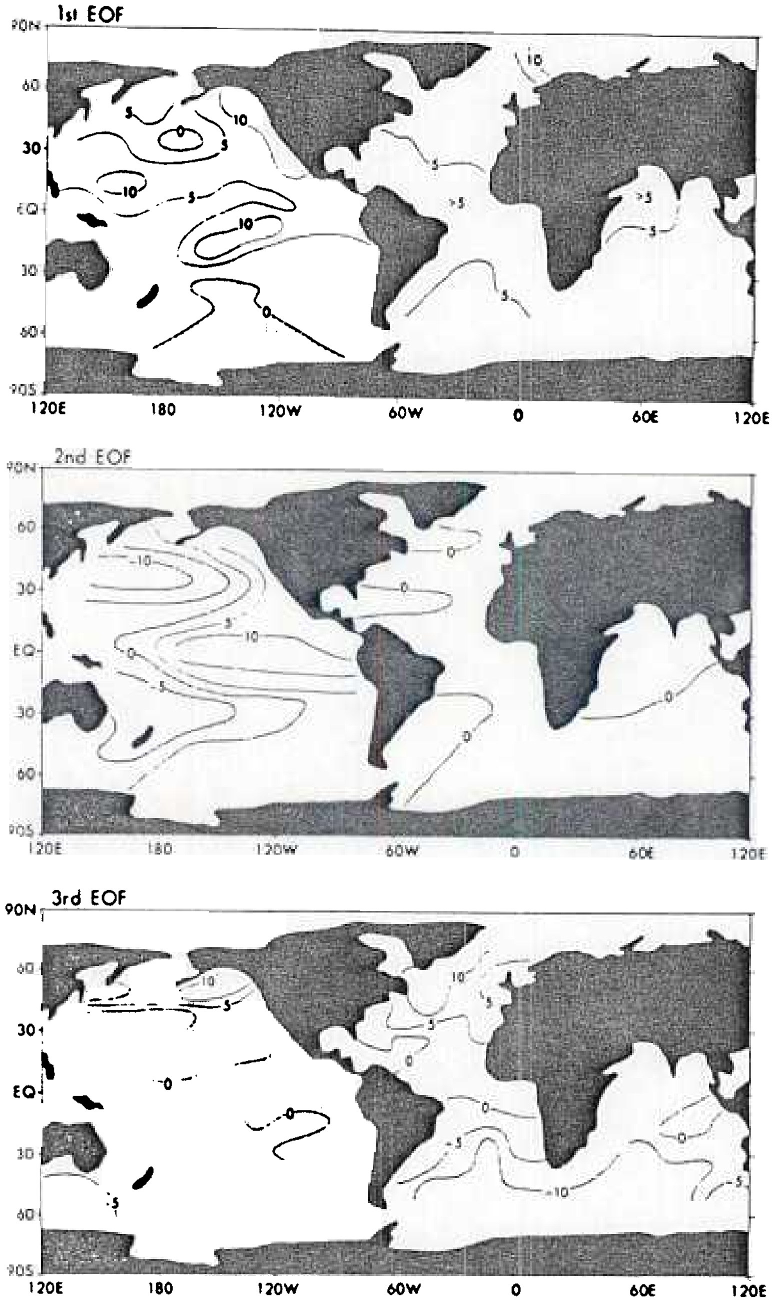

Fig. 2. Empirical orthogonal function patterns of *SST* calculated for the period 1901–1980 from the British Meteorological Office file (Folland et al., 1986a).

polar trapped surface temperature variation over land shown in Fig. 1.

The long term record of rainfall in the Sahel region of Africa (Nicholson, 1985) is plotted in Fig. 3b. As pointed out by Folland et al. (1986a) a positive correlation between African summer monsoon rainfall and the amplitude of the third EOF exists on a time scale of about 50 years. Since the correlation depends on the coherence at a very low frequency, the record is too short to make the relationship statistically significant. Folland et al. (1986b) have carried out numerical experiments with the British Meteorological Office climate model which support a direct causal link

