SEA SURFACE TEMPERATURE ANOMALIES AND PREDICTABILITY

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Abstract

The atmosphere is to a large extent forced from below through various processes. An essential part of the energy transfer to the atmosphere comes from the oceans. Through transfer of sensible heat and through evaporation and precipitation the oceans influence the atmospheric circulations. The key parameter is the sea surface temperature. Since this parameter changes in time and space it is understandable that anomalies in the sea surface temperature may be important for changes in the circulation.

Section 2 contains some simple considerations leading to estimates of the expected changes in temperature and wind caused by such anomalies. For this purpose we use linear and zonally averaged models.

Section 3 analyses the scientific problem and summarizies some recent results obtained from state-of-the art three-dimensional climate models. These model integrations give estimates of the long-term response of the atmosphere to sea surface temperature anomalies kept constant in time and contains also a description of an attempt to describe these responses in terms of linear, barotropic, forced changes for various basic states.

Section 4 describes the results of some experiments in extended-range predictions with emphasis on the response to sea surface temperarature anomalies. It is found that the anomalies have a marked influence on the model integrations, especially after about 3 weeks. The impact of realistic sea surface temperature distributions is definitely positive, resulting in improved forecasts in the extended range even to the point that individual

forecasts may reach an accuracy which is acceptable in deterministic predictions. In view of these results we investigate the accuracy of 10-day mean forecasts which for realistic temperatures turn out to be superior to those based on climatological temperature fields.
The present situation is discussed in section 5 where it is pointed out that much work needs to be done, and that the present work rests on probably unrealistic assumptions pointing to the use of coupled atmosphere-ocean models as the next step.

1. INTRODUCTION

Sea surface temperature anomalies (SSTA's), by which we mean the deviation of the actual sea surface temperature (SST) from the long term mean value, occur all the time in the oceans. The Climate Analysis Center (CAC) of the National Oceanic and Atmospheric Administration (NOAA) produces routinely maps of SSTA's over the world oceans. The magnitude of SSTA is normally a couple of degrees, but the maximum values may be as large as 5-6^{o}C. Depending on the air temperature in the layer immediately above the sea, the SSTA's provide a potential anomalous heat source which can transfer both sensible and latent heat to the lower layers of the atmosphere through heat transfer and evaporation. The object of this paper is to discuss if these anomalous heat sources and sinks have a noticeable influence on the atmospheric circulation, and, if so, how they may influence predictions on the various time-scales from short- and medium-range forecasts to long-range predictions. In short, will sea-surface temperature anomalies alter the practical limit of predictability?
It has been realized for a long time that the oceans, through transfer of sensible heat and by evaporation, provide considerable energy amounts to the atmosphere in certain geographical regions. Jacobs (1951) was one of the first

researchers to calculate the transfer of sensible heat and the evaporation on a climatological basis for the various seasons. By these methods he could present maps of the total energy gain by the atmosphere from the oceans. These amounts vary greatly from region to region, but maxima occur, especially in the winter season, over the western parts of the Atlantic and Pacific oceans strongly connected with the western boundary currents in both oceans. The total energy amounts transferred to the atmosphere this way are quite large (can in the maxima be as large as 100 Wm^{-2}) and are additions to the internal energy of the atmosphere. It was considerations of this kind which lead scientists, notably Smagorinsky (1953), to investigate the steady-state response to the heating of the atmosphere in the ultra-long wave regime. Many scientists have elaborated on this theme, especially Döös (1962).

At the moment the topic is, however, not the response to the total heating, but rather the heating created by anomalous changes in the lower boundary, i.e. changes in the sea-surface temperature. One of the treatments of this problem goes back to the pioneering work by Bjerknes (1966) who first made an analysis of the observed SSTA's in the Pacific, especially the tropical part of this ocean. Large positive SSTA's are observed here as a part of the El Nino phenomenon. His arguments, supported by his data and analysis, are that the increased sea surface temperatures will enhance the transfer of energy into the atmosphere. An immediate effect of this should be an increase in the precipitation. This argument is supported by the data. A more speculative part of the arguments is that the enhanced heating should increase the intensity of the Hadley circulation. If so, this would mean that more westerly momentum is going into the subtropical jetstream in the upper poleward part of the Hadley cell. The result can be an increase in the speed of the jetstream or a change in the position to a more southerly location. The changes in the subtropical jetstream can in turn have an

effect on the circulation pattern in the middle and higher latitudes.
A first simple analysis of the heat induced tropical circulations has been made by Gill (1980) who provided some mathematical solutions which have helped in the interpretation of many results.
Many of the investigations of the impact of SSTA's have been aimed at climatic problems and problems connected with the interannual variation. Another group of experiments is concerned with the asymptotic response of the atmosphere to a given SSTA. Examples of research papers along these lines are by Rowntree (1972, 1976).
The published SSTA by CAC for various time periods show clearly that SSTA's of several degrees exist outside the tropical region in the more northerly latitudes. Mid-latitude SSTA and their influence on the general circulation have been investigated by numerical experiments by Huang (1975) and Roads (1980), while Ratcliffe and Murray (1970) have studied the associations between SSTA in the North Atlantic and pressure changes over Europe from a more empirical point of view. Egger (1977), on the other hand, calculated the response to SSTA's using linear theory and a steady-state, quasi-geostrophic, two-level model.
A summary of these investigations has been given by the author (1982).

2. SOME SIMPLE CONSIDERATIONS

The amount of sensible heat may be computed from the formula

$$Q_s = 13.6\ (T_w - T_a) \qquad (2.1)$$

where Q_s is expressed in the unit Wm^{-2} when T is in absolute degrees. T_w is the water temperature and T_a the air temperature, see Wiin-Nielsen (1982). Converted into heating of a layer, say 10cb, we find:

$$H = 1.3 \times 10^{-2} (T_w - T_a) \qquad (2.2)$$

where H is measured in kj $t^{-1}s^{-1}$. The literature uses occasionally the unit deg day^{-1}. This is really a measure of

$$\frac{1}{c_p} H = 1.15 \times (T_w - T_a) \qquad (2.3)$$

converted to the time unit of one day.

With L the specific heat of evaporation we find for the corresponding transfer of latent heat is L Q_e when

$$Q_e = \rho_a C_d U (q_o - q_a) \qquad (2.4)$$

where ρ_a is the density of air (say 1.23×10^{-3} t m^{-3}), $C_d = 1.1 \times 10^{-3}$ the drag coefficient, U the wind speed, q_o the specific humidity of saturation and q_a the specific humidity of the air.

In the investigation by Wiin-Nielsen and Brown (1962) it was found, in an individual case over the region of the Gulf Stream, to be exactly 6. In January, 1959 at 00Z, the sum of the two processes amounted to about 0.7 kj $t^{-1}s^{-1}$. A more typical average value might be 0.1 kj t^{-1} s^{-1} which according to (2.2) corresponds to a temperature difference of 7-8 degrees.

The anomalies observed in the sea surface temperatures are typically of the order of 1-2^{o}K indicating that the additional heating or cooling is of the order of magnitude 0.01-0.02 kj $t^{-1}s^{-1}$. We are therefore talking about relatively small perturbations. Furthermore, it is of course well known that it is not the absolute amounts of internal energy, added to the atmosphere, which matters, but rather the contributions to the available potential energy. Because of the smallness of the SSTA's we can get a first idea of the effects of SSTA's by using a linearized model. This was done by Wiin-Nielsen (1975) using a quasi-geostrophic model without advection, but including a heating effect. Some additional examples are shown in the investigation by Wiin-Nielsen (1982). We may take a case in which the lowest 10 cb-layer is heated by 0.1 kj $t^{-1}s^{-1}$, while there is no heating in the remaining part of the atmosphere. The heating creates a vertical velocity profile with a maximum of 2-3 mm s^{-1} at about 90cb. The temperature changes are large in the lowest layer, but this heating effect is strongly compensated by the cooling effect created by the upward vertical velocity and the temperature actually decreases above 90cb. The geopotential decreases in the lower layers, but increases higher up. If we next suppose that the heating extends to 80cb we find, of course, larger values for the vertical velocity, the temperature and geopotential tendencies. We may thus conclude that to obtain a significant response to the heating it is necessary that the heating penetrates to high altitudes. This may be one of the reasons that the tropical heat sources seem to be more effective than mid-latitude heat sources. The convection in the low-latitude atmosphere will, because of the latent thermodynamic instability, carry the heating to much higher altitudes than the same amount of heating from below in the middle latitudes. These considerations agree also with the fact that the cyclogenetic effect is seen to be enhanced in the regions of strongest heating from below, i.e. over the western boundary currents in the oceans.

