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1. INTRODUCTION

Since 1979, a joint National Weather Service (NWS)-National Environmental Satellite, Data, and Information Service (NESDIS) team has provided an ocean feature analysis for the N.W. Atlantic (the Gulf Stream and rings) and the Gulf of Mexico (the loop current). These analyses are distributed in real time using facsimile and telecopier and are widely accepted and used by commercial fishermen, recreational boaters, commercial shippers, research oceanographers, marine meteorologists, and others. In the past, subjective forecasts have been attempted, but the rules for forecasting Gulf Stream changes are not well established. Also, the task is extremely time consuming. It is evident that some sort of a computer-derived forecast guidance is needed to make timely forecasts.

During 1983, a small ad hoc group of NWS and NESDIS personnel met to discuss the state of the art in Gulf Stream modeling. The group's primary goals were to review the literature and select models which might be adapted for NWS operations. Specifically, the group looked for techniques which could be useful in forecasting currents and temperature distributions in the region of the Gulf Stream which lies north and east of Cape Hatteras, North Carolina (Fig. 1). This report contains a summary of the literature which is pertinent to NWS goals.

2. MODELS OF GULF STREAM BEHAVIOR

In this section, we briefly describe mathematical models of Gulf Stream behavior which are likely candidiates for use in forecasting. Our literature search was limited by time considerations; we make no claim that all published papers have been scrutinized. Our aim was to choose a model which is suitable for operational forecasting, so we concentrated on only certain features of the Gulf Stream. For example, the position of the temperature front, the axis of maximum velocity, and the Gulf Stream's changes over a time scale of from 1 to 10 days are of prime importance. Papers dealing with other topics, such as physical models and Gulf Stream rings, have been excluded.

The Gulf Stream is a narrow band of swift, northward-moving flow on the westward side of the North Atlantic Ocean. It is part of the clockwise-rotating general circulation of the Atlantic and is driven by the mean winds. It has been established for some time that the Stream exists because the Coriolis parameter varies with latitude (Stommel, 1965). It was not well known why the Stream, after it leaves the shallow water of the continental shelf and heads into the deeper ocean, begins to deviate from a steady path and forms meanders which grow and eventually break off to become rings. It was initially thought (Warren, 1963) that topographic control was the dominant mechanism. Later it was recognized that an instability mechanism was at work (Hansen, 1970). It was then determined that baroclinic instability and to a lesser extent barotropic instability were the important energy sources for the meandering (Orlanski and Cox, 1973). On the whole, therefore, the most successful models of Gulf Stream behavior should be those which explicitly include baroclinic effects, generally by having two or more layers.

Gulf Stream properties have been modeled with both Lagrangian and Eulerian approaches. Lagrangian methods describe the position of the major axis of the Stream, and usually rely on a vorticity conservation theorem. Eulerian techniques describe the entire field of flow, from which the Stream is identified by some parameter such as velocity. The Eulerian approach seems to be more suitable for our purposes, but historically the Lagrangian approach appeared first. We include the Lagrangian methods for completeness and to introduce the vorticity equation.

A. Lagrangian Approaches

The earliest methods of analyzing the Gulf Stream were based on the conservation of vorticity equation. This equation comes from taking the curl of the primative equations of motion (see Holton, 1979, chapter 4.4), and neglecting the solenoidal term, the vertical shear (tilting) terms, and vertical advection of vorticity. Then, if we assume incompressible flow and consider only the planetary vorticity in the divergence term, we get the potential vorticity equation

$$\zeta_{,t} + \underset{\sim}{v}\cdot\underset{\sim}{\nabla}(\zeta + f) - f_o w_{,z} = T \quad , \tag{1}$$

where ζ = vertical component of vorticity = $\underset{\sim}{k} \cdot \underset{\sim}{\nabla} \times \underset{\sim}{v}$,
$\underset{\sim}{k}$ = unit normal in the vertical (z-direction),
$\underset{\sim}{v}$ = horizontal velocity vector,
$\underset{\sim}{\nabla}$ = horizontal gradient operator,
f = Coriolis parameter (vertical component of planetary vorticity),
f_o = Coriolis parameter at a reference latitude,
w = vertical velocity,
T = the vertical component of the torque = $\underset{\sim}{k}\cdot\underset{\sim}{\nabla} \times \underset{\sim}{R}$, and
$\underset{\sim}{R}$ = the friction term.

