POLYMODE NEWS

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In this issue:

NORTH ATLANTIC GENERAL CIRCULATION

by Carl Wunsch

Much of the discussion of the location of the various POLYMODE arrays has taken place in the context of the structure of the general circulation of the North Atlantic. Considerable attention has been paid, for example, to the hypothesis (most prominently due to Worthington, 1976) that there is an intense recirculation associated with the Gulf Stream system. Some of the Array II current meter measurements (Schmitz, 1976, 1977) suggest that such a region may exist, and may in fact be an area of particularly active eddy motion. The reality of the recirculation is difficult to doubt (e.g. Niiler et al., 1978), but Worthington's overall scheme is suspect because it violates the geostrophic assumption.

In a recent note (Wunsch, 1977), I showed that the general circulation problem could be dealt with in the context of "inverse theory." This provides a powerful technique for using real data in the context of a supposed model of the circulation. In that publication, the method was applied to a small demonstration region. I have applied the method more recently to the entire North Atlantic region west of 50°W. The solutions obtained are described in a lengthy manuscript to be published (Wunsch, 1978). Because the results might be of use in the discussions of POLYMODE planning and observations, I will briefly summarize some of the results.

Inverse theory is usually model-dependent. Given an initial model, and a data set that is possibly "inaccurate, insufficient, and inconsistent," one seeks to change the model in the simplest way to accommodate the actual observations. The models I will deal with here are purely geostrophic, and purely mass conserving. They differ only in the initially assumed reference level for the observed hydrographic data. More complex hypotheses are obviously possible. Given a set of hydrographic station pairs (see Figures 5 and 6), let v_j (p) be the velocity in station pair j as a function of pressure relative to an initial reference level, and let b_j be the true velocity at the reference level. Breaking up the western North Atlantic into a series of boxes (as shown in Figures 5 and 6), one requires (following Worthington, 1976) conservation of the following water types: total mass; and the mass contained in the layers 4°C, 4-7°C, 7-12°C, and 12-17°C.

Writing down the conservation of each of these five water types for each of the seven boxes in Figures 5 and 6 leads to 35 equations in the 235 unknowns b_j, which we write as:

$A\,\underline{b} = -\underline{\Gamma}$, $\underline{b} = \{b_i\}$, and $\underline{\Gamma} = \{\Gamma_i\}$

where Γ_i is the imbalance of a particular water type in a particular box in the initial model (taken positive if more flows into a box than out). In general, the initial model

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NOTES from the Editor

This issue begins the third year of POLYMODE News. Thus, it is timely for our annual restatement of the editorial policy of POLYMODE News.

POLYMODE News is intended to be a newsletter in support of the POLYMODE experiment. It emphasizes scientific results and the exchange of ideas and techniques. In addition, it includes information relating to non-scientific POLYMODE activities, such as meetings, committees, and experimental plans.

This newsletter is intended for quick distribution of recent measurements, results, their analysis, and ideas for scientific work. It is not intended to be a substitute for scientific journals. It is intended simply to start the communication process among those interested in POLYMODE activities. Articles in POLYMODE News may be theoretical or experimental; their inclusion depends on their appropriateness to the scientific activities and the objectives of POLYMODE. Subjects need not be confined to the North Atlantic Ocean.

Anything included in POLYMODE News is not to be quoted or published without the permission of the contributing scientist. Data is often included in preliminary form, occasionally in spite of the need for further editing or calibration corrections. In many cases, the author will have further information, will be critically evaluating the results for scientific publication, and may later find errors in preliminary material. Thus, any interested reader should communicate directly with the contributing scientist for more detail or clarification. The only exception to the above policy is that articles from the Soviet Union may be cited without the permission of the contributing scientists. This procedure has been established in response to several requests because of the difficulty of communication over long distances.

Any reference or citation to an article in POLYMODE News should be followed by the phrase "UNPUBLISHED MANUSCRIPT."

Support for POLYMODE News is provided by the Office of the International Decade of Ocean Exploration of the National Science Foundation under Grant No. OCE 76-82843. My time as editor is supported by the Office of Naval Research under Contract No. CO-197. The past and present support of these agencies is gratefully acknowledged.

--F. W.

POLYMODE OFFICE NOTES

The US POLYMODE Executive Office now has its own registered cable address. It is, appropriately enough, POLYMODE. Cables addressed to POLYMODE CAMBRIDGEMASSACHUSETTS will be delivered directly to the telex machine in the Executive Office, eliminating delays in going through the Boston Western Union Office or MIT Telecommunications Office.

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Table 1 is an updated US/USSR POLYMODE ship schedule.

ACKNOWLEDGEMENT

The POLYMODE News is produced with support from the International Decade of Ocean Exploration (IDOE) of the National Science Foundation, and the Office of Naval Research.

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Material of interest for this newsletter may be sent to either of the above at the Woods Hole Oceanographic Institution, Woods Hole, MA, 02543, Telephone (617) 548-1400, TWX 710-346-6601.

*POLYMODE is derived from the names of the USSR POLYGON experiments and the Mid-Ocean Dynamics Experiment (MODE).

WAVE TURBULENCE ON A BETA PLANE

by E. N. Pelinovsky

The main difficulty in using dimensional estimates for finding Kolmogorov-type turbulence spectra for synoptic motions of the ocean and atmosphere is the presence of dimensional constants, in addition to the spectral energy flux (the Rossby radius of deformation, the β parameter, etc.), that prevent us from obtaining an unambiguous expression for the energy spectrum. However, this difficulty can be overcome for small nonlinearity when the turbulence comes mainly from weak-interaction waves with time- and space-random amplitude and phases, so that the turbulence may be considered wave-like.

We begin with the kinetic equations in the random-phase approximation (Zaharov and L'vov, 1975; Kadomtsev and Kontorovich, 1974; Hasselman, 1968; Holloway, 1976). Exact solutions were found by the factorization method for the dispersion relation

$$\omega = qk^{a} \qquad (1)$$

where ω is the frequency, and k is the wavenumber [in particular, for gravitational (Zaharov and Filonenko, 1966), capillary (Zaharov and Filonenko, 1967), and internal (Kontorovich, 1976; Pelinovsky and Rayevsky, 1977) waves]. In these cases the spectrum can also be found from dimensional estimates (Kadomtsev and Kontorovich, 1974). There are two dimensional constants: the energy flux, P, and the constant q in the dispersion relation.

