

The signature of mesoscale eddies on sea surface temperature and its associated heat transport

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DECLARATION

I declare that the work presented in this thesis is my own and that the work of others is appropriately referenced.

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ABSTRACT

This thesis aims at analysing the impact of oceanic turbulence and air-sea interactions on the sea surface temperature (SST) of the extra-tropical oceans on spatial scales of a few hundred kilometres (the so-called “mesoscale”). Using satellite-based measurements of SST and sea level, as well as surface tracks of mesoscale oceanic cyclones and anticyclones, it is shown that turbulence does not transport heat through systematic motions of cold cyclones and warm anticyclones, as was previously thought in regions of strong mean flows like the Gulf Stream. Rather, it is suggested that heat is transported as a result of the slight phase shift between temperature and pressure fluctuations developing on the mean flow. In addition, tentative estimates of the rate at which air-sea heat fluxes damp the SST signatures of cyclones and anticyclones are provided. The weak values obtained ($\sim 20 \text{ W/m}^2\text{K}$) contrast sharply with theoretical expectations, but are in agreement with the observed long-lived thermal heat content anomalies associated with the cyclones and anticyclones. These observations provide important benchmarks for high-resolution ocean models and may moreover guide the parametrization of subgrid-scale heat transport in climate models.

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Chapter 1

INTRODUCTION

Until a couple of decades ago, ship observations were the only means of observing the ocean. Different places would thus be covered at different times, and it would typically take at least several years, or even decades, to obtain an estimate of the spatial distribution of an oceanic tracer, such as the estimate of the dissolved oxygen concentration displayed in Figure 1.1. As it was simply impossible to assess the impact of any time variations on the resulting spatial patterns, the latter were interpreted as representing time averages and the view of a nearly steady, or slowly evolving, ocean emerged. The obtained smooth and large-scale patterns, such as those seen in Fig. 1.1, were then interpreted as being the signature of a similarly smooth, large-scale and laminar oceanic flow (e.g. Wunsch, 2001).

This interpretation of Fig. 1.1 started to be challenged with the development of new observing techniques, such as open-ocean moorings that can measure time series of ocean currents at a fixed location, and floats that can track the movement of fluid parcels. These new measurements, examples of which are displayed in Figure 1.2a and b, respectively, revealed an intense time variability in ocean currents that often obscures any “time-mean” flow in which it may be embedded: In the mooring time-series of Fig. 1.2a, zonal currents are dominated by vigorous fluctuations that, although surface-intensified, are often coherent throughout the water column and evolve on a typical time scale of a few months (~ 100 days). The float displacements of Fig. 1.2b, in turn, are dominated by motions on space scales of typically ~ 100 km, which is much smaller than the smooth patterns in Fig. 1.1. Typically seen in observations, intense current fluctuations on this space scale were termed “mesoscale eddies” (without a priori assigning any dynamical nature to them), and we can think of them as the oceanic analogues to the “weather” (synoptic disturbances) in the atmosphere.

Today, space-based instruments provide near-simultaneous observations of surface currents across the globe, and their analysis has confirmed that the kinetic energy

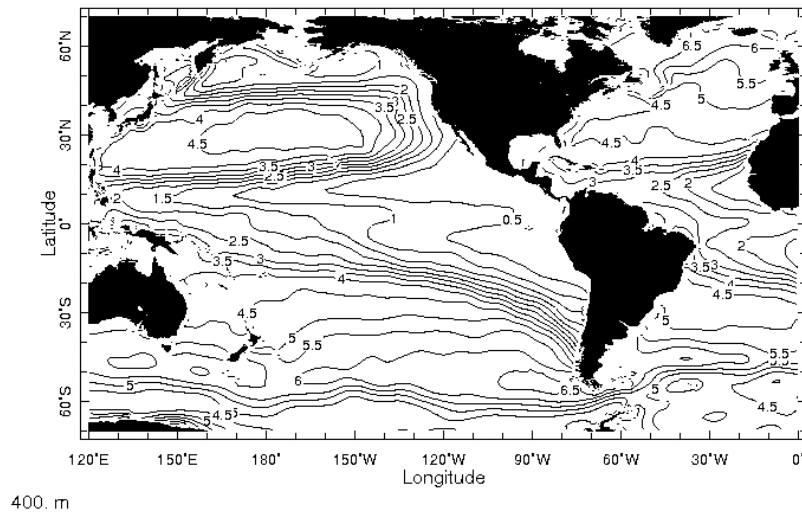


Fig. 1.1: Spatial distribution of dissolved oxygen (in ml/l) at 400 m depth in the Atlantic and Pacific oceans, from the World Ocean Atlas (Appendix B.5). Both bio-geochemical processes and oceanic transports control this distribution.

of time-varying surface flows peaks on the mesoscale and, indeed, that the transient mesoscale eddy kinetic energy is often more than an order of magnitude larger than that of the background time-mean flow (Stammer, 1997a; Scott and Wang, 2005; Ferrari and Wunsch, 2009). (In the interpretation of these satellite-based observations we however have to remember that, as will become apparent in the following chapters, neither motions on much smaller scales, nor the time-mean flow is, as of now, accurately captured by them.)

Inspired by distributions like Fig. 1.1, theories of a large-scale, laminar and near-steady ocean circulation were developed (e.g. Luyten et al., 1983). As they were strikingly successful at explaining distributions like Fig. 1.1, the observed vigorous transient eddy activity was at first regarded as mere small-scale noise, with no dynamical role in the large-scale mean circulation (apart from perhaps that of slowly diffusing properties down the latter’s mean gradients).

However time-mean oceanic transports of tracers and even mass do contain eddying contributions. Parametrizing these eddy mass transports in coarse-grid climate models, which typically resolve atmospheric eddies but not their order-of-magnitude smaller oceanic counterparts, has revealed that they are not negligible, but instead lead to zero-order improvements in the “realism” of the modelled ocean circulation and its interaction with the atmosphere. As an example, Figure 1.3 contrasts two equilibrium

