Dynamical Climatology

On the specification of surface fluxes in coupled atmosphere ocean general circulation models.

by

J.F.B. Mitchell, C.A.Wilson and C.M.Price

DCTN 7 September 1984

Meteorological Office (Met. O. 20)
London Road
Bracknell
Berkshire RG12 2SZ

DYNAMICAL CLIMATOLOGY TECHNICAL NOTE NO. 7

ON THE SPECIFICATION OF SURFACE FLUXES IN

COUPLED ATMOSPHERE OCEAN GENERAL CIRCULATION MODELS

By

J F B Mitchell, C A Wilson and C Price

Met O 20 (Dynamical
Climatology Branch)
Meteorological Office
London Road
Bracknell
Berkshire RG12 2SZ

September 1984

Submitted for publication in the proceedings of the 16th Liege Colloquium
"Coupled atmosphere-ocean models", held 7th-11th May 1984.

On the specification of surface fluxes in coupled ocean atmosphere general circulation models

Abstract

The turbulent surface fluxes of heat and momentum from a high resolution atmospheric model are presented and assessed. The errors in computing the fluxes from monthly mean atmospheric model data are calculated, and the consequences for coupled ocean models are discussed.

1. Introduction

The coupling between the ocean and the atmosphere is effected entirely by the fluxes of momentum and heat (latent and sensible) at the air-sea interface. The atmospheric circulation is dependent on the temperature, and to a lesser extent, the roughness of the ocean surface. The ocean surface temperature in turn depends on the ocean circulation which is forced by the transfer of momentum and heat from the atmosphere. Errors in coupled ocean-atmosphere simulations may arise from errors in the atmospheric model, the oceanic model, or both. In this paper, we assess a high resolution atmospheric model by comparing the simulated stress and turbulent heat flux with climatological estimates, and discuss the likely implications for coupling to an ocean model.

The fluxes of heat and momentum from both the model and climatological data are calculated using the bulk aerodynamic formulae, involving the product of a drag coefficient, surface wind and vertical gradient. In the model, the fluxes of heat and momentum from the atmosphere to the ocean are calculated timestep by timestep, and so fully take into account the correlations in time between drag coefficients, windspeed and vertical gradients. In regions where such correlations are large, the surface fluxes of heat and momentum will differ from those obtained from drag coefficients, wind speed and vertical gradients derived from time-meaned atmospheric variables. Hence an ocean driven by atmospheric fluxes accumulated timestep by timestep will be forced differently to one driven by fluxes derived from time meaned data.

This is analogous to the differences obtained in climatological estimates of surface fluxes derived from daily rather than monthly mean atmospheric variables. (Kraus and Morrison, 1966; Esbensen and Reynolds, 1981). Here we calculate the errors in model fluxes derived from monthly averaged atmospheric model data, and compare them with the errors found in parallel studies made with observational data.

A similar error may occur in bringing coupled ocean atmosphere models to equilibrium. The thermal relaxation time of the ocean is several orders of magnitude larger than that of the atmosphere. On the other hand, oceanic general circulation models are generally computationally less expensive than atmospheric models. In order to bring a coupled ocean atmosphere model to equilibrium, it has been the practice to run the atmospheric model for a short period (typically one month) with fixed ocean temperatures and then use the mean data to force the ocean model over a much longer period of time (e.g. Manabe et al 1979a, Washington et al 1980), a process often referred to as "asynchronous coupling". During the asynchronous period, the ocean is driven by the fluxes of momentum and latent heat accumulated timestep by timestep during the synchronous period. The sensible heat flux is derived from the mean low level winds, temperatures and humidities to ensure convergence to equilibrium, and so short term correlations between atmospheric variables are ignored.

2. The Model

The atmospheric model is global with 11 layers in the vertical; a limited area version was used with data from the GARP Atlantic tropical experiment (GATE) (Lyne et al 1976). It is a primitive equation model using σ (pressure/surface pressure) as a vertical coordinate, and a regular 2.5° x 3.75° latitude longitude grid. The seasonal and diurnal variation of solar radiation are represented, and the radiative fluxes are a function of temperature, water vapour, carbon dioxide and ozone concentrations, and prescribed zonally averaged cloudiness. Sea surface temperatures and sea ice extents are prescribed from climatology, and updated every 5 days.

The surface exchanges of momentum, heat and moisture are determined by

$$F_\chi = -C_\chi V(Z_1) \Delta\chi(Z_1) \qquad (1)$$

where F_χ is the mean vertical flux of χ

C_χ is the bulk transfer coefficient at height Z

$V(Z_1)$ is the mean wind speed at a specified height Z_1 above the surface but within the atmospheric boundary layer.

$\Delta\chi(Z_1) = \chi(Z_1) - \chi_o$ is the difference between the value of χ at Z_1 and its surface value χ_o.

The bulk transfer coefficients are chosen using Method I of Clarke (1970), which is based on the Monin-Obukhov similarity hypothesis for the fully turbulent boundary layer. In practice, the value of C_χ over the ocean is tabulated against the bulk Richardson number R_{iB} assuming a roughness length of 10^{-4}m. (Figure 1).

$$R_{iB} = \frac{gZ_1 \; [\Delta\theta(Z_1) + 0.61\, T\, \Delta q\, (Z_1)]}{T \qquad V^2(Z_1)} \qquad (2)$$

where g is the acceleration due to gravity, Z_1 is the height of the centre of the lowest model layer, T is a representative temperature for the bottom layer. θ, q and V are respectively the potential temperature, specific humidity and wind speed in the lowest model layer. Note that this takes into account the effect of stability, as indicated by the sharp increase in C_χ as one moves from stable ($R_{iB} > 0$) to unstable ($R_{iB} < 0$) in Figure 1. No allowance is made for an increase in surface roughness with windspeed when calculating the bulk transfer coefficient for momentum. In deriving quantities from monthly mean data, the monthly mean windspeed (average of values diagnosed each timestep) is used for $V(Z_1)$ in (1), and the monthly mean values of temperature, humidity and surface velocity are used in calculating C_χ and $\Delta\chi(Z_1)$.

3. Results and Discussion

3.1 Diagnosed fluxes

3.1.1 Momentum. The model's diagnosed monthly mean wind stress field, averaged over three years, (Figure 2) is qualitatively similar to that derived from climatological data (eg Han and Lee, 1981; Hellerman and Rosenstein, 1983). In January (Figure 2a), the maximum magnitude of westerly stress (about 2.5 dynes cm^{-2}) in the north of the major ocean basins is similar to that found in the climatological estimates. However, the easterly stress in the northern tropics exceeds 0.5 dynes cm^{-2} in only a few places whereas the climatologies suggest that it should exceed 1.0 dynes cm^{-2} over much of the tropical Pacific and Atlantic. The easterly stress in the southern hemisphere tropics also appears to be substantially underestimated. A belt of maximum westerly stress is found near 45°S in accordance with observations, though weaker. In July (Figure 2b) the model field is qualitatively similar to the climatological estimates, but generally weaker. The strength of the westerly stress around Antarctica is much closer to the observed data than in January, but there is a considerable underestimation of the easterly stress over the subtropical Atlantic and Pacific Oceans, and also over the north eastern Indian Ocean associated with the monsoon.

