HADLEY CENTRE
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AN OCEAN GENERAL CIRCULATION MODEL OF THE INDIAN OCEAN FOR HINDCASTING STUDIES

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ABSTRACT

As part of the WMO TOGA (Tropical Oceans Global Atmosphere) programme, an Ocean General Circulation Model of the Indian Ocean has been developed at the Meteorological Office with the aim of being able to hindcast the upper-level circulation. In this paper the background to this project is outlined, including an overview of the important processes in tropical ocean dynamics and the main features of the Indian Ocean circulation. A description of the model is then given, and the model simulation of the climatological seasonal cycle currents in response to climatological wind stresses and heat fluxes is discussed. In a subsequent paper the response of the model to interannually-varying forcing fields (a further step towards developing the hindcasting capability of the model) will be discussed.

AN OCEAN GENERAL CIRCULATION MODEL OF THE INDIAN OCEAN FOR HINDCASTING STUDIES

1. INTRODUCTION

In the early 1980's, the World Climate Research Program of the World Meteorological Organisation inaugurated an international coordinated program of research, called TOGA (Tropical Oceans Global Atmosphere) to run from 1985 to 1995. It is designed to investigate the interactions between the tropical oceans and the atmosphere with a view to furthering the understanding of this coupled system and ultimately to forecast its changes (WCRP, 1985)

TOGA consists of two main streams of research: observational and modelling. The principal aim of the modelling program is to produce a realistic fully coupled operational model of the tropical oceans and global atmosphere. The achievement of this objective is necessarily a step-by-step process: each of the tropical oceans is studied separately, and a hierarchy of models is used to explore the physical processes involved.

One of the contributions being made to this research by the Meteorological Office Unit at the Hooke Institute in Oxford is the development of an Ocean General Circulation Model (OGCM) of the Indian Ocean for hindcasting the upper-level circulation. In this paper a description of the model is given and an assessment made of its ability to simulate the climatological currents in the Indian Ocean in response to climatological forcing. In a subsequent paper the next step in developing the hindcasting capability of the model, namely the use of fields derived from operational Numerical Weather Prediction model analyses to force the model, will be discussed.

Before proceeding to the description of the model in Section 3, an introduction to the subject of modelling the Indian Ocean is given in Section 2. The important aspects of tropical ocean dynamics are outlined, leading to an overview of the Indian Ocean circulation; a summary of the modelling work relevant to this study is then given. This Section provides the background for both this paper and the subsequent paper in which further results are presented.

2. BACKGROUND TO INDIAN OCEAN MODELLING

2.1 Tropical ocean dynamics

Overviews of ocean circulation in the Tropics have been provided by Hastenrath (1985) and Knox & Anderson (1985). The principal driving force for surface currents in the Tropics is the wind stress. Since the TOGA programme is concerned with tropical ocean - global atmosphere interaction, the currents which are of greatest interest are those which have the maximum impact on low latitude sea surface temperatures, namely the wind-driven surface currents of the tropical regions. The World Ocean Circulation Experiment (WOCE) will seek to further understanding of the circulation in the global ocean, both in the surface layers and at depth.

The wind drives the ocean both directly and indirectly. Because of the Coriolis force the direct forcing results in Ekman drift, by which wind driven currents flow to the right of the wind direction in the Northern Hemisphere and to the left in the Southern Hemisphere. In the Tropics, two differences with respect to circulation in higher latitudes arise. Firstly, the Coriolis parameter on the Equator is zero so that the current direction there tends to be directly down-wind. Secondly, the reversal in direction of the Ekman drift relative to the wind direction across the Equator results in divergent flow away from the Equator if the wind is Easterly, convergent if Westerly; divergence leads to upwelling on the Equator, convergence to downwelling. The zero value of the Coriolis parameter on the Equator enables equatorial currents to be spun up rapidly; for example, a zonal wind stress of 0.2 dynes/cm^2 acting for one month over a depth of 50m will produce a current speed of about 1m/s.

The wind affects currents indirectly in two ways. Firstly, the curl of the wind stress produces vertical motion in the oceanic boundary layer (so-called "Ekman pumping"); this leads to vertical displacement of the thermocline, resulting in horizontal density gradients which in turn produce geostrophic currents. Secondly, changes in the wind forcing can generate waves in the surface currents. These waves provide a key mechanism by which information can be transmitted rapidly within the tropical oceans. The equatorial region forms a wave-guide for zonally-propagating waves, enabling information to be carried across a tropical ocean far more rapidly than at mid-latitudes. Rossby waves are the principal mode for westward-propagating waves and Kelvin and Yanai (mixed Rossby-gravity) waves for eastward-propagating disturbances.

Because the applied wind stress on an ocean is a time-variant property, wind-driven currents vary correspondingly. On the inter-seasonal time-scale, the wind patterns in the tropical Pacific and Atlantic remain fairly constant and nowhere are there large-scale seasonal reversals in wind direction. However, in many parts of the Indian Ocean the wind direction does reverse seasonally (see Fig. 1). This factor (together with the distinctly different geography of the Indian Ocean in comparison to that of the Pacific and Atlantic, the Indian Ocean having a northern land boundary at sub-tropical latitudes) results in the principal currents being significantly different to those in the other two oceans, especially in the equatorial and north Indian Ocean.

2.2 Indian Ocean circulation

a) Equatorial currents

During the winter monsoon (or Northeast Monsoon), an eastward Equatorial Counter Current (ECC) lies between the two westward currents - the North Equatorial Current (NEC) and South Equatorial Current (SEC) - driven by the Trade Winds. In contrast to the Pacific and Atlantic, the ECC lies to the south of the Equator, at about 5°S; this is consistent with the applied wind field in Fig.1a.

During the summer monsoon (or Southwest Monsoon), the NEC and ECC disappear and are replaced by a single eastward current, the Southwest Monsoon Current (SMC). The boundary between the SMC and the SEC lies close to the Equator.

The transition periods between the monsoons tend to be characterized by a rather more variable current pattern; the SEC is the only main current which persists though all the seasons, though its latitudinal position varies. A strong (up to 1m/s) eastward current, the Wyrtki Jet (Wyrtki, 1973), has been observed (Knox, 1976) to develop on the Equator during both transition periods as a result of strong westerly winds.

The principal sub-surface current is the Equatorial Undercurrent (EUC) which exhibits a strong semi-annual cycle. It is effectively wind-forced, being created by the sub-surface longitudinal pressure gradients which result from the surface wind-driven flow. It flows westward during the transition periods between the monsoons and eastward the rest of the year. This is in contrast to the EUC in the Pacific and Atlantic which flows eastward thoughout the year in response to the sub-surface zonal density gradient along the Equator formed mainly by the westward surface flow. In the Indian Ocean the behaviour of the EUC seems to be more complex and is not a simple function of the equatorial winds.

b) Western boundary

The western boundary current of the Indian Ocean is unique amongst the major western boundary currents of the world in that it exhibits complete reversal in direction in an annual cycle (with the exception of the boundary current in the western Pacific which also undergoes a seasonal reversal, though of smaller amplitude, in response to the reversal in the monsoon winds). During the NE Monsoon, the surface current is southward along most of the length of Africa, including a strong cross-equatorial flow. Between about April and May the flow everywhere northward of about 10°S reverses direction; the cross-equatorial flow becomes northward and a major boundary current, the Somali Current, is generated. A component of the SEC turns northward to form the cross-equatorial flow, the remainder turning southward as during the NE Monsoon.