To continue with simple considerations we may also consider the effects of a heating anomaly in a zonally-averaged model. Such an analysis will also be in the spirit of Bjerknes (1966) original speculations. For this purpose we shall adopt the two-level, quasi-geostrophic model including diabatic heating and friction. The zonally-averaged equations for this model are

$$\frac{\partial \bar{\xi}_1}{\partial t} + \frac{\partial \overline{(\xi_1 v_1)}}{\partial y} = \gamma q^2 (\bar{\psi}_T - \bar{\psi}_E) - A \bar{\zeta}_T \qquad (2.5)$$

$$\frac{\partial \bar{\xi}_3}{\partial t} + \frac{\partial (\overline{\xi_3 v_3})}{\partial y} = - \gamma q^2 (\bar{\psi}_T - \bar{\psi}_E) + A \bar{\zeta}_T - 2\varepsilon \bar{\zeta}_4$$

in which we have used the following notations:

$\bar{\xi}_1 = \bar{\zeta}_1 - q^2 \bar{\psi}_T + f$ (potential vorticity)

$\bar{\xi}_3 = \bar{\zeta}_3 + q^2 \bar{\psi}_T + f$ (" ")

$\bar{\psi}_* = \frac{1}{2} (\bar{\psi}_1 + \bar{\psi}_3)$ (stream functions)

$\bar{\psi}_T = \frac{1}{2} (\bar{\psi}_1 - \bar{\psi}_3)$ (" ")

$\bar{\zeta}_* = \frac{1}{2} (\bar{\zeta}_1 + \bar{\zeta}_3)$ (vorticities)

$\bar{\zeta}_T = \frac{1}{2} (\bar{\zeta}_1 - \bar{\zeta}_3)$ (")

$\bar{\zeta}_4 = \bar{\zeta}_* - 2 \bar{\zeta}_T$ (")

in which ψ is a stream function, ζ a vorticity and ξ a potential vorticity, while the subscripts 0, 1, 2, 3 and 4 refer to the levels 0, 25, 50, 75 and 100 cb. The notation q^2 is

$$q^2 = \frac{2 f_0^2}{\sigma P^2} \tag{2.6}$$

in which f is the Coriolis parameter, f_0 a constant value of f, $\sigma = - (\alpha/\theta) \partial\theta/\partial p$ a measure of static stability, and P = 50 cb.

The heating per unit mass and unit time is

$$H = -C_p \gamma (\bar{T} - \bar{T}_E) \tag{2.7}$$

in which C_p is the specific heat at constant pressure, γ a constant, T the temperature and T_E an externally given temperature field. Finally, A measures the internal friction and ε the friction in the planetary boundary layer. The overbar signifies a zonal average.

The system (2.5) can be closed if we specify the transports of potential vorticity in terms of the zonally averaged quantities. Since in this section simplicity is emphasized we shall make the simple assumption

$$\overline{(\xi v)} = -K \frac{\partial \bar{\xi}}{\partial y} \tag{2.8}$$

in which K is a constant exchange coefficient. In this way we get two coefficients K_1 and K_3. We use these to define

$$K_* = (K_3 + K_1)/2; \quad K_1 = (K_3 - K_1)/2 \tag{2.9}$$

Using (2.8) and (2.9) in (2.5) we obtain by addition and subtraction the following equations:

$$\begin{aligned} \frac{\partial \bar{\xi}_*}{\partial t} &= K_* \frac{\partial^2 \bar{\xi}_*}{\partial y^2} - K_T \frac{\partial^2 \bar{\xi}_T}{\partial y^2} + q^2 K_T \bar{\xi}_T - \varepsilon (\bar{\xi}_* - 2\bar{\xi}_T) \\ \frac{\partial (\bar{\xi}_T - q^2 \bar{\psi}_T)}{\partial t} &= -K_T \frac{\partial^2 \bar{\xi}_*}{\partial y^2} + K_* \frac{\partial^2 \bar{\xi}_T}{\partial y^2} - (q^2 K_T + A)\bar{\xi}_T + \gamma q^2 (\bar{\psi}_T - \bar{\psi}_E) + \varepsilon(\bar{\xi}_* - 2\bar{\xi}_T) \end{aligned} \tag{2.10}$$

In the present investigation we shall solve (2.10) by using a channel of width W with walls to the south and the north. The boundary conditions will be

$$\frac{\partial \bar{\psi}_i}{\partial y} = \frac{\partial \bar{\xi}_i}{\partial y} = 0; \; y = 0, W; \; i = 1, 3 \tag{2.11}$$

Recently White and Green (1984)pointed out that the parameterization (2.8) does not automatically lead to conservation of momentum, and they discussed other formulations which do. One may keep this remark in mind when interpreting the results.

The solutions to (2.10) using the boundary conditions (2.11) are obtained by a series expansion of the form

$$\bar{\psi} = \sum_m \Psi(m) \sqrt{2} \cos\left(m\pi \frac{y}{W}\right) \tag{2.12}$$

The details of the solution are given by Wiin-Nielsen (1986) and will not be given here, but we shall look at some examples.These were computed with the following values:

$\gamma = 1.4 \times 10^{-6}\ s^{-1}$

$q^2 = 2.25 \times 10^{-12}\ m^{-2}$

$W = 10^7\ m$

$\varepsilon = 3 \times 10^{-6}\ s^{-1}$

$A = 7 \times 10^{-7}\ s^{-1}$

$K_* = 1.4 \times 10^6\ m^2\ s^{-1}$

$K_T = 0.6 \times 10^6\ m^2\ s^{-1}$

We start with an example given in Fig. 1. The curve T_E shows a temperature disturbance with maximum value of 1° K located in the middle of the channel. The response in the model temperature T_2 is shown in the same figure. The external heating produces a temperature increase in the central part of the channel, but due to the parameterization of the eddy-transport of potential

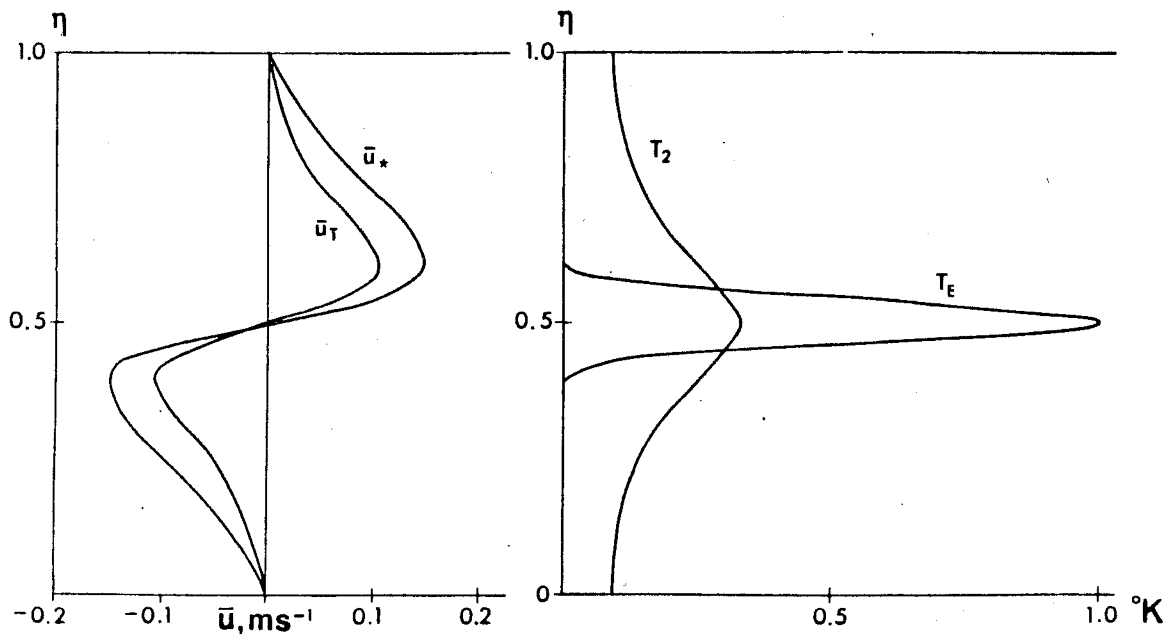


Fig.1 $T_E = T_E(\ )$ is the forcing temperature, $\overline{T}_2$ the response temperature, while $\bar{u}_*$ and $\bar{u}_T$ are the resulting wind distributions.