The energy-flux dependence of the spectrum is constrained by the kinetic equation, or, to be more precise, by the form of the nonlinear interaction. In the case of quadratic interaction, the spectrum of energy, E, is proportional to $\sqrt{P}$. Thus, the power with which q and k enter E may be found unambiguously. Here we investigate weak turbulence spectra of nondivergent barotropic Rossby waves and their relation to strong turbulence spectra.

Barotropic Rossby waves are described by the following equation:

$$\frac{\partial}{\partial t}\,\Delta\psi + \beta\frac{\partial\psi}{\partial_L} + J(\psi,\Delta\psi) = 0, \qquad (2)$$

where ψ is the current function, $\beta = d\Omega/dy =$ const, Ω is the coriolis parameter, and J is the jacobian. The solution of these equations is determined, in addition to the initial conditions, by the single dimensional parameter β. This parameter appears in the linear dispersion relation of Rossby waves:

$$\omega = -\beta k_L k^{-2}, \quad k = \sqrt{k_L^2 + k_y^2}\ . \qquad (3)$$

The conditions of nonlinear interaction of Rossby waves were analyzed in detail by Longuet-Higgins and Gill (1967). They noted that the interaction appears in the first order of the perturbation theory, so that $E \sim \sqrt{P}$. We then make use of dimensionality estimation. The determination of the velocity fluctuation spectrum

$$\overline{u^2} = \int E(k)\,dk = \int E(\omega)\,d\omega \qquad (4)$$

is followed by the dimensions of the spectral densities:

$$[E(k)] \sim L^3t^{-2},\ [E(\omega)] \sim L^2t^{-1}\ . \qquad (5)$$

The basic assumptions of turbulence theory are of an energy-transfer cascade and of local interaction. That is, we assume that energy is transferred as a result of interaction of spectral components with close wavenumbers. These assumptions correspond to the Kolmogorov hypothesis in the theory of turbulent incompressible liquids and make possible writing the kinetic equations in the form of a conservation equation:

$$\frac{\partial E(k)}{\partial t} + \frac{\partial P}{\partial k} = 0. \qquad (6)$$

Here the stationary spectrum corresponds to constant spectral energy flux. It follows from (6) that the energy flux dimension is:

$$[P] \sim L^2t^{-3}. \qquad (7)$$

Using (3), (5), and (7) we find unambiguous expressions for the energy spectra:

$$E(k) \sim \sqrt{P\beta}\,k^{-5/2},\ E(\omega) \sim \beta^{-1}\sqrt{P\omega}. \qquad (8)$$

These spectra are found in the weak nonlinearity approximation. The condition of weak nonlinearity is fulfilled if the characteristic time of wave interaction exceeds their period. This time can be estimated from (6): $\tau \sim kE(k)P^{-1}$, or, with allowance for (8):

$$\omega\tau \sim \omega^{5/2}/\beta\sqrt{P} \sim \beta\sqrt{\frac{\beta}{P}}\,k^{-5/2}\ . \qquad (9)$$

This value is large within the range of high frequencies and small wavenumbers, in which

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WAVE TURBULENCE ON A BETA PLANE (continued)

the conditions of weak turbulence of Rossby waves are fulfilled:

$$\omega \gg \omega_o \sim \beta^{2/5} P^{1/5} , \qquad k \ll k_o \sim \beta^{3/5} P^{-1/5} . \tag{10}$$

Within the range $\omega \sim \omega_o$ and $k \sim k_o$, the nonlinear effects become essential. Within the present range the spectrum may be expected (by analogy with surface gravity waves) to attain "saturation", resulting from the possible loss of wave stability, for particle velocities compared with the wave speed. This point requires a special examination because Larichev and Reznik (1976a,b) obtained strongly nonlinear Rossby waves, whose particle velocities exceeded the wave speed. The stability of the resultant solution is not yet clear.

As with surface waves, the spectrum does not depend on the energy flux, but is determined by β. Hence, the dimensionality estimation results in an expression for the saturated spectrum:

$$E(k) \sim \beta^2 k^{-5}, \; E(\omega) \sim \beta^{-2}\omega^3. \tag{11}$$

The formula for $E(k)$ was obtained by Rhines (1975) from virtually the same estimation. It is easy to make sure that with $\omega \sim \omega_o$ and $k \sim k_o$, equations (8) and (11) offer the same value, i.e., the spectra of weak turbulence coincide with the saturated spectra.

It is noteworthy that these spectra depend only on the modulus of the wave vector. To determine angular dependence, it is necessary to solve a kinetic equation. Insofar as Rossby waves are anisotropic, there should be spectra corresponding to the flux not only of the scalar quantity (of energy), but also of the vector quantity; for instance, of wave impulse in the zonal direction:

$$J_k \sim \frac{E_k}{\omega_k} \frac{k_L}{k} . \tag{12}$$

Using the continuity equation

$$\frac{\partial J_k}{\partial t} + \frac{\partial P_J}{\partial k} = 0 \tag{13}$$

and dimensionality estimation, by analogy with (8) we find spectra of weak turbulence that are in conformity with impulse constancy.

$$E(k) \sim \beta\sqrt{P_J}\, k^{-3}, \; E(\omega) \sim \beta^{-1}\sqrt{P_J}\, \omega . \tag{14}$$

Finally, using the hypothesis of enstrophy flux constancy generally applied in the theory of two-dimensional turbulence, we obtain appropriate spectra for weak turbulence:

$$E(k) \sim \sqrt{\beta P_e}\, k^{-7/2}, \; E(\omega) \sim \beta^{-2}\sqrt{P_e}\, \omega^{3/2} . \tag{15}$$

The spectra obtained are shown in Figure 4. The problem of spectral realization in (8), (14), and (15) remains to be seen and requires detailed analysis of turbulence sources and sinks, and checks on local interactions. Because of this, the interpretation of the numerical calculations of Kenyon (1967) and Rhines (1975) is difficult and the Rhines spectrum $E \sim k^{-5 \cdot 4}$ cannot be directly related to these solutions.

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DIAGNOSTIC CALCULATIONS OF CURRENT VELOCITY: A COMPARISON WITH NORTH PACIFIC DATA

by Artem Sarkisyan, Victor Drozdov and Greg Zhikharev

When planning field experiments in an ocean region, it is necessary to have a preliminary climatic current pattern. Oceanographers usually use the dynamic method for this purpose, but the results depend on the choice of reference level. We believe the diagnostic model is more accurate (Sarkisyan, 1966, 1977). The physical statement of the problem of the diagnostic calculations is as follows.