Unless otherwise stated, the x-direction will be toward the east, the y-direction to the north, and the z-direction up. A variable followed by a comma and a subscript indicates the derivative of the variable with respect to the subscript. A comma followed by a repeated subscript indicates a multiple derivative.

A common approximation of the Coriolis term is the linear variation

$$f = f_o + \beta(y - y_o), \tag{2}$$

where $\beta = f_{,y}$ at $y = y_o$, and

y_o = a reference latitude.

The use of Eq. (2) is called the beta plane approximation.

Warren (1963) used Eq. (1) to simulate the position of the Gulf Stream axis. He integrated the equation over a volume of water with sides parallel to the flow and which extended over a fixed depth. Zero vertical velocity was assumed at the top. At the bottom the flow was taken to be parallel to the ocean floor. He applied the steady, frictionless form of Eq. (1) to observed paths as defined by the surface position of the 21-degree isotherm. Because his simulated paths closely matched the data, he concluded that the flow was controlled only by changes in topography, and that stability theory was unnecessary to explain the path.

Hansen (1970) looked at synoptic positions of the Gulf Stream axes at several times and used Warren's (1963) integrated vorticity to describe the path of the Gulf Stream as defined by the position of the 15-degree isotherm at a depth of 200 meters. He found a reasonable match between observed and predicted curvatures along the path. He concluded that bathymetry was important, but it could not account for all the curvature observed. He investigated several stability theories, all of which closely predicted the wave length of maximum growth (observed to be roughly 320 kilometers). The phase speed of the most unstable waves (observed to be about 8 cm/s) was more difficult to predict.

Luyten and Robinson (1974) applied a time-averaged, frictionless form of Eq. (1) to the path as defined by the 15-degree isotherm at 200 meters. They used a scale velocity of 1.5 m/s, and a scale width of 25 kilometers. They compared velocities measured at two points with velocities inferred from their equation. They concluded that the vortex stretching, or topographic control, term was unimportant in the vorticity conservation equation as compared to the local time derivative and the advection terms.

Our review of the above three papers has shown that the Lagrangian approach is of limited value to NWS operational forecasting. This is because the method applies only to regions of the Gulf Stream where the axis can be idealized as a narrow line. Also, the method excludes the possiblility of baroclinic instability, which seems to be an important factor in determining the path.

B. Eulerian Models

The Eulerian approach describes flow properites over the entire field, and is based on the primative equations (P.E.) of fluid flow (see, for example, Pedlosky, 1979, chapter 4.2):

$$\underset{\sim}{v}_{,t} + \underset{\sim}{v}\cdot\nabla\underset{\sim}{v} + w\underset{\sim}{v}_{,z} + f\underset{\sim}{k} \times \underset{\sim}{v} = -\underset{\sim}{P} + A_z\underset{\sim}{v}_{,zz} + A_h\nabla^2\underset{\sim}{v}, \tag{3}$$

where $\underset{\sim}{P}$ = horizontal pressure gradient,
A_z = vertical eddy viscosity, and
A_h = horizontal eddy viscosity.

A mass conservation equation is also used:

$$\rho_{,t} + \underset{\sim}{v}\cdot\underset{\sim}{\nabla}\rho + w\rho_{,z} + \rho(\underset{\sim}{\nabla}\cdot\underset{\sim}{v} + w_{,z}) = 0, \tag{4}$$

where ρ = density (water mass per unit volume).

For the case of constant water density, Eq. (3) becomes

$$\text{Divergence} = \underset{\sim}{\nabla} \cdot \underset{\sim}{v} + w_{,z} = 0. \tag{5}$$

All the following Eulerian models use equations which derive from the P.E. by one or more operations. Equations for a layer are derived by integration over the vertical, so the horizontal velocity becomes either a flowrate per unit width or a vertically averaged velocity. Taking the curl of the P.E. yields an equation for vorticity, a form which is useful because the pressure term is eliminated. The vorticity equation can then be recast by making use of a stream function. Also, an integrated vorticity equation can be derived by taking the curl of the vertically-integrated equations.

