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THESIS

THE EFFECT OF STOCHASTIC SURFACE
HEAT FLUXES ON THE CLIMATOLOGY
OF THE SEASONAL THERMOCLINE

by

David C. Copley

June 1984

Thesis Advisor: Roland W. Garwood Jr.

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The Effect of Stochastic Surface Heat Fluxes on the Climatology of the Seasonal Thermocline

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I. INTRODUCTION

The object of this paper is to show the effect of stochastic surface heat flux forcing on the climatology of a model seasonal thermocline. It was theoretically demonstrated by Hasselmann (1976) that if "weather" forcing can be considered to be white noise, then the ocean acts as an integrator. The result is a red-shifted response spectrum. Frankignoul and Hasselmann (1977) developed a simple model to test this theory. Here we shall expand on their result by considering seasonal variations and mixed layer physics.

Climate researchers have long been aware of the pronounced variability associated with climatic records of sea-surface temperature (Namias, 1959; Wyrtki, 1965). This temporal variability spans a spectral range from a low frequency limit of the order of ice ages (one cycle per 100,000 years) to a high frequency limit on the order of one cycle per month.[1] Higher frequencies than one cycle per month are taken to be stochastic in this study, whereas mean quantities and climatologies will have seasonal or longer frequencies. An understanding of the origin of this diverse climatic variability is a primary goal of climate research.

Attempts have been made to link climatic changes to various external factors such as solar activity, perturbations in the Earth's orbit, or the increased atmospheric turbidity associated with volcanic eruptions. The primary difficulty with these theories is that the input-response relationships are not sufficiently pronounced to be immediately obvious upon inspection of the appropriate time series.

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[1]The limits follow the definitions adopted in GARP Publication 16, 1975.

Recent interest, both commercial and tactical, focuses on the higher frequency sea surface temperature (SST) perturbations. The 1983 El Nino (seasonal frequency phenomenon) proved disastrous to the commercial fishing industry along the west coast of the U.S. and South America. Tactically, the strategic defense of the U.S. is linked to the proper utilization of resources in the vicinity of SST anomalies and other thermal structures which alter acoustical propagation through the oceans.

Large scale SST-weather relationships have been discovered through the statistical analysis of extended time series (Bjerknes, 1966; Namias, 1969). Numerical experiments have demonstrated statistically significant modifications of weather patterns in the tropics by critically located SST anomalies. These anomalies have been studied through the use of General Circulation Models (Shukla, 1975; Rowntree, 1976) and/or Statistical Dynamical Models.

Much of the theory for ocean mixed layer modeling depends on the validity of two hypotheses. The first is that vertical mixing in the ocean mixed layer and entrainment at the thermocline occur as a result of atmospheric forcing: the heat fluxes across the air-sea surface and the surface wind stress. The second hypothesis is that the mechanical energy budget holds the key to understanding the ocean mixed layer dynamics (Garwood, 1979).

The stochastic forcing concept is another approach to understanding the ocean SST response to atmospheric forcing. Lemke (1977) applied the stochastic forcing concept to a Budyko-Sellers type model to extend the concept to a global stochastic energy model applicable for time scales from 10 to 10,000 years. Frankignoul and Reynolds (1983) examined the statistical characteristics of SST anomalies in the presence of a mean current. Their studies indicate the importance of surface heat flux and wind mixing fluctuations on the production of SST anomalies from the available North Pacific Experiment (NORPAX) data.

Hasselmann (1976) proposed the use of a stochastic forcing model to study the ocean response to atmospheric forcing. He assumed that, because the time scale of the ocean response is so much greater than the time scale of the atmospheric forcing, the atmospheric forcing can be considered to be represented as a white noise time series. Via this representation, he showed that the slow climatic variability of the ocean can be interpreted as an integral response to the random excitation provided by "weather" disturbances.

In part two of the series on stochastic climate models (Frankignoul and Hasselmann, 1977), the stochastic forcing concept was applied to a very simple SST model utilizing a fixed depth mixed layer, "copper plate" ocean model. This simplest model of the ocean mixed layer assumes a completely mixed, horizontally homogeneous layer of depth h and temperature T overlying a seasonal thermocline with prescribed temperature structure. If the effects of salinity and horizontal advection are ignored, the layer parameters T and h are governed by two prognostic equations of the form (Kraus and Turner, 1967; Denman, 1973):

$$\frac{d(hT)}{dt} = f_1(T, T_a, q, U, R) , \tag{1.1}$$

$$\frac{dh}{dt} = f_2(U, \ldots) . \tag{1.2}$$

Equation (1.1) is a heat balance equation, where f_1 denotes the sum of: (1) the total flux of latent and sensible heat through the air-sea interface (dependent on the air temperature Ta, humidity q and wind speed U via standard bulk aerodynamic formulas); (2) the short and long wave radiation R through the sea surface; and (3) the heat exchange with the layer below. Equation (1.2) is an empirical relation for the entrainment of fluid from below the mixed layer and is a function of wind speed and other variables.

Frankignoul and Hasselmann (1977) simplify the model further by retaining only equation (1.1), so that the mixed layer depth remains constant. Furthermore, only the contribution (1) to f_1 is included. Heat exchange with the atmosphere is accomplished by assuming that the air-sea temperature difference is proportional to the north-south velocity V, via the relation (Ta-T)=KV, where K=constant. Therefore, a wind from the north will cool the ocean, and conversely a wind from the south will warm the ocean. Given the above parameterizations, the rate of change of SST is,

$$\frac{dT}{dt} = C_H(1+B)\frac{\rho^a C_p^a K V|U|}{\rho^w C_p^w h} \quad . \tag{1.3}$$

The following values were assigned to the parameters of equation (1.3):

$K=0.25$ (°C/m/s)

$C_H=10^{-3}$

$B=3.0$

$\rho^a=1.25 \times 10^{-3}$ gm cm^{-3}

$\rho^w=1.0$ gm cm^{-3}

$C_p^a=0.24$ cal gm^{-1}(°C)$^{-1}$

$C_p^w=0.96$ cal gm^{-1}(°C)$^{-1}$

$h=25$ m

A mixed layer depth of 25 m was selected as representative of lower mid-latitudes (30°N) on the basis of a comparison of the observed SST cycle with predictions by a "copper plate" model, (a value of h=100 m is used for calculations at station India 59°N, 19°W). With K=0.25, the root mean square (rms) air-sea temperature difference generated by the model was 1.25 C (Frankignoul and Hasselmann, 1977).

The atmospheric model (Frankignoul and Hasselmann, 1977) is a barotropic model[2] of quasi-geostrophic turbulence developed by Kruse (1975) for numerical investigations of nonlinear transfer within large scale atmospheric motions in an equilibrium state. It provides a statistically stationary wind field from which heat and momentum fluxes are inferred using bulk aerodynamic formulae. In this model there is no direct feedback from the ocean to the atmosphere.

Kruse's model represents a non-divergent, one layer atmosphere on a β plane bounded by two latitudes, and it is zonally cyclic. The dependent variable is the stream function which obeys the equation,

$$\frac{d}{dt}\nabla^2\psi + J(\psi,\nabla^2\psi) + \beta\frac{\partial\psi}{\partial x} = \text{forcing} + \mu\nabla^4\psi - r\nabla^2\psi, \qquad (1.4)$$

where

$$\frac{d}{dt} = \frac{\partial}{\partial t} + U_0\frac{\partial}{\partial x} . \qquad (1.5)$$

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[2]Similar models are discussed by Lilly (1972) and Rhines (1975).

In equations (1.4) and (1.5), x and y are eastward and northward cartesian coordinates, U_0 is a constant velocity of 5 m/s and J denotes the Jacobian operator (Frankignoul and Hasselmann, 1977). The boundary conditions are:

$$v = \frac{\partial \psi}{\partial x} = 0 \quad \text{at } y = 0, L_y \quad , \tag{1.6}$$

and

$$\psi(x, y, t) = \psi(x \pm pL_x, y, t) \qquad p = 1, 2, \ldots \tag{1.7}$$

where, Lx=Ly=8800 km. Despite the high degree of simplification, the atmospheric model provided a statistically stationary wind field.