We consider a region of arbitrary form with arbitrary bottom relief in an extra-equatorial part of the ocean. The density field and wind stress in this region are specified. We determine sea-surface topography, ζ, at the side boundaries by the dynamic method, accepting the bottom relief as a reference level (ζ_1). The vertical velocity equals zero at the sea surface. At the sea bottom the non-slip conditions are taken. We wish to calculate the ζ-field and the current velocities in this region. Mathematically, the problem reduces to the solution of the Dirichlet problem for ζ.

We have:

$$\zeta = \zeta_1 + \zeta_2, \qquad (1)$$

$$\zeta_1 = -\frac{1}{\rho_o} \int_0^H \rho, \qquad (2)$$

$$\begin{cases} L(\zeta_2) = F, & (3) \\ \zeta_2|_\Gamma = 0, & (4) \end{cases}$$

$$\zeta|_\Gamma = \zeta_1, \text{ and} \qquad (5)$$

$$L = \Delta + a(x,y)\frac{\partial}{\partial x} + b(x,y)\frac{\partial}{\partial y} \qquad (6)$$

ρ_o is a const; ρ is the density anomaly; F is the given function of ρ and wind stress; Δ is the Laplacian; the coefficients $a(x,y)$ and $b(x,y)$ are the given functions of the bottom relief and of "β".

The equation for ζ_2 is solved numerically. After defining ζ, the flow velocity components are calculated by a simple formula.

The density field for the calculations was taken from the data of cruise 12 of R/V Dmitry Mendeleev and cruise 56 of R/V Vityaz. These ships carried out a semi-synoptic hydrographic survey from 22 June to 3 July. The polygon was in a subarctic oceanic front (North Pacific current). The distance between hydro stations equalled 20' of longitude and 1° of latitude; the measurements were made down to 2000 m. Five buoy stations with BPV current meters were deployed in the region at the same time.

The time series for velocity components were smoothed with a Tukey cosine filter. Mean velocities from the hydro survey were calculated after filtering. A more accurate analysis of the data can be found in a paper by Byshev *et al*. (1976). Interpolated values of multi-year averaged summer wind data were used. Ocean depths were taken from navigational charts. Density data were interpolated linearly along latitude to obtain data in grid points with $\delta x = \delta y = 20'$. The data were also extrapolated between 2000 m and 5000 m by the formula of Monin *et al*. (1970). The density field was smoothed in the horizontal plane to exclude the effect of inertial and tidal motions.

The resulting density field was specified at 26 levels: 0, 10, 20, 30, 50, 75, 100, 125, 150, 200, 250, 300, 400, 500, 600, 800, 1000, 1200, 1500, 2000, 2500, 3000, 3500, 4000, 4500, and 5000 m. The velocity vectors obtained by three methods are intercompared in Figures 1-3. We see that at three points the diagnostic speeds are closer to the observed ones; at one point the dynamic method is closer; and there remains one point where neither method has the advantage. There is a rapid decay with depth of the speed obtained with the dynamic method due to the reference-level hypothesis.

There are also essential discrepancies between the measured and the computed velocities caused by nonstationarity and nonlinearity. Nevertheless, it seems to us that the diagnostic calculations could serve as a useful tool for analysis of future POLYMODE data.

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NORTH ATLANTIC GENERAL CIRCULATION (continued)

is incorrect ($\underline{\Gamma} \neq 0$), and some water masses fail to be conserved. Elements of the matrix A are the areal cross-section of a particular water mass between a particular station pair. Because the system is under-determined, one chooses to modify the initial model according to a principle of maximum simplicity: choose that b_j which is as small as possible relative to the initial model (i.e., $\underline{b}^T\underline{b}$ to be as small as possible). An alternative but equivalent statement is: $\underline{b}$ should contain no components lying in the null-space of A because we have no information about that part of $\underline{b}$. The two statements are equivalent.

In practice, we take into account the presence of noise in the system; we do not want perfect conservation because the observations are not perfect. We solve the problem using the singular value decomposition (e.g., Wiggins, 1972) which allows control of the statistics, the spatial resolution of $\underline{b}$, the degree of confidence in individual observations, etc. In a future publication (Wunsch, 1978), I discuss these things in detail in the general circulation context.

Shown in Figures 5 through 8 are the circulation schemes that result for the total transport and the cold water layer for two different models. Model I (the "big Gulf Stream") has an initial reference level at the bottom; model II is more conventional and has an initial reference level at 2000 dbars (or the bottom, whichever is shallower). Both circulation schemes are purely geostrophic and conserve mass up to the estimated errors (but one needs to discuss the possibility of systematic errors as well as random errors). A recirculation region results from both models and both differ from Worthington (1976) in requiring a large circulation in a gyre to the south as well. There, the resulting eastward flow is much like that described by Price and Riser (POLYMODE News No. 36) from the SOFAR floats and should be testable from POLYMODE Array III, Cluster C. Model II yields a 4000 m velocity along 57°W that is closer than model I to current meter results of Schmitz (1977) and gives a southward flow along the western flank of the Bermuda Rise much like Schmitz (1976) found there. Model II gives a strong recirculation on both sides of the Gulf Stream, which is consistent with known westward flow along the slope/rise (Luyten, 1977), and divides the southern recirculation into smaller scales. It does, however, result in a Gulf Stream flow smaller than has been recently hypothesized, and about which one can argue.

One of the beauties of the inverse schemes is that one can usually force the solution to be consistent with any constraint one wants to impose. My own preference is for model II because it is more interesting and seems to most resemble features of the circulation (e.g., the current meter measurements) that were not separately imposed. However, on the basis of the simple model applied -- mass conservation of thermal layers, and geostrophy -- one cannot choose between them; they are both wholly consistent with the stated hypotheses.