In association with the Somali Current, a major recirculation pattern develops. The flow separates from the coast at approximately 9-10°N and forms a large gyre, the "Great Whirl" (first recorded by Findlay, 1866). A second, smaller, recirculation pattern has also been detected offshore between about 10°N and 12°N, the "Socotra Eddy". These current systems appear to be a complex result of both remote and local forcing (Lighthill, 1969 and Leetma, 1973) and may have a natural time-dependence rather than being a steady-state pattern (McCreary & Kundu, 1988).

c) Interannual variability

It is well-known that the monsoon winds, and the rainfall on the Indian sub-continent associated with the summer monsoon, vary from year to year. Far less is known about the interannual variability in the ocean currents which must result. Most of the earlier work carried out on the link between the Indian Ocean and the monsoon has concentrated on the sea surface temperatures (SSTs) in the Arabian Sea. Schott & Quadfasel (1982) observed large interannual differences in the currents in the Somali region during the onset of the SW Monsoon, and there is a correlation between low SSTs in the Arabian Sea, a wet monsoon in India and a strong two-gyre system in the Somali region. The need for greater understanding of the interannual variability in the Indian Ocean as a whole has been stressed by Knox & Anderson (1985).

A model of the Indian Ocean which seems to simulate well the climatologically observed currents in response to forcing by climatological winds should provide a suitable tool for studying this interannual variability when forced by interannually varying winds.

2.3 Indian Ocean modelling

a) General

Lighthill (1969) was the first to try to model the wind-driven circulation of the Indian Ocean. Wunsch (1977) later investigated the response of a model to a seasonally varying wind-forcing. Climatological winds have subsequently been used as forcing in a hierarchy of model types; for example, Schott et al (1988) used both a reduced-gravity model and a multi-layer model forced by the climatological winds of Hellerman & Rosenstein (1983). Many limited-area modelling studies have been carried out, with particular interest being paid to the Somali region; McCreary & Kundu (1988), for example, used an idealized wind stress pattern and a so-called 2½-layer model to produce a two-gyre system.

Only recently has much work been done using observed winds for forcing a model. Luther & O'Brien (1989) used the ship-wind dataset of Cadet & Diehl (1984) to force a reduced-gravity model of the whole Indian Ocean. The simulated currents are qualitatively realistic; the two-gyre system is reproduced in some years and not in others, a feature of interannual variability which is in agreement with observations.

In order to simulate the temperature field accurately the thermodynamics as well as the dynamics of the model upper levels must be represented realistically and vertical heat transport must be included. Therefore the forcing by observed winds of a more complex model, which represents the 3-dimensional circulation, is required.

b) OGCMs

Bryan (1969) was the first to develop a general circulation model of the ocean (an OGCM), which was later adapted by Semtner (1974) and then by Cox (1984). The use of OGCMs has been explored extensively, including their response to observed forcing. Thus, Philander & Seigel (1985) clearly demonstrated the ability of such models to reproduce the observed variations of thermal and velocity structure of equatorial oceans in response to NWP model winds, and since then, study of the tropical Pacific Ocean using OGCMs forced by actual winds has become relatively well advanced. Leetma & Ji (1988) used wind stresses both from ship reports and from operational atmospheric analyses when assimilating hydrographic and thermal data and estimates of sea surface temperatures derived from satellite radiances and ship observations into the National Climatological Centre's OGCM for operational hindcasting of the tropical Pacific. Harrison et al (1989) have further indicated the sensitivity of a Pacific OGCM to the wind stress field used, and Merle & Morlière (1988) have outlined the progress made towards a similar operational model of the Atlantic.

Use of OGCMs of the Indian Ocean has, to date, been very limited. The Meteorological Office model described here is the first one to be used for hindcasting purposes.

3. THE MODEL

A "primitive equation" ocean model is used. This takes the six so-called primitive equations of fluid motion - the x and y components of the momentum equation, the thermodynamic energy equation, the continuity equation, the hydrostatic approximation and the equation of state - and solves them numerically on a finite difference grid. The numerical scheme is that given by Cox (1984) using the finite differencing methods of Bryan (1969). The barotropic mode is excluded from the solution. It was thought reasonable to do this since the barotropic velocities in the upper levels (those of interest) are an order of magnitude smaller than the baroclinic velocities. This has the advantage of significantly relaxing the constraints on the time-step used; the time-step for the model grid used here (described below) is 72 minutes, whereas it would need to be about 2 minutes with the barotropic mode.

The complete model domain is from 27°N to 36°S and from 35°E to 118.5°E. An irregular horizontal grid is used, illustrated in Fig.2. The maximum latitudinal grid-spacing is 1°; this reduces towards the Equator from 24°N and 24°S, according to a sine function, to be 1/3° within ±1° of the Equator. Longitudinal grid-spacing is 1½° east of 64°E, decreasing westward to be ½° everywhere west of 56°E. The important dynamics of the equatorial and western boundary regions can therefore be relatively well resolved. There are 16 levels in the vertical with a resolution of 10m for the top 30m. The model has a constant depth of 4000m.

The vertical mixing scheme of Pacanowski & Philander (1981) is used, in which the vertical eddy viscosity and eddy diffusivity coefficients depend upon the local Richardson number. The horizontal eddy diffusion coefficient varies across the grid and is set to be proportional to the larger of the latitudinal and longitudinal local grid-spacing. It takes a minimum value of $2\times10^7 cm^2/s$. The coefficients of diffusivity of heat and salinity are constant across the grid, set also at $2\times10^7 cm^2/s$.

The model has closed boundaries except on the southern side. The coast of Southern Africa is artificially forced to run due south down the 35°E meridian. The open boundary conditions applied on the southern boundary are designed to keep the vertical profiles of temperature and salinity close to climatology (the climatological fields used are those given by Levitus, 1982).

For these and other model details, including stability considerations, see Gordon (1985).

4. SEASONAL CYCLE EXPERIMENT

An experiment was performed with the model in order to simulate the climatological seasonal cycle of currents in the Indian Ocean.

4.1 Experiment details

Climatological forcing fields were used. The applied heat flux is based on a Haney (1971) condition:

$$Q = Q_C + \lambda_H \, (T - T_C)$$

where Q = net heat flux

Q_C = climatological net heat flux

T = temperature at the 5m level (the uppermost gridpoint)

T_C = climatological SST

λ_H = Haney heat flux coefficient = $-35\ Wm^{-2}K^{-1}$

There is therefore a negative feedback acting on the sea surface temperature (SST), pulling it towards climatology. Note that when $T=T_c$ the heat flux is equal to the climatological value. The net surface heat flux climatology is taken from Esbensen & Kushnir (1981) and the SST climatology is from the UK Meteorological Office (Bottomley et al, 1990).

The salinity flux is of similar form and depends on the net difference between precipitation (P) and evaporation (E).

$$P - E = (P_c - E_c) + \lambda_s (S - S_c)$$

where S = the salinity at the uppermost gridpoint

$$\lambda_s = 2 \text{ mm}(\text{day})^{-1}(\text{ppt})^{-1}$$

The value of the coefficient, λ_s, is chosen so that relaxation back to the Levitus (1982) climatology occurs on the same time-scale to that of temperature; model experiments have shown that this is necessary for the temperature and salinity fields to evolve in a consistent manner, but there is no specific physical justification for it. The precipitation and evaporation climatologies are taken from Jaegar (1976) and Esbensen & Kushnir respectively.

For both the heat and salinity fluxes, the magnitude of the feedback term ($\lambda_H(T - T_c)$ and $\lambda_s(S_c - S)$ resectively) indicates the amount of deviation of the model fluxes from the prescribed climatological surface fluxes and is the size of the "anomalous" flux of heat and fresh water consequently added to or subtracted from the model. These anomalous flux fields may be used as a model diagnostic and will be discussed further in Section 4.2.