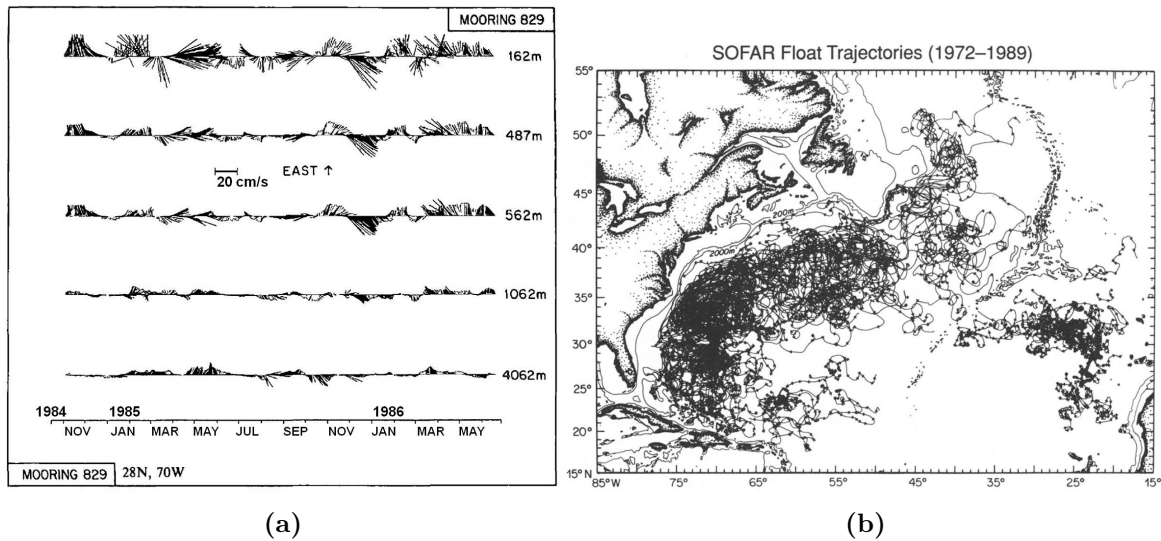


Fig. 1.2: Observations of time-varying currents: (a) Time series of the zonal current at various depths in the water column, measured by a current-meter mooring at a fixed location to the south of the Gulf Stream, from Schmitz (1989). (b) 230 neutrally buoyant floats track small-scale (~ 100 km), time-varying motions (dots every 30 days) at depths of 700m, 1100-1500m, and 2000 m in the North Atlantic, from Richardson (1993).

solutions of such a model (Danabasoglu et al., 1994), of which only the one depicted in the lower panels includes a parametrization of eddy mass transports (Gent and McWilliams, 1990). As seen in the lhs panels, the time-mean and zonally-averaged oceanic transport stream functions reveal striking discrepancies between the two integrations that are especially pronounced in the Southern Ocean (SO, approximately to the south of 40°S): In Fig. 1.3a, only the time-mean flow contributes to the transport and drives an intense overturning that reaches from the sea floor to the surface. The implied surface equatorward advection of cold and dense waters towards regions of warmer and lighter waters destabilizes oceanic water columns and thereby, as seen in Fig. 1.3b, triggers convection throughout the SO. In Fig. 1.3c, the contribution of the time-mean flow to the transport is approximately the same as in Fig. 1.3a, but here the eddy-induced transport opposes it, leading to a much weaker net transport of water masses. One implication of this is that water columns are now typically stably stratified so that, as in observations, convection becomes a highly-localized phenomenon (Fig. 1.3d).

Fig. 1.3c moreover implies that the ocean's time-variability contributes to the poleward heat transport of the climate system that is required to carry the excess heat

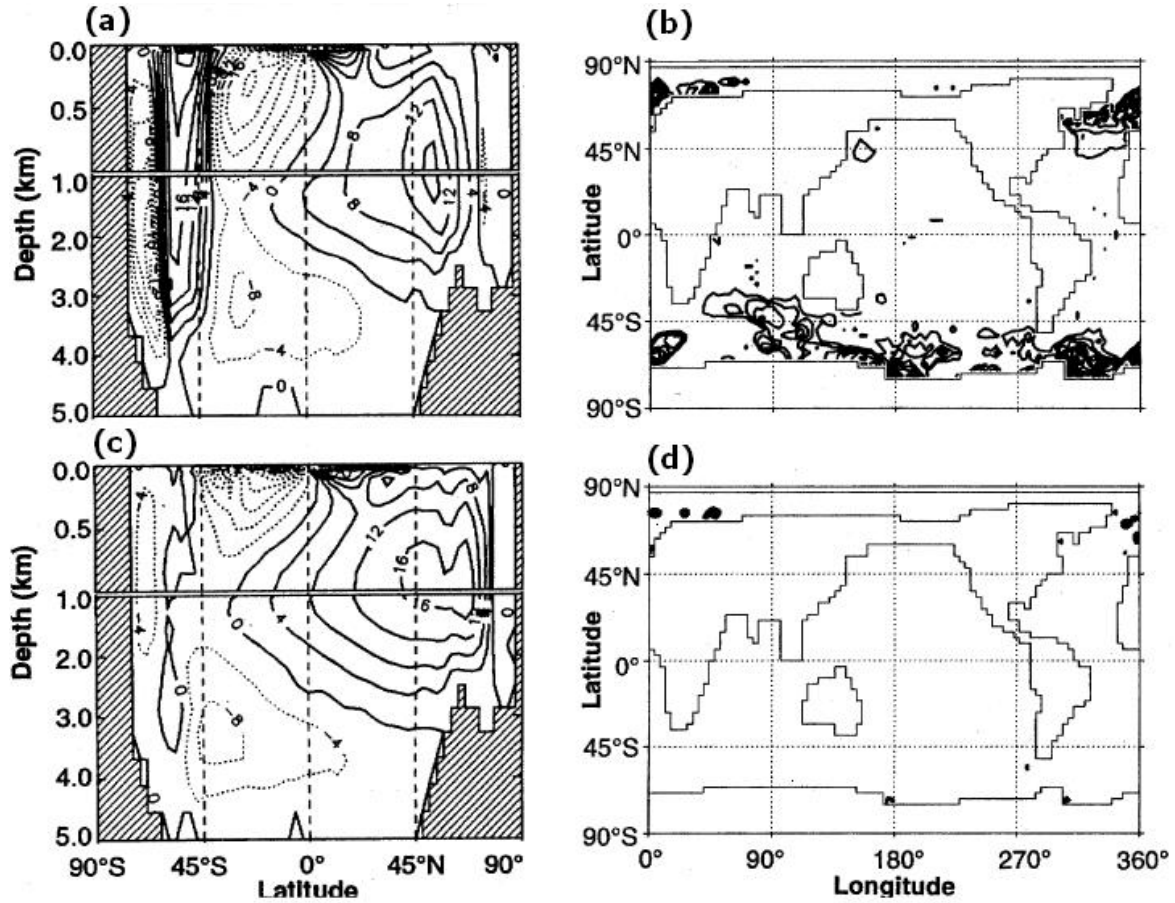


Fig. 1.3: (a,c) Time- and zonal-average oceanic transport stream function (transport between neighbouring streamlines is 4 SV ($\text{SV} = 10^6 \text{ m}^3\text{s}^{-1}$) and clockwise for continuous streamlines), and (b,d) frequency of convective adjustment in oceanic fluid columns (as percentage of all times and vertical levels, contoured every 5%), in two integrations to equilibrium of a coarse-grid ocean model with different eddy parametrizations: In (a,b) eddies stir fluid horizontally, whereas in (c,d) eddies stir fluid along potential density surfaces and moreover themselves transport mass and tracers. (Eddies thus contribute to (c), but not to (a)). From Danabasoglu et al. (1994).

gained by the Earth's equator to its heat-deficient poles. An estimate of this contribution from a near eddy-resolving model (Jayne and Marotzke, 2002) is displayed in Fig. 1.4b (thin curve) and shows that, in the model, the contribution of transients has the same order of magnitude as the total (thick curve) in both equatorial regions and the Southern Ocean.