There are two categories of Eulerian models: those with a single layer, and those with two or more layers. Single layer models can simulate meander growth caused by barotropic instability alone, while two or more layers are necessary to include baroclinic instability.

Single Layer Models

Hurlbut and Thompson (1980) describe three models they used to test parameters which caused eddy-shedding in the Gulf of Mexico's loop current. They experimented with single- and two-layer models, and with a reduced-gravity model which successfully simulated eddy-shedding with only one layer. The equations are a vertically-integrated form of Eq. (3) on a beta plane. For each layer (i = 1,2),

$$V_{i,t} + (\tilde{\nabla} \cdot \tilde{V}_i + \tilde{V}_i \cdot \tilde{\nabla})\tilde{v}_i + f\tilde{k} \times \tilde{V}_i = -\tilde{P}_i + \tilde{F}_i + A_h \nabla^2 \tilde{V}_i, \qquad (6)$$

where $\tilde{V}_i$ = horizontal flowrate vector per unit width = $H_i \tilde{v}_i$,
$\tilde{P}_1 = gH_1$,
g = gravitational acceleration,
η = deviation from mean sea level = $H_1 + H_2 - D$,
D = still water depth,
H_i = layer thickness,
$P_2 = H_2 \nabla(g\eta - g'H_1)$
$g' = g(\rho_1 - \rho_2)/\rho_1$ (reduced gravity), and
F_i = net horizontal force on layer i due to tangential stresses above and below.

The mass equation is the vertically-integrated form of Eq. (6):

$$H_{i,t} + \tilde{\nabla} \cdot \tilde{V}_i = 0. \qquad (7)$$

When vertical frictional forces are neglected and the lower layer is taken to be inert and infinitely deep, the upper layer pressure term is

$$\tilde{P}_1 = g'H_1 \tilde{\nabla} H_1.$$

Note that this term is similar to the original form of the pressure in the top layer, but with a reduced gravity. They called the resulting flow the internal mode.

The authors found that the two-layer and the reduced-gravity models were able to simulate eddy-shedding, but the barotropic (single-layer) model could not. They concluded that shedding resulted from a barotropic instability in the internal mode, although they did not perform an energy balance.

Multi-layer and Multi-level Models

Here we review the experiments which were intended to elucidate the role of barotropic instability through the use of multi-layer and multi-level numerical models. Here, multi-level is taken to refer to a placement of variables in the vertical direction which is fixed in space. As a result, the divergence, Eq. (5), can be used at each level to find a vertical velocity. Also, the primative equations apply at each level. In contrast, multi-layer refers to a placement of variables in the vertical which is not fixed but changes over time as the layers change thickness. There is, however, no mass exchange between layers, and the vertically-integrated form of the equations is applicable. This terminology is used in the following discussion, with the exception that the expressions upper layer and lower layer will apply to both multi-layer and multi-level models.

Wert and Reid (1972) investigated loop current eddy-shedding with a 2-layer, quasi-geostrophic vorticity model. They began with the vertically integrated form of Eq. (6) and took its curl to eliminate pressure. Their equation, which used a stream function, X_i, was

$$(\nabla^2 X_i)_{,z} + H_i J(X_i, [f + b\nabla^2 X_i]/H_i) - (b^{-1}/H_i)(f + b\nabla^2 X_i) w_i = T_i + A_h \nabla^4 X_i \ , \qquad (8)$$

where

$$\begin{aligned} X_1 &= H_1 + H_2 - D, \\ X_2 &= (\rho_1/\rho_2)H_1 + H_2 - D, \\ J(A,B) &= \text{the Jacobian} = A_{,x}B_{,y} - A_{,y}B_{,x} \ , \\ b &= g/f, \\ w_i &= H_{i,t} \ , \\ T_1 &= b\sigma(X_2 - X_1)/H_1 \ , \\ T_2 &= b\sigma(X_1 - X_2)/H_2 - b\sigma' X_2/H_2, \text{ and} \\ \sigma, \sigma' &= \text{interfacial friction coefficients.} \end{aligned}$$

