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Investigating Bimodal Variability of the Kuroshio with High Resolution Climate Model

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Abstract

The Kuroshio current, the western boundary current of the North Pacific subtropical gyre, exhibits bimodal variability on interannual to decadal timescales characterized by two states; a stable, non-large meander, and an unstable, large meander. This low frequency variability is a signature feature of the Kuroshio, not yet observed in other western boundary currents. The Kuroshio is one of the strongest air-sea heat loss regions in the ocean and so understanding the interannual variability of the Kuroshio is key to understanding and predicting climate in the region. There is currently no consensus about the mechanisms driving the low frequency transitions of the Kuroshio. The main question addressed in this study is whether the bimodal behaviour of the Kuroshio is a result of external time varying wind forcing, acting through a linear time-dependent Sverdrup balance, or through nonlinear processes internal to the ocean system. [Usui et al. 2013] [Knauss 2005] [Pierini et al. 2009]. We base our analysis on output from the high resolution Community Earth System Model. With a theoretical basis in the geostrophic and hydrostatic balance we compute the horizontal gradient of the mean annual sea surface height anomalies provided by the CESM and use it to determine the barotropic surface velocity in the Kuroshio region. We develope a method to track the position of the Kuroshio along the longitude where it displays the most significant path difference between its stable and unstable state. We test the hypothesis that the variability of the Kuroshio is a result of external time varying wind forcing by examining the transitions of the Kuroshio with a prescribed annual atmospheric evolution in a control integration in the CESM. Our findings suggests that the bimodal variability of the Kuroshio is not a result of external time varying wind forcing. We conduct a pertubation experiment and introduce an atmospheric forcing through a change in the zonal winds in the Southern Ocean, the effect of which can only be transmitted to the Kuroshio region through internal ocean mechanics. Our results support the hypothesis that mechanisms internal to the ocean system are linked to the low frequency bimodal variability of the Kuroshio.

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2 Introduction

The Kuroshio is the western boundary current of the North Pacific subtropical gyre. The Kuroshio flows northward from the North Equitorial Current, NEC, east of the Phillipines and Taiwan, after which it strikes east through the Tokara Strait. It then flows along the south-east coast of Japan passing through the Izu-Ogasawara Ridge, a ridge almost normal to the flow and with limited passes deeper than a 1000 meters, before it joins the eastbound northern branch of the North Pacific subtropical gyre. The subject of this study is the bimodal variability on interannual to decadal timescales of the Kuroshio along its pathway between the Tokara Strait at around 131°E and the Izu-Ogasawara Ridge at 140°E. The variability of the Kuroshio is bimodal and characterized by two states; a stable, non-large meander, and an unstable, large meander. The different states are illustrated in figure (1), where the major path difference between the two states is centered around 138°E. Once the Kuroshio forms a deep meander it can stay there for several years and this low frequency variability is a characteristic feature of the Kuroshio, not yet observed in other western boundary currents. In many other ways, the Kuroshio is similar to its better known North Atlantic counterpart, the Gulf stream. Western boundary currents are generally swift and narrow compared to the mid-ocean and eastern parts of the gyre circulation. The water transport of the Kuroshio ranges between 50 to 130 Sv, where $Sv = 10^6 m^3/s$, and the width of the Kuroshio is about 100 km. The Kuroshio is also one of the strongest air-sea heat loss regions with a heat flux in the range of $125W/m^2$. Understanding the interannual variability of the Kuroshio is therefore key to understanding and predicting climate in the region. Currently there is no consensus about the mechanisms driving the low frequency transitions of the Kuroshio. The main question we address in this study is whether the variability of the Kuroshio is induced by external time varying wind forcing, acting through a linear time-dependent Sverdrup balance, or through nonlinear processes internal to the ocean system. We investigate the variability of the Kuroshio with output from the high resolution Community Earth System Model. With a theoretical basis in the geostrophic and hydrostatic balance we use mean annual and monthly sea surface height anomalies to compute the barotropic surface velocity in the Kuroshio region. The barotropic velocities allow us to determine the position of the Kuroshio along the longitude where it displays major path difference. We then observe the effect of external and internal forcings on the variability of the Kuroshio through a control and pertubation integration in the CESM run. [Usui et al. 2013] [Talley 2011] [Knauss 2005] [Qiu and Chen 2005] [Pierini et al. 2009]

3 Background and Theory

The aim of this section is to develop a fundamental theory describing the large scale ocean circulation in the North Pacific subtropical gyre and its relation to the Kuroshio current and to setup the theoretical basis for our applied method.

3.1 The Momentum Equations

We consider a water parcel in the ocean along with the forces acting on it. Overall the forces that can accelerate a water parcel in the ocean are, the Coriolis force, the pressure gradient force, the gravitational force and frictional forces.

We describe the fluid flow by writing the equations of Eulerian motion for the water parcel. We define the acceleration of the flow past a fixed point as the sum of the acceleration of the water parcel itself



Figure 1: (a): SSH anomaly map from CESM control integration, year 21, showing the nonlarge meander of the south-east coast of Japan (b): SSH anomaly map from CESM control integration, year 17, showing the large meander. Note the significant path variation centered around 138°E.

and the momentum advection terms in eq. (1).

$$\frac{du}{dt} = \frac{\partial u}{\partial t} + u\frac{\partial u}{\partial x} + v\frac{\partial u}{\partial y} + w\frac{\partial u}{\partial z}$$
(1a)

$$\frac{dv}{dt} = \frac{\partial v}{\partial t} + u\frac{\partial v}{\partial x} + v\frac{\partial v}{\partial y} + w\frac{\partial v}{\partial z}$$
(1b)

$$\frac{dw}{dt} = \frac{\partial w}{\partial t} + u\frac{\partial w}{\partial x} + v\frac{\partial w}{\partial y} + w\frac{\partial w}{\partial z}$$
(1c)

Where u, v and w denote zonal, meridional and vertical velocity and where positive x-, y- and zdirection is eastward, northward, and upwards respectively. Simultaneously we also define the total derivative as $\frac{d}{dt} = \frac{\partial}{\partial t} + u \frac{\partial}{\partial x} + v \frac{\partial}{\partial y} + w \frac{\partial}{\partial z}$. [Talley 2011] We can write the momentum equations for the water parcel in the following way:

$$\frac{du}{dt} = fv - \frac{1}{\rho} \frac{\partial P}{\partial x} + \frac{1}{\rho} F_x$$
(2a)

$$\frac{dv}{dt} = -fu - \frac{1}{\rho}\frac{\partial P}{\partial y} + \frac{1}{\rho}F_y$$
(2b)

$$\frac{dw}{dt} = -\frac{1}{\rho}\frac{\partial P}{\partial z} - g \tag{2c}$$

In eq. (2) ρ is the density of the sea water, P the pressure, and the Coriolis parameter, f, is defined as $f \equiv 2\Omega sin(\vartheta)$, where $\Omega = 2\pi/24h$ is the angular velocity of the Earth and ϑ is the latitude. The Coriolis force is latitude dependent and is an effect of Earth's rotation. The water parcel it accelerated by a rotating Earth and since our reference system is fixed we need to introduce a compensating force, the Coriolis force. For many problems in physics the Coriolis force can be neglected, but in oceanography and meteorology the spatial scale of the flows are so large that the Coriolis force is a dominating force and must be taken into account. F_x and F_y are the frictional terms in the x- and

y-direction respectively. We write the friction as:

$$F = \frac{\partial \vec{\tau}}{\partial z} + \rho A_H \nabla^2 \vec{u}, \quad \vec{u} = \begin{pmatrix} u \\ v \\ w \end{pmatrix}$$
(3)

Where A_H is the horizontal mixing coefficient and $\vec{\tau}$ is the vertical change in horizontal stress, both as a result of windstress at the surface and bottom friction at the ocean floor. [Talley 2011] [Pedlosky 1996]

3.2 Mass Conservation and the Continuity Equation

In our attempt to describe ocean circulation one very useful constraint is that we have conservation of mass, since water is incompressible:

$$\frac{d\rho}{dt} + \rho(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z}) = 0$$
(4)

