

Coupling Global and Regional Circulation Models in the Coastal Gulf of Alaska

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ABSTRACT

As part of the US GLOBEC NE Pacific program, we are simulating currents in the Coastal Gulf of Alaska (CGOA) to explore sources of interannual and interdecadal variability. To do so, we have developed a coupled modeling system composed of linked regional and global circulation models. The regional model, configured with 13-22 km resolution in the CGOA, is forced at the surface by observed heat fluxes and wind stresses, at the continental boundaries by observed runoff, and at the open ocean boundaries by a combination of tracer climatologies and subtidal velocity and tidal elevation provided by a global finite element model. In this communication, we describe the coupled system, including its present method of inter-model coupling, describe a series of multi-year model hindcasts, compare hindcast results with Eulerian and Lagrangian field data obtained in the CGOA in fall 1996, and assess the impact of global information (barotropic subtidal velocities and tidal elevations) on the regional model under the present coupling strategy. We find that the regional model produces appropriate current systems (Alaskan Stream, Alaska Coastal Current) and scalar fields, but with mesoscale variability (of SSH and velocities) at somewhat reduced strength relative to data, and with temperature gradients somewhat larger than those observed. Barotropic subtidal information from the global model penetrates the regional model interior, supplying additional mesoscale variability, and modifying regional velocity and scalar fields in both shallow and deep areas. Tidal information exerts a significant influence on subtidal scalar and velocity structure only in specific shallow areas, where the tides (and tidal mixing) are strongest. Pending the exploration of alternate coupling schemes, we infer from these results that on a time scale of months, purely barotropic information from outside the CGOA will have a modest impact on its mean regional circulation,

but a potentially stronger impact on the statistics and details of mesoscale eddies.

Keywords: Coastal Oceanography, Dynamical Oceanography, Fishery Oceanography, USA, Gulf of Alaska, 47°N-61°N, 160°W-132°W.

1. Background

1.1. Goals of this project

A core hypothesis of the U.S. GLOBEC Northeast Pacific program is that interannual to interdecadal variability in the circulation and hydrography of the Gulf of Alaska drives changes in productivity of zooplankton in the coastal zone, with consequent effects on the feeding success of salmonids and other species in the Gulf (U.S. GLOBEC, 1996). To help establish these linkages in the CGOA, we have been developing with our biological colleagues a set of linked circulation models, coupled with a lower trophic level Nutrient-Phytoplankton-Zooplankton (NPZ) biological model and an individual based model (IBM) of salmon. Specific issues to be addressed by this suite of models include the relative importance of surface Ekman flux, flows through submarine canyons, and mesoscale eddies on cross shelf exchange, and the subsequent impacts of that exchange on plankton and fish (both through resupply of nutrients and transport of the organisms themselves).

As part of this larger effort for GLOBEC, we have thus far developed both global and regional circulation models, bathymetry and forcing datasets for each, and a method for passing both tidal and subtidal information from the global to the regional model. Initial runs have yielded prominent spatial, seasonal, and interannual differences. Here, our goals are: a) to describe the coupled global/regional physical modeling system as it exists presently; b) to assess the primary features of the physical circulation generated by those models for a specific time period and compare them with available data; and c) to assess the impact of remote barotropic information (tidal elevation and depth-averaged velocities) on the regional model behavior. In light of these

results, we consider possible improvements to the present coupling strategy.

1.2 Overview of circulation in the GOA

The Gulf of Alaska contains two major current systems: the Alaskan Current (AC) - Alaskan Stream (AS) and the Alaska Coastal Current (ACC) (Fig. 1). The AS is the intensified northern boundary of the AC; both are part of the subarctic gyre forced by cyclonic winds in the northeastern Pacific. Reed and Schumacher (1986) summarized knowledge of the AC-AS and ACC. The AS is constrained by a steep continental rise which parallels the Aleutian Island chain. It is generally steady on seasonal time scales, but varies interannually (Reed, 1984; Musgrave et al., 1992; Lagerloef, 1995, Bhaskaran et al., 1993; Matthews et al., 1992; Strub and James, 2001a,b). The ACC is driven by a widely distributed coastal source of freshwater and downwelling favorable winds (Royer, 1981; Schumacher et al., 1990). Continuity of this current in the northern and western Gulf has been established (Stabeno et al., 1995). Significant bifurcation of the ACC occurs at several locations along the coast, with offshore-flowing branches joining the AS. The ACC is stronger in the western GOA (Reed et al., 1981); presumably this is due both to the stronger local downwelling and a larger volume of accumulated runoff. It has been suggested that the baroclinic structure of the eastern ACC is too weak to support baroclinic instability (Swaters and Mysak, 1985), which is prevalent in parts of the western Gulf (Mysak et al., 1981).

The intermitently formed Sitka eddy, centered off Sitka, Alaska (Tabata, 1982), and meanders of the Alaskan Stream in the central and western Gulf (Musgrave et al., 1992; Reed and Stabeno, 1993; Stabeno and Reed, 1989; Thomson and Gower, 1998) are prominent mesoscale features

with scales of ~ 200 km. These large eddies typically drift at $< 2 \text{ cm s}^{-1}$, are predominantly anticyclonic, and are more commonly observed in spring (Crawford and Whitney, 1999).

Modeling studies have suggested that such eddies near the shelf break are intensified by ENSO warm events (Melsom et al. 1999). The Sitka eddy has been reported to be topographically generated (Swaters and Mysak, 1985), and to have average surface currents of 0.15 m s^{-1} , with a maximum of 1.10 m s^{-1} as measured by drifters.

2. Methods

2.1. Model structure

To be useful in studying cross-shelf transport, a regional model of the CGOA must sufficiently resolve baroclinic instabilities of the flow. Vertical resolution must be sufficient to allow both the decoupling of flow from topography under stratified conditions and the development of appropriate shears when the flow is baroclinically unstable. The CGOA is a downwelling area with regions of significant tidal mixing, so the model must resolve surface and bottom boundary layers, which can meet in the shallow regions. As a method of investigating interannual variations, the regional model should be informed at its boundaries with circulation and scalar fields appropriate to specific days and years. Here we briefly describe our global and regional models, which begin to achieve these intricate modeling objectives, and detail how a global model can be used to constrain the regional model simulations.

The global model. The large-scale context for our regional studies is provided by simulations

with the Spectral Element Ocean Model (SEOM; Haidvogel and Beckmann, 1999; Iskandarani et al., 2001). SEOM has been developed for the purpose of high-resolution, basin-scale modeling on unstructured global grids (Iskandarani et al., 1994). Two separate global-scale implementations of SEOM were developed to generate tidal and subtidal information for use by the regional model. In each case, the finest resolution of the horizontal grid was focussed on the North Pacific (Fig. 2). For the 2-D tidal runs, the governing equations are the shallow water equations. For subtidal runs, the governing equations are the 3-D, Reynolds-averaged, Navier Stokes equations with Boussinesq and hydrostatic assumptions. The 3-D model was implemented in layered form, with a total of five isopycnal layers. Further details of the structure, forcing, and performance of the 3-D global model are provided in Appendix A.

The regional model. For regional circulation in the CGOA, we employed the S Coordinate Rutgers University model (SCRUM) of Song and Haidvogel (1994). This free surface, primitive equation model uses curvilinear orthogonal coordinates in the horizontal, and a stretched, bottom following "s coordinate" allowing flexible spacing of vertical grid points. The latter feature is especially useful in resolving boundary layers near the surface (important for wind mixing) and near the bottom (important for tidal mixing). Initial experiments comparing a curvilinear coast following coordinate system to the present rectilinear system established the latter as the more economical choice for the highly curved CGOA coastline. Ultimately, we implemented the model on a rectilinear grid oriented at 38° to true north (Fig. 3), with bathymetry derived from ETOPO5 and finer-scale bathymetric data (Fig. 4). In SCRUM, land areas are "masked out" after the calculation of each timestep, and thus entail computational overhead. Our rotated grid is designed to efficiently cover coastal and basin areas of the GOA,

while minimizing coverage of land areas to enhance computational efficiency. Further details of the structure, forcing and performance of the regional model are provided in Appendix B. Here, we shall focus on the method employed for coupling global and regional models.

2.2 Model Coupling

Perfect open boundary conditions - ones which simultaneously allow incoming external information to enter unimpeded and internally generated signals to exit without distortion - are difficult to approximate and almost certainly impossible to fully achieve. Many variants on open boundary condition techniques have been explored, although most have considered either tidal (e.g. Flather 1976) or sub-tidal (e.g. Marchesiello et al. 2001) variability separately. In contrast, in our CGOA regional model, we have the need to simultaneously impose both tidal and sub-tidal information from the external ocean. One sensible approach is to avoid open boundaries entirely through the use of a telescoped grid and a simple closed box (see Fig. 3). This method of incorporating the global model information is in some ways an extension of the boundary condition technique of Hermann and Stabeno (1996), in which a telescoped grid and closed box were utilized for a regional model with a rigid lid. The rigid lid approximation allowed a simple specification of the total barotropic flux through the coastal domain (representing the broad influence of large-scale patterns) without constraining the small-scale details of the flow.

In the present case of our free surface model, a different technique is required. Within the telescoped box, adjacent to the finely resolved area of interest, are placed *nudging bands* where currents and tracers are nudged towards desired subtidal values, but not so strongly as to prevent the escape of (e.g. substantially reflect) any outgoing, internally generated mesoscale signals.

The revised governing equation for a nudged variable A is then:

$$D/Dt \{A[x,y,x,t]\} = (\text{dynamical terms}) + \alpha[x,y]*(A_{\text{global}} - A)$$

Here A is the nudged variable, A_{global} is the exterior value (e.g., from the global model), “dynamical terms” represents all the model physics (advection, diffusion, acceleration, etc.) and α is the nudging constant. We note that such nudging is, in fact, a very primitive form of data assimilation, and that it has been used as a component in other open boundary condition techniques (e.g. Marchesiello et al., 2001; Palma and Matano, 1998). Note also that α varies in space, having nonzero values in specified bands and being set to zero elsewhere. In the present set of models, the nudged variables are as follows: T and S (three dimensional fields from monthly climatologies), depth-averaged U and V (two-dimensional fields from the global subtidal layered model runs), and free-surface height (two-dimensional fields from the global tidal runs). The regional model treats the external (depth-averaged) mode as a separate variable, which greatly simplifies the nudging process. Although three-dimensional nudging of U and V is certainly possible, there are many unresolved issues related to the transfer of layered model information onto a level model. Here, for simplicity, we have focussed on the effectiveness of two-dimensional information, as a means of conveying tidal and subtidal information; a fuller exploration of three-dimensional layer-level coupling schemes is anticipated for future models of this region.

The external boundary values A_{global} may be based on data, the results of a large scale model simulation, or some combination of the two. For subtidal flows, the area between the *subtidal*

nudging band and the solid wall is intended to function as a free recirculation zone, satisfying continuity and possibly absorbing any mesoscale features which escape from the interior.

Relevant information from the subtidal model will include internal wave modes, in particular coastal trapped waves (CTWs) which propagate with the coast to their right. How strongly the regional model values should be nudged toward their counterparts from the global model is directly dependent on the time it takes a CTW characteristic to traverse the nudging band. For a nudging band which is 1000 km wide, and a wave speed of 1000 km d⁻¹, the time of transit is ~1 d. Hence, for CTWs, a nudging constant (α) of ~1 d⁻¹ is a reasonable choice for significant nudging of the regional model values toward their global model counterparts.

This nudging constant is also a reasonable choice for substantial damping of mesoscale signals which are leaving the finely resolved region – for example, it ensures that any CTW signal which is not present in the global model result will be significantly attenuated before entering the free recirculation region beyond the subtidal nudging band. Such outward-propagating mesoscale features, which escape the damping effects of the subtidal nudging band (or are created by the nudging to global model values), will be attenuated further by their transit across the telescoped region, through absorption into the larger-scale flows resolved by the coarser grid. Whatever signal is left may reflect off or propagate along the solid wall, and potentially encounter the subtidal nudging band a second time, where further attenuation (or reflection back out into the recirculation zone) will take place.

Our use of a fixed wall in a wind-forced model will inevitably lead to spurious (that is, not observed in the real ocean) pressure gradients and currents near that wall. While a portion of

these spurious gradients will be due to the “escaped” signals from the interior, another portion will be due to interaction of the solid walls with local wind forcing (which is applied over the entire model domain). Again, the subtidal nudging band acts as a barrier to shield the interior from these spurious signals, and a significantly large damping constant (such as the 1 d^{-1} value suggested above) will provide much of the needed attenuation.

Tidal information from the global model represents traveling waves, and a *tidal nudging band* requires nudging constants large enough to transmit this global signal to the regional model. In water that is 4000 m deep (the maximum depth of the regional model), the free wave speed of shallow water waves is approximately 200 m s^{-1} . Hence, for a tidal nudging band 1000 km wide, the time of transit for tidal characteristics crossing the band is $\sim 1\text{h}$, and a nudging constant of $\sim 1 \text{ hr}^{-1}$ is appropriate.

In order to accommodate *both* tidal and subtidal dynamics in the present model, we employ the two independent (though spatially overlapping) nudging bands simultaneously – that is, two separate areas where the regional model assimilates tidal and subtidal information, respectively, from the larger-scale model and/or from data. Tidal elevations (here, the five most energetic tidal components) are assimilated in the tidal nudging band near the western and southern walls of the (telescoped) regional model domain. Temperature, salinity, and barotropic velocity fields corresponding to desired subtidal inflows are assimilated along the subtidal nudging band further within the domain, but still external to the finely resolved CGOA (see Fig. 3).

Ideally, the tidal signals assimilated in the outer domain will be transmitted through areas of

subtidal nudging without attenuation. In practice this transmission will depend on the value of the subtidal nudging constants for scalars and velocity, the width of the subtidal nudging band, the frequencies of the tidal signals, and the phase speeds of the tidal signals. As an estimate of attenuation, consider the one-dimensional shallow-water wave equations with nudging of tidal height and subtidal velocity:

$$\eta_t = -H u_x + \alpha^T (\eta^T - \eta) \quad (1)$$

$$u_t = -g \eta_x + \alpha^S (u^S - u) \quad (2)$$

where η is the free surface height above mean fluid depth H , u is the horizontal velocity, η^T and u^S are the tidal height and subtidal velocity to be assimilated (from a large scale model and/or data), and α^T and α^S are the tidal and subtidal nudging constants, respectively. We seek approximate solutions for tidal motions within the subtidal nudging band. In a region where $\alpha^T = 0$ and where both u^S and α^S vary slowly in space, equations (1) and (2) yield:

$$\eta_{tt} - c^2 \eta_{xx} + \alpha^S \eta_t = 0 \quad (3)$$

A characteristic equation is obtained by looking for solutions of the form:

$$\eta = \eta_0 \exp[i(kx - \omega t)] \quad (4)$$

where ω may have an imaginary component. Substitution into (4) yields

$$\omega = \{-i\alpha^S \pm [-(\alpha^S)^2 + 4c^2k^2]^{1/2}\}/2 \quad (5)$$

If the nudging coefficient α^S is sufficiently small and the tidal frequency ω^T ($= kc$, with units of radians s^{-1}) is significantly large, then

$$4c^2k^2 \gg (\alpha^S)^2 \quad (6)$$

Under these circumstances, the solution for an incident tidal signal in the subtidal nudging band can be approximated as

$$\eta = \eta_0 \exp[i(kx - \omega^T t)] \exp[-(\alpha^S/2) t] \quad (7)$$

Now, integrating a tidal characteristic $(kx - \omega^T t)$ across a subtidal nudging band of width dx , the fractional attenuation of tidal amplitude due to passage through that band is

$$F = 1 - \exp[-(\alpha^S/2) dx/c] \quad (8)$$

For typical values in our model ($dx = 1000$ m, $c = 200$ m s^{-1} and $\alpha^S = 10^{-4}$ s^{-1}), F is negligible. Indeed, even if subtidal nudging occupied the *entire* finely resolved region ($dx = 4.4 \times 10^5$ m), the value of F would be less than 0.1. Hence, barring substantial reflection of the tidal signal at the edge of the subtidal nudging band, we will have clear transmission of the tidal signal into

(and back out of) the model interior.

To obtain the A_{global} values for tracers, daily T and S climatologies for each model gridpoint were interpolated from the monthly values of Levitus (Levitus et al., 1994), while daily mean subtidal velocities were obtained from the results of the global layered model. The Levitus values are themselves derived from a comprehensive set of S and T observations, averaged in time and objectively interpolated in space onto a regular, 1 degree global grid. Insufficient data exist to apply more spatially refined values, specific to a particular year, along our subtidal nudging band.

For tidal nudging, we first ran the regional model without tides (that is, with only subtidal barotropic velocity and T, S climatology nudging) and stored daily averages of sea surface height at each gridpoint. These data were added to tidal heights calculated from spatially dependent amplitudes and phases derived from the 2-D tidal run of SEOM, which included M2, S2, N2, K1, and O1 tidal components. Note that without adding the daily average height estimates from the subtidal run, the boundary of the regional model in the tidal-plus-subtidal run would be strongly nudged towards a mean zero elevation (that is, the purely tidal signal), which could potentially distort the subtidal dynamics of the interior.

For the subtidal nudging band, we apply a maximum nudging constant of $\alpha_{\text{TS}} = 0.1 \text{ d}^{-1}$ for climatological T and S and $\alpha_{\text{UV}} = 1.0 \text{ d}^{-1}$ for subtidal velocities. For the tidal nudging band, we apply the substantially larger maximum value of $\alpha_{\text{tidal}} = 1.2 \times 10^2 \text{ d}^{-1}$ for surface heights. The values of α_{TS} and α_{UV} are ramped in space, from zero at the outer solid

wall of the regional model, to their maximum values near the boundary of the telescoped and non-telescoped regions 10 gridpoints (~856 km) into the interior. Between gridpoints 10 and 17 (~120 km), they are ramped back to zero. The values of α_{tidal} are ramped as well, from their maximum value at the outer solid wall to a value of zero 17 gridpoints (~1000 km) into the interior. This ramping is designed to minimize reflection and encourage absorption of any internally generated signals.

2.3 Float Tracking

The paths of numerical drifters can be compared with drogued drifter data. For this purpose, stored daily averaged velocities from the regional model (lowpass-filtered to remove tidal and near-inertial signals) are interpolated to 5 m and 40 m, and used to generate pseudo-Lagrangian float tracks using a 4th-order Runge-Kutta routine.

3. Model experiments

We ran several model experiments to test features of the coupling scheme and performance of the regional and global models. The experiments conducted are as follows:

- 1) *Global model run*. The global 3-D model is spun up with 5 years of a repeating cycle of daily NCEP winds, spanning the period of NSCAT wind availability (Aug 1996 - July 1997). Resulting depth integrated velocities are low pass filtered and stored as daily values for use by the regional model.

2) *Regional model runs.* Here we utilize the monthly climatologies (Levitus) for T and S, daily barotropic velocities from the global model, and tidal elevations for the global model, as fields to be applied to the nudging bands in the regional model. Wind, heat flux, and coastal runoff are applied as forcing fields (see Appendix B for details), and the model is run for the period August - December 1996, when applicable SEOM results are available. The results are compared with observational data for this period.

