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The applicability of the T/S method to geopotential anomaly computations in the Northeast Atlantic

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ABSTRACT

Methods are tested for computing geopotential anomalies from temperature data in the subtropical Northeast Atlantic. Mean temperature-salinity, salinity-depth and density-depth relationships are determined for 3 × 3° squares, using hydrographic data from World Oceanographic Data Centre A. Geopotential anomalies computed from observed temperatures and salinities from these mean relationships are compared with anomalies from the original temperature and salinity data. For 0-500 dbar, geopotential anomalies can be well approximated, and the methods also work reasonably well for 0-1 000 dbar. The approximation is poor for 0-2 000 dbar. Appropriate methods for obtaining the best results in each 3 × 3° square are specified. The method is applied to a particular subset of the data.

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RÉSUMÉ

Application de la méthode T/S aux calculs d'anomalies géopotentielles dans le nord-est de l'Atlantique.

Des méthodes de calcul des anomalies géopotentielles sont appliquées à des données de température dans le nord-est de l'Atlantique subtropical. Les relations moyennes température-salinité, salinité-profondeur et densité-profondeur, sont déterminées pour des carrés de 3 × 3°, en utilisant les données hydrographiques du Centre Mondial A des Données Océanographiques. Les anomalies géopotentielles moyennes calculées à l'aide de ces relations à partir des seules températures observées sont comparées aux anomalies issues des données complètes de température et salinité. De 0 à 500 dbar, les anomalies géopotentielles sont déterminées avec une bonne précision, et la méthode peut encore s'appliquer de 0 à 1 000 dbar. Par contre, de 0 à 2 000 dbar, l'approximation est mauvaise. Les méthodes donnant les meilleurs résultats dans chaque carré de 3 × 3° sont résumées. La méthode est appliquée à un ensemble particulier de données.

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INTRODUCTION

It is usually assumed that the large-scale flow in the interior of the oceans is basically geostrophic. The shape of vertical velocity profiles can then be obtained from geopotential anomalies. Beyond that, recent developments in geostrophic current calculation methods (Stommel, Schott, 1977; Wunsch, 1978; Killworth, 1980) lead to absolute velocities and to transport estimates under certain conditions. But even the traditional method of determining geopotential anomalies relative to appropriate reference levels will often supply a reasonably good representation of the flow field.

A determination of the large scale circulation of the oceans is of relevance not only to physical oceanography, because the resulting transport contributes considerably to the global meridional heat flux (Oort, Van der Haar,

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1976) whose variations may be expected to be associated with climatic changes. Geostrophic transports will also influence the regional distribution of chemical substances, such as nutrients and oxygen, and the occurrence of small organisms in the sea.

In the North Atlantic, most observational programmes have dealt with the northern, western and equatorial parts (*see* Dietrich *et al.*, 1980; Warren, Wunsch, 1981) while knowledge of the circulation in the subtropical eastern Atlantic is still rather limited. Filling this gap would be much facilitated if the observations for determining geopotential anomalies could be reduced to temperature-depth measurements solely. This may be achieved by applying the T/S method, using the temperature-salinity relationship for a calculation of density from temperature data.

The idea of using the T/S method is not new, and was first used when the mechanical Bathythermograph began to supply large numbers of temperature profiles from the upper ocean. Stommel (1947) discussed the method, and Wyrtki and Kendall (1967) used it successfully to determine transport rates in the Pacific Equatorial Countercurrent. But only in recent years, with the advent of the XBT (Expendable Bathythermograph), have systematic studies of the validity of the T/S relationship for such applications in various regions been performed. Besides mean T/S curves (Emery, 1975; Emery, Wert, 1976), mean pressure-salinity (P/S) relationships (Emery, O'Brien, 1978) or—at boundaries between water masses—T/S relationships at selected pressure levels (Flierl, 1978) were used. The results suggest that regional differences in water mass distribution will determine the applicability of such methods. The following study is designed to testing the usefulness of mean T/S or P/S relationships for determining geopotential anomalies in the subtropical Northeast Atlantic.