vorticity the temperature increase is spread out across the channel. It is thus understandable that the change in the thermal wind will be positive in the northern half of the channel and negative in the southern half. A corresponding change is found in the vertical mean wind $\overline{u_*}$ as shown in the left part of the figure. While the arguments given above show the qualitative part of the results, we need to do the actual calculations to obtain the correct quantitative results. We note in particular that the change in the zonal wind is a small fraction of $1ms^{-1}$ when the amplitude is 1^oK

in T_E. In this regard we should remember that T_E is different from zero in a rather small region.

Fig. 2 shows a different case.

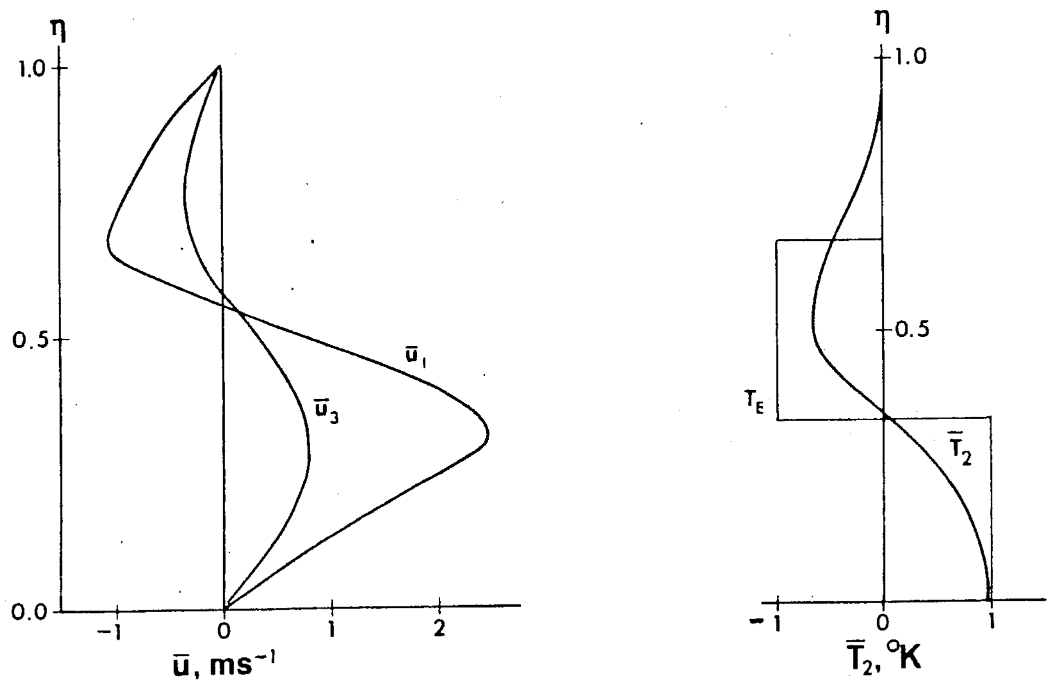


Fig.2 A case with heating close to the southern boundary adjacent to cooling in the middle of the channel. $\bar{T}_2$, $\bar{u}_1$ and $\bar{u}_3$ are the resulting temperature and wind distributions.

In this case the anomaly covers two thirds of the distance from south to north. In addition, there is a strong contrast between a positive anomaly next to the southern wall and the negative anomaly in the center of the channel. Converted to the earth this would correspond to a positive anomaly in the low latitude with a negative anomaly in middle latitudes. This distribution, although extreme, is not unlike an El Nino situation. The resulting temperature distribution is shown in the right hand part of Fig. 2. Due to the boundary condition we find $\bar{T}_2(0) = \bar{T}_E(0) = 1$. The zonal winds $\bar{u}_1$ and $\bar{u}_3$ are shown in the left part of Fig. 2. $\bar{u}_1$ is now more than $2 m s^{-1}$, a much larger value than in the first case. We find therefore, in agreement with Bjerknes' reasoning, an increase or a southward displacement of the sub-tropical jet.

We have so far been concerned with the steady state problem and have shown steady state solutions. It is, however, also of interest to investigate how long it takes for the time-dependent solution to come close to the steady state solution. To solve this problem we go back to the time-dependent equations. Exactly the same assumptions are made, and it is thus seen that we obtain a complete set of linear ordinary differential equations for each value of m in the expansion in terms of the orthogonal functions. Such equations are easy to solve. The detailed solutions are shown by the author (1986).

Fig. 3 shows the difference between the steady state solution and the actual solution as a function of time for the streamfunctions of the vertical mean flow, denoted by an asterisk, and the thermal flow for m = 2. Both

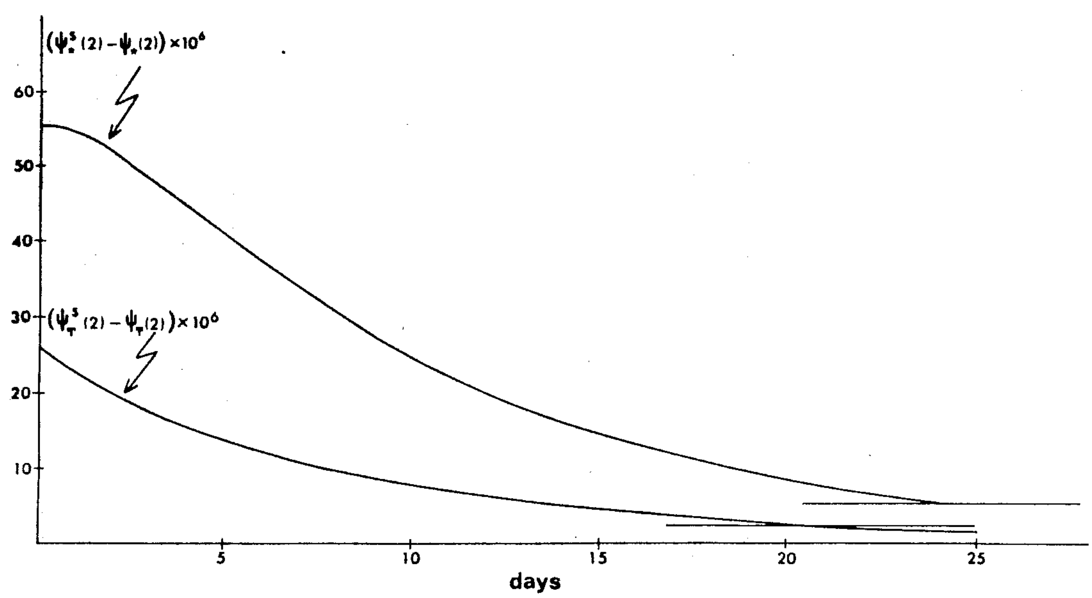


Fig.3 Time dependent solutions of the difference between the steady state and the actual state for wave number two showing that the steady state is reached after approximately 3 weeks.

quantities will naturally approach zero at very large times. The two horizontal lines in Fig. 3 show the 10% level of the initial difference. These levels are reached for the vertical mean flow after 24 days, while the corresponding figure for the thermal flow is 21 days. These results will later be compared with results obtained by non-linear integrations using medium-range prediction models.