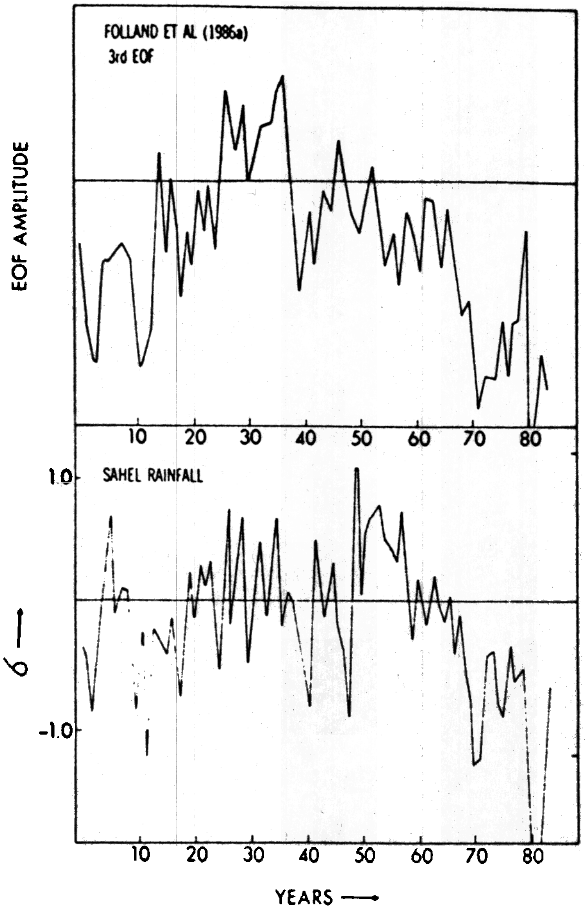

Fig. 3. (a) Amplitude of the third EOF shown in Fig. 2c. Note the correspondence with the variations of Northern Hemisphere surface temperature over land shown in Fig. 1. (b) Variations in rainfall in the Sahel region in units of standard deviations from the normal value. Based on data by Nicholson (1985).

between Atlantic *SST* anomalies of the general type given by the third EOF and Sahel rainfall.

Hasselmann's theory of climate variability

The oceans have more than a thousand times the heat capacity of the atmosphere. Heat exchange between the ocean and atmosphere is sufficiently vigorous to completely recycle the entire heat capacity of the atmosphere in less than a year. Thus the ocean must be involved in any climate variation extending over many years. The question remains, however, whether the ocean plays a purely passive role through the local uptake and release of heat in the ocean's surface mixed layer, or whether the ocean circulation is an active participant in the air–sea interaction.

Hasselmann (1974) has proposed a very simple model, which has now become a classic element of climate theory. We can think of a uniform mixed layer with a random, white noise heating at the surface associated with atmospheric disturbances. The heat capacity of the mixed layer tends to integrate the random heat input with respect to time. Hasselmann (1974) points out that a white noise input will thus be turned into a red noise *SST* spectrum by the heat capacity of the ocean. In this way a complete passive ocean could turn the random high frequency input of the atmosphere into much lower frequencies in *SST*. Recently this idea was illustrated in detailed model experiment carried out by Hansen and Lebedeff (1987). A 100 year run was carried out with the GISS (Goddard Institute of Space Studies) atmospheric general circulation model coupled to a mixed layer of the ocean whose depth varied geographically and seasonally as specified from observations. The heat transport of the ocean was also specified to be a function of season and position. No dynamic feedback by the ocean circulation is allowed by the model. In spite of this constraint, the authors found that the standard deviation of annual mean *SST* was about 0.25–0.50 °C at low latitudes and about 1.0 °C at high latitudes. This level of variability and the very long time-scales associated with it are quite consistent with the time series of surface temperature shown in Fig. 1. In an overall sense the Hansen and Lebedeff (1987) 100 year model run is a rather impressive demonstration of Hasselmann's theory of air–sea interaction. For purposes of discussion we will define this process which involves no feedbacks associated with the ocean circulation as air–sea interaction of the *first kind*.

Bjerknes' hypothesis for climate changes in the North Atlantic

Bjerknes' (1969) seminal paper on the El Niño-Southern Oscillation phenomenon is the foundation for research in tropical air–sea interaction. For reasons, which are not entirely clear, his analysis of air-sea interaction in mid-latitudes summarized in a monograph entitled "Atlantic Air-Sea Interaction" (Bjerknes, 1964) has not received the attention it deserves.

Intra-annual time-scale

The data base examined by Bjerknes consisted of North Atlantic *SST* and surface pressure patterns. As in the case of Folland et al. (1986a) the data were obtained from the files of the British Meteorological Office. Bjerknes, however, only considered the data corresponding to the period from 1890–1938. He found both an intra-annual response of the ocean to changing winds and a longer term response on the order of decades. The relatively short term response is illustrated in Fig. 4 (fig. 14 in his paper). In what Bjerknes refers to as high North Atlantic index years, the North Atlantic anticyclone is stronger and the Icelandic Low is deeper. The stronger surface zonal winds intensify the normal north–south *SST* gradient in the western North Atlantic. The intensification of a gradient results in a subtropical dipole *SST* anomaly: cold to the north of the Gulf Stream and warm to the south. A similar dipole *SST* anomaly is noted by Wallace and Jiang (1987), who point out that it is closely associated with the Western Atlantic see-saw pattern seen in 500 mb charts. They also points out that the dipole response in *SST* generally lags the wind by one or two months. A similar time-lag relationship was reported previously by Ratcliffe and Murray (1970). This type of lag suggests that the ocean is responding to the atmosphere and the *SST* patterns are not a useful predictor on a one or two month time scale. Ordinarily a lag correlation would suggest that the ocean is merely responding to the atmosphere (Davis, 1976). Wallace and Jiang (1987) point out that a lag correlation does not rule out air-sea interaction and cite recent numerical experiments at the British Met Office by Palmer and Sun (1985). These model experiments indicate that monopole *SST* anomalies in the Northwestern Atlantic can excite a train of long waves which are similar to those found by Ratcliffe and Murray (1970). This response is supported by more extensive atmospheric model sensitivity experiments carried out recently by Lau and Nath (1990), which will be discussed in more detail another section.

