POLYMODE NEWS

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In this issue:

SPATIALLY AVERAGED LOCAL DYNAMICS EXPERIMENT CTD STATIONS

by Harry Bryden and Bob Millard

During the POLYMODE Local Dynamics Experiment (LDE), most of the CTD stations extending to the ocean bottom were made on the mooring recovery cruise aboard R/V Oceanus (Ciesluk, POLYMODE News No. 71). Since top-to-bottom profiles are useful for dynamic analyses, we have averaged the ten CTD stations made during July 1979 aboard R/V Oceanus (Figure 1) at 2 dbar intervals to obtain vertical profiles of potential temperature, salinity, and potential density (Figure 2). Standard deviations of temperature and salinity about their average values are also calculated (Figure 3). The standard deviation of temperature about this spatial average is approximately half of the r.m.s. temperature fluctuation about a time average of 15 months obtained from moored temperature measurements on mooring 1. Finally, the vertical gradient of potential temperature and Brunt-Väisälä frequency are calculated from differences over 20 dbar intervals (Figure 4).

These profiles are similar to those we calculated for the MODE region (Millard and Bryden, MODE Hot Line News No. 43). Large variability as exhibited by large standard deviation occurs in regions of strong vertical gradients such as the seasonal thermocline

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MEGAMETER ATLANTIC XBT SECTIONS, II

by Maria Henke and Walter Zenk

On the occasion of the German Antarctic Expedition 1979/80, we had the opportunity to obtain new XBT sections from both hemispheres of the Atlantic. Our activities were limited by the cruise track of R/V Polarsirkel. On the southbound leg in December 1979 we concentrated on the southern area of the North East Atlantic Dynamic Studies (NEADS) and on the tropical region where a part of the Garp Atlantic Tropical Experiment (GATE) took place (Figure 5). During the return leg in March 1980, a deep XBT section from 29°S to 41°N was obtained. This section complements the 10 megameter section measured on a similar track 21 months earlier by Henke (POLYMODE News No. 59).

Sections A and B (Figure 6a,b) result from 450 m probes; section C (Figure 7) was observed by 750 m probes, except for near-shelf regions. A nominal distance of 24 n.m. was chosen in all cases, except in the equatorial region and south of the NEADS area on the return cruise where the cast density had been increased for better resolution. A total of 325 XBT observations were made. Satellite navigation was available on board the Polarsirkel.

Section C (Figure 7), taken between 28°S and 10°N, has a main thermocline that varies

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SPATIALLY AVERAGED LDE
CTD STATIONS (continued)

and the main thermocline which are separated by the region of 18° water with small variability and small vertical gradients.

We will provide to interested researchers this spatially averaged CTD station in the form of average temperature and salinity at 2 dbar intervals and a standard-level station listing.

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about 100 m. In contrast to the north summer section from 1978, this time the equatorial regime with a sharp and shallow thermocline has moved southward. The slope of the thermocline shows a well developed equatorial current system. We recognize the South Equatorial Current (6°S-4°S), the Undercurrent in 90-110 m depth a few km south of the equator, and the North Equatorial Countercurrent (7°N-8°N).

Associated with the characteristic equatorial divergence is the spreading of the 13°-14°C isotherms together with an intermediate minimum of the vertical temperature gradient known as the thermostad (Brockmann and Meincke, POLYMODE News No. 18). Presumably again caused by the difference in the seasons, the thermostad this time was more strongly developed south of the equator, while in 1978 it was more pronounced on the northern side. The thermal equator, the area of highest surface temperature, was encountered at about 3°N with values >28.5°C.

Section B (Figure 6b) was obtained four months earlier and it indicates a more northern position of the thermal equator at about 4°N. This section was obtained in an area southwest of the Cape Verde Islands where the countercurrent contributes to the formation of an anticyclonic gyre system. The associated cold water piling up in this region, seen in the thermocline at about 11°N in section B, has been called the Guinea Dome (Voituriez and Herbland, 1980).

Returning to section C (Figure 7), we recognize the lower boundary of the Warm Water Sphere, reaching a maximal depth on both hemispheres at about 15°S and 30°N. It rises to the equator and poleward of the subtropical convergence zones. Because of the two data gaps north of 10°N and 19°N, we missed the dissolution of the sharp thermocline which had been clearly observed as the subtropical convergence zone in 1978.

Two regional effects interact with the Warm Water Sphere:

1) south of 28°S in section C the Agulhas Current extension was found at exactly the same location as in 1978;

2) on the northern end (>37°N), minimal temperature gradients could be observed below 500 m. This second thermostatic layer in section C is caused by the upper temperature maximum of the Mediterranean Undercurrent flowing parallel to the shelf of Portugal.

Special attention has been given to the southern NEADS area where long-term current meter moorings have been operated by our institute (Institut für Meereskunde) since 1977 (cf. Figure 5). The region has been covered twice -- by section A (Figure 6a) and by part of section C (Figure 7). The seasonal surface cooling amounted ≈2°C. Among the aims of NEADS is the observation of eddy size and frequency in the northeast Atlantic. Although the displayed sections seem to show a maximum of undulating features in the NEADS area, it is too difficult to interpret them as eddy expressions. Further statistical treatment of these and other data sets will be necessary to get significant statements on eddy appearances in the NEADS area.

We thank N. Slotsvik for deploying the XBT probes in sections A and B. This work was supported by Deutsche Forschungsgemeinschaft, Bonn-Bad Godesberg.

Reference

Voituriez, B. and A. Herbland (1980) Comparison of the coastal and open ocean upwelling ecosystems of the Tropical Eastern Atlantic. Rapp. Proces-Verb. (in press).

PLANETARY SOLITARY WAVES: PATHOLOGY OF QUASIGEOSTROPHIC POTENTIAL VORTICITY EQUATION IN THE LONG WAVE LIMIT

by Toshio Yamagata

Many authors dealing with planetary solitary waves adopt the quasigeostrophic potential vorticity equation (PVE) as the system equation. This approach should be checked, since the quasigeostrophic PVE is only a residual equation obtained at the first order of the Rossby-number expansion, and the PVE is obtained after eliminating pressure terms by cross-differentiation. In this note I will report briefly on a class of planetary solitary waves which is lost by the above quasi-geostrophic PVE approach. Detailed analysis including topography, mean shear, and stratification will be submitted to journals shortly.