Given the changes in ρ are small, we can write the continuity equation as follows:

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0 \tag{5}$$

[Talley 2011]

3.3 The Rossby and Ekman Number

For the momentum equations we compare the acceleration and advection terms to the Coriolis force to see which terms dominate on different spatial scales. We assume that there is no pressure gradient force and no frictional forces and so the momentum equations in the horizontal reduces to:

$$\frac{du}{dt} = fv, \quad \frac{dv}{dt} = -fu \tag{6}$$

The acceleration in eq.(6) is proportional to the horizontal velocity, $U^2 = u^2 + v^2$, and normal to the direction of the flow and so produces inertial currents that move in a circle. The centrifugal force is balanced by Coriolis force:

$$\frac{U^2}{r} = Uf\tag{7}$$

The ratio between these two forces is called the Rossby number:

$$Ro = \frac{U}{fL} \tag{8}$$

Where the radius of curvature, r, has been replaced by characteristic horizontal lengthscale of the flow, L and U is the horizontal characteristic velocity. If the Rossby number is less than unity the Coriolis force becomes dominant. In many cases for the large scale flow the Coriolis force is several orders of magnitude bigger than the acceleration and advection terms and so often we can neglect them. [Knauss 2005] [Pedlosky 1996]

We can make a similar scaling argument and assess the effect of turbulent mixing when compared to the Coriolis force for large scale flows. The ratio between turbulent mixing and Coriolis force is called the horizontal Ekman number and given $\rho A_H \nabla^2 \vec{u} \propto A_H U/L^2$ it can be written as:

$$E_H = \frac{A_H U/L^2}{fU} = \frac{A_H}{fL^2} \tag{9}$$

The Ekman number is difficult to estimate, because the mixing cooefficient is difficult to estimate, but it is generally a small parameter for the large scale flow. For a characteristic L = 1000 km the mixing coefficient would have to be much larger than current estimates for E_H to be unity and so in many cases we can drop the turbulent mixing term in eq. (3). [Pedlosky 1996]

3.4 The Ekman Layer

The upper layer of the ocean is subject to windstress and this wind driven frictional layer is called the Ekman layer. The Ekman layer extends to a depth of 50 meters approximately. The two forces that govern Ekman layer are the frictional force through windstress and the Coriolis force. The momentum equations therefore reduce to:

$$-fv = \frac{1}{\rho} \frac{\partial \tau_x}{\partial z} \tag{10a}$$

$$fu = \frac{1}{\rho} \frac{\partial \tau_y}{\partial z} \tag{10b}$$

We assume for now, that bottom friction is negligible and τ_x and τ_y therefore only represent windstress. From eq. (10a) and (10b) we can derive the meridional and zonal Ekman transports, $V_E = -\frac{\tau_x}{\rho f}$ and $U_E = -\frac{\tau_y}{\rho f}$, which tell us that the Ekman transport is always perpundicular to the direction of the windstress. The transport is exactly to the right of the windstress in the northern hemisphere and to the left in the southern hemisphere due to the changing sign of f. We mostly have zonal winds, τ_x , and for our case study, the North Pacific subtropical gyre, the windstress field is dominated north of 30N by the Westerlies, inducing southward Ekman transport, and south of 30N by the Easterlies, inducing northward Ekman transport. [Talley 2011]

Next, we cross-differentiate eq. (10a) with (10b):

$$\frac{\partial}{\partial x}(fu) + \frac{\partial}{\partial y}(fv) = \frac{1}{\rho}\left(\frac{\partial^2 \tau_y}{\partial x \partial z} - \frac{\partial^2 \tau_x}{\partial y \partial z}\right)$$
(11a)

$$f(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}) + v\beta = \frac{1}{\rho} \left(\frac{\partial^2 \tau_y}{\partial x \partial z} - \frac{\partial^2 \tau_x}{\partial y \partial z}\right)$$
(11b)

$$-f(\frac{\partial w}{\partial z}) + \beta v = \frac{1}{\rho} \frac{\partial}{\partial z} \nabla \times \vec{\tau}$$
(11c)

In the last step we've used the continuity equation, eq. (5). We assume that the vertical velocity, w, must be zero at the surface and equal to the Ekman velocity, w_E , at the bottom of the Ekman layer, $w(z_{surface}) = 0$ and $w(z_{Ekman}) = w_E$. We continue by vertically integrating eq. (11c) over the Ekman layer:

$$\int_{z_{Ekman}}^{z_{surface}} f \frac{\partial w}{\partial z} dz + \int_{z_{Ekman}}^{z_{surface}} \beta v dz = \frac{1}{\rho} \int_{z_{Ekman}}^{z_{surface}} \frac{\partial}{\partial z} (\nabla \times \vec{\tau}) dz$$
(12a)

$$fw_E + \beta V_E = \frac{1}{\rho} (\nabla \times \vec{\tau}) \tag{12b}$$

We observe from eq. (12b) that the vertical velocity at the base of the Ekman layer is proportional to the curl of the windstress.

3.5 The Geostrophic and Hydrostatic Balance

We now consider the ocean interior below the Ekman layer. For large spatial scales, which we can evaluate with the Rossby and Ekman number, we can argue that the acceleration and advection terms, bottom friction and friction due to mixing are all negligible. The mixed surface layer acts as a buffer, and so we have no friction due to windstress either, and the interior ocean is therefore in horizontal geostrophic balance, where the horizontal pressure gradient force is balanced by Coriolis force. The momentum equations reduce to the geostrophic balance:

$$-fv = -\frac{1}{\rho} \frac{\partial P}{\partial x}$$
(13a)

$$fu = -\frac{1}{\rho} \frac{\partial P}{\partial y} \tag{13b}$$

Similarly, we have vertical force balance between the pressure gradient and gravity. Assumming velocity in the vertical direction is negligible $\omega \ll v, u$, which is a reasonable approximation, since Earth's radius is much greater than the fluid layer thickness. We imagine the ocean as a sheet of paper, the flow is horizontal and the fluid layer is thin compared to the area of the sheet. The momentum equation, eq. (2c) reduces to the hydrostatic balance:

$$\frac{1}{\rho}\frac{\partial P}{\partial z} = -g \tag{14}$$

[Talley 2011]

3.6 Potential Vorticity

Vorticity is defined as the curl of the velocity vector. The vorticity of fluids in motion relative to Earth's surface is called relative vorticity. We also have planetary vorticity, which is vorticity due to the rotation of the Earth. Planetary vorticity becomes important when we have long lasting motion equivalent to the rotation rate of the Earth or longer. For geostrophic flows that are almost steady, the planetary vorticity becomes an important factor. We will only consider the vertical components for both relative and planetary vorticity, since the flow is mainly horizontal, refering again to our sheet analogy from section 3.5.

The vertical component of the relative vorticity at any given point is the curl of the horizonal velocity vector at that point:

$$\zeta = \left(\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}\right) \tag{15}$$

The vertical component of the planetary vorticity is equal to the Coriolis parameter, f. The absolute vorticity simply becomes $\zeta + f$. When we define potential vorticity in a fluid we consider a water column and include not only how it rotates, but also the height of the water column itself. Potential vorticity is conserved and this is a central concept in oceanography. If the water column is squashed the relative or planetary vorticity must decrease to conserve potential vorticity. Similarly, if the water column is stretched, the relative or planetary vorticity must increase. [Talley 2011]

3.7 The Sverdrup Balance and Western Boundary Currents

We derive the Sverdrup balance, initially by treating the ocean as two different regimes; the surface Ekman layer and the geostrophic ocean interior, then seeing how they couple to form a gyre structure. Beginning with the interior, we assume geostrophy and cross differentiate eq.s 13a and 13b:

$$\frac{\partial}{\partial x}(fu) + \frac{\partial}{\partial y}(fv) = -\frac{1}{\rho}\frac{\partial^2 P}{\partial x \partial y} + \frac{1}{\rho}\frac{\partial^2 P}{\partial x \partial y}$$
(16a)

$$f(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}) + \beta v = 0$$
(16b)

$$f(\frac{\partial w}{\partial z}) = \beta v \tag{16c}$$

Where we've used the continuity equation, eq. (5), in the last step and where we define $\beta = \partial f / \partial y$. We now vertically integrate eq. (16c) from the base of the Ekman layer to the ocean floor. We assume that the vertical velocity, w, is zero at the bottom, $w(z_{bottom}) = 0$, and that $w(z_{Ekman}) = w_E$.

