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THE INFLUENCE OF THE OCEAN ON CLIMATE

Lectures presented at the twenty-eighth session of the WMO Executive Committee

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NOTE

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FOREWORD

"The influence of the ocean on climate" was selected as the theme for the scientific discussions at the twenty-eighth session (1976) of the WMO Executive Committee. The subject is of particular interest to meteorologists, since the oceans, covering as they do nearly three-quarters of the globe, have a significant influence on weather and climate.

Two outstanding lectures were presented, one by Dr. W. L. Gates and the other by Professor K. Hasselmann, both of whom are acknowledged experts in this field. The first lecture, delivered by Dr. Gates, was entitled "Modelling the ocean-atmosphere system and the role of the ocean in climate". It was followed by Professor Hasselmann's lecture, entitled "The dynamical coupling between the atmosphere and the ocean". The texts of both lectures are reproduced in the present publication.

I am pleased to have this opportunity of expressing to Dr. Gates and Professor Hasselmann the sincere appreciation of the World Meteorological Organization for the time and effort they have devoted to the preparation of these valuable papers.

D. A. Davies

D. A. Davies

Secretary-General

THE DYNAMICAL COUPLING BETWEEN THE ATMOSPHERE AND THE OCEAN

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Summary

The heat flux from the oceans to the atmosphere and the heat storage within the ocean exert a strong influence on the atmospheric circulation and its longer term variability. Internal time scales of the ocean range from 10^{-2} to 10^3 years, encompassing time scales of particular interest for climate study. Thus detailed prognostic models of the ocean are needed for the construction of comprehensive climate models. Of particular significance for climate variability is the effect of coupling the rapidly varying atmosphere to the more slowly responding ocean (and other long time-constant components of the climatic system). It is shown that the atmosphere acts as a white noise generator, producing low-frequency climate variations characterised by a red spectrum, in general agreement with observations. A complete stochastic climate model requires a deterministic prognostic description of the slowly varying components of the system together with a diagnostic statistical representation of the rapidly varying atmospheric subsystem. The emphasis this places on the dynamics of the oceans and cryosphere is not reflected in present monitoring networks designed for short-term weather prediction rather than the investigation of long-term climate variability.

1. Characteristic time scales of the ocean

The circulation systems of the atmosphere and the ocean are strongly coupled through interactions at the air-sea interface. The fluxes of latent and sensible heat from the ocean to the atmosphere represent the principal energy source of the atmospheric circulation, and, conversely, the momentum transfer from the atmosphere to the ocean, together with the heat fluxes, are the main driving forces of the oceanic circulation. Thus neither circulation system can be properly understood without some knowledge of the other. Because of the great density difference between the two media, however, the principal thermal and dynamical time scales of the atmosphere and the ocean differ by several orders of magnitude, and the view of the coupled system will therefore depend strongly on the time scale considered.

For the purposes of short-term weather prediction, the sea-surface temperature controlling the heat flux to the atmosphere may be regarded as a given, constant boundary value, and the prognostic problem lies entirely in the atmosphere. For the longer time scales of interest for climate studies, however, the roles of the atmosphere and the ocean are basically reversed: a detailed prognostic model of the ocean is now needed, whereas the relevant properties of the atmospheric circulation can be expressed in terms of strongly reduced averaged quantities.

This paper will be concerned generally with the properties of a system consisting of two coupled subsystems with widely differing response times, with particular emphasis on the time-scale range appropriate to the slowly responding system. This problem is clearly relevant not just for the ocean-atmosphere system, but generally for a climatic system containing a number of slowly responding sub-systems, such as the cryosphere, land vegetation, geochemical cycles, etc. However, the general concepts will be illustrated in terms of the ocean-atmosphere system, and for this purpose a brief summary of the principal response time scales of the ocean (to the extent that they are known) will be useful.

To a first approximation the ocean may be divided into two layers, a seasonally responding upper layer, with depths typically between 20 m and 200 m, and the rest of the ocean beneath. Wind mixing creates a fairly homogeneous layer at the top of the ocean, which is separated from the interior of the ocean by a highly stratified layer, the seasonal thermocline. The mixed layer together with the seasonal thermocline act as a very effective thermal buffer shielding the interior of the ocean from seasonal changes of the heat budget. The response of the seasonal layer is primarily controlled by vertical transfer processes and can now be reasonably well modelled (cf. Niiler, 1975). Depending on the depth of the layer, typical thermal relaxation times lie in the range of weeks to months.