These modelling examples underline that the transient mesoscale circulation cannot be regarded as simple noise, but instead plays a key dynamical role in shaping time-mean currents and distributions as displayed in Fig. 1.1.

Nonetheless, it is important to note that results as Fig. 1.3 and Fig. 1.4b are highly sensitive to the specific model set-up or the chosen values of the eddy diffusion and transport coefficients. Particularly the degree of compensation between eddy-induced and time-mean transports, which has recently been acknowledged to play a key role in setting the ocean's uptake of CO_2 and its response to changes in atmospheric forcing, varies considerably between models and, as suggested by eddy-resolving models, may not be accurately captured by current eddy parametrizations (e.g. Hallberg and Gnanadesikan, 2006; Farneti et al., 2010; Treguier et al., 2010). From an observational perspective, the relatively few existing studies consistently support an important contribution of transient eddies to poleward heat transport in the SO (deSzoeke and Levine, 1981; Phillips and Rintoul, 2000; Gille, 2003a). These are however also subject to large uncertainties and globally eddy heat transports remain relatively poorly constrained from observations (e.g. Wunsch, 1999). As illustrated in Fig. 1.4a, even the total observed ocean heat transport remains subject to large error bars (Ganachaud and Wunsch, 2003). Given the remaining large uncertainties in both modelling and observational approaches, further systematic observations are thus crucial to gain a better, and more quantitative, understanding of the suggested key roles of eddies in the oceanic circulation and climate.

Transient eddies can drive a time-average poleward heat transport across a given latitude band by moving warm blobs of water across it poleward, and cold blobs equatorward. This is the mechanism through which the eddy-induced transport implied in Fig. 1.3(c) contributes to heat transport: eddies move relatively warm surface water poleward, and relatively cold subsurface water equatorward. Eddies thereby deform the oceanic sphere of relatively light and warm waters, which, as sketched in Fig. 1.5 (white shading), is observed to overlies denser and colder deep waters (hatched) throughout mid- and low-latitudes. Specifically, as sketched by the thin black arrows in Fig. 1.5, the eddy-induced overturning acts to squeeze the light-water sphere towards

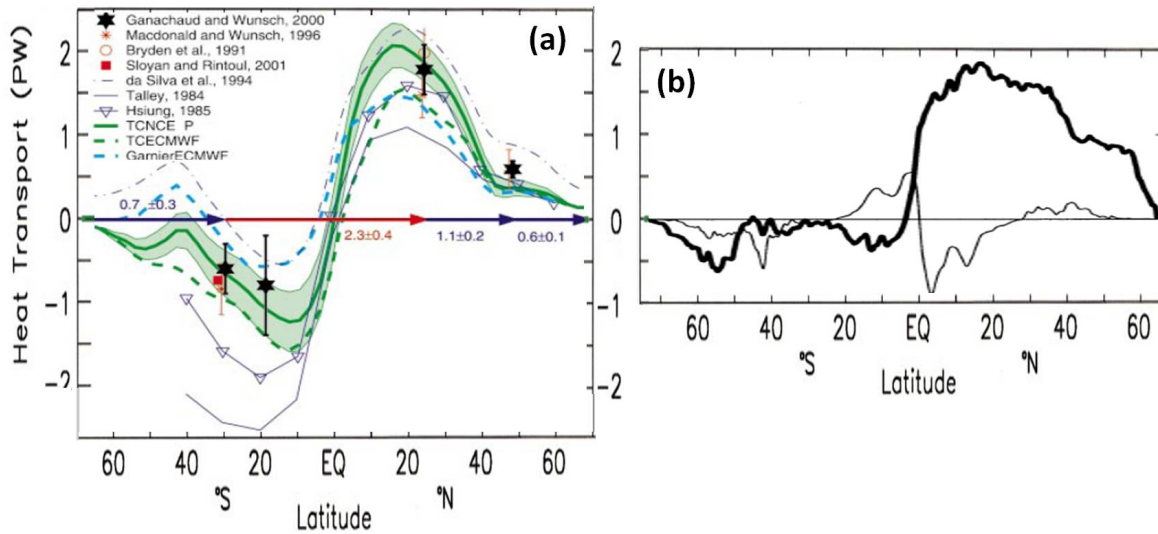


Fig. 1.4: World ocean poleward heat transport in PW ($\equiv 10^{15}$ W), as **(a)** derived from observations using various direct or inverse methods using the global heat balance (described in detail by Ganachaud and Wunsch (2003), note the large error bars), and **(b)** obtained from a $1/4^\circ$ ocean model, in which the modelled contribution to the total (thick curve) by transient motions (thin curve) can be accurately estimated, from Jayne and Marotzke (2002).

the ocean surface, thereby limiting its depth penetration (cooling the deep ocean) and enhancing its meridional extent (warming high latitudes). By this mechanism, upward and poleward eddy heat fluxes thus go hand in hand and the eddy-induced overturning acts to oppose the time-mean overturning, whose effect (sketched by the open black arrows) is to confine the warm water sphere meridionally and push it down to greater depths.

The eddy stirring can thus deform, but not directly drive exchanges of heat between the warm water sphere and its surroundings. Such irreversible heat transports can only be driven by diabatic processes (e.g. Walin, 1982; Tandon and Garrett, 1996). Thus, to close the heat budget of the warm-water sphere, that constantly gains heat from the atmosphere (as sketched by the grey wiggly arrows), deep small-scale turbulence, and eventually, molecular mixing have been evoked (Robinson and Stommel, 1959). By stirring fluid into small-scale gradients, eddies enhance the pace of this process and thereby also play a role in the heat budget of the warm water sphere. More recently the idea has emerged that in the oceanic surface mixed-layer in direct contact with the atmosphere the dissipation provided by possible eddy-driven air-sea interactions may be a far more efficient driver of the required irreversible heat transports.

