OCEANOLOGICA ACTA, 1987, N° SP

Seasonal heat content variations in the Northwestern Indian Ocean

Heat content
Indian Ocean eddies
Monsoons
Seasonal variations

Contenu thermique
Tourbillons de l'Océan Indien
Moussons
Variations saisonnières

John G. BRUCE
Naval Oceanographic Office,
Bay St. Louis, NSTL Station, MS 39522-5001, USA.

Received 17/6/85, in revised form 16/6/86, accepted 24/6/86.

ABSTRACT

The changes of heat storage of the Somali Basin in the western Indian Ocean, as shown from 55 detailed temperature sections along the western sea lane (tanker XBT program 1975-1979), are only partly accounted for by the net heat gain at the sea surface. Vertical movement of the thermocline and horizontal advection appear to be the cause of major changes both between and during the monsoons. During the southwest monsoon the large amount of heat stored in the 0 to 400-m layer during northern spring (up to $31 \times 10^8\ Jm^{-2}$) is redistributed with a heat loss of approximately $13 \times 10^8\ Jm^{-2}$ in the upper 0 to 100-m layer and the gain of a similar amount in the 100 to 400-m layer, caused to a large extent by the dynamic response of these layers to the monsoon wind stress. In late autumn this heat is then advected from the region.

Oceanol. Acta, **1987.** Proceedings International Symposium on Equatorial Vertical Motion, Paris, 6-10 May 1985, 77-83.

RÉSUMÉ

Variations saisonnières du contenu thermique dans le nord-ouest de l'Océan Indien

Les variations du contenu thermique du bassin de Somalie, déterminées à partir de 55 profils détaillés de température le long de la route occidentale de navigation (programme XBT des navires-pétroliers, 1975-1979), sont expliquées en partie seulement par l'apport net de chaleur à la surface de la mer. L'advection horizontale et verticale paraît être la cause des changements importants, à la fois entre et pendant les moussons. Pendant la mousson du Sud-Ouest, la quantité considérable de chaleur emmagasinée dans la couche de 0 à 400 m pendant le printemps de l'hémisphère Nord (jusqu'à $31 \times 10^8\ Jm^{-2}$) est redistribuée avec une déperdition de chaleur d'environ $31 \times 10^8\ Jm^{-2}$ dans la couche superficielle de 0 à 100 m, et le gain d'une quantité équivalente dans la couche de 100 à 400 m, causée dans une large mesure par la réponse dynamique de ces couches à la force de la mousson. Vers la fin de l'automne cette quantité de chaleur quitte la région par advection.

Oceanol. Acta, **1987.** Proceedings International Symposium on Equatorial Vertical Motion, Paris, 6-10 May 1985, 77-83.

INTRODUCTION

Indian Ocean heat storage is greatly influenced by the forcing signal of the semiannual monsoons. Because of the extreme conditions of evaporation and wind stirring at the sea surface imposed by the monsoons in the Somali Basin region and the intense solar heating between monsoons during northern spring and autumn, the seasonal changes in net positive heat gain through the sea surface can be relatively large. Various investigators have pointed out, however, that the local heat exchange with the atmosphere can hardly account for the large amounts of heat stored seasonally in the upper ocean at low latitudes. Düing and Leetmaa (1980) suggest that Arabian Sea coastal upwelling is a major cause of heat storage variability. In the western equatorial Atlantic, Merle (1980) points out that the annual cycle of heat content appears to be the result of vertical movement of the thermocline associated with the dynamical response of the ocean to the seasonally varying winds. In fact, he shows that the rate of heat storage change can be up to ten times larger than the net input through the sea surface. Vertical movement of the thermocline as a means of changing heat content is shown to be important by Emery (1976) also. He correlated heat content in the upper layer (0 to 250 m) with the depth of the 14 °C isotherm of the northeastern Pacific Ocean. In the northwestern Indian Ocean, because of the strong currents associated with the eddy field circulation during the monsoons (Bruce, 1979), horizontal advection is also of significance. An example of the configuration and horizontal extent of the eddies in the Somali Basin is indicated in Figure 1.