The thermal response time of the deep ocean is more difficult to estimate, as this depends on the detailed structure of the full three-dimensional heat transport. There exists a general qualitative picture, but as yet no unique quantitative description of the deep ocean circulation. Current-meter measurements of the mean circulation are difficult to make (except in restricted regions of strong currents) because of the large natural low-frequency variability of the ocean, which requires extremely long averaging periods of several years to remove. Indirect estimates based on simple one-dimensional models of the heat balance between the vertical diffusion of heat downwards and upwelling of cooler water from below yield thermal response and circulation cycle times of the order of 10^2 to 10^3 years (cf. Munk, 1964). Recent studies with geochemical tracers (Rooth, in preparation) and numerical experiments with general circulation models (e.g. Bryan and Cox, 1968) suggest that recirculation within the upper part of the ocean may reduce the cycle time for these layers by perhaps an order of magnitude. A tentative picture of the average response times of the ocean as a function of layer depth (excluding high frequency inertial-gravity modes) may therefore be sketched as in Fig. 1. Locally, this picture may be significantly modified. For example, in regions of strong currents, upwelling and in the equatorial zones, where the geostrophic restraint vanishes, a rapid response of the upper and intermediate layers of the ocean to atmospheric forcing can occur on appreciably shorter time scales of the order of days to weeks.

In summary, the dynamic variability of the ocean can be characterized by a broad spectrum of natural time scales varying from 10^{-2} to 10^3 years. If interactions of the ocean with sea ice are included, the time scale is further extended by at least one order of magnitude.

Unfortunately, our understanding of the dynamics of the oceans in the climatically most relevant time-scale range 1 - 10^3 years is still very rudimentary. Although ocean general circulation models (OGCM's) have been in use for several years, it is still not clear whether these yield a quantitatively correct description of the dynamical processes controlling the mean oceanic circulation. In part this is due to the difficulty of obtaining observations in the ocean to verify the models and test parameterizations. However, there are also more basic problems. For example, it is known that very fundamental features of the atmospheric circulation such as the three-cell meridional structure cannot be reproduced in numerical models unless baroclinic eddies are adequately resolved. Measurements have now clearly established that quasi-geostrophic eddies analogous to synoptic-scale atmospheric disturbances can be found in nearly all regions of the ocean. However, the ratio of the eddy-kinetic energy to the energy of the mean circulation in the oceans is 10 : 1, as compared with a ratio of order 1 : 1 in the atmosphere. Furthermore, because of the smaller Rossby radius of deformation in the ocean, the characteristic scale of oceanic eddies is typically 100 - 400 km, an order of magnitude smaller than in the atmosphere (Fig. 2). Thus, although they contain the major part of the kinetic energy, baroclinic eddies cannot be adequately resolved in global oceanic circulation models and must accordingly be parameterized. This problem has yet to be resolved.

In the following discussion of the evolution of a system consisting of two coupled subsystems of different characteristic time scale, it will be assumed that the prognostic equations of the slowly varying system are known. In the case of the ocean this is clearly approximately valid only for time scales up to a year. For longer time scales one will need to resort to strongly simplified models until more progress is made in the fundamental problem of modelling the general circulation of the ocean.

2. Relationship between general circulation models (GCM's), statistical dynamical models (SDM's) and stochastically forced models (SFM's)

Most theoretical climate investigations have been carried out with either high resolution general circulation models (GCM's) or highly idealised statistical dynamical models (SDM's). GCM's provide a valuable tool for sensitivity studies of the equilibrium response of the atmospheric system under different boundary conditions, but are normally too time consuming for explicit climate evolution studies. SDM's, on the other hand, are sufficiently reduced that they permit long-time integrations, but because of their simple internal structure they normally yield only a single time independent climate state for given external conditions. Thus changes in climate cannot be explained with these models as the result of internal interactions within the system, but must be attributed to changing external conditions, such as solar radiation variations, volcanic activity, etc. Although it is conceivable that climate variability is indeed externally generated, attempts to correlate observed climate changes with changes in external parameters have not been entirely convincing. The present investigation will therefore be concerned with the alternative hypothesis that climate change is caused by interactions within the climate system itself.

Self-sustained climate oscillations of a quasi-random manner may be expected, by analogy with turbulent flow and other nonlinear systems, if the internal structure of a climate model is sufficiently complex and contains a number of nonlinearly interacting components. However, a much simpler model will be considered here, in which climate variability is ascribed to the coupling of the slowly varying components of the climate system to the rapidly varying atmosphere. The atmosphere acts essentially as a white-noise generator, and climate variability arises as the low-frequency integral response of the slowly varying part of the system to this noise input. Complex internal feedback processes within the climatic system need not be postulated. In fact, climate variability with features qualitatively very similar to observations can be generated with very simple stochastically forced models based on the Budyko (1969)-Sellers (1969) model.