3. THE SCIENTIFIC PROBLEM

Based on the simple considerations given in section 2 one cannot exclude the SSTA's having an influence on the medium- and extended-range predictions. In spite of research carried out over at least the last two decades there is, however, no clear understanding of the ways in which SSTA's influence a forecast, and there is only a general understanding that the time-scale of the atmospheric response to SSTA may be of the order of 3-4 weeks.

Under the general expectation that ECMWF staff members will inform the workshop of the most recent work carried out at the Centre, it may be pertinent to describe briefly some work on the problem conducted in the U.S.A. Some general circulation experiments have been carried out with the Community Climate Model (CCM) of the National Center for Atmospheric Research (NCAR). These experiments concentrated on the El Nino phenomenon and its impact on the global circulation and are therefore climate oriented experiments rather than prediction experiments.

Blackmon et. al.(1983) and Geisler et. al. (1985) have reported on experiments carried out under perpetual January conditions. The procedure was to add a SSTA to the normal SST distribution in the CCM and to make two long integrations, one with the SSTA and one without it. The latter may be called the control experiment. The difference between the long-term means for the two experiments may be called the impact of the SSTA. It is reproduced

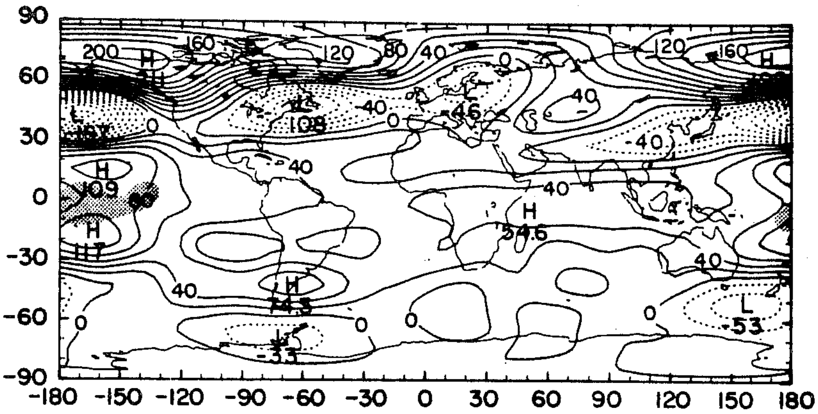


Fig.4 Anomaly in the 1200-day average 200 mb height field of the CCM caused by introducing an El Nino - Southern Oscillation composite SSTA. From Blackmon et.al. (1983). Location of the rainfall anomaly is stippled. Contern internal 20m.

as Fig. 4 of this paper and is a 1200-day average of the difference in the 200 mb height field as produced by the CCM. It should be noted that the observed SSTA was increased by a factor of 2 in these experiments, and that the experiments were carried using a resolution of R 15 in the spectral domain. Fig.4 shows that the impact of the El Nino SSTA according to these calculations is to create a series of lows in the 200 mb height field in the latitude belt of 30 to 60°N with positive height anomalies to the south, i.e. in the region 0-30°N. The main result is therefore an increase in the subtropical jetstream. There are also positive anomalies in the high latitudes which could mean a decrease in the polar front jet.

As a first result we may thus say that the experiments are by and large in agreement with the results obtained with the zonally-averaged model in section 2. On the other hand, the desire is naturally to go further and to investigate, if it would be possible to reproduce the major pattern in these long-term integrations in a different way and thereby increase our understanding of the mechanisms involved.

Once again one should turn to linear analysis, but now carried out in such a way that a complicated steady state changing in both longitude and latitude can be handled. This has been attempted by Branstator (1985 a, b) who goes back to the steady-state barotropic model with forcing. The problem is this: Assuming a steady state as the basic state we want to calculate by linear theory the changes caused by a given SSTA (such as the El Nino). Such calculations have been carried out earlier by Hoskins et.al. (1977), Simmons (1982), Simmons et.al (1983), and by Branstator (1983) himself to explore the horizontal energy propagation. It is also indicated by such investigations that the final result depends strongly on the nature of the assumed basic state.

Mathematically the problem is as follows. Let an overbar denote the basis steady state and a prime the perturbation on it. The barotropic perturbation equation is then

$$\frac{1}{a^2} J(\bar{\psi}, \zeta') + \frac{1}{a^2} J(\psi', f+\bar{\zeta}) + e\zeta' - \varepsilon \nabla^2 \zeta' = R' \qquad (3.1)$$

in which the Jacobian is

$$J(\alpha, \beta) = \frac{\partial \alpha}{\partial \lambda} \frac{\partial \beta}{\partial \mu} - \frac{\partial \alpha}{\partial \mu} \frac{\partial \beta}{\partial \lambda}\ ; \ \mu = \sin \varphi \qquad (3.2)$$

e is the term measuring the influence of the planetary boundary layer ($e = 1.65 \times 10^{-6} s^{-1}$), while $\varepsilon \nabla^2 \zeta'$ is a diffusion term which in (3.2) and in the CCM has a non-zero value only when the total wave number is larger than 15 (in which case $\varepsilon = 2.5 \times 10^5\ m^2 s^{-1}$).

We shall not here go into the details of solving the linear equation (3.1), but the method is to develop both the forcing and the desired solution in terms of spherical harmonic functions.

The basic steady state should be the long term average of the control experiment, i.e. a 1200-day average of the integration without the SSTA. It is shown in Fig. 5 together with the "observed" climatological map. It is,

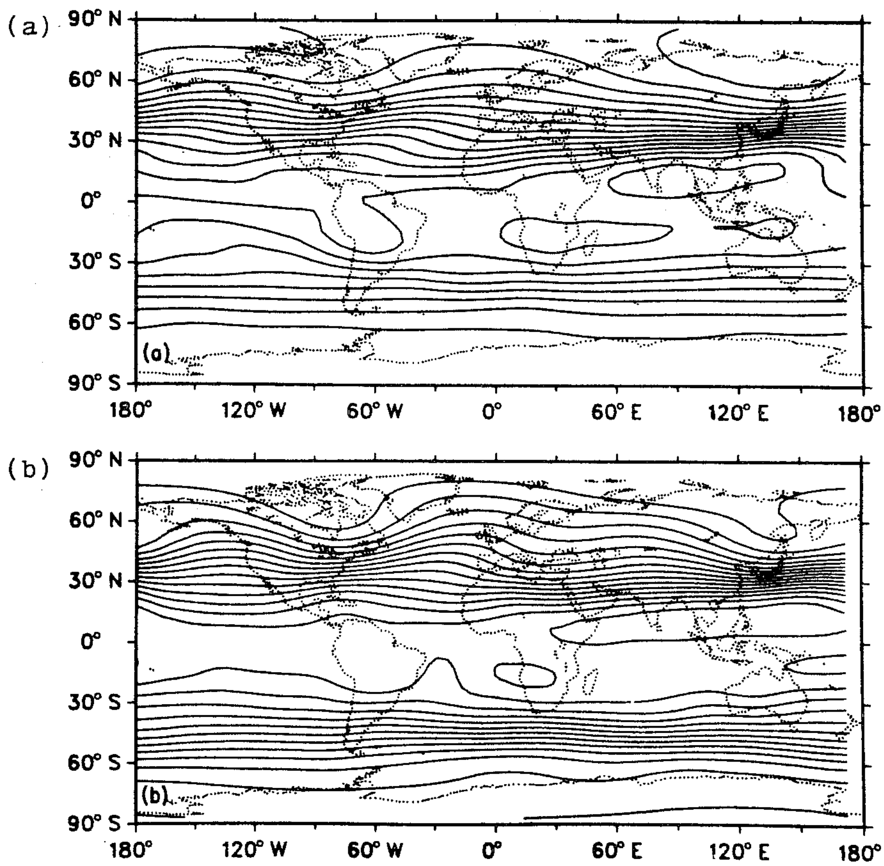


Fig. 5 (a) 1200-day mean CCM 300 mb streamfunction from a control simulation.
(b) Climatological January 300 mb streamfunction. Control internal is $1 \times 10^7\ m^2s^{-1}$.
From Branstator (1985).

however, interesting to see how much and in which way the results depend on the steady state. Branstator (loc.cit.) defined three steady states:

(1) Solid body rotation, i.e. only ψ_1^0 is maintained

(2) Zonal mean field, i.e. only ψ_n^0 is maintained

(3) Full field.

