THE SPRING TRANSITION FROM HORIZONTAL TO VERTICAL THERMAL
STRATIFICATION ON A MID-LATITUDE CONTINENTAL SHELF

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Abstract

The spring transition from wintertime horizontal thermal stratification to the summertime vertically stratified state on a wide, sloping mid-latitude continental shelf is investigated. The Sigma Z-Level Model (SZM), a new three-dimensional primitive equation numerical ocean model employing a hybrid sigma coordinate and z-level coordinate in the vertical, simulates the continental shelf region. A simple analytical model to describe the heating rate of a well mixed water column is applied to the problem to illustrate the importance of bottom slope in determining the evolution of the horizontal thermal stratification. The applicability of this simple model to continental shelf dynamics forced by realistic surface fluxes is studied with a “pseudo” two-dimensional (x-z) form of the SZM forced with daily heat and wind stress records from the West Florida shelf. A term-by-term analysis of the depth-averaged time-integrated temperature equation quantifies the contribution of each physical processes on the horizontal temperature field. The simple model used to describe the heating rate of a homogeneous water column appears as the dominant balance in the integrated temperature equation. Cross-shelf temperature advection associated with upwelling and downwelling alters the horizontal depth-integrated temperature over the shelf predicted by the simple model. Two processes describe the spring transition. The erosion of the wintertime horizontal thermal gradient occurs on a time scale of a few months when the water column is being heated and mixed periodically. The second process is the formation of the seasonal thermocline following the last strong mixing event. This occurs on a time scale of less than a week under a stabilizing heat flux and light winds completing the spring transition from horizontal to vertical thermal stratification.
1. Introduction

In temperate latitudes, wide continental shelf waters exhibit seasonal variability of thermal stratification. In the winter, the water is well mixed vertically and is stratified horizontally. The temperature gradient is consistently directed in the offshore direction. That is, the water is colder near the coast and warmer offshore with nearly vertical isotherms. Horizontal salinity gradients are also directed offshore due to freshwater runoff near the coast. The temperature and salinity gradients have compensatory effects on the horizontal density stratification such that the horizontal density gradient can be directed in either the onshore or offshore direction. In the summer, the horizontal thermal stratification is relaxed and a thermocline develops resulting in vertically stratified water with nearly horizontal isotherms. A consequence of the formation of the seasonal thermocline is a pycnocline due to the dominating effect of the vertical temperature gradient on the density field. The salinity field is in some instances consistent with vertical stratification, but at some locations is characterized by year-round offshore gradients due to the local freshwater input.

The pycnocline formation is important from a dynamical perspective. During the winter when the shelf waters are homogeneous with depth, the oceanic response to low frequency forcing over the shelf is barotropic. When a pronounced pycnocline develops, the shelf waters can behave much like a two-layer system allowing for a first baroclinic mode response [Millot, 1990]. Additional impacts of this stabilization of the water column can include the enhancement of primary productivity and the onset of a spring bloom [Mann and Lazier, 1991], and the formation of acoustical “shadow zones” [Caruthers, 1977]
The existence of two distinct regimes of stratification, the horizontally stratified wintertime regime and the vertically stratified summertime regime, has been observed on several wide mid-latitude shelves. One such well-documented location is the Middle Atlantic Bight [c.f., Bigelow, 1933; Bigelow and Sears, 1935; Beardsley et al., 1985; Burrage and Garvine, 1988; Houghton et al., 1988; Chapman and Gawarkiewicz, 1993]. Here, the wintertime stratification is horizontal in salinity and temperature with an offshore density gradient, while the summertime stratification is horizontal in salinity and vertical in temperature with a vertical density gradient (Figure 1). That is, the salinity stratification does not show strong seasonal variability and the temperature stratification is largely responsible for the presence of a pycnocline. Hydrographic data from the West Florida shelf (WFS) suggest similar seasonal stratification states [Niiler, 1976; Clarke, 1994; Weisberg et al., 1996]. Winter and summer temperature profiles show that the shelf water is strongly destratified in the winter with a horizontal temperature gradient directed in the offshore direction, and strongly stratified vertically in the summer (Figure 2). An onshore horizontal temperature gradient is evident in the summer [Weisberg et al., 1996]. Other examples of continental shelves on which similar seasonal variations of stratification have been observed include the Gulf of Lions in the western Mediterranean [Millot, 1990], the Sicilian Continental Shelf [Artale et al., 1988], Jervis Bay along the eastern Australian coast [Symonds and Gardiner-Garden, 1994], and the South Brazil Bight [Stech and Lorenzetti, 1992].

The circulation on wide shelves differs from the very steep shelves in that generally the alongshore currents on the wide shelves are weaker. Strong boundary currents along steep narrow shelves, e.g. those of the eastern Pacific Ocean and western
Gulf of Mexico, have a large impact on the local hydrography of the shelf water. On very wide shelves like the WFS, the inner and mid-shelf circulation is isolated from currents with large mean velocities offshore of the shelf break. This suggests that temperature changes on the wide shelves can be more greatly influenced by local heating than on very narrow shelves.

The classic argument to describe the spring transition from the wintertime horizontal stratification to the summertime vertical stratification has been that the relaxation of the winds during the spring combined with the increased solar heating allow a thermocline, and thus a pycnocline, to form. Though the formation of a seasonal thermocline is a fairly well understood process, it is not obvious that this can reduce a horizontal temperature gradient. Thus, some other process or processes may be responsible for altering the density gradients.