The difference in structure of an SFM and an SDM or GCM is best seen using a general notation (Hasselmann, 1976). Let the two sub-systems of the complete climate system be described by a finite number of degrees of freedom,

$$\underset{\sim}{x} = (x_1, x_2, \ldots x_p)$$

$$\underset{\sim}{y} = (y_1, y_2, \ldots y_q)$$

where the variables $x_1, x_2, \ldots x_p$ denote the rapidly varying components of the climate systems (the atmosphere), and $y_1, y_2, \ldots y_q$ the slowly varying components (the oceans, cryosphere, and land surface). The basic equations of the climate system will then be of the general form

$$\frac{dx_i}{dt} = u_i(\underset{\sim}{x}, \underset{\sim}{y}) \qquad (1)$$

$$\frac{dy_i}{dt} = v_i(\underset{\sim}{x}, \underset{\sim}{y}) \qquad (2)$$

where u_i and v_i are given functions, and it is assumed that the characteristic time scales $\tau_x = O(x_i/\frac{dx_i}{dt})$, $\tau_y = O(y_i/\frac{dy_i}{dt})$ of the systems $\underset{\sim}{x}$ and $\underset{\sim}{y}$ satisfy the two-scale inequality

$$\tau_x \ll \tau_y$$

In a typical GCM experiment, the integration time τ of the system (1), (2) lies in the range

$$\tau_x \ll \tau \ll \tau_y \qquad (3)$$

The climate variables $\underset{\sim}{y}$ can then be regarded as approximately constant, and the problem reduces to the integration of the atmospheric equations $\frac{dx_i}{dt} = u_i(\underset{\sim}{x}, \underset{\sim}{y}_o)$ under given boundary conditions $\underset{\sim}{y} = \underset{\sim}{y}_o$.

In the SDM, on the other hand, one is interested only in the slow evolution of a reduced climatic system for times of order τ_y, and equation (2) is therefore time-averaged over a period τ satisfying the inequality (3). (In addition the system (2) is normally further reduced by spatial averaging, but for simplicity of argument this ignored; it is assumed that $\underset{\sim}{y}$ refers already to the spatially averaged fields). To obtain a closed system for $\underset{\sim}{y}$ from the averaged equation (2),

$$\frac{d}{dt}\langle y_i\rangle = \langle v_i(\underset{\sim}{x}, \underset{\sim}{y})\rangle \tag{4}$$

a closure hypothesis for the nonlinear expressions involving the weather variables $\underset{\sim}{x}$ in the averaged function $\langle \underset{\sim}{v}\rangle$ has to be introduced, in order that the right hand side can be expressed as a function $\hat{v}_i$ of $\underset{\sim}{y}$ only,

$$\frac{d\langle y_i\rangle}{dt} = \hat{v}_i(\underset{\sim}{y}) \tag{5}$$

In the following, the averages in (4) and (5) will be interpreted, for formal convenience, as ensemble averages over the set of all weather states $\underset{\sim}{x}$ for given $\underset{\sim}{y}$.

Stochastically forced models may also be regarded as simplified models designed for long-time integration over periods of order τ_y. However, the essential difference of the SFM as compared with the SDM is that the fluctuating terms in the climate equations (2) are not averaged out. Writing $\underset{\sim}{y} = \langle \underset{\sim}{y}\rangle + \underset{\sim}{y}'$, $\underset{\sim}{v} = \langle \underset{\sim}{v}\rangle + \underset{\sim}{v}'$ and subtracting equation (4) from (2), the evolution of the fluctuating component is given by

$$\frac{dy_i'}{dt} = v_i'(\underset{\sim}{x}, \underset{\sim}{y}) \tag{6}$$

For short times τ satisfying (3), $\underset{\sim}{y}$ can be regarded as constant in the right hand side of (6), and the equation reduces to

$$\frac{dy_i'}{dt} = v_i'(\underset{\sim}{x}(t), \underset{\sim}{y}_o) = v_i'(t) \tag{7}$$

where v_i' is a stochastic function of t. The statistics of v_i' may be regarded as given if the statistics of $\underset{\sim}{x}$ are known.

Equation (7) is the well known equation describing the dispersion of particles of position $\underset{\sim}{y}'$ in a turbulent fluid with a given (Lagrangian) velocity field $\underset{\sim}{v}'$. It was shown by Taylor (1921) that for a statistically stationary field $\underset{\sim}{v}'$, the process $\underset{\sim}{y}'$ is non-stationary, the variance of $\underset{\sim}{y}'$ increasing linearly with time τ,

$$\langle \delta y_i \delta y_j\rangle = 2D_{ij}\tau \qquad (\tau_x \ll \tau \ll \tau_y) \tag{8}$$

where $\delta y_i = y_i'(\tau) - y_i'(\tau=0)$,

$$D_{ij} = \frac{1}{2}\int_{-\infty}^{\infty} R_{ij}(\tau)\,d\tau = \pi F_{ij}(\omega)_{\omega=0}$$

and $R_{ij}(\tau) = \langle v_i'(t+\tau)v_j'(t)\rangle$, $F_{ij}(\omega)$ represent the covariance function and spectrum of the process $v_i'(t)$ respectively.

