



# Intrinsic and forced interannual variability of the Gulf of Alaska mesoscale circulation

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## Abstract

The response of the Gulf of Alaska (GOA) circulation to large-scale North Pacific climate variability is explored using three high resolution (15 km) regional ocean model ensembles over the period 1950–2004. On interannual and decadal timescales the mean circulation is strongly modulated by changes in the large scale climate forcing associated with PDO and ENSO. Intensification of the model gyre scale circulation occurs after the 1976–1977 climate shift, as well as during 1965–1970 and 1993–1995. From the model dynamical budgets we find that when the GOA experiences stronger southeasterly winds, typical during the positive phase of the PDO and ENSO, there is net large-scale Ekman convergence in the central and eastern coastal boundary. The geostrophic adjustment to higher sea surface height (SSH) and lower isopycnals lead to stronger cyclonic gyre scale circulation. The opposite situation occurs during stronger northwesterly winds (negative phase of the PDO).

Along the eastern side of the GOA basin, interannual changes in the surface winds also modulate the seasonal development of high amplitude anticyclonic eddies (e.g. Haïda and Sitka eddies). Large interannual eddy events during winter-spring, are phase-locked with the seasonal cycle. The initial eddy dynamics are consistent with a quasi-linear Rossby wave response to positive SSH anomalies forced by stronger downwelling favorable winds (e.g. southwesterly during El Niño). However, because of the fast growth rate of baroclinic instability and the geographical focusing associated with the coastal geometry, most of the perturbation energy in the Rossby wave is locally trapped until converted into large scale nonlinear coherent eddies. Coastally trapped waves of tropical origin may also contribute to positive SSH anomalies that lead to higher amplitude eddies. However, their presence does not appear essential. The model ensembles, which do not include the effects of equatorial coastally trapped waves, capture the large Haïda and Sitka eddy events observed during 1982 and 1997 and explain between 40% and 70% of the tidal gauges variance along the GOA coast.

In the western side of the GOA basin, interannual eddy variability located south of the Alaskan Stream is not correlated with large scale forcing and appears to be intrinsic. A comparison of the three model ensembles forced by NCEP winds and a multi-century-long integration forced only with the seasonal cycle, shows that the internal variability alone explains most of the eddy variance. The asymmetry between the eddy forced regime in the eastern basin, and the intrinsic regime in the western basin, has important implications for predicting the GOA response to climate change. If future climate change results in stronger wintertime winds and increased downwelling in the eastern basin, then increased mesoscale activity (perhaps more or larger eddies) might occur in this region. Conversely, the changes in the western basin are not predictable based on environmental forcing. Eastern eddies transport important biogeochemical quantities such as iron, oxygen

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and chlorophyll-a into the gyre interior, therefore having potential upscale effects on the GOA high-nutrient-low-chlorophyll region.

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## 1. Introduction

Observational and modeling studies suggest that decadal-scale variability in the Gulf of Alaska (GOA) gyre circulation is strongly controlled by changes in the atmospheric forcing (Lagerloef, 1995; Cummins and Lagerloef, 2004). Decadal and interannual variability in wind stress over the GOA is correlated with the Pacific Decadal Oscillation (PDO) and the Southern Oscillation. Intensification of the Aleutian low during 1976–1977, also referred to as the “climate shift” (Miller et al., 1994) modified the surface ocean circulation of the GOA (Capotondi et al., 2005; Miller et al., 2005). More specifically, Miller et al. (2005) suggest that an intensification of the Alaskan Stream occurred after the 1976–1977 climate shift, although the Alaska Current and eddies generated on the eastern GOA remain mostly unchanged. In addition to local changes in surface winds, coastally trapped Kelvin waves generated in the tropics during the El Niño Southern Oscillation (ENSO) affect the GOA circulation (Meyers and Basu, 1999).

Despite the numerous observational studies over the northeast Pacific Ocean (Meyers and Basu, 1999; Crawford, 2002; Strub and James, 2002; Brickley and Thomas, 2004; Ladd et al., 2005), important aspects of the GOA dynamics (e.g., mesoscale activity) have been rarely studied or modeled, and thus are not well understood. Some modelers have nevertheless focused their studies on mesoscale circulation in the GOA, such as Haïda or Sitka eddies (Melsom et al., 1999; Di Lorenzo et al., 2005a). Okkonen et al. (2001) have also suggested that eddy variability is dominated by the local response to wind forcing along the Alaska Current (eastern GOA basin) and is non-deterministic in the Alaskan Stream (western GOA basin).

Eddies in the GOA have already been studied for several decades using available observations (Tabata, 1982). In the eastern GOA basin, three major groups of eddies have been identified according to the location of their formation, Yakutat, Sitka and Haïda (Fig. 1a). These three eddy groups share common features, such as an anti-cyclonic rotation (up to 0.4 m sea surface height anomaly) with a warmer and less saline core above 100 m depth. (Tabata, 1982; Crawford, 2002). The generation and migration of these eddies, which can survive for more than three years, is well documented in satellite observations (Thomson and Gower, 1998; Crawford, 2002). Haïda and Sitka eddies mostly propagate offshore while the Yakutat eddies propagate alongshore along the Alaskan Stream. While previous studies attribute the generation of these large eddies to baroclinic instability of the coastal currents (Thomson and Gower, 1998; Murray et al., 2001), a more recent numerical investigation by Di Lorenzo et al. (2005a) suggests that the most southern eddy, the Haïda eddy, can develop without instabilities by simple advection of buoyant water masses around Cape St. James (southern tip of Queen Charlotte Island). These advected water masses generate patches of buoyant flow with anticyclonic circulation which merge to generate larger eddies when the flow around the cape is strong. For the more northern Sitka and Yakutat eddies a more detailed modeling study of their formation has not been completed.

In addition to the eddies developed in the eastern boundary along the Canadian coast, the western basin (here defined as west of 150°W) also exhibits enhanced eddy activity. Along the Alaskan Stream, eddies have been both observed (Crawford et al., 2000) and modeled (Okkonen et al., 2001) and tend to propagate southwestward along the Aleutian peninsula. In the Shelikof sea valley (north GOA basin) Staben and Hermann (1996) find a good agreement between eddies generated by a model and observations, however their study does not diagnose the eddy dynamics.

Understanding the dynamics of eddies and other mesoscale physical processes is critical for predicting how tracer exchanges between inshore and offshore water masses will respond to changes in atmospheric and ocean state (e.g. ENSO, PDO, global climate change). Eddies have several direct implications on the climate and biology of the GOA, because of their important role in cross-shelf transport. A recent study suggest that about 35–60% of northward heat transport during winter is carried offshore by the Haïda eddies (Crawford, 2005). Crawford (2005) estimates that 15% of the total river freshwater input between the Columbian River and 54°N is transported by these eddies. Nutrient exchanges (Whitney and Robert, 2002; Ladd et al., 2005) and iron

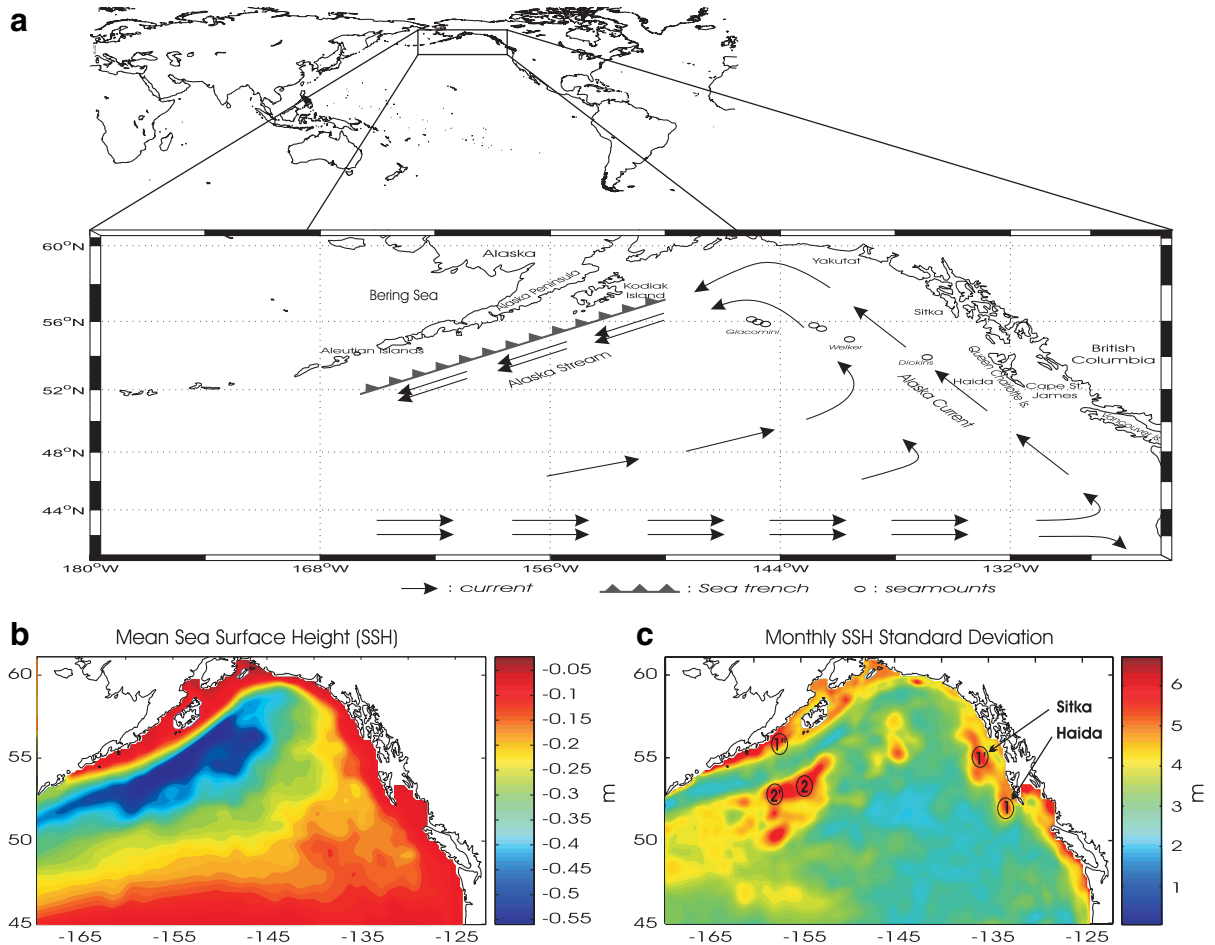


Fig. 1. (a) Map of the domain (Gulf of Alaska: GOA). (b) Mean and (c) standard deviation of the sea surface height (SSH).

transport (Johnson et al., 2005) also occur vertically in the core of eddies. The eddy induced transport is also linked to the biological productivity in the offshore high nutrient low Chlorophyll-*a* (HNLC) waters of the Alaskan gyre (Crawford et al., in press). Links between eddy transport, water mass properties and biota have also been reported in the GOA western basin (Okkonen et al., 2003) and eastern basin (Mackas and Galbraith, 2002). At the higher trophic levels the link between mesoscale processes and species distributions/abundance has not been fully established. However fish communities have been found to respond strongly to large-scale atmospheric changes, such as the 1976–1977 climate shift (Mantua et al., 1997; Litzow et al., 2006). Some of these changes may potentially be related to changes in eddy transport statistics as hypothesized by Miller et al. (2005) in a study of the Steller sea lions decline in the western GOA. Other important mechanisms for cross-shelf fluxes in the GOA include episodic upwelling, surface Ekman transport, upwelling seaward of a coastal wind jet and topographic steering (Stabeno et al., 2004).