Frankignoul and Hasselmann (1977) constructed time series of the stochastic atmospheric forcing from equation (1.3) at each gridpoint and calculated the SST changes via integration. To draw attention to the evolution of low frequencies in the SST fluctuations, the SST time series in Figure 1.1 have been low-pass filtered. Despite the simplicity of the coupled system, the white noise spectrum of atmospheric forcing results in a red-shifted SST response. Without a stabilizing negative feedback, the total SST variance is shown to grow indefinitely with time. With a properly chosen negative feedback, the response becomes asymptotically stationary.

The purpose of the present research is to explore the role of seasonality in the ocean response to stochastic forcing. It is hypothesized that the seasonal variability

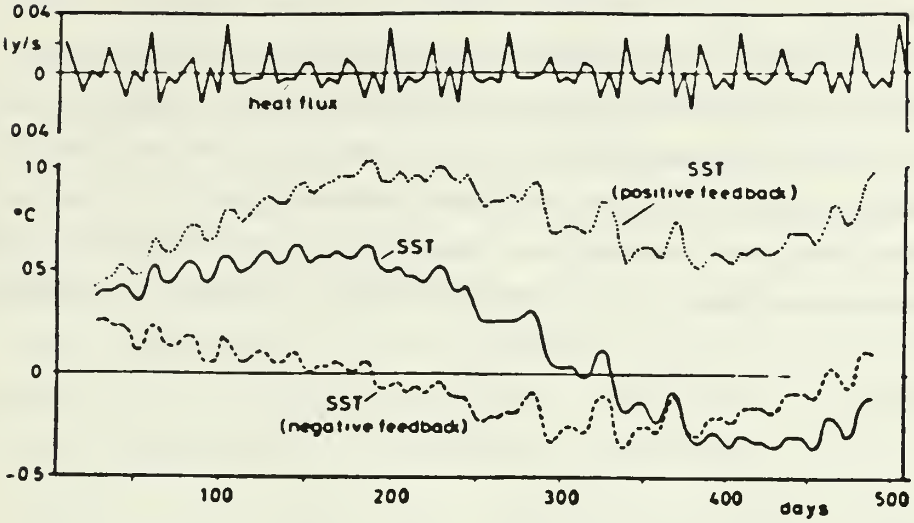

Figure 1.1 Simulated SST variations (without and with feedback effect) and total heat flux at x=0, y=Ly/4. The heat flux time series is subsampled at 5-day intervals and the SST time series are low-passed using a quadratic Lanczos filter (cut-off frequency 8 x 10^{-1} Hz) (From Frankignoul and Hasselmann, 1977).

in the surface heat flux (Q_o), as modified by a random forcing component (Q'), will cause seasonal variability in the mean mixed layer depth ($\overline{h}$) and in the variance of mixed layer depth ($\overline{h'^2}$). Variability in h will in turn modify the mean temperature ($\overline{T}$) and the temperature variance ($\overline{T'^2}$). This study is part of an investigation of increasing complexity:

1) To a constant wind stress, impose a seasonal heating/ cooling (Q_o) cycle plus a stochastic heat flux component (Q');

2) Retain the seasonal heating/cooling (Q_o) cycle minus the stochastic component, but add a stochastic component to the wind stress; and

3) Consider the combined effect of an annual heat flux cycle with both a stochastic wind stress and heat flux component. This step represents the most realistic situation, but care must be exercised to insure the proper correlation between the wind stress and surface heat flux.

This paper examines the ocean response to random atmospheric forcing as described in investigations (1) and (2) above.

II. THE STOCHASTIC MODEL

Frankignoul and Hasselmann considered a simple climatic system consisting of a slab oceanic mixed layer model which was driven by fluxes of heat and momentum across the air-sea interface. The governing equation (1.3) represents a linear approach to the study of climate variability due to stochastic forcing. The stochastic model now described is not only nonlinear in its approach, but it is designed to simulate mixed layer evolution on daily to annual time scales in response to atmospheric forcing.

A. OCEAN MIXED LAYER MODEL INCLUDING SEASONAL THERMOCLINE

Structure

A potentially significant improvement over previous studies of stochastic forcing lies in the added complexity of the ocean mixed layer model. The latest version of the Garwood (1977) one-dimensional bulk model of the mixed layer of the ocean was chosen for several reasons. First, the model was readily available and required only minor modifications for the task. Second, the model is thoroughly tested[3] and reproduces mid-latitude boundary layer dynamics. Third, the model produces solutions of the turbulent kinetic energy components and for the terms of the turbulent kinetic energy budget. Future studies will more directly study the details of stochastic forcing on the turbulence components.

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[3]It is realized that there are no exhaustive tests of a model, but a very successful study of Ocean Station Papa simulations performed in 1981 supports the dynamic consistency of the model (Garwood and Adamec, 1981).

The model explicitly parameterizes a number of mixed-layer processes ignored by other models. The ratio of the mixed layer depth to the Obukhov length scale determines the fraction of wind-generated turbulent kinetic energy available for mixing. The viscous dissipation is dependent on a local boundary layer Rossby number. The model uses separate equations for vertical and horizontal turbulent kinetic energy (see Garwood (1977) and Garwood and Yun (1979) for further details on the mixed layer model).

The Garwood model is used here to simulate the evolution of the ocean surface layer on daily to annual time scales in response to atmospheric forcing. The horizontal gradients and horizontal advection are neglected in this one-dimensional model, and the version used in this project neglects vertical diffusion within the thermocline. In addition, the salinity is held constant over the water column. The equation of state is reduced to an approximate proportionality between temperature and buoyancy.

In the absence of a vertical advection velocity, the two governing equations for the model become:

$$\frac{dT}{dt} = \frac{Q_o + Q'}{\rho_o C_p^{w} h} - \frac{\Delta T e}{h} , \qquad (2.1)$$

and

$$\frac{dh}{dt} = e , \qquad (2.2)$$

where:

T-mixed layer temperature

Q_0-mean annual surface heat flux (positive downward)

Q'-random component of the annual surface heat flux

ΔT-temperature jump across the thermocline

ρw-density of water

C_p^w-specific heat at constant pressure of water

h-mixed layer depth

e-entrainment

Q_0 is prescribed according to:

$$Q_0 = A \sin\left(\frac{2\pi n}{360}\right) \qquad n = 1, \ldots, 360 \quad , \tag{2.3}$$

Where the amplitude A is equal to 150 Watts/m^2. Q' is generated by means of a uniform random number generator and has a maximum potential amplitude equal to A. Therefore, the possible daily fluctuation in Q' is from plus to minus A.

The mixed layer temperature in this model changes as a result of the applied surface heat flux and entrainment mixing of colder water from below. Entrainment mixing of colder water from below the layer raises the potential energy of the water column. This energy for mixing is supplied by the mixed layer turbulence. Bulk layer models such as this one determine the rate of entrainment (e) from the turbulent kinetic energy budget. In the Garwood model, the time scale for viscous dissipation of turbulent kinetic energy depends on the intensity of the turbulence, the layer depth and on the planetary rotation. The rate of entrainment depends on the vertical turbulent kinetic energy available and the temperature jump ΔT at the base of the mixed layer. Entrainment becomes zero when the vertical turbulent kinetic energy approaches zero. Thus the equation for the rate of entrainment (e) becomes:

$$e = \frac{m_1 \overline{\langle w'w' \rangle}^{\frac{1}{2}} \langle \overline{E} \rangle}{\alpha g h \Delta T} \tag{2.4}$$

here

w'=vertical turbulent velocity.

$\langle \overline{E} \rangle$=mean mixed layer turbulent kinetic energy.

α=thermal expansion coefficient.

g=gravitational acceleration.

m =dimensionless constant.