The calculated steady responses are shown in Fig. 6 using a forcing which is defined below. Let ρ^2 be defined by the expression

$$\rho^2 = (\lambda - \lambda_0)^2 / \delta_\lambda^2 + (\varphi - \varphi_0)^2 / \delta_\varphi^2 \qquad (3.3)$$

in which $\delta_\lambda = 30^o$, $\delta_\varphi = 25^o$. The definition of R is therefore:

$$R = \begin{cases} 0 \; ; & \rho^2 \geq 1 \\ -f_0 D_0 (1-\rho) \; ; & \rho^2 < 1 \end{cases} \tag{3.4}$$

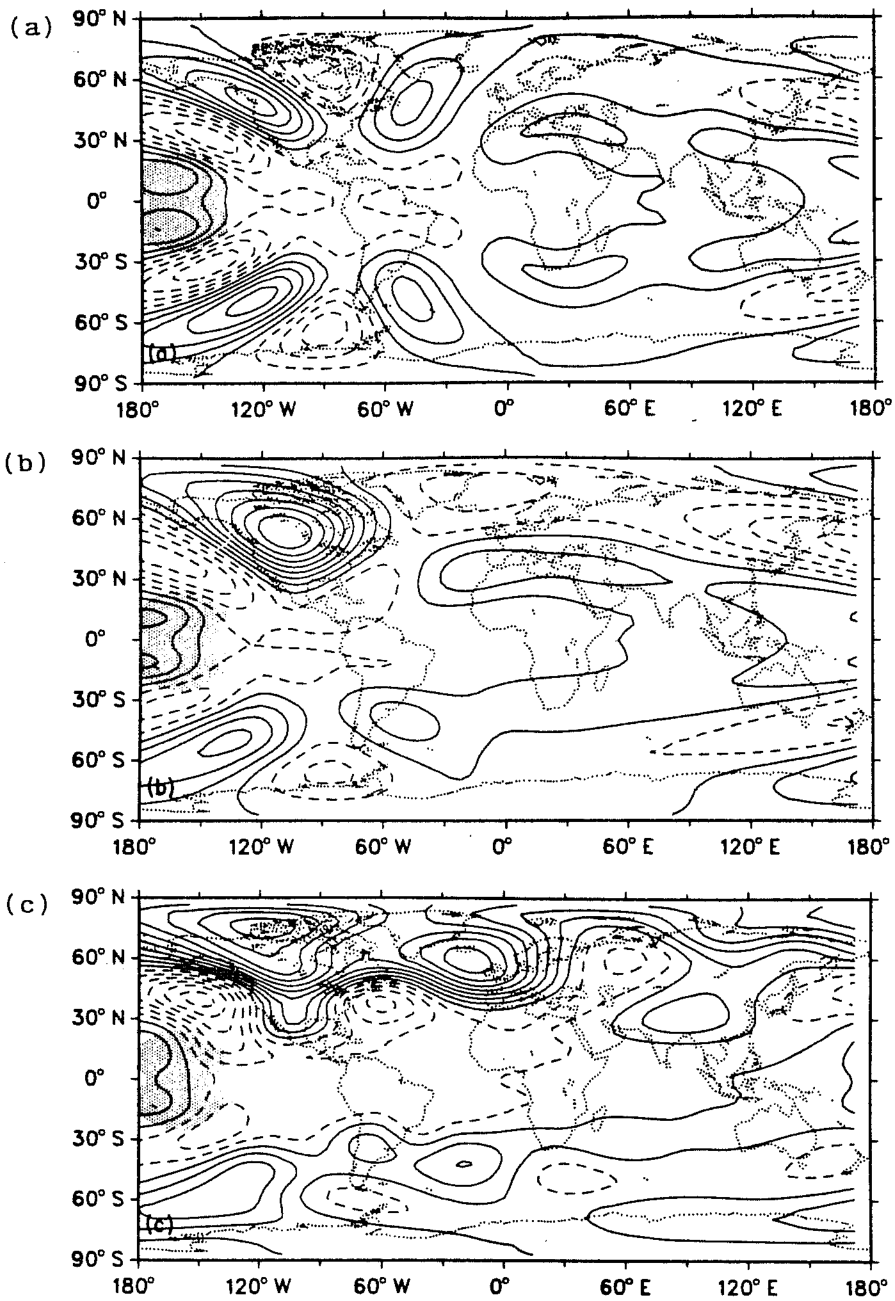


Fig.6 Steady perturbation height solutions of the linear barotropic vorticity equation forced by a vorticity source in the stippled region
(a) solid body rotation in Fig. 5 (a)
(b) zonal mean component in Fig. 5 (a)
(c) total field in Fig. 5 (a)

According to (3.4), R is non-zero in an elliptical region and changes linearly to the elliptical boundary. The responses in Fig. 6 were computed with

$\lambda_0 = 200^o$, $\varphi_0 = 0$ and $D_o = 3 \times 10^{-6}\,s^{-1}$.

Comparing Fig. 6a with Fig. 4 it is seen immediately that they are very different because the distribution in Fig. 6a is symmetrical around the equator and in Fig. 4 it is not; also Fig. 6a fails to reproduce the extensive negative anomaly across the Atlantic and stretching up to Scandinavia. Fig. 6b is somewhat better because asymmetry now appears, but the extensive negative anomaly mentioned above is still not present. Fig. 6c is a definite improvement over the other two figures because the negative anomaly is found over the Atlantic with another negative anomaly in Europe.

We may conclude that the study described above indicates that a part of the response to a given SSTA in the El Nino case can be explained in terms of a linear, barotropic analysis in which the response is calculated using the most realistic steady state. This information is useful in understanding the mechanisms behind the response to SSTA's. It goes almost without saying that the impact necessarily depends on the position of the SSTA. This is immediately understandable because the effect of an SSTA should depend strongly on the structure of the atmosphere above, i.e. its wind regime, stability conditions and moisture distribution. In spite of this reasoning Geisler et.al. (1985) seem to have arrived at the result that the response is relatively independent of the location of the SSTA.

In view of the relative success of the linear analysis described above, it would be interesting in the future to reinvestigate the response to SSTA's located in higher latitudes although some of this work was done by Egger (1977) using a two-level, quasi-geostrophic model on the beta-plane.

4. SSTA's AND EXTENDED-RANGE PREDICTIONS.

It was determined quite a long time ago that it was important to include the source of sensible and latent heat from the oceans to the atmosphere in short- and medium-range predictions. In the early years of ECMWF's experimental and operational efforts it became clear that climatological fields of sea surface temperatures were inadequate for purposes of medium-range predictions, and arrangements were made to secure up-to-date information on the distributions of this parameter as an input to the operational model. On the other hand, the initial SST's are kept constant during the whole integration of the operational model up to 10 days. Under these assumptions, and if they are valid, it would appear that any anomaly (i.e. any significant deviation from the climatological average) is automatically incorporated in the operational forecast. Any significant impact of SSTA's should thus appear in the forecast. It is of course still of great interest to investigate the impact of these anomalies on extended-range forecasts.

Based on the material covered in the previous sections we can conclude that

* the daily change in the meteorological parameters is relatively small as a direct effect of SSTA's
* the impact of a given SSTA depends on the particular atmospheric circulation which exists at the time
* the very long term impact is significant as demonstrated by climate experiments

On the other hand, it is not known with great certainty how fast the SSTA's will influence a forecast significantly, and in which way the prediction will be changed.

To explore this situation further it is necessary to conduct a large number of prediction experiments. Definitive answers cannot yet be given, but there is a great interest in various research quarters. The experiments conducted so far have, as far as I know, been made with SSTA's kept constant during the forecast. While this assumption may be justified during a medium-range forecast to 10 days, it is much more questionable for extended-range forecasts to 30 or 60 days because SSTA's may change significantly during such a time period.