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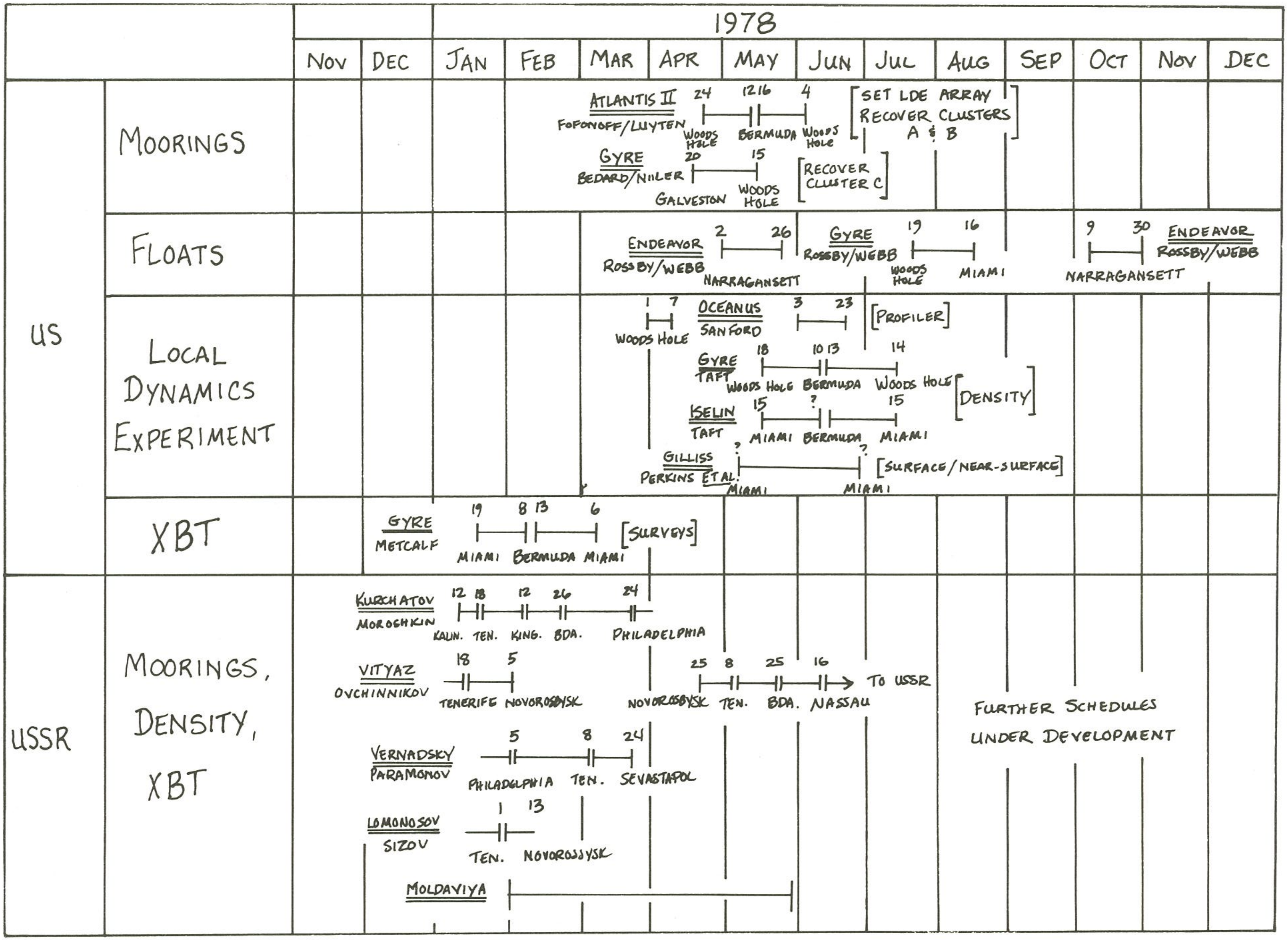
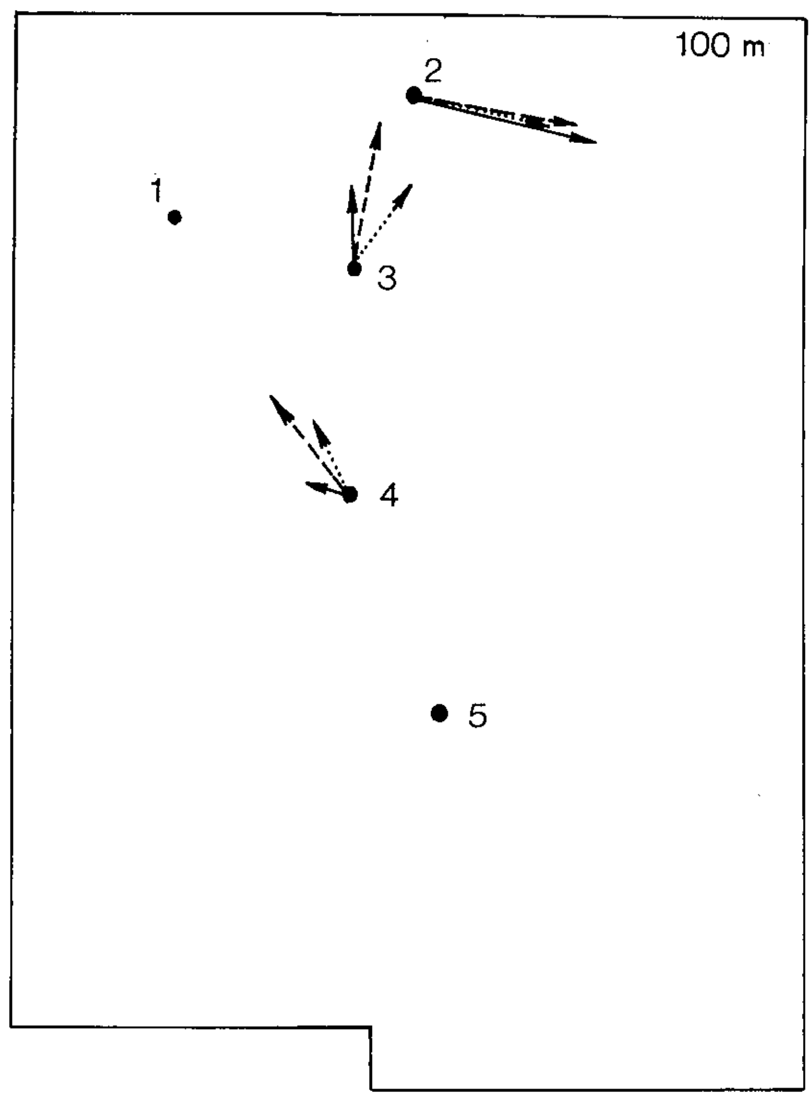


POLYMODE ship schedule.