$$\int_{z_{bottom}}^{z_{Ekman}} \beta v dz = \int_{z_{bottom}}^{z_{Ekman}} f(\frac{\partial w}{\partial z}) dz \tag{17a}$$

$$\beta V_{int} = f w_E \tag{17b}$$

Where V_{int} is the meridional transport in the interior. Combining eq. (17b) for the geostrophic interior with eq. (12b) for the Ekman Layer gives us the Sverdrup balance:

$$\beta V = \frac{1}{\rho} \nabla \times \vec{\tau} \tag{18}$$

Where V is the total meridional transport. [Pedlosky 1996]

The Sverdrup balance is a key result in explaining the gyre circulation of the ocean in terms of windstress, $\vec{\tau}$, induced Ekman transport, V, and change in planetary vorticity, $\beta = \partial f / \partial y$. As an example, the windtress in the North Pacific subtropical gyre is dominated north of 30N by the Westerlies and south of 30N by the Easterlies, inducing southward and northward Ekman transport respectively. Consequently, the Ekman transport converges along the zonal center of the subtropical gyre. This results in Ekman downwelling or pumping, as the water must go somewhere. The Ekman pumping in turn 'squashes' the water column thereby increasing the potential vorticity. To preserve potential vorticity, either the relative or planetary vorticity must decrease. The relative vorticity is small in the interior and so it is the planetary vorticity that changes by shifting the water southward. The Sverdrup transport, V, is the net meridional transport resulting from this change in planetary vorticity, which in turn is a result of Ekman pumping created by windtress $\vec{\tau}$. The large scale circulation is asymptric, with a slow flow in the interior and at the eastern boundary, but with a swift and narrow western boundary current. The closure of the circulation at the western boundary and the asymptry of the gyre cannot be explained with the linear Sverdrup balance. We know that Ekman pumping due to windstress results in an increase in planetary vorticity that, due to conservation of potential vorticity, induces an equatorward flow. For the flow to close the circulation and return north at the boundary some force must again add vorticity, and since it cannot be planetary vorticity that returns the flow north, it must be an increase in relative vorticity. This problem was addressed first by Stommel and then Munk who both included frictional terms in the momentum equations as possible solutions. Stommel considered bottom frition. This approach closes the circulation at the western boundary, but only if the bottom friction would be unrealistically high. Munk therefore suggested boundary friction. Friction at the boundary wall decreases velocity at the wall to zero, thereby injecting relative vorticity into the fluid, and it is only at the western boundary that this results in the added positive relative vorticity needed to allow the flow to move northward and close the circulation. [Talley 2011]

3.8 Geostrophic Currents

We can derive an expression for the horizontal velocity field with depth for geostrophic currents based on geostrophy, eq. (13), and the hydrostatic balance, eq. (14). We consider the geostrophic balance equation that includes the meridional velocity, eq. (13a), and cross differentiate with eq. (14). In the process we apply the Boussinesq approximation in which density variations in the horizontal direction are small compared to density variations in the vertical direction. We therefore let ρ be constant, $\rho = \rho_0$, in the geostrophic balance but variable with depth in the hydrostatic balance:

$$\frac{\partial}{\partial z}(fv) = \frac{\partial}{\partial z}\left(\frac{1}{\rho_0}\frac{\partial P}{\partial x}\right) \tag{19a} \qquad \qquad \frac{\partial}{\partial x}\frac{\partial P}{\partial z} = \frac{\partial}{\partial x} - g\rho \tag{20a}$$

$$f\frac{\partial v}{\partial z} = \frac{1}{\rho_0} \frac{\partial P}{\partial x \partial z}$$
(19b)
$$\frac{\partial P}{\partial x \partial z} = -g\frac{\partial \rho}{\partial x}$$
(20b)

We define the sea surface height, SSH or η , as the height of the water column relative to the geoid. The sea surface height is in the range of approximately 1 meter in the Pacific Ocean. This small variation in the height of the water column adds to the vertical pressure gradient. We define the depth of the ocean from bottom to geoid as h and so we can write the total vertical change in pressure as $\partial P/\partial z = \partial P/\partial \eta + \partial P/\partial h$. We isolate v in eq. (19b) and integrate over the depth of the water column, separating the integral in two parts; from z = 0 to η and from -h to z = 0.

$$v = \int_0^\eta \frac{1}{f\rho_0} \frac{\partial P}{\partial x \partial z} dz + \int_{-h}^0 \frac{1}{f\rho_0} \frac{\partial P}{\partial x \partial z} dz$$
(21a)

(21b)

We insert that $P = \rho g \Delta z$ in the first term on the right hand side and we insert eq. (20b) in the second term on the right hand side:

$$v = \frac{g}{f}\frac{\partial\eta}{\partial x} + \int_{-h}^{0} \frac{-g}{f\rho_0}\frac{\partial\rho}{\partial x}dz$$
(22a)

$$u = \frac{g}{f}\frac{\partial\eta}{\partial y} + \int_{-h}^{0} \frac{-g}{f\rho_0}\frac{\partial\rho}{\partial y}dz$$
(22b)

The first term on the right hand side of eq. (22a) and (22b) are the surface velocities or the barotropic velocities and the second terms are the baroclinic velocities. In a barotropic fluid lines of constant density, isopycnals, do not cross lines of constant pressure, isobars. The geostrophic velocity in a barotropic fluid is constant with depth. In a baroclinic fluid isobars and isopycnals are not parallel and the geostrophic velocity varies with depth. The relation between the geostrophic velocity shear and the horizontal density gradient is called the thermal wind relation and it is given by the baroclinic velocity in eq. (22). In our study we will use only the barotropic component of the absolute velocity. We compute the gradient of the sea surface height and since we know it to be proportional to the surface velocity of geostrophic currents, we can use it to locate the two meandering states of the Kuroshio. [Talley 2011] [Knauss 2005]

4 Community Earth System Model

This study is based on output from the high resolution Community Earth System Model, CESM. In the CESM the ocean is modelled with the Parallel Ocean Program version 2 or POP2. We use output from POP2 to analyse ocean variability in the Kuroshio region. In POP2 the vertical structure of the ocean is represented with 62 levels, varying from 10 m at the surface to 250 m at depth and with more centered around the thermocline. The horizontal grid spacing of POP2 is 0.1°, decreasing from 11 km at the equator to 2.5 km at higher latitudes. [Bishop et al. 2015] [Hurrell et al. 2013]. The ocean communicates with the coupler every 6 hours, where it recieves flux updates. The coupler air-sea fluxes are based on the surface layer scheme from [Large et al. 2009] that specifically aims to assemble flux parameters that improve estimates of air-sea fluxes that capture interannual variability. Based on [Large et al. 2009] a prescribed annual atmospheric evolution is applied in the CESM integration. The specific CESM run in consideration in this study consists of a 15 year long spin-up followed by an 11 year long control integration. The control integration is then followed by a perturbation experiment where the zonal wind stress in the Southern Ocean is increased by 50%. The data used in this study comes as 3600 X 2400 arrays representing a global grid with values of geophysical parameters assigned to each grid point. The parameter of interest to this study is the the annual and monthly mean sea surface height, from which we will compute the horizontal SSH gradient and make use of its proportionality to the barotropic velocity component in eq. (22a). The high resolution of the CESM is illustrated in figure 2, which shows the mean annual SSH field for the control integration, year 20. We view figure 2 in comparison to figure 3 taken from [Niiler et al. 2003], which is a sea surface height anomaly map based on observations with sea surface drifters, and we observe that the gradients of the SSH field in the CESM appears to agree well with observations.