2a) *Climatological tracers and subtidal velocity nudging.* In the first nudging experiment, the regional model is spun up via nudging during July 27-Aug 1 with daily barotropic velocities (from SEOM), and monthly T,S fields (from Levitus climatology). For this spinup period only, nudging coefficient are applied which are uniform in the finely resolved domain (1.0 d^{-1} for U,V; 0.1 d^{-1} for T,S), and linearly decrease to zero at the outer edges of the (telescoped) domain. This pattern of assimilation allows the interior to assume the SEOM velocity values, without forcing any flows through the solid walls. Subsequent to the five day spinup, the model is nudged only in the telescoped domain, with nudging coefficient values as described in section 2.2 (linearly decreasing to zero at the edge of the finely resolved domain and at the outer walls of the telescoped area). The initial spinup period provides the essential broad patterns of the flow field, including some larger eddies. Subsequently the regional model is free to develop finer scale circulation in its interior, under the influence of local wind and buoyancy forcing, and with the SEOM barotropic velocities and climatological T and S applied as horizontal boundary conditions.

2b) *Climatological tracers, subtidal velocities, and tidal elevation nudging.* After the subtidally nudged run is complete, an equivalent tidal-plus-subtidal run is performed, which has boundary conditions partially determined by the sea surface elevations of the first run as described in section 2.2. Results from this run are used for the model-data comparisons reported in section 4.2.

2c) *Climatological tracer nudging only.* The model is spun up as in case 2a). Subsequently, however, we remove all influence of the SEOM velocities, and assimilate only the climatological T,S as a horizontal boundary condition.

By contrasting these three runs, we attempt to clarify the influence of global barotropic tidal and subtidal information on the interior regional solution.

4. Results

4.1. Global model results

Figures 5 and 6 display barotropic velocities and free surface height fields from the global model interpolated to a regular latitude longitude grid for early June and early November, respectively. Coastal and shelf break currents are evident, as are meanders and eddies in the deep basin. A significant portion of the shelf break current in the northwestern GOA (the Alaskan Stream) appears to come from the deep basin in these simulations, whereas currents upstream (the Alaska Current) are weak. The Alaskan Stream is wider in the model output than is typically observed, due to the limited spatial resolution and smoothed bathymetry (see Appendix A). In early November, westward flow of the Alaska Coastal Current is spatially continuous around the

GOA. However, results for early June exhibit a significant reversal of that near coastal flow in the eastern GOA. This pattern of seasonal variation in the coastal currents in the eastern GOA has been noted by Royer (1998) and is suggested by the analyses of seasonal altimetric patterns of Strub (2001a).

4.2 Regional model results

Here we describe results from the 1996 regional run 2b, which included all three types of nudging (climatological S and T, subtidal barotropic velocity, tidal elevation). We begin with an overview of the major features generated by the model, then proceed to a more detailed comparison of model output with data in section 4.2.3.

4.2.1 Basin-scale circulation and SST

Regional model results reproduce many of the major observed features in the CGOA (Figs. 7-8). The AC-AS system and ACC are clearly evident; surface speeds achieve realistic values of 0.4 m s⁻¹ and greater. Buoyancy forcing is typically strong in the fall, contributing to the strength and continuity of the ACC in model results. Compared to surface velocities, barotropic velocities are substantially weaker, and are frequently in different directions (compare Figs 7 and 8). Note, however, that barotropic velocities at the shelf break are generally stronger than their equivalent in the global model run. A tongue of warm water penetrates west along the shelf break in model results, as has been noted in SST images and in hydrographic data (see comparisons with data in section 4.3).

4.2.2 Mesoscale circulation and eddies

Extensive mesoscale circulation features are observed in model output on the continental shelf, at the shelf break, and in the deep basin. Many of these simulated barotropic features are locked to prominent seamounts in the deep basin, and to submarine canyons at the shelf break.

Animated results (not shown) exhibit a clockwise propagation of disturbances around the seamounts. Eddy activity at the 200 km scale is most prominent in the vicinity of Sitka, AK. On the shelf, small eddies with 50 - 100 km diameter are produced, but appear less dynamic than has been observed in drogued drifter studies (Stabeno and Hermann, 1996). Eddy generation via baroclinic instability (abetted by a temporary weakening of the downwelling favorable winds) is a possible source of these smaller, nearshore eddies.

4.2.3 Comparisons with data

Here we compare model results for the fall of 1996 to available data of the following types: 1) current meter data and temperature time series from the ACC; 2) drogued drifter tracks from the continental shelf; 3) hydrographic sections corresponding to the ACC and the AS; 4) sea surface temperatures from radiometers; 5) sea surface heights from a coastal tide gauge; and 6) sea surface height perturbations from altimeters. We note areas of agreement, and suggest (here and in section 5) reasons for observed differences. For this comparison we use the model run 2b, with both tidal and subtidal forcing; subsequently (in section 4.4) we describe the sensitivity of the results to the inclusion of tidal and barotropic subtidal forcing.

4.2.3.1. SST. Model SST for October 26, 1996 has been compared with a nine-day composite SST, centered on that date, derived from the NOAA/NASA Oceans Pathfinder archives (Fig. 9; T. Strub and C. James, personal communication). This date was chosen based on the availability

of a clear image. The model reproduces broad features observed in the Pathfinder imagery: a tongue of warm water from the south, penetrating around the CGOA in the direction of the predominant, cyclonic circulation. SST signatures of 200 km-scale meanders and eddies are apparent in both model and data; similar features have been reported by Thompson and Gower (1998). The spatial mean SST produced by the model is lower than that of the data, and spatial gradients tend to be larger than the observations. This lower-than-observed spatial mean can be attributed in part to the Levitus climatology used for initialization and boundary nudging, as the Levitus values for October-November were in fact lower than the observed Pathfinder values for late October 1996.

Several small scale features of the data are captured by the model, as follows:

Structures near Kodiak Island. In both model and data, a near-coastal tongue of warm water, centered along the shelf break, bifurcates just east of Kodiak Island. In the model, shallow (<150 m) areas surrounding Kodiak Island are as much as 4 ° C colder than deeper areas; Cook Inlet is especially cold. The data exhibit the same pattern of cold waters around Kodiak Island, but with only half the amplitude of the model result.

Structures near the Sitka eddy. In the model, relatively warm waters from the south are observed penetrating to the north, along the coastline, past Sitka, AK. Just north of Sitka, a tongue of ~2 ° C colder water streams out from the coast to the south and west. Warmer water swirls anticyclonically around the edge of this colder tongue; we identify this anticyclonic circulation as the “Sitka eddy” (Tabata, 1982) of our model. In the data, a tongue of warm water

perpendicular to the coast is observed penetrating out from approximately the same location as the cold tongue of the model. In both model and data, this tongue appears to represent coastal water. In the model, SST is $\sim 3^\circ\text{C}$ colder at the coast than in deeper waters; the data also exhibit cooler waters at the coast, but the temperature difference is only ~ 1 degree C.

4.2.3.2. Hydrographic sections. During fall 1996, hydrographic sections were measured across the ACC (at the southwest exit of Shelikof Strait) (Fig. 10) and the AS (at the shelf break, southwest of Kodiak Island) (Fig. 11). In both cases, the model exhibits similar structures in temperature and salinity fields at locations and times corresponding to the hydrographic sections. Across the ACC section, both model and data exhibit highest temperatures at the coastline and near the ocean surface. The model exhibits slightly warmer surface temperatures than the data. In the model, strongest vertical gradients are near the surface; in the data, there is a thermocline at 50-100m depth. Both model and data exhibit freshest waters at the surface in the middle of the strait, rather than at the coast, with slightly stronger vertical gradients in the model. Across the AS section, both model and data exhibit a thermocline at approximately 50 m depth; the data exhibit a stronger vertical gradient. In the model, SST is slightly warmer to the north; in the data, it is slightly warmer to the south. Both model and data exhibit fresher waters near the shelf break, relative to values further offshore. The weaker vertical gradients in the model are partially the result of limited vertical resolution, especially in the deeper areas (see Appendix B).

A snapshot of model output for early October, 1996, at a location corresponding to the GAK line (Fig. 12), reveals several features typical of fall observations from hydrographic data collected between 1997 and 2000 (T. Weingartner, personal communication). A warm pool of water (10°

C) spans the shelf, with strongest thermocline at ~50 m depth. A low salinity core (< 30 psu) is evident near the coast, and a secondary low-salinity core is evident near the shelf-break. All of these features are present in the hydrographic data. A subsurface maximum in T, similar to that produced by the model, appears in early December data from the GAK line (T. Weingartner, personal communication).

4.2.3.3. Time series of subsurface temperature. A time series of temperature for the period August 10 - September 25 was obtained from a mooring in the center of Shelikof Strait (Fig. 13). The model exhibits a gradual warming trend, with warmest temperatures and strongest gradients near the surface. The data exhibit a similar warming trend, but with more temporal variability; most likely this is due to the passage of 20 km scale eddies not resolved by the model (see comments below). As in the October 26 SST comparison, the model results here are generally colder (by ~ 1 degree C) than the data.

4.2.3.4. Current meter data. Current meter data from the same mooring are representative of fall conditions in the ACC (Fig. 14). Surface flows are to the southwest, with a reversal to northeastward flow at depth. As in the temperature data, fluctuations at periods of 3-5 days are common; for the velocity record, these result from both fluctuations in wind stress (due to the passage of storms) and the passage of mesoscale eddies. The former should be captured by the present model; the latter is presumably underestimated by the model due to its limited horizontal resolution. (A model with finer horizontal resolution does capture more of this eddy variability; see Hermann and Stabeno, 1996; Stabeno and Hermann, 1996). While the data exhibit greater

variance and stronger amplitudes than the model over the period August 3 - September 1996, specific events, such as strong downstrait flow events in September, appear in both model and data.

There are multiple potential reasons for discrepancies between the modeled and observed ACC, especially at daily time scales. First, winds in Shelikof Strait are frequently ageostrophic (flowing down the pressured gradient between two mountain ranges on the Alaska mainland and Kodiak Island). This fact can lead to strong differences between measured winds and those calculated from large-scale atmospheric analyses (Stabeno et al., 1995a). The NCEP winds presently used for this model do not include such fine-scale ageostrophic effects; therefore much of the direct downstrait wind forcing is lost. Second, the true bathymetry is rather severe (steep slopes) and the model depths are necessarily smoothed, because of stability issues. Third, the coarse resolution of the grid, together with its implied viscosity and the proximity of the no-slip boundaries, may artificially retard the flow.

Despite these limitations, a comparison of time-mean vertical profiles, averaged over August 3 - September 26, 1996 (Fig. 15), reveals similar vertical shears in both model and data, with a reversal from downstrait (southwestward) to upstrait (northeastward) velocity at ~160 m in each case. As in Fig. 14, the data exhibit stronger mean downstrait (that is, negative alongshelf) flow than the model.

4.2.3.5. Drogued drifter tracks. Model floats were released at locations corresponding to drogued drifter data on DOY 238 (August 26), 1996 (Fig 16). In the data, the drifter furthest

offshore did not lie in the Alaskan Stream, whereas the corresponding model float was vigorously advected by the model Stream. Conversely, one of the mid-shelf drifters was ultimately entrained into the Stream, whereas the corresponding model float was not. Such differences are expected, as the precise location of the stream does vary due to eddy activity. Note, however, that the ultimate speeds of real versus model floats entrained in the stream are very similar. A second mid-shelf drifter was observed to drift towards the coast over the 60-day period; the model float executes a similar onshore path after encountering an eddy, but achieves landfall further to the southwest. A nearshore drifter followed a sinuous path through the Shumagin Islands; the corresponding model float remained close to its release point. Again, a divergence of actual versus model paths is expected due to the variable nature of the circulation and the inability of this model (or any other, absent extensive data assimilation) to reproduce all the spatial details.

4.2.3.6. Sea surface heights. Model SSH has been compared with TOPEX/POSEIDON/ERS-2 analyses for October 8, 1996 (Fig. 17), obtained from the web site maintained by the Colorado Center for Astrodynamic Research (CCAR) at the University of Colorado (R. Leben, 2001, personal communication). These results must be viewed with caution, as the TOPEX analysis is designed primarily to capture mesoscale *time variability* of SSH about its climatological mean, and not its absolute magnitude. Since we do not have a suitable climatology from the model to subtract off the total signal, we have plotted the TOPEX result from CCAR, including their estimate of dynamic height (obtained via climatological hydrographic data), referenced to 1500 m. Similar spatial scales (~200 km) are apparent for much of the mesoscale energy in both model and data. Both model and data exhibit the same broad depression associated with the Subarctic

Gyre. The TOPEX image exhibits many closed eddies in the interior; the model exhibits fewer such closed features, but many strong meanders. Overall, there is less mesoscale energy in the model output than in the data. As noted earlier, the largest mesoscale feature in the model is a large, anticyclonic eddy near Sitka, AK. A smaller feature is present in the data, between 56 and 57 ° N at ~138 ° W; it is unclear whether or not this corresponds to any Sitka eddy feature actually present. Indeed, there may be particular difficulties in capturing a stationary eddy feature with mesoscale TOPEX data, as much of the eddy signature would end up in the climatological background. There is some evidence for offshore migration of the Sitka eddy, however, from model studies (e.g. Melsom et al., 1999), and the smaller feature in the data may be the result of that migration. It is presently unknown whether the larger eddy in the data, at 54.5 ° N, 137 degrees W, itself corresponds to an offshore migration of the Sitka eddy, or an eddy formed at some other location.

4.2.3.7. Time series of sea surface height. Raw surface elevation data from a tide gauge at Sitka, Alaska were filtered with successive applications of simple 24-hour and 25-hour moving averages, and corrected for atmospheric pressure using data from a nearby weather station, to reveal the subtidal signal. This series was compared with the corresponding free surface elevation data from model runs, both with and without tidal and subtidal nudging from the global model (Fig. 18). As with the TOPEX spatial comparison above, the amplitude of the model time series was typically weaker than the amplitude of the data, but the timing of specific events was nonetheless well-reproduced, especially the events of September 10-11, 17-18, and 25-26. Note that the event of September 17-18, in particular, appears in both model and data for the velocity

series from Shelikof Strait (see Fig. 14).

4.3 Sensitivity of the regional model to remote barotropic influence.

Although subtidal velocity information provided by the global model can be expected to influence the regional model most strongly near the subtidal nudging band, dynamics including coastal trapped waves (CTWs), Rossby waves, simple advection, and diffusion can be expected to carry boundary information into the interior, as well. Indeed, one strong motivation for using global model results (or real data) at the boundary of the regional model is to provide just such information about remotely forced CTWs to the regional model interior.

A Hovmuller (distance versus time) plot of the difference in coastal sea level between experiments 2a and 2c (the subtidally “forced” – i.e. nudged - and “unforced” regional runs) (Fig 19) illustrates the propagation of 10-day period signals from the southern nudging band into the interior of the model as CTWs, with a phase speed of $\sim 1000 \text{ km d}^{-1}$. This value is somewhat higher than phase speeds calculated for CTWs off Washington and Oregon ($\sim 500 \text{ km d}^{-1}$, e.g. Battisti and Hickey, 1984), and also higher than phase speeds calculated for coastal propagation in SEOM itself ($\sim 100\text{-}200 \text{ km d}^{-1}$). Given the higher latitude of our region, and its steep, irregular topography, a closer match with lower latitude data is not expected. A different phase speed in the regional model, relative to the global model, is also expected, insofar as the global model uses a quite different bathymetry than the regional model.

To more explicitly reveal the influence of the barotropic nudging, snapshots of SST, surface velocity, SSH and depth-averaged velocity are presented for the “unforced” run (run 2c), and the run with subtidal forcing only (run 2a) (Fig 20). Also presented are difference maps representing the effects of subtidal forcing (“subtidal only” run minus “unforced” run) and the effects of tidal forcing (“subtidal and tidal” run minus “subtidal only” run) (Fig. 21). Note that climatological T and S are used as a boundary condition in all these runs; hence the differences reflect only the addition of our particular forms of barotropic information.

First, note how the subtidally forced runs exhibit higher sea levels in the interior (and hence lower sea levels at the coast, relative to the interior), compared to the unforced run. Both surface and barotropic velocity differences indicate that information from the global model has slightly weakened the coastal flows, yielding an underestimate (by 20%) of near-surface flows at the location of current meter mooring in Shelikof Strait. This may be due to the fact that in coastal areas, barotropic velocities from the global model (Figs. 5 and 6) are generally weaker than those produced by the regional model (Fig. 8), due in large part to the deeper bathymetry used by the global model.

Second, note that the addition of the subtidal barotropic information has significantly modified both surface and barotropic velocity patterns, and the SST, by altering the positions of interior eddies. In many locations these differences are as large as 0.2 m s^{-1} and 1 degree C, respectively.

Finally, note how the subtidal signals have in fact *added* mesoscale variability to the interior. Mesoscale eddies and meanders in the SSH field are in fact more numerous, and of greater

amplitude, after addition of the subtidal boundary information (compare Figs. 17a and b). In particular, note the two closed eddy features in the northern GOA ($57\text{-}59^{\circ}\text{N}$, $144\text{-}146^{\circ}\text{W}$), produced when subtidal forcing is added. The time variance in SSH was also calculated, for the period Sep 13 – Oct 13. Averaged over the areas shown in Fig 17, this approximate measure of Eddy Kinetic Energy was $\sim 28\%$ higher in the subtidally-forced case, as compared with the unforced case.

The addition of tides to the subtidally forced run does not make any substantial difference to the results in most areas. However, a significant impact is observed in shallow areas around Kodiak Island and Cook Inlet, where strongest tides have been observed. In those areas, SST is modified by ~ 1 degree C and surface currents are modified in geostrophic response to those changes in the scalar fields. The barotropic subtidal velocities are slightly altered, possibly representing a weak tidal residual.

Time series of SSH at the Sitka tide gauge are shown for these sensitivity experiments (Fig. 18). The model run with subtidal and tidal barotropic forcing produced a time series with slightly stronger amplitude than the run without either of those influences, in conformance with the SSH snapshot results. A run with subtidal, but no tidal, forcing produced a time series indistinguishable from that of the subtidal-plus-tidally forced run. Apparently, the addition of barotropic subtidal information adds variance to the time series of SSH (and improves the overall fit with observations), just as it appears to add spatial variance to the SSH pattern.

5. Discussion and Conclusions

No ocean model, absent extensive data assimilation, is capable of replicating all the details of the true ocean, even at the mesoscale. Here, we have nudged information as a boundary condition over limited areas, using global model results (as opposed to measured information). Hence, we can at best hope to replicate time-mean features of the flow, and aspire to capture some, but not all, of the time variability at any location (that is, “statistically correct” fields). With these caveats, our initial experiments with various configurations of two circulation models for the Gulf of Alaska suggest that this area can be realistically modeled using a regional scale model.

Our regional model with ~13-20 km horizontal resolution and 20 vertical z coordinate levels captures much of the spatial and seasonal pattern of CGOA circulation. Both the AC-AS and ACC current systems are easily identified in model output. The model also captures the dominant 200 km scale variability observed in the Gulf. It displays especially strong mesoscale dynamics in the vicinity of the Sitka eddy. Many observed SST patterns are reproduced, with gradients somewhat larger than observed. Additional runs (not shown here) have reproduced the seasonal weakening/strengthening of the coastal current in the CGOA over a full model year. Interannual differences have also been evident between different years simulated; for instance, spring 1997 has generally higher SST than spring 1995, as was observed in the CGOA at Shelikof Strait (J. Schumacher, personal communication). This is important because it implies that using this model, we will be able to investigate effects of the physical regime shift that has been documented in the GOA system.