The underestimation of the strength of the surface stress in an atmospheric model which is to be coupled to a dynamical model of the ocean is a serious shortcoming. For example Bryan et al (1975), and Washington et al (1980) attribute the anomalous oceanic warming and the consequent underestimate of sea-ice cover round Antarctica in their coupled ocean-atmosphere models in part to the weakness of the westerly circumpolar flow, and the associated equatorward Ekman surface drift. Similarly, a coupled model will fail to produce sufficient equatorial upwelling, and the pronounced east-west surface temperature contrast which characterizes the equatorial Pacific if the atmospheric model does not produce sufficient easterly stress. In higher northern latitudes, a weakening of the surface stress will contribute to a weakening of the western boundary currents. (See Bryan, Gates; this volume?).

One reason for the underestimate of the surface stress by atmospheric models may be lack of horizontal resolution. In general, the surface flow becomes stronger as resolution is increased, though in the northern hemisphere in winter, the mid latitude depression belt becomes excessively deep. (See, for example Manabe et al., 1979b). Similar trends are found in the present model. If the horizontal diffusivity, included to remove computational stability, is also reduced as the resolution is enhanced, the energy in the transient flow is increased, and is in closer agreement with observations.

The Meteorological Office model produces stronger than observed westerly flow in northern mid-latitudes in January, yet apparently still simulates the magnitude of the surface stress correctly (Figure 2a). The drag coefficient fixed for momentum does not include a dependence on wind strength (except in the wrong sense through the bulk Richardson number in (2)). There is a growing body of evidence that the momentum drag coefficient should increase with wind speed, to allow for the increase in surface roughness (Wu (1982) has attempted to produce a simple empirical relationship between wind strength and the momentum drag coefficient). Increasing the drag coefficient for high wind speeds would undoubtedly increase the surface stress in the model. However, the atmospheric flow will be decelerated as a result, reducing the net increase.

3.1.2 Turbulent Heat Fluxes. The turbulent heat fluxes (sensible and latent heat, Figure 3) are generally similar in both magnitude and geographic distribution to available climatological estimates. (Budyko, 1963; Bunker, 1976 Esbensen and Kushnir, 1981). Peak values of over 200 Wm^{-2} occur in winter along the sea ice margin, and, in the northern hemisphere, off the eastern seaboard of continents. Large values also occur in the tropics, though there is a minimum along the equator, particularly in the East Pacific, due to the local surface temperature minimum associated with oceanic upwelling. In the summer hemisphere, the fluxes in middle and high latitudes are small. Note the large cooling of the ocean off India associated with the summer monsoon. Further discussion of the simulation of sea surface temperatures using the model's turbulent heat flux is given by Gordon and Bottomley (this volume).

3.2 Errors in sampling due to the use of monthly mean data

The surface fluxes of momentum and turbulent (sensible plus latent) heat flux were

(a) accumulated timestep by timestep (diagnosed) as presented in Section 3.1.

(b) calculated in retrospect from monthly time means of low level temperature, wind and humidity using the model's boundary layer algorithm (derived).

The difference (derived - diagnosed) indicates the likely error introduced into climatological estimates by using monthly mean data, or in the surface forcing of an ocean model coupled asynchronously to an atmospheric model.

3.2.1 Surface stress

A comparison of Figure 2 (diagnosed stress) and Figure 4 (derived -diagnosed stress, with the contour interval reduced by a factor of ten) indicates that magnitudes of diagnosed and derived stresses differ by about 10%, though locally the differences exceed 20%. In general, the difference in stresses is in the opposite direction to the diagnosed fields, indicating a consistent underestimation of the stress due to using monthly mean data. This is in agreement with the findings of Esbensen and Reynolds (1981) using observational data, who found that the monthly mean stress at various weather ships was larger when estimated from data when divided into 16 direction categories, each with a mean wind speed, as opposed to a single monthly mean wind speed and direction. (Note that here, the "derived" surface stress is calculated from

$$\underset{\sim}{\tau} = \rho C_D |V| \underset{\sim}{V}$$

where $|V|$ is wind speed averaged over each model timestep, but $\underset{\sim}{V}$ is the monthly mean vector wind, whereas Esbensen and Reynolds appear to have used

$$\underset{\sim}{\tau} = \rho C_D |V|^2 \hat{\underset{\sim}{V}}$$

where $\hat{\underset{\sim}{V}}$ is the unit vector in the direction of the mean vector wind. Since $|V| \geq |\underset{\sim}{V}|$, our "derived" wind stress will tend to be smaller than using their "direction only" wind rose method).

3.2.2 Turbulent fluxes

Differences (Figures 5a,b; 6a,b) are generally less than $10Wm^{-2}$, though they exceed $20Wm^{-2}$ locally. There are few points where the difference exceeds 10%, in agreement with Esbensen and Reynolds findings using observational data. The sign of the difference varies, and so cannot be minimised by a uniform fractional change in the magnitude of the fluxes. Note the general underestimation of the surface cooling in the vicinity of the sea-ice margins. This is probably due to the correlation between large (unstable) drag coefficients and large air-sea temperature differences which occur when equatorward winds advect cold air from sea-ice over the warm ocean. The surface cooling may be underestimated in a similar fashion in an ocean model forced by prescribed low level atmospheric data (wind, temperature and humidity), leading to an underestimation of the extent of sea ice and intensity of meridional circulation.

There is also a consistent over-estimation of the heat flux in lower latitudes (30°N and 30°S in December, January, February; 20°N and 35°S in June, July, August).

4. Concluding remarks

The UK Meteorological Office 11-layer model appears to underestimate the surface stress over the ocean. This may be due to insufficient horizontal resolution and a lack of a wind speed dependence in the calculation of the momentum drag coefficient. The simulated turbulent heat fluxes are broadly similar to climatologicl estimates. As found in observational studies, the use of monthly mean data can lead to an underestimation of the surface stress (by up to 20%), and a geographically dependent bias in the surface fluxes (up to 5 or 10 Wm^{-2}). If more accuracy is required for coupled simulations, the wind stress should be accumulated from the atmospheric model timestep by timestep, as should the turbulent heat flux in the case of synchronous coupling. An asynchronously coupled ocean-atmosphere model will converge to a slightly different equilibrium to a synchronously coupled model, due to the inevitable bias in the estimate of the surface fluxes. Although this bias is generally less than 10%, it is particularly pronounced along the sea-ice margins, where small differences in simulations could be amplified by the strong feedback between temperature and albedo.

References

Bryan K, Manabe S and Pacanowski R L 1975 A Global Ocean-Atmosphere Climate Model Part II. The oceanic circulation. J Phys Oceanogr. 5, 30-46.

Budyko, M I Ed. 1963, Atlas of the Heat Balance of the Earth, Akad. Nauk SSSR, Prezidium. Mezhduvedomstvennyi Geofiz, Komitet.

Bunker A F 1976, Computation of surface energy flux and annual air-sea interaction cycles of the North Atlantic Ocean. Mon. Wea. Rev. 104, 1122-1140.

Clarke, R.H., 1970, Recommended methods for the treatment of the boundary layer in numerical models. Australian Met. Mag. 18, 51-73.

Esbensen S.K., and Kushnir, Y 1981, The heat budget of the global ocean. An atlas based on estimates from surface marine observations. Climate Research Institute Report 29, Oregon State University Corvallis Oregon.