The salinity structure and variability on continental shelves is largely a consequence of local freshwater runoff and differs from one shelf to another, but the distinct winter and summer regimes of temperature stratification are a defining feature of wide mid-latitude continental shelf temperature fields. Additionally, the summertime vertical thermal stratification has a dominating effect on the density field. The formation of the wintertime horizontal temperature gradient by convective cooling has been examined in detail [Symonds and Gardiner-Garden, 1994; Pringle 1998], but how this gradient is eroded and reversed is not yet fully understood. The purpose of this investigation is therefore to determine the processes responsible for the spring transition of the thermal stratification on a mid-latitude continental shelf.
The paucity of observations with adequate spatial resolution and temporal sampling concurrent with the spring transition makes this an ideal subject for a modeling study. First, a simple analytical model for the temperature change of a well mixed water column subject to a surface heat flux is applied to the problem of relaxing a horizontal temperature gradient by application of a spatially uniform heat flux. This model cannot be used to explain the complete transition, as it cannot simulate a vertically stratified ocean. Next, the problem is investigated using a recently developed primitive equation ocean model with a hybrid sigma and z-level vertical coordinate useful for modeling a shelf slope and escarpment. A realistic simulation is run with fluxes and topography representative of the WFS. A term-by-term analysis of the depth-averaged time-integrated temperature equation highlights the applicability of the simple mixed-layer model to the problem, but also demonstrates that horizontal advection of temperature can significantly alter the horizontal temperature gradient under certain conditions; in this case, a downwelling event.

2. A simple model

Consider a water column subject to a (positive downward) net surface heat flux \( Q(t) \) (W/m²) and insulated at the bottom. There is no horizontal mixing or horizontal advection. The mean temperature \( T \) of a water column of height \( H \) will change at the rate

\[
\frac{dT}{dt} = \frac{Q(t)}{\rho_0 c_p H}
\]  

(1)

Here, \( c_p \) is the specific heat of the water and \( \rho_0 \) is the average density. Since the time rate of change of temperature is inversely proportional to the water column depth, a shallower
water column will heat more rapidly than a deeper water column subject to the same positive surface heat flux.

Noh and Fernando [1995] demonstrated by way of a laboratory experiment that a fluid column subject to a stabilizing surface buoyancy flux, but mixed vigorously enough to prevent vertical stratification, had a decrease in density with time that was uniform with depth. Thus, if the buoyancy flux takes the form of a heat flux, a well mixed water column will be heated uniformly. Under the assumptions made in the formulation of (1), a vertically mixed body of water over a sloping bottom subject to a spatially uniform heat flux will, at all depths, have a temperature gradient tendency positive in the direction of the shallower water.

As an example, consider a vertically mixed body of water over a sloping bottom of depth \( H(x) = mx \). The change in temperature from some time interval \( t_0 \) to \( t \) will be (neglecting free surface variations)

\[
\Delta T = \frac{\int_{t_0}^{t} Q(t) dt}{\rho_0 c_p mx}
\]

That is, \( \Delta T \) will increase hyperbolically toward the coast. Obviously, this becomes unphysically large in extremely shallow water unless one considers a negative feedback of the surface heat flux to the sea surface temperature. Nevertheless, the process described by (1) and (2) can reasonably explain the formation of the cross-shelf temperature gradient on a continental shelf subject to surface cooling in the absence of cross-shelf eddy transports [Symonds and Gardiner-Garden, 1994; Pringle, 1998].
An example representative of the WFS demonstrates how a spatially uniform constant heat flux could result in the relaxation of a wintertime horizontal temperature gradient. The cross-shelf temperature difference measured in the winter of 1973 over the WFS near the Southwest Florida coast could be reduced from 3° C to nearly 0° C by calculating ΔT from (1) applying a constant heat flux of 30 W/m² for 65 days (Figure 3). This example is clearly too simplistic to represent all the processes that may be responsible for altering the shelf water stratification. It neglects horizontal and vertical motions, stabilization of the water column thus confining the heat input to the surface, and spatial variations in heat flux. Nevertheless, it demonstrates that the simplest thermodynamic forcing and ocean physics could account for the relaxation of the offshore temperature gradient. The question remains whether or not (1) can adequately describe the erosion of the offshore temperature gradient without the assumptions made in developing this simple model. The roles of cross-shelf advection and mixing must be considered in the presence of realistic forcing and dynamics of continental shelf waters. Additionally, the vertical stratification of the shelf water as observed at the end of the spring season must be explored as well.