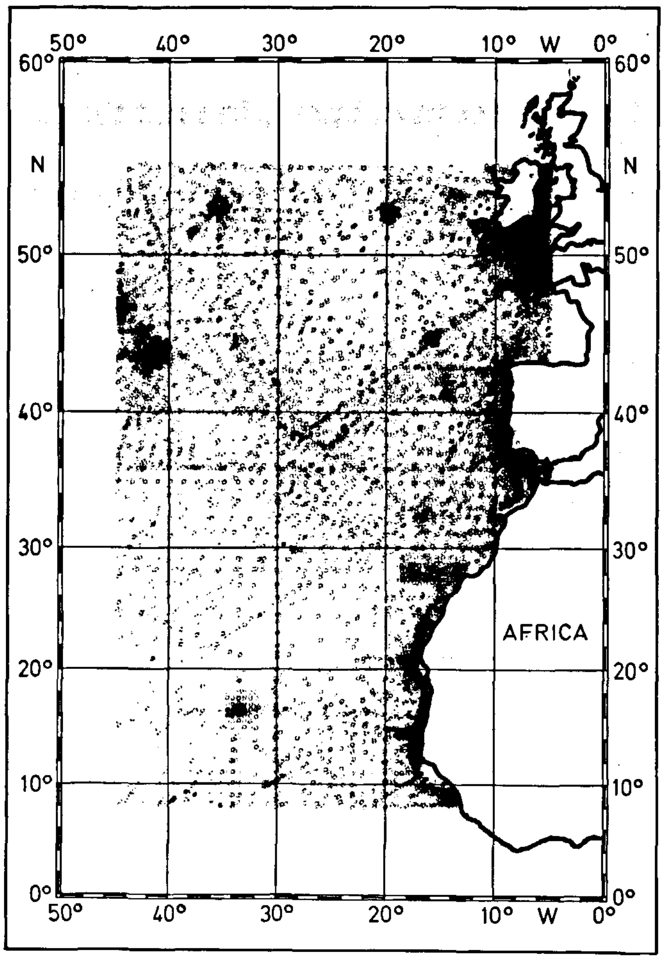

Figure 1

Distribution of hydrographic stations in WODC-A (1980) *data set in the area bounded by 8°N, 55°N and 5°W, 45°W.*

THE DATA SET

Hydrographic station density is highly variable in the Northeast Atlantic. The distribution of stations between 8°N and 55°N latitude and 5°W and 45°W longitude whose data were archived at World Oceanographic Data Centre (WODC) A in 1980 is shown in Figure 1. Except for near-coastal positions, station density is low in the subtropics, particularly in the eastern part. Because of the low number of stations in the region of interest, it was decided to use the original WODC data set as a basis, despite its data quality deficiencies, rather than the already edited sets with reduced numbers of stations which exist at some laboratories.

The region selected for this study is shown in Figure 2. Subsets in $3 \times 3°$ squares were used for testing the applicability of T/S and P/S relationships. In order to exclude shallow shelf water samples, only stations including data for depth levels deeper than 100 m were considered near the African coast. Also, stations without salinity data and with dubious salinities apparent from spikes or systematic deviations were rejected. This led to a reduction from 6232 to 4385 stations. The resulting number of stations per $3 \times 3°$ square are given in Figure 2.

In determining the mean T/S curves for each square, procedures were used which further eliminated data points of apparently poor quality. The following method was used: first, temperature data were grouped in one-degree intervals, and the mean salinity and standard deviation for each group were calculated for each square. Secondly, the temperature interval was reduced to 0.2°C, and all salinities exceeding twice the standard deviation determined in the first step were excluded before averaging. The resulting mean T/S distribution still displayed much scatter and was therefore smoothed with a procedure similar to that used by Emery and Wert (1976), applying twice a moving average of 7 data points. The procedure ensures a reasonable compromise between smoothing and preserving the main water mass characteristics of the area. To obtain smoothed data at 0.05°C intervals for geopotential anomaly computations, a spline procedure was applied for interpolating between the resulting T/S points. The results are summarized in Figure 3. A corresponding method was used to determine the mean P/S distributions at 2 dbar intervals. Details of the procedures can be found in Stramma (1981).