The corresponding relation in the spectral domain is given by

$$G_{ij}(\omega) = \begin{cases} \dfrac{F_{ij}(0)}{\omega^2} & (\tau_y^{-1} \ll \omega \ll \tau_x^{-1}) \qquad (9a) \\ 2\pi F_{ij}(0)\delta(\omega)\tau & (\omega \lesssim \tau^{-1}) \qquad (9b) \end{cases}$$

where $G_{ij}(\omega)$ is the spectrum of the response δy_i.

Equations (8), (9a,b) are restricted to times τ satisfying (3), during which the process v_i' can be regarded as stationary and the diffusion coefficient is constant. The stochastic forcing model predicts that for this time range, without feedback, the climate variance increases at a constant rate. Thus from the viewpoint of the SFM, the problem of climate variability is not to discover various forms of positive feedback which could enhance the small variations introduced into the system externally - as in the usual SDM model - but rather to consider the negative feedback mechanisms which must be present in the system to control the continual growth of climate variability generated by the stochastic weather forcing.

If a stabilising linear negative feedback is included, equation (7) takes the form

$$\frac{dy_i'}{dt} = v_i' - \lambda y_i' \qquad \lambda = \text{const} > 0 \qquad (10)$$

of a first-order auto-regressive process. In this case, the variance $\langle\delta y_i \delta y_j\rangle$ approaches a limiting value for large τ, and the (non-integrable) ω^{-2} pole in the response spectrum (9a) is removed (cf. Fig. 3).

In the general case of nonlinear feedback, the evolution of the climate system can no longer be described completely by the first and second moments. A closed treatment then requires consideration of the evolution of the complete climate probability density function $p(\underset{\sim}{y})$ in the climate phase space $\underset{\sim}{y}$. This is governed by a Fokker-Planck equation, an extended form of the Liouville equation describing the conservation of probability in phase space in which the effect of the stochastic forcing is represented by additional diffusion terms (Wang and Uhlenbek, 1945).

The Fokker-Planck equation provides a physical basis for the discussion of climate predictability. It can be shown that despite the stochastic nature of atmospheric forcing, climatic systems with stabilising feedback still have a certain degree of predictability (Hasselmann, 1976). This is because the feedback terms, in contrast to the stochastic forcing, represent deterministic functions of the instantaneous climatic state. However, for a statistically stationary climatic system the deterministic feedback and stochastic forcing terms are generally of the same order, and the resultant predictive skill is maximally of order 50 %.

3. Applications

(a) Sea Surface Temperature (SST) Variability

One of the simplest models of the ocean which has been used for long-term air-sea interaction studies is the "copper plate" model (e.g. Salmon and Hendershott, 1976). Only the upper mixed layer is considered, and it is assumed that the layer is isothermal, of constant depth, and insulated from the deeper layers. The model provides a rough first-order description of the oceanic thermal response for time scales from a week up to a few years (cf. Thompson, 1976). The rate of change of temperature of the ocean is directly proportional to the local heat transfer across the sea surface. Fig. 4 shows a comparison of the SST spectrum computed from this model with observations taken at the Atlantic Weather Ship INDIA (from Frost, 1975). Good agreement is found if the linear feedback parameter is chosen as $\lambda = (4.5 \text{ months})^{-1}$. Note that the ω^{-2} relation for higher frequencies follows immediately from the assumption of a white input spectrum at low frequencies, in accordance with the two-scale hypothesis (3). Fig. 5 shows that the spectrum of the atmospheric fluxes of latent and sensible heat are indeed white, as postulated. Quantitatively, the levels of the input and response spectra are consistent in this example if a mixed-layer depth of 100 meters is assumed, in reasonable agreement with observation. Further examples of SST variations forced by short-time-scale random atmospheric fluxes are given in Frankignoul and Hasselmann (1977).

(b) A Stochastically Forced Budyko-Sellers Model

Fig. 6 shows the climate variance spectra computed from a zonally averaged, stochastically forced Budyko-Sellers model (Lemke, 1977). In the usual non-stationary SDM version of this model, the rate of change of the climatic state is governed by the net time-averaged radiation balance and the divergence of the meridional heat flux. Stochastic forcing terms were introduced into the model by retaining the fluctuating components of the various fluxes in addition to the mean values. The zero-frequency levels of the variance spectra of the fluxes, which determine the diffusion matrix D_{ij}, were not available in published form, but reasonable estimates based on the data of Starr and Oort (1973) yielded climate variance spectra in general qualitative agreement with observations. The computations were carried out both for a single-layer "copper-plate" model and for a two-layer model. In the latter case the lower layer was represented as a homogeneous thermally conducting medium with a thermal conductivity $K = 1$ cal cm^{-2}sec^{-1}. The difference between the two computations underscores the need for more sophisticated models of the deep ocean circulation.