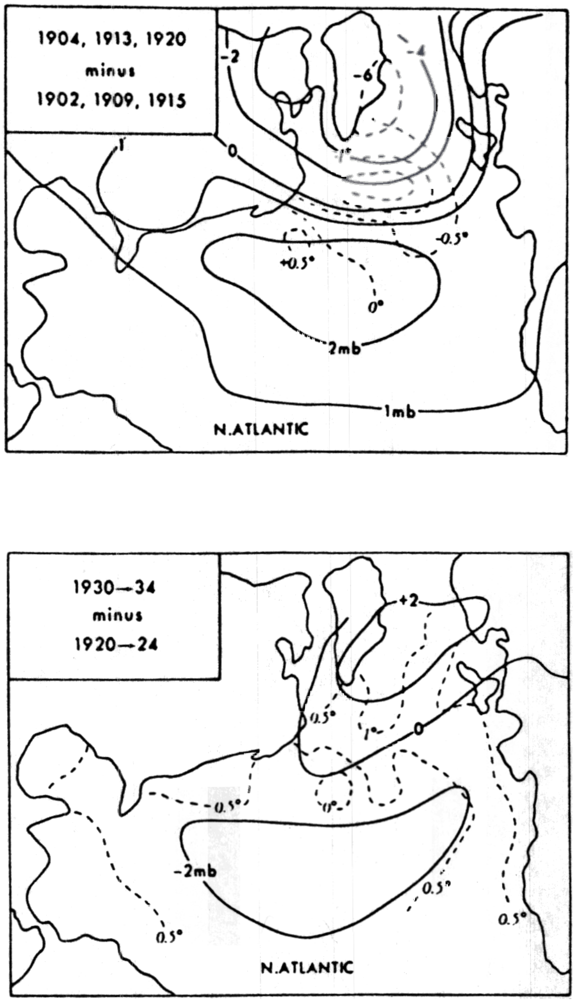

Fig. 4. *SST* and surface pressure anomalies from Bjerknes (1964). (a) The difference between isolated years of high and years of low North Atlantic Index, where the North Atlantic Index is defined as the difference surface pressure between the Azores and Iceland. (b) The difference between an average over a period of persistent high index and a period of persistent low index (1920–1924 minus 1930–1934).

Interdecadal time-scales

In his analysis of the North Atlantic data set Bjerknes noted that the pattern of *SST* response to prolonged periods of increased or decreased zonal winds in the Western Atlantic appeared to be different from the simple dipole *SST* pattern found in connection with higher frequencies. This is illustrated in Fig. 4b (Bjerknes' fig. 7) which shows the difference between the period 1930–1934 and the period 1920–1924. The pressure difference between the Azores and Iceland became much less in the 1930's than it had been in the 1920's. Westerly winds became very much weaker from 1930 to 1934 and there was wide spread warming over the entire northern area of the North Atlantic. The dipole *SST* pattern is absent, with diffuse warming in the Gulf Stream area as well as in the subpolar gyre region. Bjerknes notes that "...the northern waters probably reached their all-time high temperature maximum, since 1880, during the four year sequence of low index years 1939–1942.".

Bjerknes points out that most of the poleward heat transport by the oceans in the Northern Hemisphere is carried out by the North Atlantic. His concept of poleward heat transport in the Atlantic was based on heat balance calculations by Sverdrup (1957). Sverdrup's poleward heat transport show the effects of the Atlantic "conveyor belt", a net northward heat transport across the equator. Sverdrup's estimate of poleward heat transport at 24° N in the Atlantic is within the error bars of recent measurement (Bryden and Hall, 1980; Roemmich, 1980; Roemmich and Wunsch, 1985). Bjerknes hypothesized that during certain decades poleward transport of heat was stronger than in others, with a compensating decrease in poleward heat transport by the atmosphere. Were will designate the type of response in which the ocean plays an active role as air–sea interaction of the *second kind.* Since the North Pacific is not a site of deep water formation, he argued that the North Atlantic was favored as an area where low frequency, active coupling between the atmosphere and the thermohaline circulation could take place. The Bjerknes hypothesis does not rule out the North Pacific as a site for the more passive type of ocean response suggested by the Hasselmann (1974) model.