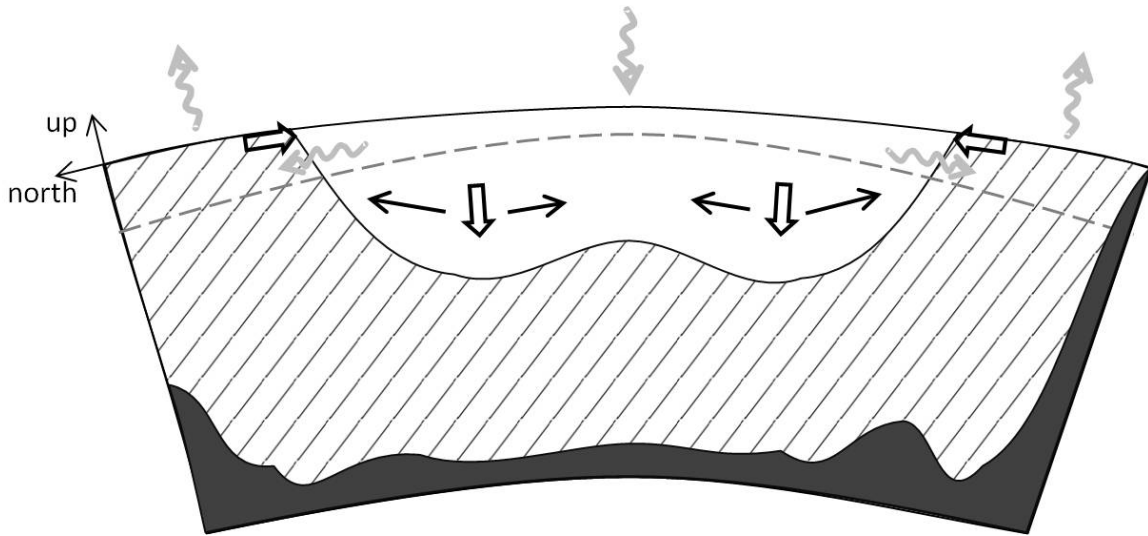


Fig. 1.5: The ocean’s warm and light water sphere (white shading) is enclosed by cooler and denser water (hatched) at depth and, as possible only on a rotating planet, on its poleward sides. As schematized by the black arrows, its shape is set by wind-driven (time-mean) transports (open arrows) and eddy-induced transports (thin arrows). As schematized by the wiggly gray arrows, its heat content can only be modified by diabatic processes: air-sea heat exchanges may thus be the dominant driver of its variations, driving its time-mean heat gain, and, through eddy-induced modulations, its heat loss towards colder waters. Inspired from observations, Marshall et al. (2002) and Greatbatch et al. (2007).

This points to a possibly special role of mixed-layer eddy heat fluxes in the oceanic circulation. Whereas from an observational perspective, eddy transports are poorly characterized even at the surface, hints for an important role of mixed-layer eddy fluxes are provided by theories (e.g. Marshall and Radko, 2003) and eddy-resolving simulations (e.g. Hallberg and Gnanadesikan, 2006). The basic prerequisite for these is a systematic association of eddying circulations with surface thermal anomalies, without which even the most vigorous turbulence could not lead to heat transport. However this association has been questioned by theoretical and modelling results (Bretherton, 1982; Barsugli and Battisti, 1998), which suggest that at the surface any mesoscale thermal anomalies should be quickly eroded by air-sea heat exchanges. This clearly calls for a systematic observational quantification of eddy surface temperature anomalies and their exchanges with the atmosphere, to shed light on the observed role of mixed-layer eddy fluxes and to guide efforts to accurately represent them in

climate models, which is currently a very active area of research (Ferrari et al., 2008; Danabasoglu et al., 2008; Sallee and Rintoul, 2011).

Eddy signals in the sea surface temperature (SST) field are also of interest in their own right. Indeed, apart from purely frictional effects, they determine what the atmosphere knows about the ocean. Does the atmosphere merely see a mirror image of its own time variability overlaid on a smooth large-scale SST field and thus experience a “laminar” ocean at its lower boundary, or does it instead know about the vigorous mesoscale turbulence that dominates the kinetic energy in the ocean? This question is of particular interest given the recent discovery of ocean-atmosphere coupling over mesoscale features in the time-mean SST field in midlatitudes, where on larger spatial scales the ocean is thought to respond to zero-order passively to forcing by atmospheric variability (e.g. Kushnir et al., 2002). Narrow mesoscale fronts in the time-mean SST field, an example of which is seen by the contours in Fig. 1.6a, are not only strongly coupled to the atmosphere’s lower boundary layer (Sweet et al., 1981; Chelton et al., 2004; Small et al., 2008), but the impact of this coupling extends far higher into the tropopause, where it has been shown to have a clear impact on the occurrence of deep atmospheric convection and associated rain patterns (Minobe et al., 2008, 2010; Czaja and Blunt, 2011). Indeed, modelling results show that the rain band “anchored” by the SST front in Fig. 1.6a,b disappears in its absence Fig. 1.6c (Minobe et al., 2008). Whether transient mesoscale eddy signals in the SST field are systematic, or not, thus determines whether there is scope for transient midlatitude ocean-atmosphere coupling, as well.

A systematic observational characterization of the SST signatures associated with *transient* mesoscale eddies, complementing previous regionally scattered observations, is therefore clearly outstanding – as a key piece in the puzzle of understanding the role of oceanic turbulence in the climate system (through mixed-layer heat transports and mesoscale ocean-atmosphere interactions), and moreover as observational benchmark against which to test the realism of both eddy-resolving ocean-only models and coupled models that today move towards mesoscale resolution, as well as to guide new parametrizations of eddy heat transport and midlatitude ocean-atmosphere interactions in climate models.

In this thesis, we aim to provide such an estimate by taking advantage of global satellite observations of (balanced) surface motions and SST, whose accuracy and time-space resolution only recently became adequate for studies of the oceanic mesoscale.