The nonlinear shallow-water equations on the mid-latitude beta-plane are

$$u_t + uu_x + vu_y - fv = -g\zeta_x \quad ,$$
$$v_t + uv_x + vv_y + fu = -g\zeta_y \quad , \qquad (1)$$
$$\{(H+\zeta)u\}_x + \{(H+\zeta)v\}_y + \zeta_t = 0 \quad .$$

Here the Coriolis parameter f is given by

$$f = f_0 + \beta y \quad , \qquad (2)$$

and the mean fluid depth H is constant. The other notations obey the conventional usage. We adopt a channel ocean with north-south extent L. Let L, $\beta^{-1}L^{-1}$, U, and $g^{-1}fLU$ denote a length scale, a time scale, a velocity scale, and a scale of surface elevation. Then we obtain the nondimensional equations:

$$\delta u_t + R_0(uu_x+vu_y) - (1+\delta y)v = -\zeta_x \quad ,$$
$$\delta v_t + R_0(uv_x+vv_y) + (1+\delta y)u = -\zeta_y \quad , \qquad (3)$$
$$u_x + v_y + F\{\delta\zeta_t + R_0(\zeta u)_x + R_0(\zeta v)_y = 0 \quad ,$$

where δ ($= \beta L f_0^{-1}$) is the beta parameter, F ($= f_0^2 L^2 g^{-1} H^{-1}$) is the rotational Froude number, and R_0 ($= U f_0^{-1} L^{-1}$) is the Rossby number. Hereafter, we assume $\delta << O(1)$ and $R_0 \sim \delta^2$. The former assumption is valid in mid-latitude for L smaller than $O(10^3 Km)$. The latter assumption can be satisfied well in mid-oceans.

Introducing the stretched coordinates:

$$X = \delta(x - c_0 t) \quad ,$$
$$T = \delta^3 t \quad , \qquad (4)$$

where c_0 is the long-wave speed defined later, we can write (3) in the form:

$$-\delta^2 c_0 u_X + \delta^4 u_T + \delta^2(\delta u u_X + v u_y) - (1+\delta y)v = -\delta\zeta_X ,$$
$$-\delta^2 c_0 v_X + \delta^4 v_T + \delta^2(\delta u v_X + v v_y) + (1+\delta y)u = -\zeta_y , \quad (5)$$
$$u_X + v_y + F\{-\delta^2 c_0 \zeta_X + \delta^4\zeta_T + \delta^3(\zeta u)_X + \delta^2(\zeta v)_y\} = 0 .$$

We seek an asymptotic solution to (5) in the form:

$$\begin{pmatrix} u \\ v \\ \zeta \end{pmatrix} = \begin{pmatrix} u^{(0)} \\ v^{(0)} \\ \zeta^{(0)} \end{pmatrix} + \delta \begin{pmatrix} u^{(1)} \\ v^{(1)} \\ \zeta^{(1)} \end{pmatrix} + \delta^2 \begin{pmatrix} u^{(2)} \\ v^{(2)} \\ \zeta^{(2)} \end{pmatrix} + \ldots \qquad (6)$$

Substituting (6) into (5) gives, to (δ^2)

$$\zeta^{(0)}_{yyX} - (F + \frac{1}{c_0})\zeta^{(0)}_X = 0 \quad , \qquad (7)$$

with the boundary condition:

$$\zeta^{(0)}_X = 0 \text{ at } y = 0 \text{ and } 1 \; . \qquad (8)$$

The eigensolution is

$$\zeta^{(0)} = A(T,X)\sin m\pi y \;, \qquad (9)$$

with the eigenvalue (phase speed of the long wave):

$$c_0 = \frac{-1}{m^2\pi^2 + F} \quad . \qquad (10)$$

Proceeding to $O(\delta^3)$, we get

$$\zeta^{(1)}_{yyX} - (F + \frac{1}{c_0})\zeta^{(1)}_X = 2A_X(m\pi\cos m\pi y + Fy\sin m\pi y) , \quad (11)$$

with the boundary condition:

$$\zeta^{(1)}_X = -c_0 m\pi A_X \cos m\pi y \text{ at } y = 0 \text{ and } 1 \; . \quad (12)$$

The boundedness of $\zeta^{(1)}$ requires

$$\frac{1}{2}c_0 F A_X = 0 \; . \qquad (13)$$

Thus we find $A_X = 0$ (which means $\zeta^{(1)} = 0$) at this order unless $F = 0$. The case when the lateral divergence is negligible ($F = 0$) is discussed later. Proceeding further to $O(\delta^4)$, we obtain

$$\zeta^{(2)}_{yyX} - (F + \frac{1}{c_0})\zeta^{(2)}_X = \mathrm{F} \qquad (14)$$

with the boundary condition

$$\zeta^{(2)}_X = c_0 m\pi y A_X \cos m\pi y - m^2\pi^2 A A_X \cos^2 m\pi y , \quad (15)$$

where F is tedious and shielded. Applying the solvability condition and integrating by parts by

(continued page 5)

PLANETARY SOLITARY WAVES (continued)

use of (15), we find the evolution equation for A:

$$A_T + a_1 A_X + a_2 A A_X + a_3 A_{XXX} = 0 \ , \qquad (16)$$

where

$$a_1 = c_0(2c_0^2 m^2\pi^2 + \frac{1}{3}c_0 m^2\pi^2 + \frac{3}{2}c_0 - \frac{1}{2m^2\pi^2} + \frac{1}{3}) \ , \qquad (17a)$$

$$a_2 = c_0\{1-(-1)^m\}\{\frac{8}{3}c_0 m^3\pi^3 + \frac{20}{3}m\pi + \frac{2F(c_0 m^2\pi^2+2)}{3m\pi}, \qquad (17b)$$

$$a_3 = -c_0^2 \ . \qquad (17c)$$

Equation (16) can be reduced to the well known K-dV equation by using the transformation:

$$\begin{pmatrix} X \\ T \end{pmatrix} \rightarrow \begin{pmatrix} X - a_1 T \\ T \end{pmatrix} \ . \qquad (18)$$