4. Implications for the construction of climate models

Both SDM's and SFM's require parametrisation of the rapidly responding atmospheric system $\underset{\sim}{x}$ in order to obtain closed equations for the slowly varying climate components $\underset{\sim}{y}$. In the case of SDM's the dependence of $\langle v_i \rangle$ on the $\underset{\sim}{x}$-statistics in equation (4) must be expressed in terms of $\underset{\sim}{y}$; for SFM's one needs in addition the functional dependence of the diffusion coefficient D_{ij} on $\underset{\sim}{y}$. The quantities $\langle v_i \rangle$ and D_{ij} will clearly depend not only on the atmospheric system $\underset{\sim}{x}$, but also on the variables and type of internal coupling used in the climate model - as illustrated by the two very simple but structurally quite different examples discussed above. It should be noted, however, that only very limited statistical information of the atmospheric system is in fact needed, and that this information can be obtained by relatively short numerical experiments with atmospheric GCM's based on integration times τ in the intermediate range (3). A conceptually straightforward procedure for deriving closed equations for an SFM would be to carry out a series of sensitivity experiments with a high resolution GCM for a number of different climate boundary conditions $\underset{\sim}{y}$, and to parametrise the inferred values $\langle v_i \rangle$, D_{ij} as empirical functions of $\underset{\sim}{y}$.

According to this view of climate modelling, the prognostic equations of the climate system are thus limited to the slowly varying components, which in the usual atmospheric GCM are treated as fixed boundary conditions, whereas the entire atmospheric system is parametrised in terms of mean feedback and stochastic forcing coefficients. Despite this considerable reduction in information through a statistical treatment of the atmosphere, however, it will still be impossible in practice to simulate all the interactions of the slowly varying system $\underset{\sim}{y}$ in adequate detail. The structure of a numerically feasible stochastic climate model will therefore necessarily take the form shown in Fig. 7, where $\underset{\sim}{y} = (\underset{\sim}{y}_p, \underset{\sim}{y}_d)$ is subdivided into a prognostic subsystem $\underset{\sim}{y}_p$ and a parametrised diagnostic subsystem $\underset{\sim}{y}_d$, in analogy with the similarly unavoidable subdivision of the atmospheric system $\underset{\sim}{x}$ in atmospheric GCM's.

5. Conclusions

In summary, one is faced with two basic tasks in understanding and modelling the climatic system in the range of time scales in which coupling with the ocean is important:

(1) The construction of physically realistic but nevertheless simple models of the global ocean circulation which can be integrated numerically for the long time periods in question (and the incorporation of these models within more comprehensive climate models including other slowly responding components of the climate system).

(2) The determination of the atmospheric feedback and random forcing coefficients which drive the ocean circulation, and their parameterization in terms of the ocean circulation itself.

Both of these developments are still in an early stage. However, it is already clear that ultimately both of these problems can be tackled successfully only if there exists an adequate long-time data base against which climate models can be tested. Ship reports, weather ship data and long time series from a few oceanic stations have already proven invaluable for studies of long-term ocean-atmosphere interaction. In fact, it can be argued that real progress in this field has been achieved only on the basis of such long-term data. However, for a serious investigation of global ocean-atmosphere interactions on climatic time scales the existing data base is extremely sparse. By far the greatest proportion of the WWW network is concentrated on the land surfaces, which cover less than one third of the earth's surface. This distribution has developed in response to the requirements of short-term weather prediction. However, in the climate problem the role of the oceans and the atmosphere is reversed: it is the interactions within the slowly responding ocean system and other long time-constant components of the climatic system which ultimately determine long-period climatic changes. Thus the optimal distribution of resources for permanent recording stations on land and in the oceans will need to be reconsidered with respect to the requirements of a future climate programme.

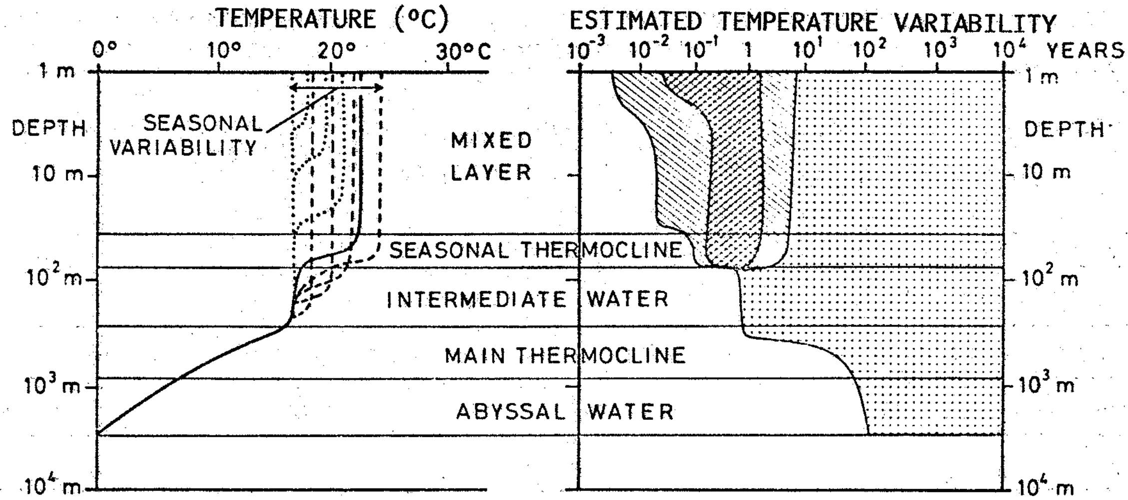

Figure 1 - Estimated time scales of variability of the ocean. Heavier shading denotes larger levels of variance