We have also found that both tidal and subtidal barotropic information from a global model can be effectively incorporated into the regional model as a boundary condition, through the use of suitably designed nudging bands located outside the region of interest. Spurious effects of rigid wall dynamics are avoided by allowing free compensatory flow in the telescoped region outside the subtidal nudging band. A proper choice of nudging coefficients minimizes the interference between the applied tidal and subtidal signals.

Model results are affected, in a specific and limited manner, by the inclusion of barotropic boundary information from the global model. We find the regional model, with or without barotropic boundary forcing, generates appropriate current systems (AS, ACC) and scalar fields, but with less spatial (mesoscale) variability of subtidal SSH than is observed, and with less temporal (e.g. daily-to-weekly) variability of subtidal currents and SSH than is observed. Barotropic subtidal information from the global model penetrates the regional model interior, enhancing the spatial and temporal variability in most (but not all) areas, and modifying regional velocities and scalar fields (both surface and subsurface) in both shallow and deep areas. Barotropic tidal information exerts a significant influence on subtidal scalar and velocity structure only in specific shallow areas, where the tides (and tidal mixing) are strongest. Pending the exploration of alternate coupling schemes, we infer from these results that on a time scale of months, purely barotropic information from outside the CGOA will have a modest impact on its mean regional circulation, but a potentially stronger impact on the statistics and details of mesoscale features.

These results are an important step towards our goal of exploring the effect that interannual and

decadal changes in GOA circulation have on lower trophic level dynamics and fishery populations. However, the true coastal current (that is, the ACC) has a strong baroclinic structure, and the use of only barotropic information may be inadequate to convey remote influences to this structure. Further, the global model bathymetry was typically deeper than the corresponding bathymetry in the regional model; for a strongly baroclinic flow, the depth-averaged velocity from the global model might provide an artificially weak signal for the regional model under our present coupling scheme, and so artificially dampen the locally generated baroclinic flows.

It is possible that a larger impact of global barotropic information would be observed with one or more of the following changes: 1.) The barotropic velocity used to nudge the regional model could be adjusted to account for differences in underlying bathymetry between global and regional models (i.e, we could nudge to the value of barotropic transport produced by the global model, rather than to the value of barotropic current). Note, however, that this may lead to overestimates of the remote influence on coastal areas, just as the present method may underestimate that influence; 2.) The nudging constant could be increased, or the nudging bands could be expanded to more fully spin up wave signals for propagation into the interior. Note, however, that the former approach may create a strong barrier to the free passage of internally generated signals out of the regional model domain; 3.) The tidal and subtidal nudging bands could be further separated, to guard against any possible interference between tidal nudging and subtidal nudging.

Future studies will explore these issues in more detail, as we gain more experience with both

barotropic and baroclinic coupling methods. We believe that our model results can be further improved relative to data by including the influence of baroclinic information from the global model. Specifically, we plan to generate multiyear simulations with the global layered model, and project information about the first baroclinic mode from those results onto the first baroclinic mode of the regional model in the subtidal nudging band. Intensification of current speeds to more closely match observed current magnitudes and spatial focus of the AS, AC and ACC will also require even higher horizontal resolution in the regional model and ageostrophic wind forcing. New regional model codes, which take advantage of parallel computer architectures, will allow higher resolutions to be implemented for our multiyear runs in GLOBEC. New studies to generate and compare fine-scale winds with NCEP products in the CGOA are also now underway in GLOBEC.

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APPENDIX A. The Global Model

A large scale context for our regional studies is provided by simulations with the Spectral Element Ocean Model (SEOM; Haidvogel and Beckmann, 1999; Iskandarani et al., 2001). SEOM has been developed for the purpose of high resolution basin scale modeling on unstructured global grids (Iskandarani et al., 1994). The governing equations are the 3 D, Reynolds averaged, Navier Stokes equations with Boussinesq and hydrostatic assumptions. Lateral subgridscale mixing of momentum is parameterized using the shear and mesh size dependent formulation of Smagorinsky (1963), which has proven to be highly effective on these horizontally heterogeneous grids. The resulting class of large scale circulation models has several significant virtues over those using more traditional approaches, including complete geometric flexibility, regionally selective horizontal resolution, and the ability to avoid open boundary conditions by use of global grid refinement.

The spectral element circulation model has now been applied in its reduced gravity form to a variety of test problems and global oceanic/atmospheric applications. When applied to a now standard suite of shallow water test problems on the sphere, SEOM is shown to be highly competitive with other numerical models, including those based on spherical harmonic methods (Taylor et al., 1997). Oceanic applications on global, non uniform grids show that these favorable properties are maintained in the presence of continental geometry and highly unstructured elemental meshes (Haidvogel et al., 1996).

Here, for economy, SEOM has been implemented on a global grid in layered form with a total of five isopycnal layers (Fig. 2). Vertical transfer of momentum is represented with weak (linear) interfacial drag and (nonlinear) stress laws. Following Hurlburt et al. (1996), outcropping of the layers is avoided by mass sharing between layers as a minimum "entrainment thickness", here taken to be 40 m, is reached. Table A1 gives the relevant parameters (resting layer thicknesses and reduced gravities) used in the global simulation. A significant limitation of the non-outcropping layered model is that topographic variations must be contained within the lowermost layer. We have done so here by clipping topography (obtained from ETOP05) at 200 m (minimum) and 5000 m (maximum), and then multiplying the resulting topographic excursions above 5000 meters by 0.85. Some of the effects of this topographic shrinkage are noted in the main body of the text.

As a first test of behavior, the global layered model has been forced with a repeating cycle of NCEP winds, corresponding to the period of NSCAT wind availability (August 1996 through July 1997). A comparable simulation using NSCAT winds has also been prepared, as a basis for a future sensitivity study. No explicit thermodynamic forcing is included; therefore, the resulting simulations can at most represent the wind driven component of the large scale, low frequency circulation.

SEOM is highly scalable on parallel computing platforms (Curchitser et al., 1998). The simulations reported in the text have been obtained on a Beowulf type cluster of Sun Ultra 5 workstations maintained at the Institute of Marine and Coastal Sciences, Rutgers University. On this system, a year's simulation requires approximately 6 CPU days when run on 12 processors.

APPENDIX B. The Regional Model

For regional circulation in the CGOA, we employed the S Coordinate Rutgers University model (SCRUM) of Song and Haidvogel (1994). We implemented the model on a rectilinear grid oriented at 38° to true north. The horizontal grid has telescoped edges to reduce the computational overhead associated with horizontal boundary conditions. The grid has 145 by 113 horizontal gridpoints with 17 telescoped gridpoints on the southern and western boundaries (Fig. 3). The finely resolved area of the model domain reaches from Queen Charlotte Island in the south to Unimak Pass in the west. The telescoped region continues to the south end of Vancouver Island, and to Amukta Pass in the west. Grid resolution varies from 13-22 km in the finely resolved area to 200 km near the western and southern walls. As described in section 2.2, the telescoped regions here serve primarily to recirculate flows into and out of the area of interest. A no-slip condition is applied at the solid walls.

The model has 20 vertical s-coordinate levels. We optimized the s-coordinate feature of SCRUM to achieve quasi uniform spacing near the surface. In the shallowest (50 m) areas, spacing is uniformly 2.5 m over the entire water column. In the deepest (4000 m) areas, spacing is 3.6 m - 10.0 m over the top 50 m of the water column, with closest spacing nearest the surface. This quasi-uniform spacing will facilitate efficient coupling with the biological models.

SCRUM is written in highly vectorized code, and most of the simulations were obtained on vector architectures (CRAY J932) at the Arctic Region Supercomputing Center. However, recent

simulations have been obtained on an equally fast, single-processor (Alpha XP-1000) workstation at PMEL. On the CRAY platform, simulation of one model year requires approximately 7 CPU days; on the local workstation, 5 CPU days are required.

B.1. Bathymetry

Model bathymetry was interpolated from a specially developed 5 minute bathymetric map of the CGOA, based on ETOPO5 and other sources. While ETOPO5 has the advantage of broad coverage, it is well known to be inaccurate in many coastal areas. More accurate bathymetric data was obtained from two different sources and used to improve ETOPO5. Nearshore data was derived from the National Ocean Service (NOS) Hydrographic Data Base, error checked and gridded to 30 seconds by National Geophysical Data Center (NGDC) and distributed as the TerrainBase data collection. These data are focused on specific coastal areas such as Cook Inlet. Offshore data was derived from Smith and Sandwell (1997), who collected and verified coastline and marine ship track data from many sources, and distributed that data as part of their 2 minute measured and estimated digital topographic map. Though their estimated bathymetry (based partly on gravity anomalies) contains too much noise to be useful on the continental shelf, the measured bathymetry was easily extracted and used to improve ETOPO5 values offshore.

Data from these two detailed sources were combined and interpolated to a 5 minute grid using Global Mapping Tools (GMT). To reduce computational effort, the interpolations were done for 10 deg. by 10 deg. areas that overlap by .25 deg. The interpolated grid points match those of ETOPO5 so that when they were combined, ETOPO5 seamlessly supplied data in areas where the detailed bathymetry data set was lacking. The final bathymetry is particularly accurate in

areas of high data resolution, such as along the shelf break. It constitutes a major improvement over ETOPO5, especially in shallow shelf areas such as the Trinity Banks southwest of Kodiak Island. After interpolation to the SCRUM grid, the bathymetry was cropped to 50 m (minimum) and 4000 m (maximum), and filtered with six passes of a Shapiro filter, to improve numerical stability of the model. The result is shown in Fig. 4. Even after filtering, the result is considerably more accurate than ETOPO5 alone.

B.2. Heat Flux and Wind Stress

Suitable values of wind and heat flux in specific years were obtained from the NCEP/NCAR Global Reanalysis Project. NCEP products include a global data set of atmospheric variables, obtained by combining a global spectral model with historical data. Their model has been run for the years since 1958, and the output is available online. Resolution of these data is roughly 2° . Temporal resolution is 6 hours, but we have chosen to use daily averages as input to our regional circulation model.

Daily average NCEP/NCAR values for latent and sensible heat net flux, and net longwave and shortwave radiation were summed to provide total heat flux from the ocean. Daily average U wind and V wind at 10 m height above the ocean surface were converted to wind stress using the simple formula:

$$\tau = \rho_a * C_d * U10 * |U10|$$

where τ is the wind stress in N m^{-2} , U10 is the vector of wind speed in m s^{-1} , ρ_a is the air

density and $C_d = .0012$.

B.3. Freshwater Input

Salinity controls the density field more strongly than temperature for much of the year in the CGOA; freshwater input is greatest in October and smallest in March (Royer, 1982), and is primarily from distributed sources. Monthly values of integrated freshwater runoff along segments of the coastline were derived by Royer (1982 and pers. comm.) from snowpack, precipitation and temperature data. The data represent runoff from areas seaward of the coastal mountain range. To supplement this estimate with freshwater contribution from inland areas, we utilized Copper and Suisitna River discharge data from the US Geological Survey (USGS). Because there are few complete years of data, a monthly climatology was computed for each river using all the available data. These large rivers supply a small amount of freshwater (~10% of the total), relative to the line sources (~90% of the total), but their peak discharge is earlier in the year.

B.4. Mixing

Vertical mixing is parameterized as a function of local shear and stability, using Mellor Yamada level 2.0 closure (Mellor and Yamada, 1974). Background diffusivity and viscosity were 10^{-4} and $10^{-3} \text{ m}^2 \text{ s}^{-1}$, respectively. Horizontal mixing of both scalars and momentum is calculated with a biharmonic operator, and scaled by the local grid spacing as described in Hermann and Stabeno (1996). Mixing is computed along geopotential surfaces, rather than along sigma surfaces (Haidvogel and Beckmann, 1999).

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Figure Captions

Fig. 1. Overview of circulation in the Gulf of Alaska.

Fig. 2. Layout of quadrilateral elements for the layered implementation of the Spectral Element Ocean Model (SEOM). Structure within each quadrilateral is represented with a polynomial basis set of order eight. The resulting average "grid spacing" is approximately 25 km around the periphery of the North Pacific Basin, and increases to about 100 km elsewhere.

Fig. 3. Layout of the telescoped rectilinear grid for our regional implementation of the S Coordinate Rutgers University Model (SCRUM). Regions of maximum tidal and subtidal nudging, respectively, are indicated.

Fig. 4. Smoothed bathymetry used for the regional model simulations. In this and subsequent regional model results figures, the axes are aligned with those of the model (that is, rotated 38° relative to true north), and units are model gridpoints in the two coordinate directions (λ and ϕ). Distance between successive gridpoints ranges from 13-22 km.

Fig 5. 3-D Global model SSH (contoured, m) and barotropic velocity (m s^{-1}) for DOY 155 (June).

Fig 6. 3-D Global model SSH (contoured, m) and barotropic velocity (m s^{-1}) for DOY 304

(November).

Fig 7. Regional model surface velocities (m s^{-1}) and temperature ($^{\circ}\text{C}$, shaded) for October 8 (DOY 282), 1996, with boundary nudging of subtidal and tidal information from the global model.

Fig. 8. As in Fig. 7, for barotropic velocities (m s^{-1}) and sea surface height (contoured, m).

Fig. 9. Comparison of model SST (upper panel) with observed SST (lower panel) from Pathfinder weekly composite ($^{\circ}\text{C}$), for October 26, 1996.

Fig. 10. Modeled (upper) and measured (lower) T (left) and S (right) sections for the Alaska Coastal Current at the southwest end of Shelikof Strait. In each plot, the Alaskan Peninsula is to the left, and Kodiak Island is to the right. Bottom is marked with a heavy line. Model gridpoints and CTD station locations are marked with asterisks.

Fig. 11. Modeled (upper) and measured (lower) T (left) and S (right) sections for the Alaskan Stream near the Shumagin Islands. Maximum water depth is 3000 m. Model gridpoints and CTD station sections are marked with asterisks.

Fig. 12. Modeled sections of T (upper left), S (upper right), and alongshore velocity (lower left) at the GAK line. Bottom is marked with heavy line. Model bathymetry in the GAK region is also shown (shaded and contoured, m) with Kenai Peninsula (black) and approximate location of

GAK line (heavy line).

Fig. 13. Time-depth comparison of modeled (upper) and measured (lower) temperature at the center of Shelikof Strait, for August and September, 1996.

Fig. 14. As Figure 13 for alongshelf velocities. Negative velocities are to the southwest.

Fig. 15 Time averaged velocity profiles for the data shown in Fig. 14.

Fig. 16. Comparison of modeled (top) and measured (bottom) drogued drifter tracks at 40 meters depth near the Shumagin Islands. Stars mark track beginning and triangles mark track end.

Fig. 17. Comparison of model SSH with observed SSH from TOPEX/ERS, for October 28, 1996. TOPEX estimate includes climatological dynamic height, referenced to 1500 m. a) model results with barotropic (tidal and subtidal) boundary forcing; b) model results without barotropic boundary forcing; c) TOPEX data.

Fig. 18 Subtidal sea surface height at a tide gauge at Sitka, AK (solid line) (corrected for local atmospheric pressure), compared with corresponding signals from model runs with (dashed line) and without (dotted line) nudging of global velocity data. In each case, the time mean for this interval has been subtracted from the time series.

Fig. 19. Hovmuller diagram (distance versus time) for the difference in sea surface height (m)

produced by boundary nudging to global model subtidal velocities.

Fig. 20. Results as in Figs 7-8 for model runs with no tidal or subtidal barotropic nudging (upper) and subtidal barotropic nudging only (lower). SST ($^{\circ}$ C) and surface velocity (m s^{-1}) are on left, SSH (m) and barotropic velocity (m s^{-1}) are on right.

Fig. 21. Difference in velocity and temperature fields between: model with subtidal barotropic nudging versus model with no subtidal barotropic nudging (upper); model with subtidal and tidal barotropic nudging versus model with subtidal nudging only (lower). Layout and variables as in Fig. 20. Note different length scale for vectors, relative to Fig. 20.

Table A1. Mean interface depths (m) and reduced gravity at interfaces (m s^{-2}) for the layered implementation of the Spectral Element Ocean Model.

Interface depth, m	Gravity at interface m s^{-2}
0.	9.81
135.	0.020
320.	0.009
550.	0.004
800.	0.002

Figure 1

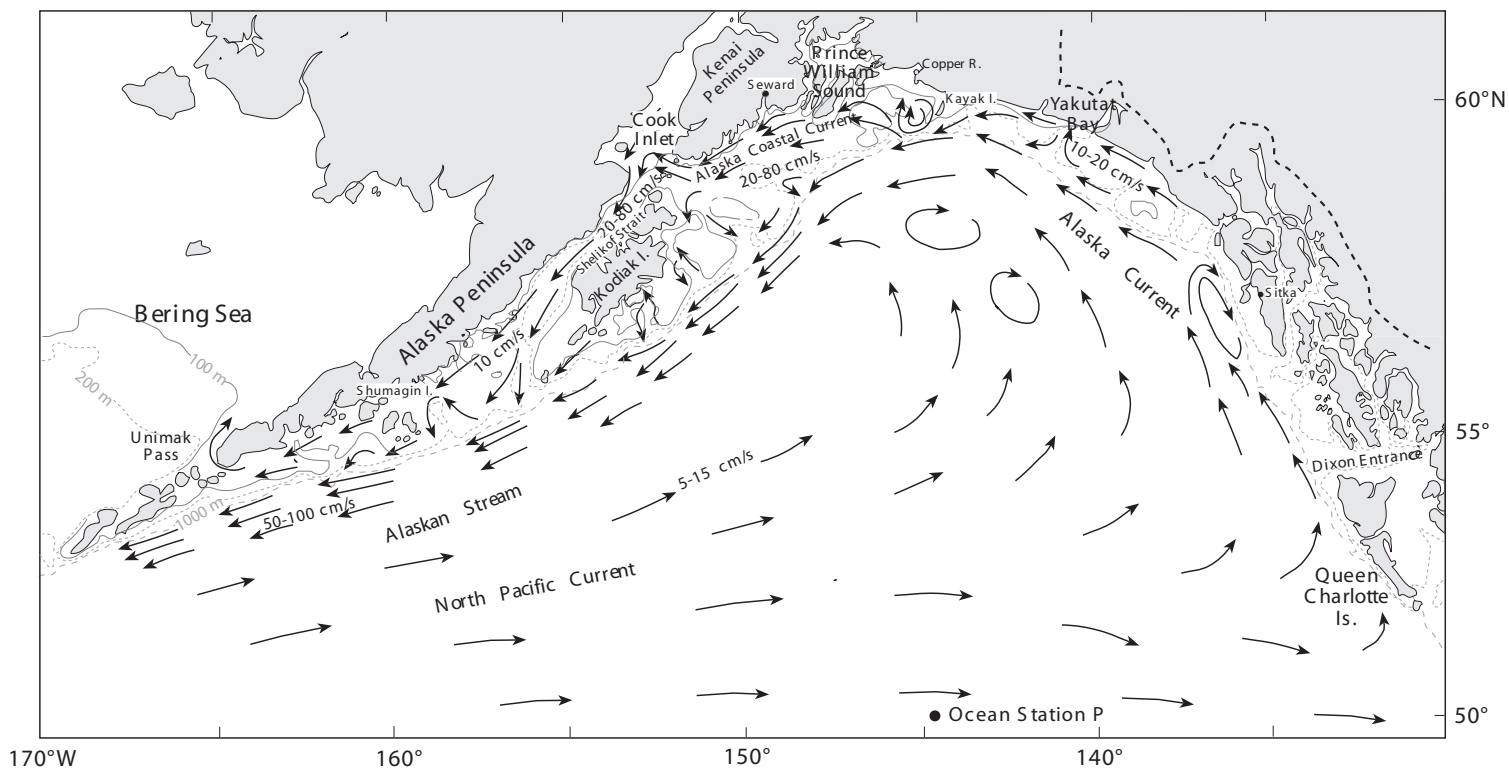


Fig. 1. Overview of circulation in the Gulf of Alaska.