The applied climatological wind stress is the monthly mean fields given by Hellerman & Rosenstein (1983), who calculated the data on a 2° grid from over 100 years' surface wind observations.

The starting fields used for the experiment were the climatological temperature and salinity fields for September derived from observations by Levitus (1982) and zero velocities. The model was run for just over three years. After a two-year spin-up period, by which time the model had settled to the applied fluxes to a depth of 200-300m, the final year (November to October) was taken to represent the seasonal cycle.

4.2 MODEL RESULTS

a) Currents

The model simulation of the seasonal cycle is compared with estimates of current climatology from observations. The current climatologies used are those given by Hastenrath (1985) (taken from Düing, 1970), Knox & Anderson (1985) (taken from Deutsches Hydrographisches Institut, 1960) and an analysis by Rao et al (1989) of the ship-drift data-set produced by Cutler & Swallow (1984); the observed surface currents for January and July taken from Rao et al are illustrated in Fig.3.

The aim of the present study is to simulate the wind-driven circulation and SSTs. The focus of attention is therefore on the upper level currents, the currents at the uppermost velocity grid-point (5m) being used for this purpose. Monthly mean currents are presented; this facilitates easy comparison with data from other sources. Because the model SSTs are strongly influenced by the SST climatology used in the heat flux parameterization scheme, they will not be discussed here.

The principal features of the Indian Ocean circulation are well reproduced by the model in response to forcing by seasonally varying climatological winds. Fig.4 shows the simulated surface currents for three months - January, July and May.

The equatorial surface currents compare well with observations. In January, the North Equatorial Current, extending to just south of the Equator, and the strong Equatorial Counter Current can be seen in Fig.4a. Maximum current velocities are realistic: in excess of 60 cm/s for the NEC, around 40 cm/s for the ECC, and 20-30cm/s for the South Equatorial Current. The latitudinal positions of these currents are also consistent with observations. In July (see Fig.4b), the maximum velocity in the Southwest Monsoon Current of over 80 cm/s is perhaps too strong, but the weaker flow in the SEC, situated well to the south of the Equator at around 10°S, is well simulated. In both May and September the Wyrtki Jet reaches a maximum velocity of 100cm/s (see Fig.4c) which compares well with observations.

In the region of the Somali Current, the seasonal reversal in direction of the flow along the western equatorial boundary is reproduced. Northward flow from where the SEC meets the African coast, across the Equator and into the Somali Current can be seen in Fig.4b; a velocity maximum in the Somali Current in excess of 100 cm/s is reasonable. However, two weaknesses in the circulation are that the recirculation associated with the "Great Whirl" is obscured in the uppermost model levels (see below), and the model fails to reproduce the Socotra Eddy at all, possibly because the island of Socotra is not represented and the model resolution is lower at that latitude and longitude.

The tendency of the model to obscure the Great Whirl in the upper levels highlights a general model weakness, namely that the Ekman component of the simulated surface currents is too strong. In the case of the Somali Current, where the wind stress is towards the northeast, the Ekman current is in an eastward direction; the recirculation, which involves westward flow, is not reproduced. At lower model levels, however, the gyre is simulated; this is illustrated in Fig.5 which shows the currents at 35m for July. The problem seems to be related to the way in which the downward transfer of momentum from the wind stress is parametrized in the model, but it is not well understood and attempts to eliminate the problem by using other methods to apply the wind stress have been unsuccessful.

In a qualitative sense the principal aspects of the sub-surface currents in the equatorial region are well reproduced, though lack of data makes quantitative comparison difficult. Fig.6 shows a depth-time plot of zonal velocity at 70°E on the Equator. There is a marked semi-annual cycle in the Equatorial Undercurrent, as would be expected.

The model simulation of the currents in the deep ocean is not reliable for three reasons. Firstly, there is no barotropic component to the flow, so the deep currents in the model must balance the baroclinic velocities in the upper levels, which, in general, is unrealistic. Secondly, the model spin-up time for the deep ocean is far greater than that for the upper layer and so the density distribution will not have reached equilibrium. Thirdly, the model ocean has a flat bottom and does not represent bottom topography. No analysis will therefore be made of the deep currents.

b) Anomalous heat and water fluxes

The fields of anomalous fluxes of heat and fresh water are indicators, as described above, of the deviation of the model from the prescribed surface climatological fluxes. This deviation can result from inaccuracies in one or all of the following quantities: the climatological fluxes of heat and fresh water; the climatological SST and surface salinity fields to which the model surface fields are being forced; and the model simulation of surface temperature and salinity.

The anomalous heat flux over most of the model domain is found to be generally in the range 0 to -35 Wm^{-2}, which is equivalent to the SST being too warm by 0 to 1°C. These anomalies are as likely to be a result of inaccuracies in the climatological fields as in the model simulations. The only instances of large anomalous heat fluxes (of magnitude much greater than 35 Wm^{-2}) are in areas of strong upwelling, where the model SSTs are significantly cooler than the climatological SSTs; this is illustrated in Fig.7 which shows the anomalous heat flux for July in which the upwelling along the Somali coast during the Southwest Monsoon may be clearly seen. The main reason for the large anomalous fluxes here is that the climatological fields do not resolve the coastal upwelling regions.

The anomalous water flux principally indicates inconsistencies between the flux of precipitation-minus-evaporation (P-E) and the climatological surface salinity field. This is most apparent in the Bay of Bengal since the land surface runoff associated with the Ganges River is not included in the (P-E) climatology; Fig.8 shows the anomalous water flux for July.

5. CONCLUSIONS

An Ocean General Circulation Model of the Indian Ocean has been developed with a view to being able to hindcast the upper-level circulation.

When forced by climatological winds and heat fluxes the model reproduces well the seasonal cycle of the principal surface currents, the only major weakness being in the Somali region. This ability of the model to simulate successfully the seasonal cycle gives confidence in its suitability as a tool for proceeding to an investigation of the interannual variability in the currents in the Indian Ocean in response to actual winds and heat fluxes. For this purpose forcing fields derived from the operational Numerical Weather Prediction model analyses of the Meteorological Office and the European Centre will be used. This will be discussed in a subsequent paper.