Heat absorbed from the atmosphere by the ocean in tropical regions has to be exported meridionally out of the region to higher latitudes. In the case of the Arabian Sea and Somali Basin which extend to about 25°N, most of this heat must leave southward across the Equator.

METHODS

In order to examine what is probably the region subjected to the greatest seasonal variability in the Arabian Sea, a series of 55 temperature sections of 63 XBT stations each (0-450-m depth) have been examined. Exxon tankers were used as ships of opportunity along the western sea lane enroute from Cape Town to the Persian Gulf (Bruce, 1979) between October 1975 to December 1979. The sections along the sea-lane (Fig. 1) extend from 2°S to 22°N and station spacing is relatively close particularly in the Somali Basin region (approximately 15 to 20 nautical miles), thus the small-scale details of the boundaries with strong temperature gradients occurring during periods of maximum upwelling in the southwest monsoon are sampled. As will be seen, it becomes important when estimating the heat content in a region of strong currents and eddies, that closely spaced sampling be obtained.

The stations falling only along the 2 to 12°N portion of the sea-lane have been used in order to keep sampling within the Somali Basin and north of the Equator. The eddies which develop during the southwest monsoon within the Somali Basin, particularly the northern eddy (Bruce, 1979), seasonally have strong horizontal gradients and vertical displacements of the thermocline. From the XBT sections used here

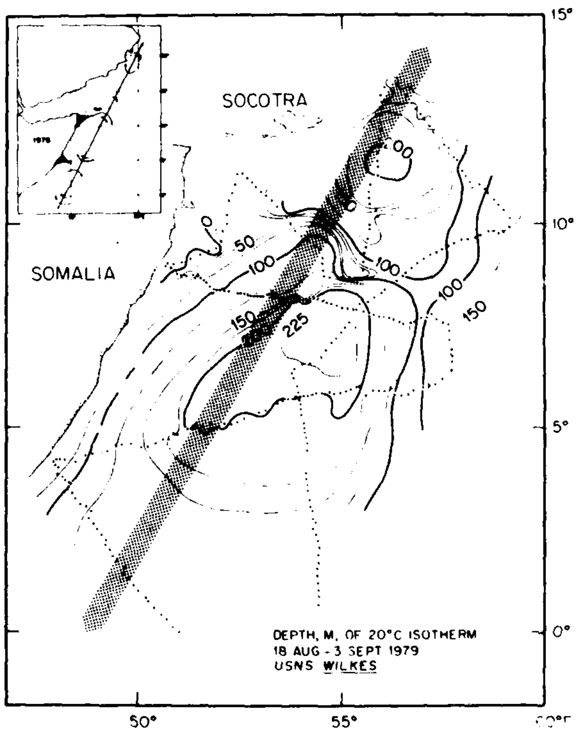

Figure 1

Western Indian Ocean sea-lane used by Exxon tankers along which XBT stations were located shown as long line paralleling coast in inset (upper left) and as stippled arrow in main figure. Inset shows estimate of eddy patterns during 1976 southwest monsoon. Main figure shows eddy pattern as indicated by depth of 20°C isotherm during the 1979 southwest monsoon. Dots show XBT station locations.