Thus we find

$$A_T + a_2 A A_X + a_3 A_{XXX} = 0 \ . \qquad (19)$$

The above equation is further reduced to the canonical form by the transformation:

$$\begin{pmatrix} A \\ X \\ T \end{pmatrix} \rightarrow \begin{pmatrix} \frac{6a_3^{1/3}}{a_2}\eta \\ a_3^{1/3}\xi \\ T \end{pmatrix} , \qquad (20)$$

if a_2 is non zero. The result is

$$\eta_T + 6\eta\eta_\xi + \eta_{\xi\xi\xi} = 0 \ . \qquad (21)$$

The progressive wave solution which moves with the speed η in the ξ direction and satisfies the condition: $\eta(\xi\rightarrow\pm\infty) \rightarrow 0$ is obtained for positive λ by integrating (21) twice with respect to ξ. One solution is

$$\eta = \frac{\lambda}{2}\,\mathrm{sech}^2\{\frac{1}{2}\sqrt{\lambda}(\xi - \lambda T - \xi_0)\} \ , \qquad (22)$$

and the other one is

$$\eta = \frac{\lambda}{2}\,\mathrm{cosech}^2\{\frac{1}{2}\sqrt{\lambda}(\xi - \lambda T - \xi_0)\} \ . \qquad (23)$$

We should note here that a_2 vanishes when the cross channel mode m is even. Therefore the evolution is reduced to the linear equation. However, it is possible to balance the linear dispersion with the nonlinearity by adopting longer time and length scales. The remedy is given in a later paragraph.

For the nondivergent case ($\delta^2 >> F$) we must restart from (11). The solution is obtained as

$$\xi_X^{(1)} = A_X(-C_0 m\pi \cos m\pi y + y \sin m\pi y), \qquad (24)$$

where

$$C_0 = -\frac{1}{m^2\pi^2} \ . \qquad (25)$$

To $O(\delta^4)$, we find

$$\xi_{yyX}^{(2)} - \frac{1}{C_0}\xi_X^{(2)} = G \ , \qquad (26)$$

with the boundary condition

$$\xi_X^{(2)} = -m^2\pi^2 A A_X \cos^2 m\pi y \text{ at } y = 0 \text{ and } 1 \ , \qquad (27)$$

where G is shielded. The solvability condition yields

$$A_T + b_1 A_X + b_2 A A_X + b_3 A_{XXX} = 0 \ , \qquad (28)$$

where

$$b_1 = \frac{m^2\pi^2}{3} + \frac{1}{2} \ , \qquad (29a)$$

$$b_2 = \frac{2}{m\pi}\{1 - (-1)^m\} \ , \qquad (29b)$$

$$b_3 = -C_0^2 \ . \qquad (29c)$$

Equation (28) is reduced to

$$A_T + b_2 A A_X + b_3 A_{XXX} = 0 \ , \qquad (30)$$

by use of the transformation:

$$\begin{pmatrix} X \\ T \end{pmatrix} \rightarrow \begin{pmatrix} X - b_1 T \\ T \end{pmatrix} . \qquad (31)$$

Here, as $F \rightarrow 0$, it is apparent from (17) and (29) that (a_1, a_2, a_3) do not tend to (b_1, b_2, b_3) and so a limit: $F \rightarrow 0$ is singular or, in other words, the order-by-order perturbation scheme and the limit $F \rightarrow 0$ are not interchangeable in the present problem.

So far we cannot obtain the balance between the linear dispersion and the nonlinearity for even m modes. Here we discuss the case by adopting the different length and time scales as

$$X = \delta^2(x - C_0 t) \ , \qquad T = \delta^6 t \ . \qquad (32)$$

Then (3) is rewritten as

$$-\delta^3 C_0 u_X + \delta^7 u_T + \delta^2(\delta^2 u u_x + v u_y) - (1+\delta y) = -\delta^2\xi_X,$$

$$-\delta^3 C_0 v_X + \delta^7 v_T + \delta^2(\delta^2 u v_x + v v_y) + (1+\delta_y) = -\delta_y,$$

$$\delta^2 u_X + v_y + F\{-\delta^3 C_0\xi_X + \delta^7\xi_T + \delta^4(\delta u)_x + \delta^2(\delta v)_y\} = 0. \qquad (33)$$

Expanding variables by powers of δ as before and substituting them into (33), we find, to $O(\delta^3)$, (7) and (8) again. Thus the solution is

$$\xi^{(0)} = A(T,X) \sin m\pi y \ , \qquad (34)$$

(continued page 6)

with

$$c_0 = \frac{-1}{m^2\pi^2+F} \quad . \tag{35}$$

Proceeding to $O(\delta^4)$, we find (11) and (12) again. Therefore $A_X = 0$ ($\zeta^{(1)} = 0$) at the order if $F = 0$. Similar discussions at $O(\delta^5)$ and at $O(\delta^6)$ give $\zeta^{(2)} = 0$ and $\zeta^{(3)} = 0$ respectively. Finally, to $O(\delta^7)$, we obtain

$$\xi^{(4)}_{yyX} - (F+\frac{1}{c_0})\ \xi^{(4)}_X = H, \tag{36}$$

with the boundary condition

$$\xi^{(4)}_X = m\pi y^2 \cos m\pi y\ (c_0 y A_X - 3m\pi A A_X \cos m\pi y). \tag{37}$$

Applying the solvability condition, we find

$$A_T + c_1 A_X + c_2 AA_X + c_3 A^2 A_X + c_4 A_{XXX} = 0, \tag{38}$$

where

$$c_1 = 2c_0\{c_0(c_0 m^2\pi^2+1) - \frac{3c_0}{2m^2\pi^2} + (c_0 m^2\pi^2+1)(\frac{1}{10} - \frac{1}{2m^2\pi^2} + \frac{3}{4m^4\pi^4}) - \frac{1}{m^2\pi^2} + \frac{3}{2m^4\pi^4} + Fc_0 m\pi(\frac{m\pi}{6} + \frac{1}{4m\pi} + \frac{2}{3})\}, \tag{39a}$$

$$c_2 = -2c_0\{3c_0 m^3\pi^3+12m\pi+2m\pi(c_0 m^2\pi^2+1) + 6m\pi + F(-\frac{m^2\pi^2}{4} - \frac{4}{3m\pi} + \frac{3}{4m^2\pi^2} - \frac{1}{2})\}, \tag{39b}$$

$$c_3 = c_0 m^4\pi^4\{4 + \frac{1}{4}(c_0 m^2\pi^2+1)\}, \tag{39c}$$

$$c_4 = -c_0^2 \ . \tag{39d}$$

By use of the transformation:

$$\begin{pmatrix} X \\ T \end{pmatrix} \rightarrow \begin{pmatrix} X - c_1 T \\ T \end{pmatrix}, \tag{40}$$

equation (38) is reduced to

$$A_T + c_2 AA_X + c_3 A^2 A_X + c_4 A_{XXX} = 0 \ . \tag{41}$$

It should be noted that the quadratic nonlinearity is comparable to cubic nonlinearity in (41). The above mixed *K-dV* and *MK-dV* equation is further reduced to the canonical form:

$$\eta_T + 6\eta\eta_\xi + 6\delta_* \eta^2 \eta_\xi + \eta_{\xi\xi\xi} = 0 \ , \tag{42}$$

by use of the transformation

$$\begin{pmatrix} A \\ X \\ T \end{pmatrix} \begin{pmatrix} \frac{6c_4^{Y_3}}{c_2}\eta \\ c_4^{Y_3}\xi \\ T \end{pmatrix}, \tag{43}$$

where

$$\delta_* = \frac{6c_3 c_4^{Y_3}}{c_2^2} \quad . \tag{44}$$

The progressive wave solution which moves with the speed λ in the ξ direction and satisfies the condition: $\eta(\xi\to\pm\infty) \to 0$ is obtained for positive λ by integrating (42) twice with respect to ξ. Thus, for $0 < \eta < -(1-\sqrt{1+\delta_*\lambda})/\delta_*$ we have the solution:

$$\eta = \frac{\lambda}{C\cosh^2\{\frac{1}{2}\sqrt{\lambda}(\xi-\lambda T-\xi_0)\}+D\sinh^2\{\frac{1}{2}\sqrt{\lambda}(\xi-\lambda T-\xi_0)\}}, \tag{45}$$

here

$$C = \sqrt{1+\delta_*\lambda} + 1 \ , \tag{46a}$$

$$D = \sqrt{1+\delta_*\lambda} - 1 \ . \tag{46b}$$

For $-(1+\sqrt{1+\delta_*\lambda})/\delta_* < \eta < 0$, we have the solution:

$$\eta = -\frac{\lambda}{C\sinh^2\{\frac{1}{2}\sqrt{\lambda}(\xi-\lambda T-\xi_0)\}+D\cosh^2\{\frac{1}{2}\sqrt{\lambda}(\xi-\lambda T-\xi_0)\}}. \tag{47}$$

The non-divergent limit ($F \to 0$) is not uniform as in the preceding paragraph and it is necessary to obtain $\zeta^{(1)}$, $\zeta^{(2)}$, and $\zeta^{(3)}$. However the result is similar to (41), and so we do not show the explicit form.

Now let us start tentatively from the barotropic potential vorticity equation, which is written as

$$\delta\ \partial_t\ (\nabla^2-F)\ \xi + R_0 J\ (\xi,\nabla^2\xi) + \delta\xi_x = 0. \tag{48}$$

Here J denotes the Jacobian operator and ∇^2 denotes the two dimensional Laplacian operator. Focusing our interests on the case:

$$R_0 = \delta^2 \ , \tag{49}$$

as before and introducing the same time and length scales as (4), we obtain

$$(-c_0\ \partial_X + \delta^2\partial_T)(\delta^2\partial^2_{XX} + \partial^2_{yy} - F)\ \xi + \delta(\xi_y\partial_x - \xi_x\partial_y)(\delta^2\partial^2_{xx} + \partial^2_{yy})\xi + \xi_x = 0. \tag{50}$$

(continued page 7)

PLANETARY SOLITARY WAVES (continued)

Expanding ζ in an asymptotic series in δ, we have, to $O(\delta^0)$, (7) and (8) again. Therefore the eigensolution and the eigenvalue are identical with (9) and (10) respectively. However, to $O(\delta)$, we have the equation:

$$\xi^{(1)}_{yyx} - (F + \frac{1}{C_0})\ \xi^{(1)}_x = 0, \qquad (51)$$

with the boundary condition:

$$\xi^{(1)} = 0 \quad \text{at } y=0 \text{ and } 1. \qquad (52)$$

We should choose here the solution forced by the lower order one. Thus, we decide

$$\xi^{(1)} = 0. \qquad (53)$$

To $O(\delta^2)$ we obtain the equation:

$$\xi^{(2)}_{yyx} - (F + \frac{1}{C_0})\ \xi^{(2)}_x = -(\frac{-1}{C_0^2}A_T + A_{xxx})\ \sin m\pi y, \qquad (54)$$