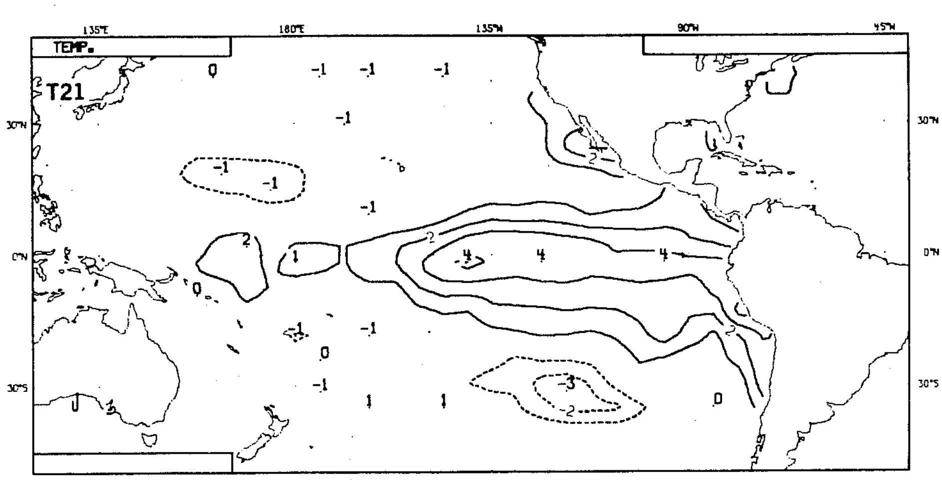
The description in the following will be restricted to a single case which has been analysed by Cubasch and Wiin-Nielsen (1986); it is an El Nino experiment. Fig. 7 shows the representation of the observed SSTA using spectral components up to T 21, T 42 and T 63. It is seen that T 63 is necessary to represent the maximum correctly.

The experiment was conducted as follows:

1. Integrations of the operational ECMWF model up to 60 days were made using a SST equal to its climatological average, marked CLIM in the figures.

2. Integrations were made to 60 days with SST equal to its climatological value <u>plus</u> the SSTA, marked POS in the figures.

3. Integrations were made also with SST equal to its climatological value <u>minus</u> the SSTA, marked NEG in the figures.



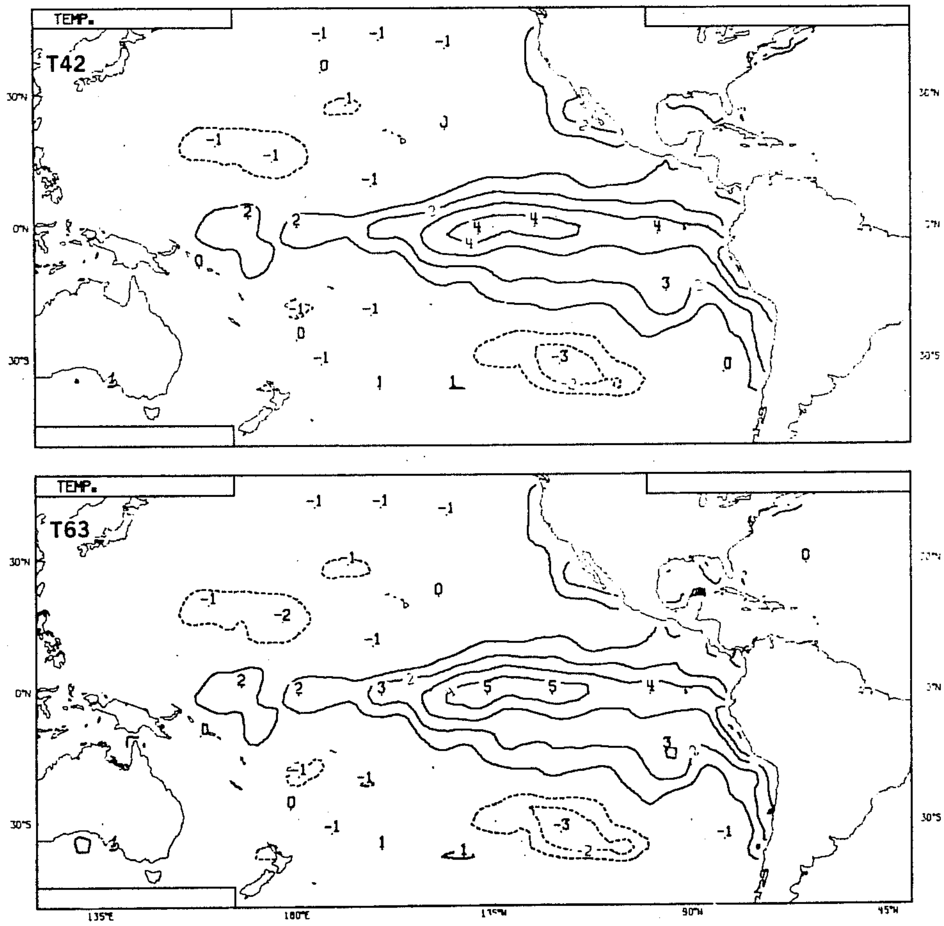


Fig.7 The observed El Nino anomaly as seen by the different resolutions, contour internal 1°C.

Note that we used the mature phase of the 1982/83 El Nino event with operationally assimilated data.Fig. 8 shows the predicted anomaly, averaged for the period 11-40 day, at 500 mb. The predicted results are for the three resolutions T 21, T 42 and T 63, while the last map is the observed anomaly. As we have seen in the climate experiment the anomalies are of rather large magnitude amounting in the observed map to about 280m in the Gulf of Alaska. Comparing the three resolutions it is evident that the T 21 integration does not result in magnitudes which compare favorably with the observed