In order to further clarify the mean, seasonal and interannual response of the GOA mesoscale circulation to atmospheric changes, we conduct numerical investigations with a high-resolution ocean model. Three ensembles, forced by the last 55 years of NCEP/NCAR wind stress reanalysis (Kalnay et al., 1996) and heat fluxes corrected using the NOAA extended SST (Smith and Reynolds, 2004), are used to separate the intrinsic and forced variability in the GOA. The primary goals of this study are to (1) understand how the surface circulation and eddy statistics are controlled by atmospheric forcing at the seasonal and interannual timescale, and (2) separate the intrinsic and forced variability in the GOA.

This paper is organized as follows. Section 2 describes the model experiments and the domain used for this study. Section 3 discusses the mean circulation in the GOA and its interannual modulations by changes in surface forcing during different phases of the PDO. Section 4 looks at the seasonal and eddy circulation in the western and eastern basin. A summary is provided in Section 5.

## 2. Modeling approach and domain

The Regional Ocean Modeling System (ROMS, [Haidvogel et al., 2000](#); [Shchepetkin and McWilliams, 2005](#)) is used to investigate the forced and intrinsic variability of the Gulf of Alaska circulation. ROMS is a free-surface, hydrostatic, eddy-resolving primitive equation ocean model that uses stretched, terrain-following coordinates in the vertical and orthogonal curvilinear coordinates in the horizontal. The hydrostatic primitive equations are solved using a split-explicit time-stepping scheme, which allows the separation of the barotropic and baroclinic components of the momentum equations with internal and external time steps. Different resolutions and different domains have previously been used with ROMS and provide reliable results for the North Pacific Ocean ([Marchesiello et al., 2003](#); [Di Lorenzo et al., 2005b](#); [Curchitser et al., 2005](#); [Curchitser et al., 2005](#)). Given the success of these previous ROMS applications, this study uses similar model configuration options. In particular we use the KPP scheme for vertical mixing in the surface boundary ([Large et al., 1994](#)) and implicit horizontal mixing associated with the third order upstream bias scheme ([Shchepetkin and McWilliams, 1998](#)).

The grid extends northward from 25N to 61N and westward from 111W to 179W. The average spatial resolution is between 19 km (South) to 13.4 km (North) with 30 vertical levels (with higher resolution at the upper ocean layer), that is to say a total number of 1,697,280 points (latitude  $\times$  longitude  $\times$  depth =  $272 \times 208 \times 30$ ). Although this resolution is of the same order of the Rossby deformation Radius, the model is eddy permitting and is able to capture qualitatively the large-scale Haida and Sitka eddies (diameters  $\geq 200$  km) along the eastern side of the GOA and the eddy centers along the Alaskan Stream. The domain of interest for this study includes all of the Gulf of Alaska from Washington State and presents open boundaries in the south and west. Along these open boundaries we use a modified radiation boundary condition ([Marchesiello et al., 2001](#)) together with nudging to observed monthly mean climatologies ([Levitus et al., 1994](#)). The nudging is stronger (timescale of 1 day) if the direction of the flow is inward and weaker (timescale of 1 year) for the outflow.

In order to avoid large pressure gradient errors ([Mellor et al., 1994](#)), the model bathymetry is obtained by a smooth interpolation from David T. Sandwell and Walter H. F. Smith bathymetry ([Smith and Sandwell, 1994](#); [Sandwell and Smith, 1997](#)).

Experiment 1 ([Table 1](#)) consists of three ensembles, each forced by the last 55 years of NCEP/NCAR (National Center for Environmental Prediction/National Center for Atmospheric Research) wind stress reanalysis ([Kalnay et al., 1996](#)) and by heat fluxes corrected using the NOAA (National Oceanic & Atmospheric Administration) extended SST ([Smith and Reynolds, 2004](#)). The freshwater surface fluxes are set to a corrected monthly climatology derived during an 80 year long spinup run in which the surface salinity is nudged to the observed climatology and the resulting flux correction is saved.

Experiment 2 differs from Experiment 1 in the surface forcing conditions, which are prescribed to the monthly mean climatology. The initial and boundary conditions remain unchanged. We refer to Experiment

Table 1  
Table of experiments

	Run length	Initial conditions	Open boundaries	Surface forcing (monthly-means)
Experiment 1	163 years (53 + 55 + 55)	<a href="#">Levitus et al. (1994)</a>	Temperature, salinity and geostrophic velocity derived from <a href="#">Levitus et al. (1994)</a> + nudging	55 years wind stress of NCEP/NCAR heat fluxes corrected with NOAA extended SST fresh water fluxes from corrected climatology
Experiment 2	163 years (53 + 55 + 55)			monthly climatological forcings derived from Experiment 1 forcing functions



2 as the “Unforced Run”, in that there is no forcing function with periodicity longer than the seasonal cycle. In contrast, Experiment 1, which is forced by the time dependent NCEP winds and NOAA SST, is referred to as the “Forced Run”.

For the initial condition, zonal and meridional velocity, sea surface height, temperature, and salinity are obtained from Levitus data (1994). The initial conditions of the ensembles 2 and 3 use respectively the final states of ensemble 1 and ensemble 2. At the open boundaries all model integration use the modified radiation condition (Marchesiello et al., 2001) and nudging to prescribed monthly climatologies for temperature, salinity and geostrophic velocity derived from Levitus et al. (1994). The geostrophic velocity is computed using the bottom as the reference level of no motion.

### 3. Mean and interannual circulation

The circulation in the GOA is mainly described by the subarctic gyre (Fig. 1). On the eastern boundary of the GOA basin, the North Pacific Drift diverges into the southward California Current and the northward Alaska Current. The Alaska Current is rich in meanders and eddies, such as Haïda and Sitka along the eastern boundary. As the Alaska Current flows along the Alaska Peninsula it becomes the Alaskan Stream, which has a more developed and richer eddy field associated with the strong instability of the mean currents (Melsom et al., 1999). Near the Kodiak Island, the Alaskan Stream is narrow (about 50 km) and strong ( $50 \text{ cm s}^{-1}$ ) (Stabeno et al., 2004). The currents are principally in geostrophic balance and consequently follow lines of constant dynamic height with the largest values on the right. The model mean circulation, represented in Fig. 1b by the sea surface height (SSH), is mostly consistent with this view described in the literature. In the eastern basin we find a broad northward current, corresponding to the Alaska Current, and an intensification of surface currents along the Aleutian Islands in the location of the Alaskan Stream. A map of the SSH standard deviation (Fig. 1c) shows high variance in regions of intense eddy variability along the southern edge of the Alaskan Stream, along the eastern basin, and in the coastal regions where the variability is directly forced by changes in the atmospheric winds (a discussion follows in Section 4). On the eastern boundary, maxima of standard deviation are found in the generation sites of the Haïda and Sitka eddies. A corresponding region of high variance is not found at the location of the Yakutat eddy, which is not well represented in the model.

A closer inspection of the model mean momentum budget for the surface layer confirms (Fig. 2a) that the dominant balance is geostrophic, with a stronger Alaskan stream ( $0.4 \text{ m s}^{-1}$ ) following the topographic slope in the northwestern basin and a weak northward flow in the eastern basin. The model geostrophic flow along the coast results from the ocean adjustment to the mean downwelling conditions in the eastern basin and upwelling along the topographic slope in the western basin. This is inferred from the model Ekman currents averaged over the surface layer (Fig. 2b) and the corresponding vertical fluxes of temperature (Fig. 2c), which show regions of surface divergence in the west and convergence in the east. A detailed analysis of the temperature budget shows that the dominant balance is between the vertical and the horizontal flux. The next largest term in the budget, although still small compared to the advection terms, is the surface heat flux (Fig. 2d). On average the surface heat fluxes are warming the Gulf of Alaska except in regions of strong downwelling (see blue areas in Fig. 2c and d) where the strong horizontal convergence of surface warmer water tends to heat the atmospheric marine boundary layer and cool the ocean surface.

To investigate the interannual modulation of the mean circulation we compute the first Empirical Orthogonal Function (EOF) of the model SSHa, where the anomalies are defined by removing the seasonal cycle. The resulting EOF 1 (Fig. 3a) explains 30% of the total variance and has the same structure as the mean circulation. Therefore temporal modulations in the amplitude and sign of this pattern, which are inferred from the Principal Component 1 (PC 1; Fig. 3e) of the three ensemble members, reflect interannual and decadal changes in the mean circulation. The spread of the PCs 1 of the ensemble members is very small suggesting that changes in the mean circulation are mainly forced. Indeed strong interannual modulations of the PCs 1 correspond to changes in the monthly PDO index with a significant correlation of 0.64 (Fig. 3f). Intensification of the gyre scale circulation is evident after the 1976–1977 climate shift, as well as during 1955–1960, 1965–1970 and more gradually during 1990–1998. Weakening of the circulations is evident during 1970–1975 and 1986–1991.