3. A realistic model

3.1 Description of the Sigma Z-Level Model

The spring transition of thermal stratification is investigated further using the Sigma Z-Level Model (SZM) [Martin et al., 1998] to realistically simulate the ocean dynamics on a continental shelf, specifically the WFS. The modeled region consists of a continental shelf with a shelf break leading to an open boundary toward the deep ocean.
The SZM uses the hydrostatic, Boussinesq, and incompressible approximations, and allows for a free surface. The model equations are

\[
\frac{\partial u}{\partial t} = -\nabla \cdot (uv) + f v - \frac{1}{\rho_0} \frac{\partial p}{\partial x} + \nabla_h (A_m \nabla_h u) + \frac{\partial}{\partial z} (K_M \frac{\partial u}{\partial z})
\]

(3)

\[
\frac{\partial v}{\partial t} = -\nabla \cdot (vv) - f u - \frac{1}{\rho_0} \frac{\partial p}{\partial y} + \nabla_h (A_m \nabla_h v) + \frac{\partial}{\partial z} (K_M \frac{\partial v}{\partial z})
\]

(4)

\[
\frac{\partial p}{\partial z} = -\rho g
\]

(5)

\[
\nabla \cdot \mathbf{v} = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0
\]

(6)

\[
\frac{\partial \zeta}{\partial t} = -\frac{\partial ((\zeta + H)u)}{\partial x} - \frac{\partial ((\zeta + H)v)}{\partial y}
\]

(7)

\[
\frac{\partial T}{\partial t} = -\nabla \cdot (\mathbf{v} T) + \nabla_h (A_H \nabla_h T) + \frac{\partial}{\partial z} (K_H \frac{\partial T}{\partial z})
\]

(8)

\[
\frac{\partial S}{\partial t} = -\nabla \cdot (\mathbf{v} S) + \nabla_h (A_H \nabla_h S) + \frac{\partial}{\partial z} (K_H \frac{\partial S}{\partial z})
\]

(9)

\[
\rho = \rho(T, S, z).
\]

(10)

The boundary conditions at the surface are

\[
\rho_0 K_M \frac{\partial u}{\partial z} = \tau_x
\]

(11)

\[
\rho_0 K_M \frac{\partial v}{\partial z} = \tau_y
\]

(12)

\[
\rho_0 K_H \frac{\partial T}{\partial z} = \frac{Q_x + Q_y + Q_z + Q_s}{c_p} = \frac{Q}{c_p}
\]

(13)

\[
K_H \frac{\partial S}{\partial z} = S\bigg|_{z=0}(E_v - P_r)
\]

(14)
and at the bottom,

\[ K_M \frac{\partial u}{\partial z} = c_b u |v| \]

\[ K_M \frac{\partial v}{\partial z} = c_b v |v| \]

\[ K_H \frac{\partial T}{\partial z} = 0 \]

\[ K_H \frac{\partial S}{\partial z} = 0. \]

In these equations, \( T \) is the potential temperature (henceforth referred to simply as temperature), \( S \) is the salinity, \( A_M \) and \( A_H \) are the horizontal eddy coefficients for momentum and for scalar fields, respectively, \( K_M \) and \( K_H \) are vertical eddy coefficients for momentum and scalar fields, barred variables are the depth-averaged variables, \( \zeta \) is the elevation of the free surface above the undisturbed value at \( z = 0 \), and \( H \) is the bottom depth. In the surface boundary conditions, \( \tau_x \) and \( \tau_y \) are the \( x \) and \( y \) components of the surface wind stress, \( Q_n \), \( Q_b \), \( Q_s \), and \( Q_l \) are the net downward short wave, long wave, latent and sensible surface heat fluxes, and \( Q \) is the net surface heat flux. \( E_r \) and \( P_r \) are the surface evaporation and precipitation rates, and \( c_p \) is the specific heat of seawater.

The bottom stress is parameterized by a quadratic drag law with drag coefficient \( c_b \) that is calculated by

\[ c_b = \max \left( \frac{K^2}{\log^2 \left( \frac{\Delta z}{2z_0} \right)}, c_{b_{\text{min}}} \right) \]
\( \Delta z_b \) is the bottom layer thickness, \( z_0 \) is the bottom roughness, \( \kappa = 0.4 \) is von Karman’s constant, and \( c_{b-} \) is a minimum value for \( c_b \).

The vertical eddy coefficients are calculated from the Mellor-Yamada Level 2 (MYL2) turbulence closure scheme [Mellor and Yamada, 1982], which has been shown to give turbulence mixing scales and turbulent layer depths comparable to the more computationally demanding Mellor-Yamada Level 2.5 turbulence closure scheme [Martin et al., 1998]. Horizontal eddy mixing coefficients are specified as the maximum of a background value \( A_0 \) and the value needed to keep the grid-cell Reynolds number below a maximum specified value \( R_n \). For example, in the x-direction,

\[
A_H = \max(A_{H0}, \frac{u \Delta x}{R_n})
\]

\[
A_M = \max(A_{M0}, \frac{u \Delta x}{R_n})
\]

A similar form of the model equations in SC can be found in Blumberg and Mellor [1987]

The SZM uses a staggered Arakawa C grid, and a hybrid sigma / z-level vertical coordinate as described in the Appendix. Spatial interpolations are second-order centered averages and spatial gradients use second-order centered differences. The model employs a standard leapfrog time-stepping scheme in which the advection, baroclinic pressure gradient, and Coriolis terms are centered in time at the current time step \( n \). Horizontal diffusion terms are computed at \( n-1 \), and vertical mixing terms are treated implicitly. An Asselin [1972] filter is applied to the model fields at the end of each time step to damp out the computational mode when using leapfrog time differencing. The
free-surface mode is calculated implicitly, such that the surface pressure gradient and divergence terms in the surface elevation equation have a component at the time level \( n+1 \).

