ASPECTS OF THE NORTHERN BERING SEA ECOHYDRODYNAMICS

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INTRODUCTION

The Northern Bering Sea is a relatively shallow basin limited by the Bering Strait to the north and St Lawrence Island to the south (Fig. 1). The flow passing through the Bering Strait, from the Pacific Ocean to the Arctic Ocean, penetrates the Northern Bering Sea through the Strait of Anadyr, to the west of St Lawrence Island, and by the Strait of Shpanberg, to the east. More than 60% of the mean northward transport of water through the Bering Strait is derived from the "Anadyr Stream", a subsidiary of the Bering Slope Current which flows around the coasts of the Gulf of Anadyr, following the 60-70 isobaths, to the Anadyr Strait and the western part of the Shpanberg Strait (Coachman et al, 1975). The proportion of that stream which goes through the Strait of Anadyr or skirts St Lawrence Island, as well as the orientation, with respect to the Strait's axis, and seasonal variations of the entering flow, is likely to have a strong influence on the subsequent deployment of that flow in the Northern Bering Sea and in the Chukchi Sea.

Observations suggest that the Anadyr stream is the main source of nutrients and biological productivity in the Northern Bering Sea (Walsh et al, 1985).

In preliminary studies for the ISHTAR Research Project (e.g. Walsh et al, 1985), a series of numerical simulations were performed, with 2D barotropic and 3D baroclinic mathematical models, to test this hypothesis and determine if the residual circulation pattern in the Northern Bering Sea was indeed compatible with observed biological data (Walsh and Dieterle, 1986; Nihoul et al, 1986).

The results of these exploratory simulations confirm the general trend of the Anadyr Stream to spread to the east after passing the Anadyr Strait; the nutrient rich Anadyr waters deploying eastwards and progressively fostering biological productivity in the whole basin (Fig. 2, 3, 4).

Studies of the year-to-year variability of the flow pattern reveal however the existence of occasionally strong secondary flows in the form of eastwards propagating interleaving layers of frontal origin. These layers may contribute significantly to the cross-stream diffusion of nutrients and subsequent
Fig. 1. The Northern Bering and Chukchi seas including Bering Strait.

It is shown in the following that the main features of these layers can be explained by a simple model of baroclinic instabilities predicting length scales and time scales in excellent agreement with the observations.

A scenario of the Northern Bering Sea Ecohydrodynamics, including the general circulation pattern and the local frontal secondary flows, is then presented as a working hypothesis to be tested by field surveys and mathematical models.
Fig. 2. Current pattern in the top layer of the Northern Bering Sea for a total flow through the Bering Strait of $1.8 \times 10^6 \text{m}^3\text{s}^{-1}$. 
Fig. 3. Current pattern in the middle layer of the Northern Bering Sea for a total flow through the Bering Strait of $1.8 \times 10^8 \text{m}^3\text{s}^{-1}$.
Fig. 4. Current pattern in the bottom layer of the Northern Bering Sea for a total flow through the Bering Strait of $1.8 \times 10^6 \text{m}^3\text{s}^{-1}$.
The dateline ergocline

The marine system is characterized by fairly well-defined "spectral windows", i.e. domains of length-scales (inversely, wave-numbers) and time scales (inversely, frequencies) associated with identified phenomena. These windows may correspond to eigenmodes of the system (internal waves, inertial oscillations, Rossby waves, El Niño ...) or external forcing (annual or daily variations of insolation, tides, storms, atmosphere climate changes ...) (e.g. Monin et al 1977, Nihoul 1985).

In general, time scales and length scales are related and it is customary to associate high frequencies and high wave numbers, small frequencies and small wave numbers although the association may be different for eigenmodes and forced oscillations.

The transfer of energy between windows is effected by non-linear interactions.

Chemical and ecological interaction processes can also be characterized by specific time scales and the comparison between these time scales and those of hydrodynamic phenomena indicates which processes are actually in competition in the sea (Nihoul 1984, Denman and Powell 1984).

Obviously, at hydrodynamic scales much smaller than interaction scales, very little interaction takes place over time of significant hydrodynamic changes and basically the constituents are transported and dispersed passively by the sea. On the other hand, hydrodynamic processes with time scales much larger than interaction scales scarcely affect the dynamics of interactions over any time of interest.