Evidence for "active" ocean–atmosphere coupling in the North Atlantic on decadal time-scales

It is not at all clear whether there is enough evidence available at the present time to discriminate between ocean–atmosphere coupling of the "first" or "second" kind in the North Atlantic. The answer may require the much more intensive monitoring network which is being planned in connection with WOCE (World Ocean Circulation Experiment Implementation Plan, 1988). However, there are several interesting lines of evidence which are worthy of attention. Palmer and Sun (1985) point out that *SST* anomalies in the Northwest Atlantic have a distinctive character. Unlike the tropical *SST* anomalies associated with El Niño they are strongly modified by winter-time convective overturning. They are often related to a premature, or a delayed onset of winter convection. Once winter convection has taken place, the anomaly tends to disappear as surface temperature is reset to the temperature of the deep water which has a much smaller variability. Data on *SST* anomalies averaged over a 10 degree latitude and 20 degree longitude rectangle centered at 45° N and 50° W from Palmer and Sun (1985) are plotted in Fig. 5. The ordinate is amplitude of the *SST* anomaly. Only anomalies greater than 0.5° C for the months of November through February are included and the month is indicated by the appropriate letter. In some years anomalies persisted through several winter months. It is clear from Fig. 5 that high latitude temperature anomalies will be much "noisier" than tropical *SST* anomalies. Positive and negative anomalies tend to be grouped in certain decades. Negative anomalies are clustered in the 1915–1925 period, while positive *SST* anomalies are present in the 1950–1960 decade.

In the Labrador Sea area the salinity is very important in determining the surface density field. Taylor and Stephens (1980) describe the normal seasonal cycle of surface salinity. The highest values of surface salinity in the Labrador Sea area are attained in winter when deep convection mixes the water column. Minimum salinity is attained at the surface is summer. Taylor and Stephens (1980) demonstrate that surface freshening takes place primarily as a result of a non-local process, consisting of fresh water advection from the vicinity of Greenland rather than a local excess of rainfall over evaporation. An extreme surface freshening event took place in the Labrador Sea in the late

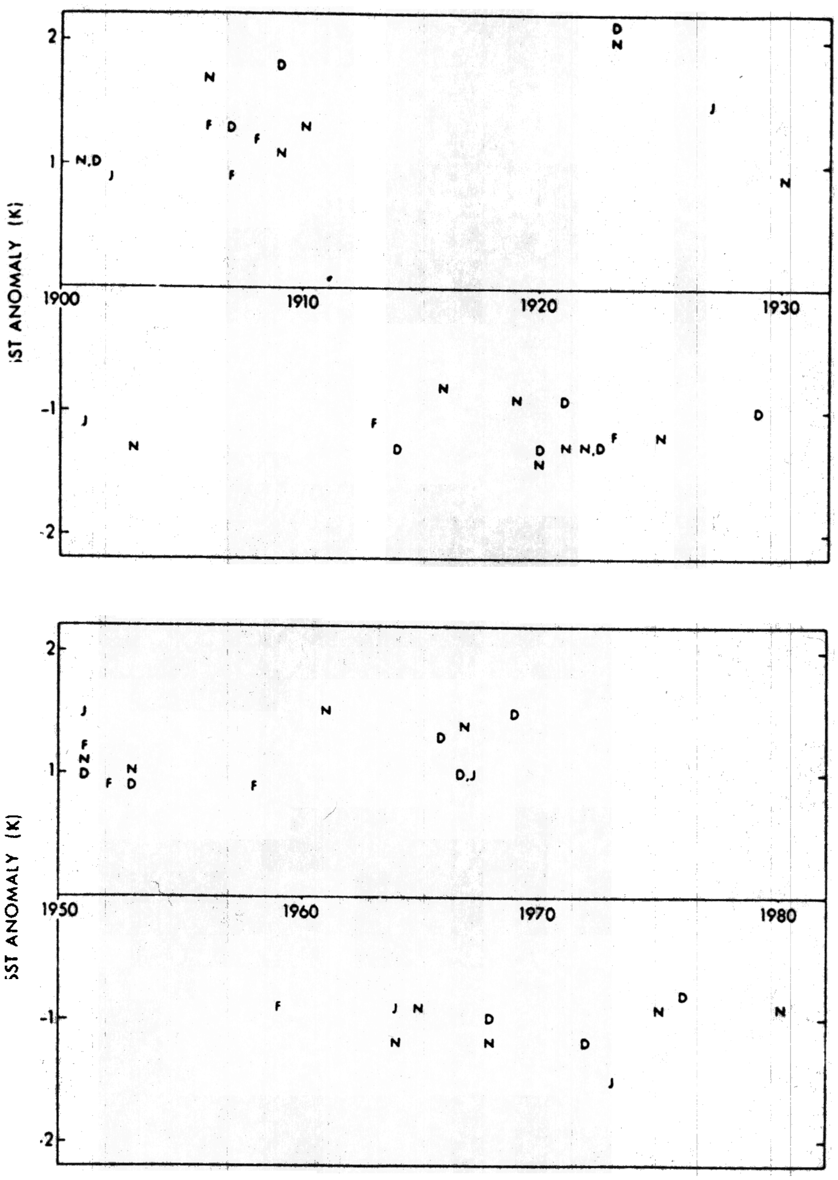

Fig. 5. November through February *SST* anomalies averaged over 40° N–50° N and 40° W–60° W in the Atlantic. Positive anomalies are plotted by month above the axis and negative anomalies below the axis. Only anomalies greater than 0.5° C are included. Several dots in one year indicate persistent *SST* anomalies. The data show how anomalies tend to be switched off by winter overturning which resets the temperature of surface waters to that of the deep water.