Esbensen S K., and Reynolds, R.W., 1981, Estimating monthly averaged air-sea transfer of heat and momentum using the bulk aerodynamic method. J. Phys. Oceanogr. 11 457-465.

Han, Y.J., and Lee, S.W, 1981 A new analysis of monthly mean wind stress over the global ocean. Climate Research Institute Report No 26, Oregan State University, Corvallis, Oregon.

Hellerman, S and Rosenstein, M. 1983, Normal Monthly Wind Stress over the World Ocean with Error Estimates. J Phys. Oceanogr. 13, 1093-1104.

Kraus, E.B., and Morrison, R E, 1966, Local interactions between the sea and the air at monthly and annual timescales. Quart. J. Roy. Meteor. Soc 92 114-27.

Lyne, W.H., Rowntree, P.R., Temperton, C. and Walker, Julia, 1976 Numerical modelling using GATE data Met. Mag. 105, 261-271.

Manabe, S., Bryan, K., and Spelman, M.J., 1979a, A Global Ocean-atmosphere climate model with seasonal variation for future studies of climate sensitivity. Dynamics of Atmospheres and Oceans, 3 393-426.

Manabe, S., Hahn, D.G. and Holloway, J.R., Jr. 1979b Climate Simulations with GFDL Spectral Models of the atmosphere: Effect of Spectral Truncation Report of JOC Study Conference on climate models: Performance, intercomparison and sensitivity studies. GARP Publications Series No 22. 41-94

Washington, W.M., Semtner, A.J. Jr., Meehl, G.A., Knight, D.J., and Mayer, T.A., 1980, A general circulation experiment with a coupled atmosphere, ocean and sea-ice model. J. Phys. Oceanogr, 10, 1887-1908.

Wu, J., 1982 Wind stress coefficients over sea surface from breeze to hurricane. J. Phys. Oceanogr. 12, 9704-9706.

Figure Captions

1. Surface layer bulk transfer coefficients used in the Meteorological Office 5-layer model.

2. Model surface stress (3 year mean). Contours every 0.5 dyne cm^{-2} (0.05 Nm^{-2}). Shaded where less than 0.5 dyne cm^{-2}. Arrows show direction only.

(a) January (b) July

3. Model turbulent (sensible + latent) heat flux (3 year mean) Contours every 20 Wm^{-2}.

(a) December, January and February, (b) June, July and August.

4. Error in model surface stress due to using monthly mean data. (3 year mean). Contours every 0.5 dyne cm^{-2}. Shaded where less than 0.5 dyne cm^{-2}. Arrows show directions only.

(a) January (b) July

5. Error in model turbulent heat flux due to using mean data (3 year mean). Contours every 5 Wm^{-2}; shaded where underestimated. December, January and February. (a) Northern Hemisphere (b) Southern Hemisphere.

6. Error in model turbulent heat flux due to using mean data (3 year mean). Contours every 5 Wm^{-2}, shaded where underestimated. June, July and August (a) Northern Hemisphere (b) Southern Hemisphere.

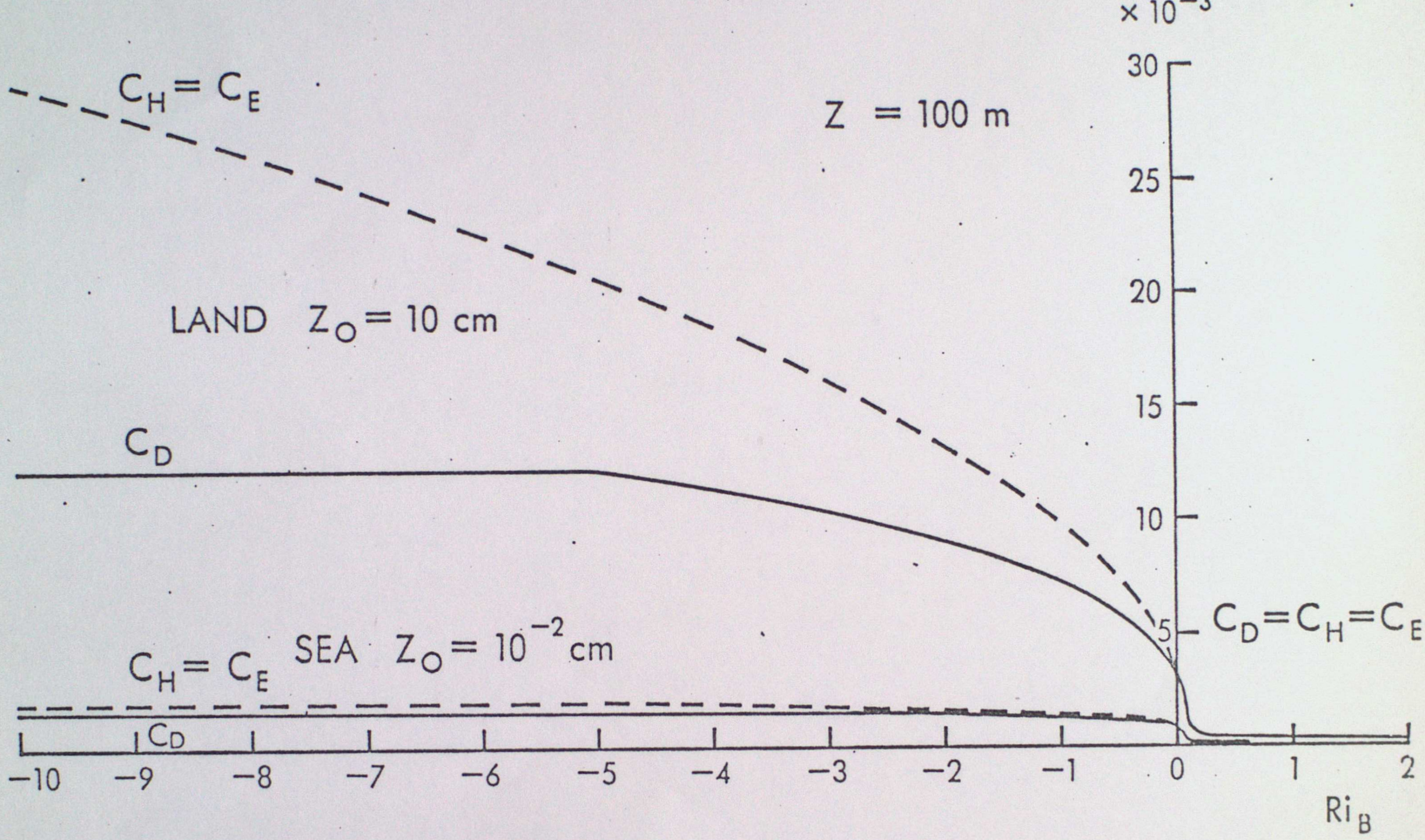


Figure 1 Surface layer bulk transfer coefficients derived from Monin-Obukhov similarity theory and used in the MO 11-level model.