In particular, we wish to provide an observational perspective on the following key

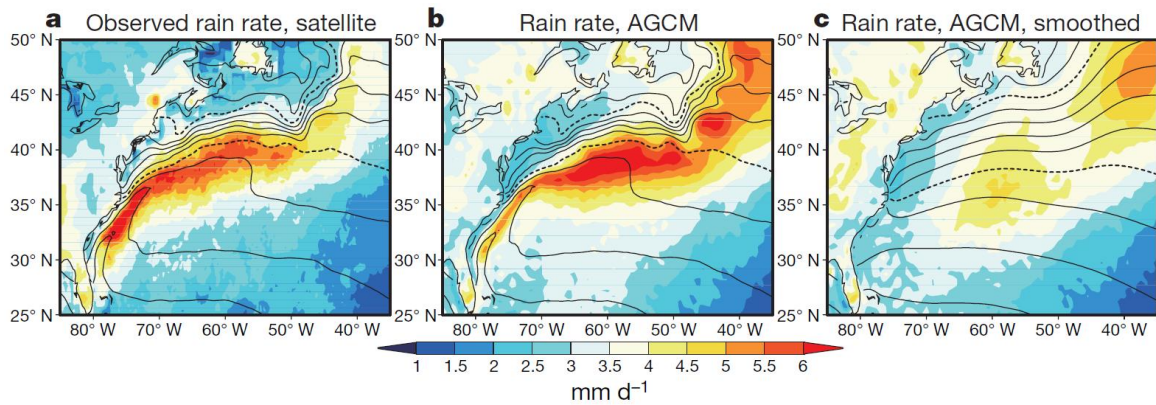


Fig. 1.6: (a) A satellite-based climatology of rain-rate (colours) and SST for the North Atlantic, from Minobe et al. (2008). (b) and (c) show the rain-rate modelled by an atmospheric general circulation model, forced with (b) the observed and (c) a smoothed SST field.

questions:

- Are transient mesoscale motions associated with systematic SST signals, and do these make an important contribution to the observed variability of sea surface temperature? (This will be discussed in chapter 4.)
- How quickly do eddies lose their SST signals to the atmosphere? (This will be addressed in chapter 5.)
- What characterizes eddy SST signals and what can we learn from them about mixed-layer eddy heat transports and their mechanisms? (This will be addressed in chapter 6.)

The original aspect of this thesis is thereby the combination of two contrasting observational view-points: on the one hand, we will address the impact of eddies as observed at any given fixed location in space across which eddies move, and complement this, on the other hand, by establishing the eddying signal that is observed when travelling with any individual eddy. Here the latter perspective can be adopted thanks to an automated eddy tracking algorithm that has been recently developed by Chelton et al. (2011b, hereafter referred to as CSS11). Together with the observational datasets, this will be described in chapter 3.

Although observations are available and the analysis is readily extended across the whole globe, here we propose to focus on North Atlantic (NA) and Southern Ocean (SO) basins, as they feature two of the world's most vigorous currents and associated eddy fields (the Gulf Stream (GS), and the ACC), which not only provide interesting

examples for two contrasting dynamical set ups (a western boundary current, and a circumpolar current in a channel), but moreover both play a fundamental role in the climate system: whereas the ACC connects not only the three world oceans with each other but also the deep ocean to the atmosphere, over the GS a large amount of heat carried northward by the ocean in the Southern Hemisphere and at lower Northern Hemisphere latitudes is transferred to the atmosphere, where it fuels the development of atmospheric eddies in the midlatitude storm tracks.

Before commencing, chapter 2 provides some background on the dynamics of sea surface temperature anomalies and introduces one of the main mechanisms through which eddies are generated, impact the ocean's mean state and transport heat.

Chapter 2

BACKGROUND

This chapter introduces the basic physical principles underlying the driving questions of this thesis. Section 2.1 provides an introduction to the generation of mesoscale eddies and their mutual relationship with the observed large-scale circulation in NA and SO, section 2.2 an overview of the ocean-atmosphere boundary layer system and the dynamics of sea surface temperature.

(If not otherwise cited the material presented in this chapter is either my own or can be found in textbooks of atmosphere and ocean dynamics, such as Marshall and Plumb (2008), Cushman-Roisin (1994) or Gill (1982).)

2.1 The generation of mesoscale eddies

2.1.1 The energy reservoir of the oceanic thermocline

Figure 2.1 maps a five-year average of the height of the sea surface (SSH) above the Earth’s geoid, observed in NA and SO. If the ocean were at rest, its water masses would be distributed in a way that minimizes potential energy. The ocean surface would then coincide with a level surface of equal geopotential (the geoid), everywhere normal to the effective acceleration due to gravity, which includes centrifugal forces due to the Earth’s rotation. The non-zero values of SSH or η (measured along the local vertical $\mathbf{k}$ with respect to the geoid at $z = 0$) seen in Fig. 2.1 thus reflect departures from this state of rest. As their amplitude varies by less than 2 m over the global ocean, they are challenging to observe amid the much larger undulations observed in the geoid ($\mathcal{O}(100\text{m})$). Chapter 3.2 introduces how this can be done (and the specific dataset used in Fig. 2.1 and our confidence in it are further discussed in Appendix B.4).

On the scales resolved in Fig. 2.1, and down to a few tens of kilometres, the moving ocean is to a very good approximation (e.g. Pedlosky, 1987) in the same equilibrium as static fluids, the hydrostatic equilibrium, in which the increase of pressure p with

depth is simply given by the (incremental) weight of the fluid column:

$$-dp = \rho g dz. \quad (2.1)$$

Ignoring atmospheric pressure variations¹, deflections of the sea surface η thus correspond to changes in ocean pressure (at $z = 0$):

$$\delta p|_{z=0} = \rho_0 g \delta \eta, \quad (2.2)$$

in which $\rho_0 \sim 1000 \text{ kg/m}^3$ is an average density, and their spatial gradients accelerate oceanic motions ($\mathbf{u}$), as of the horizontal momentum equation for shallow fluids (such as ocean and atmosphere) on a rotating planet:

$$d_t \mathbf{u} + f \mathbf{k} \wedge \mathbf{u} = -\nabla p / \rho_0 + \mathbf{F}. \quad (2.3)$$

In this, $d_t = \partial_t + \mathbf{u} \cdot \nabla + w \partial_z$, where w is upward flow, denotes the time derivative following a fluid parcel (Lagrangian derivative), $f \mathbf{k} \wedge \mathbf{u}$ is the horizontal component of the Coriolis acceleration ($f \equiv 2\Omega \cdot \mathbf{k} = 2\Omega \sin \phi$, in which Ω is the Earth's angular velocity and ϕ latitude), ∇ is the horizontal gradient operator and $\mathbf{F}$ are frictional forces per unit mass.