The mean mixed layer turbulent kinetic energy $\langle \overline{E} \rangle$ is a function of both the surface wind stress and the surface heat flux. The relative distribution of the turbulent components is particularly sensitive to the sign and magnitude of the surface buoyancy flux. Positive buoyancy flux produces $\langle \overline{w'w'} \rangle$ directly and increases vertical turbulent kinetic energy and entrainment. Negative buoyancy flux inhibits vertical mixing. Entrainment and mixing of water from below occurs only when there is sufficient turbulent kinetic energy available in the mixed layer. In this case there is active entrainment and mixed layer evolution is given by the prognostic equation (2.2).

Under conditions of insufficient turbulent kinetic energy for mixing, the mixed layer retreats immediately to a depth hr. This retreat depth is determined by assuming steady state conditions for the turbulent kinetic energy budget and letting $\langle \overline{w'w'} \rangle$ approach zero. The retreat depth hr is given by:

$$hr = 2\rho_0 C_p^w \frac{m_1 U_*^3 - m_1 \langle \overline{E} \rangle^{\frac{3}{2}}}{\alpha g Q} , \qquad (2.5)$$

where

U* is the friction velocity at the water surface.

m_1 and m_3 are dimensionless model constants.

With the prescribed surface heat flux $(Q=Q_0+Q')$ and the friction velocity U*, the model calculates $\langle \overline{E} \rangle$ and $\langle \overline{w'w'} \rangle$ and the mixed layer temperature and depth from (2.1) to (2.5).

1. Reduction to "Copper Plate" Form

The Garwood (1977) model is easily converted to a simple slab or "copper plate" model as used by Frankignoul and Hasselmann (1977). This is accomplished by setting the mixed layer depth (h) equal to some constant value. The result is that equation (2.2) becomes:

$$\frac{dh}{dt} = 0 \quad . \qquad (2.6)$$

Specifically, this implies infinite thermal conductivity (strong mixing) between Z=-h and Z=0, but zero conductivity (no mixing) below this. Given this condition, equation (2.1) reduces to:

$$\frac{dT}{dt} = \frac{Q_0 + Q'}{\rho_0 C_p^w h} \quad . \qquad (2.7)$$

Equation (2.7) is effectively the form of equation (1.3) that Frankignoul and Hasselmann used.

The primary difference between (2.7) and (1.3) lies in the determination of the heat transfer between the ocean and atmosphere. Frankignoul and Hasselmann (1977) parameterize the heat transfer as a function of the wind direction, while Garwood (1977) specifies the surface heat flux directly. However, both (2.7) and (1.3) will produce equivalent stochastic forcing. The advantage of (2.7) is that the rate of change of SST is now dependent upon both the annual cycle and the random forcing, and facilitates later application with a predictive equation for mixed layer depth.

Reducing Garwood's model to a simple slab model permitted the recreation of the results of Frankignoul and Hasselmann (1977) for the case without negative feedback. These results will be discussed further in Chapter III.

B. ATMOSPHERIC FORCING INCLUDING AN ANNUAL CYCLE

The ocean model is driven by an annual period sinusoidal surface heat flux (Q_o), a stochastically varying heat flux (Q') and a constant wind stress. For computational convenience a model year is taken to be 360 days. The model is initialized at "spring transition" corresponding approximately to the vernal equinox with a mixed layer depth of 100 meters, a mixed layer temperature of 6 C and an underlying homogeneous water column of 5 C. The magnitude of the annual period surface heat flux was chosen so that the surface temperature would rise approximately 8 C and the seasonal thermocline would shallow to approximately 50 meters at the times of maximum SST and maximum downward surface heat flux respectively. The necessary amplitude of the annual heat flux to achieve this was 150 Watts/m^2 in the case of a 1 dyne/cm^2 wind stress. The salinity is specified as a constant 33 ‰ and it has no dynamical significance in this study. The initialization described approximates conditions at Ocean Station Papa at 50°N, 145°W. To allow the model to reach a statistical equilibrium, the model is actually integrated to 720 days, and the second model year is used for calculations and graphic display.

The stochastic forcing is supplied by a random number generator. A known but different seed value is specified for each case, which allows duplication of the random forcing series. The stochastic forcing is applied to the surface heat flux such that the random forcing can on any day equal (positively or negatively) the amplitude of the annual surface heat flux cycle.

C. CALCULATION OF STATISTICAL VARIABLES

At this point it becomes necessary to define the nomenclature to be used in discussions of the model results. Model outputs are sea-surface temperature, mixed layer depth, the average temperature variance $\overline{T'^2}$ and average mixed layer depth variance $\overline{h'^2}$. The variables T_o and h_o represent the ocean response in temperature and mixed layer depth to the annual period sinusoidal heat flux cycle without the stochastic component. The differences $T-T_o$ and $h-h_o$, between the mean surface temperature or mixed layer depth with random forcing and the mean surface temperature or mixed layer depth without random forcing provide insight into the effects of the stochastic forcing theory.

An overbar denotes an ensemble mean of the daily values of the variable divided by the number of cases summed. Thus, the output statistics represent daily averages over some climatological time frame. The variance of temperature and mixed layer depth can be computed from the following equations obtained after performing a Reynolds decomposition.

$$\overline{T'^2} = (\overline{T^2}) - (\overline{T})^2 \quad , \tag{2.8}$$

and

$$\overline{h'^2} = (\overline{h^2}) - (\overline{h})^2 \quad . \tag{2.9}$$

Thus, the variances are the result of the difference between mean quantities.

III. THE STOCHASTIC MODEL EXPERIMENTS

A. PROCEDURES

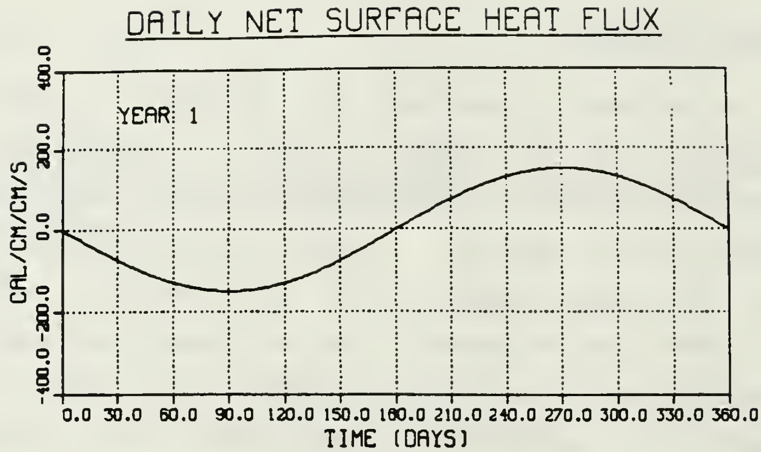
1. General Method

The stochastic forcing is applied to 101 cases, which corresponds to 101 model years. The first case or model year represents the nonstochastic case that determines the daily value of T_0 and h_0. The next 100 cases contain a daily random forcing component to the annual surface heat flux cycle. From these cases, the averages and variances of the temperatures and mixed layer depths are calculated. The average temperature and mixed layer depth are computed daily via a simple summation over the 100 stochastically forced cases. For the variance computation see equations (2.8) and (2.9).

2. Reproducing Frankignoul and Hasselmann's Result

Frankignoul and Hasselmann (1977) applied the stochastic forcing concept to a simple slab or "copper plate" ocean model as given by equation (1.3). The results obtained from the Garwood model as in equation (2.7) confirm the hypothesis of Hasselmann (1976) that the ocean acts as an integrator of the white noise atmospheric forcing and produces a red-shifted response spectrum, that is the spectral response of the model is shifted toward the lower frequencies, and appears geophysical in nature (Figure 1.1). Frankignoul and Hasselmann further showed that, without a stabilizing negative feedback, the total SST variance grows indefinitely with time, as suggested by Hasselmann (1976).