SURFACE WIND STRESS   DIAGNOSED
JANUARY MEAN

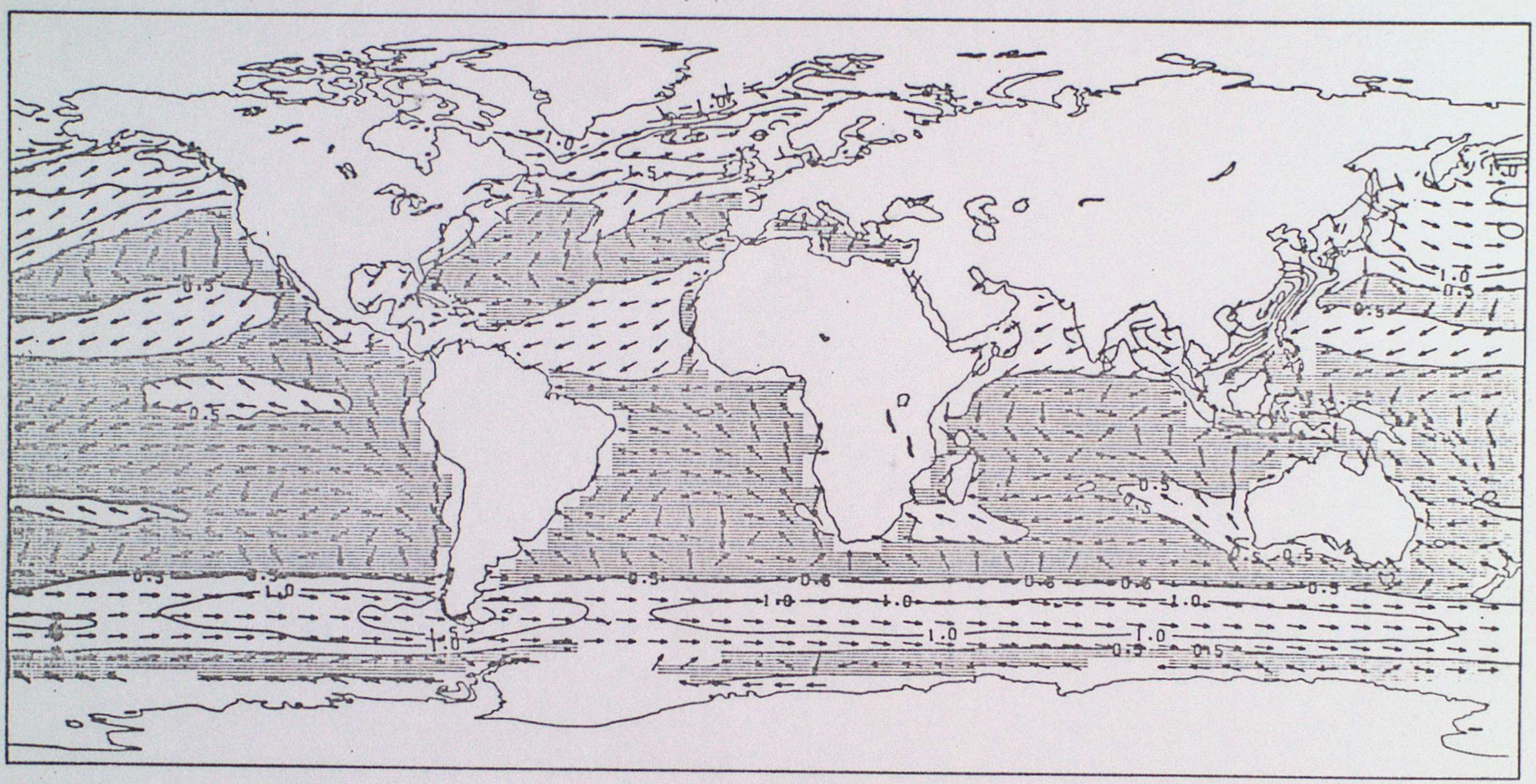


Figure 2a