Composite maps of the difference in surface winds (Fig. 3c and d) suggest that the gyre intensifies during anomalous southeasterly winds (stronger Aleutian low) and weakens when the winds are predominantly

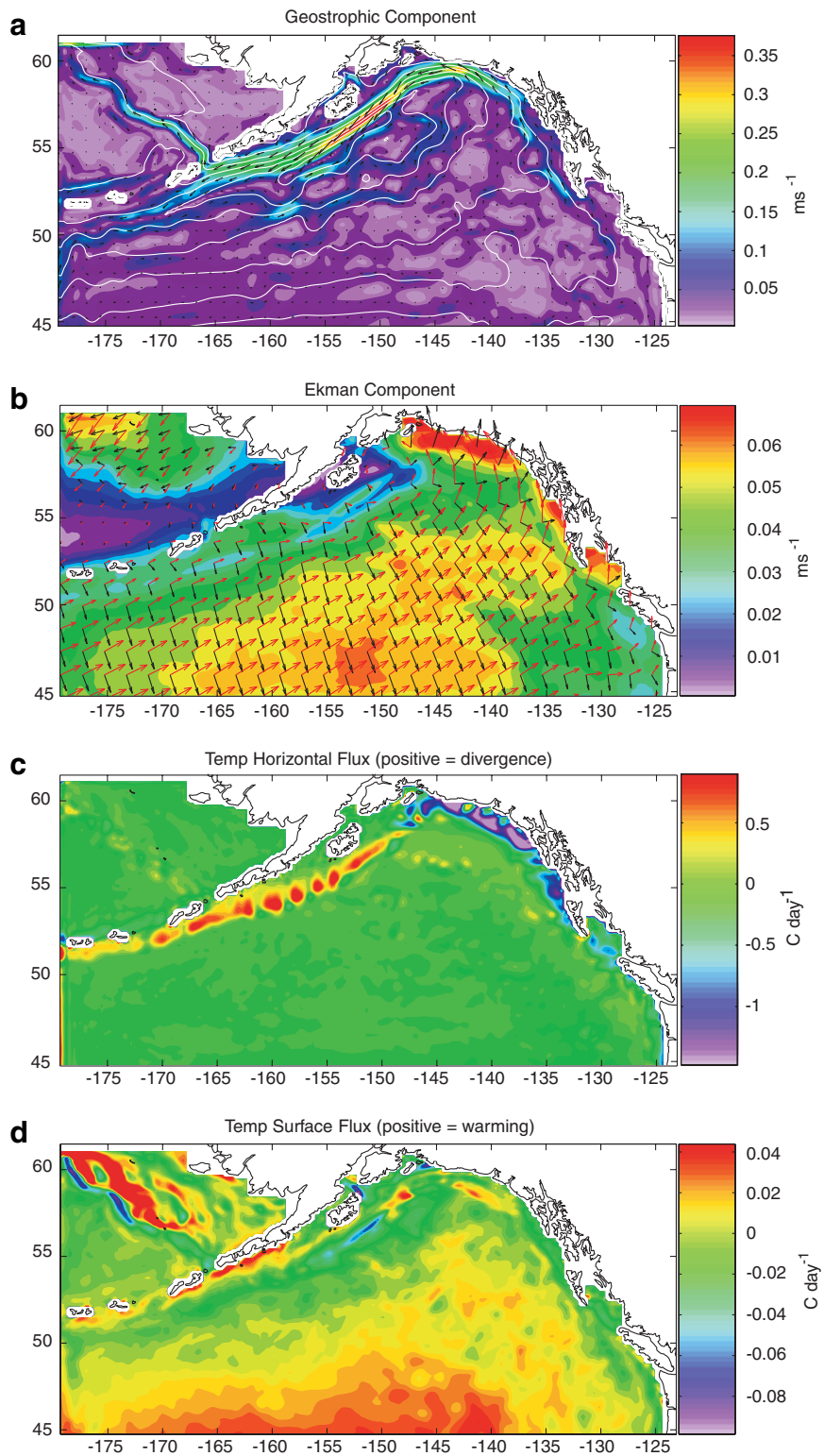


Fig. 2. (a) Mean geostrophic surface currents, with mean SSH contours (white lines). (b) Mean Ekman transport in the model surface layer. The red arrows indicate the direction and intensity of mean surface wind stresses. (c) Temperature horizontal flux, positive regions indicate upwelling. (d) Temperature surface flux, positive indicates that the ocean is warming due to the atmospheric heat flux.

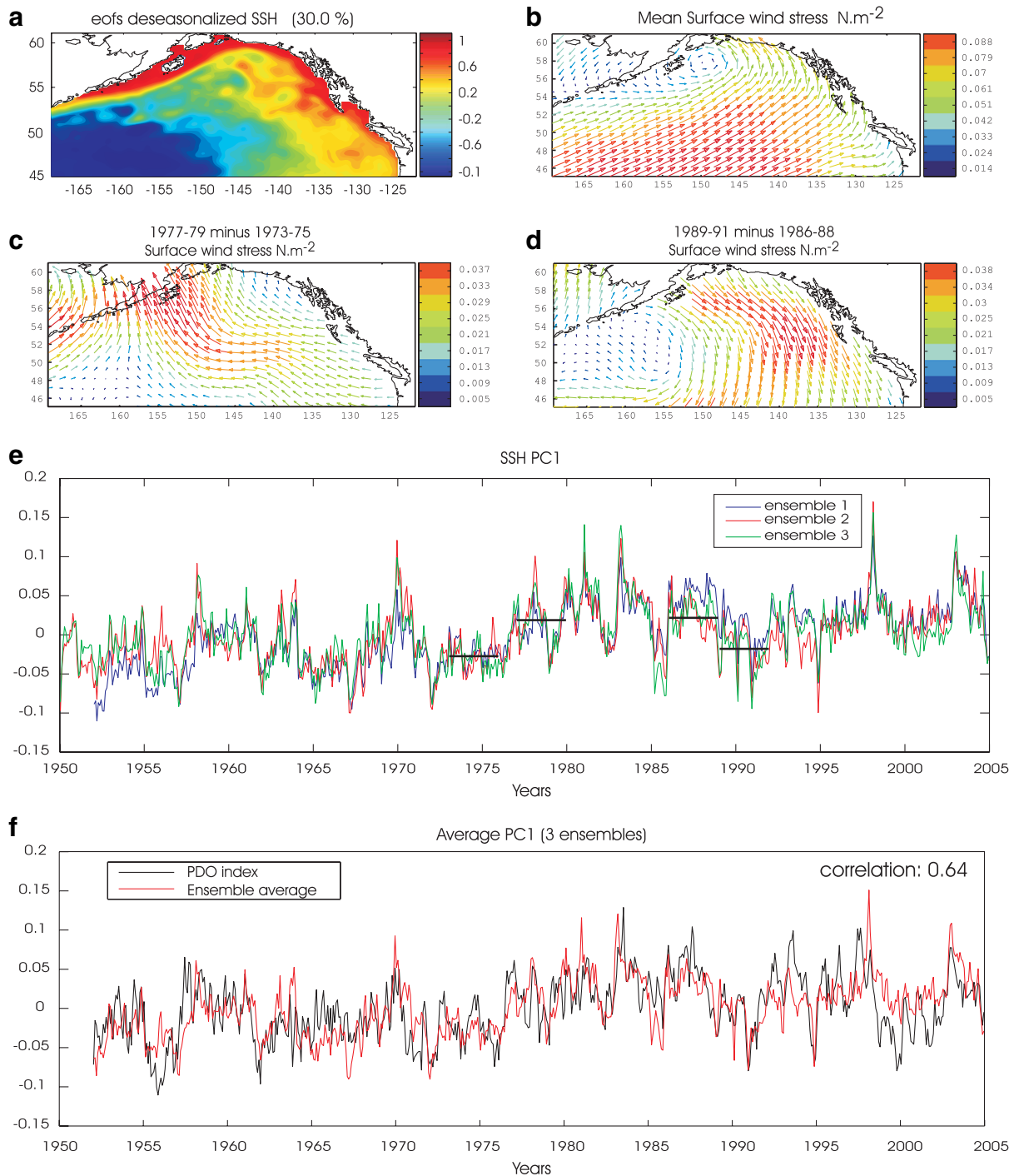


Fig. 3. (a) First empirical orthogonal function (EOF 1) of the sea surface height after removing the seasonal cycle. EOF1 explains 30.0% of the total variance. (b) Mean surface NCEP wind stress. (c) and (d) show the difference in surface wind stress for selected periods (black horizontal lines in (e)). (e) PC 1 for the three forced model ensembles. (f) Comparison between PDO index and PC1 ensemble average.

anomalous northwesterly. A detailed analysis of the model momentum and tracer budgets reveals that these extrema in the mean circulation are indeed forced by changes in the surface wind stresses (Fig. 4 illustrates an example for the 1976 shift). In particular we find a strengthening of the gyre when the Ekman transport tends

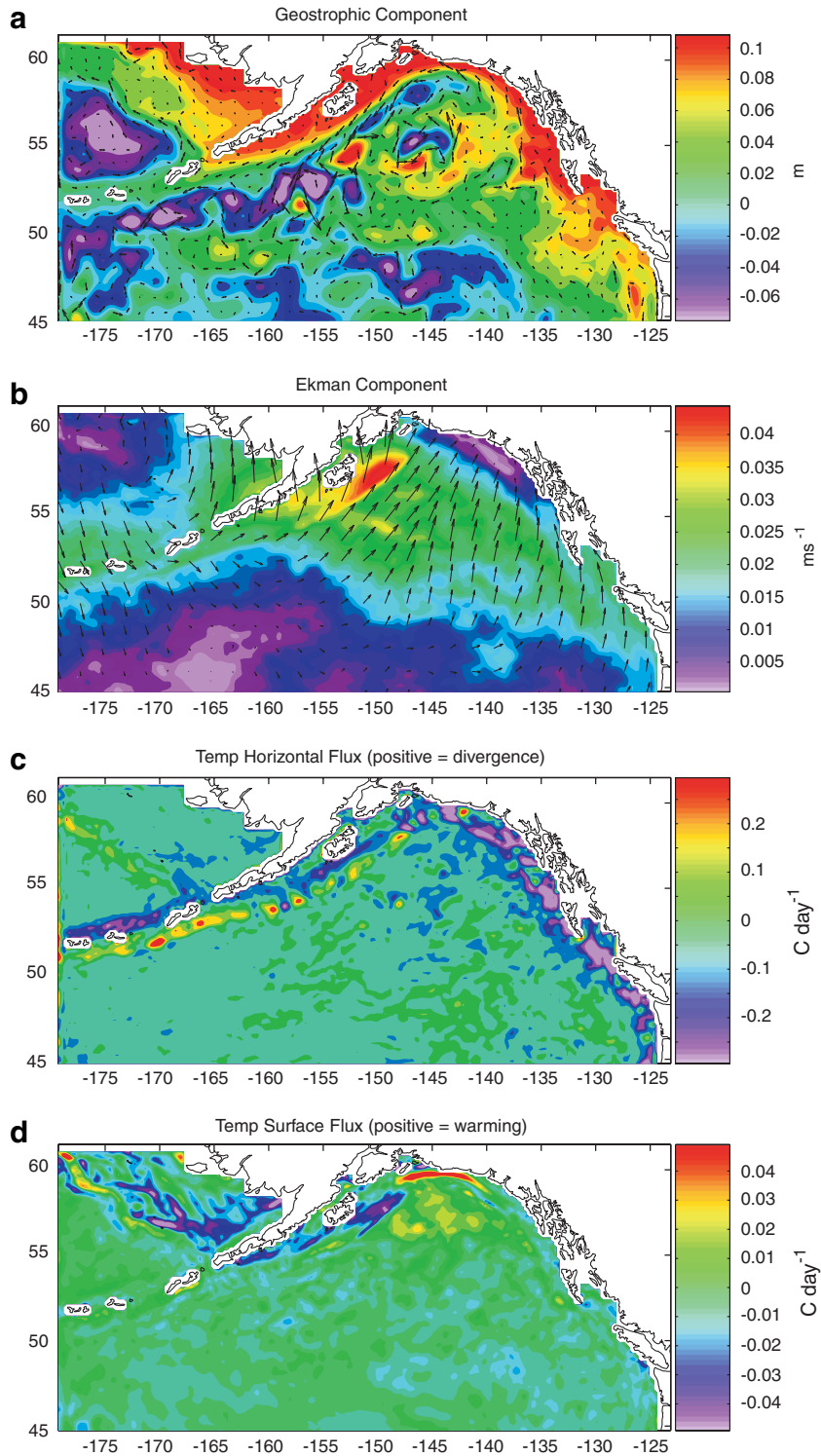


Fig. 4. Difference maps for the periods (1977–1980) minus (1970–1975). (a) Difference in SSH and Geostrophic surface currents. (b) Difference in Ekman transport in the model surface layer. (c) Difference in temperature horizontal flux, negative regions indicate stronger downwelling conditions after 1976. (d) Difference in temperature surface flux, positive indicates that the ocean is warming due to the atmospheric heat flux.



to converge mass towards the coastal boundary in the central Gulf of Alaska (Fig. 4b). This condition is also associated with stronger than usual downwelling along the eastern boundary (Fig. 4c) and upwelling along the Alaskan Stream. The contribution of surface heat fluxes is small compared to the mechanical forcing.