Figure 2: Mean annual SSH map based on output from the Communty Earth System Model, the control integration, year 20. The Kuroshio region along the coast of Japan is nicely resolved and we observe the large meander and the structure of the Kuroshio Extension jet as it passes through the Izu-Osagawara Ridge to join the eastern flow of the North Pacific subtropical gyre. The reader can easily see that there are some issues plotting the SSH field based on the tripole grid of the CESM, notably in the Atlantic region and in the easternmost part of the Pacific. This is due to technical difficulties and the author apologises.

5 Method

Our objective is to compare the Kuroshio in the high resolution CESM with the real current in terms of mean and variability. Initially, we wish to see whether the model is in good agreement with observations. Secondly, we test the hypothesis that external time varying wind forcing is the driver for the transition between the two states of the Kuroshio. Thirdly, we test the hypothesis that interior ocean mechanics are responsible for the transition of the Kuroshio. We wish to develop a method to determine if the Kuroshio is in the large-meandering state or the non-large meandering state for a



Figure 3: Mean absolute sea level between the years 1992-2002, measured with surface velocity drifters. Taken from Niiler et al. 2003.

given model year in order to compare the bimodal variability with observations and see the effect of an atmospheric forcing pertubation.

5.1 Control Integration

We begin by processing the annual mean SSH for the ctrl integration, model years 16-26. The data is imported into Matlab and plotted for each model year. The area chosen is 25N-42N and 130E-145E in the Kuroshio region of the coast of Japan and centered around the meandering structure, see figure 4. We choose, at first simply by eye-sight, the longitude where the spatial difference between the two states is most pronounced, which we decide is at approximately 138°E. We want to determine the position of the Kuroshio by finding the strongest point of the flow. First, as a precursory exercise, we simply use the maximum value of the SSH as an indicator for the strength of the flow. Along the longitude 138°E we then compute the maximum value of the SSH for every model year. This allows us to record the location of the Kuroshio for the given year and determine which of the two states the Kuroshio can be said to be in. The resulting spatial distribution is shown in figure 5a.

Now, we improve our method. Firstly, by finding a more consistent way of selecting the longitude where the variance between the two states has its maximum. Secondly, by computing the gradient of the SSH rather than just using the maximum value. The maximum value of the SSH is not the most appropriate indicator of velocity at the sea surface. If we assume that the geostrophic balance holds, as discussed in the theory section, we can apply the barotropic component of eq. (22) and by computing the gradient of SSH obtain a more dynamically correct expression for the velocity of the Kuroshio current at the sea surface, since the gradient of the SSH is directly proportional to the velocity. We start by addressing the selection of the longitude of largest spatial difference between states. We compute the variance of the SSH field between the 11 model years using matlab function *var* defined by eq. (23).

$$var(SSH) = \frac{1}{N-1} \sum_{i=1}^{N} |SSH_i - \overline{SSH}|^2$$
(23)

Where N=11 and $\overline{SSH} = \frac{1}{N} \sum_{i=1}^{N} SSH_i$. We select the longitude for which the variance is maximal. The resulting value for the ctrl-integration is 137.65°E.

We now compute the gradient of the SSH between grid points along the selected longitude based on the forward difference approximation.

$$\frac{\partial SSH}{\partial y}(i) = \frac{SSH(i) - SSH(i+1)}{\Delta y}$$
(24)

We begin by determining Δy , the distance between the grid points in the y- or meridional direction. The difference in latitude between grid points is in degrees, ϕ . We convert to radians and then apply eq. (25) to find grid spacing in meters.

$$\Delta y = \Delta \phi \times R_E \tag{25}$$

Where the radius of the Earth is $R_E = 6.371 \times 10^6 m$.

As mentioned in the CESM description section 4.1, the grid spacing varies with location on the sphere, with maximum values at the equator and decreasing towards the poles and so ϕ also varies. Since we work with a comparatively small area, we wish to test if we can assume even spacing for simplification. We find the latitudinal difference, ϕ , between the grid points along longitude 137.65°E and compute the standard deviation with matlab function *std* defined by eq. (26)

$$std(\phi) = sqrt \frac{1}{N-1} \sum_{i=1}^{N} |\phi_i - \overline{\phi}|^2$$
(26)

Where N=100. This yields a standard deviation of 0.0028 and we conclude that we may approximate the problem and choose even grid spacing. The mean value of ϕ is 0.09 and we calculate $\Delta y = 10km$. We then compute the gradient of the annual mean SSH for the control integration along longitude 137.65°E. The resulting distribution of maximum ssh gradient is shown in figure 5b.

So far we have treated annual mean ssh for the ctrl integration. Now, we should like to make sure that the transitions of the Kuroshio is indeed inter-annual and that we have not missed any events affecting the transition happening on smaller timescales. We therefore repeat our procedure for the monthly mean SSH for the control integration. We consider only variations that are independent of any seasonal cycle. We remove seasonal cycles by computing the mean January, February, March and so on and then subtracting these seasonal means from each month belonging to it, that is e.g. for every January the mean January is subtracted. In this way, we create new arrays for every month in the 11 year period, free of seasonal cycles. The data is then processed using the same method as for the annual mean SSH. Results are shown in figure 6

5.2 Pertubation Experiment

From year 27 to year 42 in the model run a pertubation in the atmospheric forcing is introduced into the system; the zonal winds are increased by 50% in the Southern Ocean. We wish to assess the effect, if any, on the transition of states of the Kuroshio.

The data is imported and plotted as can be seen in figure 7. The aim is the same as for the control integration; to determine the variability of the two states of the Kuroshio, this time after a pertubation. Before repeating the same procedure as in section 5.2.1 for the tau integration, we wish to ascertain that the pertubation itself has not shifted the selected longitude where the spatial variance between the states is most pronounced. We again compute the variance of the SSH field, now between the 16 tau-group model years and select the longitude for which the sum of the variance is maximal. The resulting value for the tau-integration is 137.25°E. For the sake of simplicity we retain the value of the longitude from section 5.2.1 of 137.65°E, as the two values do not vary greatly. We compute the gradient for both annual and monthly mean SSH for the tau-group along the longitude 137.65°E. The ditributions can be seen in figure 8.

6 Results

6.1 Comparison of Model output with Satellite Altimeter Observations

In this first part of our analysis we compare our model results with measurements by satellite altimeter to see whether the CESM model produces realistic partitioning of the Kuroshio current between its two possible states. We compare the annual mean SSH field from the ctrl-run with altimeter data from Ocean Topography Experiment(TOPEX)/Poseidon, Jason-1 and European Remote Sensing Satellites 1 and 2 (ERS-1/2) taken from [Qiu and Chen 2005]. The satellite data contains yearly averaged SSH fields from the years 1993 to 2004. Beginning in 1993 the Kuroshio is settled in the large meandering state. From 1994-95 it is stable in the non-large meandering state but from 1995 the path length increases and the latitudinal position decreases and the Kuroshio transitions into the large meander again. Following this the system stays in the unstable state for roughly 7 years between 1995-2001 before transitioning to the non-large meander in 2002-2003, and thereafter again settling in the large meander in 2004. When comparing the temporal characteristics of the Kuroshio in the Topex/Poseidon satellite measurements with the control integration in the CESM, they seem reasonable. In our control integration the Kuroshio settled in the large meander twice in the space of 11 years, between years 16-18 and 23-26, and once in the non-large meander from year 19-22, see figure 4 and 5a. An unstable period of 7 years and a stable period for 4 years seem consistent when comparing the observation sample with the CESM control integration. One other thing to remark on in the comparison between model and observations, is that when considering interannual variability we would generally need a longer timeseries than our short 11 year control integration to confirm that model and observations are in agreement. Observations are however limited by time span of satellites in orbit.