SURFACE WIND STRESS DIAGNOSED
JULY MEAN

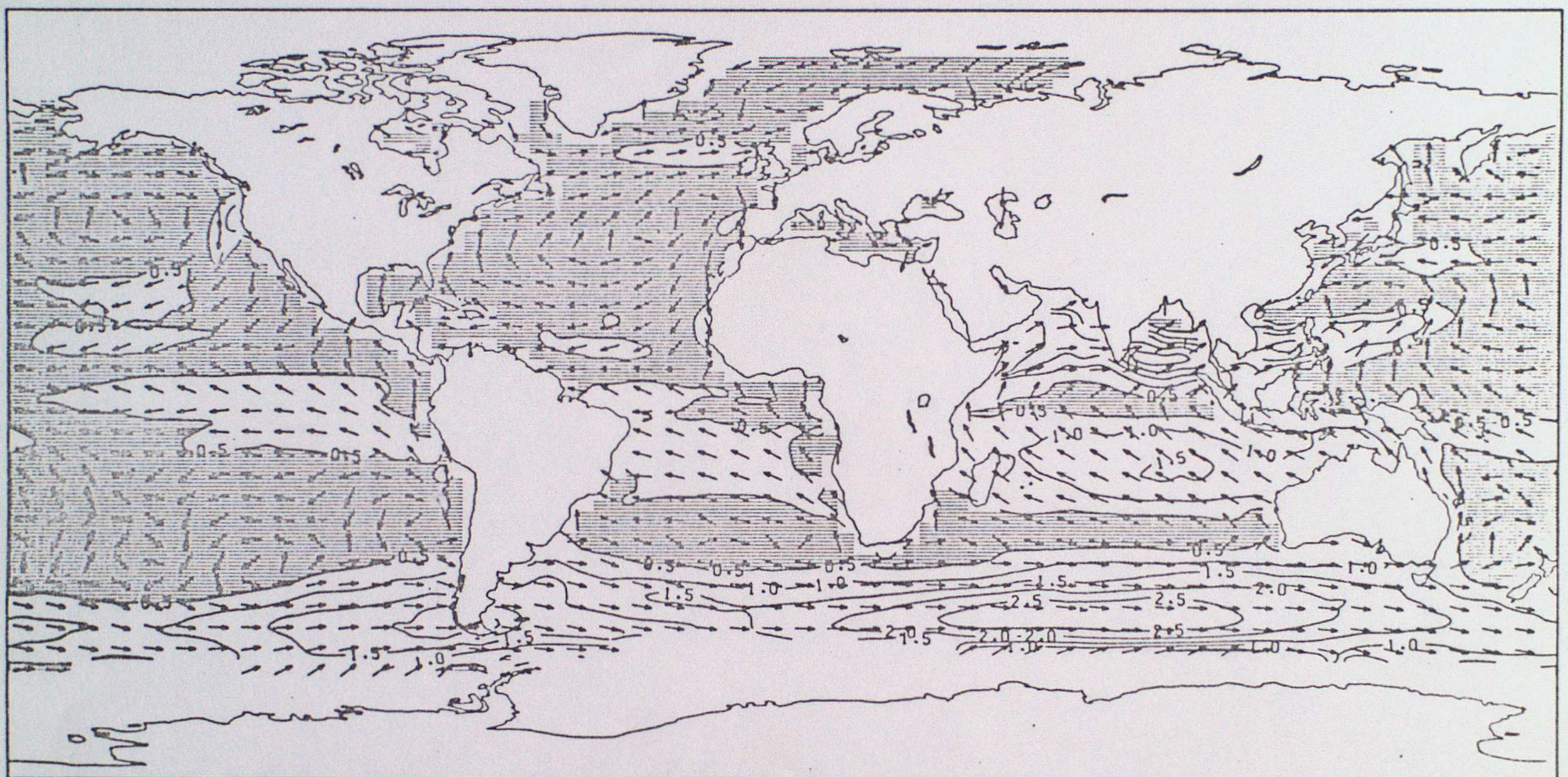
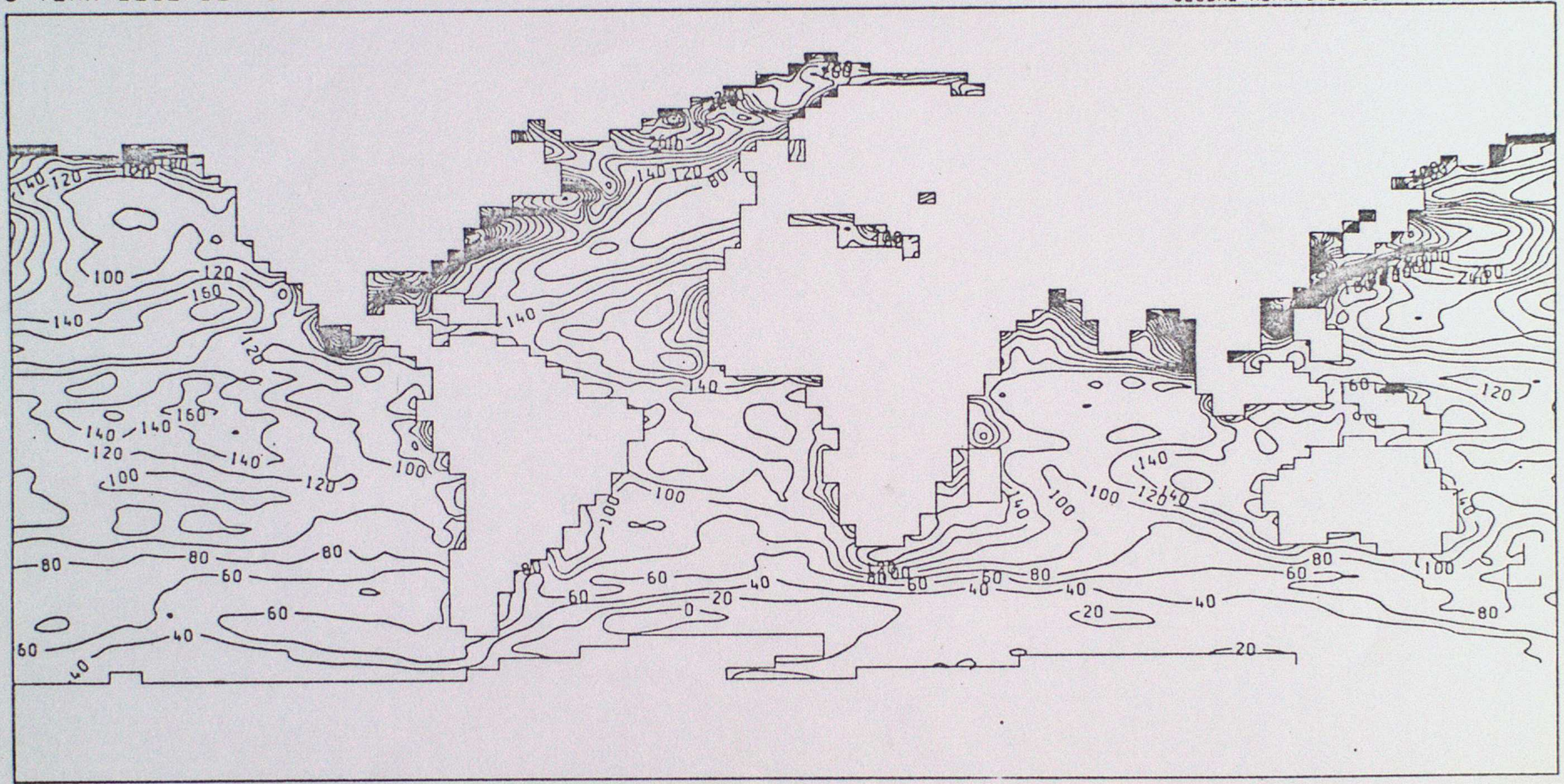