The adjustment process of the gyre scale circulation in response to the interannual and decadal changes of the PDO and ENSO can be summarized as follows. When the GOA experiences stronger anomalous southeasterly winds (positive phase of the PDO as well as ENSO) there is net convergence of surface mass (downwelling conditions) in the central and eastern coastal boundary induced by Ekman transport (Fig. 4b). This generates higher coastal SSHa (Fig. 4a) and lowers the coastal isopycnals. The resulting geostrophically adjusted currents (Fig. 4a) intensify the gyre scale circulation. The opposite situation occurs during stronger anomalous northwesterly winds (negative phase of the PDO). In the next section we continue to explore the ocean adjustment process to changes in the forcing by investigating how the changes in the winds are reflected in the eddy scale circulation.

#### 4. The seasonal and interannual eddy field

To gain confidence in the degree of realism of the model simulation, we compare the model seasonal SSH anomalies to the one derived from the AVISO satellite maps (Fig. 5). These maps (<http://www.jason.oceanobs.com>) merge data from T/P or JASON-1+ERS-1/2 and Envisat satellites to produce sea level anomalies at weekly resolution from October 1992 to January 2005 on a  $1/3^\circ \times 1/3^\circ$  Mercator grid. An annual oscillation of the SSHa along the coast is evident in both the model and satellite data. Consistent with Strub and James (2002) observational results, the SSHa along the coast is positive during the winter months and negative in summer, leading to a strong SSHa gradient (stronger downwelling current) in winter and weak SSH gradient (weak current) in summer. From the analysis of the model budget terms, we find that this oscillation is driven from the seasonality in the wind stress and consequent Ekman currents. Following the winter downwelling conditions along the coast in the eastern basin, a clear westward propagation of the SSHa is evident in both model and satellite during March through July. As the positive SSHa detaches from the coast a rich eddy field develops off the shelf. Anticyclonic eddies become a prominent feature, particularly in the model where the signature of the seasonal development of the Haïda and Sitka eddies is stronger when compared to satellite observations. In the western basin we also find a stronger mesoscale signature in July and August. Although both model and AVISO seasonal SSHa maps use the same time range, from January 1993 to December 2004, the satellite maps reveal an overall higher SSH anomalies in the gyre interior between August and October and lower between January and May.

The seasonal propagation of eastern eddy field is also reflected in the eddy kinetic energy (EKE) phase map (Fig. 6b). Along the coast the seasonal maxima is reached in January when the stronger current develop on the shelf in response to the stronger downwelling conditions. During late spring/summer the EKE peak is found further away from the coast in the eastern basin and reflects the seasonal development of the eddy field. During the fall the seasonal peak moves further offshore in the gyre interior. The amplitude of the seasonal EKE is strongest on the shelf where the currents are directly forced by the winds (Fig. 6a).

##### 4.1. Forced regime in the Alaska current region

To better isolate the development of the large anticyclone along the eastern boundary we perform an analysis of the model SSHa EOFs 1 and 2 along the east coast (Fig. 7). Consistent with the seasonal maps, the EOFs shows a spatially coherent excitation of SSHa along the coast (EOF 1) followed by offshore propagation (EOF2). The major anticyclonic eddies in EOF 2, Haïda and Sitka, are captured by the model and exhibit substantial interannual variability as evident by the corresponding PCs (Fig. 7c). Though the two EOFs are uncorrelated, a 1 year low pass filter of PC1 and PC2 (Fig. 7c) shows a 200 days lag average, which confirms that the development of the two anticyclonic cores in EOF 2 (corresponding to Sitka and Haïda eddies) follow the SSHa of EOF 1. The two EOFs explain more than 54% of the variance. The same EOF analysis is performed for the specific region around the Queen Charlotte Islands, where Haïda eddies are generated (figures not shown here). From this analysis we infer the speed of eddy propagation of  $1.3 \text{ km day}^{-1}$ , which is consistent with previous estimates of  $1.12 \pm 0.35 \text{ km day}^{-1}$  by Gower and Tabata (1993). The PCs show



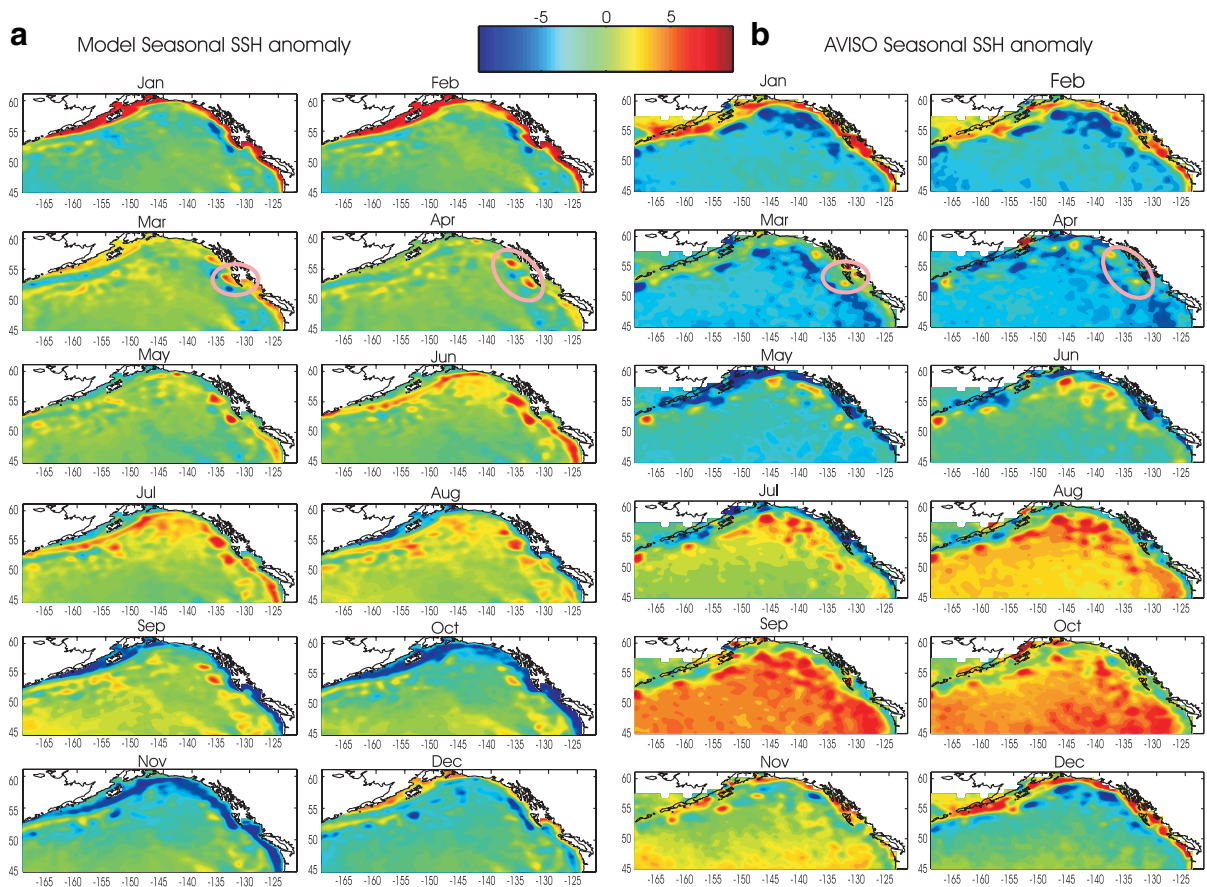


Fig. 5. Model (a) and AVISO (b) SSH seasonal anomalies derived from 1993 to 2004.

that some of the largest events over the entire record correspond to strong El Niño years, 1982–1983 and 1997–1998. A table showing the year and month when large eddy events are found (Fig. 7d) reveals that large Haïda and Sitka eddies are mainly generated during winter months (December, January, February, March) and are phase locked with the seasonal cycle. Therefore, interannual changes in the amplitude of the generated eddy are reflected in interannual modulation of the seasonal cycle.

To further clarify the relationship between the modeled eddies and El Niño years, we perform the same EOF analysis for the three model ensembles and isolate the forcing patterns corresponding to large eddy events (Fig. 8). We find that PCs 1 of all ensembles are correlated above 0.9 (Fig. 8b and h) corroborating the hypothesis of a strongly forced regime. For the Haïda region, a composite analysis of the winter wind patterns during times when PC1 is greater than 1 and less than  $-1$ , shows that the winds are predominantly southwesterly (Fig. 8c and d). However if we take the differences between the composites and the mean winds we find that during times of higher amplitude eddies ( $PC\ 1 \geq 1$ ) the winds are strongly downwelling (Fig. 8e) and therefore capable of exciting stronger SSHa. In contrast during times of weaker amplitude eddies the winds have upwelling favorable anomalies (Fig. 8f). The same analysis performed for the Sitka region leads to the same results (Fig. 8g–i), mainly that stronger eddies are excited by stronger winter downwelling winds. An inspection of the composite map of the winter wind stress anomalies during the 1982 and 1997 El Niños (Fig. 9) confirm the existence of stronger downwelling winds, therefore explaining the observed correlation between El Niño and large eddy events. In contrast the La-Niña events of 1970 and 1988 show upwelling favorable winds and reduced eddy amplitude.