6.2 Analysis of Atmospheric Forcing Response

In figure 4 we see annual mean SSH fields in the Kuroshio region from the CESM control integration, years 16-26. During the control integration the Kuroshio is settled in the large meandering structure from year 16-18 and again from year 23-26 and then transitions to the non-large meander in the years 19-21. In year 22 we have what appears to be a large meandering structure, or perhaps a transitionary structure. From the distribution of the latitudinal position of the Kuroshio in figure 5 we can see that for the maximum annual mean SSH along constant longitude 137.65°E the Kuroshio settles in the large meander 8 years out of 11 and in the non-large meander for a duration of 3 years. For the distribution of latitudes with maximum annual mean horizontal SSH gradient the Kuroshio settles in the large meander 7 years and in the non-large meander in 4 years. The distribution of the bimodal variability of the Kuroshio is therefore sensitive to the choice of tracking method. There is a general shift of the position of the Kurioshio to higher latitudes for the more dynamically correct horizontal gradient results and we also see greater spatial variability of the unstable state between the years 19-22. Most importantly, we also observe, that when the CESM is run with a prescribed annual atmospheric evolution, that is, with no change in the annual cycle of the external time varying wind forcing, we still see bimodal variability. Our results show that the variability of the Kuroshio is *not* affected by external time varying wind forcing, acting through a linear time-dependent Sverdrup balance.

The distribution of latitudes with maximum monthly mean horizontal SSH gradient for the control integration, see figure 6, shows us how the position of the Kuroshio fluctuates within the annual cycle and we are presumably with our method recording the transitionary positions of the Kuroshio when it is between states from year 18-19 and year 22-23.

Figure 7 shows the annual mean SSH fields from the CESM pertubation integration, years 27-42. Upon increasing the zonal windstress in the Southern Ocean by 50%, we see a marked difference in the variability of the Kuroshio as it seems to predominantly settle in the large meander after pertubation. The distribution of latitudes with maximum annual mean horizontal SSH gradient for the pertubation



Figure 4: Mean annual SSH fields in the Kuroshio region from the CESM control integration, years 16-26. The Kuroshio is settled in the large meandering structure from yr 16-18 and again from yr 23-26. The Kuroshio transitions to the non-large meander in the yrs 19-21. In year 22 we have perhaps a transitionary structure, as it is not a well defined large meander.

integration in figure 8a show that the Kuroshio is in the large meandering structure 14 out of 16 years, in the non-large meander in 2 years. This leaves out 1 year of the integration, and it is most likely year 40, where the variability of the position of the large meandering state, prevents us from capturing its southernmost position with our method. For the distribution of latitudes with maximum monthly mean horizontal SSH gradient in figure 8b we note that the spatial variability of the Kuroshio during the annual cycle differs from the annual mean SSH distribution as it occasionally moves to high latitudes close to the coast of Japan. Based on our results one may speculate that the zonal winds in the Southern Ocean could have an effect on the bimodal variability of the Kuroshio. This remote atmospheric forcing can only be transmitted to the Kuroshio region through internal oceanic mechanics, for reasons that are outside the scope of this study. Our results therefore support the hypothesis that mechanisms internal to the ocean system are linked to the low frequency bimodal variability of the Kuroshio.

7 Discussion

7.1 Method Evaluation

There are several points in our method that might be worth while evaluating. Firstly, we determine the position of the Kuroshio based on the barotropic velocity when the ocean is largely baroclinic. Since we only wish to the find position of Kuroshio at the surface and not analyse its vertical structure, we assume it is an acceptable approximation. Secondly, we compute the gradient of the annual mean SSH field in accordance with the barotropic component of eq. (22), and exploit that the horizontal SSH



Figure 5: (a): Distribution of latitudes with maximum annual mean SSH along constant longitude 137.65°E for the control integration, years 16-26. From the distribution we see that the Kuroshio settles in the large meander 8 yrs and in the non-large meander in 3 yrs. (b): Distribution of latitudes with maximum annual mean horizontal SSH gradient along constant longitude 137.65°E for the control integration, years 16-26. From the distribution we see that the Kuroshio settles in the large meander 7 yrs and in the non-large meander in 4 yrs. We observe that the distribution of the bimodal variability of the Kuroshio is sensitive to the choice of tracking method. There is a general shift of the position of the Kurioshio to higher latitudes for the horizontal gradient results and we also note greater spatial variability of the unstable state between the yrs 19-22. We also observe, that when the CESM is run with a prescribed annual atmospheric evolution, with no change in the annual cycle of the external time varying wind forcing, we still see bimodal variability.



Figure 6: Distribution of latitudes with maximum monthly mean horizontal SSH gradient along constant longitude 137.65°E for the control integration, years 16-26. For the monthly annual mean SSH horizontal gradient we are able to see how the position of the Kuroshio fluctuates within the annual cycle and presumably we are capturing the transition positions between states from yr 18-19 and 22-23.



Figure 7: Mean annual SSH fields in the Kuroshio region from the CESM pertubation integration, years 27-42. We see a significant difference in duration and in transition between states after we have pertubed the system. After the pertubation the Kuroshio seem to predominantly settle in the large meander.



Figure 8: (a): Distribution of latitudes with maximum annual mean horizontal SSH gradient along constant longitude 137.65°E for the pertubation, tau, integration, years 27-42. The Kuroshio is in the large meandering structure 14 out of 16 yrs, in the non-large meander in 2 yrs (b): Distribution of latitudes with maximum monthly mean horizontal SSH gradient along constant longitude 137.65°E for the pertubation, tau, integration, years 27-42. From the monthly distribution, we see the spatial variability of the Kuroshio during the annual cycle as it occasionally moves close to the Japanese coast at high latitudes in comparison to the annual mean SSH distribution.

gradient is proportional to the barotropic velocity. However, the proportionality constant contains the Coriolis parameter, f, which increases with latitude. Across the whole latitudinal band from 27.6°N to 36.7°N in which we perform our forward difference approximation the change in f is 25%. Since f changes with latitude, we might expect the strongest point of the flow to be shifted and so this might have an affect on how we have determined the position of the Kuroshio. Thirdly, our choice to only evaluate along a single longitude, makes it difficult to capture some of the transitional positions of the Kuroshio, but we accept this, as our main goal was to simply evaluate whether external or internal mechanics induced a transition of state for the Kuroshio.

7.2 CESM Validation

An evaluation of the SSH field generated by the CESM is provided in [Small et al. 2014], where the CESM SSH field is compared with satellite measurements from the Archiving, Validation and Interpretation, AVISO, merged product. The standard deviation of SSH field, σSSH , between the CESM and AVISO observations show, that model and observations compare well in general. The variability of the SSH field is examined and the results show that the variability is weaker compared to observations in most locations except for the tropical Pacific and Indian Ocean. A further analysis of the variability was parted in three temporal regimes; less than 90 days representing variability of mesoscale activity, 90-400 days representing the variability of the annual cycle, and greater then 400 days representing the low-frequency variability. Variability in all three regimes compares well between CESM and AVISO, except for the mesocale variability in the Kuroshio region, which is stronger than observations indicate. However, since we are investigating low frequency variability in the Kuroshio region, this does not affect our results.

8 Conclusion

In this study we have investigated two possible driving mechanisms for the low frequency transitions of the Kuroshio resulting in bimodal variability. We have based our analysis on output from the high resolution Community Earth System Model. With a theoretical basis in the geostrophic and hydrostatic balance we have computed the horizontal gradient of the annual and monthly mean sea surface height anomalies provided by the CESM to compute the barotropic surface velocity in the Kuroshio region. We have developed a method to track the position of the Kuroshio along the longitude where it displays the most significant path difference between its stable and unstable state. We have tested whether the bimodal variability of the Kuroshio is induced by external time varying wind forcing, acting through a linear time-dependent Sverdrup balance, by investigatin the transitions of the Kuroshio in a control integration with a prescribed annual atmospheric evolution. We find that the bimodal variability of the Kuroshio is not induced by an external time varying wind forcing. We conduct a pertubation experiment and introduce an atmospheric forcing in the Southern Ocean, the effect of which can only be transmitted to the Kuroshio through internal ocean mechanics. Our results support the hypothesis that mechanisms internal to the ocean system are linked to the low frequency bimodal variability of the Kuroshio.