Figure 2b

TURBULENT HEAT FLUX - DIAGNOSED
3 YEAR DECEMBER.JANUARY.FEBRUARY

GLOBAL MEAN OVER SEA POINTS 127.999

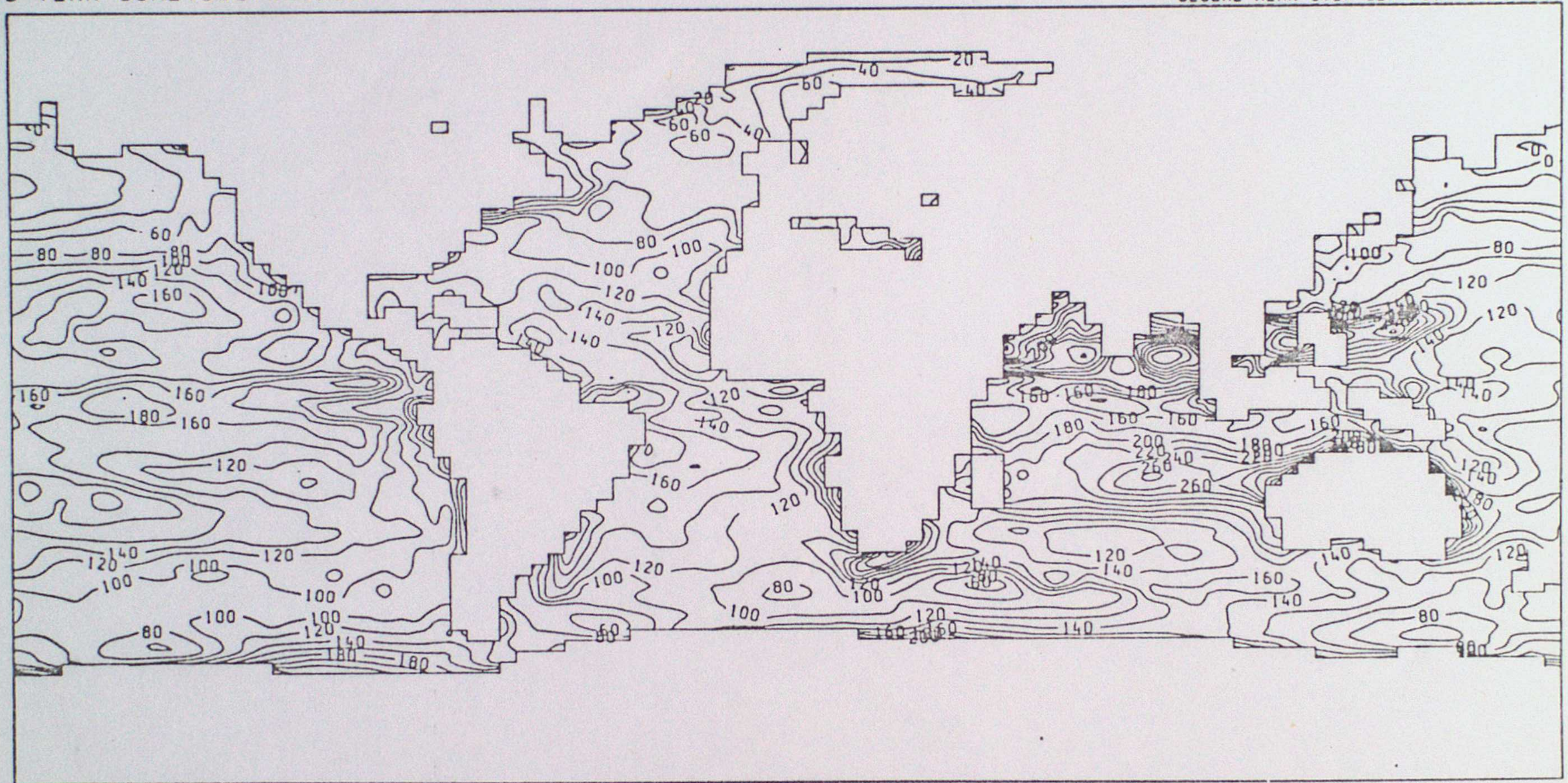


CONTOUR INTERVAL: 20 MKS UNITS

Figure 3a

TURBULENT HEAT FLUX - DIAGNOSED
3 YEAR JUNE.JULY.AUGUST

GLOBAL MEAN OVER SEA POINTS 128.396



CONTOUR INTERVAL: 20 MKS UNITS

Figure 3b

SURFACE WIND STRESS DERIVED-DIAGNOSED