Excitation of positive SSHa along the coastal boundary can also be attributed to coastally trapped Kelvin waves of equatorial origin. Indeed Meyers and Basu (1999) indicate that Kelvin waves are a source of

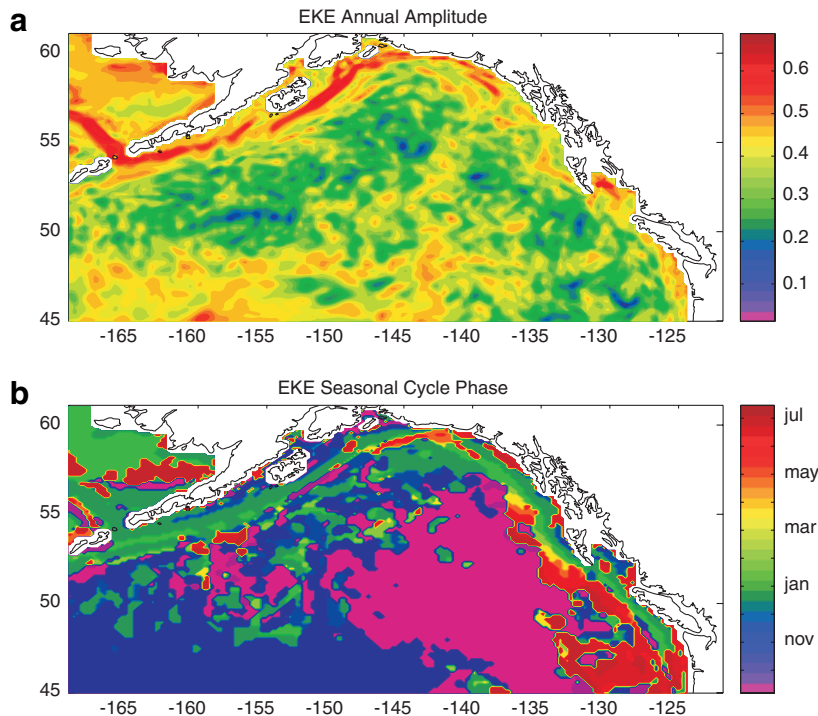


Fig. 6. (a) Annual amplitude of the eddy kinetic energy (EKE) and (b) EKE seasonal cycle phase.

interannual variability in the coastal current along western North America. However, in this study we used a monthly mean climatology southern open boundary condition from [Levitus et al. \(1994\)](#), which does not include these tropically generated Kelvin-like waves. Nevertheless a comparison of the model coastal sea level with in situ records and satellite data in various coastal locations (Sitka, Neah Bay, Seward, Yakutat, Kodiak and Sand Point; [Fig. 10](#)) indicate that the model hindcast is able to explain a very large fraction of the in situ sea level variance. For example at Sitka the model explains 41% of the in situ observations variance, which is more than the satellite (21%). At other locations, except Yakutat, the model generally compares more favorably than the satellite with in situ data explaining between 40% and 70% of the tidal gauges variance. The poor performance of the satellite data in the nearshore region has been reported before by [Strub and James \(2002\)](#) who performed analysis of coastal tide gauge and altimeter data to generate a merged sea level anomaly product for the Northeast Pacific. At Yakutat the correlation is not very high as the model does not reproduce Yakutat eddies very well. High correlations with in situ data are also found at coastal locations along the Alaskan Stream. From these results, which are consistent with [Strub and James \(2002\)](#) observational results, we conclude that most of the interannual variability in the eastern boundary eddies and over the shelf, is locally forced by the winds, although coastally trapped waves may still contribute to additional variance.

We now discuss a dynamical framework to understand the relationship between the forcing and the development of large scale eddies in the eastern boundary. Previous studies suggest that the seasonal and interannual development of the mesoscale eddy field results from instability associated with the seasonal mean coastal currents and buoyancy gradients ([Thomson and Gower, 1998](#); [Melsom et al., 1999](#); [Okkonen et al., 2001](#); [Murray et al., 2001](#)). However other studies suggest alternative generating mechanisms that rely on linear dynamics ([Di Lorenzo et al., 2005a](#)). For this reason we attempt to reframe the current and previous findings in the context of instability of linear Rossby waves. This follows from the observation that EOF 1 and EOF 2 ([Fig. 7](#)) are consistent with a linear excitation and propagation of a Rossby wave. [LaCasce and Pedlosky \(2004\)](#) show using an idealized setup, that Rossby waves excited at the eastern coastal boundary are subject to baroclinic instability, in particular when the growth rate of the instability is fast compared to the time it takes for the wave to cross the ocean basin. Because the phase speed of Rossby waves decreases with

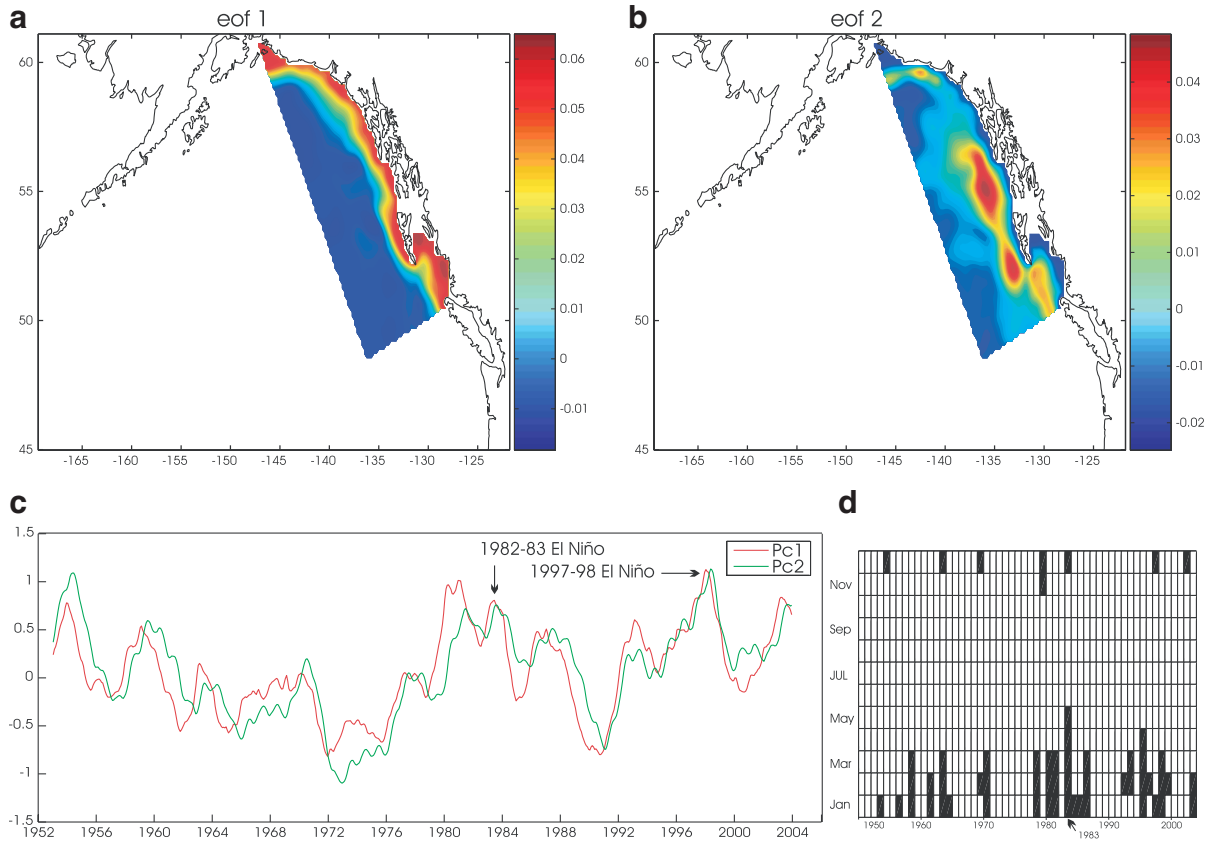


Fig. 7. (a) EOF 1 and (b) EOF 2 of SSHa. (c) The red curve (green curve) is the principal component associated to EOF1 (EOF2), using a 1 yr lowpass filter. (d) Matrix showing the times when large Haida and Sitka eddies (EOF2) are formed in the model simulation.

increasing latitude, first mode baroclinic Rossby wave phase speeds are extremely slow in the Gulf of Alaska (about 1–2 km/day from Chelton and Schlax (1996)). Indeed during winter, following the strong downwelling condition, the linear adjustment process to the positive sea level perturbations (seasonal in Fig. 5a, interannual in Fig. 7) leads to radiation of Rossby waves that would take about 6 months to propagate away from the coastal boundary where they are excited. This timescale agrees with the observed and modeled SSHa propagation away from the east coast during winter and spring (Fig. 7). However, LaCasce and Pedlosky (2004) theory predicts that the excited Rossby waves become unstable at higher latitudes and will not have time to coherently propagate in the gyre interior. Both model and observations (Chelton and Schlax, 1996) support this view and do not show a meridionally coherent westward phase propagation.

To verify if the LaCasce and Pedlosky (2004) theory is applicable for the Gulf of Alaska, we compare the propagation timescale of the waves with the typical growth rate of baroclinic instability in the Gulf as inferred by the Eady index (Lindzen and Farrell, 1980) defined:

$$\sigma_E = 0.31 \frac{f}{N} \frac{du}{dz}$$

where  $f$  is the Coriolis parameter ( $s^{-1}$ ),  $N^2 = -\frac{g}{\rho_0} \frac{d\rho}{dz}$  represents the static stability ( $s^{-2}$ ) and  $\rho$  the density of water ( $kg\ m^{-3}$ ). In the computation, we consider  $u$  as the current speed. The Eady index,  $\sigma_E$ , is then integrated between the surface and 100 m meters depth. Although  $\sigma_E$  describes the baroclinic instability of a zonal basic-state flow (Hoskins and Valdes, 1990) it still provides a rough estimate of the typical growth rate of the instabilities and is similar to the criteria used by LaCasce and Pedlosky (2004) which use the ratio between vertical shear and Rossby deformation radius. Fig. 11 shows the model monthly mean maps of the Eady index, which exhibit a clear