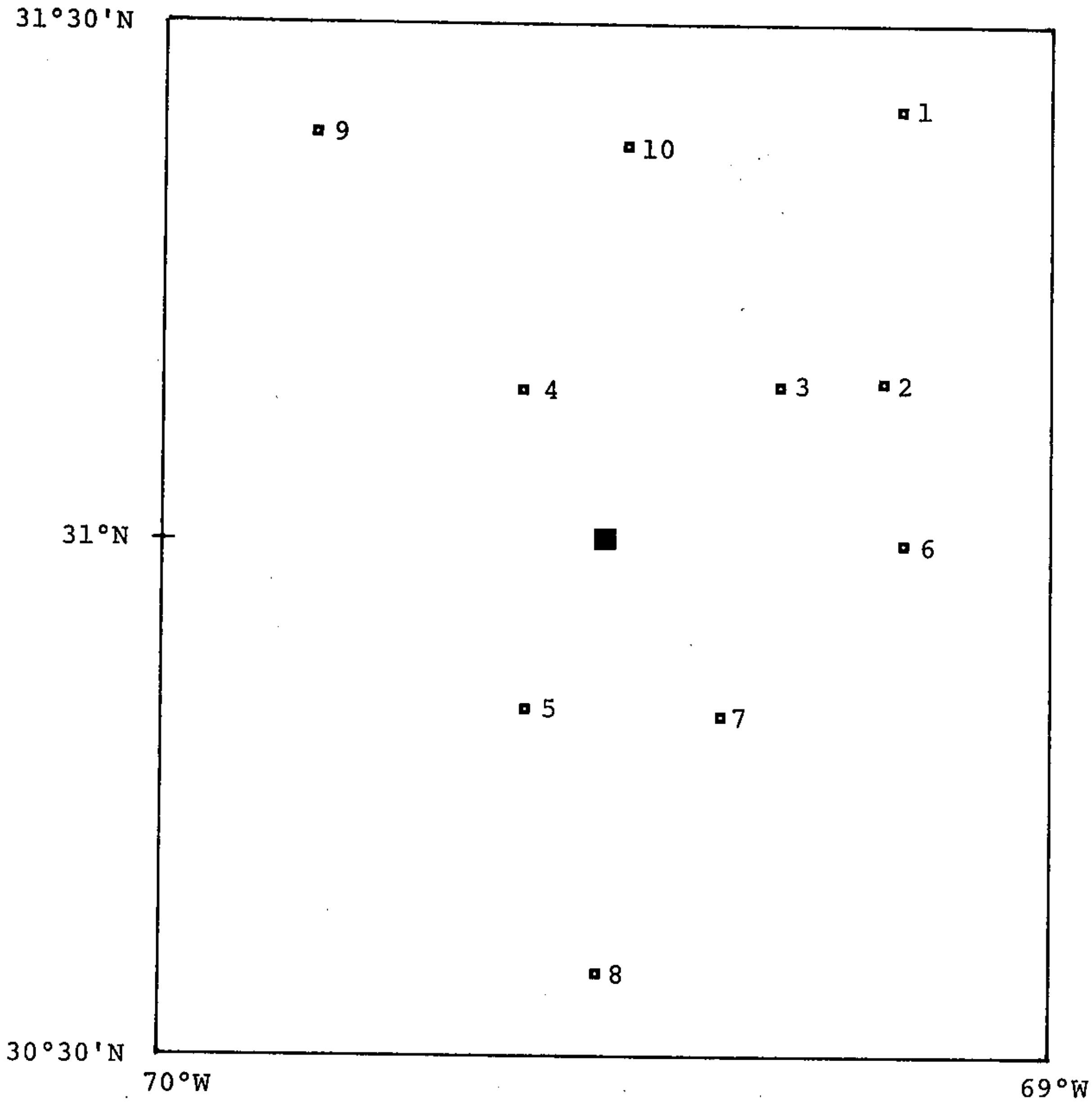
with the condition:

$$\xi^{(2)} = 0 \quad \text{at } y=0 \text{ and } 1. \qquad (55)$$

The solvability condition yields the evolution equation for A:

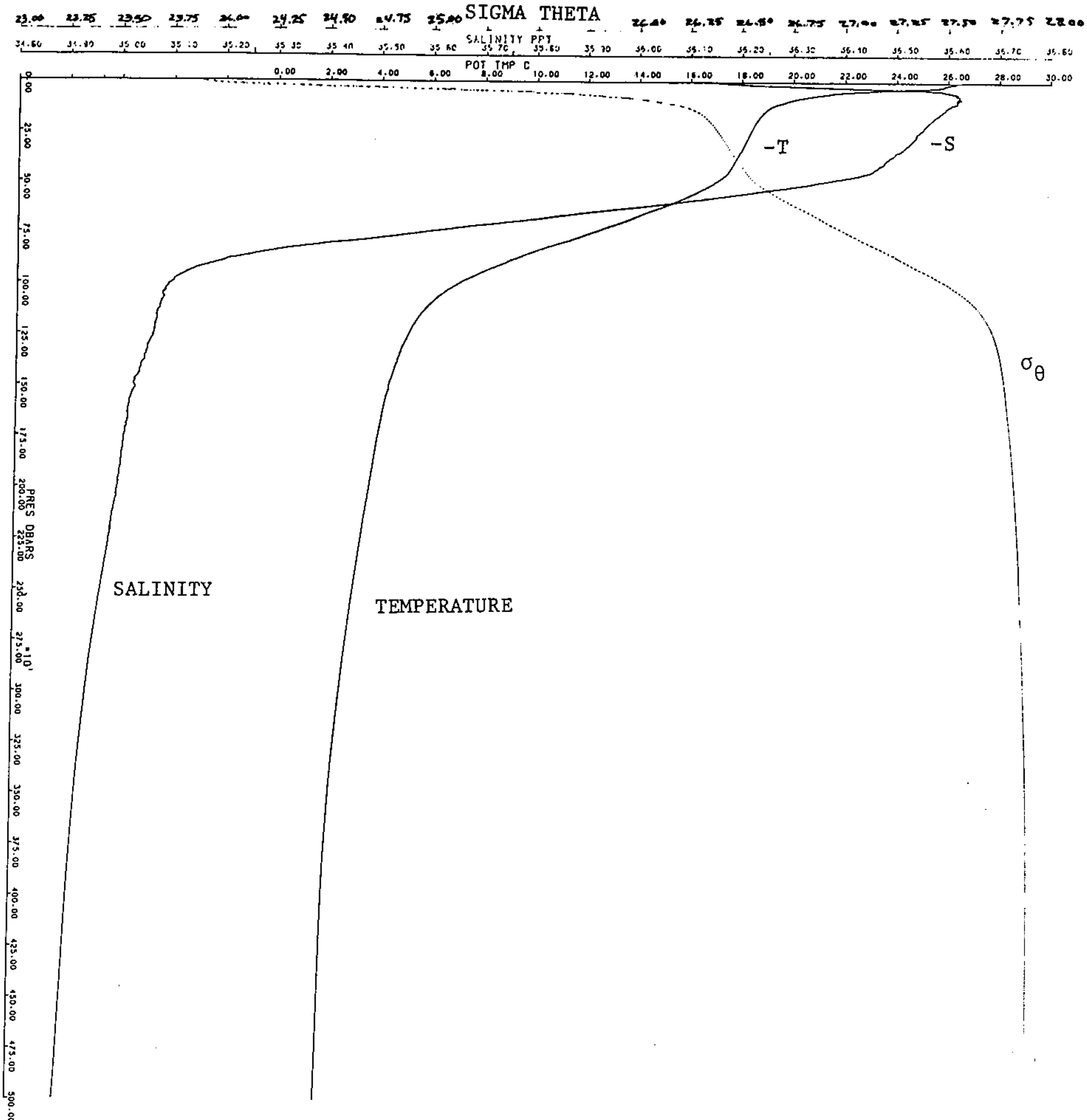
$$A_T - C_0^2\ A_{xxx} = 0. \qquad (56)$$

Equation (56) is always the linear dispersive equation independent of the mode number m. This consequence is quite misleading. The issue exists on the fact that the ageostrophic and the resulting nonlinear effect are underestimated in the long wave limit if we use the quasigeostrophic potential vorticity equation. We should remember the procedure by which quasigeostrophic PVE is derived. In order to obtain quasigeostrophic PVE, we have already performed an R_0-expansion assuming $\delta \lesssim R$. Assuming that the flow is two-dimensional and nondivergent, we can derive the barotropic potential vorticity equation similar to (48) with $F = 0$ without the assumption about the order of δ and R_0. In the latter case, however, ζ denotes the streamfunction, not the surface elevation. If we adopt $\zeta = 0$ as the condition at the channel boundaries, we again reach the linear evolution equation (56); a misleading result.