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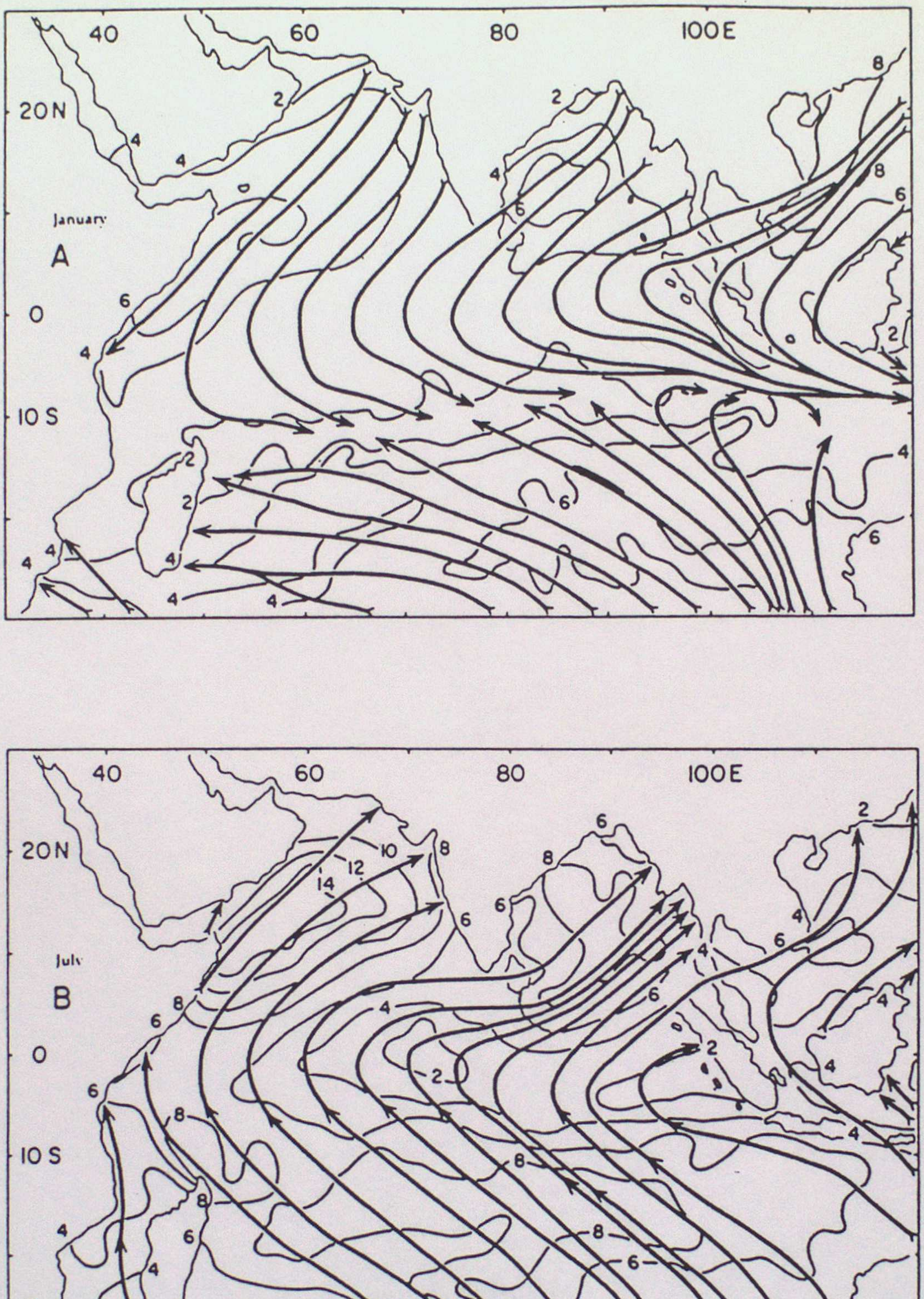
a



b

Fig.1 Surface wind field over the tropical Indian Ocean in (a) January and (b) July. Streamlines, and isotachs in m/s. (From Hastenrath & Lamb, 1980).

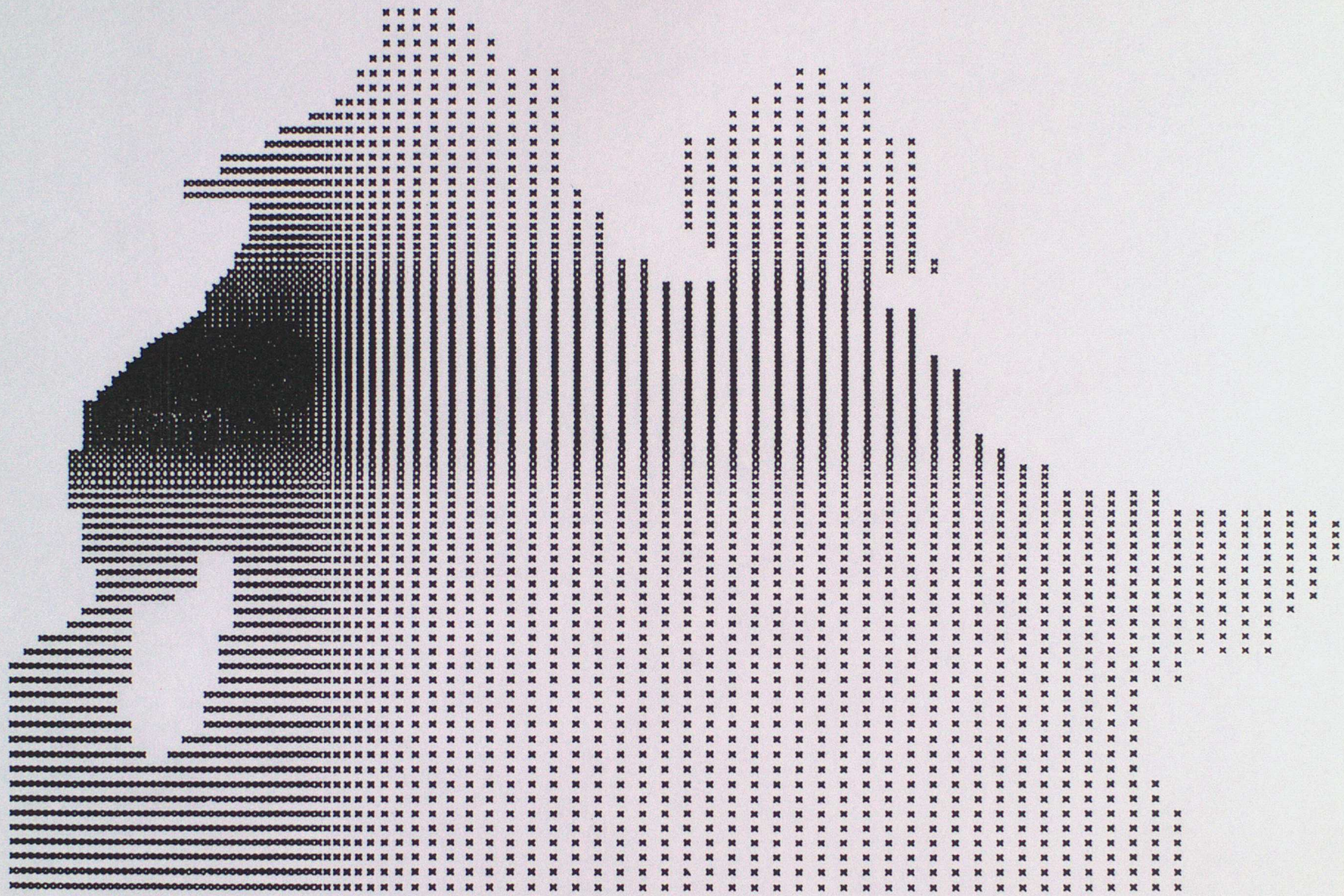
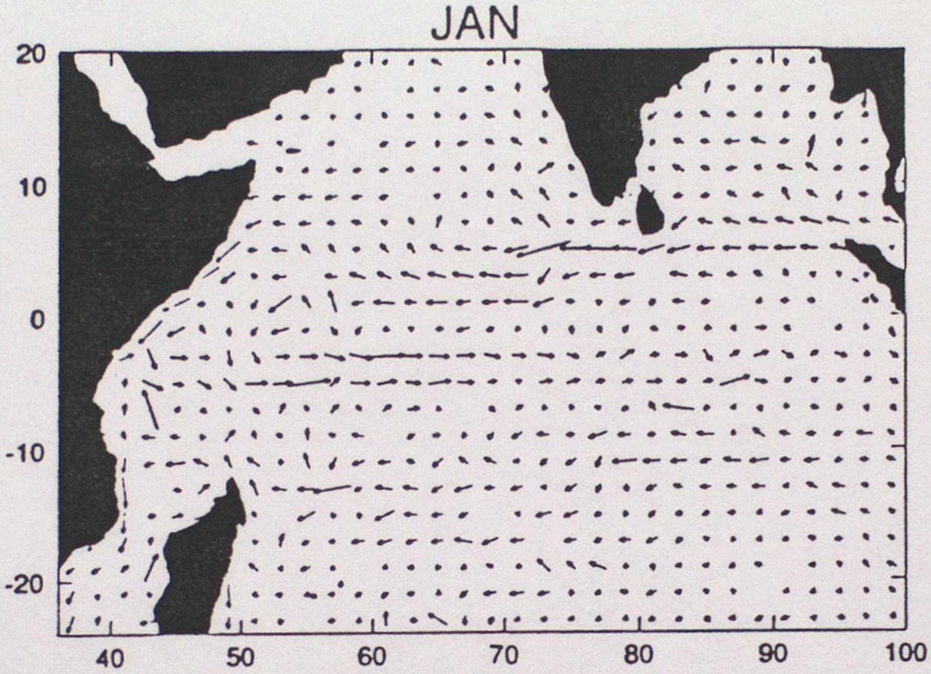