Figure 3.1 displays the daily model response to the sinusoidal annual heat flux cycle without stochastic



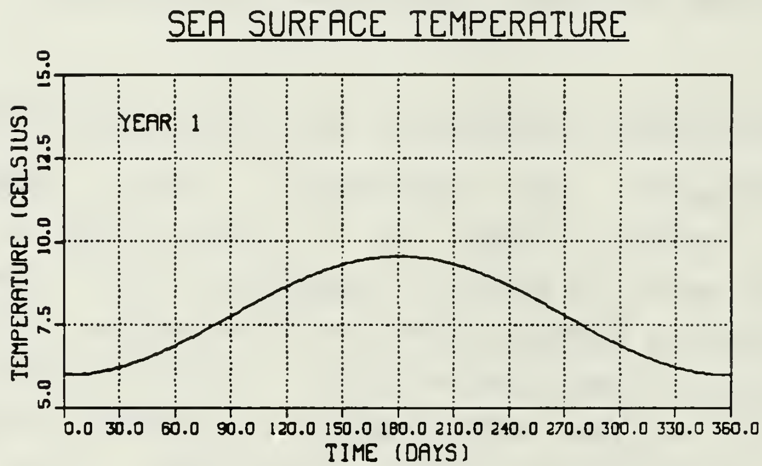


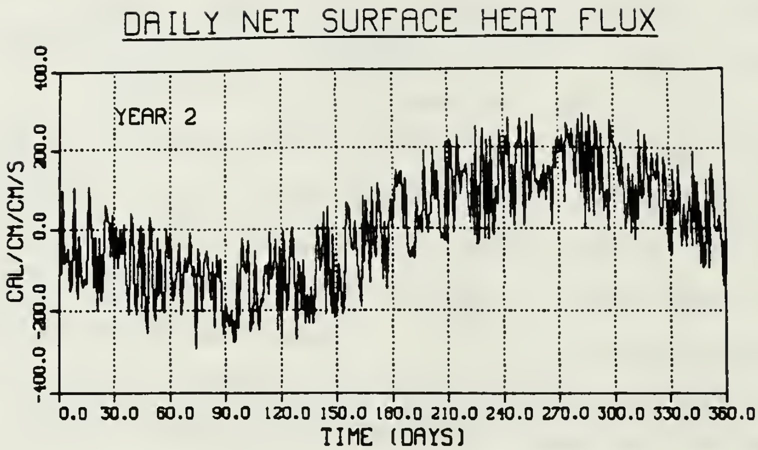
Figure 3.1 Response to annual cycle. a) The daily net surface heat flux (positive upwards). b) T_o, the ocean response to the annual period surface heat flux without stochastic forcing; h_o equals 100 m.

forcing. It can be seen that the annual surface heat flux cycle begins at vernal equinox, and adds heat to the mixed layer with the maximum heating rate occurring at 90 days. Heat continues to be added to the mixed layer for an additional 90 days, and at day 180 the cycle reverses and heat is withdrawn from the ocean, with the maximum withdrawal rate on day 270. Heat continues to be removed from the ocean until the cycle is complete on day 360.

Although the one-dimensional model is designed so that the mixed layer temperature will rise approximately 8 C with realistic mixed layer depths, it is evident that fixing the mixed layer depth reduces the ocean temperature response to only approximately 3.5 C. Since the mixed layer depth is constant there is no interaction between the minimum depth at the time of maximum surface heat flux. Instead, the heat input is spread over the entire 100 m, and the temperature increase is only about half of the expected value.

The effects of the stochastic forcing for a typical case can be seen in Figure 3.2. The SST would statistically be expected to be larger than T_o one half of the time and smaller than T_o the other half. In this case the SST is slightly smaller than T_o.

Figure 3.3 shows the significant temperature statistics calculated from the model. Hasselmann (1976) and Frankignoul and Hasselmann (1977) showed that the SST variance in the absence of negative feedback will grow indefinitely. The upper graph of Figure 3.3 confirms this result. The variance of the temperature perturbation can clearly be seen to exhibit a ramp type growth over the period of the experiment. Also, there is no statistically significant difference between the average temperature with random forcing and the average temperature response to the annual cycle alone, $T-T_o$. If the experiment was repeated an infinite number of times, the values of $T-T_o$ would be zero.



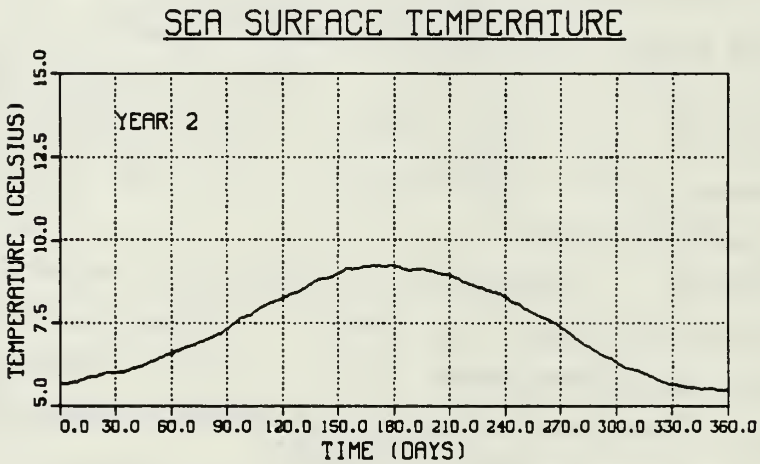
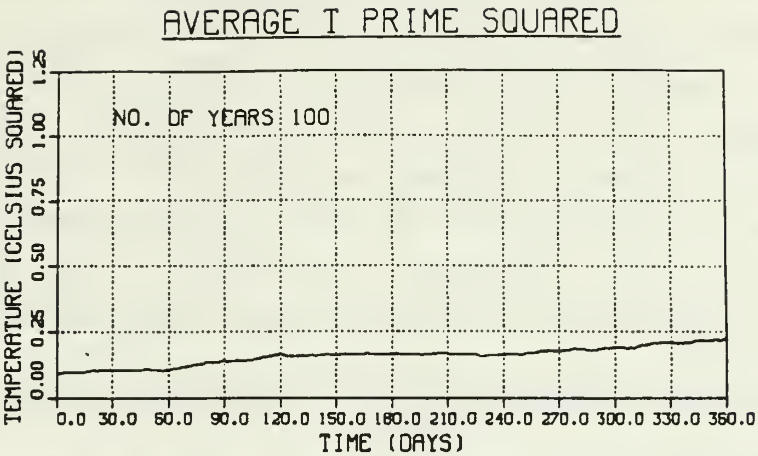


Figure 3.2 Typical year with stochastic forcing. a) The same annual heat flux cycle shown in Figure 3.1, but with a random daily heat flux component Q' added. b) The SST over the model year. The mixed layer depth h=100 m.



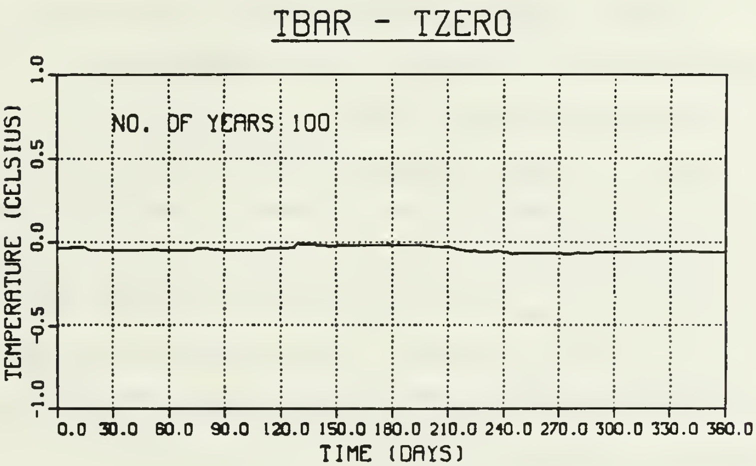


Figure 3.3 Statistical temperature variables. a) The mean temperature variance of 100 stochastically forced cases. b) Plot of $T-T_0$.

This result only holds for the case where the mixed layer depth is fixed, which is a physically unrealistic situation.

From the above results, it can be said that the modified Garwood model in slab form correctly duplicates the linear time rate of change of SST developed by Frankignoul and Hasselmann (1977). Because the Garwood model is not linear and the forcing is completely specified, it is not possible to compute an adequate negative feedback term, without first determining the climatology, to confirm further studies made by Frankignoul and Hasselmann. In the next section, we explore the role of seasonality in the ocean response to stochastic forcing using the one-dimensional model with entrainment into the mixed layer.