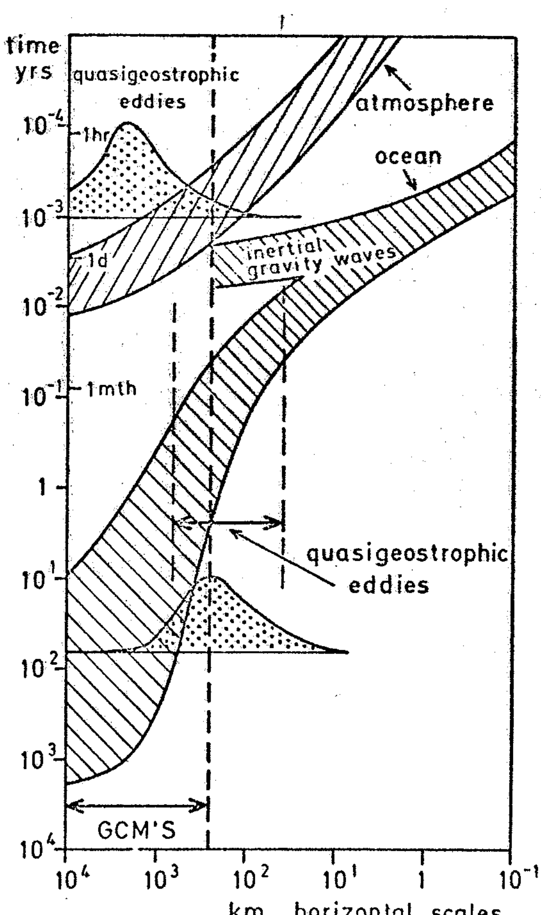

Figure 2 - Frequency-wavenumber bands of variability in the atmosphere and the ocean. Stipled areas indicate variance spectra of quasi-geostrophic eddies

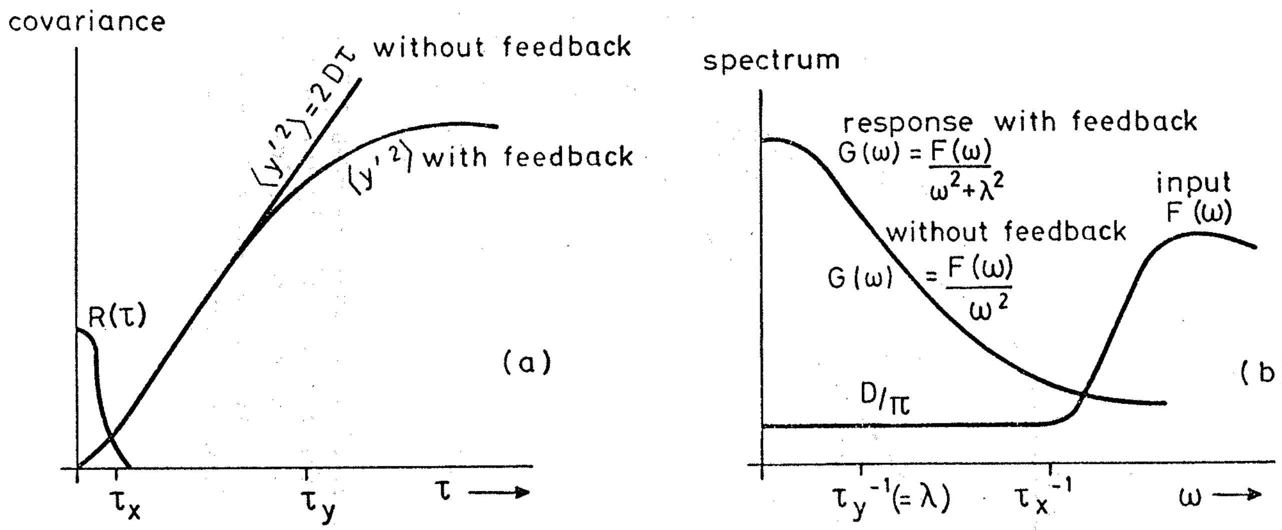

Figure 3 - (a) Variance and (b) spectral response of climate system to stochastic atmospheric forcing with and without linear feedback