1960's. The data from Ocean Weather Ship BRAVO shown in Fig. 6 indicates that there is a period of years in which the surface waters had a low enough salinity to inhibit convection entirely. This period coincided in time with a rapid cooling of the entire Northern Hemisphere relative to the Southern Hemisphere. In later studies, Taylor and Stephens (1980), Dickson et al. (1988) and Lazier (1988) concluded that the very large salinity anomaly, which caused the failure of winter convection in the Labrador Sea, subsequently drifted to the east. According to the analysis of Dickson et al. (1988), the same anomaly could be traced all around the counterclockwise path of the Subpolar gyre of the North Atlantic. It appeared in the Norwegian Current area about a decade after the

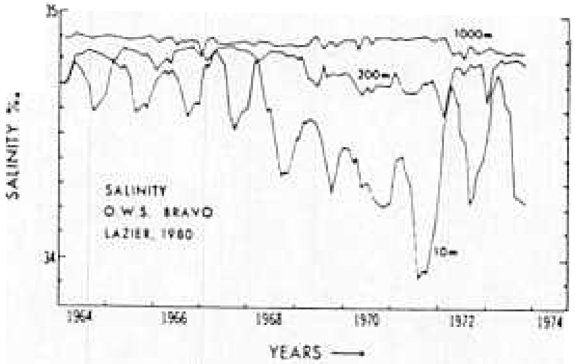

Fig. 6. The salinity record at O.W.S. "Bravo" in the Labrador Sea (Lazier, 1980). Note how winter convection fails for a few years and then resumes in the early 1970's.

event shown in Fig. 6. Dickson et al. (1988) speculate that the anomaly may have originated from an excess discharge of sea ice by the Greenland Current.

In summary, data on the salinity and associated *SST* anomalies in the Northwest Atlantic in recent decades appears to be inconsistent with a purely local mechanism which could be explained on the basis of a simple model of the Hasselmann (1974) type discussed previously. Nonlocal effects, bringing surface waters from other locations, may determine when winter convection takes place, producing large deviations from the normal seasonal cycle.

Modeling air–sea interaction in the North Atlantic

Since only limited data is available for analyzing decadal climate variations, climate models are vital to interpret the scanty historical record. Folland et al. (1986b) have carried out a very interesting investigation of the relation between Atlantic *SST* anomalies and African drought using the British Meteorological Office climate model. They took North Atlantic *SST* patterns corresponding to years of good rainfall in the 1950's and *SST* patterns corresponding to drought conditions in the 1970's and determined an anomaly pattern from the *SST* difference. To determine the atmospheric sensitivity to this anomaly pattern the model was integrated over a period of four months corresponding to spring and summer when most of the rainfall takes place in the Sahel region. The response of the model gave the observed differences in rainfall in approximately the correct locations. The agreement is quite remarkable and supports the idea that Atlantic *SST* anomalies play a large role in Sahel drought cycles as suggested by the empirical data of Folland et al. (1986a) shown in Fig. 2 and Fig. 3. As yet there has not been a concerted effort to check these results at other research centers.

The third EOF shown in Fig. 2 indicates that the *SST* anomalies in the subpolar North Atlantic is particularly important. We can see the time history of *SST* in this region in the form of a Hoevmoeller diagram in Fig. 7 based on COADS (Woodruff et al., 1987), data averaged temperature from 45°N to 55°N is shown as a function of longitude and time. The longitude span is the interval from 60°W to 30°E, the entire width of the North Atlantic at these latitudes. The linear trend over the entire period has been removed at each point and the data has been low-pass filtered. Periods less than a year and a half have been removed using a nonrecursive digital filter with a Lanczos window (Hamming, 1977). Figure 7 gives an overview of the long term climatic changes in the subpolar North Atlantic. Major features are a cold period from 1900 to 1930 followed by a warm period in the 1930's. There was another warm period in the 1950's, terminated by an cooling in the late 1950's and another cooling event in the late 1960's. Bursts of cold or warm water seem to originate in the western North Atlantic. Evidence for eastward propagation is present in some periods, but is not consistent over the whole record. The data is not exactly comparable to the anomalies of Palmer and Sun (1985) shown in Fig. 5, which is just for the winter months and represents an average from 40°N–50°N rather than from 45°N–55°N. The data in Fig. 5 is drawn from the British Meteorological Office data set while Fig. 7 is based on the COADS data set. Nevertheless, in the 1950–1975 period there is general agreement in the 40°W–60°W longitude interval near the western boundary. A warm event in the early 1950's is present in both figures and the return to colder sea surface temperatures in the 1970's is also clearly shown in both. The corre-