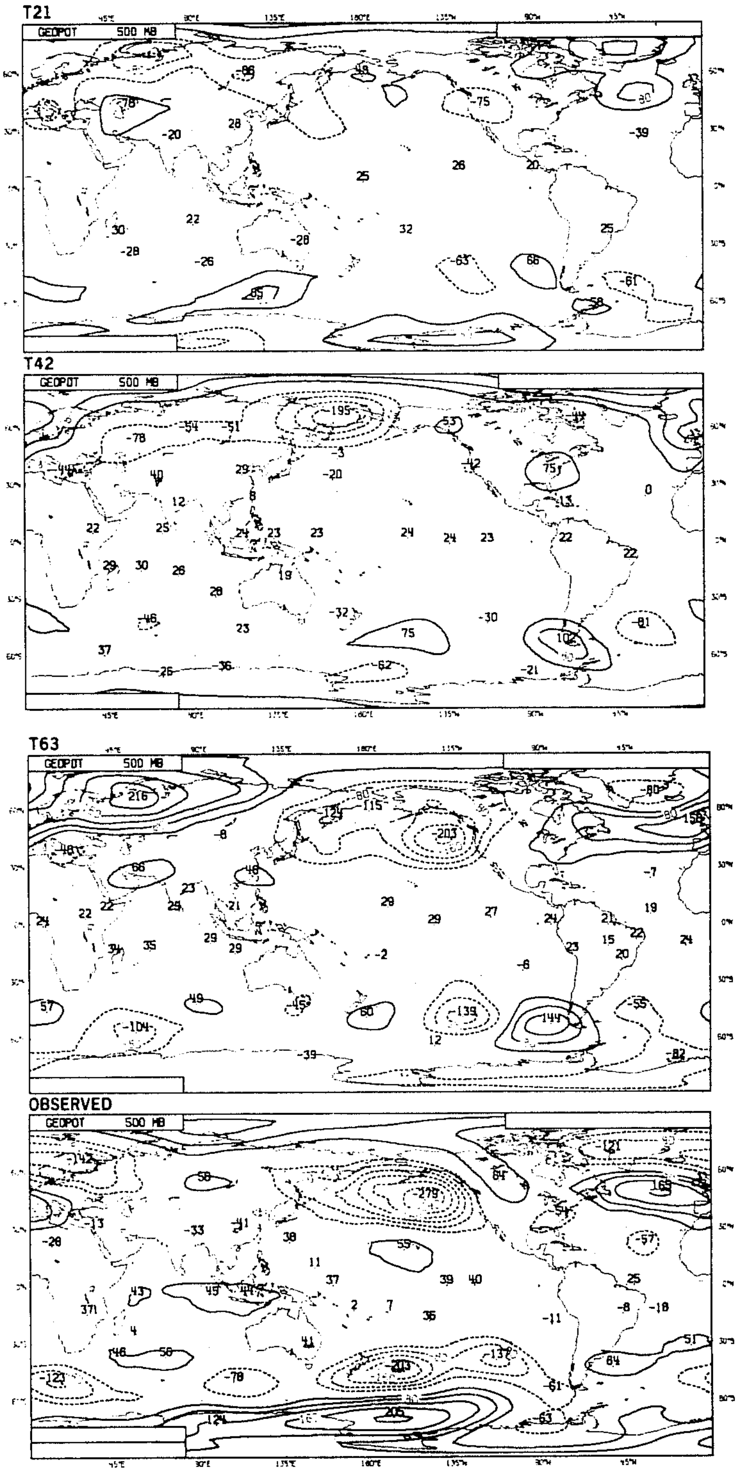


Fig.8 The anomaly of the 500 mb height field averaged for days 11 to 40 for the El Nino case.

anomalies. The magnitude of the predicted anomalies increaseswith increasing resolution. Comparing the T 63 results with the observed map we see very good results because the large negative anomaly in the Gulf of Alaska is predicted, albeit with somewhat smaller magnitude than observed.

Similarly, the positive anomaly across the Atlantic accompanied by a negative anomaly over Greenland are also caught rather well. The major error is that the predicted large positive anomaly over the northern part of U.S.S.R. is too intense and not in the correct position, while the negative anomaly observed over Scandinavia is unpredicted.

Fig. 9 shows the anomaly correlation coefficient as a function of time to

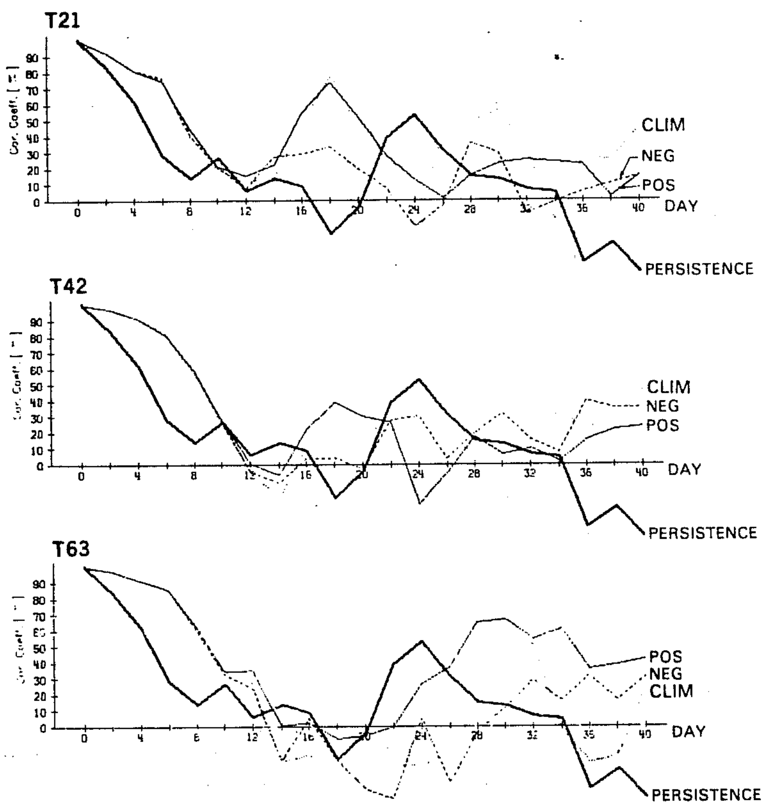


Fig.9 The anomaly correlation for the 500 mb height field and wavenumbers 1-3 in the region 20.5 - 82.5°N. El Nino case.

40 days. Concentrating on the results for T 63 we may note the following features:

* the SSTA has no impact during the first 10 days
* the POS SSTA is the best forecast during all 40 days
* the NEG SSTA does not have predictive value after approximately 8 days

* all curves indicate the normal loss of deterministic predictability after 8 days

* both CLIM and POS show positive values after 22 days with a broad maximum around 32-25 days. POS is better than CLIM in the period.

Assuming that this single case is representative (which may not be the case) we could conclude that the response to the SSTA is observed after about 22 days, say 3 weeks, and that the signal from the SSTA has a definite tendency to improve the prediction. In this single case it is found that the improvement is so large that the anomaly correlations on individual days reach a level of about 0.6. This phenomenon has been called the recovery of the forecast.

In view of these results, and because individual smaller meteorological systems will be unpredictable on this time scale, it was decided to calculate 10-day mean anomaly correlation coefficients. They are shown in Fig. 10 which indicates for T 63 that the observed SSTA (POS) produces the best forecast among the three integrations. In this figure we have introduced a horizontal line, called NORM, with a value slightly less than 0.3. This level was selected as the lowest value for a useful forecast using correlations for 10 day mean maps. It corresponds to the level of o.6 used as the lower limit for useful daily deterministic forecasts. The value was derived from multi-year climate forecasts by asking how much skill the model climate possesses in predicting the 10 day means in the period considered in our study. There were 12 cases. The measure of skill was obtained by averaging the 12 coefficients and adding 1.96 times the standard deviation. Measured in this way it is only the POS prediction which has skill during the whole period.

Since the recovery period starts after 3 weeks it was decided to look at the 10 day period, 21-30 days. Looking again at the T 63 results, we show in Fig. 11 the 500 mb mean height anomaly, the control (CLIM) and the El Nino

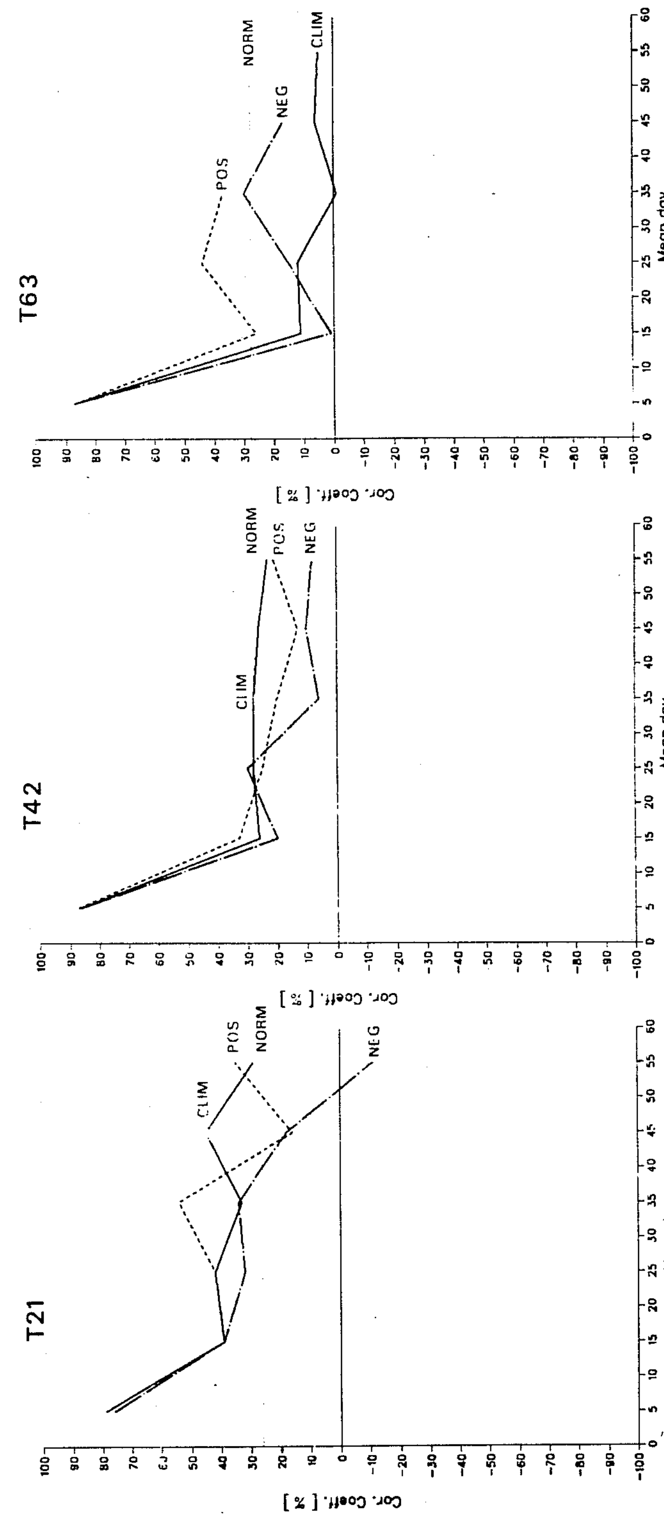


Fig.10 The 10-day mean anomaly correlations for the 500 mb height field. El Nino case.