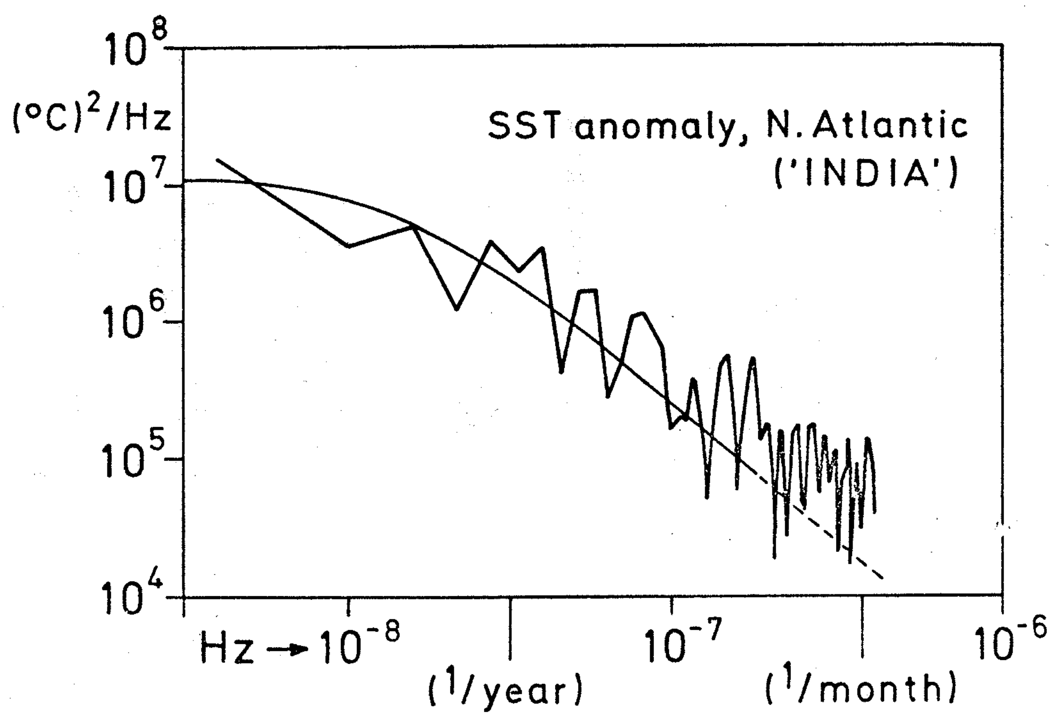

Figure 4 - Computed variance spectrum of SST anomaly (without seasonal term) for linear feedback parameter λ = (4.5 months)$^{-1}$ (from Frankignoul and Hasselmann, 1977) and observed spectrum at Ocean Weather Station INDIA (from Frost, 1975)

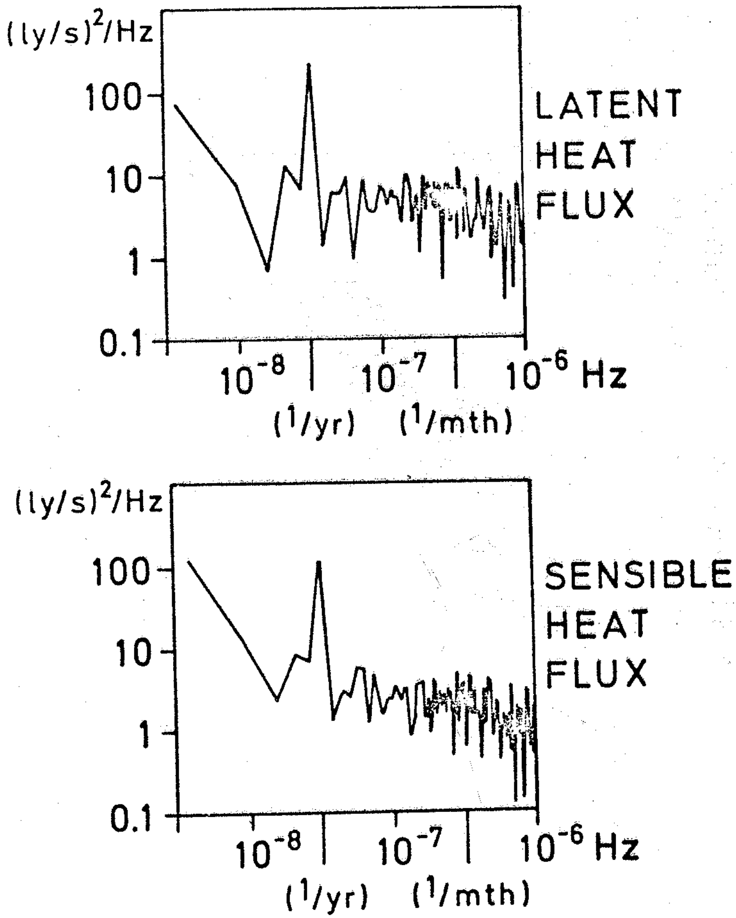

Figure 5 - Variance spectrum of latent and sensible heat flux at OWS INDIA (from Frost, 1975). Except for the annual line, the spectra are white, as required for the computation of the theoretical SST spectrum shown in Figure 4. The spectral levels in Figures 4 and 5 are mutually consistent if a mixed-layer depth of 100 m is assumed