Fig.2 The Indian Ocean model horizontal grid.

Fig.3 Observed mean monthly ship drift currents for (a) January and (b) July from analysis by Rao et al (1989) of ship drift climatology of Cutler & Swallow (1984).



-50cm/s

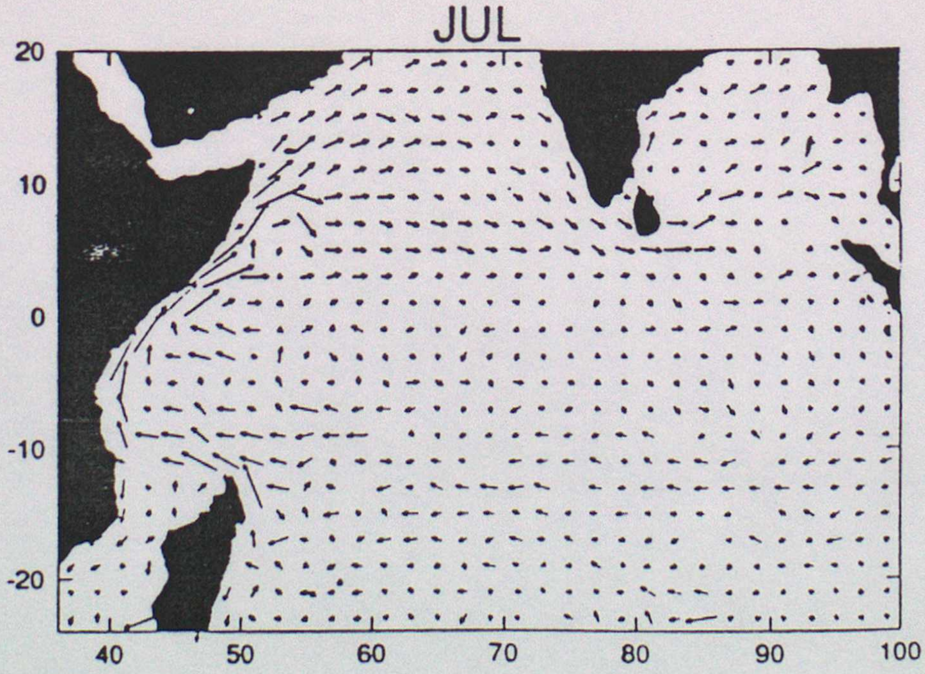
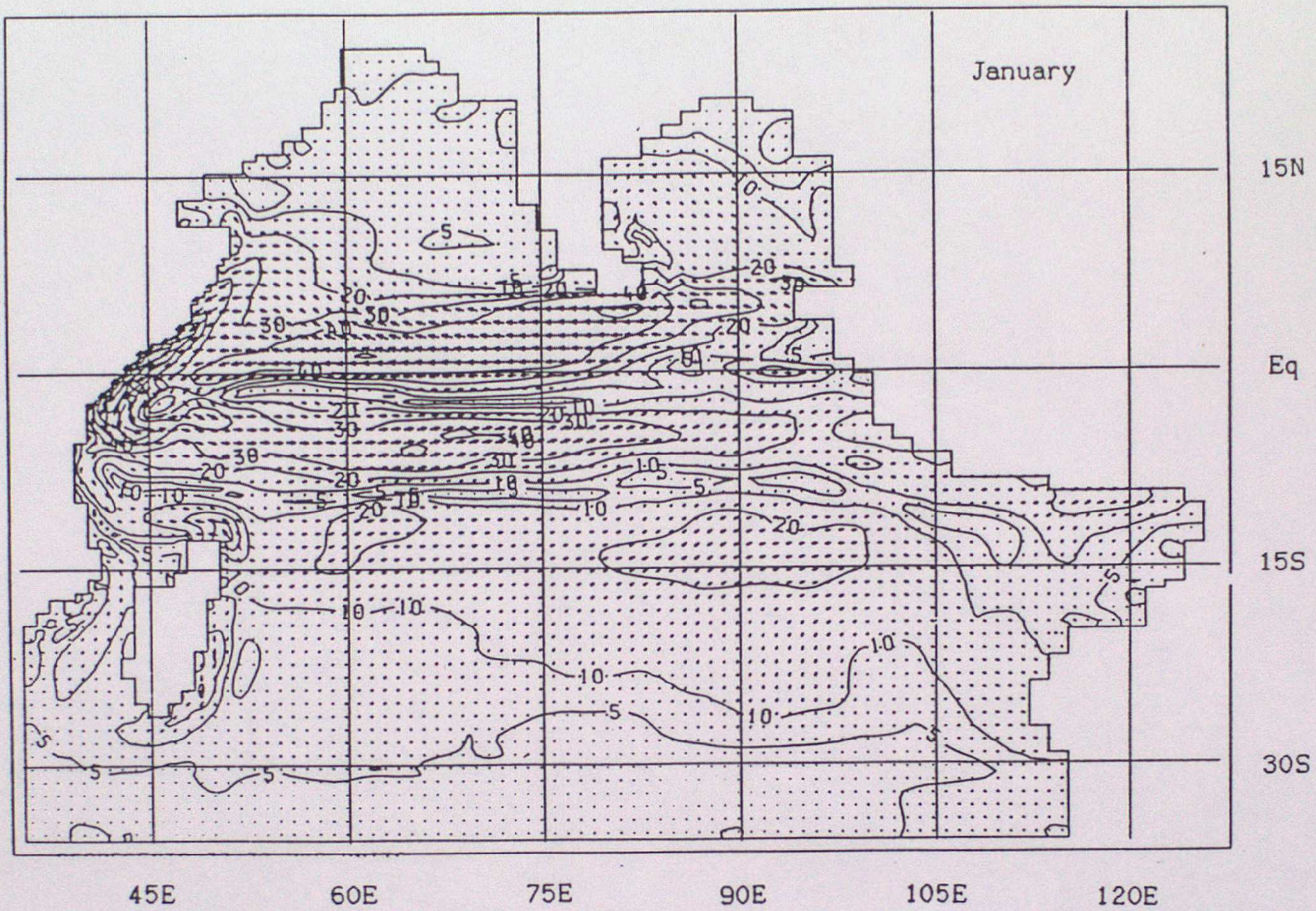


Fig.4 Model seasonal cycle: 5m currents for (a) January, (b) July and (c) May.