Figure 2

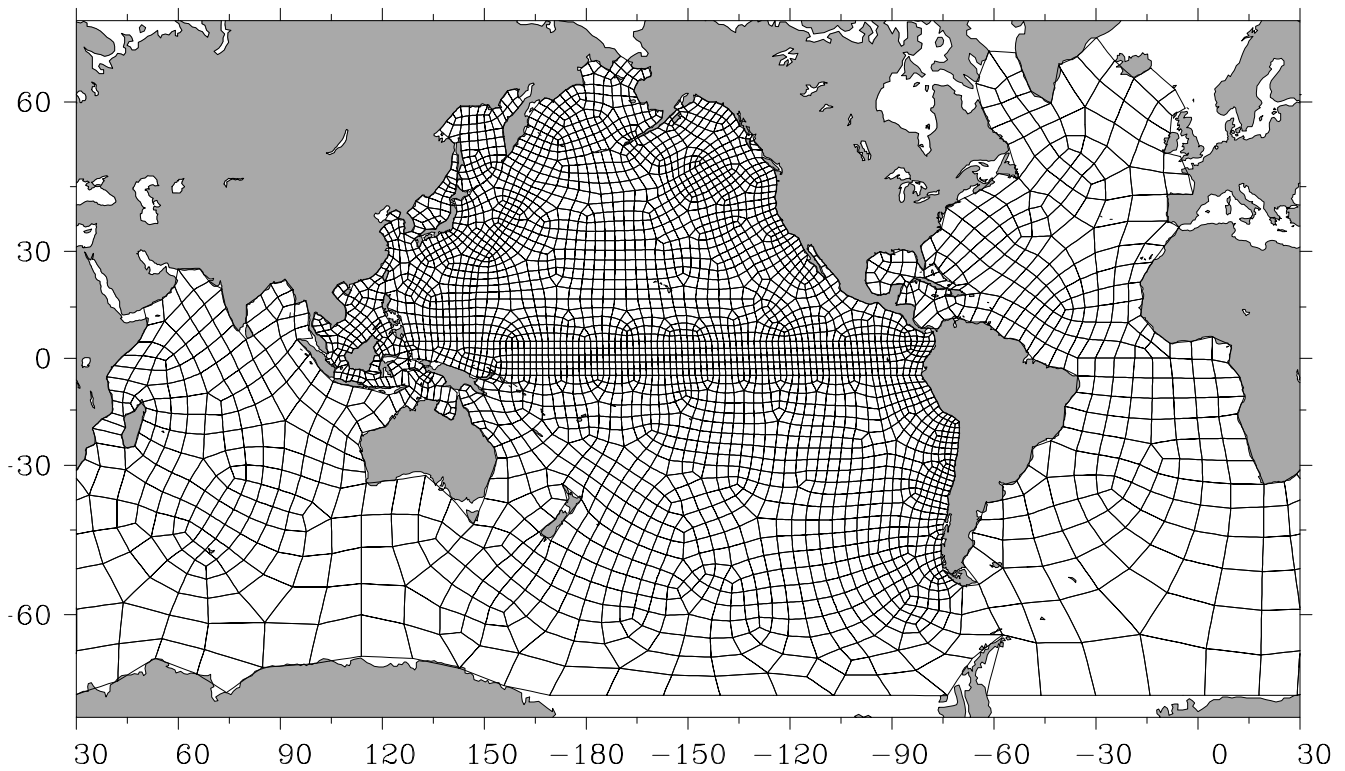


Fig. 2. Layout of quadrilateral elements for our layered implementation of the Spectral Element Ocean Model (SEOM). Structure within each quadrilateral is represented with a polynomial basis set of order eight. The resulting average "grid spacing" is approximately 25 km around the periphery of the North Pacific Basin, and increases to about 100 km elsewhere.

Figure 3

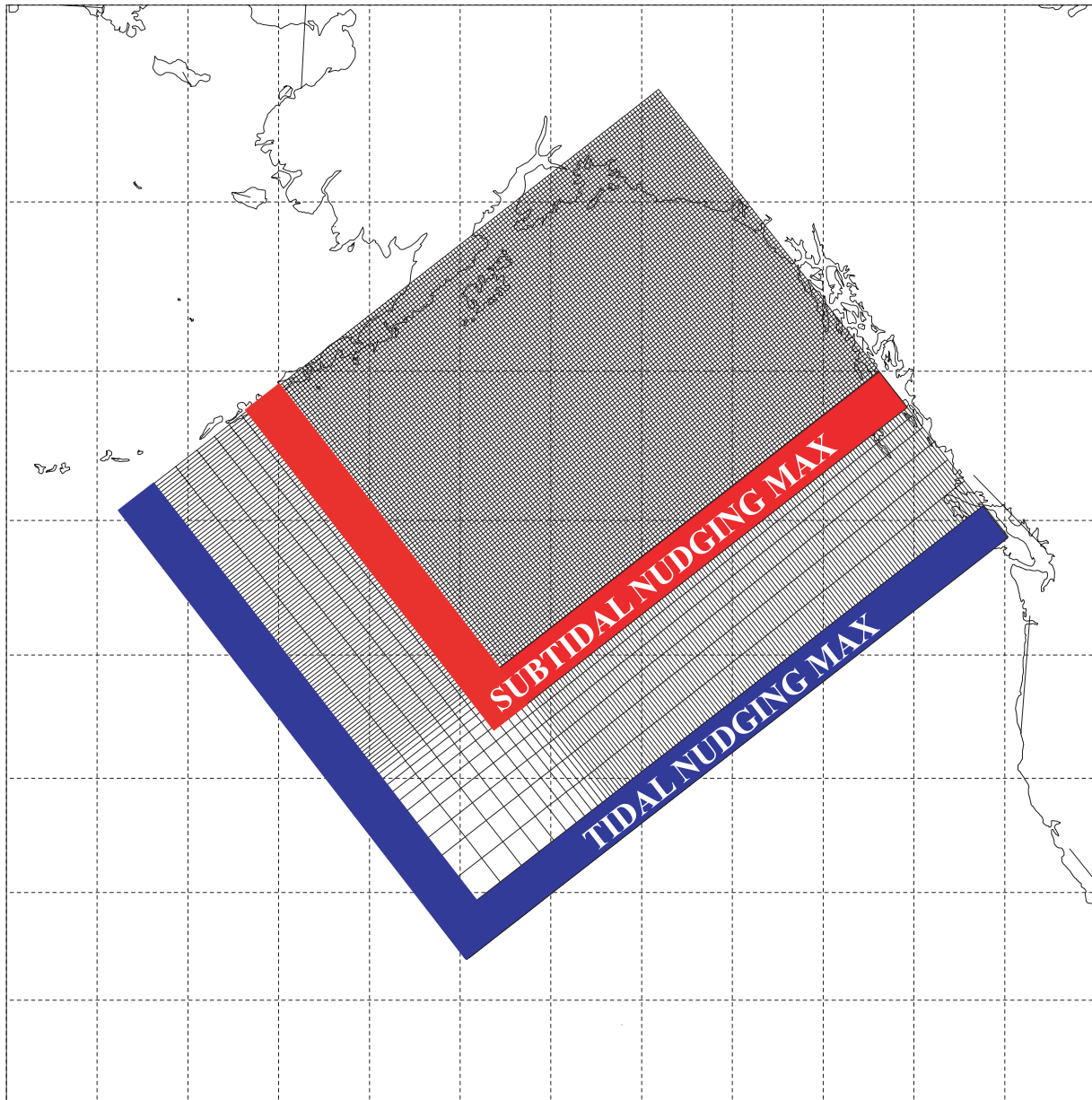


Figure 4

Bathymetry at Model Gridpoints

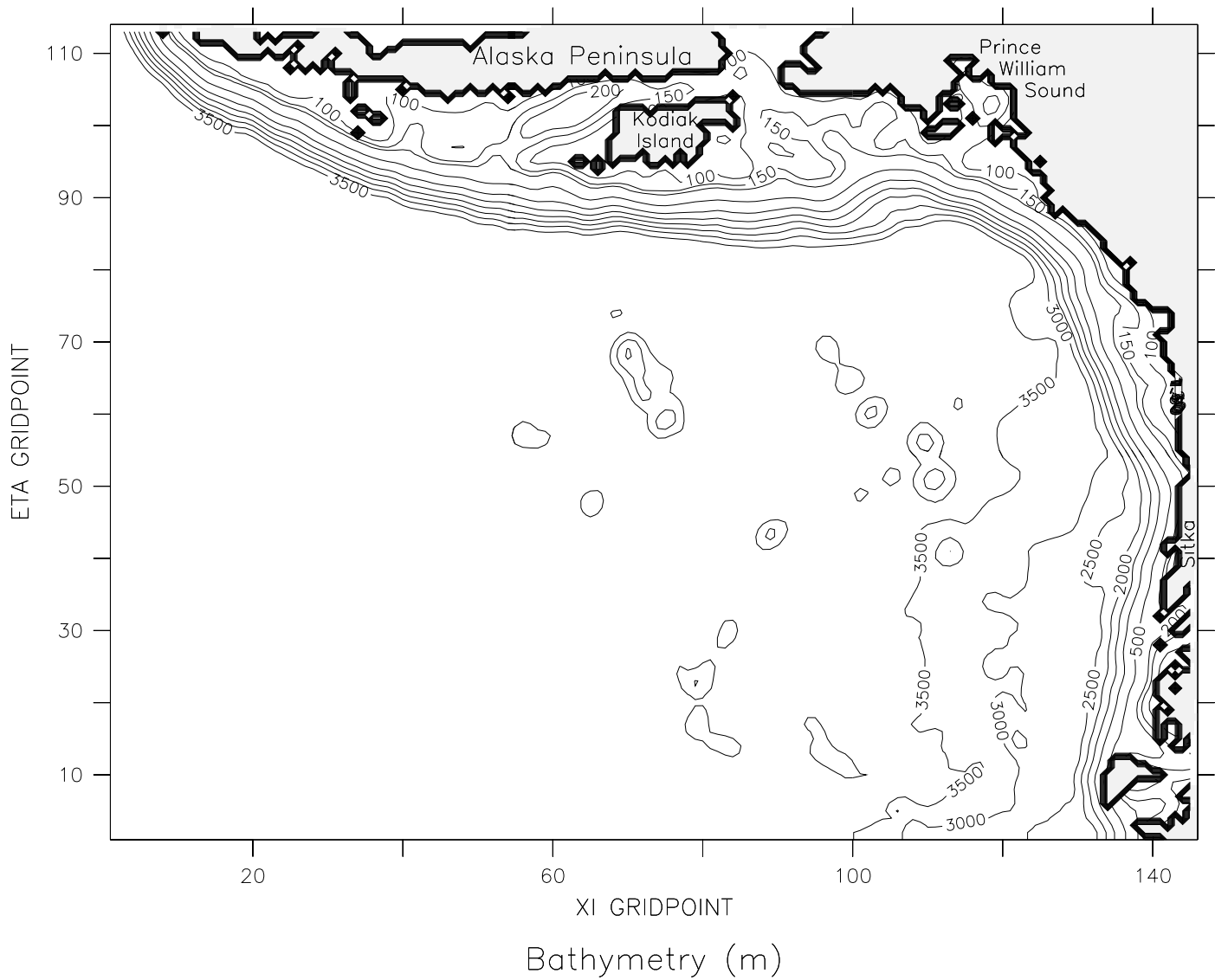


Figure 5

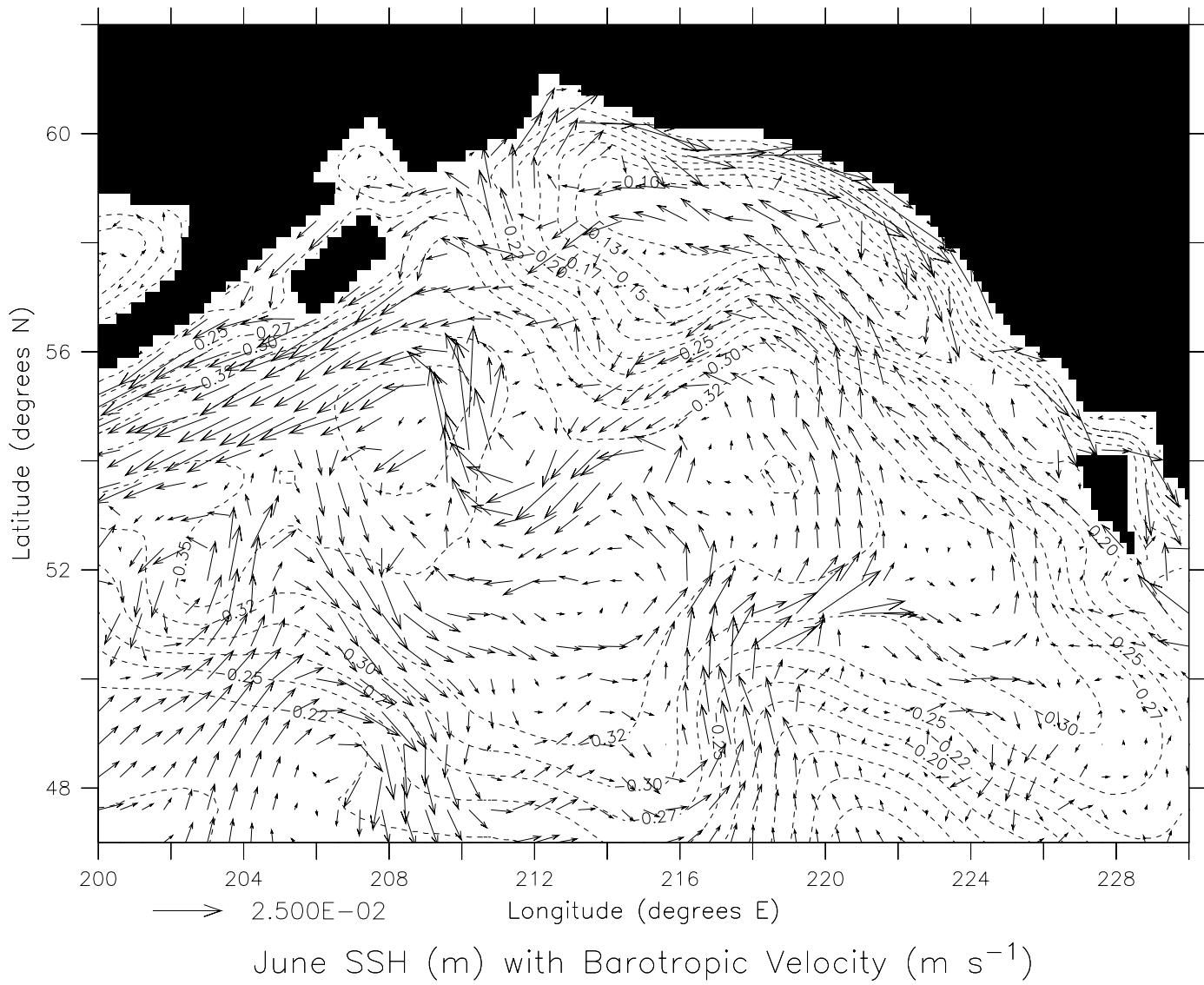
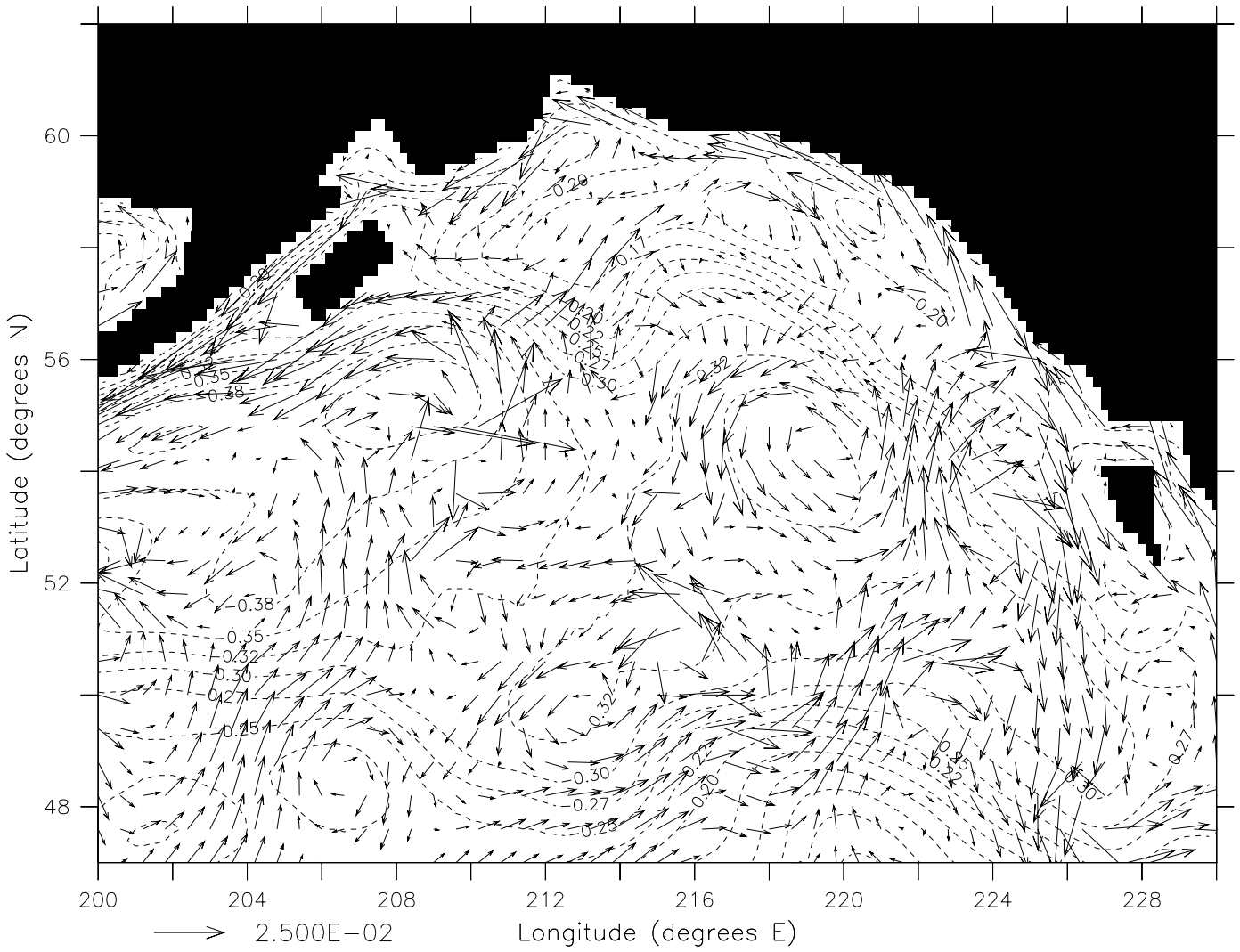


Figure 6



November SSH (m) with Barotropic Velocity (m s^{-1})