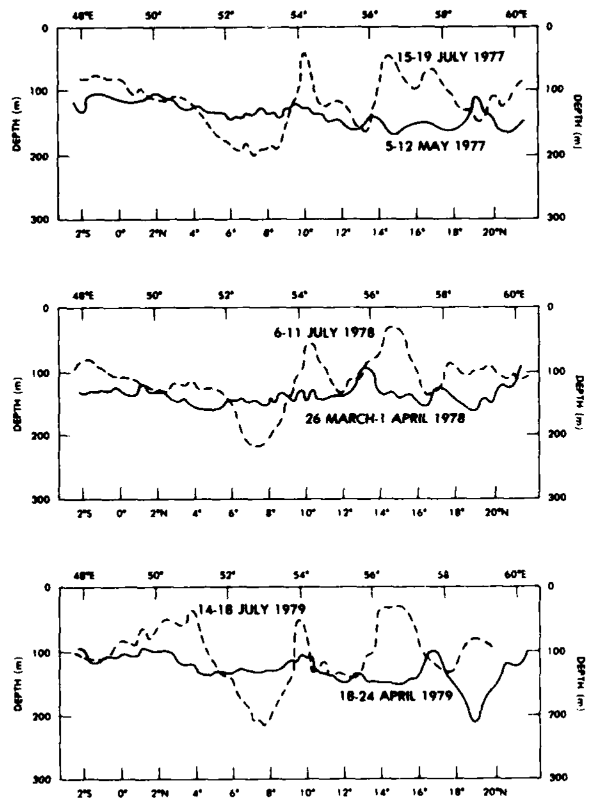

Figure 2

Examples of depth (m) of 20°C isotherm (mid-thermocline) of pre-monsoon (March, April, May) and southwest monsoon (July) conditions along tanker sea-lane (Fig. 1).

(Bruce, 1981) and other surveys (Bruce, 1968 ; 1973 ; Swallow *et al.*, 1983) there is convincing evidence that the northern eddy occurs each year. It usually starts to develop in the surface layer during May, reaches maximum energy in July and August, with horizontal dimensions of approximately 400 to 600 km, and remains in the northern Somali Basin until approximately December (and during some years remains through the following northeast monsoon (Bruce, Volkmann, 1969). The large Socotra eddy (Bruce, 1979) appears annually, as well as the variable eddy structure developing to the east of the northern Somali eddy (Bruce, Beatty, 1985), to the northeast of the Socotra eddy, the eastern part of the Gulf of Aden, and off the Arabian coast. Thus the Somali Basin eddy region, although considerably more energetic than other regions, might be considered representative of the pronounced southwest monsoon eddy field in the western Indian Ocean. The 2 to 12°N section used here extends into the colder water (approximately 4-5°N) to the south of the eddy. This southern cold water, associated with the southern turn-off region of the Somali current, can be relatively well developed during certain years such as 1979 (Swallow *et al.*, 1983).

A comparison of the vertical changes in the thermocline that occur in pre-monsoon conditions with those found during the southwest monsoon is shown in Figure 2. Because of dynamic forcing in response to the wind, numerous regions of upwelled colder water into the upper thermocline and mixed layer occur in the boundary zones between anticyclonic eddies, while within the eddies the warm water extends downward deeper than the mean pre-monsoon thermocline depth.

HEAT GAIN AT SEA SURFACE

The seasonal distribution of net heat gain at the sea surface with strong semiannual peaks is the result of gain from long-wave minus short-wave radiation and also minus evaporative and sensible heat loss. Of the latter two, evaporative loss is by far the larger, having maxima during each monsoon, thus reducing the net heat gain to approximately zero values at these times.

The net heat gain at the sea surface from different sources is given in Figure 3. There is relatively close agreement between the Bunker (1976) and Colborn (1975) curves at the semiannual maxima, whereas there is slightly closer agreement with the Bunker and H & L (Hastenrath, Lamb, 1979) curves at the minimum values. There is considerable spacial variability in the heat gain values mapped in the Somali Basin (*see* Hastenrath, Lamb, 1979), particularly during the monsoon period, and thus these means for the basin would have a relatively high standard deviation at this time (estimated at approximately ± 20 % of the mean). The Bunker values (unpublished for the Indian Ocean) determined by the method discussed by Bunker (1976) have been used in this study and were felt to be the most accurate for depicting heat gain in this region. The net annual heat gain at the sea surface for the Somali Basin is positive amounting to : Bunker, 80 Wm^{-2} ; Colborn, 67 Wm^{-2} ; and Hastenrath and Lamb, 58 Wm^{-2}. This gain must then be exported from the region by various means.