3. Stochastic Forcing with Mixed Layer Physics

For this experiment, the governing equations are (2.1), (2.2), and the rate of entrainment is given by equation (2.4). This study presents a significant departure from the slab models used by Hasselmann (1976) and Frankignoul and Hasselmann (1977), because the mixed layer temperature is no longer independent of the mixed layer depth. Given a constant wind stress, the mixed layer depth will shallow during periods of strong downward surface heat flux, which also causes an increase in mixed layer temperature. Thus on an annual basis, the effects of the heat flux and shallow mixed layer depth will reinforce each other and generate larger perturbations in temperature. The atmospheric forcing and model initializations will remain as stated for the previous study.

Figure 3.4 displays the model response to the annual period atmospheric surface heat flux without a stochastic forcing component. As stated earlier the atmospheric forcing amplitude is designed to allow the SST to increase by approximately 8 C and the mixed layer depth to shallow to

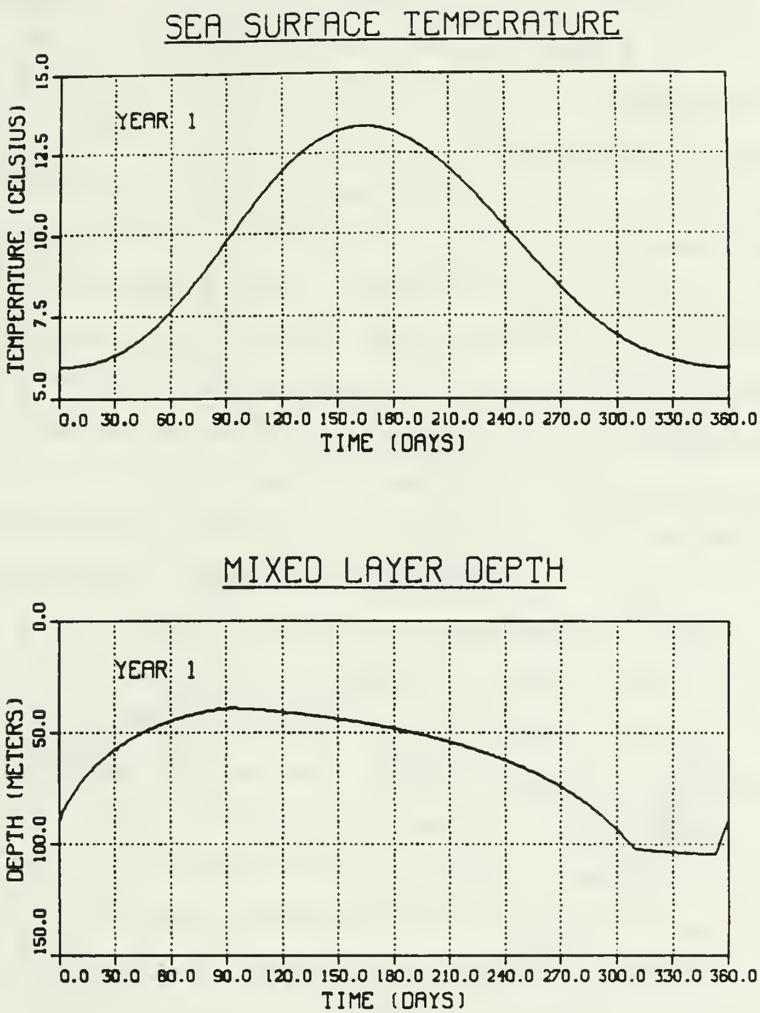


Figure 3.4 Annual cycle without stochastic forcing. a) The daily SST response to the annual cycle. b) the daily mixed layer depth response to the annual cycle.

approximately 50 m. These conditions are met in this version of the model.

The SST reaches a maximum at approximately day 170 which indicates a 80 day lag from the point of maximum downward heat flux. As the second model year of each case is used for calculations, the mixed layer depth (Figure 3.4) at vernal equinox is slightly shallower than the 100 m used to initialize the model. The depth continues to shallow and reaches minimum values at about day 100. For the remainder of the model year, the mixed layer depth gradually deepens until approximately day 350. From day 310 to day 350, the mixed layer depth has achieved the greatest depth that turbulent mixing will allow, and it is entraining the top of the permanent thermocline. During the last 10 days, the mixed layer begins to shallow toward the initial value.

Upon comparing Figures 3.4 and 3.1, it is apparent that the maximum SST is greater when the mixed layer depth is allowed to vary. As the atmosphere adds heat to the ocean, the mixed layer becomes more stable, and the potential energy of the system increases. Increasing the potential energy of the mixed layer is at the expense of the turbulent kinetic energy, which causes the mixed layer depth to retreat according to equation (2.5). The initial shallowing of the mixed layer corresponds to the period of reduced surface cooling in late winter and early spring when entrainment ceases because of the reduced buoyancy production of turbulent kinetic energy and a relative increase in the dissipation of turbulent kinetic energy. Equation (2.1) can be reduced, for this case with forced heat fluxes, to a ratio of the atmospheric heat flux and the mixed layer depth. As the time rate of change in temperature is inversely proportional to the mixed layer depth, the shallowing of the mixed layer coupled with the downward heat flux produces a rapid increase in temperature with time.

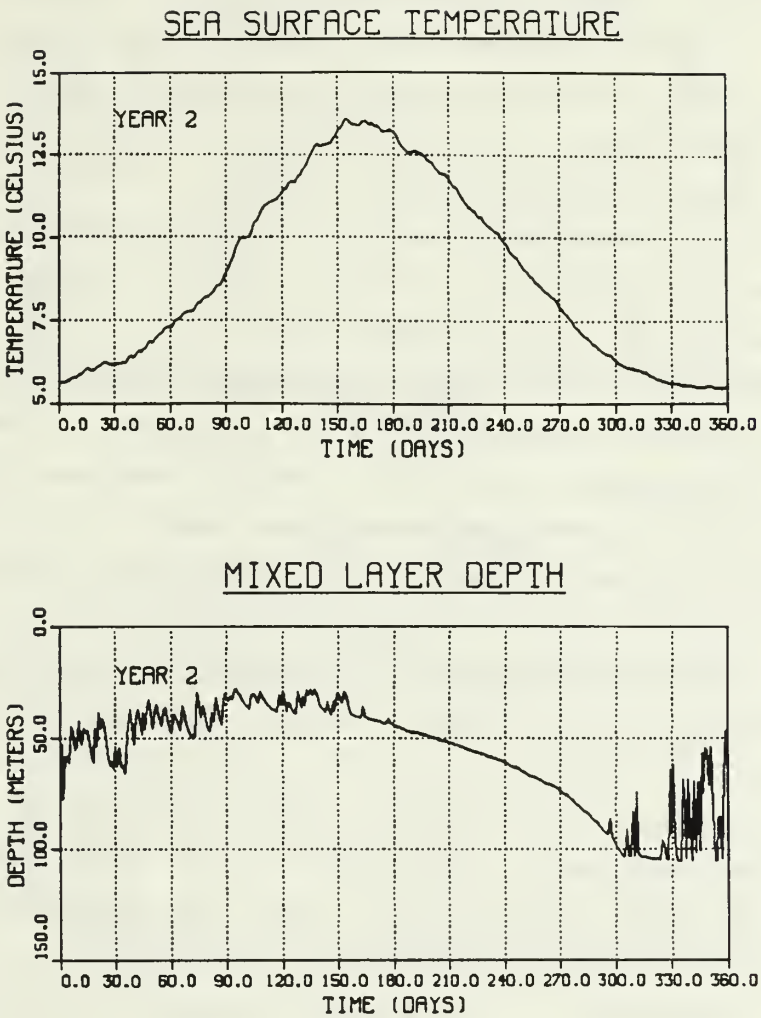


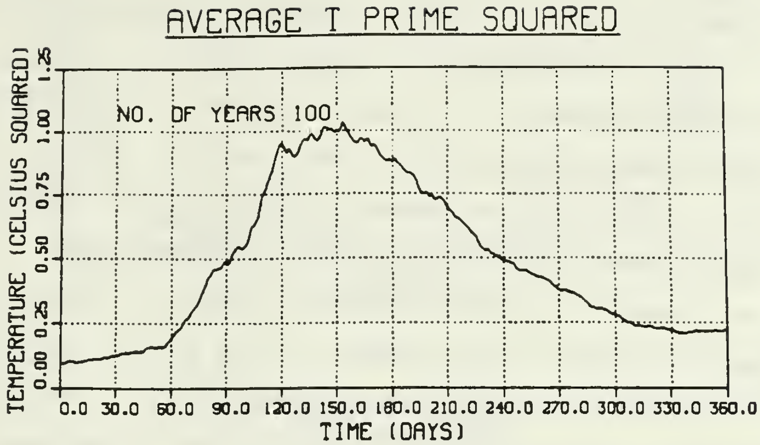
Figure 3.5 Typical year with stochastic forcing-full model. a) The SST over the model year. b) The mixed layer depth over the model year.