Fig 7

Coastal Gulf of Alaska, 1996

SCRUM TIME : 282.54 Modified Julian Days

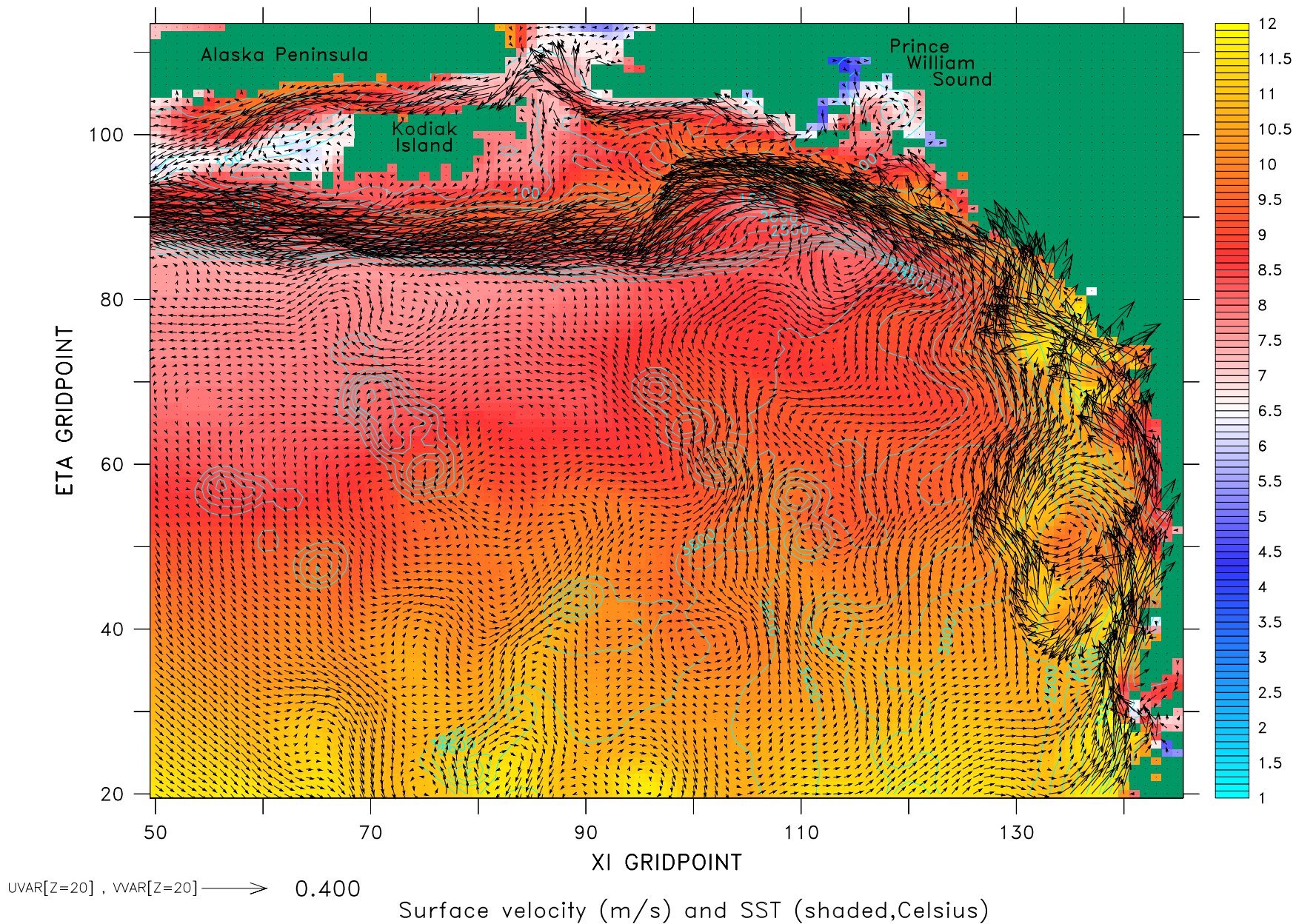
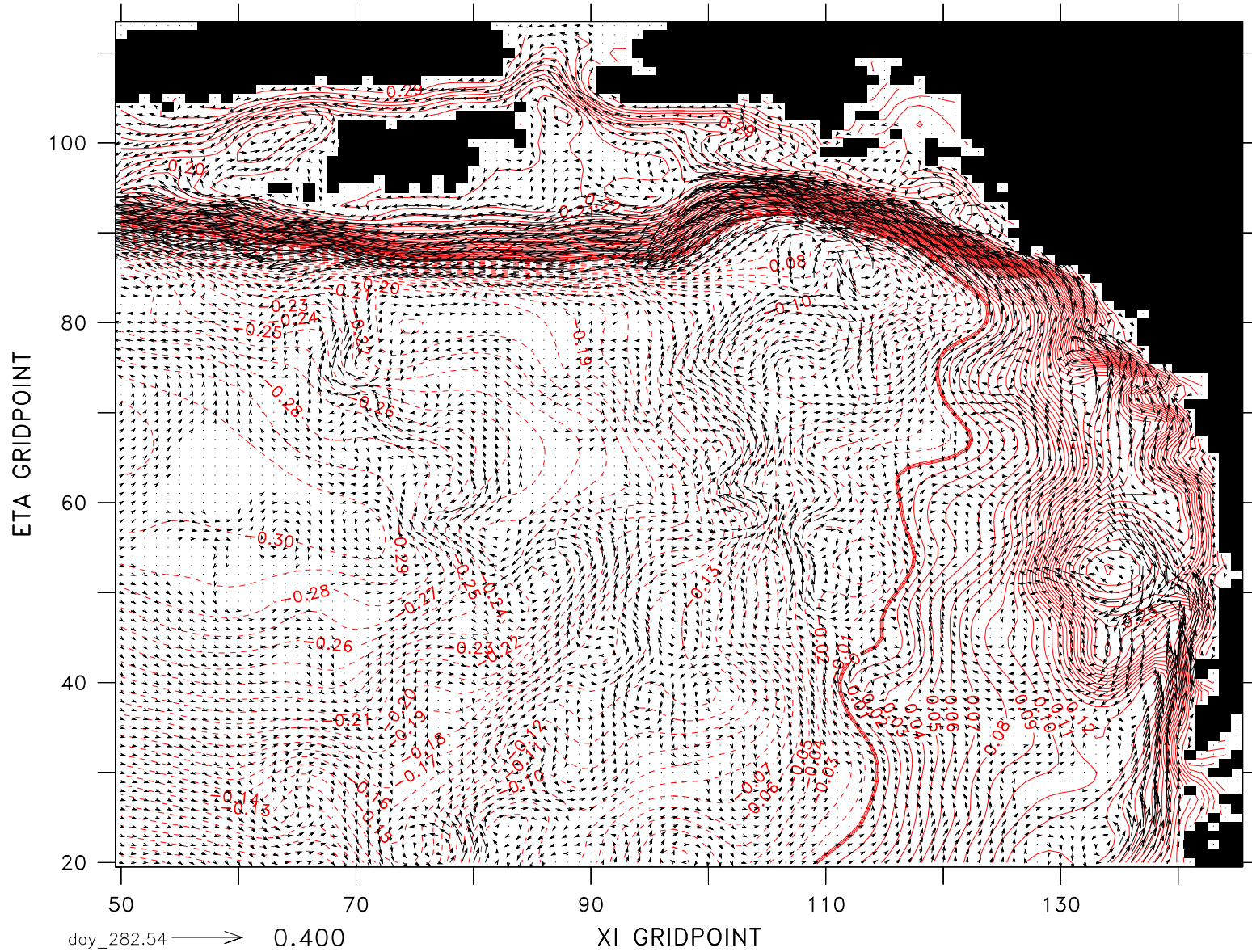


Fig 8

Coastal Gulf of Alaska, 1996

SCRUM TIME : 282.54 Modified Julian Days



Barotropic velocity (m/s) and SSH (shaded,m)

Figure 9

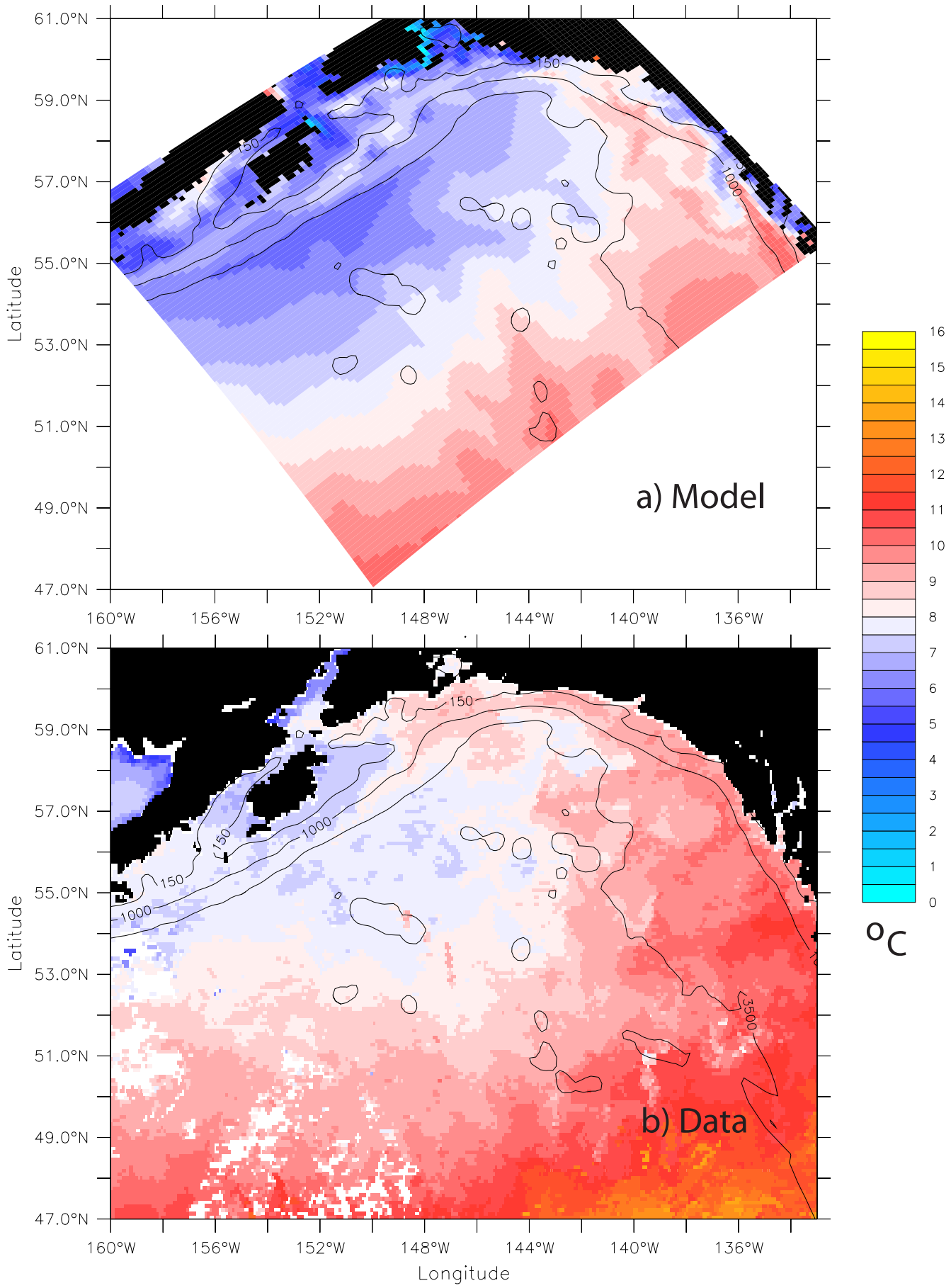
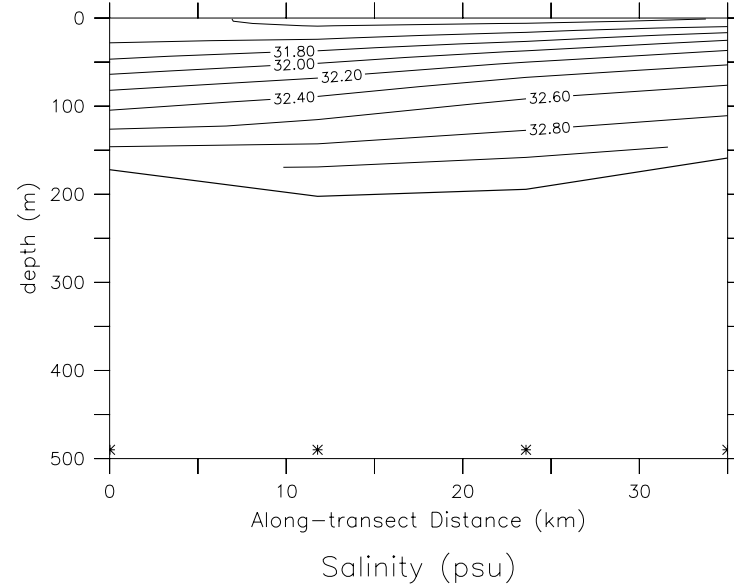
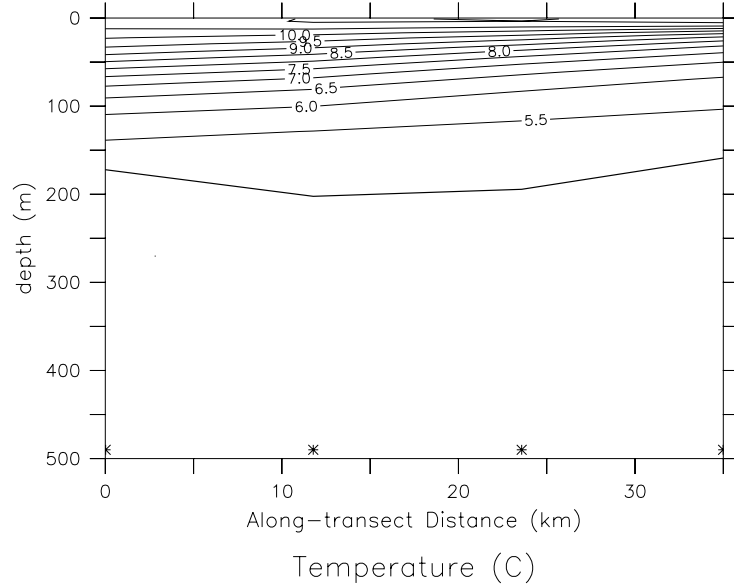


Fig 10

Model Output



Data Transect

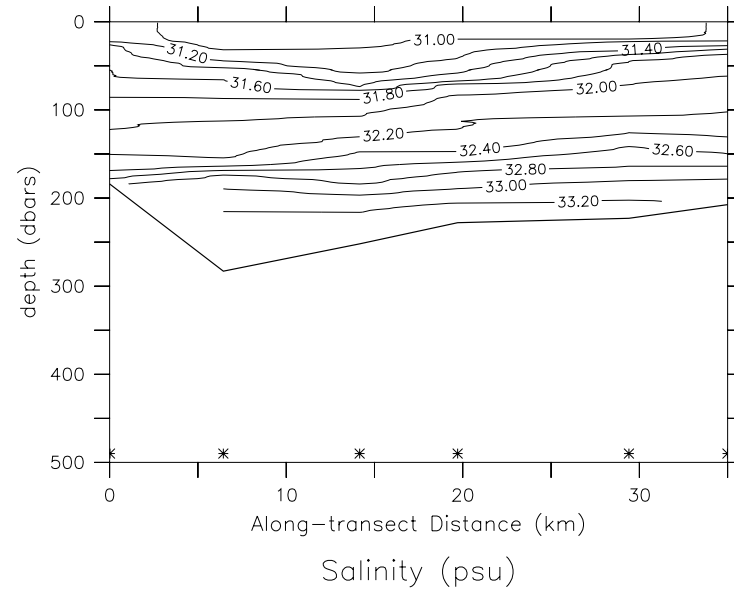
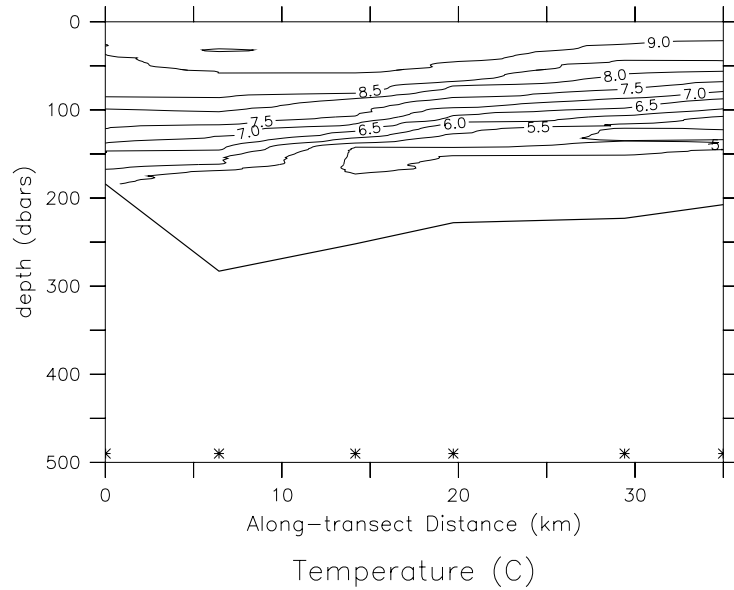
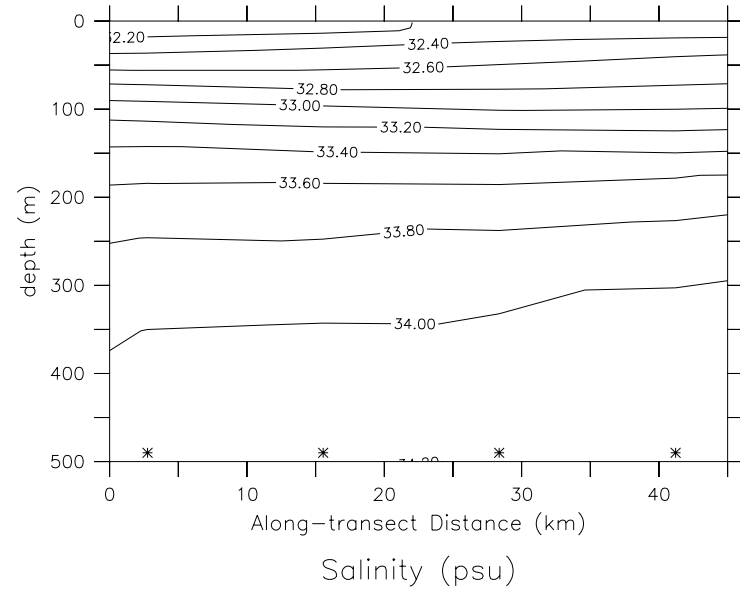
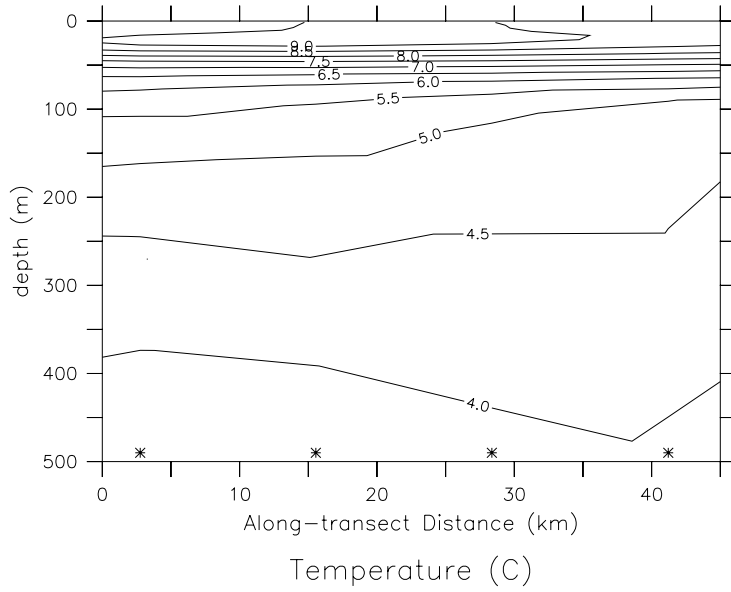