For large-scale ($L \sim 1000 \text{ km}$) and mesoscale ($L \sim 100 \text{ km}$) flows at midlatitude ($f \sim 10^{-4}$), the first term on the lhs of (2.3) is typically at least one order of magnitude smaller than the second. Their ratio $d_t \mathbf{u} / f \mathbf{k} \wedge \mathbf{u} \sim U^2 / L / f U \approx U / f L$, termed Rossby number Ro , only approaches unity if typical flow speeds U reach $fL \geq 10 \text{ m/s}$, which is much larger than observed ($U \sim 1 \text{ cm/s}$ to 1 m/s). Currents driven by pressure gradients are thus to zero order geostrophic:

$$\mathbf{u}_g = \mathbf{k} \wedge \frac{\nabla p}{\rho_0 f}, \quad (2.4)$$

in a close balance between pressure gradient and Coriolis forces. They flow along lines of constant pressure or, at the surface, from (2.2), along isolines of η , as displayed in Fig. 2.1. The high sea surface elevation seen in Fig. 2.1a in the NA subtropics is therefore associated with a basin-scale anticyclonic circulation (opposing the Earth's rotation projected onto the local vertical, or clockwise in the Northern Hemisphere (NH)), referred to as subtropical gyre. The low sea surface observed at higher latitudes is associated with the cyclonic flow of the subpolar gyre. In gyre interiors, sea surface slopes are gentle, typically $\sim 10 \text{ cm}$ over a distance L of $\sim 10^\circ$. Surface currents $u_g = \frac{g \delta \eta}{f L}$

¹ or accounting for them by an inverse barometer correction to η , as described in chapter 3.2.2

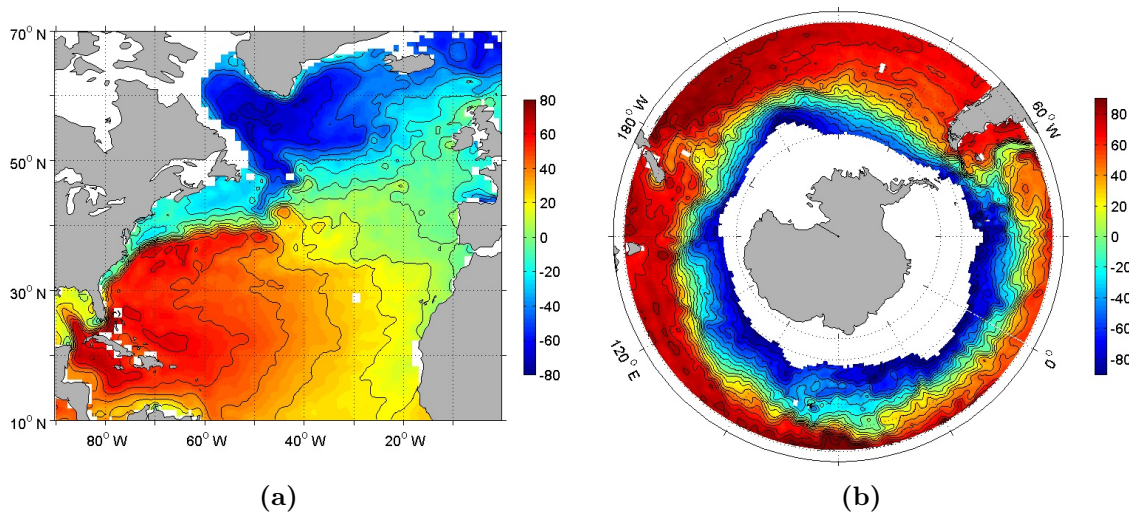


Fig. 2.1: 5½-year (June 2002–December 2007) average sea surface height η , in cm, obtained by merging satellite altimeter (cf. chapter 3.2), surface drifter and in-situ temperature and salinity observations, as described in Appendix B.4. The contour interval is 10 cm and basin means are removed.

$\sim 1 \text{ cm s}^{-1}$ are thus slow. Towards the western side of the NA basin η contours bunch together, and the sea surface can easily drop by 50 cm within 1° . The associated fast flows ($\sim 50 \text{ cm/s}$) are typical for the NA western boundary currents (the poleward flowing Gulf Stream (GS) and the equatorward flowing Labrador current) and their zonal extensions. In the SO (Fig. 2.1b), the sea surface drops by more than 1 m from the Southern Hemisphere’s subtropical gyres towards Antarctica, and this step-like drop supports the fast and narrow jets of the Antarctic Circumpolar Current (ACC).

If the ocean were a homogeneous fluid, the surface flow patterns indicated by (2.4) and Fig. 2.1 would extend to the sea floor without modification (the flow would be depth-independent, or barotropic). The ocean is however observed to be stratified, with light and warm water overlying cold and dense water. As seen in Fig. 2.2, surfaces of constant potential density², referred to as isopycnals, are most tightly spaced in the top $\sim 1 \text{ km}$ of the water column, the ocean’s pycnocline (or thermocline, as density variations are mainly controlled by temperature – salinity plays a role mainly in cold high-latitudes). Fig. 2.2 also shows that, in this thermocline, isopycnals are not purely horizontal, but deflect from level surfaces. The resulting horizontal density variations drive depth variations in the geostrophic flow. As seen by taking ∂_z (2.4) and using

² Potential density is the density of a parcel with measured salinity S and temperature measured after its adiabatic lift to the surface, and is conserved by a parcel in adiabatic flow.

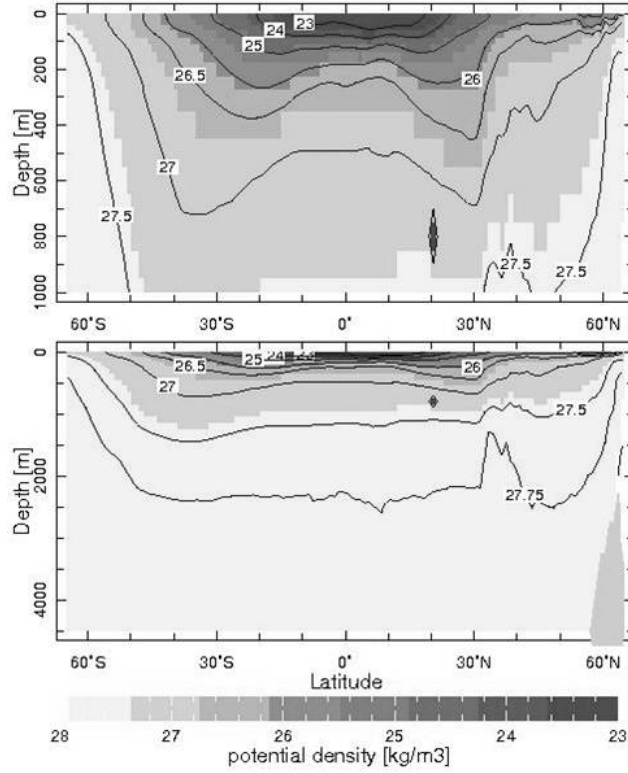


Fig. 2.2: Zonal-average of the climatological potential density (as deviation from 1000, in kg/m^3), referenced to the surface, as given by the Levitus World Ocean Atlas 1994 (<http://iridl.ldeo.columbia.edu/>).

(2.1), this baroclinic flow, or geostrophic shear:

$$\partial_z \mathbf{u}_g = -\mathbf{k} \wedge \frac{g \nabla \rho}{\rho_0 f}, \quad (2.5)$$

approximately follows density (or temperature) contours on level surfaces, and is therefore also referred to as “thermal wind”.