The quasi-geostrophic approximation is the use of the stream function velocity, which is geostrophic, in all terms of the vorticity equation except the one involving the divergence. Wert and Reid used a layer mass equation (7), and the stream function such that

$$\underset{\sim}{V}_i = b\underset{\sim}{k} \times \underset{\sim}{\nabla} X_i \ . \qquad (9)$$

The authors recast the equation set by subtracting the lower layer equation from the upper layer equation to get a baroclinic mode. They used a variable-depth basin with a top layer thickness of 100 meters, a grid size of 37 kilometers, and a time step of 1.5 hour. They were able to simulate loop current meandering and growth, eddy-shedding, and subsequent eddy decay. However, their numerical scheme required a large amount of computer time. It seems likely that this time can be reduced by approximating several of the terms in Eq. (8) without altering the results.

Orlanski and Cox (1973) simulated a coastal jet on the continental shelf with vertical velocity shear and a stable density distribution. They used a 15-level, rigid-lid P.E. model with Eq. (3), but with

$$P = \rho_o^{-1} \underset{\sim}{\nabla} p,$$

p = pressure,

$_o$ = reference density, and

$f = f_o$.

They used a mass balance like Eq. (4), plus a hydrostatic balance

$$p_{,z} = -\rho g, \tag{10}$$

and a heat balance equation

$$\theta_{,t} + \underset{\sim}{v} \cdot \underset{\sim}{\nabla}\theta + w\theta_{,z} = (A_z/\delta)\theta_{,zz} + A_h\nabla^2\theta, \tag{11}$$

where θ = absolute temperature, and
δ = 1 if $\rho_{,z} \geq 0$
0 if $\rho_{,z} < 0$.

An equation of state was necessary

$$\rho = \rho_o[1 - \alpha(\theta - \theta_o)], \tag{12}$$

where α = thermal expansion coefficient, and
θ_o = reference temperature.

They found that perturbations inserted into the basic flow amplified due to baroclinic instability. They reached this conclusion by constructing an energy budget. They defined mean quantities as the spacial average in the direction parallel to the coast. The eddy kinetic energy increase was due primarily to a decrease in the eddy potential energy, which is the defining characteristic of baroclinic instability. Their region of study, however, was limited to the continental shelf and shelf break, and excluded the open ocean.

Holland and Lin (1975a, b) investigated the wind-driven flow in a two-layer, rectangular, rigid-lid, uniform-depth basin simulating the North Atlantic. The upper layer was 1000 meters thick; the lower was 4000 meters. They used the vertically-averaged equations (6) and (7), but with $F_i = 0$. They found that mesoscale eddies (of Gulf Stream ring size) were the principle means of redistributing kinetic energy throughout the basin. The eddies took approximately 90% of their energy from the potential energy of the system. The flow reached a statistical steady state after 800 to 900 days, when the eddies had a mean wavelength of 440 kilometers, and propagated westward at 6 cm/s. The deep flow was driven by the eddies, not the mean flow. The horizontal viscosity had to be low enough to permit instability, and a grid size of not greater than 20 kilometers was necessary for eddy formation. This grid size allowed 20 points per wavelength. The length scale of the eddies was established by the radius of deformation,

$$R_d = (gH_1H_2/(H_1 + H_2))^{1/2}/f,$$

which for this case was 55 kilometers. The eddies were generated in the gyre's return flow. However, the flow was strongly dependent on the lateral boundary conditions (slip or no-slip). A no-slip condition added extra vorticity to the flow, which distorted it.