Fig 11

Model Output



Data Transect

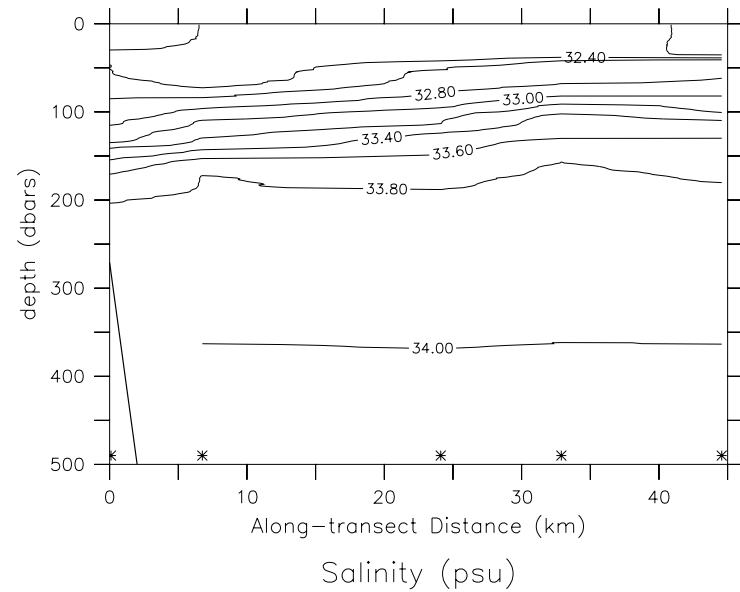
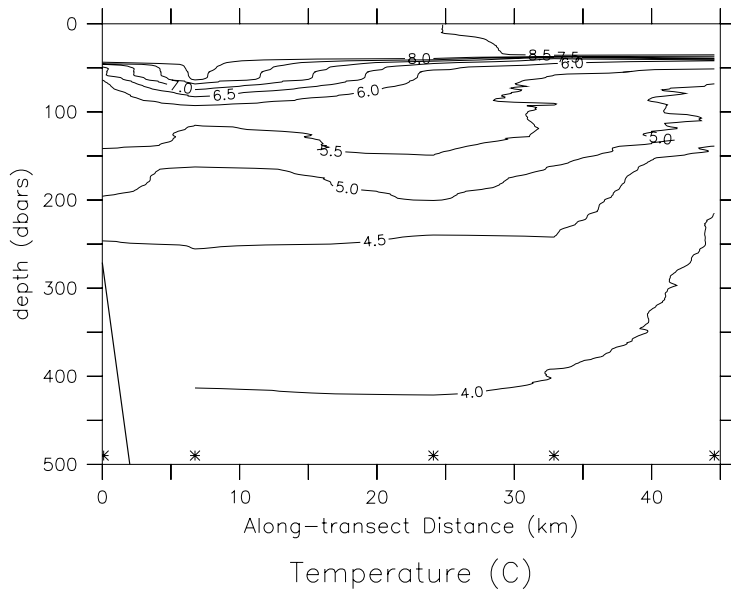


Fig 12

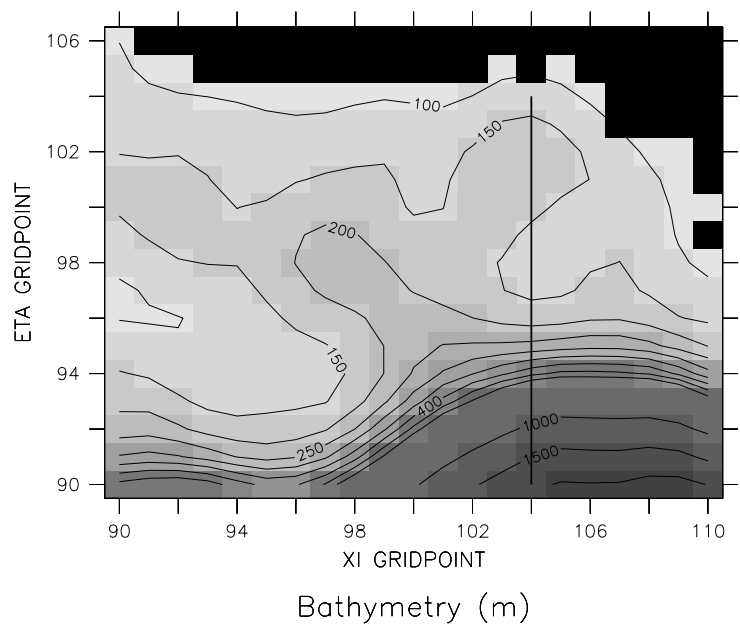
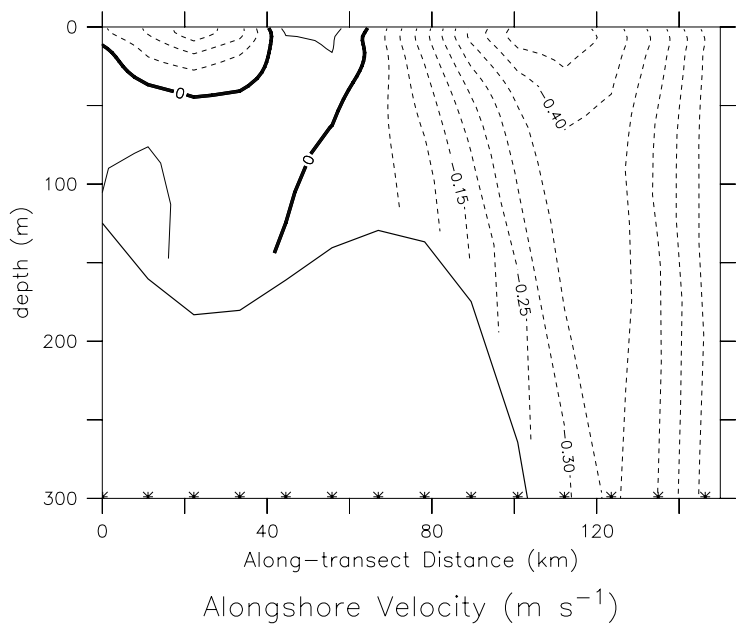
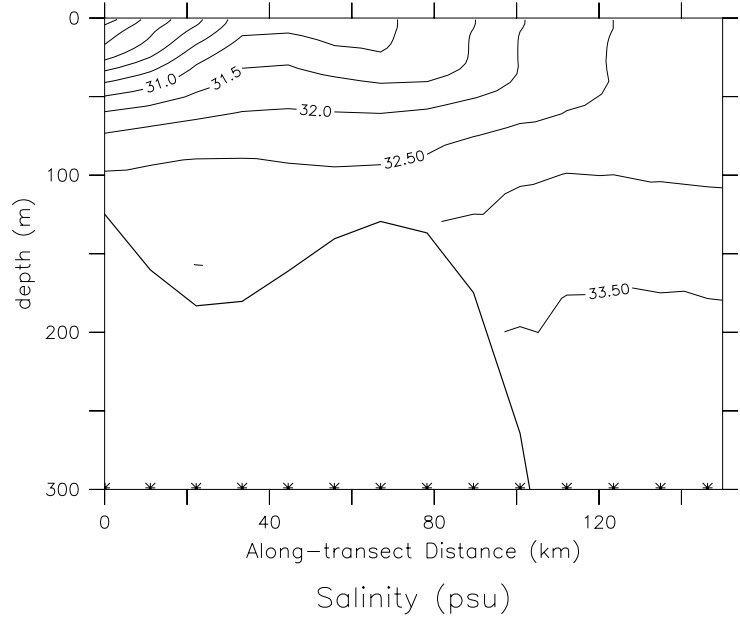
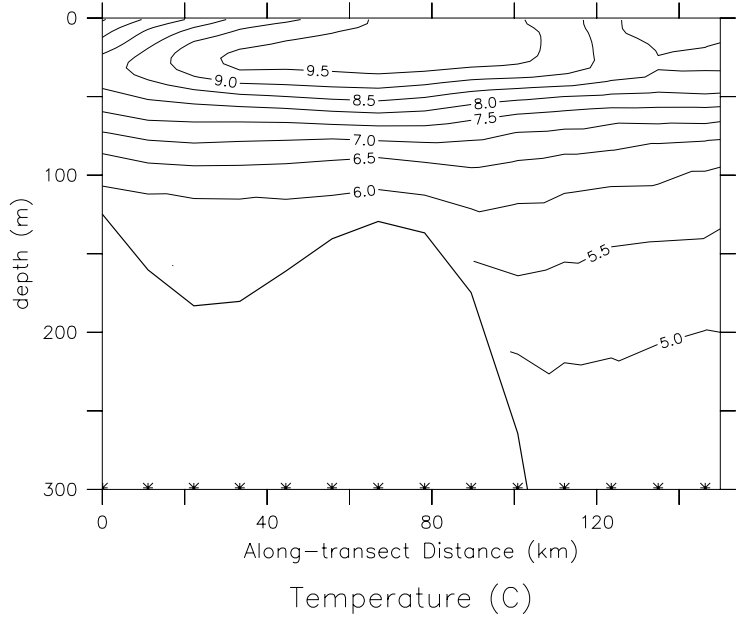
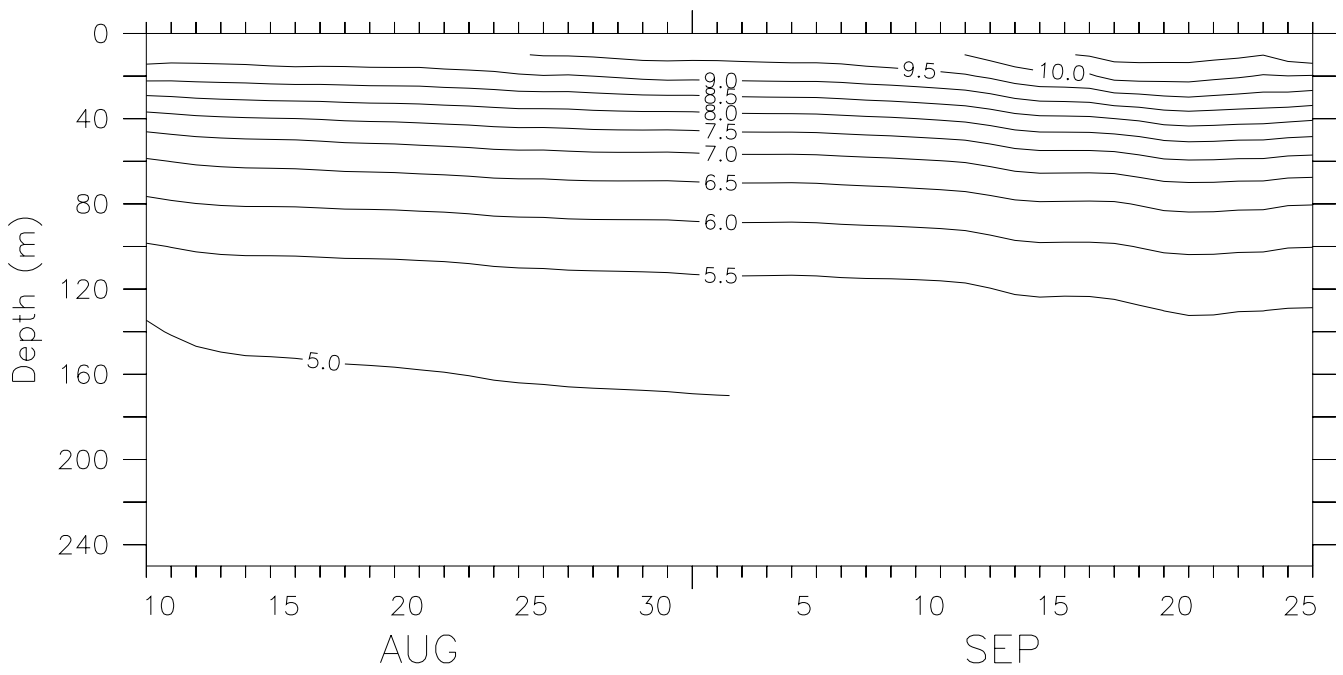
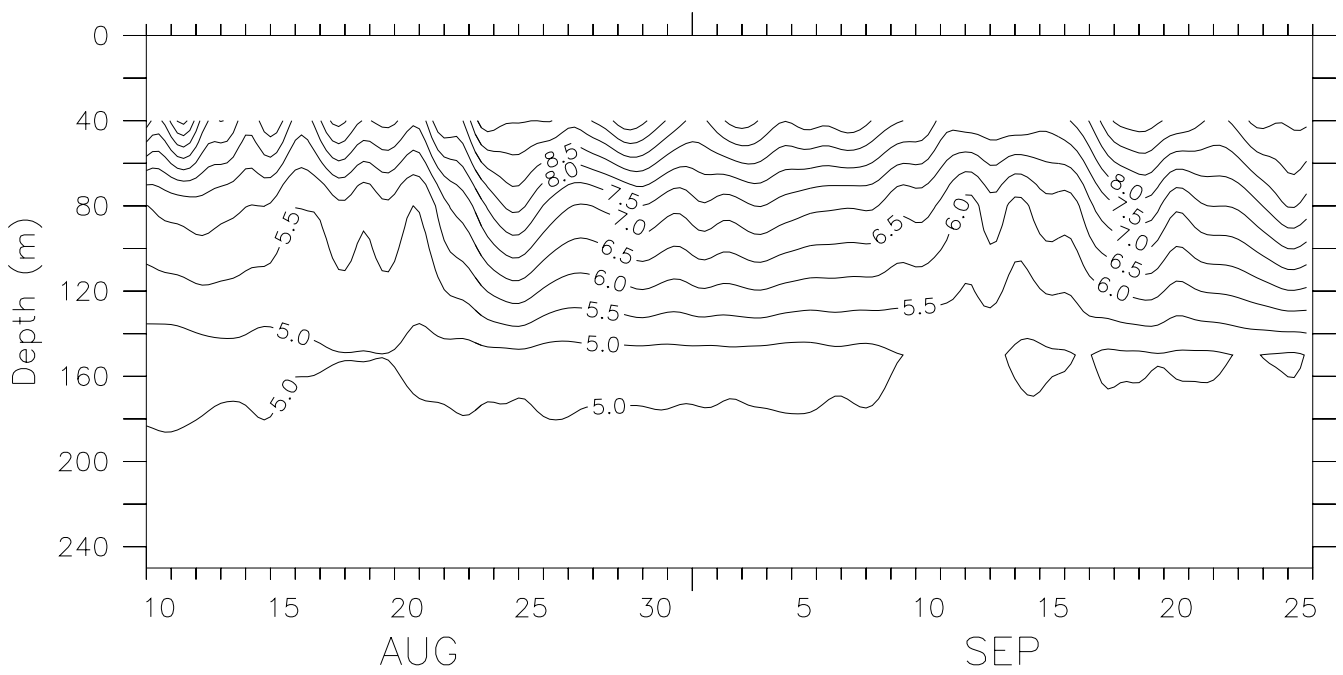


Figure 13

Temperature in Shelikof Strait



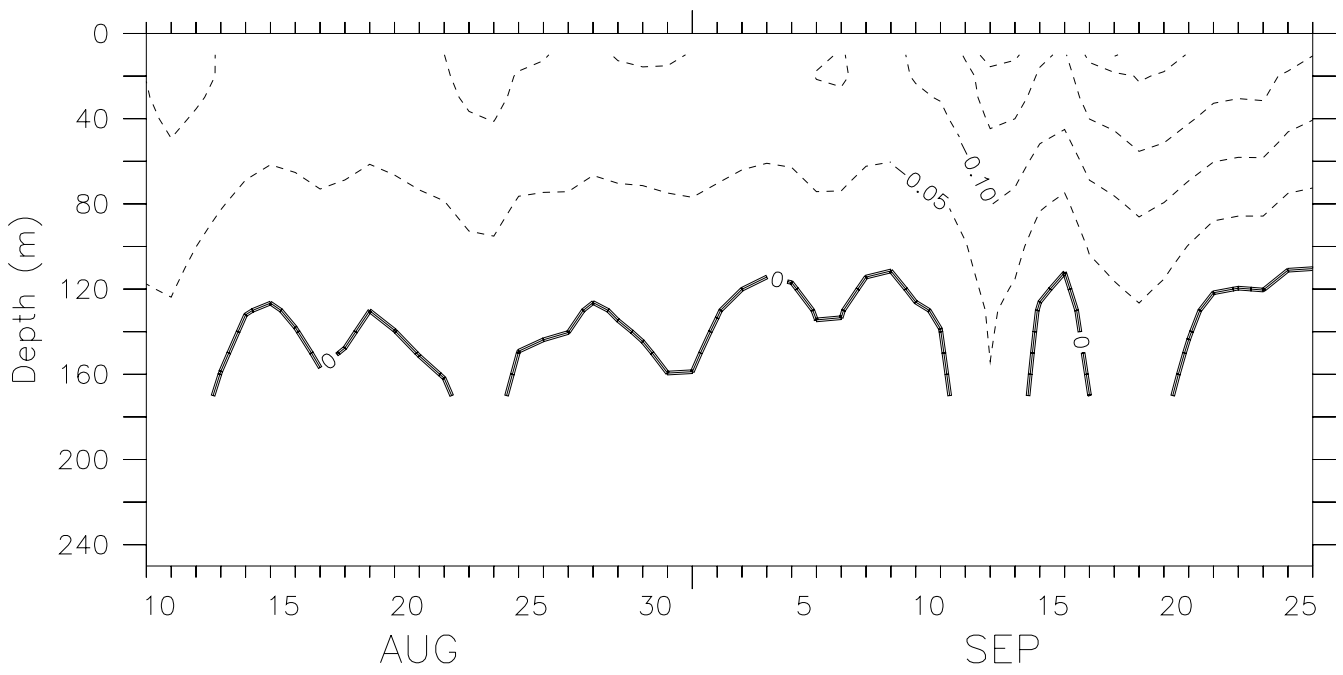
Modeled Temperature (C)