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A Matlab Script

```
1 clear all
2 close all
```

```
2
3
  \% Investigating the variability of the Kuroshio current with CESM output.
4
     Focus on bimodal feature; the large and non-large meander.
  % Control (ctrl) integration of 11 years, prescribed annual atmospheric
\mathbf{5}
     evolution wind stress in CESM integration. Parameter: mean sea surface
      height anomaly, annual + monthly.
  \% Pertubation (tau) integration of 16 years, with prescribed annual
6
     atmospheric evolution and introducing pertubation by 50% increase in
     zonal windstress in the Southern Ocean. Parameter: mean sea surface
     heigth anomaly, annual + monthly.
  %An introduction to processing big data was an objective and an integrated
7
      part of this project. Consequently, the following description is
     stepwise and detailed with regard to the %applied software and data
     processing.
  %We begin by installing PuTTY, which is a Secure shell, SSH, client for
8
     windows. Putty is primarily used for remote access to Unix and Linux
     servers, in our case the ikos/atlas %server at the Physics Department
     of Copenhagen University. The NetCDF data can be processed using
     NetCDF Operators, NCO, which is a set of operators developed at NCAR
     for analysis % and manipulation of gridded scientific datasets. NCO
     operators take netCDF files as input, perform a given operation, and
     then produce a netCDF file as output. Data is accessed %through PuTTY
     and processed in UNIX terminal using both NetCDF operators and by
     creating scripts run in matlab. We also install Xming, a display
     server needed for viewing Matlab %figures on windows. After creation,
     plots and figures are retrieved from the remote server with PuTTY
     Secure Copy, PSCP. PSCP is a Secure CoPy, SCP, client and a freeware
     program % for the Windows command line processor.
  %Seasonal cycles are removed from the SSH monthly means for the control
9
     and pertubation integration. The computations are performed using
     NetCDF operators. The operator ncra is used % for averaging and ncdiff
     is used for subtraction. The individual data files are concatenated
     into a single matrix with the operator nercat, whereupon we select the
      Kuroshio region %by cutting the matrix down with the operator ncks,
     in order to bring down the run time of the Matlab code.
10
  % Importing data, control integration, 11 annual mean ssh anomaly, years
11
     16 to 26. Defining landmass as NaN.
  ctrl annual ssh = zeros(3600, 2400, 11);
12
  ctrl_annual_long = zeros(3600, 2400, 11);
13
  ctrl_annual_lat = zeros(3600, 2400, 11);
14
  a = 16;
15
  for i = 1:11
16
  ctrl_annual_ssh(:,:,i)=ncread(['/gfy/gfy-1/cdz293/CESM.T62_t12.002/ctrl.g.
17
     e11.G.T62 t12.002.pop.h.SSH.00' num2str(a, '%02d') '.annualmean.nc'], '
```

```
SSH');
   ctrl_annual_long(:,:,i)=ncread(['/gfy/gfy-1/cdz293/CESM.T62_t12.002/ctrl.g
18
      . e11.G.T62 t12.002.pop.h.SSH.00 ' num2str(a, '%02d') '.annualmean.nc'], '
      TLONG');
   ctrl annual lat (:,:,i)=ncread (['/gfy/gfy-1/cdz293/CESM.T62 t12.002/ctrl.g.)
19
      e11.G.T62 t12.002.pop.h.SSH.00' num2str(a, '%02d') '.annualmean.nc'], '
      TLAT');
  a = a+1;
20
   end
21
   ctrl annual ssh(ctrl annual ssh==-1)=nan;
22
   ctrl_annual_ssh(ctrl_annual_ssh>1e20)=nan;
23
   \operatorname{ctrl} annual \operatorname{lat}(\operatorname{ctrl} annual \operatorname{lat}=-1)=\operatorname{nan};
^{24}
   ctrl_annual_long(ctrl_annual_long==-1)=nan;
25
26
  \% Importing data, control integration, 132 monthly mean ssh anomaly, years
27
        16 to 26. Defining landmass as NaN.
   ctrl monthly ssh = ncread (['/climgeo/ssh/ctrl.kuro.16-26.nc'], 'SSH'); % or
^{28}
        '/climgeo/ssh/ctrl.16-26.nc'
   ctrl monthly long = ncread (['/climgeo/ssh/ctrl.kuro.16-26.nc'], 'TLONG');
29
   ctrl monthly lat = ncread (['/climgeo/ssh/ctrl.kuro.16-26.nc'], 'TLAT');
30
   \operatorname{ctrl}_{\operatorname{monthly}_{\operatorname{ssh}}} = -1 = \operatorname{nan};
31
   \operatorname{ctrl} monthly \operatorname{ssh}(\operatorname{ctrl} monthly \operatorname{ssh}>1e20)=\operatorname{nan};
32
   ctrl_monthly_lat(ctrl_monthly_lat==-1)=nan;
33
   ctrl monthly \log(\text{ctrl monthly } \log = -1) = \text{nan};
34
35
  % Importing data, pertubation integration, 16 annual mean ssh anomaly,
36
      years 27 to 42. Defining landmass as NaN.
   tau annual ssh = zeros(101, 101, 16);
37
   tau annual long = zeros(101, 101, 16);
38
   tau annual lat = z e ros (101, 101, 16);
39
   b = 27;
40
   for j = 1:16
41
   tau annual ssh(:,:,j) = ncread(['/climgeo/ssh/tau.' num2str(b, '%02d')'.
42
      annualmean.kuro.nc'], 'SSH');
   tau annual long(:,:,j) = ncread(['/climgeo/ssh/tau.' num2str(b, \%02d')'.
43
      annualmean.kuro.nc'], 'TLONG');
   tau_annual_lat(:,:,j) = ncread(['/climgeo/ssh/tau.' num2str(b, '%02d')'.
44
      annualmean.kuro.nc'], 'TLAT');
   b = b+1;
45
   end
46
   tau annual ssh(tau annual ssh==-1)=nan;
47
   tau annual ssh(tau annual ssh > 1e20) = nan;
48
   tau annual lat(tau annual lat==-1)=nan;
49
   tau annual \log(\tan annual \log = -1) = \tan;
50
   plot tau annual ssh = zeros(3600, 2400, 16);
51
   plot_tau_annual_long = z eros(3600, 2400, 16);
52
   plot tau annual lat = z e ros (3600, 2400, 16);
53
  b1 = 27;
54
  for j = 1:16
55
```

```
plot tau annual ssh(:,:,j) = ncread(['/climgeo/ssh/tau' num2str(b1, '%02d')]
    '.annualmean.nc'], 'SSH');
plot tau annual long(:,:,j) = ncread(['/climgeo/ssh/tau' num2str(b1, '%02d'
   ) '. annualmean . nc '], 'TLONG');
plot tau annual lat (:,:,j) = ncread(['/climgeo/ssh/tau' num2str(b1,'%02d')]
    '.annualmean.nc'], 'TLAT');
b1 = b1+1;
end
plot tau annual ssh(plot tau annual ssh==-1)=nan;
plot tau annual ssh(plot tau annual ssh>1e20)=nan;
plot_tau_annual_lat(plot_tau_annual_lat==-1)=nan;
plot tau annual \log(\text{plot tau annual }\log=-1)=\text{nan};
% Importing data, pertubation integration, 192 monthly mean ssh anomaly,
   years 27 to 42. Defining landmass as NaN.
tau_monthly_ssh = ncread(['/climgeo/ssh/tau.kuro.27-42.nc'], 'SSH');
tau monthly long = ncread (['/climgeo/ssh/tau.kuro.27-42.nc'], 'TLONG');
tau monthly lat = ncread (['/climgeo/ssh/tau.kuro.27-42.nc'], 'TLAT');
tau monthly ssh(tau monthly ssh==-1)=nan;
tau monthly ssh(tau monthly ssh>1e20)=nan;
tau_monthly_lat(tau_monthly_lat = -1) = nan;
tau monthly \log(\tan \text{ monthly } \log = -1) = \tan;
\% Selecting areas in Kuroshio region for illustrative figures and for
   dataanalysis.
lat min1 = 1470; % 27.6197N
lat min2 = 1440; % 24.9290N
lat_max1 = 1570; \% 36.6857N
lat max2 = 1630; % 41.7393N
\log \min 1 = 2430; \% 132.7593E
long min2 = 2400; % 129.7459E
long max1 = 2530; % 142.8179E
long max2 = 2550; \% 144.8318E
\% Figure 1: pcolor map of Kuroshio Region showing large meander, crtl year
    17, and non-large meander, ctrl year 21.
map ssh1 = pcolor(ctrl annual long(long min2:long max2, lat min1:lat max2)
   ,2) ', ctrl_annual_lat(long_min2:long_max2,lat_min1:lat_max2,2) ',
   ctrl annual ssh(long min2:long max2,lat min1:lat max2,2)');
map\_ssh1.EdgeAlpha=0
axis xy
xlabel('\fontsize{16} Longitude')
ylabel('\fontsize{16} Latitude')
```

```
<sup>91</sup> e=colorbar ('Location', 'SouthOutside');
```

```
92 e.Label.String = '\fontsize {16} Mean SSH [cm]';
```

```
93 axes ('Position', [.16 .015 .7 .1], 'Visible', 'off');
```

```
94 % Figure 2: pcolor map of ctrl integration year 20 (choosen at random) to compare resolution of CESM with Niller et al. from 2003.
```

```
95 figure
```