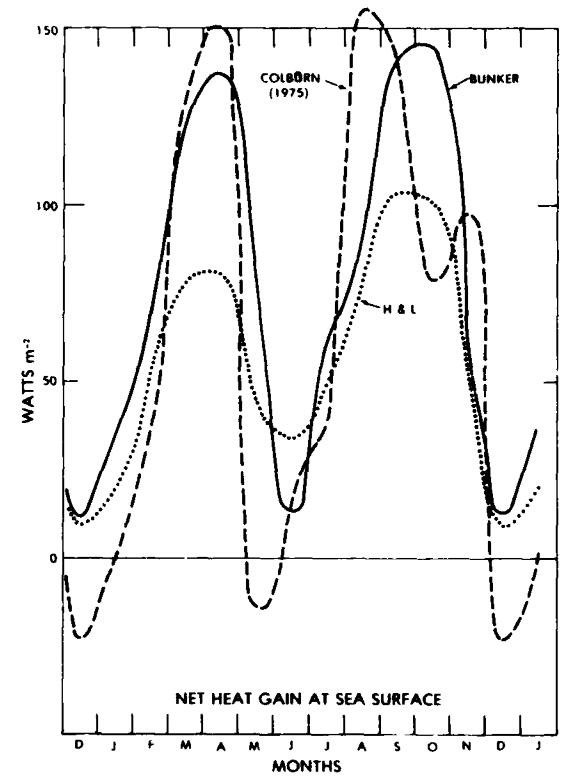

Figure 3

Net heat gain at the sea surface in Somali Basin. Bunker distribution calculated according to Bunker (1976) within basin region approximately 2°N-12°N ; H & L is Hastenrath and Lamb (1979), their Figure 16 ; Colborn (1975), his Figure 25.

HEAT CONTENT

The seasonal changes in the heat content of the upper layer (0 to 100-m) for the 1975-1979 observational period are shown in Figure 4. In this layer the largest of the positive semiannual signals commences at the end of each northeast monsoon during the time of the northern spring sea-surface heat gain (Fig. 3). After

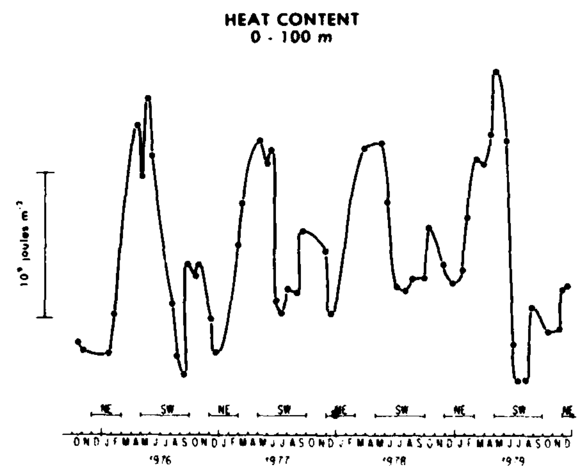

Figure 4

Heat storage from XBTs in the 0-100 m layer between 2°N-12°N in the Somali Basin from October 1975 to December 1979 along tanker sea-lane (Fig. 1) with duration of northeast (NE) and southwest (SW) monsoons given. Vertical scale length is 10^9 Jm^{-2}.

the southwest monsoon begins to develop, a large heat loss in this layer occurs, reaching minimum heat content toward the end of the southwest monsoon. Also at this time the surface heat gain approaches a minimum. The second of the semiannual signals occurs during the northern autumn surface heat-gain maximum and then decreases near the start of the northeast monsoon.

The monthly means of the heat content from the 1975-1979 sections for the 0 to 100-m, 100 to 200-m, and 200 to 400-m layers, each adjusted to its annual mean (Fig. 5), all indicate a large January-February

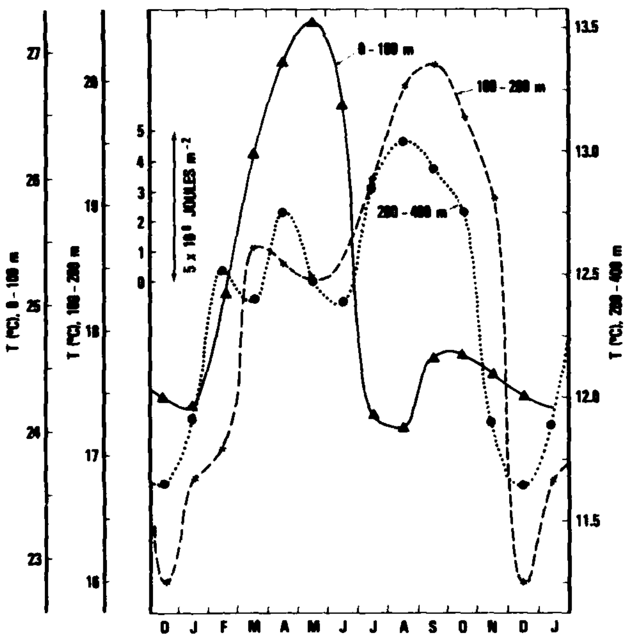

Figure 5

Heat storage 0-100 m, 100-200 m and — 200-400 m layers (indicated by mean layer temperature, °C) between 2°N-12°N in the Somali Basin (Fig. 1) from average of October 1975 to December 1979 XBT sections. Scale represented by arrow (5 × 10^8 joules m^{-2}) applies to all curves.