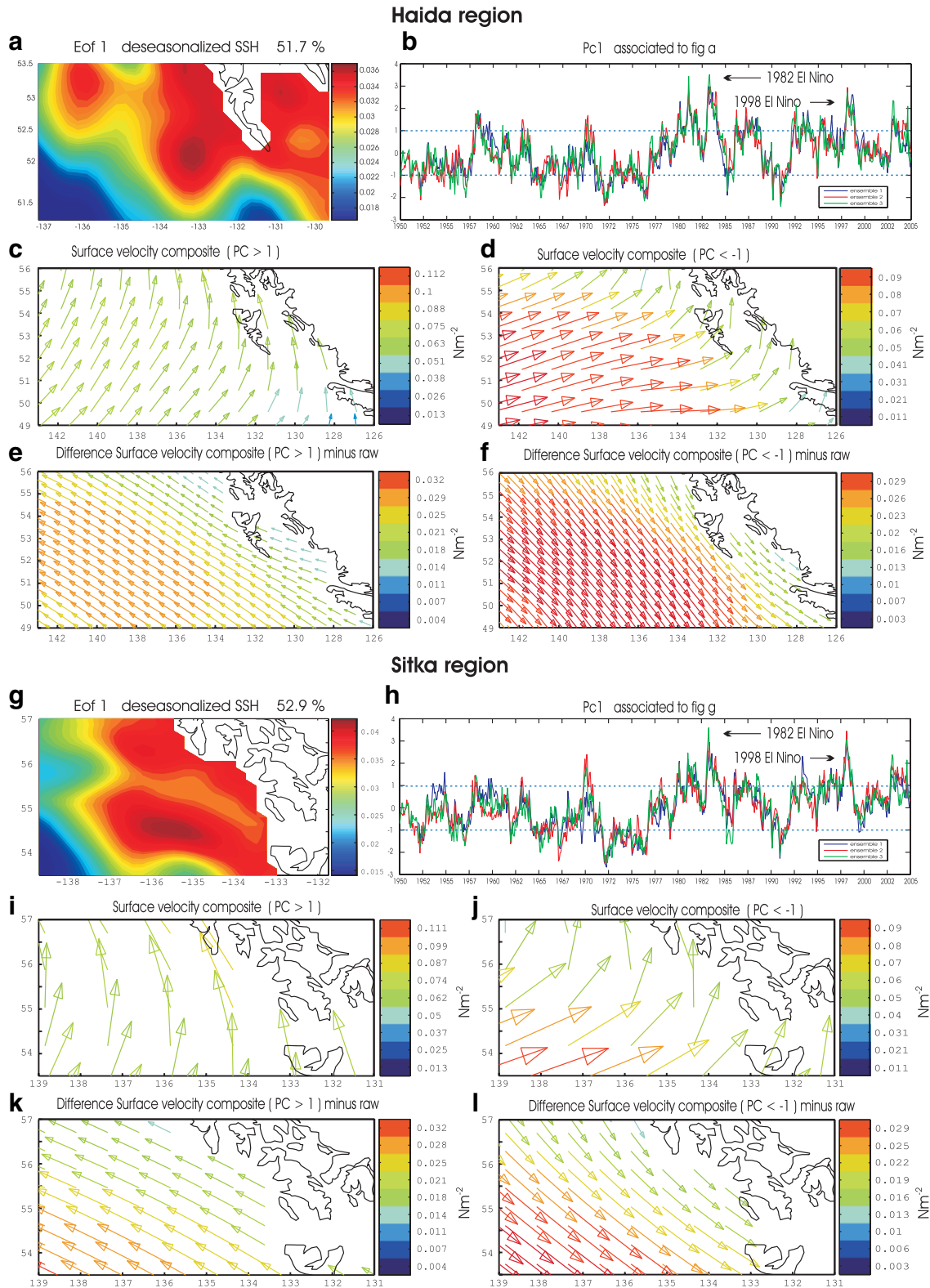


Fig. 8. (a) EOF 1 and (b) PC 1 of SSHa in the region of formation of Haida eddies. (c) Wind stress composite maps for PC 1 > 1, (d) for PC 1 < -1. (e) Wind stress anomaly composite maps for PC 1 > 1, (f) for PC 1 < -1. (g-l) same as (a-f) but for Sitka region.



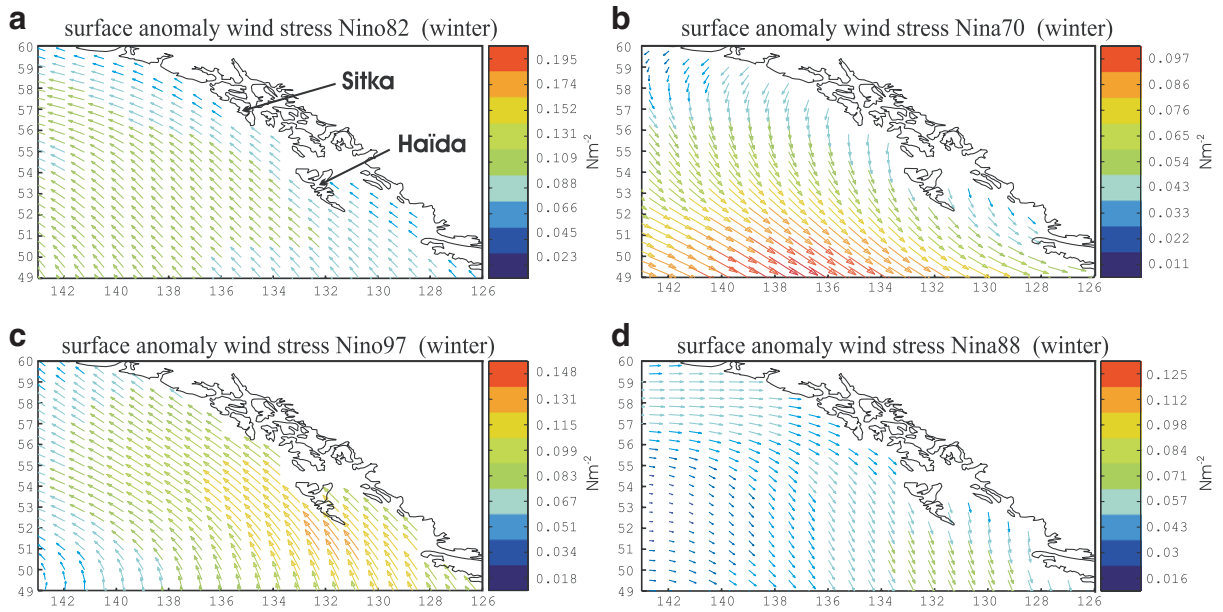


Fig. 9. NCEP surface wind stress anomaly (with respect to the 1950–2004 period) for the El Niño winters (a) 1982–1983, (c) 1997–1998 and (b) La Niña winters 1970–1971, (d) 1988–1989.

seasonal cycle with higher baroclinicity during winter. Between October and April along the coast where the SSHa is perturbed by the seasonal peak of downwelling winds, the Eady index has values between 4 and  $7 \times 10^{-6} \text{ s}^{-1}$ . The corresponding timescale for the growth rate of the instability is 2–6 days, which is much too fast for the coastal disturbances to radiate away as Rossby waves.

The reoccurrence of large anticyclones at the same regional locations (e.g. Haïda, Sitka and Yakutat) is not explicitly accounted for by the LaCasce and Pedlosky (2004) theory. Nevertheless it is likely that during the linear phase of the ocean adjustment process to the wind induced isopycnal perturbations (EOF 1 and 2 Fig. 7), the positive SSHa pattern is strongly influenced by the coastal and topographic geometry. This geographical focusing combined with the LaCasce and Pedlosky (2004) theory could explain why the amplitude of the eastern basin eddy field is proportional to the perturbation energy of the downwelling winds. Specifically during downwelling conditions in winter, the early response of the coastal ocean is linear with the onset of stronger poleward currents associated with the high coastal SSHa, and with the initial radiation of Rossby and topographically trapped waves. However because of the fast growth rate of the instability and the geographical focusing, most of the perturbation energy is locally trapped until converted into the eddy field. Although more detailed studies are needed, this energy trapping mechanism would explain the strong relationship between the perturbation amplitude of the winds and eddies.

#### 4.2. Intrinsic regime in the Alaskan stream region

While the development of mesoscale eddies on the eastern GOA basin is modulated by changes in the wind stress, eddies along the Alaskan Stream (western basin) do not correlate with atmospheric forcing and their variability may be intrinsic (Okkonen et al., 2001). To separate the intrinsic and forced variability in the GOA we conduct an additional long-term (300 years) model run forced by climatological monthly mean surface fluxes (momentum, heat and freshwater; Experiment 2; Table 1; “Unforced run”). We then separate the SSH variance of the “Forced Run” and “Unforced Run” into seasonal (defined as all frequencies  $\geq$  seasonal) and interannual (all frequencies lower than seasonal) (Fig. 12a, b, d and e). First inspection of Fig. 12a–c shows that both runs exhibit similar seasonal variances with higher values along the coast where the seasonal winds drive the coastal current. A map of the difference in seasonal variances (Fig. 12c) shows that the



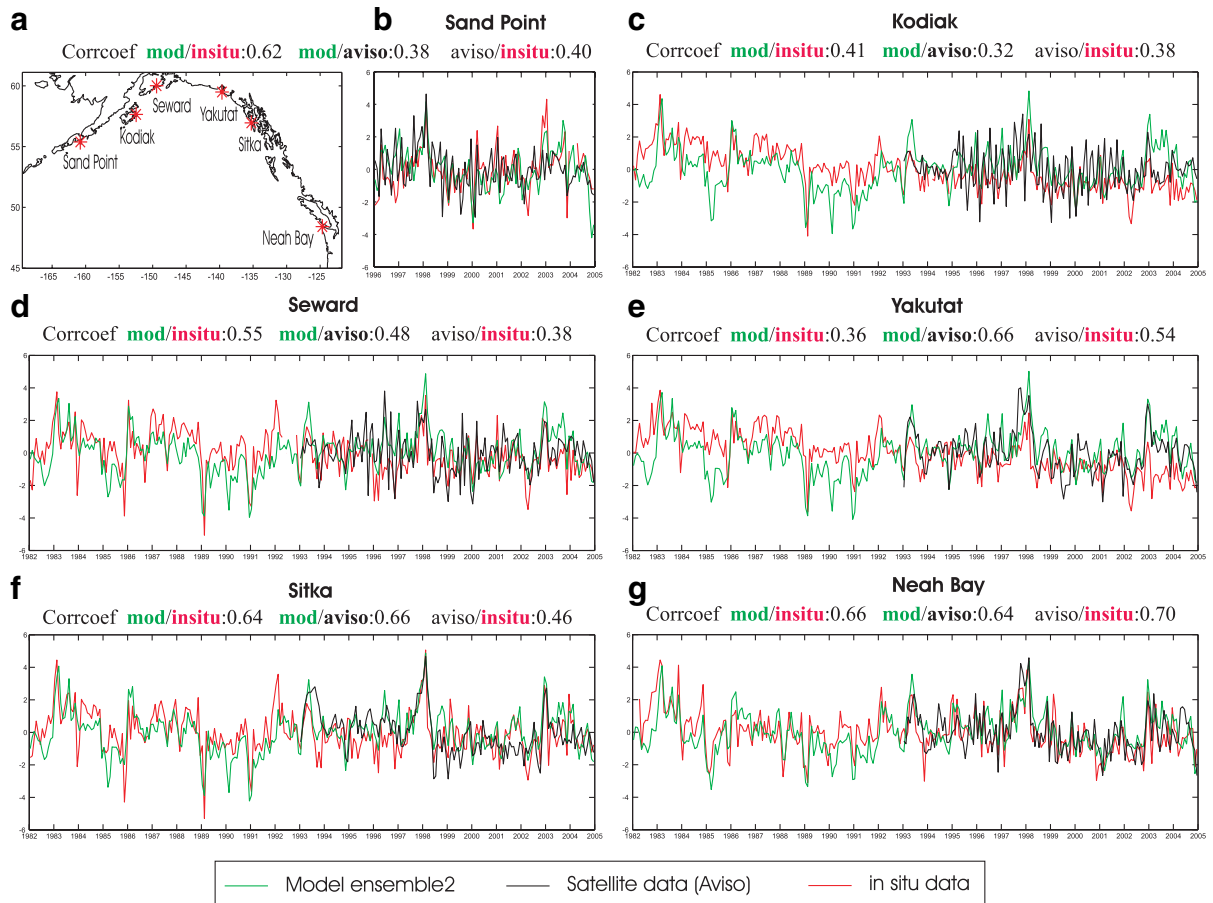


Fig. 10. Comparison of the coastal deseasonalized SSH for Sand point, Kodiak, Seward, Yakutat, Sitka, and Neah Bay between the model output (green line), AVISO/TOPEX (black line) and GLOSS “fast delivery” data (red).