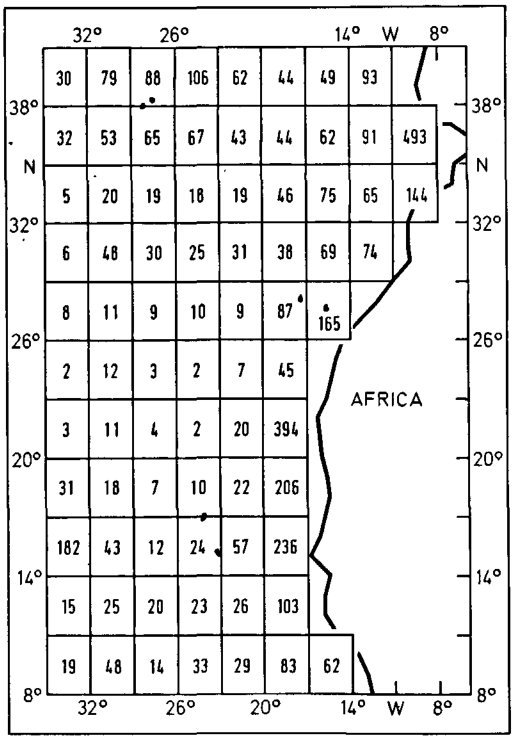

Figure 2
Area of this study with 3° x 3° squares and numbers of stations selected for determining mean relationships.

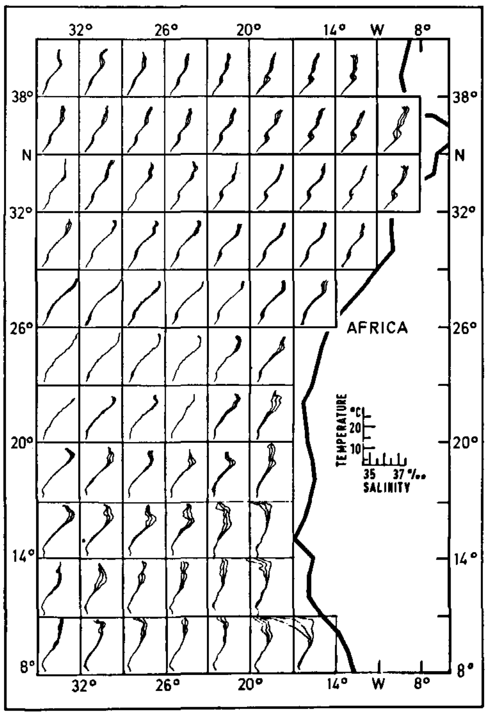

Figure 3
Mean temperature/salinity and standard deviation curves.

COMPUTATION OF GEOPOTENTIAL ANOMALY

Three different methods were used for computing geopotential anomalies. First, densities from temperature/salinity/depth triples led to "normal geopotential anomalies". Secondly, salinities were determined from temperature data by using the mean local T/S relationship, resulting in "T/S geopotential anomalies". Thirdly, salinities were derived from temperature data and the mean local P/S curve, leading to "P/S geopotential anomalies". To ensure appropriate data close to the reference levels 500, 1 000 and 2 000 dbar, only stations with data at a nominal depth corresponding to these levels ± 2 m were used. Another prerequisite was a minimum of 6 data levels with a maximum vertical separation of 250 m in the water column above the reference levels.

In addition to the geopotential anomaly d, the RMS deviation $\overline{\Delta d} = (\Delta d^2/(n-1))^{1/2}$ of the deviations Δd between normal and T/S or P/S geopotential anomalies for n profiles was computed. The term $\overline{\Delta d}$ provides a measure of the quality of approximation when using the T/S or P/S method. According to Fomin (1964), measurement errors lead to the following typical errors in the computation of normal geopotential anomalies: 1 dyn.cm (10^{-1} m^2.sec.$^{-2}$) for 0-500 dbar (10^{-2} MPa), 2 dyn.cm for 0-1 000 dbar, and 4 dyn.cm for 0-2 000 dbar. Additional errors will occur as a result of inadequate sampling in the presence of internal waves. It seems therefore appropriate to follow Emery (1975) and Emery and Wert (1976), considering the quality of approximation sufficient with $\overline{\Delta d} \leqslant 4$ dyn.cm with 500 or 1 000 dbar and $\Delta d \leqslant 5$ dyn.cm with 2 000 dbar reference level. It is conceded that such limits are somewhat arbitrary.