increase. However, after the February-March period the layers below 100 m change relatively little until June when the southwest monsoon has commenced. At this time it appears that heat previously stored in the 0 to 100-m layer is then distributed downward. By the middle of the southwest monsoon (July) a loss of approximately 13 × 10^8 Jm^{-2} in the 0 to 100-m layer and a gain of approximately 13 × 10^8 Jm^{-2} in the 100 to 400-m layers occurs, suggesting a pronounced vertical transfer of heat downward to a large extent by the ocean's dynamic response to the monsoon. The strong sea surface (and mixed layer) warming during northern spring followed by the downward transfer of heat to depths of 200 m and below during the late southwest monsoon (July-September) is clearly illustrated in Figure 6.

HEAT CONTENT RATE

The rate of heat-storage change minus the sea-surface heat gain (Fig. 7) indicates the degree of convergence or advection of heat during the year. In the entire 0 to 400-m layer maximum convergence (heat import) occurs during January-March, with values reaching over 280 Wm^{-2}, to a large extent caused by heat import in the 100 to 200-m layer. Note that the 0 to 100-m layer during December-April has relatively little input from advection (amounting to slightly over 60 Wm^{-2}, probably negligible) indicating that in this layer the net surface heat gain (December-April) of approximately 10 × 10^8 Jm^{-2} from the atmosphere accounts for nearly all of its increase in heat content of approximately 11 × 10^8 Jm^{-2}. The heat gain of the 0 to 400-m layer from December-April amounts to approximately 31 × 10^8 Jm^{-2} or about 3 times the heat furnished from the atmosphere.

During June-August the downward movement of heat from the 0 to 100-m layer to the deeper layers is shown, suggesting that a relatively small amount of heat is exported horizontally from the region, whereas the vertical movement of the thermocline becomes important at this time. The reverse appears to be the case during the October-December period when a large heat export by horizontal advection from the 0 to 400-m layer amounting to over 600 Wm^{-2} occurs, which is comparable to the Gulf Stream heat export at the sea surface during January-February (Bunker, 1976). If the net heat gained at the sea surface for the September-December period amounting to approximately 13 × 10^8 Jm^{-2} is added to that lost in the 0 to 400-m layer, approximately 27 × 10^8 Jm^{-2}, the total comes to approximately 40 × 10^8 Jm^{-2} which must be advected from this region. Relatively small losses, however, occur in the 0 to 100-m layer, as may be seen.

In part, heat might be advected away by subsurface southward flows such as those described by Quadfasel and Schott (1983). Strong zonal flow near the Equator also might remove heat as suggested by the semiannual maxima in surface dynamic height (1°N-1°S) along the tanker sea-lane where it crosses the Equator (Fig. 8). From the equatorial region heat would then be advected into the southern hemisphere. Moored current meter records on and near the Equator indicate a pronounced semiannual signal of eastward and westward flow which extends considerably deeper than the 400 m level discussed here (Luyten, Roemmich, 1982). The signal occurs as far east as Gan (approximately 73°E ; Knox, McPhaden, 1980).

CONCLUSIONS

It would appear that the energy of the eddy field, particularly during the southwest monsoon, plays a significant role in the storage and in the vertical and horizontal transfer of heat in the western Indian Ocean, especially in the Somali Basin. Merle (1980) has found a somewhat similar situation in the western equatorial Atlantic.