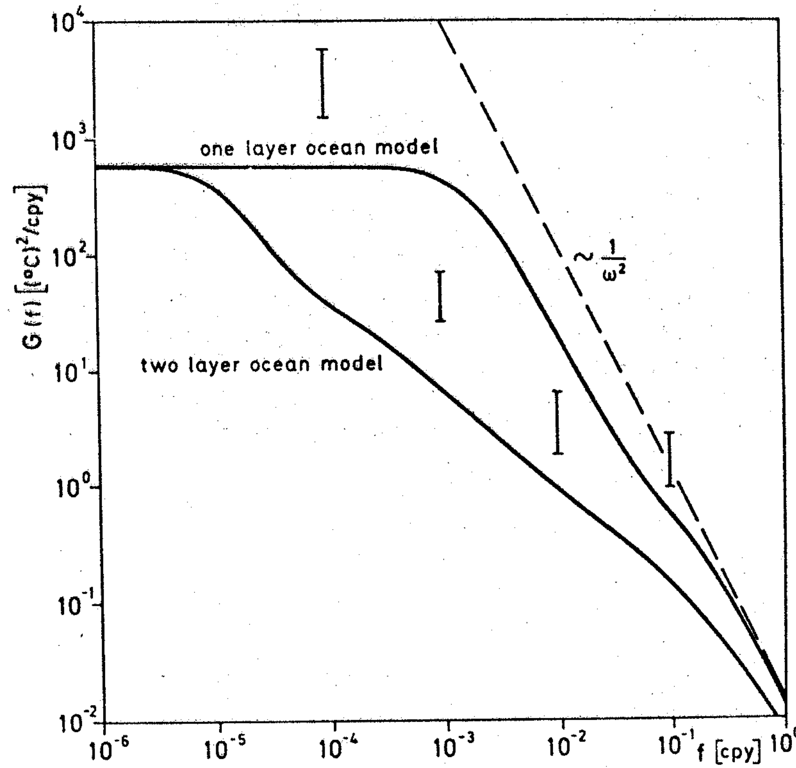

Figure 6 - Variance spectra of zonally-averaged temperature fluctuations for a stochastically-forced Budyko-Sellers model with a one- and two-layer ocean. Vertical bars denote observed temperature variations for central England (from Kutzbach and Bryson, 1974)

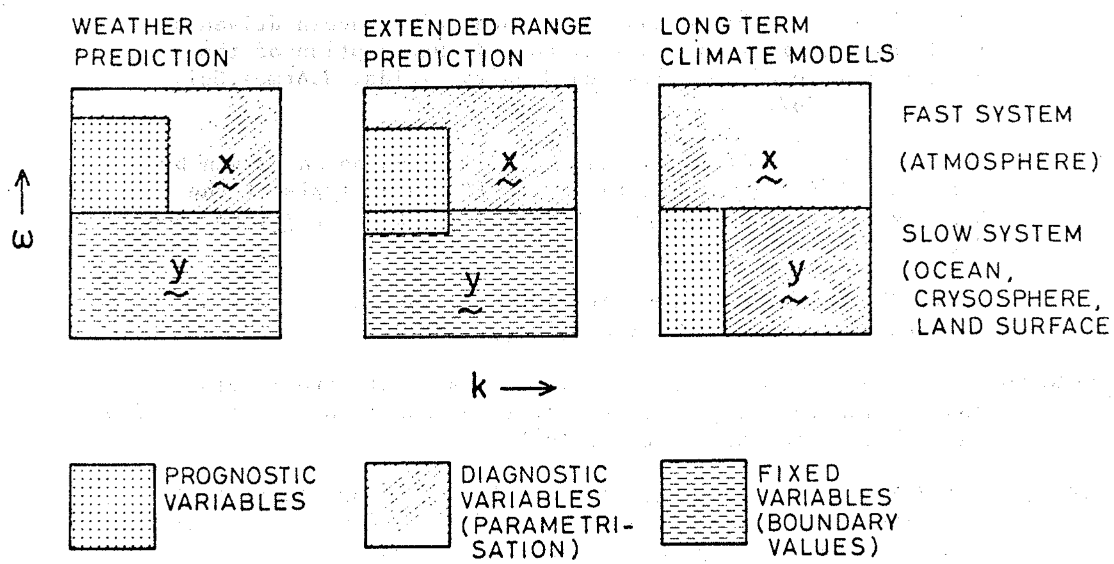

Figure 7 - Schematic structure of prognostic models for short-term weather prediction, extended range weather prediction and long-term climatic models

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