“Unforced Run” has higher variance at the generation sites of Haïda and Sitka eddies. This is expected because the large interannual eddy events in these regions are phase locked with the seasonal cycle (see previous discussion of Fig. 7), therefore some of the seasonal variance is shifted to the lower frequencies in the “Forced Run”. If we now compare the maps for the interannual variance we find that the “Unforced Run” (Fig. 12e) has strong variance in regions of high eddy activity in both the western and eastern basin. Since this model run has no forcing on periodicities higher than seasonal, regions of high variance must be associated with internal ocean (intrinsic) variability. To assess the degree of intrinsic variance we take the difference in variance between the Forced and Unforced (Fig. 12f). Three important features can be isolated. First we observe that in the western basin (Fig. 12f) there is no significant difference in interannual variance between the Forced and Unforced run. This confirms that most of the variance in the region of high eddy activity along the Alaskan Stream is intrinsic in nature. A second important feature in the difference map is in the shelf region along the western basin coastline. Here the “Unforced Run” has much less variance than the “Forced Run”. This confirms that interannual changes in shelf circulation are mostly forced by the winds and it explains why the model “Forced Run” ensembles are so favorably correlated with in situ tide gauges (as shown before in the discussion of Fig. 10). The third feature is in the eastern basin, where the “Forced Run” has higher variance along the eastern boundary. This result agrees with the findings of the previous section showing that interannual changes in the winds force large anticyclones eddy such as Haïda and Sitka.

To further isolate the intrinsic and forced variability we compare timeseries from all Forced model ensembles in four selected regions of eddy activity, two in the western basin and two in the eastern basin (Fig. 13a, c,

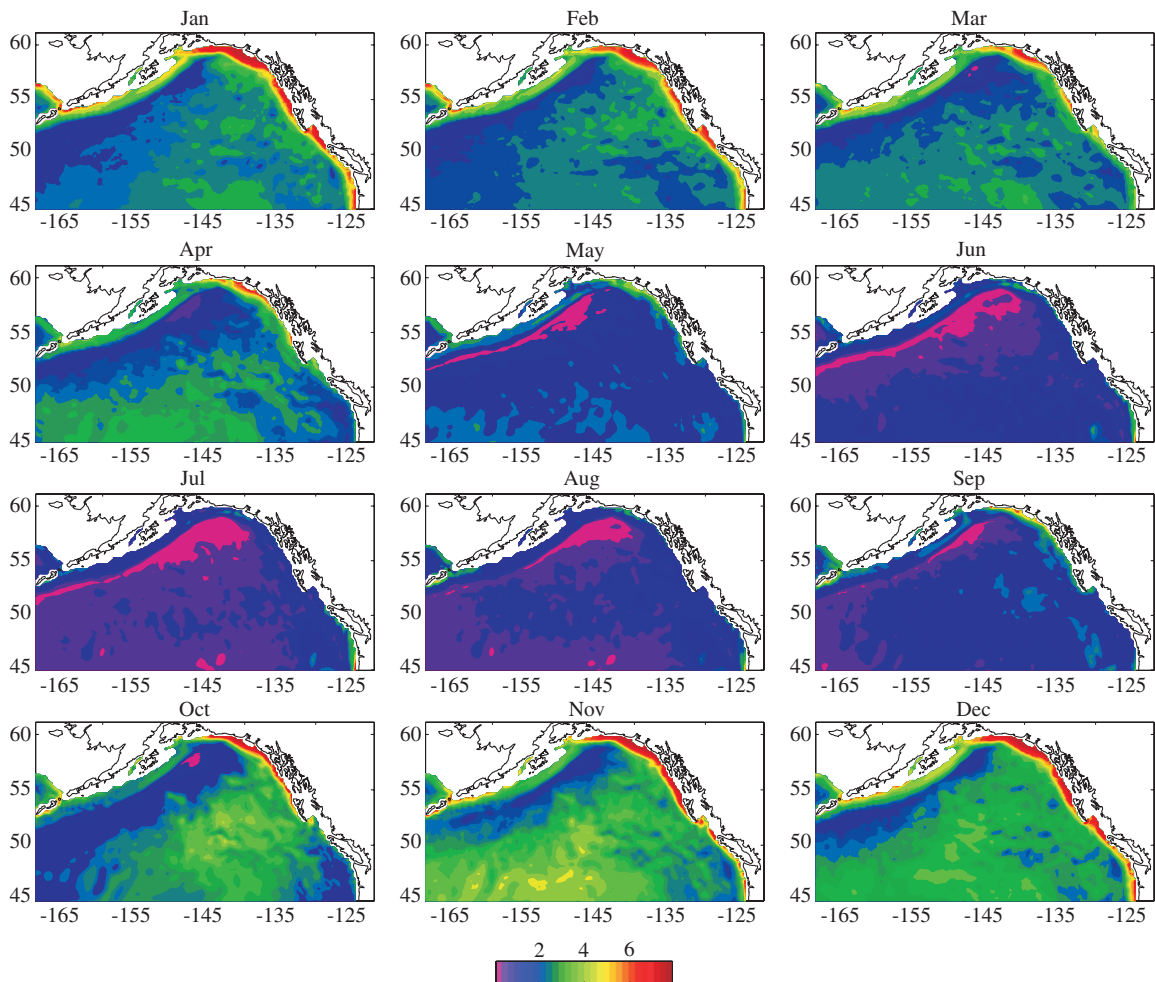


Fig. 11. Seasonal variability of the Eady index, units in  $10^{-6} \text{ s}^{-1}$ .

e and g). These timeseries show that in the eastern basin (Fig. 13a and c) the spread of the ensemble members is very low with shared variance above 42%. This confirms again the existence of a forced regime. In contrast in the western basin the ensemble members share little variance (Fig. 13e and g) and are mostly uncorrelated with each other. Some low frequency variability after 1990 is shared by all the ensemble members in the western basin, with an indication of a negative trend, however the amount of explained variance by this component is small. Interestingly a comparison with satellite data (not shown) reveals a similar negative trend over the last 10 years, however we did not pursue this aspect further as the temporal data coverage is short.

Although eddy variability in the western basin is intrinsic, one may expect that changes in variance from year to year can be related to changes in the mean ocean conditions or forcing function. To explore this concept we compare the 4-year running standard deviation for all ensembles (Fig. 13, right column). On the eastern boundary inspection of the timeseries show a strong degree of correlation among all ensembles implying that interannual changes in eddy variance are also forced. In contrast in the western basin the spread among the ensemble members is high, although there are periods in the record when the ensembles are closer to each other. This reinforces the idea that changes in the forcing have a weak impact in modulating the eddy variance, along the Alaskan Stream, on interannual timescales.

We also investigate if ENSO teleconnected atmospheric forcing may play any role in changing the conditions for baroclinic instabilities. We quantify the ENSO effect by correlating the instability growth rate (the

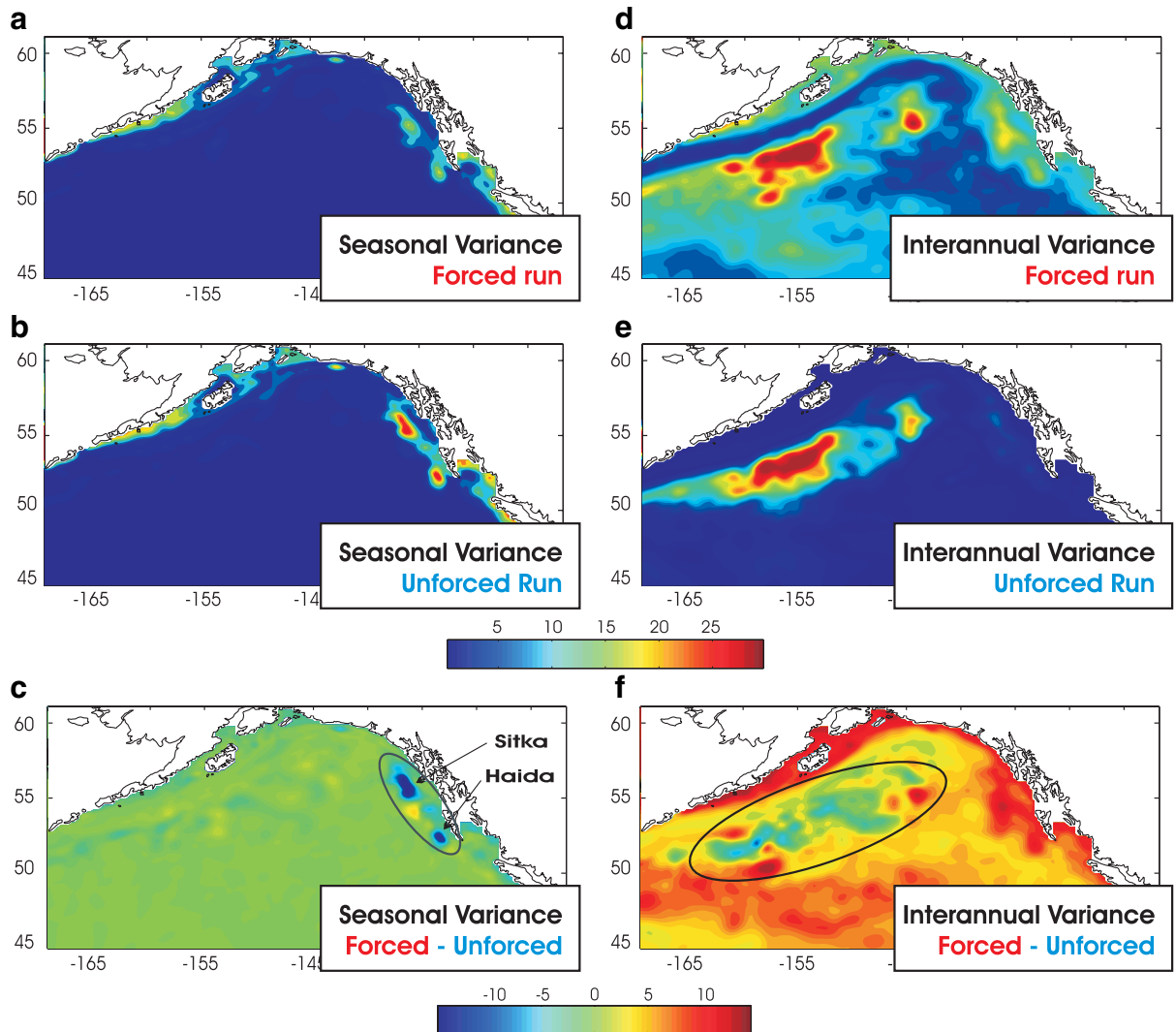


Fig. 12. Model SSH variances ( $\text{m}^2$ ) for the run forced by monthly NCEP wind stress (first row) and for the run forced by climatology monthly wind stress (second row). The difference between the two runs is shown in the 3rd row. The first (second) column shows the seasonal (interannual) variance.