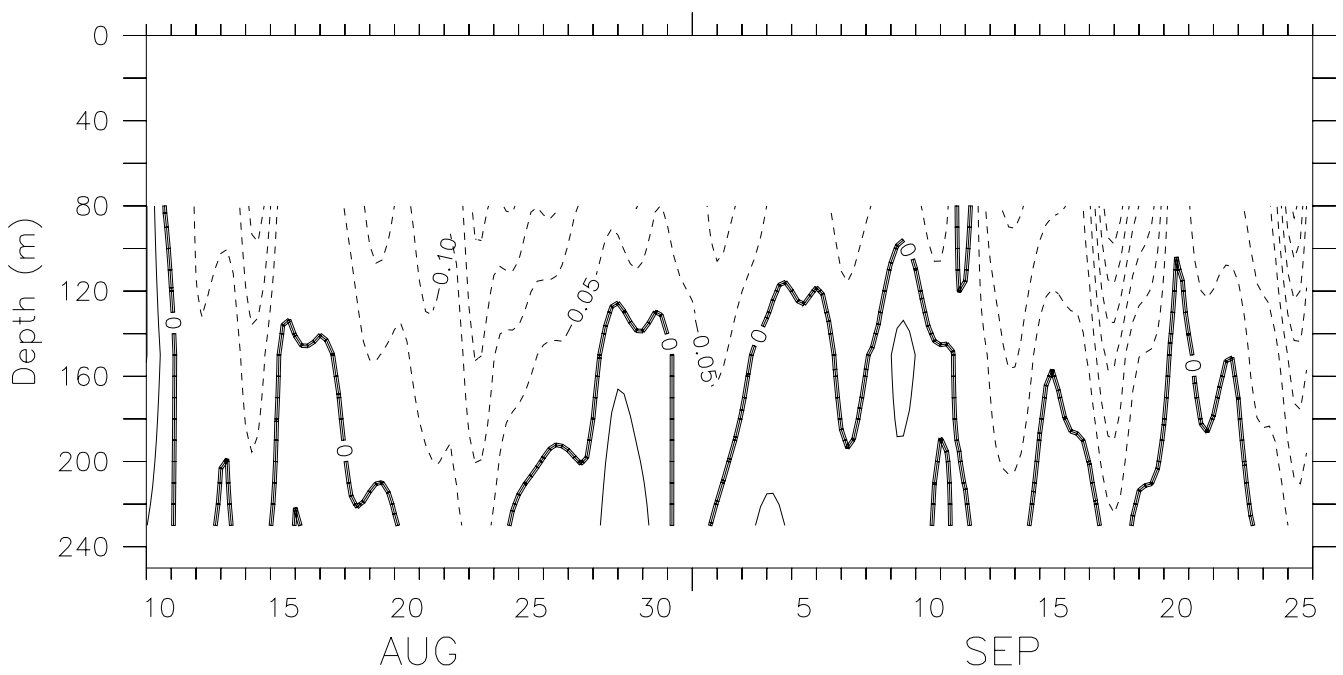
Measured Temperature (C)

Figure 14

Alongshelf Velocity in Shelikof Strait



Modeled Velocity (m s^{-1})



Measured Velocity (m s^{-1})

Figure 15

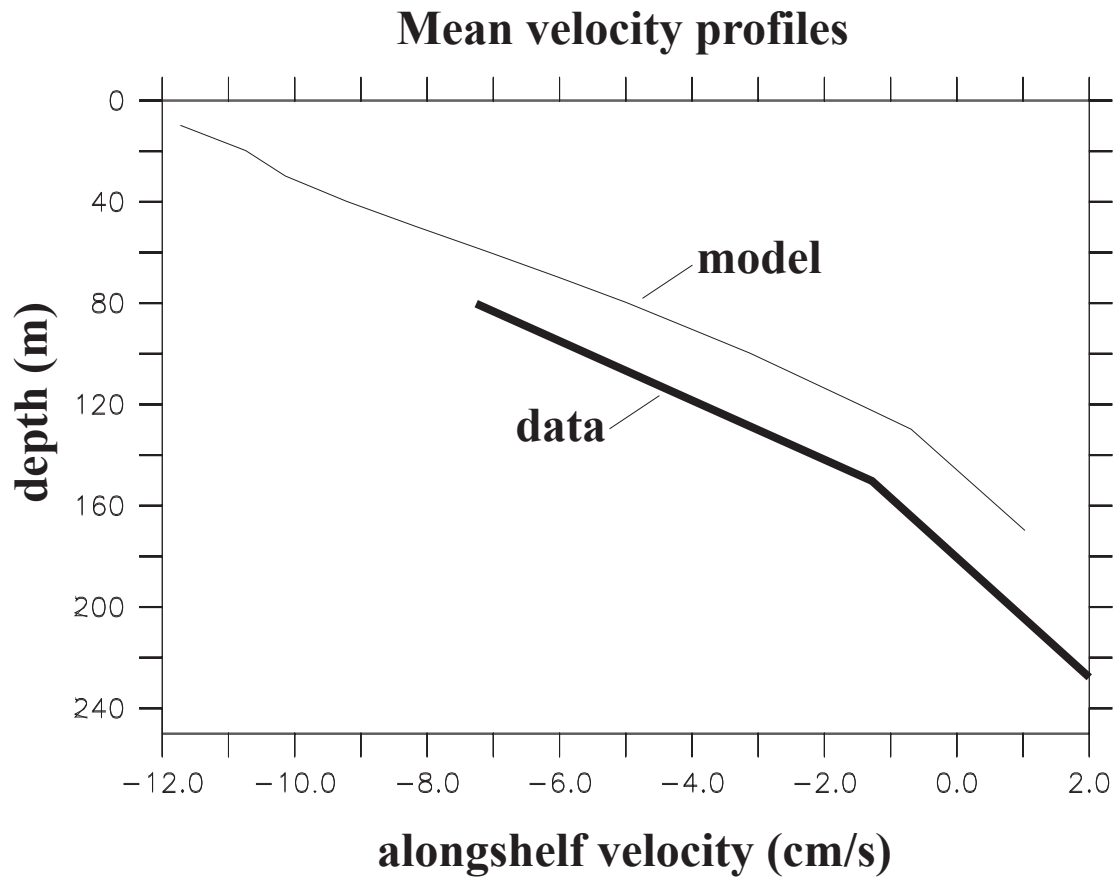


Figure 16

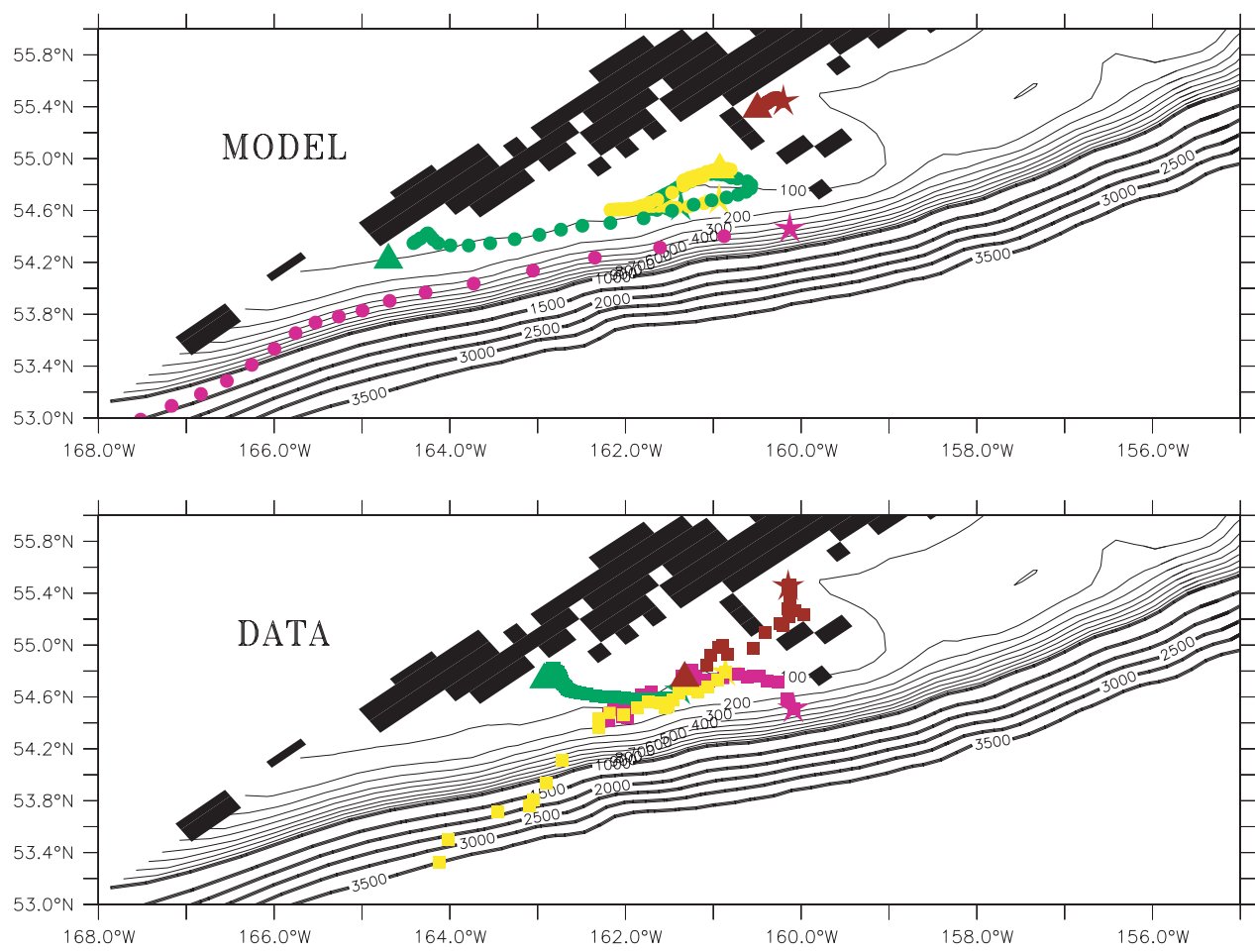
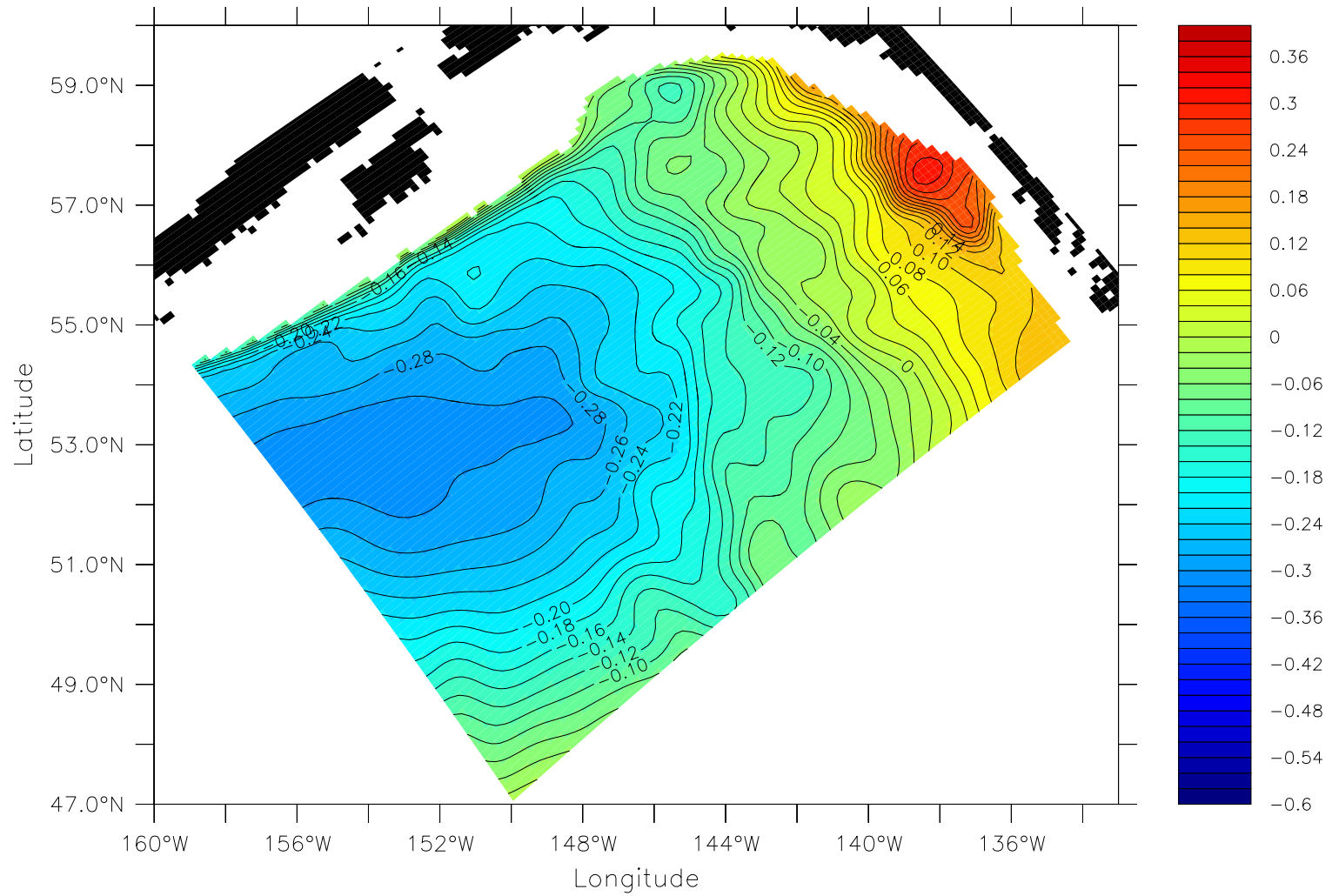
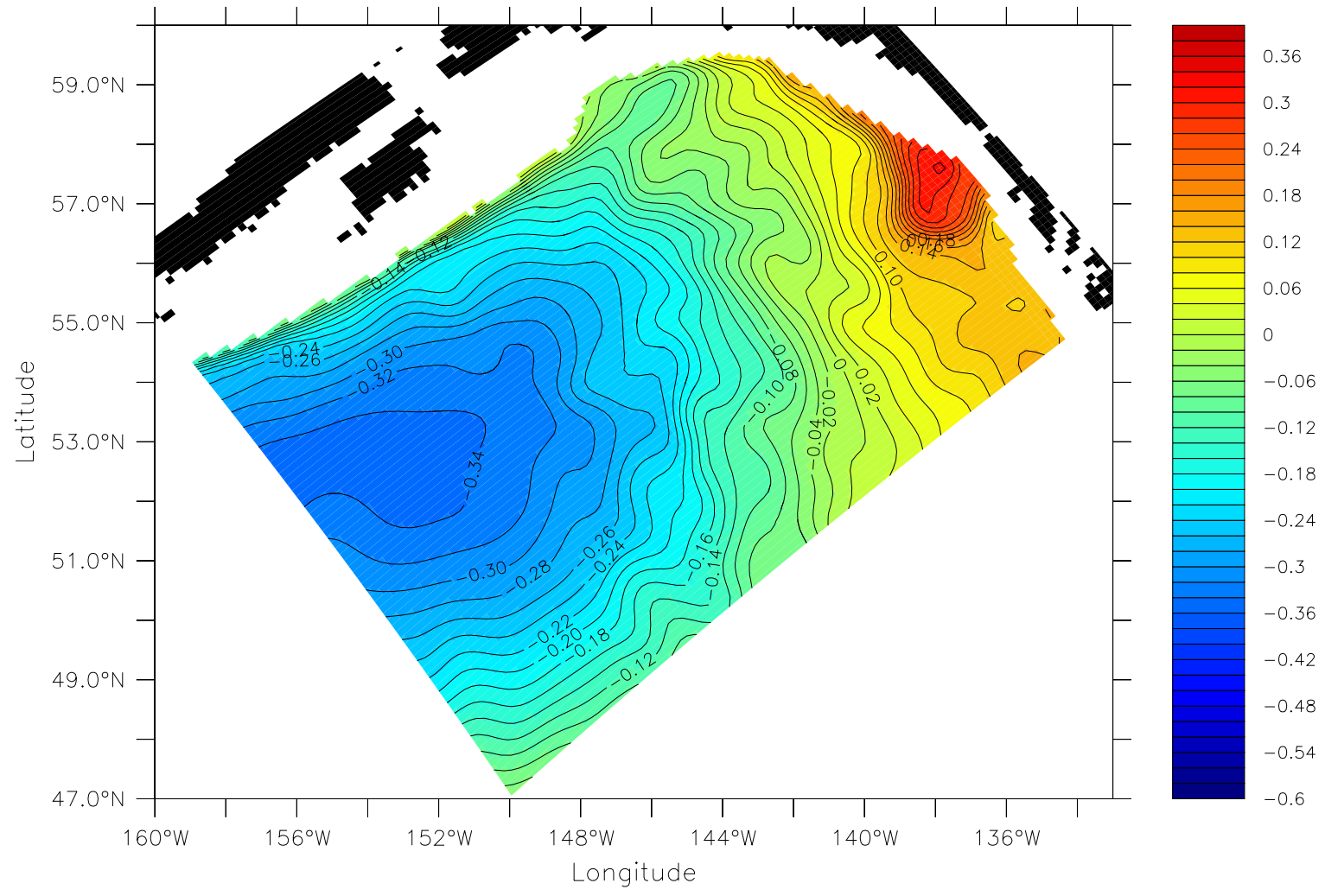