For the experiments used in this investigation, the SZM is run in a "pseudo" two-dimensional mode by employing a total of four grid points (two interior grid points) and periodic boundary conditions in the alongshore direction. This allows the use of a three-dimensional model for a two-dimensional problem without actually converting the model to two-dimensional form. The domain extends from the coast across the shelf and shelf-break to an open boundary at the deep ocean. All scalar fields at the open boundaries are relaxed to specified values for inflow and Orlanski radiation [Orlanski, 1976] is used for outflow. The constants and model parameters used in this study are shown in Table 1.

3.2 The West Florida Shelf simulation

The WFS is a wide continental shelf on which the spring transition in temperature stratification has been documented [Florida Institute of Oceanography, 1975]. To study the mechanisms responsible for the spring transition, the SZM is used to realistically simulate the region. For this experiment the slope of the continental shelf and surface forcing time series are chosen to represent typical conditions of the WFS. The model topography is defined by a 7 m depth at the first ocean grid point next to the coast and a bottom slope \( m \) of

\[
m = \begin{cases} 
1.04762 \times 10^{-4}, & x < 21 \text{ km} \\
3.36562 \times 10^{-4}, & 21 \text{ km} < x < 203 \text{ km} \\
18.5000 \times 10^{-4}, & 203 \text{ km} < x < 273 \text{ km} \\
0, & 273 \text{ km} < x 
\end{cases}
\] (22)
where \( x \) is the distance from the coast. For \( x > 273 \text{ km} \), the bottom is 200 m deep to the open boundary 413 km offshore. The domain is situated at 26° N with the shoreward direction aligned toward the east. The shelf region from the 10 m isobath to the shelf break will be the focus of the model results (Figure 4).

The ocean model is forced with spatially uniform daily NCEP Reanalysis [Kalnay, et al., 1996] wind stresses and net surface heat fluxes from 1974 at 27.6° N, 84.4° W, a proximal point from the gridded product at mid-shelf (Figure 5). This record was selected for illustrative purposes due to its numerous easily recognized frontal passages. The purpose of this study is to investigate whether there is a simple solution to the problem, that is, can one-dimensional mixing and stratification of a shelf region subject to only spatially uniform forcing account for the spring transition. Obviously, cross-shelf variation in heat fluxes will affect the local rate of change of temperature, but a good physical understanding of the processes responsible for the spring transition can be gained by considering the simple case of spatially uniform forcing. The important property of the surface forcing time record used in this study is that it resolves in time the atmospheric synoptic scale.

The wintertime weather pattern is dominated by the passages of synoptic-scale cold fronts, which move toward the southeast and are characterized by a clockwise rotation of the local wind vector. Cold, dry air masses following the passages of cold fronts are associated with northerly winds with speeds typically 10 to 15 m/s and give rise to negative heat fluxes to the ocean’s surface. These cold fronts become less frequent and produce less intense surface cooling of the ocean late in the season. Springtime meteorological conditions are typified by periods of light winds and
stabilizing heat fluxes followed by frontal passages with associated strong winds and
often periods of surface cooling. A running average of the net surface heat flux shows a
transition from surface cooling to a stabilizing heat flux in the early spring. Though the
spring cold fronts are often associated with a negative heat flux, the heat flux averaged
over several days remains positive for most of the season.

The ocean model is started from rest with constant salinity of 35 PSU and is
initially linearly vertically stratified in temperature from 15° C at 200 m to 22° C at 70 m,
and the upper 70 m is at a uniform temperature of 22° C. The offshore open boundary is
relaxed to this temperature profile on a time scale of three days. Surface momentum and
heat fluxes from the 1974 data record are then applied from year day 1 (January 1) to day
58. This allows sufficient time for the model to achieve a horizontally stratified state
typical of wintertime conditions over the shelf and to adjust to dynamic equilibrium. Day
58 follows the last frontal passage before the transition to a stabilizing heat flux. This
transition is considered to be the beginning of the springtime weather pattern for this
study and marks the beginning of the transformation from the wintertime horizontal
stratification to the summertime vertical stratification.

3.3 The role of vertical mixing and stratification

At Day 58, the transition marking the beginning of the warming heat fluxes, the
ocean is mixed thoroughly to the bottom over the shelf with a cross-shelf temperature
difference of 7° C (Figure 6a). This is a reasonable horizontal thermal gradient as shown
in earlier hydrographic studies of this region [Florida Institute of Oceanography, 1975].
During the next ten days, the light winds allow a thermocline to form between 5 m and 10
m under the stabilizing heat flux (Figure 6b). The heat input is confined only to the surface layer since the thermocline suppresses vertical mixing, thus preventing the vertical transport of heat to deeper water. The horizontal temperature gradient remains even in the presence of the thermocline. A weak cold front accompanied by a brief period of destabilizing heat flux reaching $-180 \text{ W/m}^2$ and a maximum wind stress magnitude of $0.1 \text{ N/m}^2$ passes by day 76. The convective cooling and wind stirring weaken the thermocline and the surface mixed layer deepens to 10 m to 15 m over the shelf (Figure 6c). Where the depth of the surface mixed layer intersects with the bottom boundary layer, the water column is well mixed. The ocean is vertically mixed offshore to the 30 m isobath, and the vertical stratification over the deeper shelf is only about 1°C.