Eady Index) with the NINO34 index (Fig. 14). The spatial pattern of the correlation shows that along the eastern boundary and northern Gulf of Alaska, an increase in the instability growth rate is positively correlated with positive ENSO phases. In contrast along the Alaskan stream in the region of high SSHa eddy variance, and in the gyre interior, we find a negative correlation indicating that during El Niño the growth rate is reduced. However the amount of negative correlation is insignificant, which agrees with the hypothesis of the intrinsic nature of eddy variability in the western basin.

## 5. Summary and conclusions

The adjustments of the Gulf of Alaska circulation to changes in large scale climate forcing has been investigated using three high resolution ensembles of a regional ocean model forced with the 1950–2004 NCEP/NCAR winds and NOAA SST corrected heat fluxes. Consistent with previous descriptions of the circulation we find that the mean currents are mainly geostrophic (Fig. 2a) and maintained by Ekman downwelling

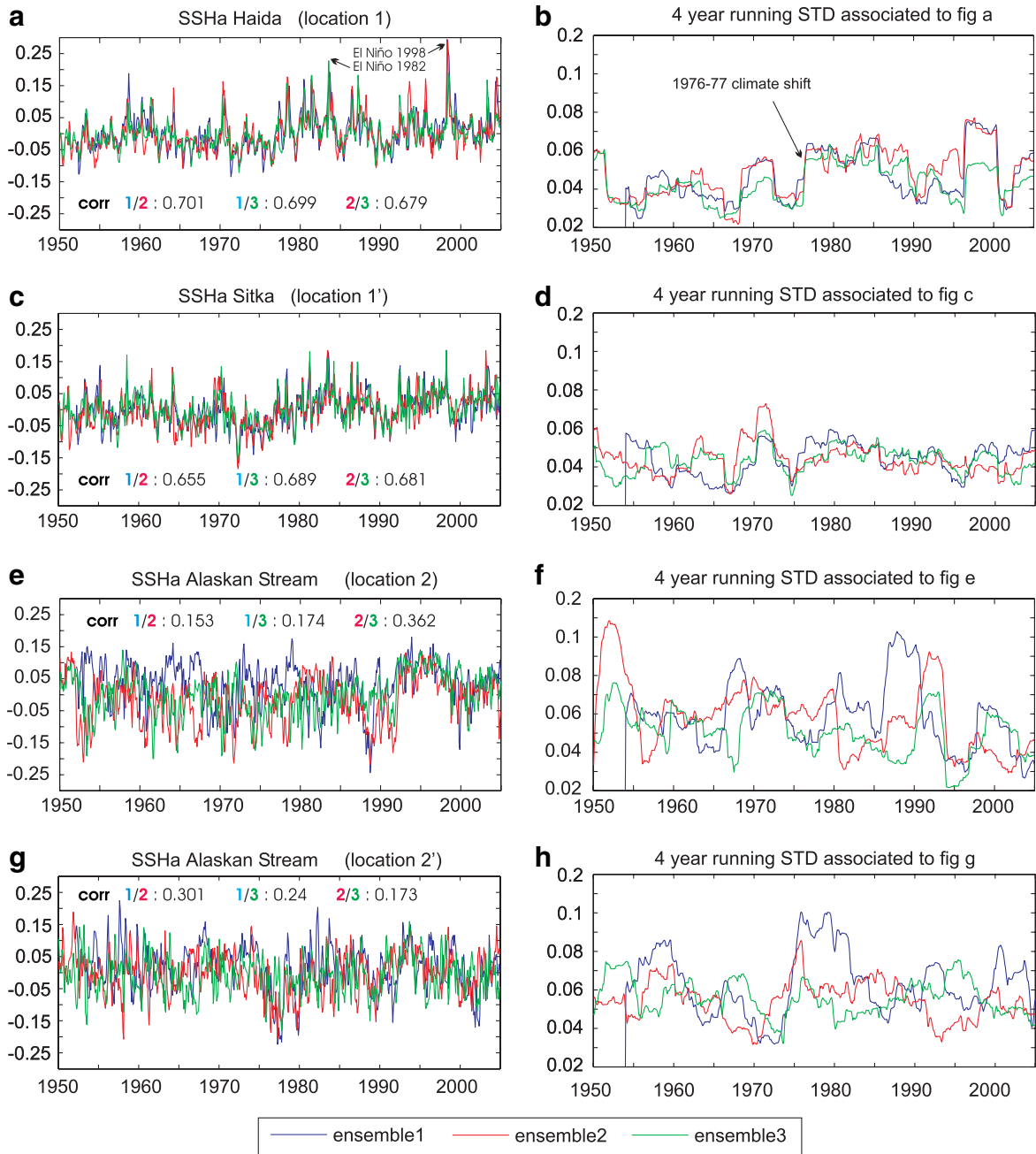


Fig. 13. For the three ensembles (1, 2, 3), (a), (c), (e) and (g) show the SSHa respectively in the location 1, 1', 2 and 2' represented on Fig. 1c. (b), (d), (f) and (h) show the 4 year running standard deviation associated to (a), (c), (e) and (g). (corr = correlation coefficient).

conditions in the eastern basin and upwelling along the Alaskan Stream (Fig. 2b and c). On interannual and decadal timescale the mean circulation is strongly modulated by changes in the large scale climate forcing associated with PDO and ENSO variability (Fig. 3). The leading mode of spatial and temporal variability of the mean currents in the three model ensembles is significantly correlated with the PDO (0.64) and shows strong peaks during El Niño (e.g. 1997–1998). Intensification of the gyre scale circulation is found after the 1976–1977 climate shift, as well as during 1955–1960, 1965–1970 and more gradually during 1993–1995. A

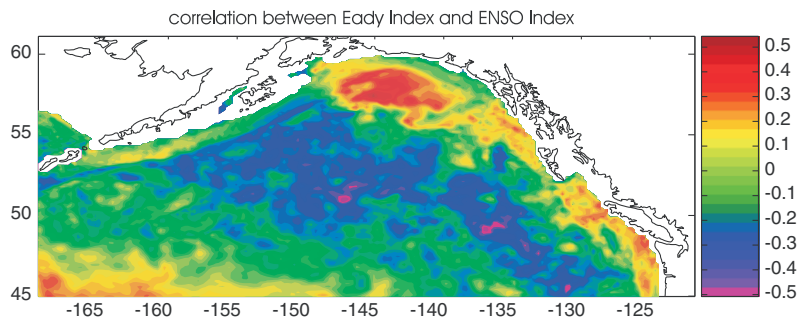


Fig. 14. Map of the correlation between Eady index and ENSO index.

detailed analysis of the model momentum and tracer budget reveals that when the GOA experiences strong anomalous southeasterly winds (positive phase of the PDO after the 1976 shift and the 1997 ENSO) there is net Ekman convergence of surface mass (downwelling conditions) in the central and eastern coastal boundary (Fig. 4b). This generates higher SSHa (Fig. 4a) and lower isopycnals along the coast. The resulting geostrophically adjusted currents (Fig. 4a) intensify the gyre scale circulation. The opposite situation occurs during strong anomalous northwesterly winds (negative phase of the PDO).

Interannual changes in the surface winds are also reflected in modulations of the seasonal development of high amplitude anticyclonic eddies along the eastern basin, in particular at the location of the Haïda and Sitka eddies. The generation of the large interannual eddy events during winter-spring, is phase-locked with the seasonal cycle and is consistent with a quasi-linear Rossby wave response to stronger winter southwesterly winds typical of El Niño (Figs. 7 and 8). The high correlation between the amplitude of the eastern basin eddy and the perturbation energy of the downwelling winds is explained by the combined effect of geographical focusing and instability of Rossby waves at higher latitudes (LaCasce and Pedlosky, 2004). Specifically during downwelling anomaly conditions, the early response of the coastal ocean is linear with the onset of stronger poleward currents associated with the high coastal SSHa, and with the initial radiation of Rossby and topographically trapped waves (Fig. 7). However because of the fast growth rate of the instability (Eady index maps Fig. 11) and the geographical focusing associated with the coastal geometry, most of the perturbation energy in the Rossby wave response is locally trapped until converted into the eddy field.

In the western basin, the high core of interannual eddy variability south of the Alaskan Stream (Fig. 12d) is not correlated with large scale forcing and appears to be intrinsic. Indeed a comparison of the forced and internal interannual variability from the three forced ensembles and a multi-century-long integration forced only with the seasonal cycle of winds, heat and freshwater, shows that the intrinsic variance alone explains most of the eddy variability (Fig. 12f). The asymmetry between the eddy forced regime in the eastern basin and the intrinsic regime in the western basin is also confirmed by timeseries of the ensemble members SSHa. These show that along the eastern boundary the spread is very low with shared variance above 42% (forced regime). In contrast in the west, the ensemble members share little variance (Fig. 13e and g) and are mostly uncorrelated (intrinsic regime), although some low frequency co-variability is evident after 1990.

We also explored the possibility that ENSO atmospheric teleconnections may impact the degree of baroclinicity in the western basin. However a spatial correlation analysis between the Eady index, which is a measure of the instability growth rate, and NINO34 index shows high correlations only along the eastern boundary and northern Gulf of Alaska but not in the centers of high eddy activity in the western basin.

Although the eddy variability in the western basin is intrinsic, interannual changes in the shelf circulation are forced by the winds (Fig. 12f). A comparison of SSHa between tidal gauges and the three model ensembles forced with the NCEP winds, reveals that the model is able to explain between 40% and 70% of tidal gauges variance. This finding suggests that most of the shelf variability on interannual scale does not strongly depend on equatorially trapped waves, but rather on downwelling (upwelling) conditions associated with the large scale wind stress patterns.

This study provides a dynamical framework to interpret and predict how the GOA circulation has and will respond to large scale climate forcing. The model analyses suggest that the adjustment to large scale climate



forcing of the coastal and gyre-scale mean circulations is predictable. On decadal and interannual timescale the most important aspect of the forcing is the overall strength of downwelling conditions along the coast, although the role of freshwater forcing has not been fully investigated. River runoff along the GOA is known to contribute to the cross-shore buoyancy gradients, which in turn strengthen the shelf circulation. Therefore decadal variation in river runoff may also lead to additional variability in the currents and their stability properties. This study also supports the idea of an east-west asymmetry in the response of the mesoscale eddy circulations. In particular eddy activity in the eastern basin is strongly correlated with changes in downwelling favorable winds, while in the western basin the variability is mainly intrinsic. The link between stronger eddy in the eastern basin and downwelling winds has important climate implications. It allows predicting changes in the cross-shore transport of important biogeochemical quantities such as iron, oxygen and chlorophyll-a, which in a global climate change scenario may contribute to feedbacks in the HNLC region of the GOA. Within the dynamical and modeling framework presented in this study, further analyses with higher resolution nested models will better resolve the mixing statistics associated with the eddy field in eastern and western basin. Model representations of the GOA circulation are in fact very sensitive to the coastal and topographic geometry. As such the model results presented so far should be interpreted more qualitatively rather than quantitatively.

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