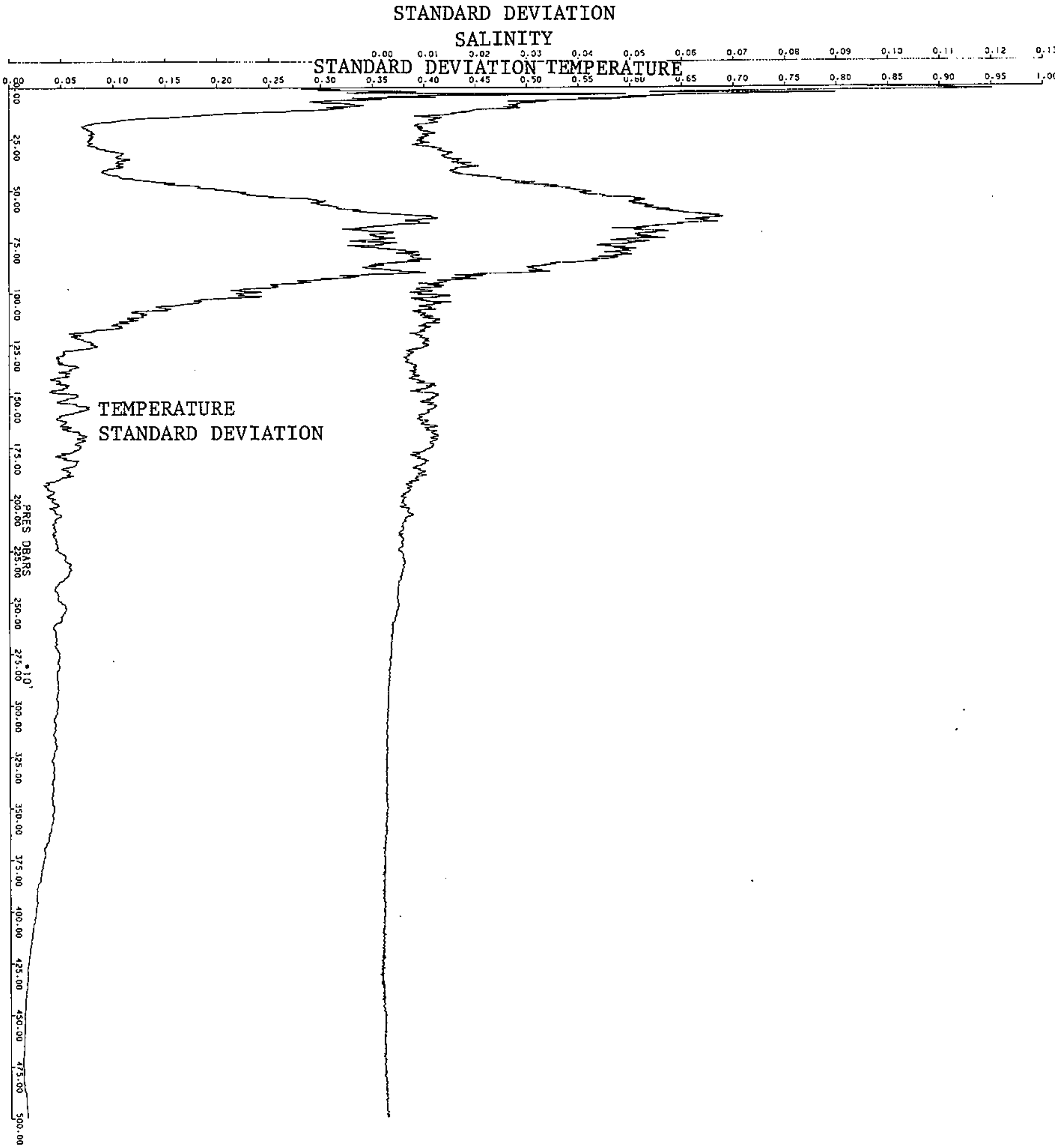
Positions of CTD stations taken on R/V Oceanus, July-August 1979. ■ denotes center of LDE.

Figure 1 (Bryden and Millard)