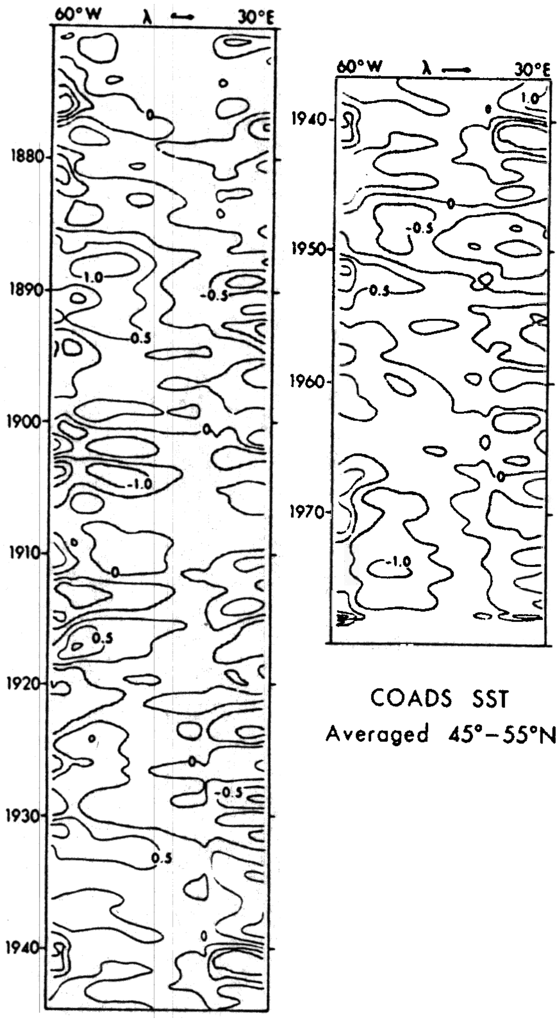

Fig. 7. *SST* averaged over the interval 45°N–55°N as a function of time and longitude in the North Atlantic. From the COADS data set, low-pass filtered and detrended.

spondence is not as good in the earlier period from 1900–1930.

In a recent study Lau and Nath (1990) have used the atmospheric climate model of the Geophysical Fluid Dynamics Laboratory to test atmospheric sensitivity to *SST* patterns. A thirty year integration was carried out with historical *SST* patterns specified for the global ocean from 60°N to 38°S as a function of season. Using EOF and cross-correlation analysis two marine areas are identified as being associated with the strongest atmospheric signals. One of these is the area of the coast of Newfoundland, which was the focus of earlier studies by Ratcliffe and Murray (1970) and Palmer and Sun (1985). The pattern of atmospheric response to the *SST* anomalies in this western North Atlantic area appear to be very similar to that found by Palmer and Sun (1985), but the statistical significance is much greater as the results are based on an extended thirty year integration. Although observations show that at frequencies of one year or less atmospheric disturbances tend to lead *SST* changes, the models show that a positive ocean feedback can take place. Further studies are needed to provide more details on the air–sea interaction and to separate the effects of intra-annual variability and multi-year, low frequency variability.

The 30-year integration of Lau and Nath (1990) confirms both the empirical findings of Wallace and Jiang (1988) and the numerical experiments of Palmer and Sun (1985) for winter-time response to Atlantic *SST* patterns. Unfortunately the summer precipitation climatology of the model over Africa is not enough to allow Lau and Nath (1990) to draw any conclusions concerning the effect of Atlantic *SST* patterns on summer monsoons in the Sahel region. Extension of the Folland et al. (1986b) study of African drought awaits sensitivity studies with higher resolutions models.

Gradually a modest success is being achieved in simulating global climate with coupled models of the ocean and atmosphere. Some very interesting results have been achieved recently by Manabe and Stouffer (1988). The atmospheric component of their model is the same climate model used in the sensitivity run of Lau and Nath (1990). In this particular study seasonal variations of solar isolation are not included, and cloudiness is specified from data rather than predicted. Lower boundary conditions for the atmospheric model over continents are provided by a simplified land surface model, which includes the effects of snow cover and water storage and runoff. Over the ocean lower boundary, conditions on *SST* are provided by an ocean model which is based on finite differences rather than spherical harmonics due to the complicated geometry of the World Ocean. It has 12 layers with the highest resolution in the upper ocean. Temperature and salinity are predicted separately and the local density is calculated from a realistic equation of state for sea water.

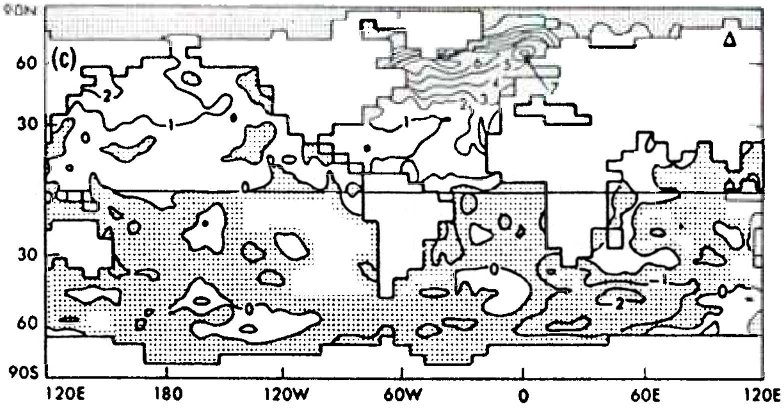

Fig. 8. The difference in sea surface temperature two model equilibrium climates found in a coupled model by Manabe and Stouffer (1988). The difference between the case of an active and inactive thermohaline circulation in the model North Atlantic.