Though the cold front passing by day 76 is accompanied by a negative heat flux, the time-integrated heat flux since the transition to the springtime warming period in day 58 remains positive. From day 58 to day 76, the time-averaged heat flux is 32.3 W/m². This heat input to the surface of the ocean is distributed evenly with depth when the water column is mixed by the winds associated with the frontal passage. Since this heat flux is spatially uniform, from (1) one expects the shallower water to have a larger increase in temperature. This effect is apparent by the reduction of the cross-shelf temperature difference from 7°C on day 58 to 5°C following the passage of the cold front on day 76. The mixing of the water column provides for warming at all depths so the horizontal temperature gradients are reduced equally at all depths over the inner shelf.

Following the frontal passage, the water column again stratifies under the stabilizing heat flux and light winds. This cycle of stabilization of the water column by
surface heating and mixing of the water column resulting in uniform heating with depth continues throughout the spring. The squared buoyancy, or Brunt-Väisälä, frequency

\[ N^2 = \frac{-g}{\rho_0} \frac{\partial \rho}{\partial z} \]  

(23)

is a measure of the stratification of a water column (here, \( \rho \) is not a function of pressure). \( N^2 = 0 \) indicates a mixed water column, while a stratified water column has a positive \( N^2 \).

A time series of depth-averaged \( N^2 \) shows the cycle of stabilization and mixing of the water column in response to the surface fluxes of heat and momentum (Figure 7). The cycle is most apparent over the inner shelf where the mixing reaches the bottom. At each well mixed state following a frontal passage, the horizontal temperature difference is weakened and eventually reversed (Figure 6c-e). The horizontal temperature gradient has been reduced to nearly zero by day 132. Additional heating and weak mixing later in the season reverse the temperature gradient such that it becomes positive toward the coast. The thermocline is allowed to fully develop after the last frontal passage of the season (Figure 6f) completing the transition from the wintertime offshore temperature gradient to the summertime vertical temperature gradient.

It has been reported that the transition from the winter to summer regime is thought to be a rapid event that occurs soon after the last cold front of the season [Florida Institute of Oceanography, 1975], though the transition has never been precisely captured by observations. This present study concludes, however, that the transition is a two-step process. The relaxation and reversal of the horizontal temperature gradient (the first step) actually takes place throughout the spring with a time scale of a few months. The
formations of the transient thermocline between fronts and following the last front (the second step) are, however, rapid events with a time scale of only a few days.

A description of the mechanism responsible for the formation of the transient thermocline is useful for achieving an understanding of the time scale for the water column stabilization. The gradient Richardson number

\[ R_i = \frac{g \frac{\partial \rho}{\partial z}}{\left( \frac{\partial u}{\partial z} \right)^2 + \left( \frac{\partial v}{\partial z} \right)^2} \]  

(24)

represents the ratio of the density stratification and the vertical shear of the horizontal velocity (again, \( \rho \) is not a function of pressure). A negative \( R_i \) indicates an unstably stratified water column and turbulent mixing will intensify. A "small" \( R_i \) indicates that the shear production of turbulent kinetic energy dominates the stabilizing effects of buoyancy, and turbulent mixing is significant. A \( R_i \) that is considered "large" indicates that turbulent mixing is suppressed by the density stratification. The \( R_i \) above which mixing and turbulence ceases is known as the critical Richardson number. This has been estimated empirically to be \( R_c = 0.23 \). Where \( R_i \) exceeds \( R_c \), the turbulence is suppressed and the eddy diffusivity is reduced. When subject to a stabilizing buoyancy input, this will induce an even stronger stratification further suppressing turbulence. This feedback mechanism leads to the formation of a thermocline across which both the turbulent kinetic energy flux and the buoyancy flux are diminished.

The role of the Richardson number is illustrated for the stabilization of the inner shelf water beginning on day 116 following a frontal passage (Figure 8). On day 116, the inner shelf water is completely mixed vertically out to the 30 m isobath. Here, there is no
depth where $R_t > R_c$ so turbulent mixing maintains the homogeneity of the water column. A reduction in the wind stress and an increase in the heat flux over the next several days allow the water column to begin to stratify at mid-depth. $R_c$ is exceeded so turbulent mixing of heat to the bottom is prevented. Thus, heating is confined only to the surface rapidly strengthening the vertical temperature gradient. The thermocline shallows as the wind stress magnitude reaches a minimum on day 119. A well-defined thermocline is apparent over the inner shelf on day 120.

3.4 The role of cross-shelf temperature advection

A feature that is consistently apparent in the temperature fields throughout the record following day 90 is a cold pool located at mid-shelf on the bottom. This cold pool has an effect on the depth-averaged temperature that is discrepant to what one would expect from (1). It is therefore necessary to elucidate the origin and persistence of this cold pool. The cold pool is formed during the week centered on day 91 (Figure 9). The previous two weeks of stabilizing heat flux has stratified the shelf water so it behaves much like a two-layer system. A five-day period of southerly winds (Figure 5) causes downwelling resulting in warm offshore surface waters flowing toward the coast and offshore flow of cold coastal water below the thermocline. This cold pool is advected far enough offshore that it has no contact with the surface during subsequent mixing events. It becomes essentially insulated due to the limited heat flux across the thermocline so it remains a persistent feature, and is still evident as a coherent structure on day 150 (Figure 6f). The quantitative effect of this event on the thermal structure of the shelf water, and
when cross-shelf temperature advection becomes important, is examined later in the next section.