JANUARY MEAN

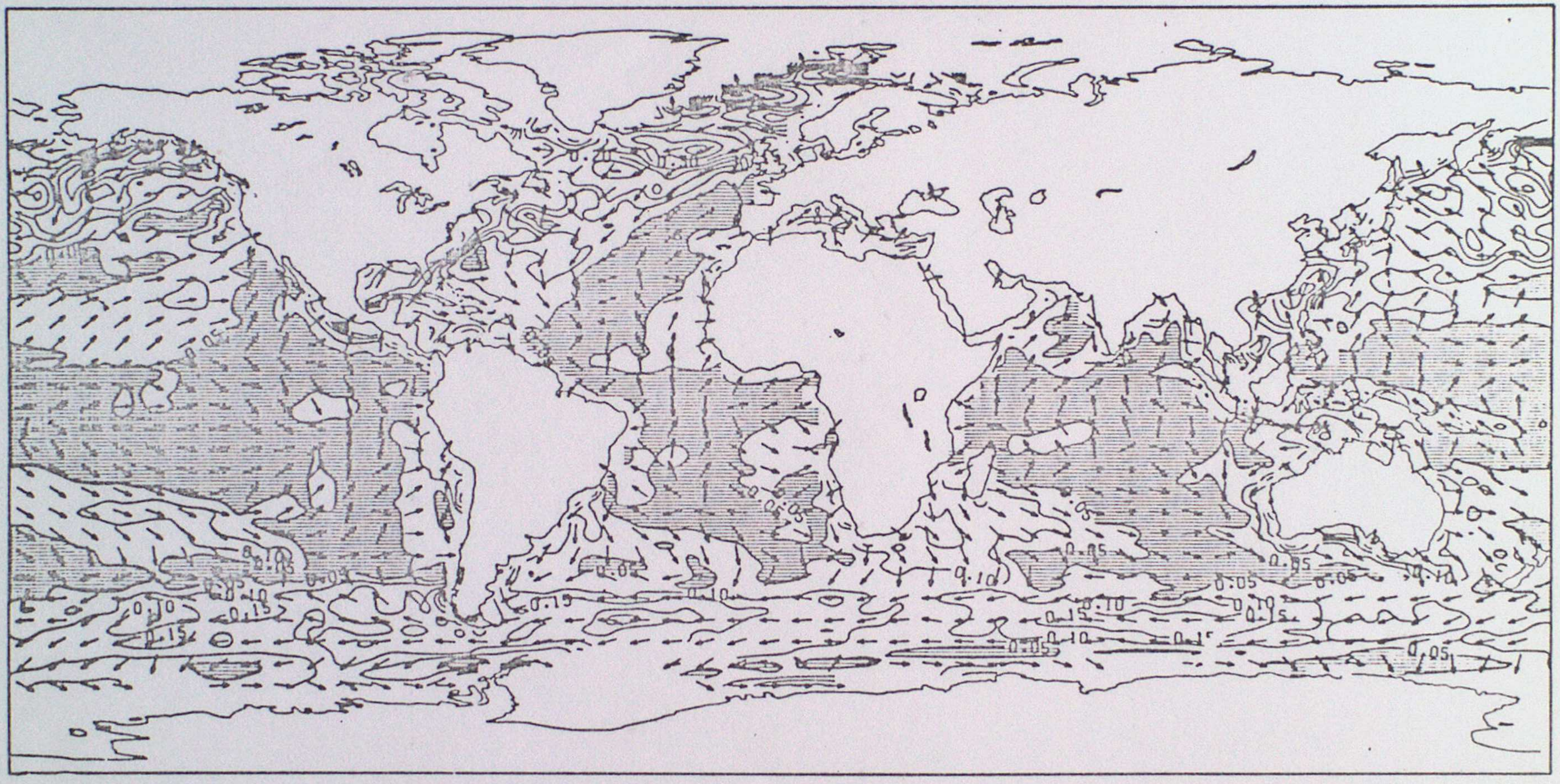


Figure 4a

SURFACE WINDS STRESS   DERIVED-DIAGNOSED
JULY MEAN

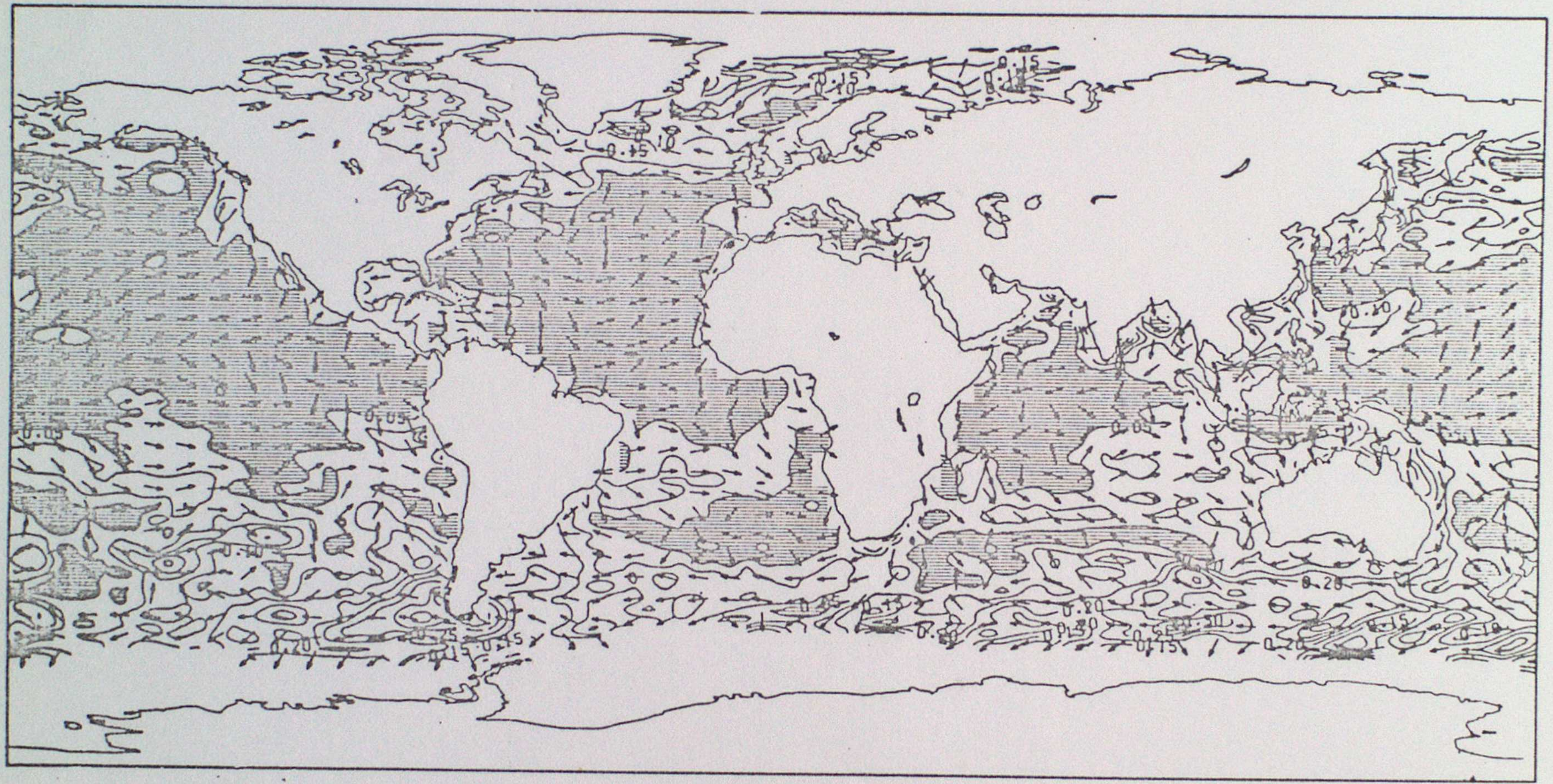


Figure 4b

Figure 5a

TURBULENT HEAT FLUX - DERIVED-DIAGNOSED
3 YEAR DECEMBER.JANUARY.FEBRUARY

GLOBAL MEAN OVER SEA POINTS -0.414

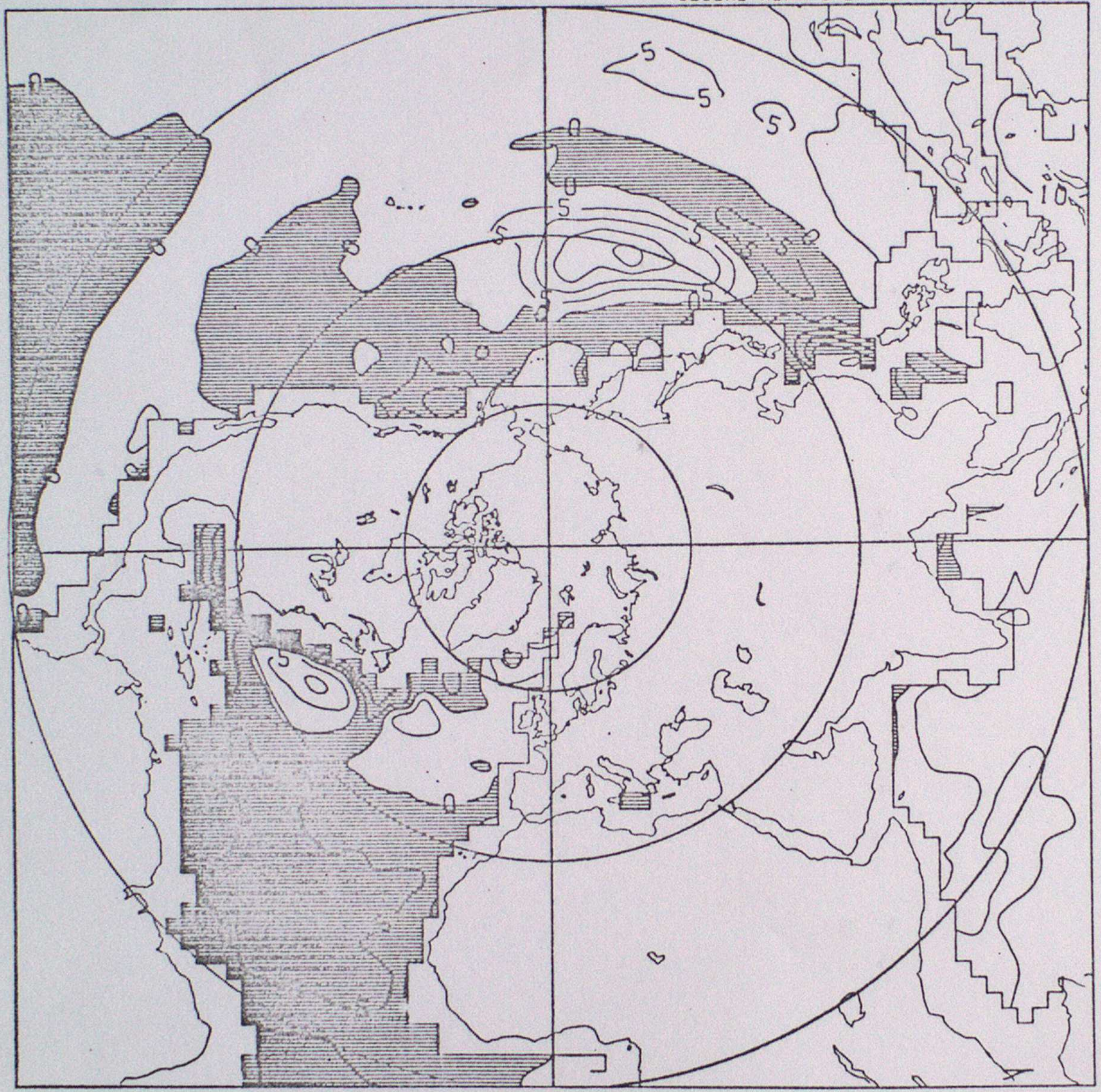


Figure 5b

TURBULENT HEAT FLUX - DERIVED-DIAGNOSED