a



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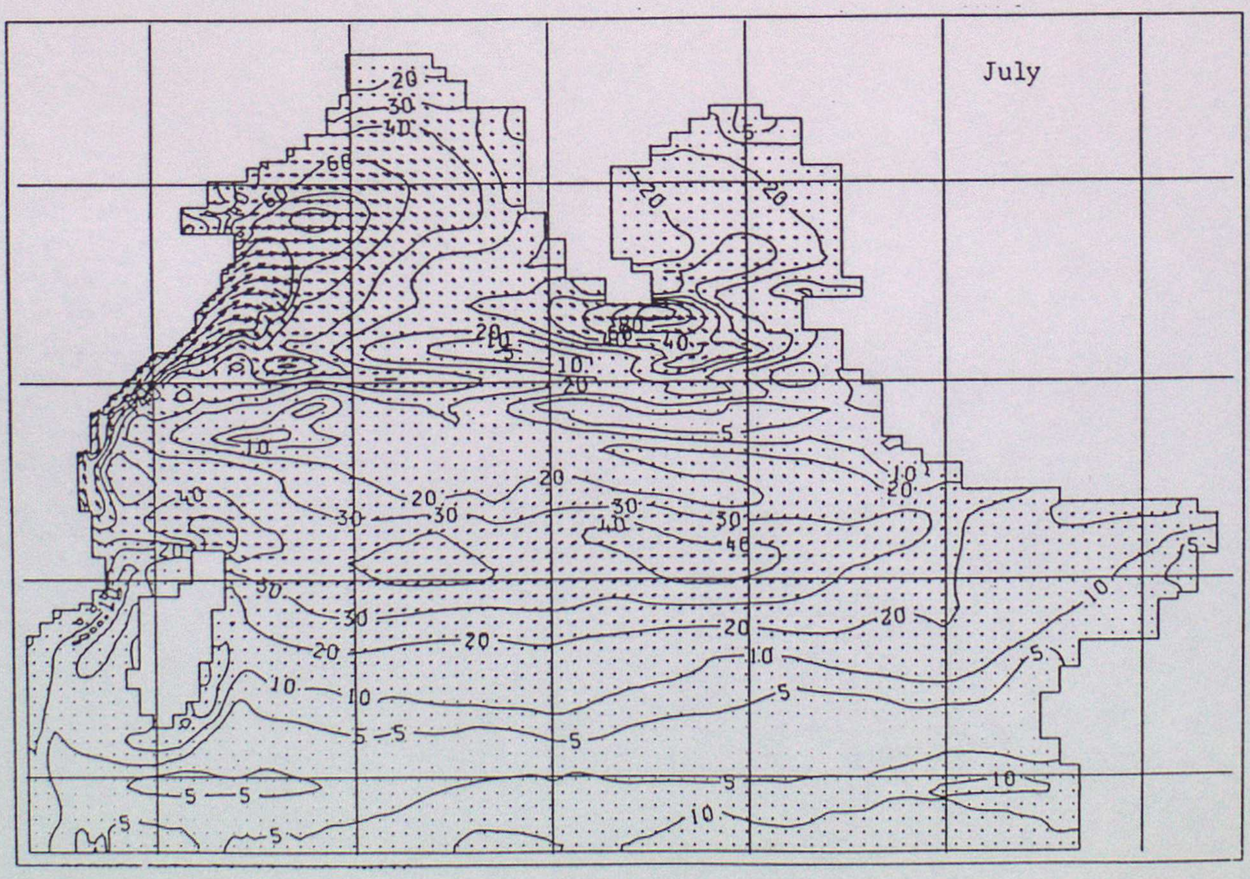

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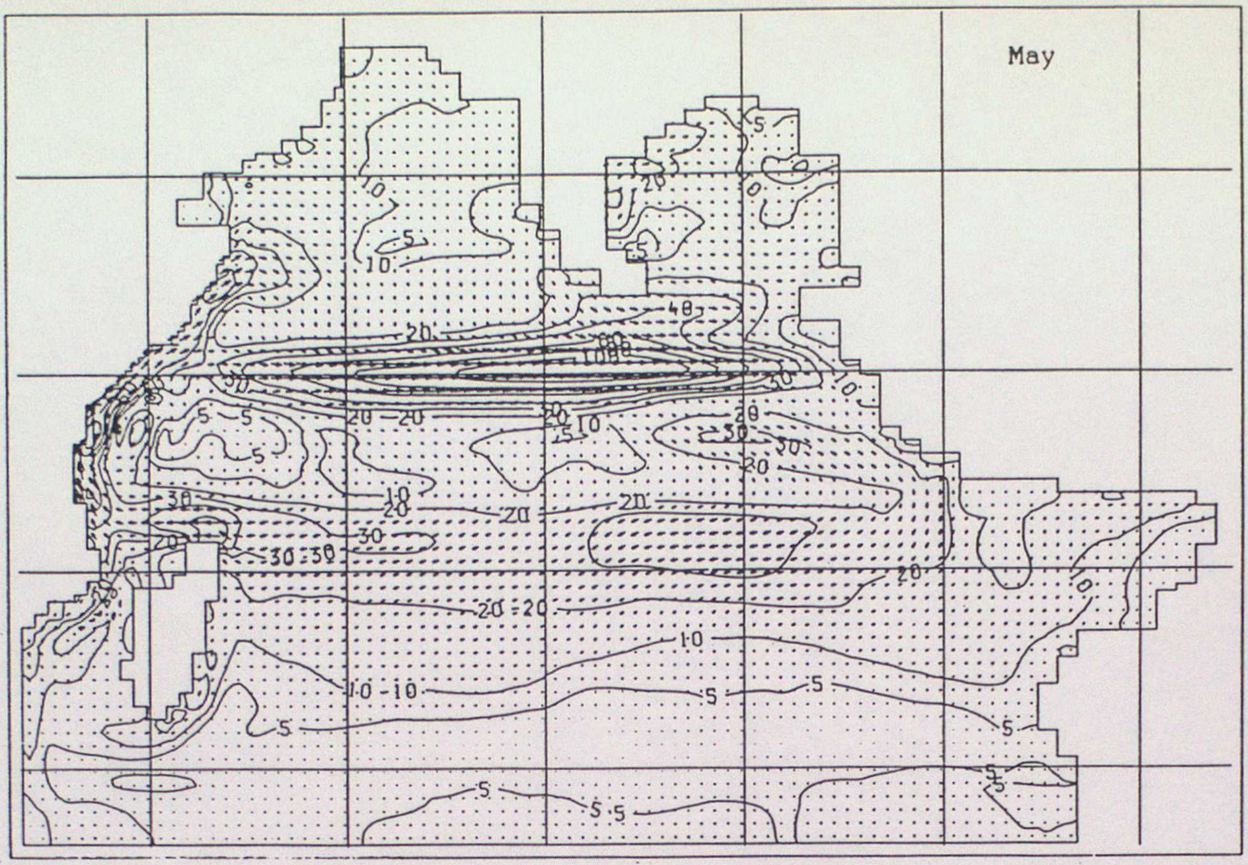


Fig.5 Model seasonal cycle: 35m currents for July.