Fig 17a



Modeled sea surface height (m), Oct. 8, 1996

Fig 17b



Modeled sea surface height (m) without subtidal influence, Oct. 8, 1996

Figure 17c

TOPEX/ERS Analysis Oct 9 1996

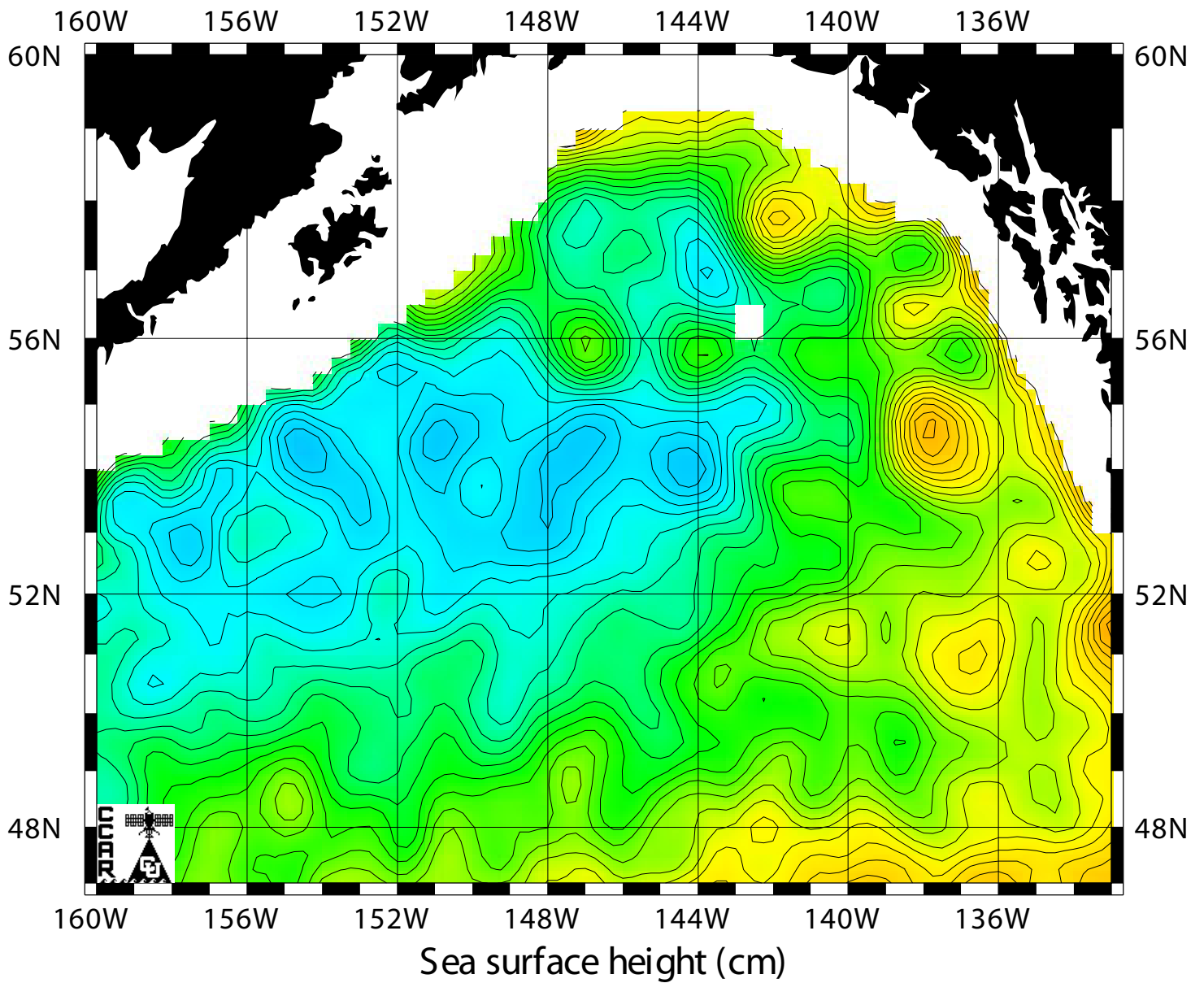
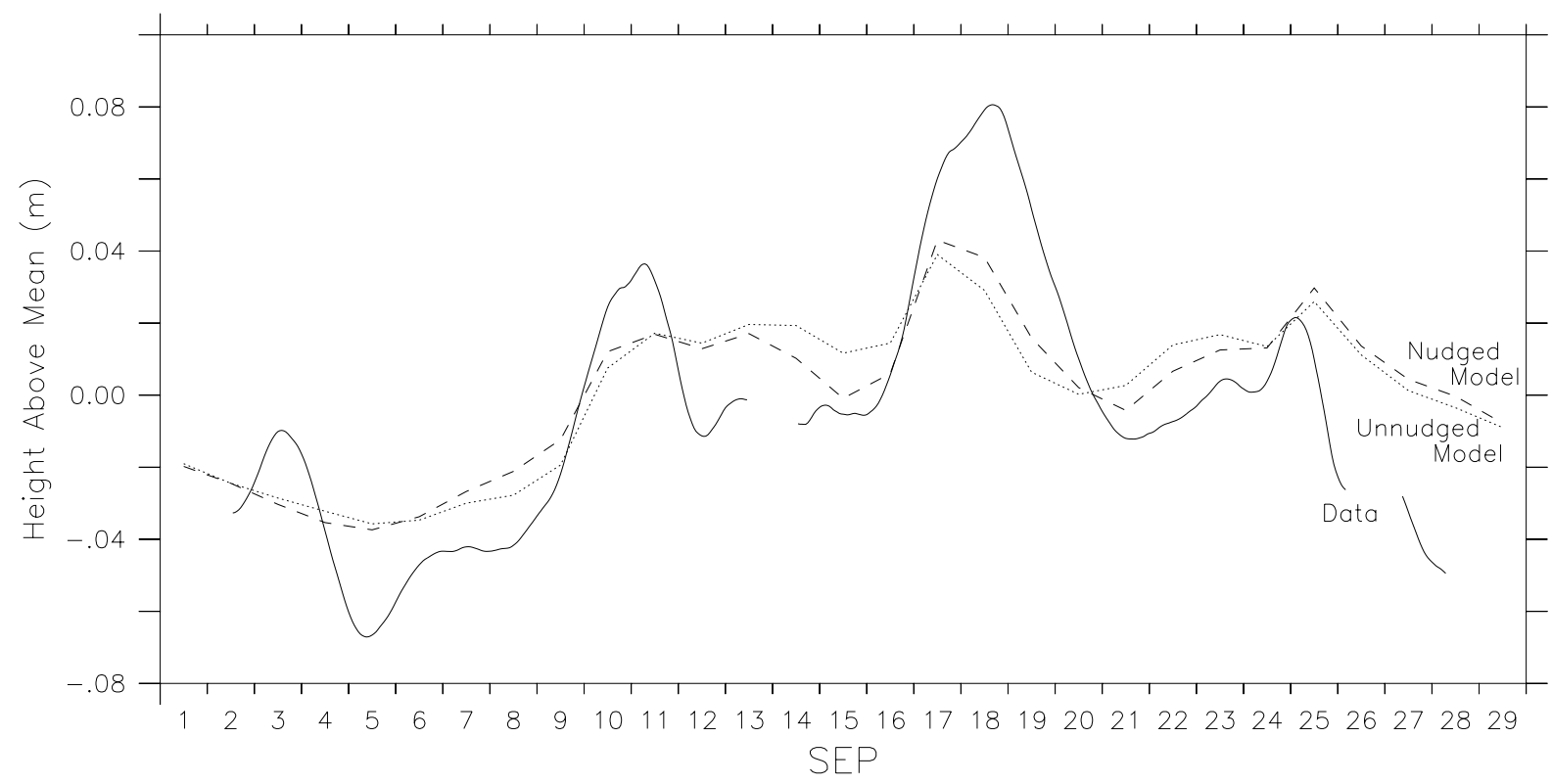
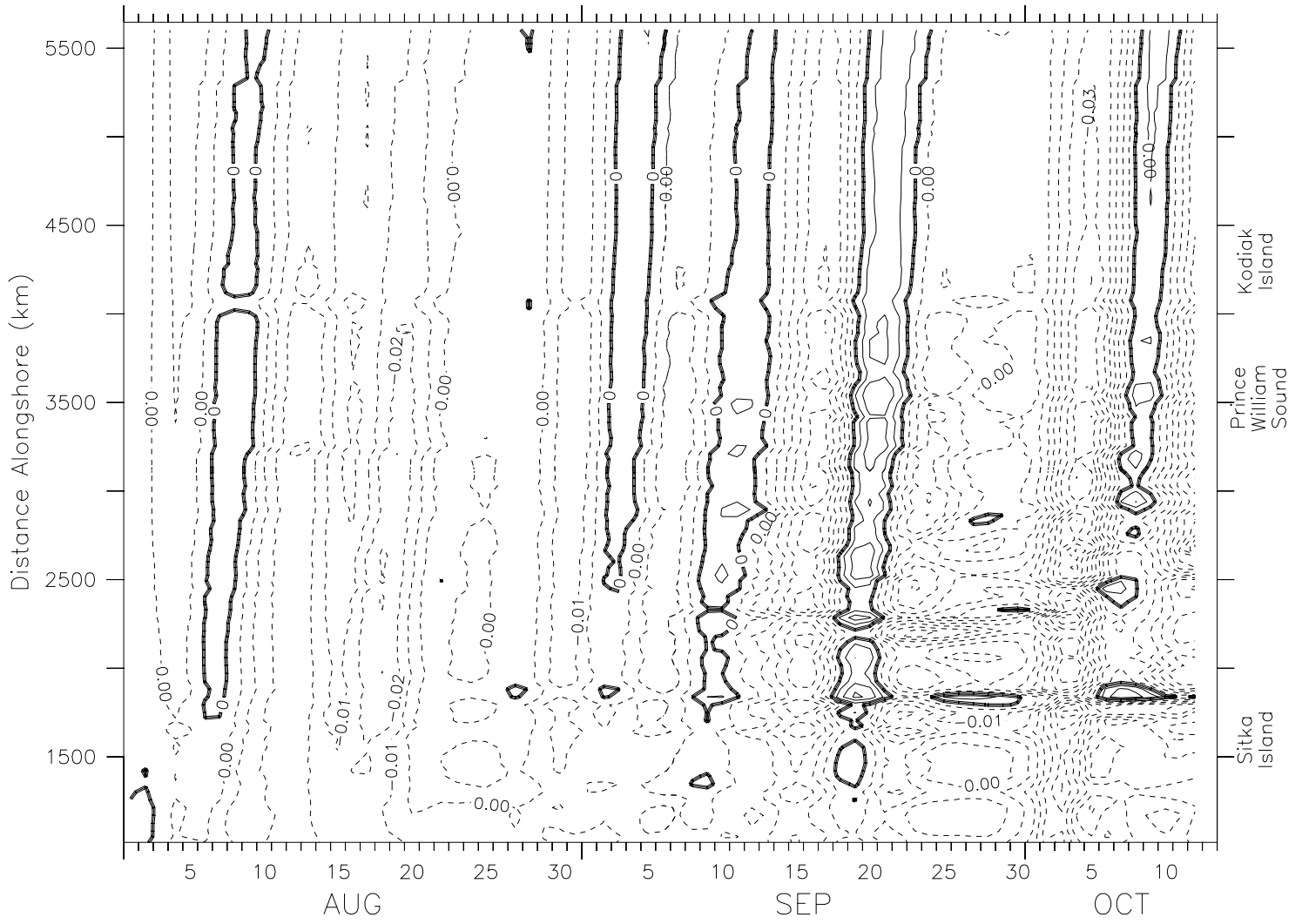


Fig 18



Subtidal Sea Surface Height, 1996

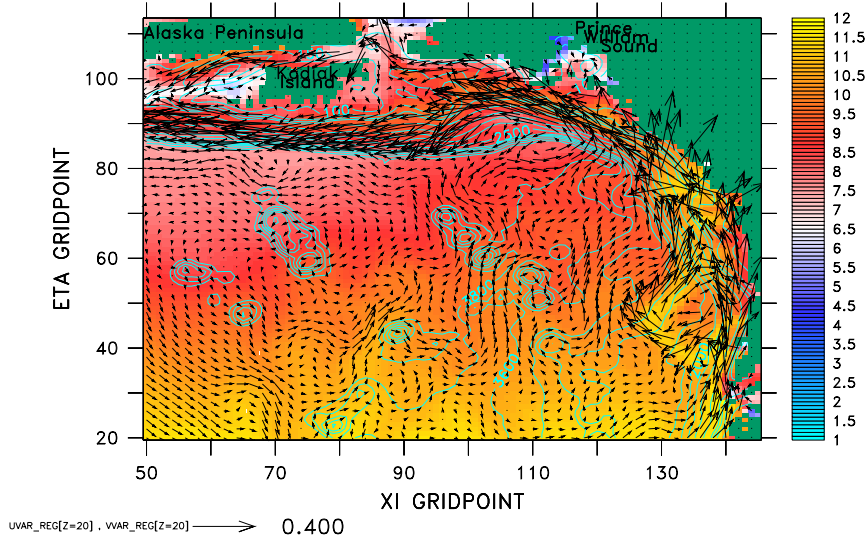
Figure 19



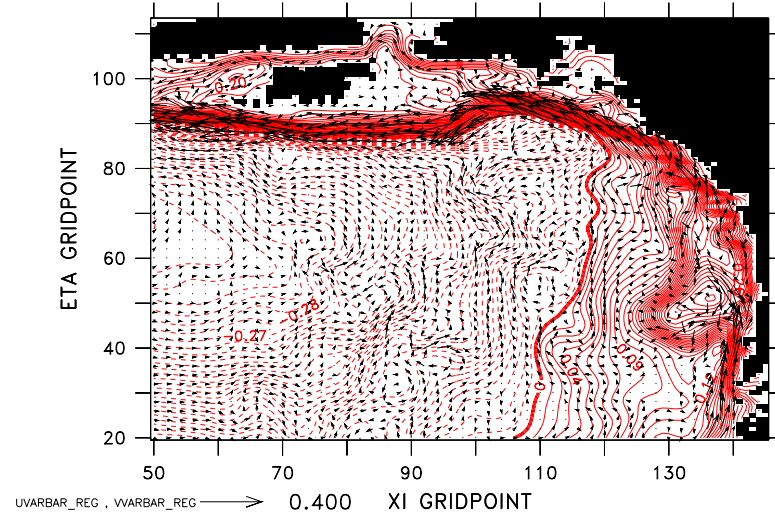
Contours of Subtidal Sea Surface Height (m)

Fig 20

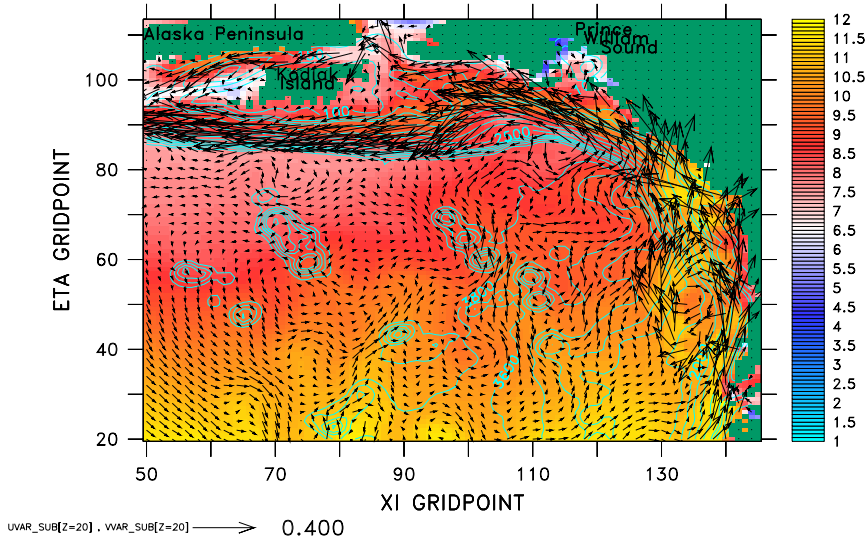
SST NO TIDAL/SUBTIDAL FORCING



SSH NO TIDAL/SUBTIDAL FORCING



SST WITH SUBTIDAL FORCING



SSH WITH SUBTIDAL FORCING

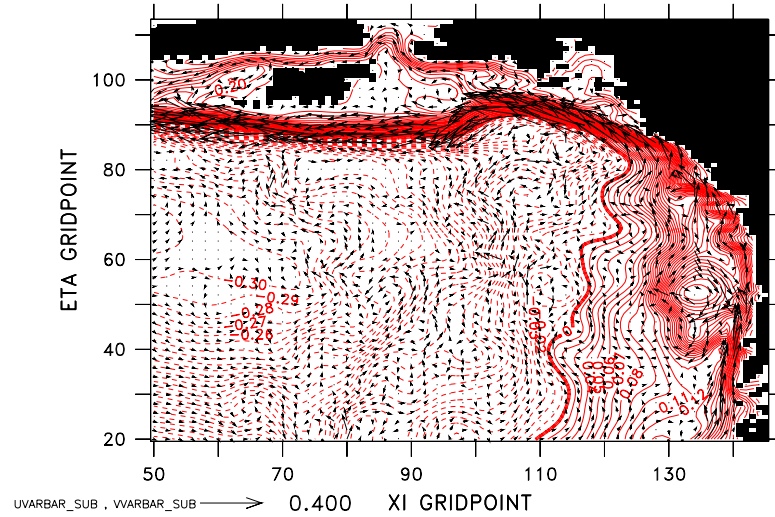
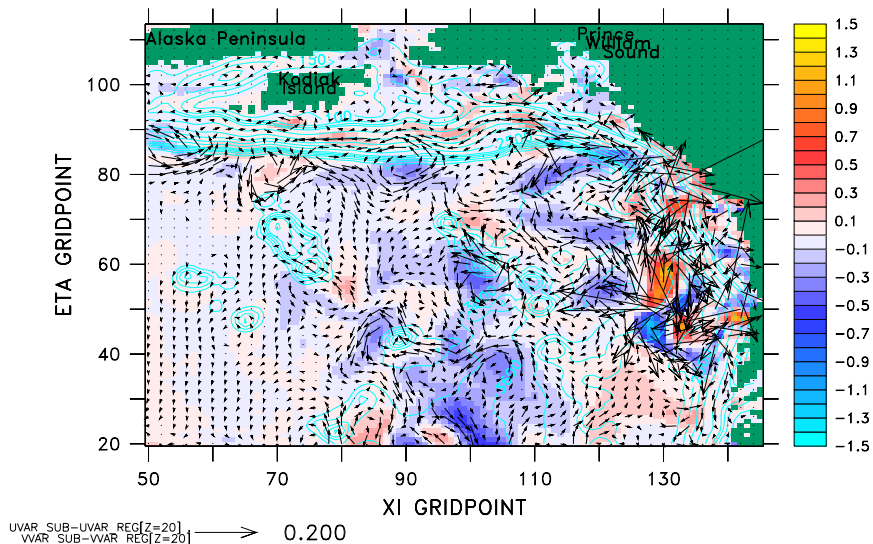
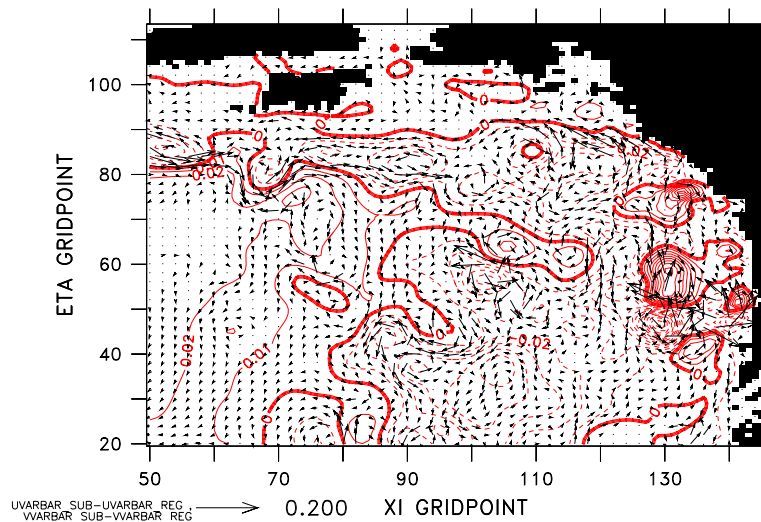


Figure 21

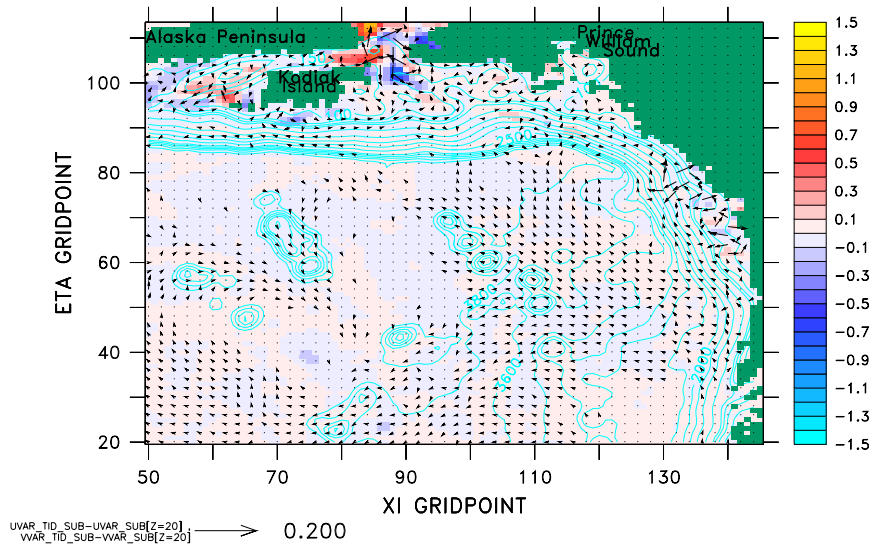
SST CHANGE SUBTIDAL FORCING



SSH CHANGE SUBTIDAL FORCING



SST CHANGE TIDAL FORCING



SSH CHANGE TIDAL FORCING