3.5 Analysis of the T-equation

It was earlier suggested that (1) might be useful in describing the change in horizontal stratification on a sloping continental shelf during the spring. When the assumptions associated with (1) are violated, as is the case when horizontal advection exists, it is no longer obvious that the simple model provides an adequate explanation to the problem. A quantification of the physical processes affecting the heating of the shelf water aids in demonstrating if the one-dimensional dynamics modeled by (1) can adequately explain the erosion of the horizontal thermal gradient, and whether or not other processes may be important.

The change in depth-averaged temperature during a time interval \( t_0 \) to \( t_1 \) is found by depth-averaging the model temperature equation (8) subject to boundary conditions (13) and (14) and integrating in time yielding

\[
\Delta \overline{T} = \frac{1}{\rho_0 c_p} \int_{t_0}^{t_1} \left[ \int_{h}^{H} \frac{\partial T}{\partial x} \, dz \right] dt - \int_{t_0}^{t_1} \left( \int_{0}^{h} \left( \frac{w}{\partial z} \right) \, dz \right) dt + \int_{t_0}^{t_1} \left( \frac{\partial}{\partial x} \left( A_H \frac{\partial T}{\partial x} \right) \right) dt + \int_{t_0}^{t_1} \left( \frac{T(\zeta) - \overline{T}}{h} \frac{\partial h}{\partial t} \right) dt
\]

(25)

where

\[
\overline{()} = \frac{1}{h} \int_{-h}^{h} () \, dz
\]

(26)

and \( h = H + \zeta \). Equation (1) appears once again in a time-integrated fashion by equating the LHS of (25) with term A. A is the result of depth-averaging the vertical mixing term.
in the temperature equation and applying the surface and bottom boundary conditions. \( B \) and \( C \) give the change in depth-averaged temperature due to horizontal and vertical advection of temperature. \( D \) is the change in the average temperature of a water column due to horizontal mixing. \( E \) owes its existence to variations in the free surface and has a negligible contribution to the temperature change of a water column, as will become evident.

Since equation (25) describes the change in depth-averaged temperature, it can be used to investigate the change in horizontal stratification at all depths when the limits of time integration correspond to well mixed states. Using the results from the WFS simulation, the calculation of each term of (25) illustrates the contribution of each physical process responsible for the change in horizontal stratification between periods of vertical mixing (Figure 10). Because the horizontal temperature gradient is reduced to nearly zero at day 132 (Figure 6e), the results are shown for limits of time integration corresponding to the well mixed states at day 58 and day 132, a 74-day integration.

It is clear that \( A \), the vertical mixing term, accounts for the majority of the temperature change. The horizontal and vertical advection terms \( B \) and \( C \) act in the same sense on the temperature change of the water column and are largely responsible for the remainder of the heating. The horizontal mixing term \( D \) has little effect \((O(0.5° C))\) on the temperature change. The residual temperature change from \( E \) is a maximum of 0.26° C, supporting \textit{a posteriori} the decision to consider this term insignificant. This calculation suggests that the simple model (1) can yield good physical understanding of the process responsible for the reduction of the wintertime horizontal temperature stratification by surface heating and vertical redistribution of the heat by mixing of the
water column, even in the presence of three-dimensional flow on the shelf (though there are no along-shelf variations in this model).

For the West Florid Shelf experiment, the balance $LHS = A$ describes the change in the depth-averaged horizontal stratification on the continental shelf to within a 31.0% error. The average agreement on the shelf is to within 11.4%. The strongest deviations are toward the inner shelf where the water column tends to heat faster than calculated by (1). This additional heating is due to horizontal and vertical temperature advection, and primarily occurred during a downwelling event in this experiment as discussed previously (Figures 9, 11).

Cross-shelf temperature advection becomes important when the shelf water is stratified and an alongshore wind induces cross-shelf Ekman transport resulting in coastal upwelling/downwelling. To see this, consider a two-dimensional cross-shelf model as before only represented by a two-layer system with upper layer temperature $T_1$, cross-shelf velocity $u_1$, thickness $h_1$, and correspondingly labeled variables for layer 2. The vertical temperature difference is $\delta T = T_1 - T_2$. The local change in depth-averaged temperature due to cross-shelf temperature advection in this two-layer system is

$$
\frac{\partial \bar{T}}{\partial t} = -\frac{1}{h_1 u_1 + h_2 u_2} \left[ h_1 \frac{\partial T_1}{\partial x} + h_2 \frac{\partial T_2}{\partial x} \right]
$$

By mass conservation, $h_1 u_1 + h_2 u_2 = 0$. The upper layer approximates the Ekman layer so $u_1$ can be thought of as the Ekman layer-averaged velocity cross-shelf component $u_E$, allowing one to write

$$
\frac{\partial \bar{T}}{\partial t} = -\frac{h_1}{H} u_E \frac{\partial \delta T}{\partial x}
$$

or, substituting for the Ekman transport $h_1 u_E$. 