In the case of the Somali Basin, there appear to be three stages in the heat transfer process. First, the large import of heat commences near the beginning of the northeast monsoon. The 100 to 400-m layer appears to receive heat by advection, whereas in the 0

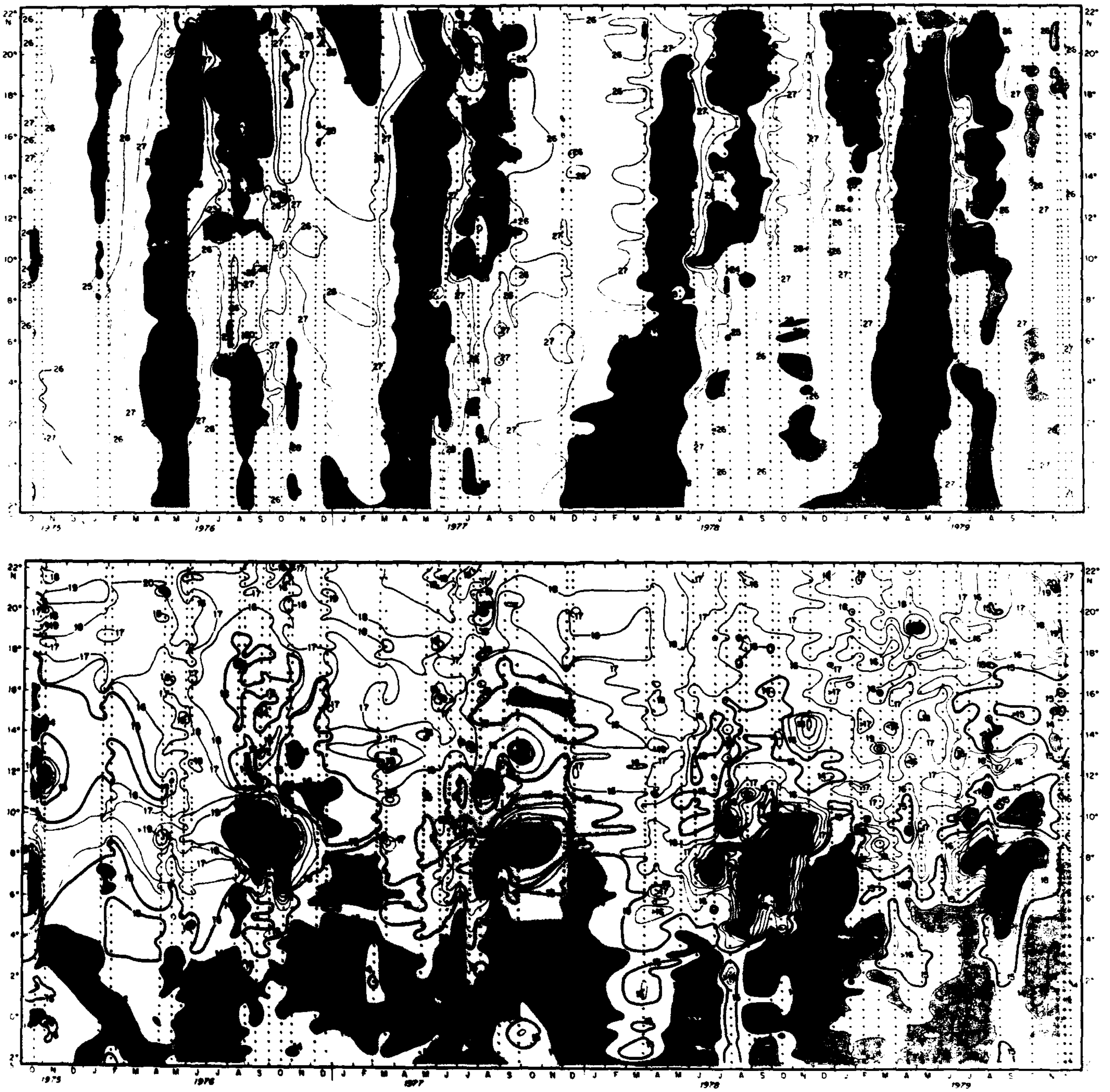

Figure 6

Temperature, °C, of sea surface (upper) and at 200 m depth (lower) along XBT sections from 2°S-22°N (see Fig. 1) for period from October 1975 through December 1979.