56

57

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89

90

```
map ssh = pcolor(ctrl annual long(:,:,5)', ctrl annual lat(:,:,5)',
96
       \operatorname{ctrl} annual \operatorname{ssh}(:,:,5)');
   map ssh.EdgeAlpha=0
97
   coast = load('coast.mat')
98
   mapshow(coast.long,coast.lat,'DisplayType','polygon','FaceColor','black')
99
   mapshow(coast.long+360,coast.lat,'DisplayType','polygon','FaceColor','
100
       black ')
   axis xy
101
   xlabel( '\fontsize {16} Longitude ')
102
   ylabel('\fontsize {16} Latitude')
103
   e=colorbar('Location', 'SouthOutside');
104
   e.Label.String = '\fontsize {16} Mean SSH [cm]';
105
   axes('Position', [.16 .015 .7 .1], 'Visible', 'off');
106
   \% Figure 3: plotting control integration years 16-26 to illustrate that
107
      CESM captures bimodal feature with prescribed annual atmospheric
       evolution.
   figure
108
   string = { '\fontsize {16} mean SSH, ctrl year 16 ', '\fontsize {16} mean SSH,
109
       ctrl year 17', '\fontsize {16}mean SSH, ctrl year 18', '\fontsize {16}mean
       SSH, ctrl year 19', '\fontsize {16} mean SSH, ctrl year 20', '\fontsize
       {16}mean SSH, ctrl year 21', '\fontsize {16}mean SSH, ctrl year 22', '\
       fontsize {16} mean SSH, ctrl year 23', '\fontsize {16} mean SSH, ctrl year
       24', '\fontsize {16} mean SSH, ctrl year 25', '\fontsize {16} mean SSH, ctrl
        year 26';
   c = 1;
110
   for k = 1:11
111
   subplot(3,4,k);
112
   pcolor ctrl =
113
   pcolor(ctrl annual long(long min2:long max2, lat min2:lat max2, k)',
114
       ctrl annual lat(long min2:long max2,lat min2:lat max2,k)',
       ctrl annual ssh(long min2:long max2,lat min2:lat max2,k)');
   pcolor ctrl.EdgeAlpha=0
115
   title (string{k})
116
   axis xy
117
   set (gca, 'xTick', 130:5:140)
118
   set (gca, 'xTickLabel', { '\fontsize {16}130E', '\fontsize {16}135E', '\fontsize
119
       \{16\}140E'\})
   set (gca, 'yTick', 25:5:40)
120
   set (gca, 'yTickLabel', { '\fontsize {16}25N', '\fontsize {16}30N', '\fontsize
121
       \{16\}35N', '\setminus fontsize\{16\}40N'\}
   c = c + 1;
122
   end
123
   axes ('Position', [.16 .015 .7 .2], 'Visible', 'off');
124
   d=colorbar('Location', 'SouthOutside');
125
   xlabel(d, \{ ' \setminus fontsize \{16\}SSH [m] ' \})
126
   set (gca, 'fontsize', 16)
127
   \% Figure 4: plotting tau integration years 27-42 to illustrate on bimodal
128
       feature in CESM with prescribed annual atmospheric evolution and
       introduced pertubation.
```

```
string = { '\fontsize {16} mean SSH, tau year 27', '\fontsize {16} mean SSH, tau
130
        year 28', '\fontsize {16} mean SSH, tau year 29', '\fontsize {16} mean SSH,
        tau year 30', '\fontsize {16}mean SSH, tau year 31', '\fontsize {16}mean
       SSH, tau year 32', '\fontsize {16} mean SSH, tau year 33', '\fontsize {16}
      mean SSH, tau year 34', '\fontsize \{16\} mean SSH, tau year 35', '\fontsize
       {16}mean SSH, tau year 36', '\fontsize {16}mean SSH, tau year 37', '\
       fontsize {16}mean SSH, tau year 38', '\fontsize {16}mean SSH, tau year 39
       ', '\fontsize {16} mean SSH, tau year 40', '\fontsize {16} mean SSH, tau
       year 41', '\fontsize {16} mean SSH, tau year 42' };
   c = 1;
131
   for k = 1:16
132
   subplot(3, 6, k);
133
   pcolor tau = pcolor (plot tau annual long(long min2:long max2, lat min2:
134
       lat max2,k)', plot tau annual lat(long min2:long max2, lat min2:lat max2
       , k) ', plot tau annual ssh(long min2:long max2, lat min2:lat max2, k) ');
   pcolor tau.EdgeAlpha=0
135
   title (string{k})
136
   axis xy
137
   set (gca, 'xTick', 130:5:140)
138
   set (gca, 'xTickLabel', { '\fontsize {16}130E', '\fontsize {16}135E', '\fontsize
139
       {16}140E'})
   set (gca , 'yTick ' ,25:5:40)
140
   set (gca, 'yTickLabel', { '\fontsize {16}25N', '\fontsize {16}30N', '\fontsize
141
       \{16\}35N', '\setminus fontsize\{16\}40N'\}
   c = c + 1;
142
   end
143
   axes ('Position', [.16 .015 .7 .2], 'Visible', 'off');
144
   d=colorbar('Location', 'SouthOutside');
145
   xlabel(d,{ '\fontsize{16}SSH [m] '})
146
   set (gca, 'fontsize', 16)
147
148
   \% Selecting longitude where the variance in ssh is largest between the
149
       control integration years, i.e. where bimodal feature is most
       pronounced. Using this select longitude for all further analysis.
   var ctrl annual ssh = var(ctrl annual ssh(long min1:long max1, lat min1:
150
       lat \max(1, :), 0, 3, \operatorname{`omitnan'});
   sum_var_ctrl_annual_ssh = sum(var_ctrl_annual_ssh);
151
   [a1, a2] = \max(sum var ctrl annual ssh);
152
   index_long = long_min1 + a2;
153
   index long1 = 47; \% a2 = 47
154
   select long = ctrl_annual_long(index_long,lat_min1,1); % value 137.65E
155
   \% Checking that the large meander is still at its 'widest' at 137.65\mathrm{E}
156
       after pertubation.
   var_tau_annual_ssh = var(tau_annual_ssh(:,:,:), 0, 3, 'omitnan');
157
   sum var tau annual ssh = sum(var tau annual ssh);
158
   [b1, b2] = \max(\text{sum var tau annual ssh});
159
   index \log 2 = \log \min 1 + b2;
160
   select long 2 = tau annual long (b2, 1, 1); % value = 137.25
161
```

```
162
   \% Finding for the selected longitude the latitudes where the gradient in
163
      ssh is at its max. for ctrl integration and tau integration, annually
      + monthly.
   \% 1. determining distance between latitudinal grid points (spacing of
164
       horizontal grid varies with location on sphere, but as I work with a
       relatively small area, and want to test if I can approximate even
      spacing)
   index ctrl annual lat grid = ctrl annual lat (long min1:long max1, lat min1:
165
      lat max1,1); % just checking value for 1st ctrl group year.
   phi = index_ctrl_annual_lat_grid (47,1:end-1)-index_ctrl_annual_lat_grid
166
       (47, 2: end); % result = values are approx. 10 km (spacing between