Inspired by Fig. 2.2, it is tempting to simplify the ocean for a moment as a 2-layer fluid, as depicted in Fig. 2.3: a light layer of density ρ_1 and depth $\eta + h$, representative of the thermocline, overlies a dense layer (of density ρ_2), with a density contrast $\Delta\rho = \rho_2 - \rho_1$ of typically 3 kg/m^3 , as evaluated from Fig. 2.2.

Comparing Fig. 2.2 and Fig. 2.1 reveals that density surfaces tend to deflect downwards ($\delta h > 0$) in regions of high SSH ($\delta\eta > 0$) such as the subtropical gyres, and vice versa in the subpolar gyres. Integrating (2.1) from the sea surface to a depth $z \ll -h$ shows that these isopycnal deflections tend to cancel out pressure gradients below the thermocline:

$$\frac{\nabla p_2}{\rho_0} = g \nabla \eta - g \frac{\Delta\rho}{\rho_0} \nabla h. \quad (2.6)$$

They do so exactly, if $\delta h = \frac{\rho_0}{\Delta\rho}\delta\eta$, that is if thermocline deflections are $\approx 1000/3$ times larger than deflections in η . On gyre scales ($\delta\eta \sim 50$ cm, from Fig. 2.1) this corresponds to $\delta h \sim 150$ m (and is just what we observe in Fig. 2.2).

In that case, large-scale geostrophic flows are restricted to the thermocline, and their speed scales as $U = \frac{g}{f} \frac{\Delta\rho}{\rho_0} \frac{\delta h}{L}$, which is ≈ 3 cm/s for the above parameters and a gyre-scale L over which h deflects of ≈ 1000 km. The associated average kinetic energy of an oceanic column

$$KE \equiv \overline{\int_{-H}^{\eta} \frac{\rho}{2} \mathbf{u} \cdot \mathbf{u} dz} \approx \frac{\rho}{2} \overline{U^2 h} \approx \frac{g\Delta\rho}{2} \overline{\delta h^2} \left(\frac{L_d}{L} \right)^2, \quad (2.7)$$

where over-bars denote area-averaging, is ≈ 500 J/m², and can be expressed in terms of the ratio between L and the deformation scale of the 2-layer system $L_d = \frac{\sqrt{g(\Delta\rho/\rho_0)h}}{f}$ (≈ 50 km).

The potential energy ($PE = \overline{\int_{-H}^{\eta} \rho g z dz}$) available by flattening the interface between the two layers (i.e. going from Fig. 2.3b to a) is given by:

$$APE = PE(\delta h, \delta\eta) - PE(\delta h = \delta\eta = 0) \approx \frac{g\Delta\rho}{2} \overline{\delta h^2} + \frac{g\rho_0}{2} \overline{\delta\eta^2} \approx \frac{g\Delta\rho}{2} \overline{\delta h^2}. \quad (2.8)$$

In purely baroclinic flows, the available potential energy thus exceeds the kinetic energy by a factor $\left(\frac{L}{L_d} \right)^2$, which on gyre-scales amounts to typically $\left(\frac{1000 \text{ km}}{50 \text{ km}} \right)^2 = 400$, a striking result perhaps first derived by Gill, Green and Simmons (1974, GGS in the following). The observed bowl- and dome-shaped thermocline deflections of oceanic gyres and the ACC thus represent a very large reservoir of available potential energy ($\sim 10^5$ J/m²) that is typically a few hundred times larger than the kinetic energy of the associated baroclinic currents.

2.1.2 Energy source

This energy reservoir is generated and constantly replenished by the stress ($\boldsymbol{\tau}$) exerted by atmospheric surface winds on the (moving) ocean surface.

A climatology of $\boldsymbol{\tau}$, as observed by satellite scatterometry (Appendix B.3), is displayed by the arrows in Fig. 2.4. On basin scales, atmospheric surface wind speeds $|\mathbf{u}^a|$ (~ 10 m s⁻¹) are roughly two orders of magnitude larger than the observed gyre-scale ocean surface current speeds $|\mathbf{u}|$, and $\boldsymbol{\tau}$ therefore reveals the signature of the main atmospheric wind systems. Atmospheric boundary layer bulk formulae relate

$$\boldsymbol{\tau} = \rho^a c_D |\mathbf{u}^a - \mathbf{u}| (\mathbf{u}^a - \mathbf{u}) \quad (2.9)$$

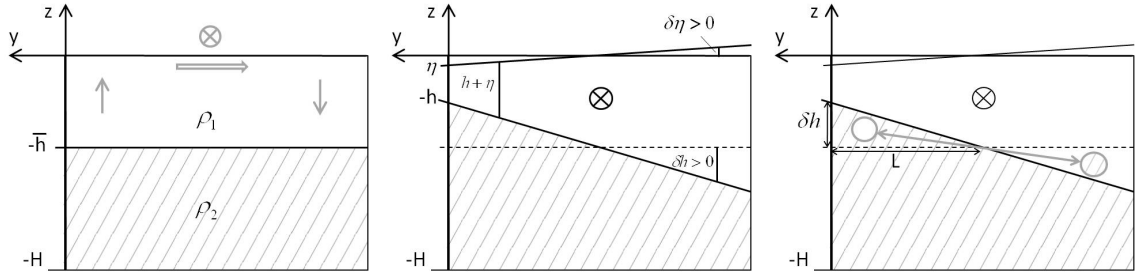


Fig. 2.3: A two-layer ocean: **(a)** in a state of rest, and **(b)** with only the upper layer in a balanced geostrophic motion U (into the page). The grey symbols moreover depict in **(a)**: how a surface wind stress τ (into the page) inputs APE into the thermocline (via an equatorward Ekman transport V_{ek} and ensuing Ekman suction $w_{ek} > 0$ on its poleward, and Ekman pumping $w_{ek} < 0$ on its equatorward flanks, leading to the isopycnal deflection in b), and in **(c)**: how this APE is transformed into eddy energy through slantwise convection/baroclinic instability, which leads to poleward and upward “heat” transport. (Figure applies to NH, y points poleward, otherwise notation as defined in text.)

to the air-sea speed difference via a drag coefficient c_D , for which a range of formulations exist to express its dependence on wind-speed and stratification (stability) (e.g. Large et al., 1994; Fairall et al., 2003). Taking typical values of air-density ρ^a as 1 kg m^{-3} and c_D as 10^{-3} , then yields a typical wind stress magnitude of 0.1 N m^{-2} , as observed. In the NA (Fig. 2.4a), mid-latitude westerlies overlie the GS and its north-westward extension, the NA drift (as indicated by the SSH contours), and low-latitude trade winds overlie the westward return flow of the subtropical gyre. In the SO (Fig. 2.4b), vigorous SH westerlies overlie the ACC. This shows that the wind stress, via the frictional force it exerts below the ocean surface, transfers kinetic energy from the large reservoir of atmospheric winds to the gyre scale geostrophic circulation: $\partial_t KE = \overline{\tau \cdot \mathbf{u}} > 0$ (as obtained from (2.7) and (2.3) with $\mathbf{F} = \frac{1}{\rho_0} \frac{\partial \tau}{\partial z}$).