Figure 3.5 is for a typical model year under the influence of both seasonal and stochastic forcing. The daily random heat flux component Q' is identical to the forcing in Figure 3.2. In all of the graphs the annual cycle is evident. The first significant result can be seen in the graph of mixed layer depth for a typical year in Figure 3.5. It is evident that the daily variations in mixed layer depth can be as great as 50 m during portions of the year. Therefore, a linear model of stochastic forcing with fixed layer depth is unrealistic. The temperature perturbation is dependent on the interaction between the annual plus stochastically forced heat fluxes and the perturbations in the mixed layer depth.

Figure 3.6 shows the temperature statistics calculated from the 100 cases in the experiment. There is a phase shift associated with the time of maximum SST, with the maximum SST occurring about two weeks later for the stochastically forced cases. The difference between this result and that obtained for the slab model, can be attributed to the combined effects of a changing mixed layer depth and the evolving SST. This curve does retain a strong annual cycle.

The temperature variance curve of Figure 3.6 retains the increased amplitude throughout the year observed in Figure 3.3 for the slab model, but the influence of the annual cycle is clearly evident and produces a 1 C maximum temperature variance. The growth visible in the variance graph is a result of the lack of any negative feedback applied to the model.

To further delineate the temperature differences, the graph of $T-T_o$ reveals that the stochastically forced average temperature is approximately 0.75 C higher than the temperature derived from the annual cycle alone. This result is significant. Both curves have a strong annual



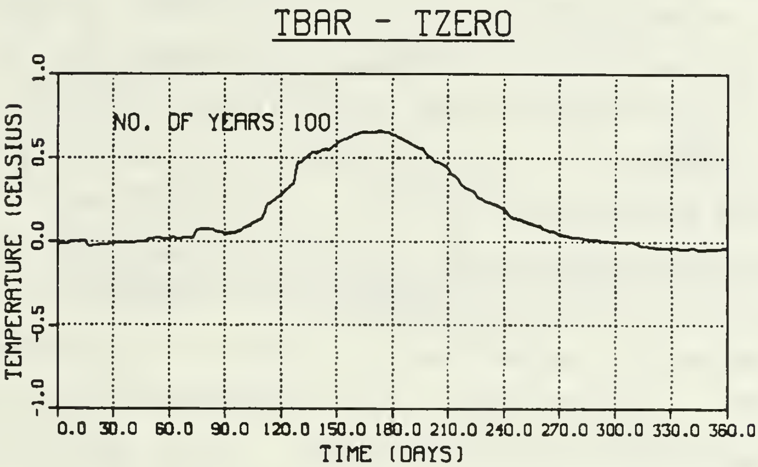
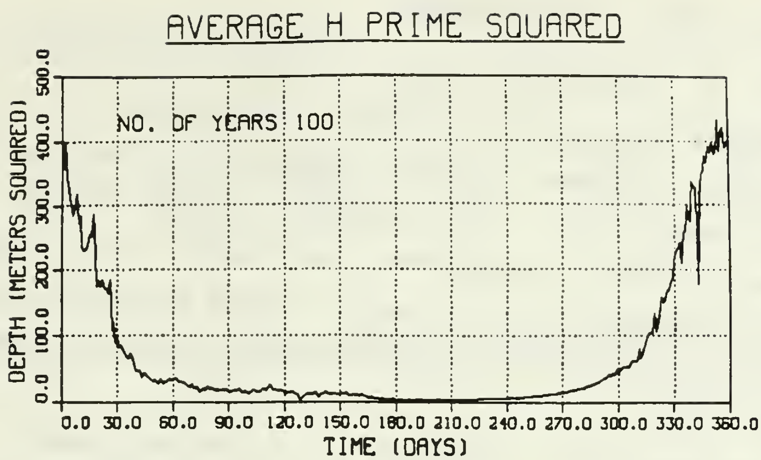


Figure 3.6 Temperature statistics-full model. a) The variance of the temperature perturbation. b) The difference between the mean temperature resulting from the stochastically forced cases and the ocean response to the annual cycle T_o.

cycle, which indicates that the seasonal variations cannot be ignored when studying SST perturbations.

Figure 3.7 shows the mixed layer depth statistics analogous to those computed for temperature. It is apparent that the average mixed layer depth with stochastic forcing is smaller than the mixed layer depth produced by the annual cycle alone. A considerable phase shift in the minimum mixed layer depth occurs. The stochastically forced depth minimum occurs approximately one month later than the minimum in h_o. It is this phase difference which produces a corresponding phase shift in the temperature statistic, $T-T_o$. As a result of this phase shift a second harmonic becomes evident in the $h-h_o$ curve. The final effect of note pertains to the mixed layer depth variance. The mixed layer exhibits greater fluctuations during the shallowing regime than the deepening regime. This phenomenon can be explained through an examination of the turbulent kinetic energy budget. In the spring the near-zero heat flux Q_o, allows the mixed layer depth to be controlled by Q', the random heat flux component. At this time, h' will show a large variation in proportion to Q'. In addition, there is a small temperature jump (ΔT) across the seasonal thermocline and, as can be seen in equations (2.2) and (2.4), the time rate of change of mixed layer depth is inversely proportional to ΔT. In the fall, the temperature jump across the seasonal thermocline is very large. This greatly restricts mixed layer depth fluctuations. Thus the shallowing regime has a greater mixed layer depth variance than does the deepening regime. The curve of $h-h_o$ is always negative, except for a random phenomenon occurring around day 20, which indicates that the stochastically forced mixed layer is on average 20 m shallower than the mixed layer produced solely by the annual cycle.



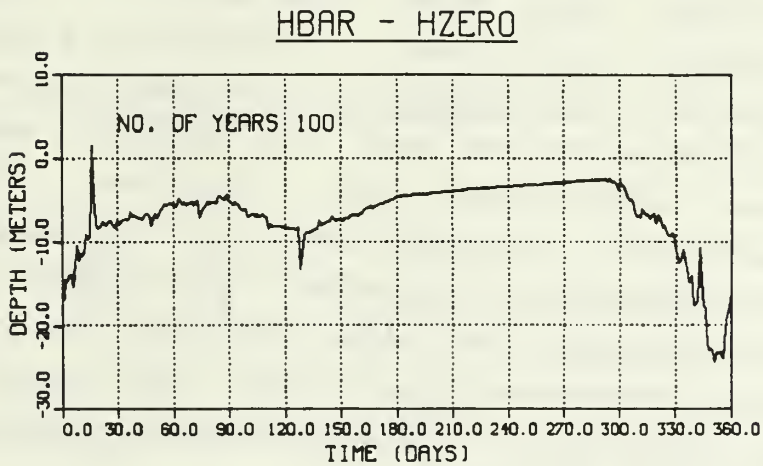


Figure 3.7 Mixed layer depth statistics-full model. a) The variance in the mixed layer depth over the experiment. b) The difference between the average mixed layer depth when randomly forced and the mixed layer depth which is the product of the annual cycle alone, $h-h_o$.