Mesoscale, synoptic scale and seasonal scale processes in the 10⁻⁶ - 10⁻⁷ s⁻¹ range of frequencies form what one tends to call now the "weather of the sea" while longer time scale phenomena affect both the oceanic and the atmospheric climates.

Independently of climate problems, the year-to-year variability of the marine system, associated with globalscale processes, may be an important element in forecasting the reserves (population dynamics) and permissible hauls of the basic commercial fish (Nihoul 1985).


Although year-to-year variability is a climatic effect, one of its main consequences is a modification of the typical sea-weather patterns. This may result from changes in the general oceanic circulation and energy transports or (and) from similar changes in the atmosphere with differences in typical atmospheric-weather patterns over the marine area which is being investigated.
Thus, some exceptional years, rather important modifications will be observed in meso- and synoptic scale processes characteristic of a particular time of the year and these will entail changes in the flow field and in many chemical and biological processes which depend on transport and dispersion.

A typical illustration of this phenomenon can perhaps be found in a sequence of six remote sensing photographs taken on June 18, June 19, July 15, July 16, August 1 and August 30, 1984. These photographs show a marked plume of cold water, originating near Cape Chukotski on the Soviet Coast and spreading in the Northern Bering Sea under the effect of the general Northward circulation and lateral instabilities, eddies and extrusions progressing towards the East (Fig. 5).

Fig. 5. Thermal image of the Northern Bering Sea for 16 July 1984.
(Cold water in white)

Considering the climatological wind field for this time of the year, this plume could be due to a rather intense upwelling bringing, to the surface, cold bottom water with high nutrient concentrations.
The extruding layers which have typically a width of the order of 10 km, in the early stages of development, widen progressively as they flow eastwards, spreading the nutrient rich water over the Northern Bering Sea.

As shown in the next section, the formation of such layers can be explained by a baroclinic instability of the frontal edge of the cold plume.

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**Fig. 6.** Distribution of temperature °C at 5 m observed by "Brown Bear", 26 July-28 August 1960 (from Fleming and Hegarty, 1966).
Evidence of similar fronts, in the region and general direction of the dateline in the Northern Bering Sea can be found in several field surveys (e.g. Coachman et al. 1975, fig. 6) but their intensity can be highly variable and, in this respect, the 1984 summer situation may have been exceptional.

Nevertheless the dateline front, when it occurs, constitutes an extremely efficient "ergocline" and the cross-front transport by the extruding layers is equivalent to a rather intense lateral mixing, extending the region of biological production and determining to a large extent the amount of organic matter which is ultimately transporter to the Chukchi Sea and further.

A simple model of baroclinic instability

The problem of baroclinic stability has been extensively studied and many models, with various degrees of sophistication and numerical skill, can be found in the literature (e.g. Eady 1949, Stone 1966, 1970, 1971, Tang 1971).

In a simple form, appropriate to the dateline ergocline in the Northern Bering Sea, the problem can be described as the determination of the conditions of instability of a depth-dependent horizontal current, flowing parallel to constant buoyancy surfaces in a region of significant buoyancy gradients (Fig. 6). The fastest growing mode of the linear stability problem gives rise to the observed extruding layers and its length-scale sets the width of these layers in the initial stages of development.

The basic equations applicable to this problem are the inviscid Boussinesq equations, viz.

\[ \nabla \cdot \mathbf{v} = 0 \]  
\[ \frac{\partial \mathbf{v}}{\partial t} + \mathbf{v} \cdot \nabla \mathbf{v} + f \mathbf{b} + \nabla \beta = - \frac{\partial p}{\partial z} + \beta \frac{\partial b}{\partial x} \]  
\[ \frac{\partial b}{\partial t} + \mathbf{v} \cdot \nabla b = 0 \]

where \( \mathbf{v} \) is the velocity, \( f \) the Coriolis frequency, \( b \) the buoyancy, \( \beta = - \frac{\partial p}{\partial z}, \) \( g \) the acceleration of gravity, \( \rho \) the density (\( \rho_0 \) its constant reference value), and where

\[ q = \frac{\rho}{\rho_0} + g x_3 \]

Only two remote sensing photographs were available and clear enough for other years: 18 July 1982 and 13 June 1983. If one goes looking for it, one can see, on them, traces of the same coastal upwelling, resulting in a small, quickly diffused plume carried along into the Northern Bering Sea but the event is of no comparable importance and may be smeared in the general water circulation pattern.
(\(e_3\) is the unit vector along the vertical axis pointing upwards).