3 YEAR DECEMBER.JANUARY.FEBRUARY

GLOBAL MEAN OVER SEA POINTS -0.414

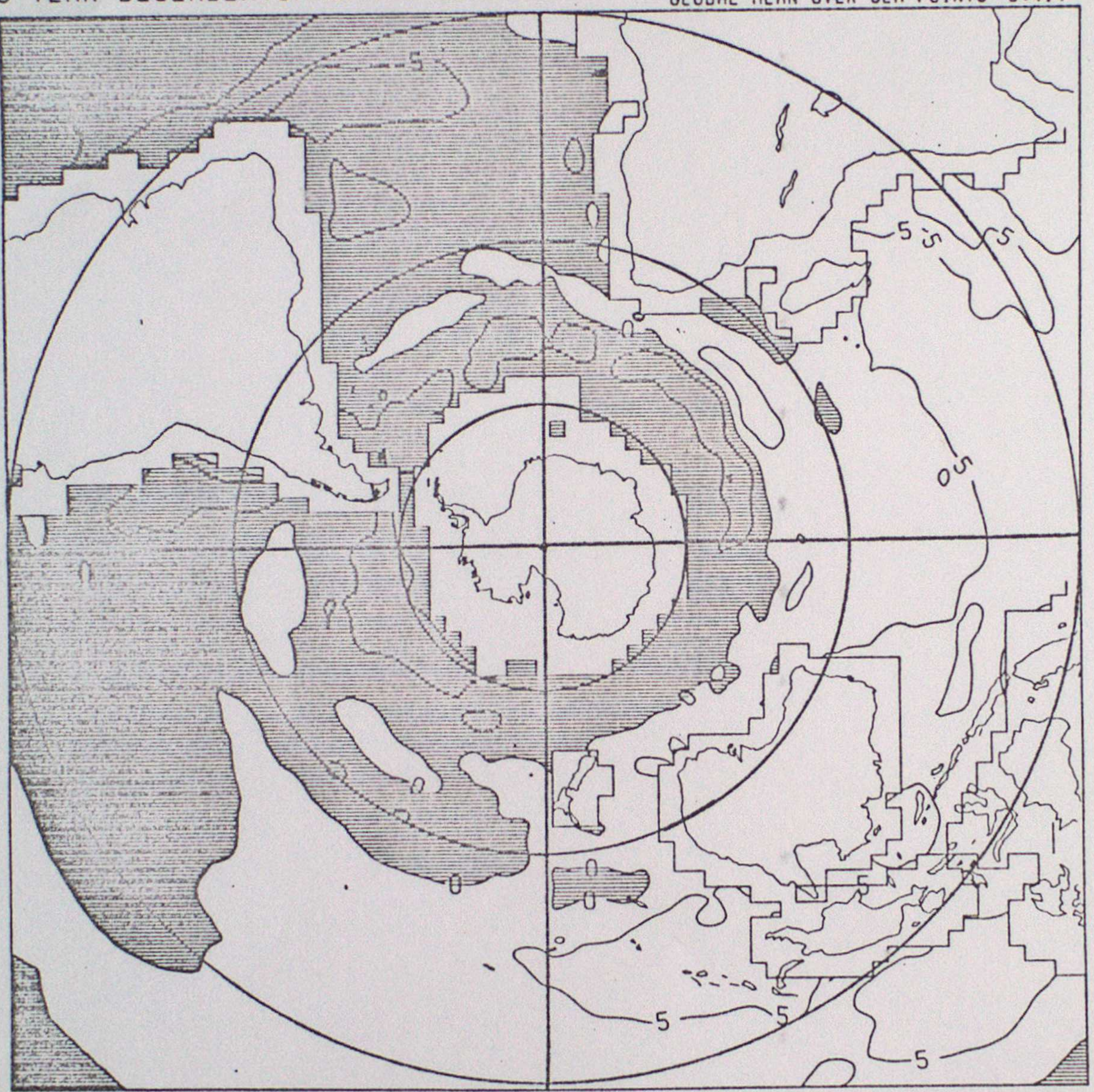


Figure 6a

TURBULENT HEAT FLUX - DERIVED-DIAGNOSED
3 YEAR JUNE.JULY.AUGUST

GLOBAL MEAN OVER SEA POINTS -0.145

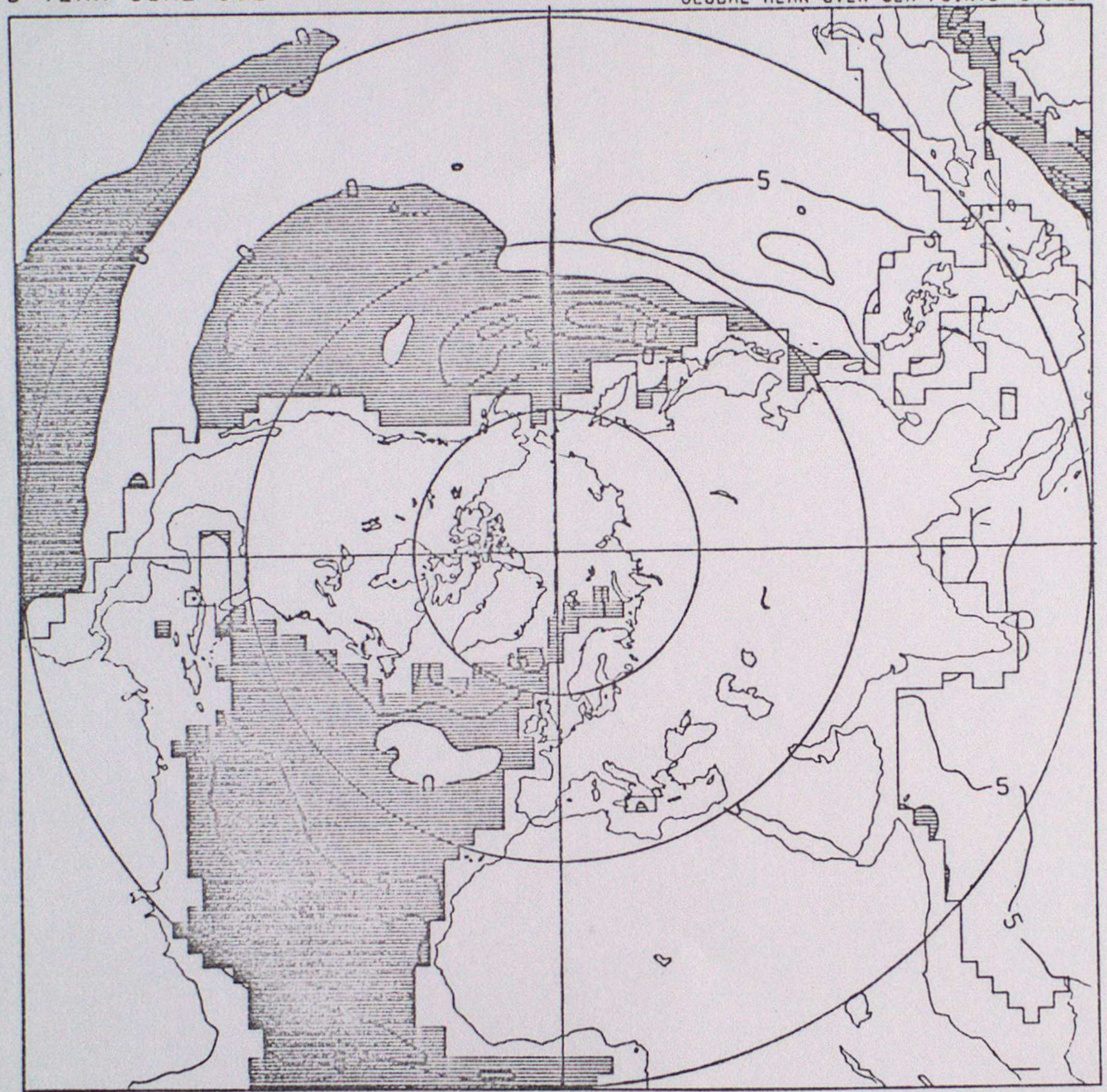


Figure 6b

TURBULENT HEAT FLUX - DERIVED-DIAGNOSED
3 YEAR JUNE.JULY.AUGUST

GLOBAL MEAN OVER SEA POINTS -0.145