Rhines (1976, 1977) studied the evolution of a jet in a two-level, rectangular, uniform-depth basin. He used the vorticity conservation equations at each level with a higher-order horizontal eddy viscosity. With a stream function, Eq. (1) becomes

$$(\nabla^2\chi_i)_{,t} + J(\chi_i,\nabla^2\chi_i) - f_o w_{,z} = T_i \; . \tag{13}$$

For the vertical velocity, he used continuity, Eq. (4), for zero divergence, so that

$$w = \bar{H}/(\rho_2 - \rho_1)[\rho_{,t} + \underset{\sim}{v} \cdot \underset{\sim}{\nabla}\rho],$$

where $\bar{H}$ is the mean of the layer depths. If we combine the above relation with the hydrostatic Eq. (10) and the pressure-stream function relation, which is

$$\underset{\sim}{\nabla} p_i = \rho_i f_o \underset{\sim}{\nabla} X_i,$$

we get, for the top layer,

$$w_{,z} = (\bar{H} f_o/g')X_{i,zz}.$$

The final equation at each level is

$$\{\partial/\partial t + J(X_i, \ )\}[\nabla^2 X_i + \varepsilon_i f_o^2/(g'H_i)(X_2 - X_1)] + \beta X_{i,x} = T_i \ , \qquad (14)$$

where $\varepsilon_i = 1$ if $i = 1$
$= -1$ if $i = 2$,
$T_i = A_2\nabla^2 X_i + A_h\nabla^4 X_i + A_6\nabla^6 X_i$, and
A_2, A_6 = horizontal eddy viscosities.

The solution, obtained by a Fourier expansion of the stream function, showed how eddies coalesced into larger features. Baroclinic instability cascaded energy into the barotropic flow, so that the lower layer eventually behaved like the upper layer. Two-dimensional turburlence acted differently from three-dimensional turbulence since vortex stretching in the vertical is limited. Two-dimensional eddies, although they initially take energy from the mean flow, tend to coalesce over time and can return energy to the mean flow. Meanders began to develop in 20 days and broke up by 40 days. The flow was nearly barotropic after 60 days. The Gulf Stream was hardly distinguishable in the velocity field, but was readily apparent in the density field.

Semtner and Mintz (1977) studied the flow in the western North Atlantic Ocean. The region was modeled as an enclosed basin, driven by a multiple gyre wind stress distribution and surface heating. They solved the primative equations at five levels with the rigid-lid assumption. They used the momentum equations (3), but with a biharmonic diffusion formulation, so that in Eq. (3) $A_h\nabla^2 V$ is replaced by $-B_h\nabla^4 V$. The biharmonic form differed from the Laplacian form in that it smoothed only at the smallest wavelengths. The grid size was 37 kilometers. They included a mass equation (5), a hydrostatic relation (10), a thermal balance (11), and an equation of state (12).

They found that meanders, which developed over the continental slope, got 70% of their energy from the mean potential energy and 30% from the mean kinetic energy. This showed that baroclinic instability was the dominant mechanism. Eddies tended to be reabsorbed into the mean flow out over the abyssal plain.

Holland (1978) investigated the North Atlantic circulation with a two-level, quasi-geostrophic vorticity equation. The system was driven by wind, and included bottom friction but not interfacial friction. He used a vorticity equation like Eq. (13), except that the torque term was

$$T_i = A_h \nabla^4 X_i - B_h \nabla^6 X_i + \begin{cases} H_1^{-1} \underset{\sim}{\nabla} \times \underset{\sim}{\tau}_s & \text{for } i = 1 \\ -\sigma \nabla^2 X_2 & \text{for } i = 2, \end{cases}$$

where $\underset{\sim}{\tau}_s$ = the surface wind stress.

The vertical velocity comes from a thermal balance equation, and is

$$(g'/f_o)w = (X_2 - X_1)_{,t} + J(X_1 - X_2, [H_1 X_1 - H_2 X_2]/(H_1 + H_2)).$$

The author's main conclusion was that mesoscale eddies were important for three reasons. They limited the amplitude of the upper-level mean flow, they propagated energy downward, and they drove the deep flow through the downward momentum flux. In addition, the model ran an order of magnitude faster than the Holland and Lin (1975a, b) P.E. version.

Semtner and Holland (1978) simulated the North Atlantic circulation in an enclosed basin, and compared their results to those of their earlier P.E. (Semtner and Mintz, 1977) and vorticity models (Holland, 1978). They found that the essential features of the energy balance, including the role of baroclinic instability, could be simulated by the two-level model, provided the upper level thickness was properly set. The P.E. model showed that the upper layer had to be roughly 100 meters thick, or equal to the jet thickness. Also, the time savings can be significant.