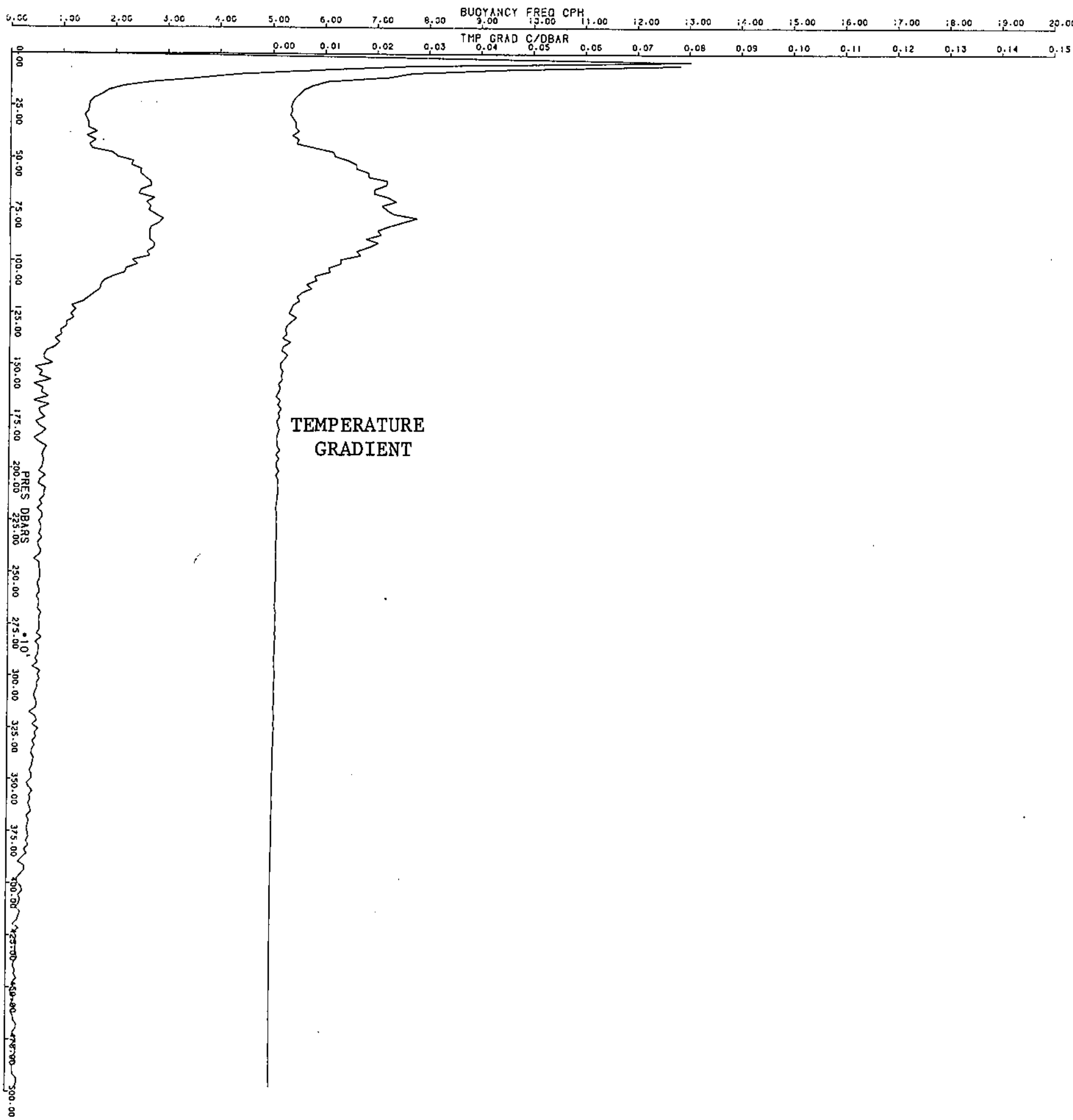
Vertical profiles of potential temperature, salinity, and potential density.

Figure 2 (Bryden and Millard)



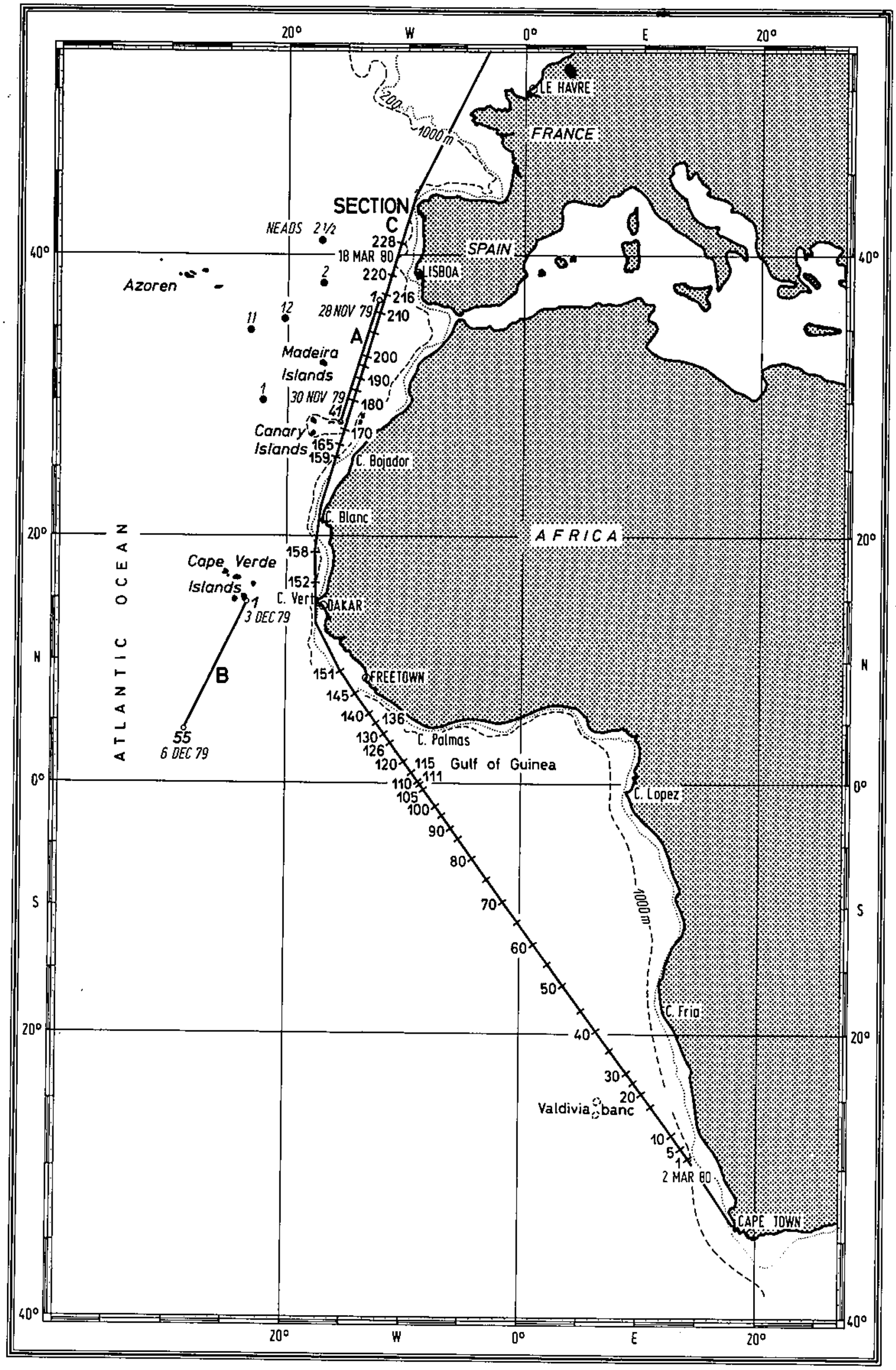
Vertical profiles of the standard deviation of temperature and salinity about their spatially averaged values.