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\[ \frac{\partial T}{\partial t} \sim -\frac{\tau'}{f \rho_0 H} \frac{\partial \delta T}{\partial x} \]

The cross-shelf gradient of \( \delta T \) results from an erosion of the top-to-bottom temperature difference near the coast due to upwelling/downwelling, or by intersecting top and bottom mixed layers. Compared to the typically dominating effect of surface heat flux, the cross-shelf advection of temperature contributes significantly to the local heating of a water column on the shelf when

\[ \left| \frac{\tau'}{f \rho_0 H} \frac{\partial \delta T}{\partial x} \right| \sim \left| \frac{Q(t)}{c_p \rho_0 H} \right| . \]

Using values representative of the period of downwelling centered on day 91 from the WFS experiment (\( \tau' = 0.04 \, \text{N/m} \), \( \frac{\partial \delta T}{\partial x} = 5 \times 10^{-5} \, ^\circ\text{C}/\text{m} \) at \( H = 20 \, \text{m} \), \( Q = 100 \, \text{W/m}^2 \)), both terms are of comparable size (\( O(10^{-6} \, ^\circ\text{C}/\text{s}) \)). Therefore one would expect cross-shelf temperature advection to be important, as has been shown (Figure 11) and discussed above. It is clear from (30) that the LHS will be small when the shelf water is well mixed (\( \delta T = 0 \)) so the surface heat flux will dominate. Thus the cross-shelf temperature advection term only becomes important when the ocean is stratified, and there is an alongshore component to the wind stress resulting in cross-shelf Ekman transport. This is consistent with observations of the North Carolina Shelf heat budget and a similar analysis presented by Austin [1999]. It should be noted, however, that an alongshore wind will also drive an alongshore current which would result in alongshore temperature advection in the presence of an alongshore temperature gradient, but this is not considered in the two-dimensional model.
4. Discussion

The process by which the wintertime horizontal temperature gradient is eroded is not entirely unlike the process by which it forms. In the winter, a negative surface heat flux causes convective mixing enhancing the wind mixing. Over the shelf, the mixed layer reaches the bottom resulting in a homogeneous water column. The negative surface heat flux is distributed uniformly with depth bringing about a greater temperature reduction in shallower water than in a deeper water column. Thus, the nearshore water cools faster than the deeper shelf water resulting in the characteristic wintertime horizontal thermal gradient.

Previously, it was thought that the light winds and solar heating in the spring allowed the thermocline to form and it was not well understood how the horizontal temperature gradient was relaxed or reversed. The fallacy of this explanation lies in the fact that light winds and solar heating are equated with springtime weather over the temperate latitude continental shelves. It is the case, rather, that the heat input to the surface is intermittently redistributed throughout the water column by vigorous wind mixing during the course of the spring season. It is this realization that suggests that the simple model (1) describing the heating rate of a homogeneous water column may apply to the problem of the evolution of the temperature field throughout the spring.

The process responsible for the formation of the offshore thermal gradient in the winter reverses in the spring, except that convective mixing is limited only to brief events. Strong winds associated with frontal passages are responsible for the periodic mixing of the water column. When appropriate conditions exist in the spring such that the ocean is being heated, but still occasionally vigorously mixed by wind forcing, this
mechanism breaks down the horizontal offshore temperature gradient on a sloping shelf. These meteorological conditions are often found over mid-latitude continental shelves. Late winter and springtime frontal passages are followed by periods of positive downward heat flux and light winds. This allows the upper part of the water column to heat rapidly. As a new front passes, the water column is promptly mixed to the bottom over the inner shelf redistributing the heat from the surface uniformly throughout the water column. The shallower inshore water is heated more rapidly than the deeper offshore water. After a series of these warming events and frontal passages, the offshore temperature gradient is relaxed or even reversed due to the faster heating of the shallower water. After the last strong wind event of the season, a thermocline can begin to form within a few days with a reduced or reversed offshore horizontal temperature gradient.

This paper has explored the likely dominant mechanisms responsible for the evolution of the temperature field on mid-latitude shelves during the spring season. The simulation used here, however, neglects the role of along-shelf temperature advection, the role of salinity, and the spatial differences of heat fluxes across the shelf. The latter can have a profound impact on the heating rate of the shelf water. There is physically a negative feedback in the heat flux to the ocean from the sea surface temperature. That is, there is an increase in the outgoing latent heat flux as the surface warms. This implies that during the early spring when the offshore water is warm relative to the cold inshore water, the net downward heat flux should be larger inshore than offshore. The result will be a faster relaxation of the horizontal thermal gradient than calculated with the spatially uniform fluxes. Therefore, this simplest time-dependent thermal forcing results in what
can be thought of as an upper bound for the time scale of the relaxation of the cross-shelf temperature difference.

It should be noted that not all mid-latitude shelves exhibit the spring transition from horizontally stratified wintertime conditions to vertically stratified summertime conditions as described here. As counterexamples, the Oregon and California continental shelves exhibit a spring transition from a vertically stratified winter regime to a horizontally stratified summer regime. This is because the temperature field is largely a result of summertime coastal upwelling and wintertime coastal downwelling [Gilbert et al., 1976; Lentz, 1987; Dever and Lentz, 1994]. That is, the cross-shelf advection of temperature is the dominant mechanism for the evolution of the temperature field throughout the spring.