Table 1



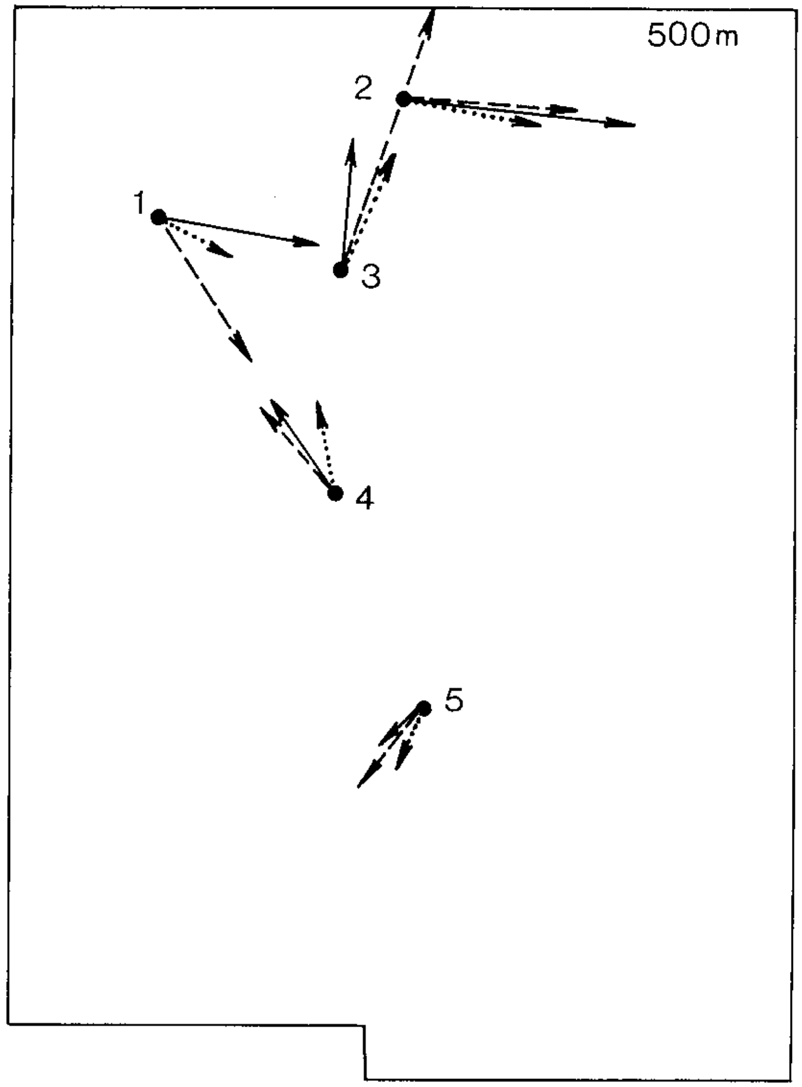
Flow velocity vectors at the 100 m level:

⟶ observed data;

----> calculated with the diagnostic relations (1) to (5);

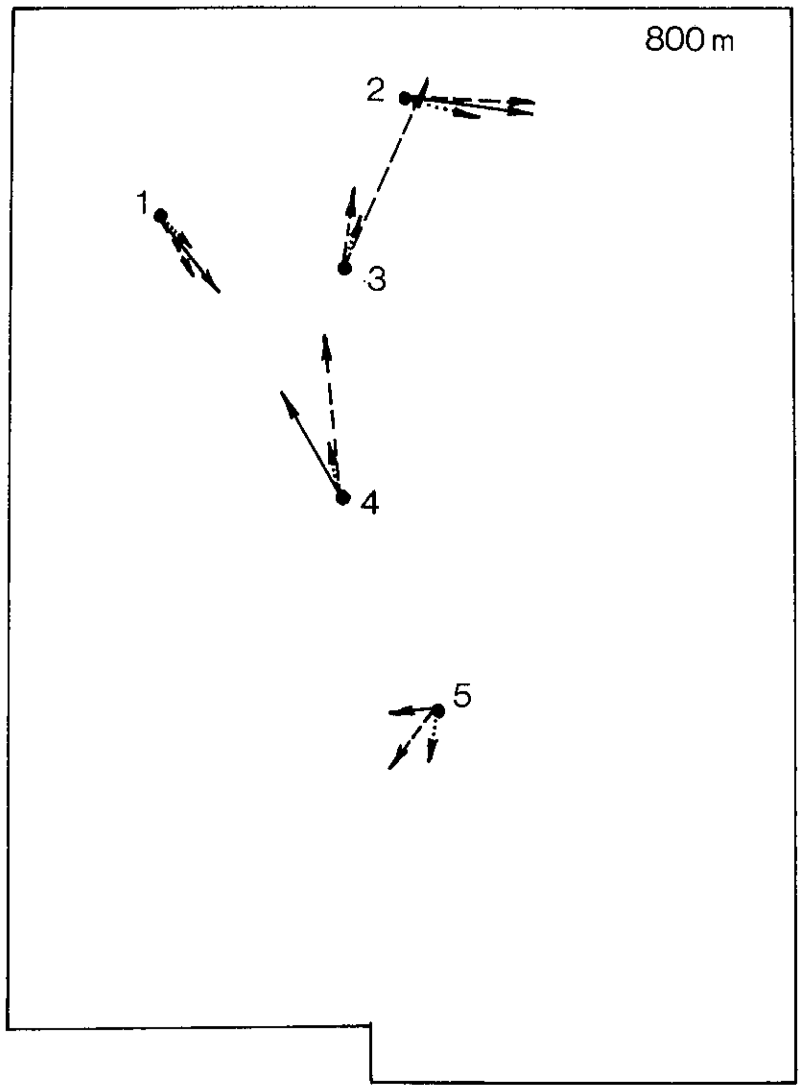
······> calculated with the dynamic method using a 1500 m reference level.

Figure 1 (Sarkisyan *et al*.)