Ikeda (1981) looked at the growth of a single-wavelength meander in a two-level, uniformly sloping bottom, open sided basin. He used the vorticity equations (14) in dimensionless form, but with $T_i = 0$, and

$$\beta' = \beta + (f_o/H_2)\, D_{,y} \;,$$

in the bottom layer. There was no flow out the northern and southern boundaries, and the east and west boundary conditions were identical, forcing a periodic solution. A small meander was imposed on the mean flow, and the meander amplified until an eddy was shed. For the flat-bottom case, the eddy detached only when beta was non-zero. When beta was zero, a bottom which sloped down to the south created a topographic beta effect which also had to be non-zero to detach the eddy. If the slope was too large, the eddy was not shed, indicating that barotropic instability is influential.

Ikeda and Apel (1981) simulated meanders in a two-level jet in an open, rectangular basin. The stream function and vorticity boundary values were specified on the western side. An outflow condition was set on the eastern side. They used the equations of Ikeda (1981). An initial small meander with a 200-km wavelength was added to the mean flow, and it amplified as it propagated downstream. They found that the nonlinear advective terms tended to stabilize the smaller wavelength meanders. The velocity in the bottom layer was important in controlling the instability in two ways. If it were too high, the meanders propagated out of the basin before shedding eddies.

Also, it had to be low enough to allow baroclinic instabililty. They found that the outflow condition eventually began to distort the interior flow.

Chao and McCreary (1982) examined the bimodality on the axis of the Kuroshio with the two-level model of Holland (1978), but with an inert lower layer. They used the beta plane approximation and biharmonic mass diffusion. The operative boundary condition was set at the eastern boundary. They found that the path of the Kuroshio was moderately dependent on the transport and highly dependent on the spin-up conditions. Because the lower layer was inert, the model effectively had only a single layer and baroclinic instability was suppressed.

3. A CANDIDATE MODEL

This brief review of existing models has revealed some of the more successful approaches and the range of parameters likely to give useful results. A two- or three-level, quasi-geostrophic vorticity model like that of Rhines (1977) or Ikeda (1981) seems to be the best approach because it is relatively simple and fast, and it incorporates baroclinicity. This type of model consists of an equation for the stream function at each level, and the mass equation to find the vertical velocity. The method of Holland (1978) uses a thermal equation for the vertical velocity; it is not known what advantages this offers.

The area to which we would like to apply a model includes the North Atlantic Ocean from Cape Hatteras (34°N) to the Nova Scotian rise (45°N), and east from North Carolina (76°W) to 50°W (see Fig. 1). A grid size of 20 kilometers should be suitable. This means the north-south number of gridpoints is 60, and the east-west number is 90. For a time step, Semtner and Holland (1978) used values up to 6 hours for their model which had a grid size of 37 kilometers.

The thickness of each layer must be specified. Semtner and Holland (1978) found that the upper layer thickness in their two-layer model was an important parameter in determining the flow. Their results were more realistic when the thickness was reduced from 500 meters to 200 meters. However, the stability of the flow in the model with the thinner layer was more sensitive to topographic influences. The authors tested a three-layer model which removed the sensitivity by confining the topographic effect to the lowest layer.

For initial conditions, this model would require temperature (or density) and velocity (or stream function) data at all gridpoints in the horizontal plane and at all levels.

There may be a problem specifying the proper boundary conditions. Most of the studies were performed on enclosed basins, so that mass transport across the sides was zero. This condition is appropriate only when a very large area is modeled, such as one covering most of the North Atlantic Ocean. Confining the domain of interest to the smaller area described above means using an open boundary. Orlanski and Cox (1973) and Ikeda (1981) solved the problem by prescribing periodic conditions at the open sides; this solution is not appropriate to our problem. Ikeda and Apel (1981) specified the stream function for the input condition. For the output condition at a boundary perpendicular to the x-direction, they used

$$X,_t + cX,_x = 0.$$

The phase speed, c, was estimated from changes in the fields near the open boundary. The authors report that the flow was distorted by the use of this condition. Chao and McCreary (1982) specified the stream function at the outflow boundary, which they placed at a geographical location where the Kuroshio was well-defined. Unfortunately, the Gulf Stream is not well-defined at its down-stream extremity. Much care will be required for specification of the outflow boundary condition.