T/S relationships must be expected to be poor in the upper layers, due to surface heat and water fluxes. Indeed, temperatures and salinities in this data set vary strongly near the surface, but when comparing T/S points at selected depths it appeared that they were generally on the same density ($\sigma = [\rho - 10^3]$kg m^{-3}) line. This suggests that in those layers a mean σ/P curve might be appropriate for approximating the geopotential anomaly. In other words, although water mass properties change with time, geostrophic transports are not much affected by these changes in the upper layer. Mean σ/P curves for the depth range 0-400 m were calculated, using the method that had been applied for T/S and P/S curves. Applicability of the σ/P relationship was first tested in the square bounded by 14-17°N and 25-28°W. The error was minimized to $\overline{\Delta d} = 2.4$ dyn.cm when taking σ/P between 80 and 350 m and T/S elsewhere, reducing the error with the T/S method by one-third. Subsequently, this type of calculation was carried out for all 3 x 3° squares, leading to "σ/T/S geopotential anomalies". If $\overline{\Delta d}$ appeared unusually high in any of the determinations, the original data series used for the mean relationships were plotted and inspected. In several cases stations were identified which had spikes or systematic deviations from the typical values. Such apparently erroneous data were deleted. The results are summarized in Table 1. It appears that the T/S method leads to good results for 500 dbar in all but 3 squares (one near the coast and two near the Cape Verde Islands). The error is usually below 2 dyn.cm north of 23°N. South of this latitude, the strong variability in the tropical salinity maximum water introduces larger errors, but still below 4 dyn.cm.

Table 1

Number of 3 × 3° *squares with errors* $\overline{\Delta d}$ *below levels stated, using the different methods described in the text.*

Reference levels	0-500 dbar	0-1 000 dbar	0-2 000 dbar
Error levels	(≤4 dyn.cm)	(≤4 dyn. cm)	(≤5 dyn.cm)
No. of squares with at least one usable station	78	78	74
No. of squares with error above given level:			
T/S	3	16	34
P/S	22	44	50
σ/T/S	1	15	33
Best method (*see* Figures 4 *a*, *b*)	0	8	23

Not unexpectedly, the P/S relationship leads to much less satisfying results, with sufficient quality only in two-thirds of the 500 dbar calculations. The σ/T/S method usually leads to satisfactory results, with anomaly errors below 4 dyn.cm in all but one (near-coastal) square for this reference level. This third method improves the results considerably south of 23°N where the transition occurs between North and South Atlantic Central Water (Tomczak, 1981), leading to large variances in the respective T/S relations as indicated in Figure 3. When selecting the optimum method for each square, the error is below 4 dyn.cm everywhere: in the Azores-Canaries region, it is even ≤2 dyn.cm.

Results for 1 000 dbar are not as good; some 20% of the squares have errors ≥4 dyn.cm, with little improvement when changing from the T/S to the σ/T/S method. In addition to the tropical salinity maximum