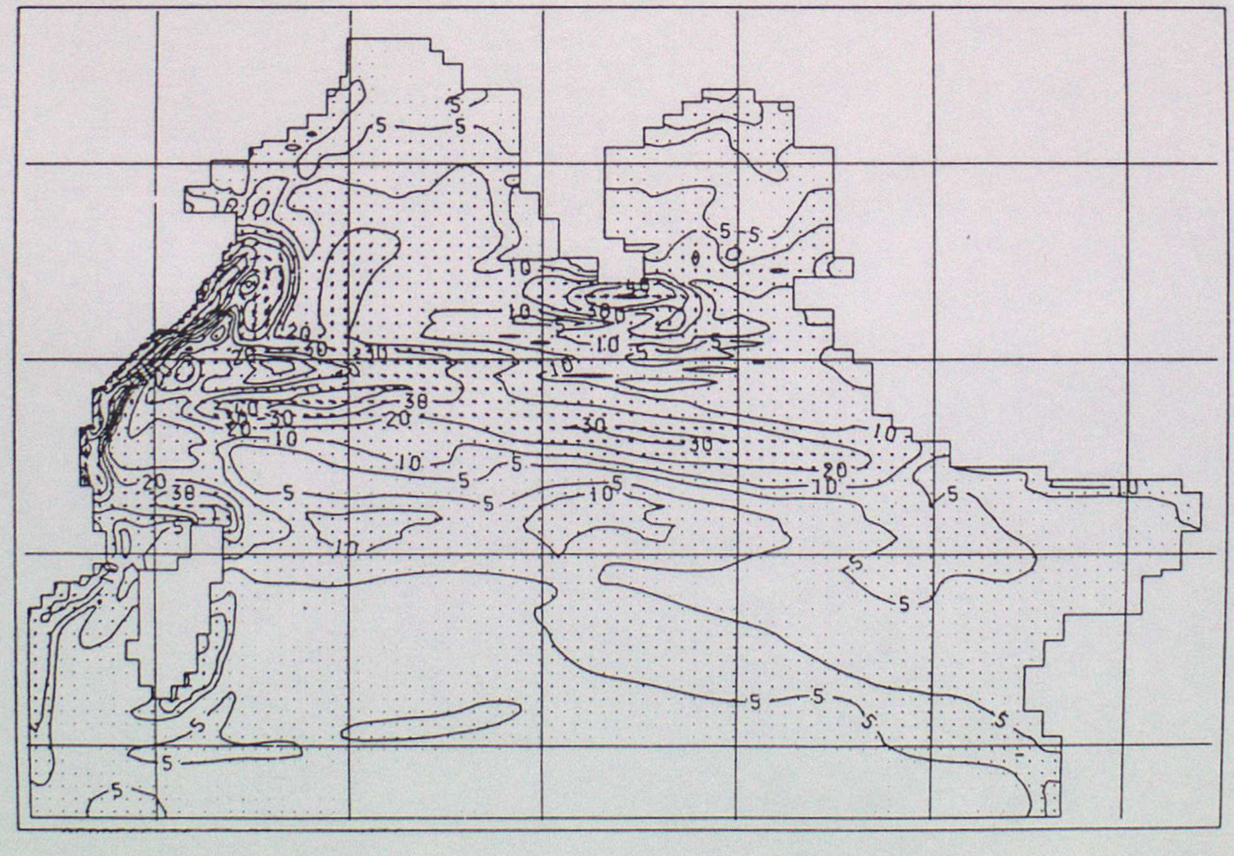
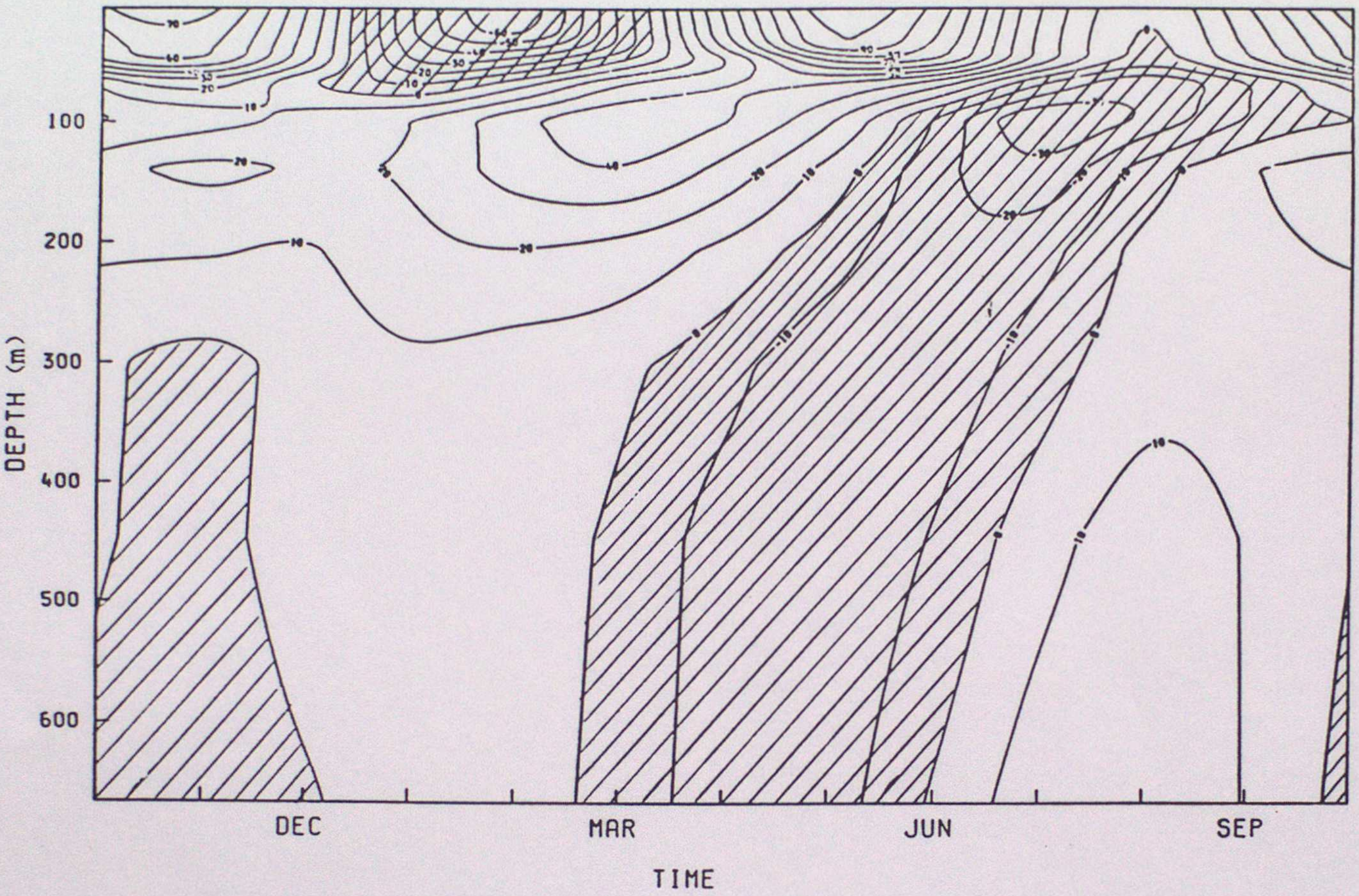


Fig.6 Model seasonal cycle: depth-time plot of zonal velocity (cm/s) at 70°E on the Equator (westward values shaded).