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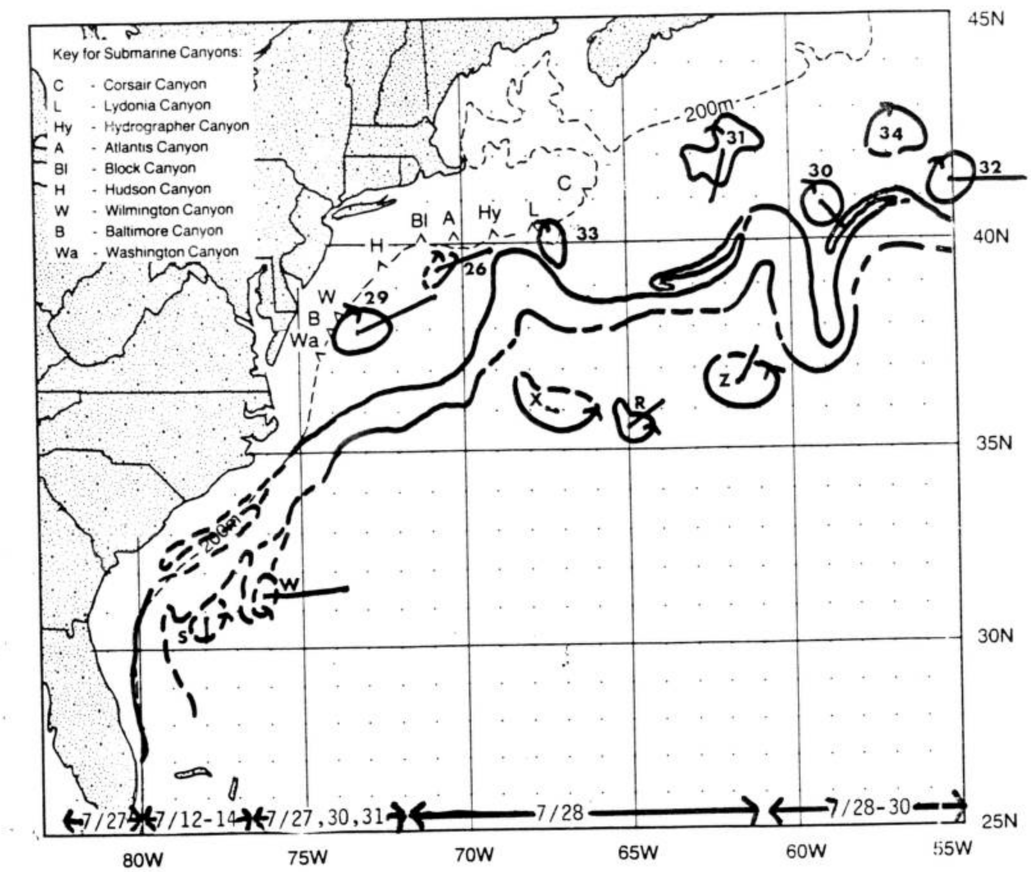

Figure 1. Surface feature chart for July 1983 as prepared by NWS and NESDIS.[1] The solid, more or less continuous line shows the position of the Gulf Stream front (also called the north wall). The front separates the cool shelf waters to the north from the warmer Gulf Stream water to the south. The roughly parallel, but more broken, line below the front defines the south wall, which separates the Gulf Stream from the slightly cooler water of the tropical Atlantic. A large meander system is visible near 40°N from 57°W to 62°W. A smaller meander loop is centered near 39°N, 68°W. Seven rings are shown to the north of front, and five to the south. These rings, which propagate slowly westward, are created when meanders grow large and become separated from the main flow.

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[1]U.S. Department of Commerce, 1983: <u>Oceanographic Monthly Summary</u>, 3, (7) 23 pp.