      latitudinal grid points)
   std phi = std (abs (phi (:))); \% = 0.0028
167
   mean_phi = std ( abs ( phi ( : ) ) ); \% = 0.0930
168
   delta y = 10000; % the distance between two latitudinal grid points in
169
      meters
170
   \% 2. Computing horizontal SSH gradient based on the forward difference
171
      approximation ...
   % I. ctrl integration annual mean SSH
172
   ctrl_annual_ssh_grad = (ctrl_annual_ssh(:, 1:end-1,:)-ctrl_annual_ssh(:, 2:end-1,:))
173
      end,:)) ./ delta y;
   N1 = 11;
174
   ctrl_annual_lat_hist1 = zeros(1,N1);
175
   ctrl annual lat hist_2 = zeros(1, N1);
176
   f = 1;
177
   for i = 1:N1
178
   [c1, c2] = \max(ctrl_annual_ssh(index_long, lat_min1: lat_max2, i));
179
   [d1, d2] = \max(\text{ctrl annual ssh grad}(\text{index long}, \text{lat min1:lat max1}, i));
180
   index ctrl annual_lat1 = lat_min1 + c2;
181
   index ctrl annual lat 2 = lat min 1 + d2;
182
   ctrl_annual_lat_hist1(f) = ctrl_annual_lat(index_long,
183
      index ctrl annual lat1, i);
   ctrl annual lat hist2(f) = ctrl annual lat(index long,
184
      index ctrl annual lat2, i);
   f = f + 1;
185
   end
186
   % II. ctrl integration monthly mean SSH
187
   ctrl_monthly_ssh_grad = (ctrl_monthly_ssh(:,1:end-1,:)-ctrl_monthly_ssh
188
      (:, 2: end, :)) ./ delta y;
   N2 = 132;
189
   ctrl monthly lat hist = z e ros(1, N2);
190
   g = 1;
191
   for j = 1:N2
192
   [e1, e2] = \max(ctrl_monthly_ssh_grad(index_long1, :, j)); \% all latitudes,
193
       since the data has allready been cut using NCO
   index ctrl monthly lat = e2;
194
```

```
\operatorname{ctrl} monthly lat \operatorname{hist}(g) = \operatorname{ctrl} monthly \operatorname{lat}(\operatorname{index} \operatorname{long}),
195
       index ctrl monthly lat);
   g = g + 1;
196
   end
197
   % III. tau integration annual mean SSH
198
   tau annual ssh grad = (tau annual ssh(:, 1: end -1,:) - tau annual ssh(:, 2: end
199
       ,:)) ./ delta y;
   N3 = 16;
200
   tau annual lat hist = zeros(1, N3);
201
   h = 1:
202
   for k = 1:N3
203
   [f1, f2] = max(tau_annual_ssh_grad(index_long1, :, k));
204
   index tau annual lat = f2;
205
   tau annual lat hist(h) = tau annual lat(index long1, index tau annual lat, k
206
       );
   \mathbf{h} = \mathbf{h} \! + \! 1;
207
   end
208
   % IV. tau integration monthly mean SSH
209
   tau monthly ssh grad = (tau monthly ssh(:, 1:end-1,:)-tau monthly ssh(:, 2:
210
       end,:)) ./ delta y;
   N4 = 192;
211
   tau monthly lat hist = z eros(1, N4);
212
   m=1:
213
   for i = 1:N4
214
    [g1,g2] = max(tau_monthly_ssh_grad(index_long1,:,i)); % all latitudes,
215
       since the data has allready been cut using NCO
   index tau monthly lat = g2;
216
    tau monthly lat hist(m) = tau monthly lat(index long1,
217
       index tau monthly lat);
   m = m+1;
218
   end
219
220
   \% H1: ctrl integration, histogram of latitudes along constant longitude,
221
       137.65E, where annual mean ssh has maximum value.
   nbins = 7
222
   H1 = histogram(ctrl annual lat hist1, nbins)
223
    set (gca, 'xTick', 27:1:35)
224
    set (gca, 'xTickLabel', { '27N', '28N', '29N', '30N', '31N', '32N', '33N', '34N', '35N
225
       '})
    set (gca, 'fontsize', 16)
226
    title ({ '\fontsize {16} Latitude of maximum annual mean SSH'; ' ctrl years
227
       16 - 26'
    ylabel('\fontsize {16} Years')
228
    xlabel( '\fontsize {16} Latitude ')
229
    axis([27 \ 35 \ 0 \ 7])
230
   % Hist2: ctrl integration, histogram of latitudes along constant longitude
231
       , 137.65E, where gradient of annual mean ssh has maximum value
   nbins = 7
232
   H2 = histogram(ctrl annual lat hist2, nbins)
233
```

```
set (gca, 'xTick', 27:1:35)
234
   set (gca, 'xTickLabel', { '27N', '28N', '29N', '30N', '31N', '32N', '33N', '34N', '35N
235
       '})
   title ({ '\fontsize {16} Latitude of maximum gradient of annual mean SSH'; '
236
       ctrl years 16-26'})
   ylabel('\fontsize {16} Years')
237
   xlabel('\fontsize {16} Latitude')
238
   axis ([28 35 0 7])
239
   % Hist3: ctrl integration, histogram of latitudes along constant longitude
240
       , 137.65E, where gradient of monthly mean ssh has maximum value
   nbins = 7
241
   H3 = histogram(ctrl monthly lat hist, nbins)
242
   set (gca, 'xTick', 27:1:35)
243
   set (gca, 'xTickLabel', { '27N', '28N', '29N', '30N', '31N', '32N', '33N', '34N', '35N
244
       '})
   set (gca, 'fontsize', 16)
245
   title ({ '\fontsize {16} Latitude of maximum gradient of monthly mean SSH'; '
246
       \operatorname{ctrl} years 16-26'})
   ylabel('\fontsize {16} Months')
247
   xlabel( '\fontsize {16} Latitude ')
248
   axis([27 35 0 50])
249
   \% Hist4: tau integration, histogram of latitudes along constant longitude,
250
        137.65E, where gradient of annual mean ssh has maximum value
   nbins = 7
251
   H4 = histogram(tau annual lat hist, nbins)
252
   set (gca, 'xTick', 27:1:35)
253
   set (gca, 'xTickLabel', { '27N', '28N', '29N', '30N', '31N', '32N', '33N', '34N', '35N
254
       '})
   set (gca, 'fontsize', 16)
255
   title ({ '\fontsize {16} Latitude of maximum gradient of annual mean SSH '; '
256
       tau years 27-42')
   ylabel('\fontsize {16} Years')
257
   xlabel('\fontsize {16} Latitude')
258
   axis ([28 33 0 13])
259
   % Hist5: tau integration, histogram of latitudes along constant longitude,
260
        137.65E, where gradient of monthly mean ssh has maximum value
   nbins = 7
261
   H5 = histogram(tau_monthly_lat_hist, nbins)
262
   set (gca, 'xTick', 27:1:35)
263
   set (gca, 'xTickLabel', { '27N', '28N', '29N', '30N', '31N', '32N', '33N', '34N', '35N
264
       '})
   set (gca, 'fontsize', 16)
265
   title ({ '\fontsize {16} Latitude of maximum gradient of monthly mean SSH'; '
266
       tau years 27-42')
   ylabel( '\fontsize {16} Months')
267
   xlabel('\fontsize {16} Latitude')
268
   axis ([28 35 0 90])
269
```