For the purpose of this study, taking into account that the length-scale of the perturbation is much smaller than the characteristic scale of variation of the basic flow, one may assume that the latter is horizontally uniform and extending to infinity. The buoyancy and velocity fields, in the unperturbed state, are then given by

\[ \nabla b_0 = -m^2 e_2 + n^2 e_3 \]  

(5)

\[ f e_3 \wedge \frac{\partial v}{\partial x_3} = -\nabla b_0 \]  

(6)

i.e.

\[ v_0 = (U + \frac{m^2}{f} x_3) e_1 \]  

(7)

where \(n\) and \(m\) are, respectively, the "vertical" and "horizontal" Brunt-Väisälä frequencies, i.e.

\[ n^2 = \frac{db}{dx_3^2} \quad m^2 = \frac{db}{dx_2^2} \]  

(8),(9)

and where \(U\) is a constant of integration representing an eventual regional large scale flow directed along the front (i.e. with the approximation made in the Northern Bering Sea, parallel to the coast).

Assuming constant depth \(h\) and constant values of \(m\) and \(n\), one can define a length-scale

\[ \lambda = m^2 h f^{-2} \]  

(10)

and the following non-dimensional variables and parameters

\[ x = x_1 \lambda^{-1} \quad y = x_2 \lambda^{-1} \quad z = x_3 \lambda^{-1} \]  

(11),(12),(13)

\[ \tau = ft \quad \tau = \tilde{f} \frac{m^2}{h^2} U^{-1} \]  

(14),(15)

\[ a = \tilde{f} \frac{m^2}{h^2} U^{-1} \quad \omega = \tilde{\omega} \frac{m^2}{h^2} U^{-1} \]  

(16),(17)

\[ u = \tilde{u} U^{-1} \quad v = \tilde{v} U^{-1} \]  

(18),(19)

\[ r = f^2 n^2 m^{-4} \quad a = \tilde{U} \frac{m^2}{h} \]  

(20),(21)
where \( \ast \) denotes a perturbation of the basic state.

The non-dimensional form of the Boussinesq equations for the perturbation may then be written, in the quasi-hydrostatic approximation,

\[
\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0 ; \quad \frac{\partial \pi}{\partial x} = \alpha 
\]  
(22), (23)

\[
\frac{\partial u}{\partial t} + (s + z) \frac{\partial u}{\partial x} + \omega - \nu = - \frac{\partial \pi}{\partial x} 
\]  
(24)

\[
\frac{\partial v}{\partial t} + (s + z) \frac{\partial v}{\partial x} + \omega = - \frac{\partial \pi}{\partial y} 
\]  
(25)

\[
\frac{\partial w}{\partial t} + (s + z) \frac{\partial w}{\partial x} - \nu + \nu \omega = 0 
\]  
(26)

The perturbation are assumed to be small and to have time and space dependences of the form \( \Phi(x) \exp \{ i (\alpha x + \beta y - \omega t) \} \) where \( \omega \) denotes a complex frequency (instability occurs when the imaginary part of \( \omega \) is positive) and where \( \Phi \) is the appropriate amplitude, function of \( x \).

Substituting in eqs (22) - (26) and solving for \( W(z) \) (the amplitude of \( \omega \)), one obtains

\[
\zeta (\zeta^2 - 1) \ddot{W} + 2(1 - i \gamma \zeta) \dot{W} + W[r(1 + \gamma^2) \zeta + 2i \gamma] = 0 
\]  
(27)

where

\[
\zeta = \alpha x - \omega + \alpha \delta 
\]  
(28)

\[
\gamma = \frac{\beta}{\alpha} 
\]

Eq. (27) belongs to the class of Heun's equations. It must be solved subject to the boundary conditions

\[
W = 0 \quad \text{at} \quad z = 0 \quad (29)
\]

\[
W = 0 \quad \text{at} \quad z = 1 \quad (30)
\]

The boundary value problem set by eqs. (28), (29) and (30) leads to a general complex dispersion relation of the form
\[ R \left( \omega, \omega_i, a, \gamma, r \right) = 0 \]  \hspace{1cm} (31)

where \( \omega_r \) and \( \omega_i \) are respectively the real and imaginary parts of \( \omega \).