B. INTERPRETATIONS

1. "Slab Model" Experiment

With a constant depth mixed layer the results obtained confirm those of previous researchers. The ocean does respond to white noise atmospheric forcing with a "red-shifted" spectrum. The annual cycle affects only the average temperature, not the variance in temperature. However, the temperature variance does display the indefinite growth predicted by Hasselmann, since there is no negative feedback.

The assumptions used in this experiment produce a linear equation for the change in T'. There are no temperature-mixed layer depth interactions. This simple model eliminates the possible contribution of the applied annual period heat flux forcing. The results would be the same if the random surface heat flux component were applied to a constant surface heat flux. Therefore, the initial model assumptions of Hasselmann and Frankignoul and Hasselmann oversimplify the problem. To produce more realistic results requires the added complexity associated with the nonlinear dependence between mixed layer depth and SST.

For simplicity, assume that the time rate of change of SST is given as follows:

$$\frac{dT}{dt} = \frac{Q}{h} \quad . \tag{3.1}$$

By means of a Reynolds decomposition it can be shown that it is possible to linearize equation (3.1) only if $|h'| \ll \overline{h}$. Expanding equation (3.1) to include the average and perturbation terms results in the following expression

$$\frac{\partial \overline{T}}{\partial t} + \frac{\partial T'}{\partial t} = \frac{\overline{Q} + Q'}{\overline{h} + h'} = \frac{\overline{Q} + Q'}{\overline{h}\,(1 + h'/\overline{h})} \quad . \tag{3.2}$$

A binomial expansion gives

$$\left(1 + \frac{h'}{\overline{h}}\right)^{-1} = 1 - \frac{h'}{\overline{h}} + \left(\frac{h'}{\overline{h}}\right)^{2} - \cdots \quad \approx 1 - \frac{h'}{\overline{h}} \quad . \tag{3.3}$$

Therefore the approximate form of equation (3.1) becomes

$$\frac{\partial \overline{T}}{\partial t} + \frac{\partial T'}{\partial t} = \frac{\overline{Q} + Q'}{\overline{h}} \left(1 - \frac{h'}{\overline{h}}\right) \quad . \tag{3.4}$$

If $|h'| << \overline{h}$, equation (3.4) reduces to

$$\frac{\partial \overline{T}}{\partial t} + \frac{\partial T'}{\partial t} = \frac{\overline{Q}}{\overline{h}} + \frac{Q'}{\overline{h}} \quad : \tag{3.5}$$

Averaging each term produces

$$\overline{\frac{\partial \overline{T}}{\partial t}} + \underbrace{\cancel{\overline{\frac{\partial T'}{\partial t}}}}_{0} = \overline{\frac{\overline{Q}}{\overline{h}}} + \underbrace{\cancel{\overline{\frac{Q'}{\overline{h}}}}}_{0} \quad . \tag{3.6}$$

Subtracting (3.6) from (3.5) results in the following linear equation for the change in T'

This equation is the functional form used by Hasselmann (1976). It is evident that the temperature perturbation

$$\frac{\partial T'}{\partial t} = \frac{Q'}{\bar{h}} \quad . \tag{3.7}$$

will be the integral of the right hand side. Thus the ocean acts as an integrator of the random forcing. It is equally evident that the annual heat flux cycle plays no part in the determination of the temperature perturbation.

2. <u>One-Dimensional Model Experiment</u>

Suppose $|h'|$ is not much less then $\bar{h}$. Expanding equation (3.4) gives

$$\frac{\partial \bar{T}}{\partial t} + \frac{\partial T'}{\partial t} = \frac{\bar{Q}}{\bar{h}} - \frac{\bar{Q}h'}{\bar{h}^2} + \frac{Q'}{\bar{h}} - \frac{Q'h'}{\bar{h}^2} \quad . \tag{3.8}$$

Averaging each term produces

$$\overline{\frac{\partial \bar{T}}{\partial t}} + \cancelto{0}{\overline{\frac{\partial T'}{\partial t}}} = \frac{\bar{Q}}{\bar{h}} - \cancelto{0}{\frac{\overline{\bar{Q}h'}}{\bar{h}^2}} + \cancelto{0}{\frac{\overline{Q'}}{\bar{h}}} - \frac{\overline{Q'h'}}{\bar{h}^2} \tag{3.9}$$

Subtracting (3.9) from (3.8) gives

$$\frac{\partial T'}{\partial t} = \frac{Q'}{\bar{h}} - \frac{\bar{Q}h'}{\bar{h}^2} - \frac{Q'h'}{\bar{h}^2} - \frac{\overline{Q'h'}}{\bar{h}^2} \quad . \tag{3.10}$$

Here the time rate of change of T' depends on the correlation between the random forcing component and the perturbation mixed layer depth and on the relationship between the average heat flux and the mixed layer depth perturbation. If, as in the second experiment, $|h'|$ is not much smaller than $\bar{h}$, a linear relationship for T' is not possible.

The results of these two experiments clearly demonstrate the complex relationships existing in a single phase of ocean response. The linear approach to stochastic forcing does bring out some basic responses, but ignores some very significant interactions. The ocean does respond as an integrator of the random forcing and the variance in SST does grow indefinitely when no feedback is applied, but the strong influence of the annual cycle is missed when a slab model is used. In addition, the dynamics of the slab model are such that the average of the stochastically forced climatology is equivalent to the annual cycle climatology. This is not realistic. The two climatologies are not identical. The random forcing component imparts enough extra energy to the ocean mixed layer to raise the temperature by 0.75 C during the summer, and it delays the temperature maximum and depth minimum until later in the year. These effects are important.

C. STOCHASTIC WIND STRESS EXPERIMENT

The next step in the investigation of the stochastic forcing theory is to examine the model response to a wind stress composed of a mean component and a random perturbation. Previous investigations into the importance of the wind stress in the development of SST anomalies indicate this is an important forcing mechanism (Camp and Elsberry, 1978; Frankignoul and Reynolds, 1983).

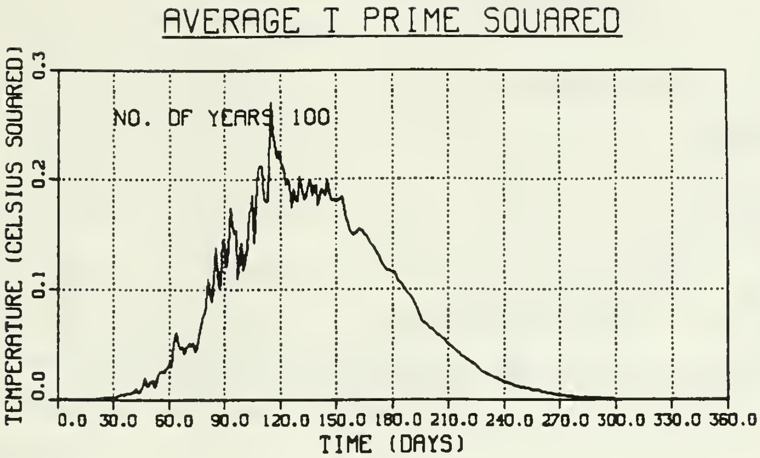
1. General Method

A modified version of the Garwood (1977) model described in the previous sections is used for this investigation. The model is modified such that there is no longer a stochastic heat flux component Q', but the annual period heat flux cycle remains. The kinematic wind stress U^{*2} is decomposed into a mean component $\overline{U^{*2}}$ and a stochastic component $(U^{*2})'$. The mean kinematic wind stress is 1.0 cm^2/s^2 and the random component is allowed a daily variation of plus or minus 0.9 cm^2/s^2. In all other respects the procedure is the same as that used in the stochastic heat flux investigations.