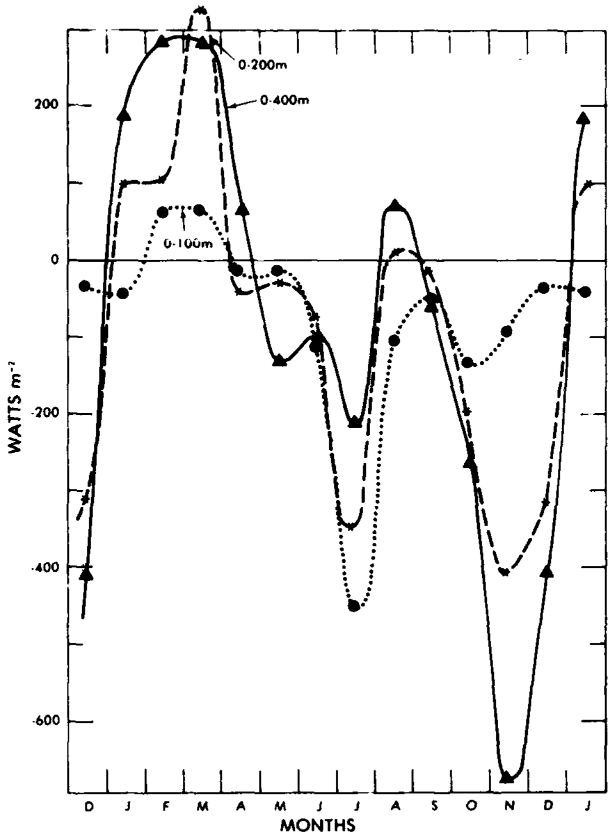

Figure 7

Rate of heat storage minus sea surface heat gain from Bunker's values (Fig. 3). Curves represent residual caused by horizontal advection and vertical displacement and diffusion of heat.

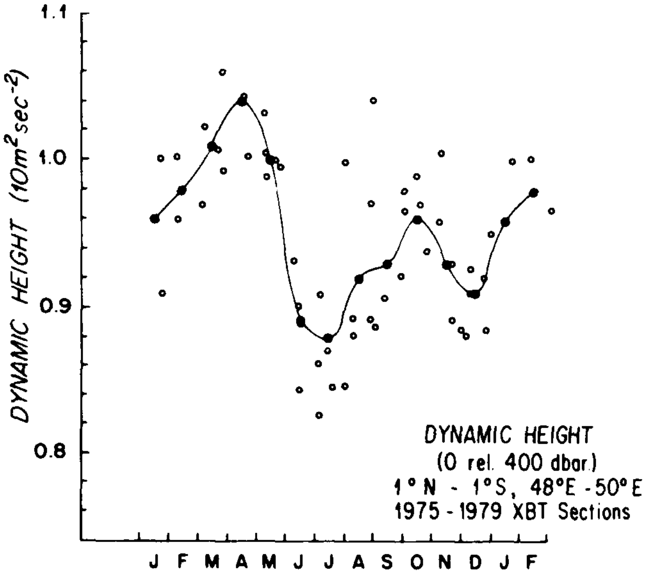

Figure 8

Geopotential anomaly of sea surface (0 rel. 400 dbar) at equator from 1975-1979 tanker XBT observations. Fixed T-S relation used in determinations (Bruce, 1979). Curves connects monthly means (solid dots). Standard deviation of " individual " observations (open dots) is ± 0.04 ($10\ m^2\ s^{-2}$). Each " individual " observation is generally an average of 5 stations between 1°S-1°N along sea-lane (Fig. 1).

to 100 m layer, heating correlates well with the strong net surface gain in northern spring. Next, as the southwest monsoon commences, much of the heat stored in the upper 100 m is moved vertically downward to deeper layers while cooler upwelled water, uplifted by the dynamic forcing of the wind, removes heat from the upper layer (Fig. 9). Lastly, in the third stage, post-southwest monsoon cooling commences in the deeper layers with relatively rapid advection of heat from the 100 to 400-m layer.

Acknowledgements

Partial support by Office of Naval Research Contract N00014-76-C-0071, NR 083-004. Also thanks are due to the University of Cape Town Oceanography Dept. and Exxon Corp. for help in data collection.