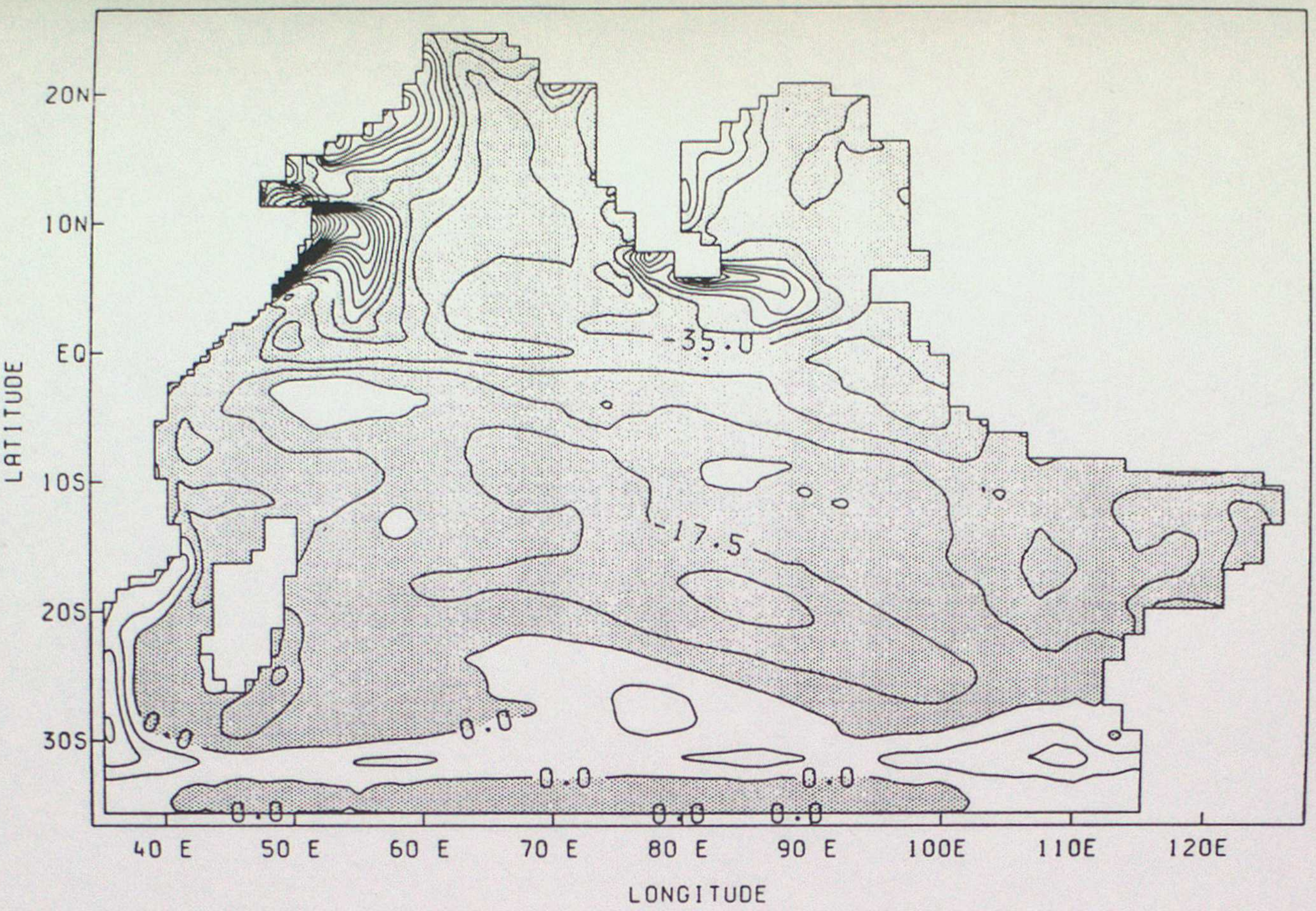


Fig.7 Model seasonal cycle: anomalous heat flux (Wm^{-2}) for July. *Positive* values correspond to downward heat flux. Contour interval 17.5 Wm^{-2} and negative values shaded.

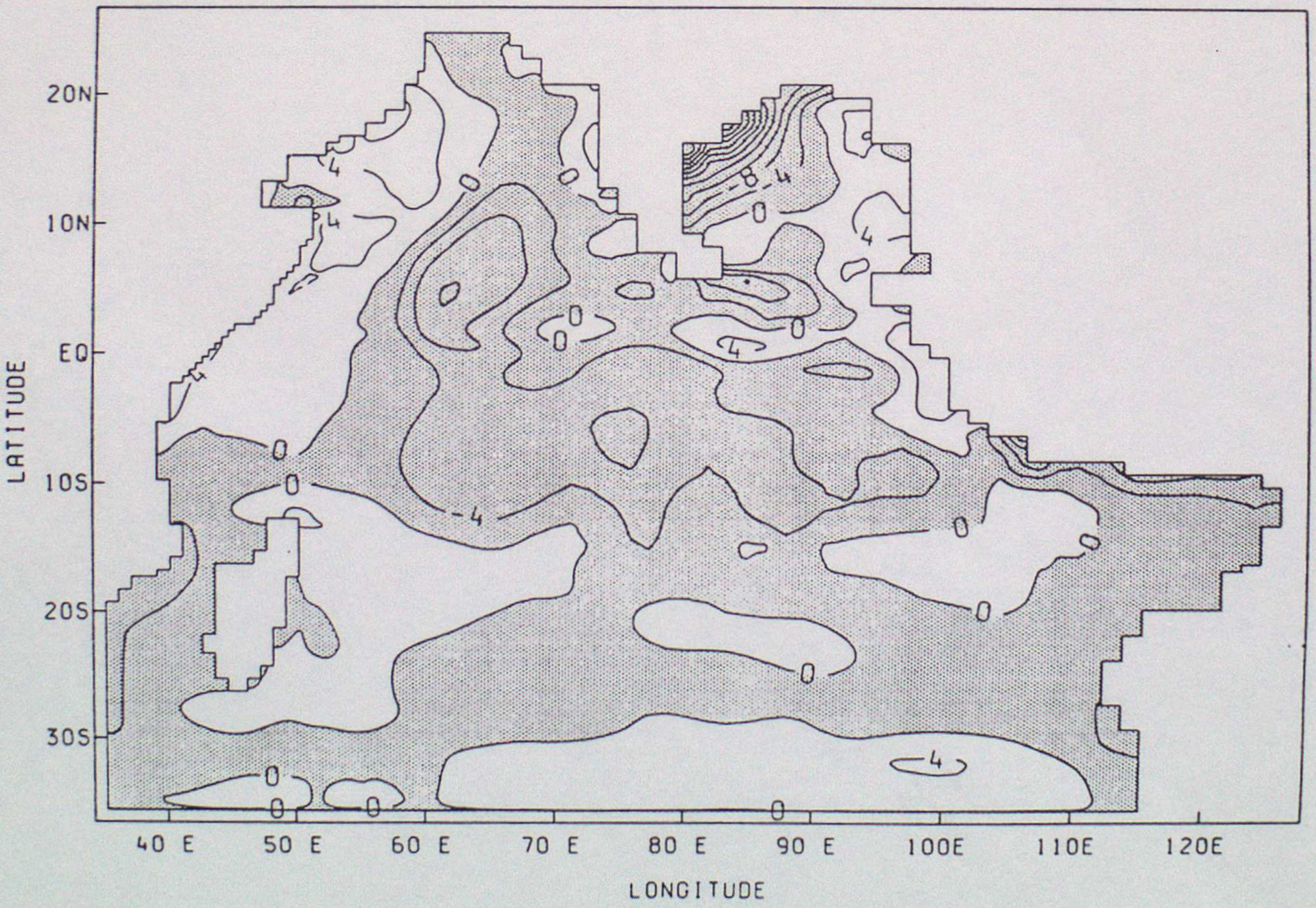


Fig.8 Model seasonal cycle: anomalous fresh water flux (mm $(day)^{-1}$) for July. *Negative* values correspond to downward fresh water flux. Contour interval 4 mm $(day)^{-1}$ and negative values shaded.