More detailed descriptions of the atmospheric and ocean models are given in Bryan et al. (1975) and Manabe et al. (1975). The method of coupling to obtain climate equilibrium states is described by Manabe and Bryan (1969) and Bryan (1984).

Over ocean areas the ocean model and sea ice model specify the lower boundary conditions for the atmosphere. Upper boundary conditions for the ocean model are fluxes of momentum, heat and moisture calculated in the atmospheric model. To compensate for biases in both the atmospheric and ocean models a fixed adjustment is added to

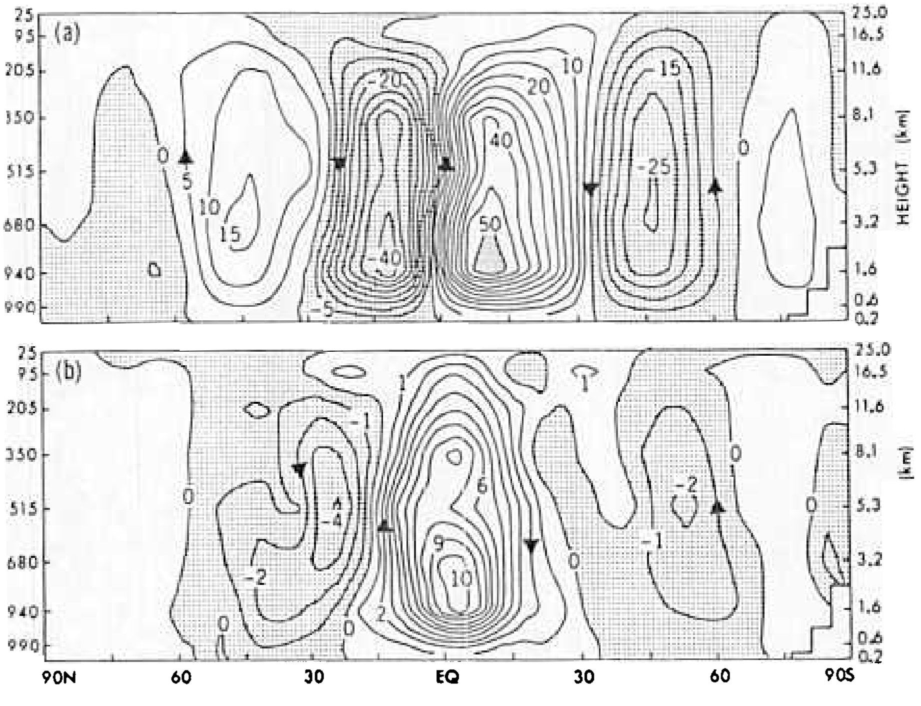

Fig. 9. (a) The total meridional overturning in units of megatons per second in the atmospheric component of the coupled model of Manabe and Stouffer (1988). (b) The difference in overturning between the case with the active North Atlantic thermohaline circulation and the case without.

the moisture flux. Without this adjustment the coupled system tends toward a climate with a very inactive Atlantic thermohaline circulation and cold and fresh water water on the surface of the North Atlantic. Climate drift is nearly eliminated from the initial state by an extended initial integration of the ocean and atmospheric components with nonsynchronous coupling.

Manabe and Stouffer (1988) find that the corrected model has two stable climatic states. One state has an active thermohaline circulation in the North Atlantic and the other state has not. The differences between the two climates are a measure of the very profound effects of the North Atlantic Ocean circulation on global climate and particularly on the climate of the Northern Hemisphere. *SST* differences between the active North Atlantic climate and the inactive case are shown in Fig. 8. The largest differences are naturally in the North Atlantic area, but sizeable differences also exist over the entire polar and subpolar zones of the Northern Hemisphere, including a secondary maximum in the northwestern sector of the Pacific. Comparing with Fig. 2 we see that the pattern of temperature change has some resemblance to the pattern of the third EOF of *SST* of Folland et al. (1986a). Both patterns are generally positive in the North Atlantic and 'Pacific and generally negative in the Southern Hemisphere.

The change in meridional circulation in the atmosphere of the climate model of Manabe and Stouffer (1988) is shown in Fig. 9. The pattern for an active North Atlantic thermohaline circulation is shown in Fig. 9a. This is the case that corresponds more closely to the Earth's present climate. The difference between the active and inactive cases is shown in Fig. 9b. Note the active case favors a slight shift of the rising branch of Hadley cells into the Northern Hemisphere . This favors rainfall in the subtropical zone of the Northern Hemisphere. On the other hand, ascending motion between 40° N–50° N is weakened, while the Ferrel cell in the Southern Hemisphere becomes stronger. The relative low spatial resolution of the atmospheric model does not allow really reliable predictions of geographical patterns of rainfall and many features of the distribution of rainfall and rainfall anomalies differ from those of Folland et al. (1986b). However, the zonal shift of the meridional circulation shown in Fig. 9b is a more robust feature.