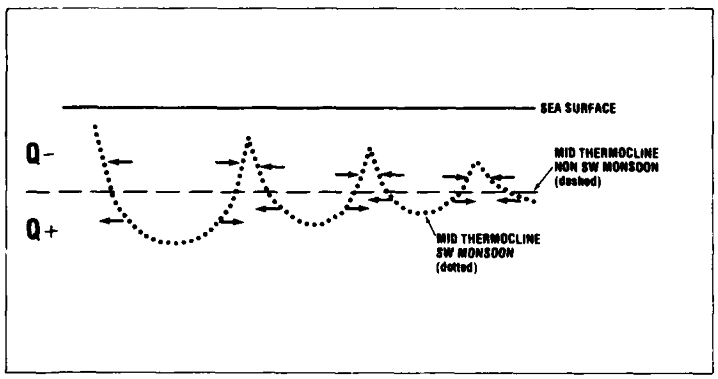

Figure 9

Schematic diagram suggesting mechanisms of heat redistribution occurring during the southwest monsoon within the Somali Basin-Northwestern Indian Ocean thermocline (see Fig. 2). Large eddy on left would represent the northern eddy in the Somali Basin (" Great Whirl "), adjacent would be the Socotra eddy, then to the right the smaller Arabian Sea eddies lying to the northeast. Cold upwelled water now above the previous mid-thermocline level (dashed) absorbs heat (Q−) as shown by arrows, whereas heat flows out (arrows) from base of warm eddies, thus warming (Q+) lower thermocline layer.

REFERENCES

Bruce J. G., 1968. Comparison of near surface dynamic topography during the two monsoons in the Western Indian Ocean, *Deep-Sea Res.*, **15**, 665-677.

Bruce J. G., 1973. Large scale variations of the Somali current during the southwest monsoon, 1970, *Deep-Sea Res.*, **20**, 839-846.

Bruce J. G., 1979. Eddies off the Somali coast during the southwest monsoon, *J. Geophys. Res.*, **84**, 7742-7748.

Bruce J. G., 1981. Variations in the thermal structure and wind field occurring in the western Indian Ocean during the monsoons, Technical Report 272, Naval Oceanographic Office, NSTL, MS 39522-5001, USA.

Bruce J. G., Volkmann G., 1969. Some measurements of current off the Somali coast during the northeast monsoon, *J. Geophys. Res.*, **74**, 8, 1958-1967.

Bruce J. G., Beatty W., 1985. Some observations of the coalescing of Somali eddies and a description of the Socotra eddy, *Oceanol. Acta*, **8**, 2, 207-219.

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

Colborn J. G., 1975. *The thermal structure of the Indian Ocean*, Univ. Press of Hawaii, 173.

Düing W., Leetmaa A., 1980. Arabian Sea cooling : a preliminary heat budget, *J. Phys. Oceanogr.*, **10**, 307-312.

Emery W. J., 1976. The role of vertical motion in the heat budget of the upper northeastern Pacific Ocean, *J. Phys. Oceanogr.*, **6**, 299-305.

Hastenrath S., Lamb. P. J., 1979. *Climatic atlas of the Indian Ocean, Part II*, The University of Wisconsin Press, 93.

Knox R. A., McPhaden M. J., 1980. Profiles of velocity and temperature near the Indian Ocean equator, SIO Ref. Ser. No. 80-1, Scripps Inst. Oceanogr., Univ. California, USA.

Luyten J. R., Roemmich D. H., 1982. Equatorial currents at semi-annual period in the Indian Ocean, *J. Phys. Oceanogr.*, **12**, 406-413.

Merle J., 1980. Seasonal heat budget in the equatorial Atlantic Ocean, *J. Phys. Oceanogr.*, **10**, 464-469.

Quadfasel D. R., Schott, F., 1983. Southward subsurface flow below the Somali current, *J. Geophys. Res.*, **88**, 5973-5979.

Swallow J. C., Molinari R. L., Bruce J. G., Brown O. B., Evans R. H., 1983. Development of near-surface flow pattern and water mass distribution in the Somali basin in response to the southwest monsoon of 1979, *J. Phys. Oceanogr.*, **13**, 1398-1415.