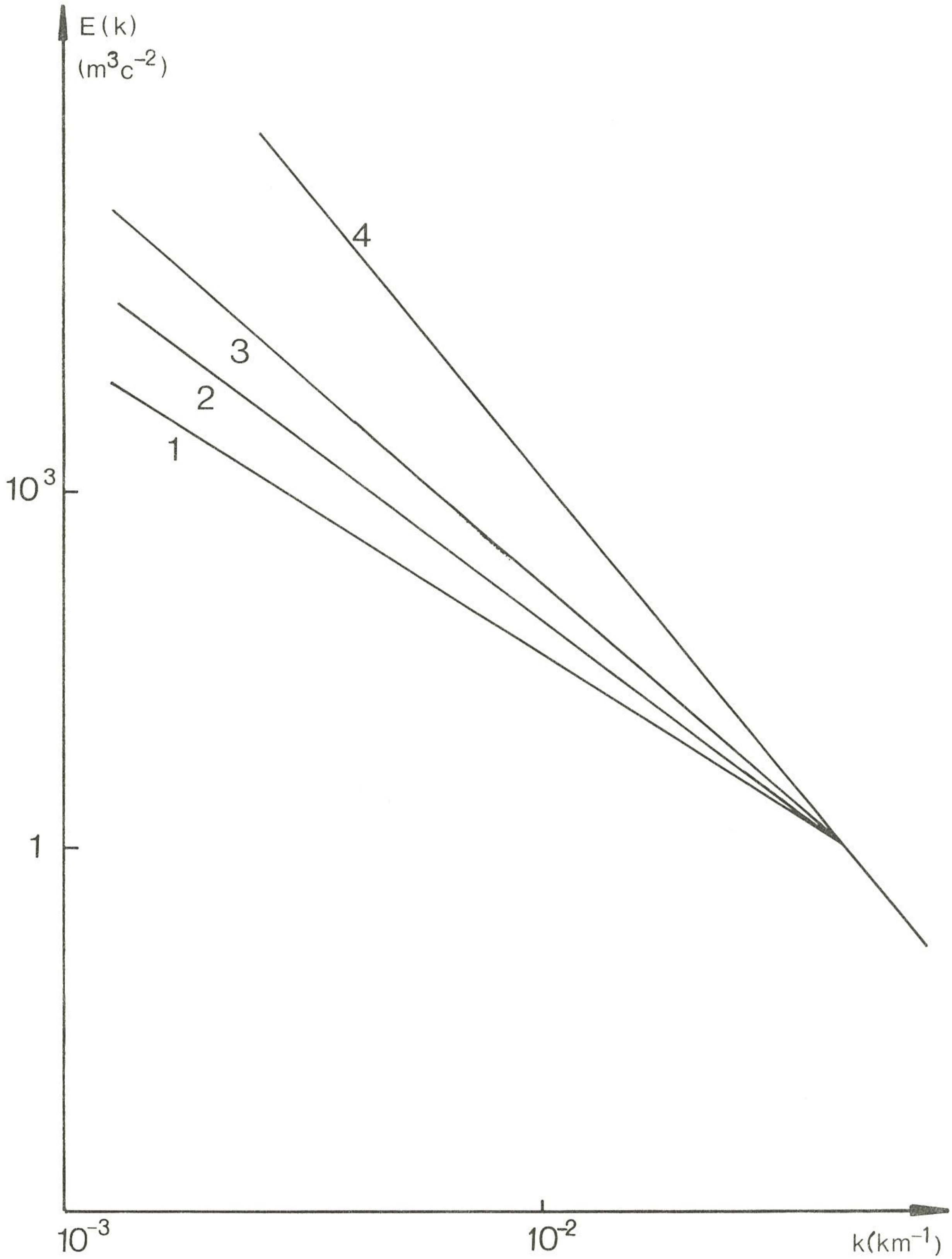
Flow velocity vectors at the 500 m level.
Symbols are the same as in Figure 1.

Figure 2 (Sarkisyan *et al.*)



Flow velocity vectors at the 800 m level.
Symbols are the same as in Figure 1.

Figure 3 (Sarkisyan et al.)



Spectrum of velocity turbulence pulsations:

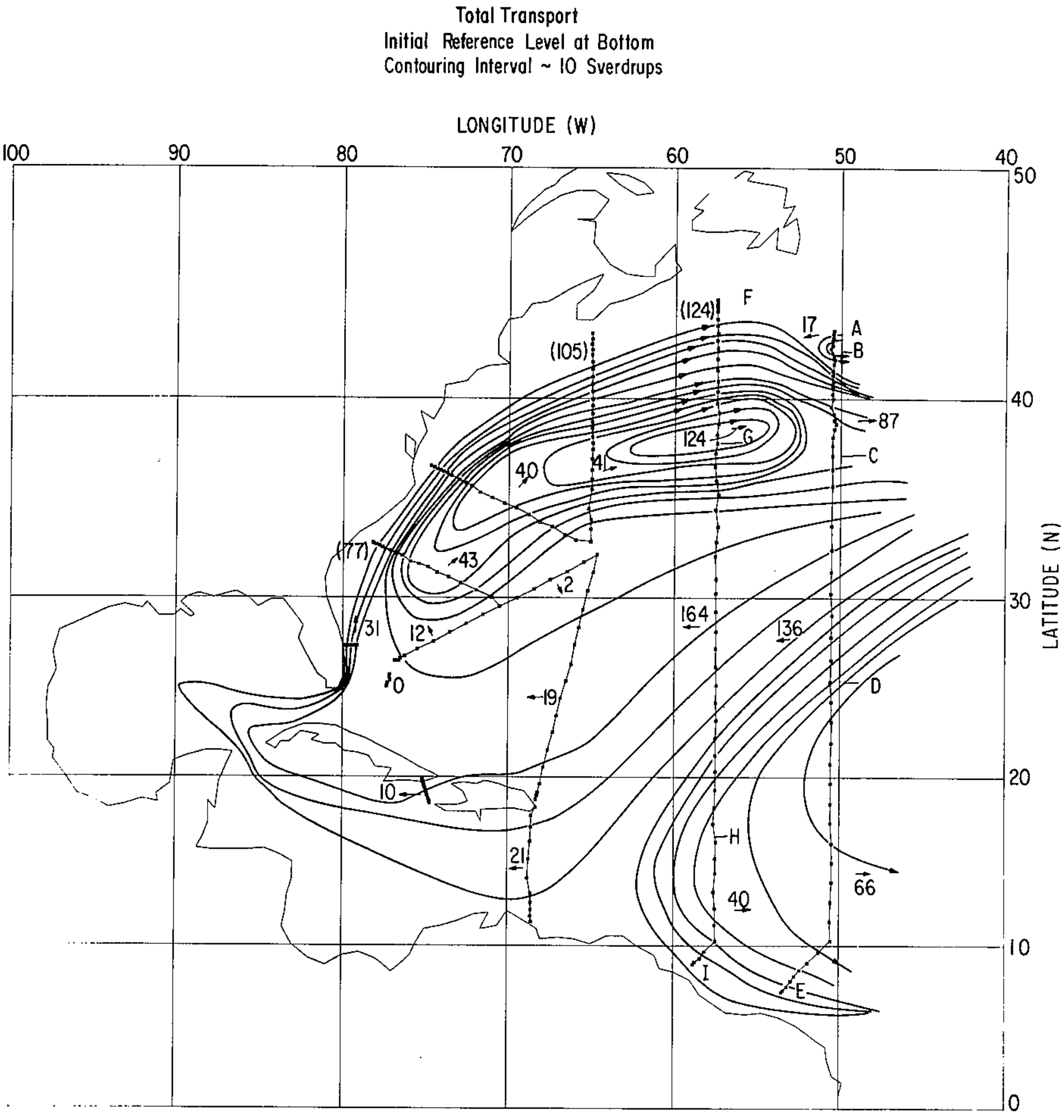
1) at the constant energy flux ($P = 10^{-7}$ cm^2/c^3);