Conclusions

How can natural variations of climate be discriminated from anthropogenic greenhouse warming? This is an urgent scientific question. In this review we stress the importance of gaining a better understanding of natural variability through modelling and the study of climate data. The historical record of surface temperature shows that natural low frequency variability can be classified into very broad categories. One category, the El Niño phenomenon, consists of equatorially trapped climate variations which are associated with east–west shifts of deep convection in the Pacific region and global changes in the intensity of the hydrologic cycle. The other broad category has a period of several decades and tends to be amplified toward the pole. Because the earlier record in high latitudes of the Southern Hemisphere is missing, low frequency, polar trapped variability can only be seen in the data for the Northern Hemisphere. Due to its ultra low frequency this category of natural variability will be the most difficult to discriminate from anthropogenically induced climate change. It is obvious that the oceans with their enormous heat capacity will play an important role in decadal climate variations. The question is whether the oceans play an active or passive role in global air–sea interaction on decadal time-scales. A passive role, which we call air–sea interaction of the first kind, can be summarized in a very simple way by a model suggested by Hasselmann (1974). In Hasselmann's model a purely local process is assumed. A uniform ocean mixed layer responds to random heating by the atmosphere in such a way as to produce a red noise *SST* spectrum.

Data on *SST* variations in the North Atlantic indicate that seasonal variations in stratification may play an important role. Anomalies are most pronounced at the beginning of the winter season and appear to be associated with premature or delayed winter overturning. At Ocean Weather

Ship "Bravo" in the Labrador Sea Lazier (1980) noted that changes in surface salinity were large enough to arrest winter overturning for a few years in the late 1960's. Taylor and Stephens (1980), Dickson et al. (1988) and Lazier (1988) found evidence that in the decade following the interruption of water overturning at "Bravo" the surface salinity anomaly drifted eastward across the North Atlantic, appearing in the area of Iceland a decade later. In summary, actual *SST* anomalies in the Northwestern Atlantic appear to be much more complex than any purely local theory can account for. It appears that changes of stratification, associated with advection of salinity anomalies near the surface can greatly amplify *SST* anomalies by disturbing the normal seasonal cycle of winter convection and summertime warming.

Bjerknes' (1969) classical paper on the El Niño–Southern Oscillation won widespread attention because it presented a rather compelling model of large-scale interaction of the ocean and atmosphere in terms of the equatorial Walker Cell. Bjerknes' (1964) paper on Atlantic climate variations may not have received the attention it deserves, because the hypothesis that North Atlantic climate variations are due to large scale interaction was not supported by a convincing physical mechanism. Before Bjerknes (1964) published his study Stommel (1962) had already formulated a theory for multiple equilibrium states of a thermohaline circulation, but it is only recently through the work of Rooth (1985), F. Bryan (1986) and Manabe and Stouffer (1988) that the real implications of Stommel's (1962) paper become better known.

Recent coupled ocean–atmosphere calculations provide a possible model of a global air–sea interaction process which can simulate some aspects of the data, but not the detailed geographic distribution of rainfall. Manabe and Stouffer (1988) find that a coupled ocean–atmosphere model with a corrected water flux to the ocean exhibits two climatic states. One has an active thermohaline circulation in the North Atlantic and the other does not. The active overturning leads to a relative warm subpolar North Atlantic and a shift of the thermal equator into the Northern Hemisphere. Ascending motion in the monsoonal zone of the Northern Hemisphere is intensified and the Ferrel cell of the in temperate zones is weakened. This particular coupled model calculation did not include seasons. However, one way of interpreting Manabe and Stouffer's (1988) result is that a warm Atlantic would displace the thermal equator to the north and intensify the global monsoon, since most of the continental area is in the Northern Hemisphere. A change in the intensity of the monsoon persisting over several years would produce widespread changes in rainfall in the tropics and at higher latitudes.

Lack of data will never allow us to reconstruct the detailed mechanisms of past climatic events. Thus in an absolute sense it will never be possible to confirm or reject the Bjerknes hypothesis that periodic changes in the poleward heat transport of the ocean are responsible for the North Atlantic climate variability over the last century. We conclude, however, that much more can be done through the meticulous construction of new historical data sets and more detailed simulation with models of the ocean and atmosphere, used alone or in a coupled mode. Bjerknes' hypothesis is extremely interesting in its own right, but is should be pointed out that detailed understanding and monitoring of the high latitude ocean areas could lead to advances in diagnosis low-frequency climate trends, whether the air–sea interaction was the first kind, as envisioned in Hasselmann's (1974) simple climate model, or of the second kind as proposed by Bjerknes. Since the amplitude of natural climatic variations is so large, understanding low frequency, large-scale air–sea interaction may be one of the most important elements in distinguishing natural from anthropogenic climate change.

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