5. Summary

The sloping bottom is responsible for the different heating rates across the shelf. The temperature change described by (1) is inversely proportional to the water column depth so that a shallower water column will heat more rapidly than a deeper water column subject to the same positive heat flux. This has been previously well understood but not applied to the problem of the relaxation of the wintertime horizontal thermal gradient. A realistic simulation of the WFS using the SZM forced by daily heat and momentum fluxes at the surface shows the applicability of the simple analytical model to the problem. The computation and comparison of each term of the depth-integrated time-averaged temperature equation quantifies the significance of the physical processes that can alter the temperature field. The simple model used to describe the heating rate of a
mixed water column subject to a surface heat flux appears as the dominant balance in the integrated temperature equation. Calculation of the depth-averaged temperature change from the surface heat flux using (1) agrees to the SZM model results for the WFS simulation to within 31% over the shelf. The spatially averaged error is only 11.4%. Cross-shelf advection becomes important when the ocean is stratified and forced by an alongshore wind causing upwelling or downwelling. The WFS experiment suggests this contribution is episodic and plays a secondary role during the spring season.

The spring transition of the thermal stratification over a wide sloping shelf in a temperate location is described by two processes. The first is the erosion of the horizontal thermal gradient as calculated by (1), and modified by temperature advection and only weakly by horizontal mixing. This occurs on a time scale of a few months when the water column is being heated and mixed periodically. The second process is the formation of the seasonal thermocline following the last strong mixing event of the spring season. This occurs on a time scale of less than a week under a stabilizing heat flux and light winds completing the spring transition from horizontal to vertical thermal stratification.
Acknowledgements

Sincere appreciation goes to Paul Martin for the development of, and assistance with, the Sigma Z-level Model. The comments of two anonymous reviewers proved most helpful for revision of the paper. Steven Morey received financial support for this work from the Department of Defense through the National Defense Science and Engineering Graduate Fellowship and the University Fellowship from FSU. The ONR Physical Oceanography Program provides a Secretary of Navy Grant to J. J. O’Brien as the base support for the Center for Ocean – Atmospheric Prediction Studies (COAPS).
Appendix

Sigma coordinates (SC), or terrain following coordinates, are often employed in ocean models over coastal and shelf regions. SC are defined such that a \( \sigma \) surface is everywhere a constant fraction of the total water depth. That is,

\[
\sigma = \frac{z - \zeta}{\zeta + H}
\]

where \( z = 0 \) at the resting surface and is positive upward, \( \zeta \) is the elevation of the free surface, and \( H \) is the undisturbed depth of the water column. Thus, \( \sigma \) varies from 0 at the free surface to -1 at the bottom. The use of a SC system is ideal over the shelf because it provides for increased vertical resolution in shallow water and prevents the need for exactly matching the model topography to fixed-depth grid cells. Near steep topography such as the shelf break, however, computing the horizontal pressure gradients in SC can produce large errors. The pressure gradient in SC is the sum of two terms. In the \( x \)-direction (neglecting deviations in the free surface),

\[
\left( \frac{\partial p}{\partial x} \right)_\sigma = \left( \frac{\partial p}{\partial x} \right) - \frac{\sigma}{H} \frac{\partial p}{\partial \sigma} \frac{\partial H}{\partial x}
\]

The first term on the RHS involves the gradient of pressure along a constant sigma surface and the second involves the gradient of the bottom topography. Near steep topography these two terms will be large, comparable in magnitude, and opposite in sign. Thus, a small error in computing either term can result in a large error in computing the pressure gradient term [Haney, 1991].

The horizontal pressure gradient can be accurately computed by using z-level, or fixed-depth, coordinates (ZC). ZC cannot, however, provide for increased vertical resolution in
shallow water. In addition, implementing a free surface requires special treatment of the surface layer. *Martin et al.* [1998] combined ZC and SC in the SZM developed at the Naval Research Laboratory. SC are used down to a given depth, and ZC extend below this fixed depth to the bottom. Using the hybrid vertical coordinate, the transformation from ZC to SC is given by

$$\sigma = \frac{z - \zeta}{\zeta + \min(H, z_s)}$$

(A3)

where $z_s$ is the specified depth below which ZC are used. Now, $\sigma$ varies from 0 at the free surface to -1 at the bottom interface of the lowest sigma layer. Each sigma layer is a fixed fraction of the total depth of the sigma layers. This hybrid vertical coordinate system allows for more accurate computation of the pressure gradient near a sharply sloping bottom, as well as for the benefits of SC on the shelf. For the WFS simulation, the vertical grid is chosen to use SC to the depth of the shelf break (70 m), and ZC below.
References


Florida Institute of Oceanography, Compilation and summation of historical and existing physical oceanographic data from the eastern Gulf of Mexico, Report to Bureau of Land Management, Contract #08550-CT4-16, St. Petersburg, FL, 97pp., 1975.


Table 1. Ocean model constants and parameters.

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
<th>Value</th>
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</thead>
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<tr>
<td>$\Delta x$</td>
<td>Horizontal grid spacing</td>
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<tr>
<td>$\Delta t$</td>
<td>Time step</td>
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<tr>
<td>$f$</td>
<td>Coriolis parameter</td>
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<td>$A_{M_0}$</td>
<td>Minimum horizontal eddy coefficient for momentum</td>
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<td>$z_0$</td>
<td>Bottom roughness</td>
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<tr>
<td>$c_{b_{\text{min}}}$</td>
<td>Minimum bottom drag coefficient</td>
<td>0.0025</td>
</tr>
</tbody>
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Depth-Averaged Change in Temperature from Year Day 58 to 132

- Model A
- Model B
- Model C
- Model D

Temperature (°C) vs. Distance from 10m isobath (km)