2) at the constant impulse flux (P_J = $0_1 3$ cm^3/c^2);

3) at the constant enstrophy flux ($P_e = 10^{-20}$ c^{-3}); and

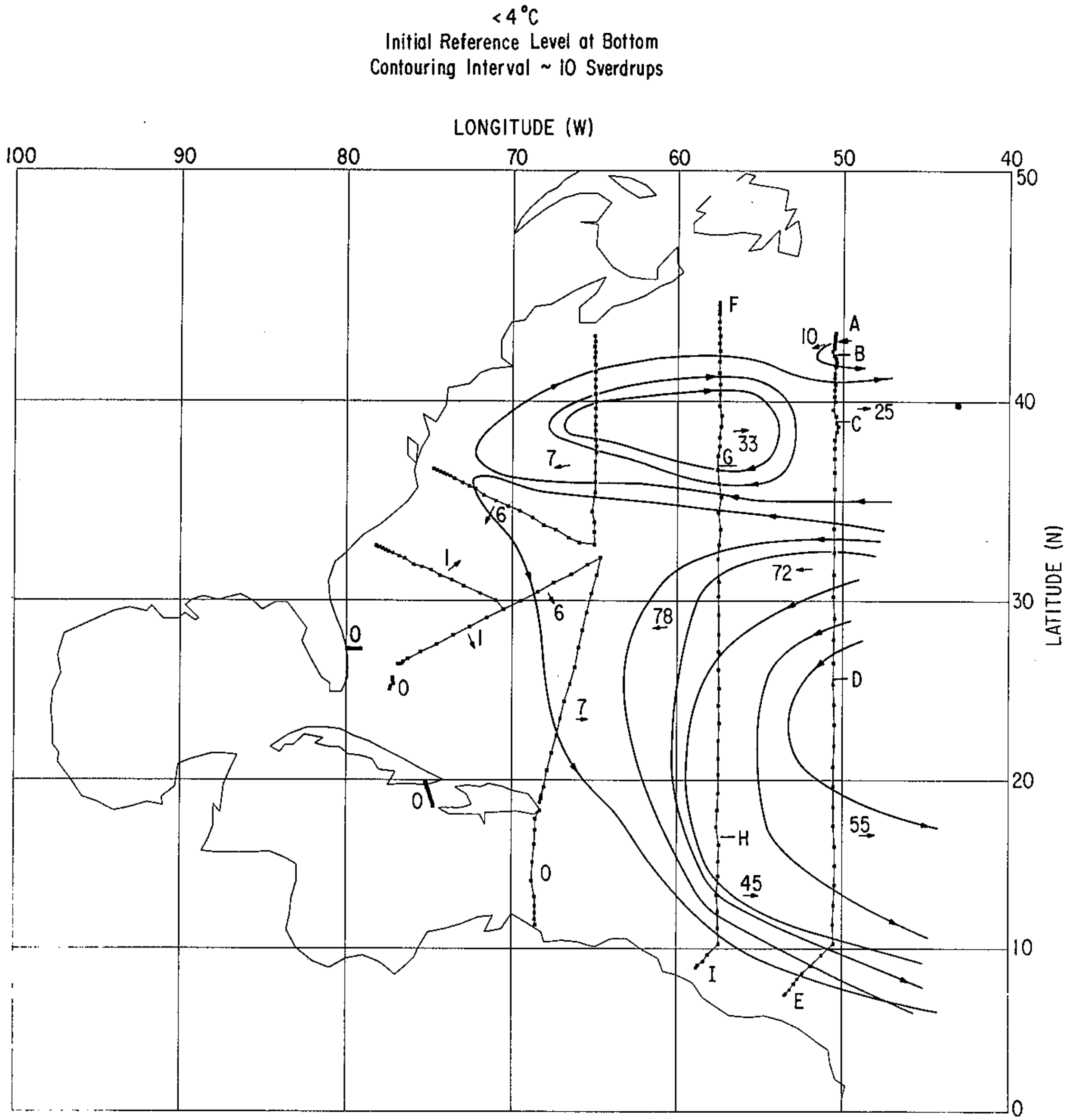
4) saturated spectrum.

Figure 4 (Pelinovsky)