2. Stochastic Wind Stress with Seasonal Thermocline

This experiment isolates the model response to stochastically forced wind stress and a simple sinusoidal annual surface heat flux while retaining a variable mixed layer depth. As in previous investigations, the first case possesses no stochastic forcing to determine the base or zero state. The model is then run 100 years and the output averaged to approximate climatology. The temperature variance (Figure 3.8) does not appear to display an indefinite growth in spite of the fact that there is no feedback applied to the model. Such growth, although still present, is insignificant at the scale of the figure. Additionally, the maximum temperature variance occurs about day 112, approximately 40 days prior to the maximum average SST. The magnitude of the temperature variance is much smaller than for the stochastic heat flux experiment. The annual cycle is evident in both curves of Figure 3.8. The mean temperature response to stochastic wind forcing is smaller than the temperature response to the annual cycle alone. It can be seen that the temperature of the stochastically forced cases is about 0.25 C smaller than T_o, the temperature response to the annual cycle alone.

Figure 3.9 shows the mixed layer depth statistics for the random wind stress experiment. The magnitude of the variance of the mixed layer depth in the shallowing regime is greater than it was for the heat flux experiment (Figure 3.7). During the deepening regime, the variance in $\overline{h'^2}$



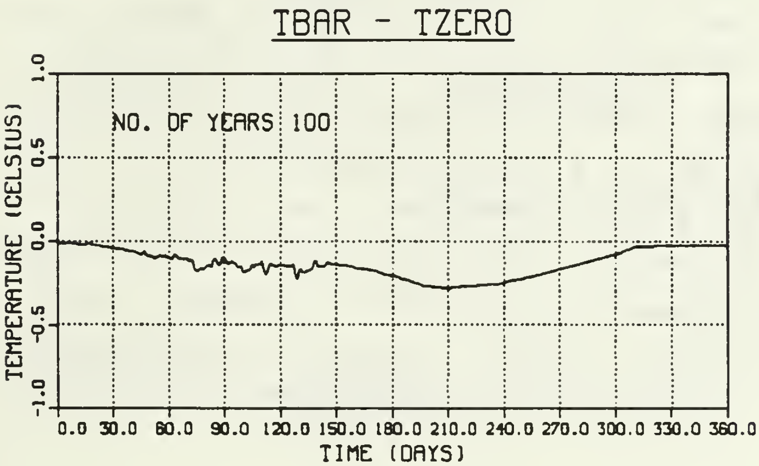


Figure 3.8 Stochastic winds-temperature statistics. a) Variance of the temperature perturbation. b) The difference between the mean temperature resulting from the stochastically forced cases and the temperature response to the annual cycle.

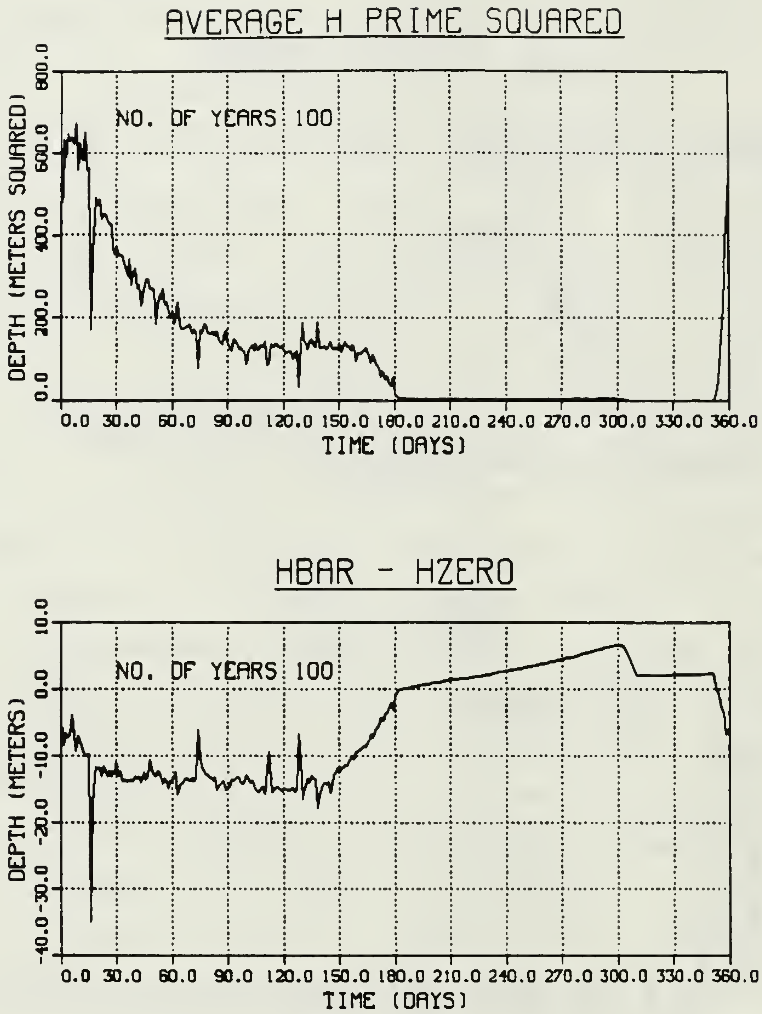


Figure 3.9 Stochastic winds-mixed layer depth statistics. a) The variance in the mixed layer depth over a 100 year climatology. b) The difference between the mixed layer depth when randomly forced and the mixed layer depth response to the annual cycle.

approaches zero. This phenomenon can be explained through an examination of the turbulent kinetic energy budget (see discussion for Figure 3.7). Examination of the lower graph of Figure 3.9 reveals that the stochastically forced mixed layer depth is about 15 m smaller than h during the shallowing regime. In comparison with Figure 3.7, it is evident that the mean mixed layer depth with stochastic wind stress forcing is about 5 m smaller than with stochastic heat flux forcing. During the deepening regime, however, the mean mixed layer with stochastic wind forcing is actually deeper than h the model response to the annual cycle. Therefore, the stochastic wind stress component causes the mixed layer to deepen more rapidly through additional generation of turbulent kinetic energy that is not completely compensated by dissipation.

In conclusion, it can be stated that the stochastic forcing concept when applied to the wind stress U^{*2} does affect the SST response of the ocean model by as much as 0.4 C. The stochastic wind stress forcing reduces SST on the whole because of enhanced mixing. The random wind stress forcing becomes an important source of turbulent kinetic energy during the deepening regime in the late summer and early winter. The mixed layer deepens more rapidly and achieves a greater depth than without the stochastic wind stress component.

IV. SUMMARY AND RECOMMENDATIONS

The Hasselmann stochastic forcing theory was verified for a slab model, and then it was modified to include a mixed layer depth model. This enabled the exploration of the interaction between stochastic and annual period forcing of the ocean. The simple linear prediction equation for T' that was developed by Frankignoul and Hasselmann was no longer applicable because of nonlinear interactions between h' and T'. The mixed layer depth perturbations are significant in comparison with the average mixed layer depth. The temperature perturbation is a function not only of the stochastic heat flux component, but also of the interaction between the annual heat flux cycle Q_o, the stochastic forcing component Q', and the mixed layer depth perturbation. Of considerable significance is the fact that with the stochastic forcing the climatological SST response is 0.75 C higher than the corresponding ocean response to the annual cycle alone. This is consistent with the climatological mixed layer depth being smaller when random forcing is applied. Furthermore, the minimum mixed layer depth occurs almost one month later in the year than the ocean response to the annual cycle alone. Therefore, the stochastic forcing component of the atmospheric heat flux greatly influences the climatology of both the SST and mixed layer depth.

In confirmation of Elsberry and Camp (1978), we show that the wind stress with a stochastic component produces a cooler (by 0.25 C) ocean surface than would be observed without the random component. This is because the mixed layer is deeper on the average due to the generation of additional turbulent kinetic energy. Interestingly, however, the climatological mixed layer depth is reduced in value during much of the cooling season.

Recommended future effort will be to apply the stochastic forcing simultaneously to the wind stress and the heat flux. This represents the most realistic situation, but care must be exercised to insure the proper correlation between the random forcing of each variable, because wind stress and heat flux are not independent. This is the first step to analyzing the effect of stochastic forcing on the individual turbulence components.

The SST response of the ocean on climatological time scales is dependent on both annual cycles and stochastic forcing components. The interactions are complex and nonlinear. Only by examining the parts is it possible to understand the whole, but care must be exercised to insure that the simplifications used do not misrepresent the actual interactions.

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