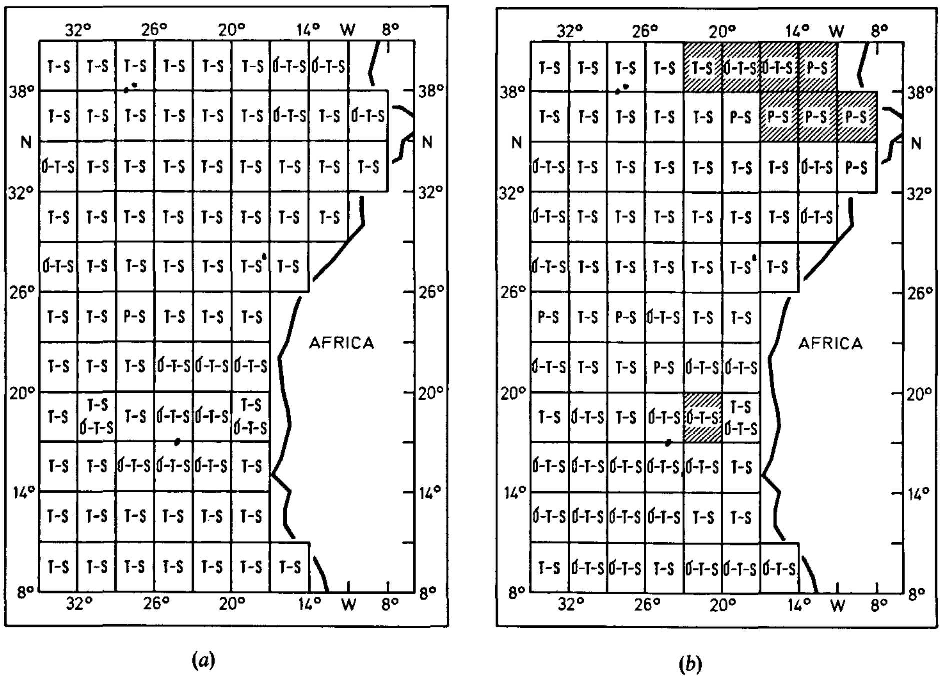

Figure 4

Best methods for computing geopotential anomalies: 0 to 500 dbar (a), 0 to 1,000 dbar (b). Squares with the geopotential anomaly exceeding 4 dyn.cm are shaded.

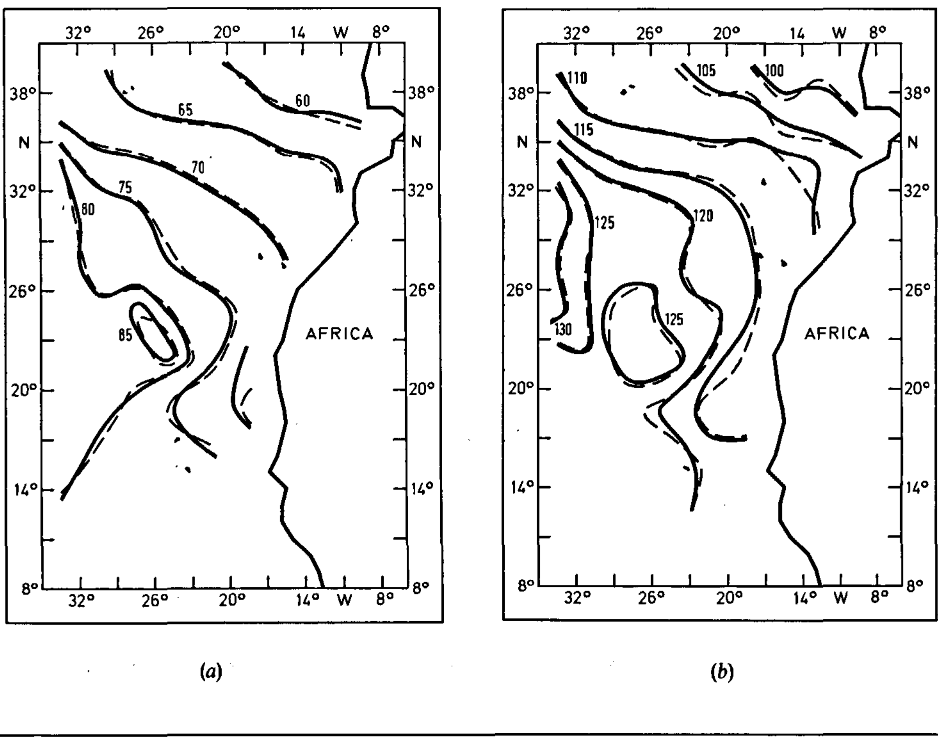

Figure 5

Topographies of geopotential anomalies in dyn.cm. Solid lines obtained from density data, dashed lines from methods given in Figure 4. 0 to 500 dbar (a), 0 to 1,000 dbar (b).

water, the variability of the Mediterranean water causes larger errors near the Portuguese coast. For 2 000 dbar levels the errors are usually high; neither method is adequate. No further improvement could be obtained when grouping the hydrographic data into 4 seasonal intervals and applying 4 mean relationships according to season. This negative result may be due to the uneven distribution of data over the year and to low station density in much of the region.

In order to check whether the methods indicated in Figure 4*a* and *b* and the normal method actually lead to similar current patterns, topographies of geopotential anomaly are presented in Figures 5*a* and *b*. The patterns compare quite well for both the 500 and 1 000 dbar reference levels.