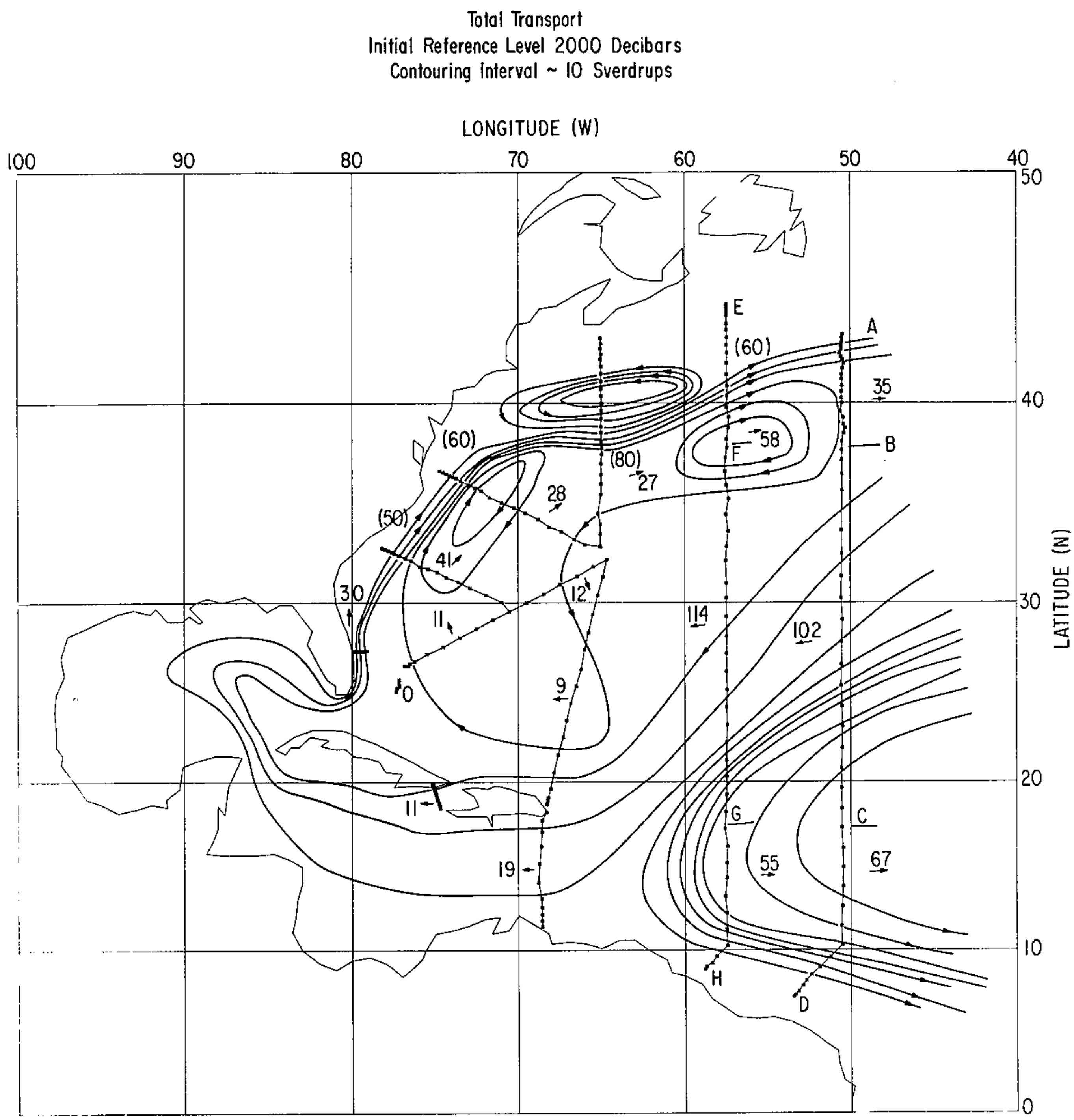


These hydrographic stations divide the ocean into several boxes, each of which must have a balanced mass flow for each water type. Shown here is total transport for model I where the initial reference level was at the bottom. Numbers with arrows denote final transports across the various sections (A-B, B-C, etc. on long meridional lines). Numbers in parentheses are total "Gulf Stream" transports including recirculations.

Figure 5 (Wunsch)

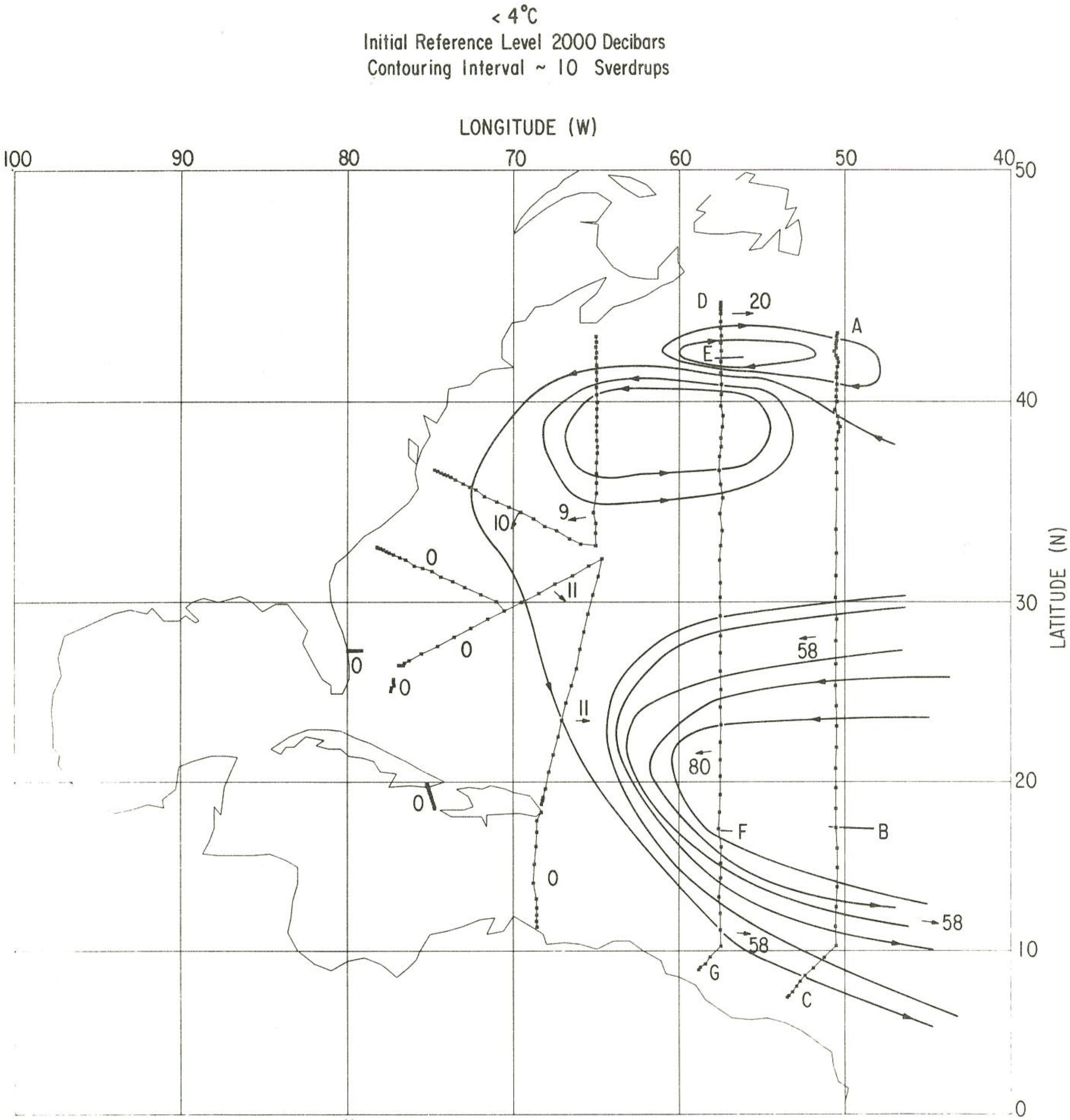


Same as Figure 5, except shown here is transport of water colder than 4°C.
Figure 6 (Wunsch)



Total transports resulting from model II with initial reference level at 2000 dbars. Parenthetical numbers are maximum transports assignable to the Gulf Stream.

Figure 7 (Wunsch)



Same as Figure 7, except shown here is transport of water colder than 4°C.

Figure 8 (Wunsch)