Figure 3 (Bryden and Millard)



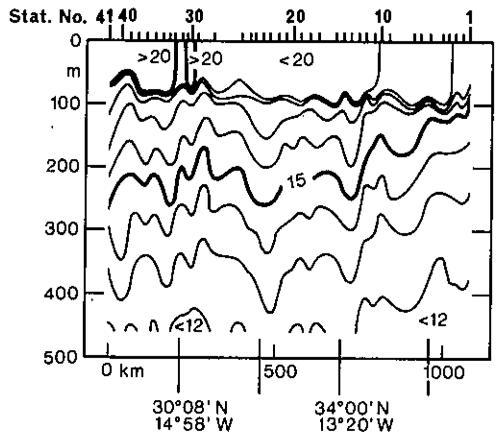
Vertical profiles of Brunt-Väisälä frequency and vertical gradient of potential temperature.

Figure 4 (Bryden and Millard)

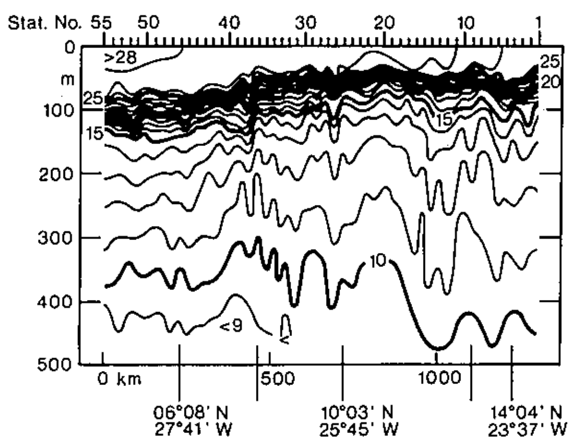


R/V Polarsirkel track towards Antarctica in December 1979 (XBT sections A, B) and returning from South Africa in March 1980 (XBT section C). Numbers indicate XBT stations. Some NEADS current meter positions are included.

Figure 5 (Henke and Zenk)



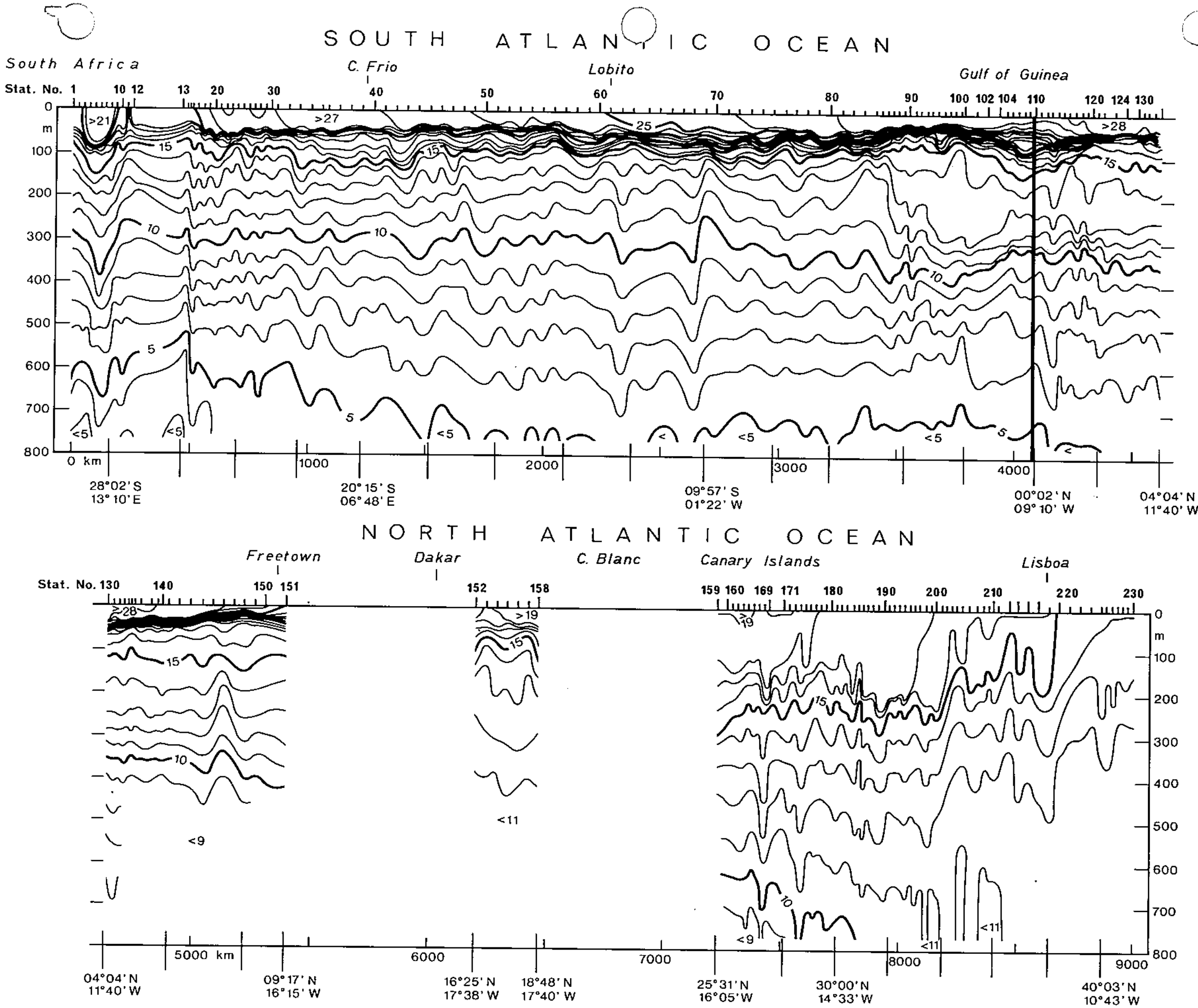
a



b

a) XBT section A (Cape Vincent to Canary Islands) in December 1979; b) XBT section B (southwest of Cape Verde Islands) in December 1979.

Figure 6 (Henke and Zenk)



XBT section C (28°S to 40°N) taken in March 1980.

Figure 7 (Henke and Zenk)