APPLICATION TO A PARTICULAR SUBSET OF THE DATA

Saunders (1982) used several zonal hydrographic sections to compute basin-wide mean meridional transports between the mid-Atlantic ridge and the eastern coast using the Killworth (1980) model for determining a layer of no motion. The section crossing the area of this study at 32°15'N was selected for an application of the T/S method between 34°13'W and 13°24'W also with consecutive basin-wide averaging over the density slopes. The density was calculated at 27 selected depth levels between the surface and 2 500 m, using first temperature-salinity-depth data and secondly temperature data and mean T/S relationships. The

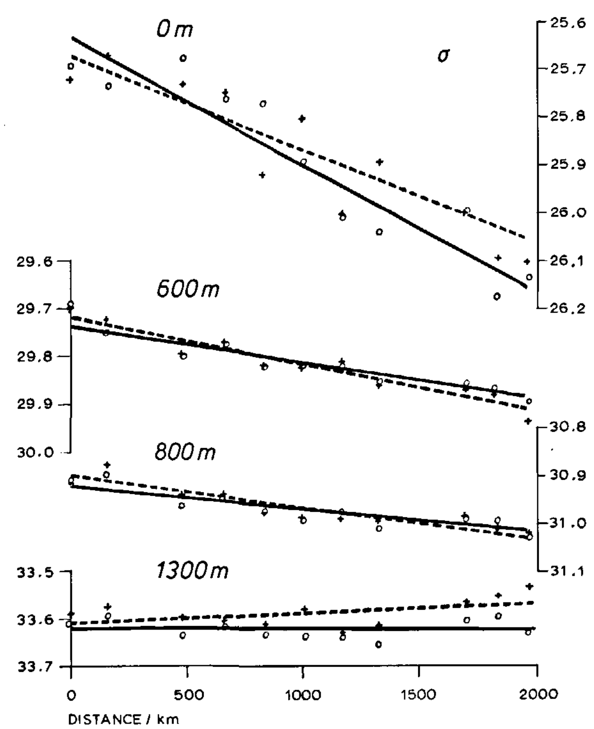

Figure 6

Density deviations $\sigma = (\rho - 10^3)$ $kg.m^{-3}$ at selected levels of zonal section and linear fit obtained from temperature-salinity-depth data (circles, solid line) and from temperature-depth data and mean T/S curve (crosses, dashed line).

Table 2

Meridional cumulative volume transports in 10^6 m^3/sec. calculated from Discovery section (December 1957) at 32°15'N between 34°13'W and 13°24'W. Reference level: 1 300 dbar.

Depth (m)	With density obtained from S-T-P	T-P and T/S	Error (%)
0- 200	−4.22	−4.37	+3.6
0- 500	−7.60	−7.85	+3.3
0- 800	−8.87	−8.94	+0.8
0-1 000	−9.10	−8.84	−2.9
0-1 500	−9.13	−8.70	−4.7
0-2 000	−8.93	−8.77	−1.8

section is close to the border of the 3 × 3° squares at 32°N. Use of the 32°-35°N T/S relationships led to unsatisfactory results. Considerable improvement was achieved when averaging the T/S curves of the squares bordering 32 to 35 and 29 to 32°N. Samples for some depth levels of the two resulting density data sets are presented in Figure 6, together with the least-squares linear fit. 1 300 dbar was selected as a level of no motion because of the small slope in the density distribution at that depth. Currents and cumulative transports are given in Table 2 and Figure 7. The errors introduced by the T/S method are below 5% of the cumulative transport values obtained from the actual density data. The method leads to satisfactory results in this case.

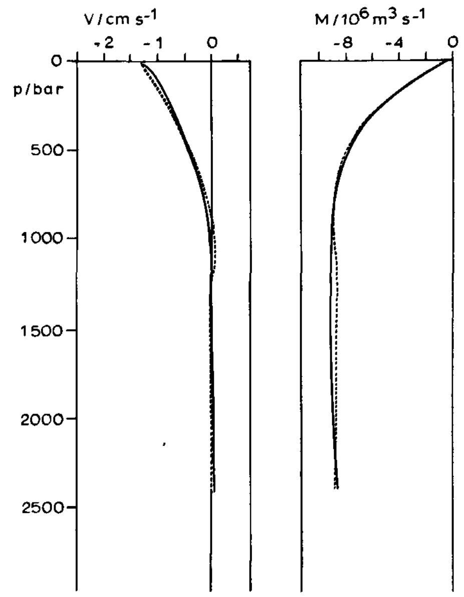

Figure 7

Geostrophic meridional velocity V *relative to 1,300 dbar (left side) and the corresponding cumulative volume transport* M *(right side). Solid line is obtained from temperature-salinity-depth data, dashed line from temperature-depth data and mean* T/S *curve.*

Acknowledgements

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