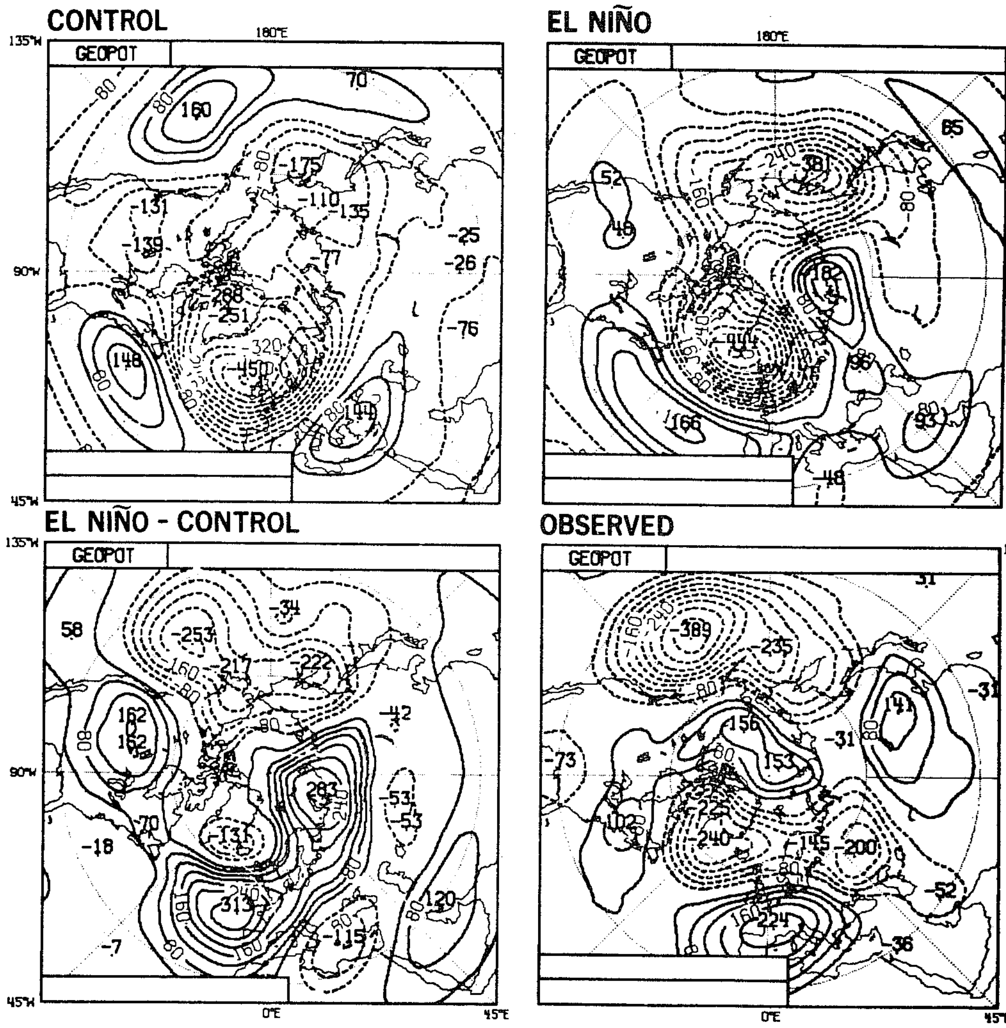


Fig.11 The 500 mb mean height field anomaly averaged for days 21 to 30, T63, El Nino case.

(POS) forecasts. These maps are shown in the upper part of the figure. The chart in the lower left corner shows the impact of the El Nino anomaly. The lower right corner contains the observed anomaly averaged for the same 10 days. It is evident that the El Nino prediction is better than the control forecast, but that it leaves much to be desired.

Fig. 12 shows that especially the T 63 forecasts are systematically better when realistic SST's are used.

We conclude:

* The SST has an impact on the forecasts in the extended range, but

practically no impact during the first 10 days.

* A resolution corresponding to at least T 63 is necessary

* A positive impact is found for realistic SST

* The forecast may recover after about 3 weeks

* The extended forecasts are obscured by the systematic errors in the model.

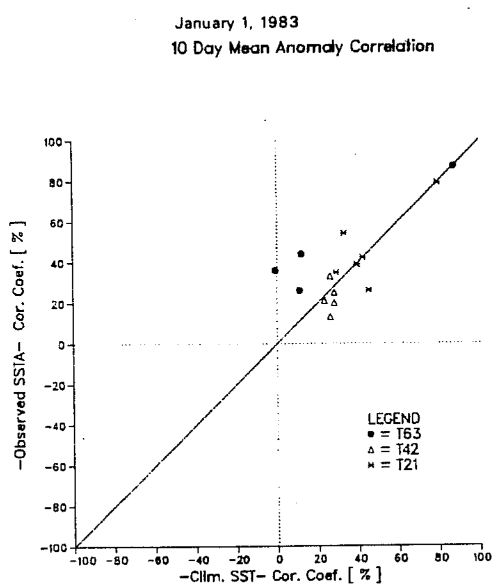


Fig.12 Intercomparisons of the skill of the 10-day means between the experiments with the positive anomaly and the climatological SST. El Nino case.

5. CONCLUDING REMARKS.

The investigations of the atmospheric response to the interactions with the oceans through the transfer processes in the field of extended-range prediction have only just begun. There is sufficient evidence at this time to conclude that the impact of sea-surface temperature anomalies on the global circulations in all latitudes is significant in the extended-range predictions, but probably rather insignificant, except locally, in the medium-range time interval. The preliminary results also suggest that the observed SSTA's, especially in the tropical latitudes, play a major rôle in explaining rather drastic changes in global circulation pattern due to the rather large anomalies created by the SSTA's with a time scale of 3-4 weeks.

A major question for future research is how complete the models need to be to give a realistic treatment of the physical and hydrodynamical processes which are involved in the complicated interactions between the two major parts of the complete system, i.e. the oceans and the atmosphere. The present approach which keeps the SSTA's constant in time for periods of several weeks is probably not good enough. The interactions between ocean and atmosphere giving a realistic description of a complicated phenomenon such as an El Nino event should involve two way interactions and not only a stagnant ocean interacting with the atmosphere.

Eventually we should, in a single model which couples the ocean to the atmosphere, be able to account for the processes which create the SSTA, calculate the atmospheric response, and thereby also predict the atmospheric and oceanic changes which remove the SSTA and brings the ocean back to a normal state or to another disturbed state.

With these thoughts in mind it becomes important to understand both the atmospheric and oceanic processes to a degree which will permit the formulation of an adequate model. The design of coupled models should continue, but it is equally important to search for analytical models or low-order, idealized systems which can reproduce the principal aspects of large sea surface temperature anomalies, their creation, maintenance and destruction. Before we do this we cannot expect to draw the full benefit of the predictive value of the existence of these abnormal conditions.

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