Separating the real and imaginary parts of eq. (31) and eliminating \( \omega_r - \omega_i \), one finds \( \omega_i \) as a function of \( a, \gamma \) and \( r \). The perturbation for which \( \omega_i \) is maximum has the largest growth rate and generates the observed cross-ergocline secondary flows.

In the Northern Bering Sea, observations (e.g. Coachman et al. 1975, Sambrotto 1984) indicate that

\[ n^2 \sim 10^{-5} \quad ; \quad m^2 \sim 10^{-13} \quad ; \quad f^2 \sim 10^{-8} \]

hence

\[ r = f^2 n^2 m^{-1} = \frac{\partial n_0}{\partial x_3} \left( \frac{\partial x_0}{\partial x_3} \right)^2 \sim 1 \]

For values of the Richardson number of that order, one can show (e.g. Stone 1966, 1970, Happel et al. 1986) that the maximum growth rate is obtained for

\[ \gamma_{\text{max}} \sim 0 \quad ; \quad a_{\text{max}} \sim 1 \]

The typical wave-length of the fastest growing perturbation is then given by

\[ \lambda = \frac{2\pi}{a_{\text{max}}} = \frac{2\pi m^2 h}{a_{\text{max}} f^2} \]

i.e., taking \( h \sim 50 \text{ m} \),

\[ \lambda \sim 10 \text{ km} \]

in agreement with the observations.

The secondary flow pattern of eastward extruding layers in the Northern Bering Sea may thus presumably be attributed to the baroclinic instability of the dateline ergocline which is formed, in well-defined environmental conditions, at the edge of a plume of cold upwelled water passing through the Anadyr Strait.
A scenario for the Northern Bering Sea Ecohydrodynamics

It is now believed that nutrients are essentially brought to the Northern Bering Sea by the Anadyr Stream and that biological production in the area is closely related to the deployment and residence time there of Anadyr Stream waters, in different environmental conditions.

The occasional occurrence of a marked upwelling plume swept along in the Northern Bering Sea as an unstable frontal current and the subsequent development of extruding layers, flowing eastwards, contribute, in exactly the same way, to the lateral diffusion of the nutrients and one may argue that the productivity of the Northern Bering Sea depends on the intensity and the variability of both the primary and secondary flows.

It is illuminating, in this respect, to examine the results of the Second Soviet-American Expedition in the Bering Sea, 27 June - 31 July 1984, a period characterized, as pointed out before, by a well-marked ergocline event (Sambrotto 1984).

The expedition had selected four polygons of observations; three of which were more or less distributed along the Anadyr Stream. While Polygon I upstream was characterized by relatively high surface temperatures (> 6°C) and high nutrient concentrations, increasing with depth, the transect between Polygons II and III showed a decrease in surface temperature down to 2°C, a depletion of nutrients in surface waters due to active phytoplankton production, high concentrations of ammonia in the bottom layers indicating active microbial degradation of organic matter (as confirmed by microbiological studies) and high rates of organic sedimentation. The additional observation of a larger birds' population at Polygon III suggest that the food chain has completely developed along the Anadyr Stream before it penetrates the Northern Bering Sea and the following ecohydrodynamic scenario appears quite plausible.

Marine productivity in the Northwestern Bering and Chukchi Seas is achieved in two successive phases: one downstream of the shelf-break and the other downstream of St. Lawrence Island. The first involves intense primary production along the "Anadyr stream", with sedimentation of organic matter, both south of Cape Chukotski and St. Lawrence Island, and, generally speaking, in the outer-lagoon of the slowly revolving water in the Gulf of Anadyr's secondary gyre-flow (Fig. 7). The second phase develops north of the Anadyr Strait, spreading to the eastern side of the Northern Bering Sea to a variable extent, due to interannual changes in wind forcing. The eastward excursion of nutrient enriched water is then a function, not only of the intensity and direction of the inflowing current from the Gulf of Anadyr, but also of the intensity of the coastal upwelling, the resulting plume development, and frontal instabilities.
At this stage, the scenario described above can be nothing more than daring hypotheses. A more thorough investigation requires more data (and in particular, more remote sensing data and meteorological information) and the development of mathematical models growing to full three-dimensional maturity.

This will hopefully be the program of the second phase of the ISHTAR project.
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REFERENCES