However, if the wind is to drive the observed large-scale thermocline circulation, it is not enough if it imparts kinetic energy to the ocean. It also has to set up, the intimately related, but – as shown above – typically more than 100 times larger available potential energy of the basin-scale baroclinic flow.

To see how this is achieved, let’s go back to the momentum equation (2.3). Wind-induced frictional forces generate a turbulent ocean surface boundary layer (with a depth on the order of 100 m), the Ekman layer, in which, for low Rossby number

flow, the vertically-integrated (ageostrophic) horizontal transport $\mathbf{U}_{ek}$ (in m^2s^{-1}) is directly related to the surface wind stress as

$$\mathbf{U}_{ek} = -\mathbf{k} \wedge \frac{\boldsymbol{\tau}}{\rho_0 f}. \quad (2.10)$$

Forced by the wind-stress curl, horizontal convergences and divergences of this Ekman transport $\mathbf{U}_{ek}$ generate vertical velocities at the base of the Ekman layer

$$w_{ek} = \boldsymbol{\nabla} \cdot \mathbf{U}_{ek} = \boldsymbol{\nabla} \wedge \frac{\boldsymbol{\tau}}{\rho_0 f} \quad (2.11)$$

that transmit information on atmospheric winds to the oceanic interior.

A climatology of w_{ek} is coloured in Fig. 2.4. Interestingly, w_{ek} acts like a magnifying glass for the small-scale features present in $\boldsymbol{\tau}$. Contrary to basin-scales, these are the signature of vigorous *oceanic* jets and also of mesoscale structures in the time-mean SST field, as only recently discovered (e.g. Chelton et al., 2004). Their appearance in $\boldsymbol{\tau}$ reflects the frictional damping of the former, and, as shown in chapter 5, the thermal damping of the latter, at the sea surface. The basin-scale patterns of w_{ek} show Ekman pumping ($w_{ek} < 0$) throughout the subtropical gyre of the NA, which results from the convergence of equatorward V_{ek} driven by mid-latitude westerlies and poleward V_{ek} driven by the trade winds. Poleward of the main westerlies Ekman transports diverge and lead to Ekman suction ($w_{ek} > 0$) in the subpolar gyre. The typical magnitude of w_{ek} of $\approx 40\text{m/year}$ corresponds to an Ekman transport $V_{ek} = \tau/\rho_0 f \approx 1\text{ m}^2\text{ s}^{-1}$ (as driven by a typical midlatitude wind stress of 0.1 N m^{-2}) that converges on gyre scales, so that $w_{ek} \approx V_{ek}/1000\text{ km} \approx 10^{-6}\text{m/s}$, as observed.

Whereas the possible role played by the small-scale features observed in w_{ek} is a question of active research, its gyre-scale pattern are well known to play a fundamental role in the wind-driven circulation of the NA thermocline. To see this consider that in a stratified and near-adiabatic ocean any forced vertical motions, such as w_{ek} , will act to either push light surface waters downward or pull dense bottom waters towards the surface. As illustrated in Fig. 2.3a,b, subpolar Ekman suction thus lifts and subtropical pumping lowers the thermocline, both increasing the magnitude of thermocline deflections $\overline{\delta h^2}$ and thereby, from (2.8), the available potential energy of the system. As seen in Fig. 2.4b, the same mechanism is observed to be at work in the SO. SH westerlies drive intense equatorward Ekman transports that diverge poleward and converge equatorward of the ACC. They thereby set up the cross-ACC pressure gradient and, via the associated upward and downward w_{ek} , the thermocline tilt that maintains the ACC. (By bringing up oceanic bottom water in contact with

the atmosphere, here w_{ek} also plays a fundamental role for the global, deep meridional overturning circulation.) Wind-driven Ekman pumping and suction thus generates the observed thermocline slopes and the associated (kinetic and) potential energy reservoir. From Fig. 2.4, $w_{ek} \approx 50$ m/year, and from our scaling analysis in section 2.1.1, $\delta h \approx 150$ m (cf. also Fig. 2.2). Departing from a state of rest, it would thus take ≈ 3 years for the wind to build up the observed APE and associated circulation in the wind-driven thermocline (cf. GGS for a more thorough derivation of this time-scale considering the energy balance).

Moreover, consider taking $-f\nabla \cdot \mathbf{u}_g$ (2.4), while noting that $\frac{df}{dy} = \frac{2\Omega \cos(\phi)}{R_{earth}} = \beta$ and using the continuity equation for near-incompressible flow ($\nabla \cdot \mathbf{u} + \partial_z w = 0$), to obtain the geostrophic vorticity equation

$$-f\nabla \cdot \mathbf{u}_g = f\partial_z w = \beta v_g. \quad (2.12)$$

Integrating (2.12) from just below the Ekman layer to the sea floor, neglecting bottom friction and slopes, yields

$$fw_{ek} = \beta V_g. \quad (2.13)$$

This shows that w_{ek} drives gyre-scale geostrophic transports³, equatorward in the subtropical gyre ($w_{ek} < 0$) and poleward in the subpolar gyre, and thereby sets the wind-driven gyres into motion. (The required return flow can, also from vorticity arguments, only occur on the western boundary of the basin, which gives rise to the observed intense western-boundary currents (WBCs).) As introduced by Luyten et al. (1983), water parcels that experience Ekman pumping are advected down- and equatorward, so that their potential vorticity, which has been set at the sea surface, remains constant after their subduction. This wind-forced “ventilation” thereby plays an important role in setting the density structure of the thermocline.

2.1.3 Energy extraction

In a quasi-steady state, the APE input driven by the wind needs to be balanced by a process that extracts APE from the mean thermocline by moving light waters back up and dense waters back down.

Whereas in non-rotating fluids this release would be straightforward (and driven by pressure gradients), on the rotating Earth the Coriolis force stabilizes the observed isopycnal slopes (via (2.5)). There is nevertheless an efficient mechanism to extract the

³ For reference, adding βV_{ek} to (2.13) yields the Sverdrup (1947) flow $\beta V = \nabla \wedge \frac{\tau}{\rho_0}$.

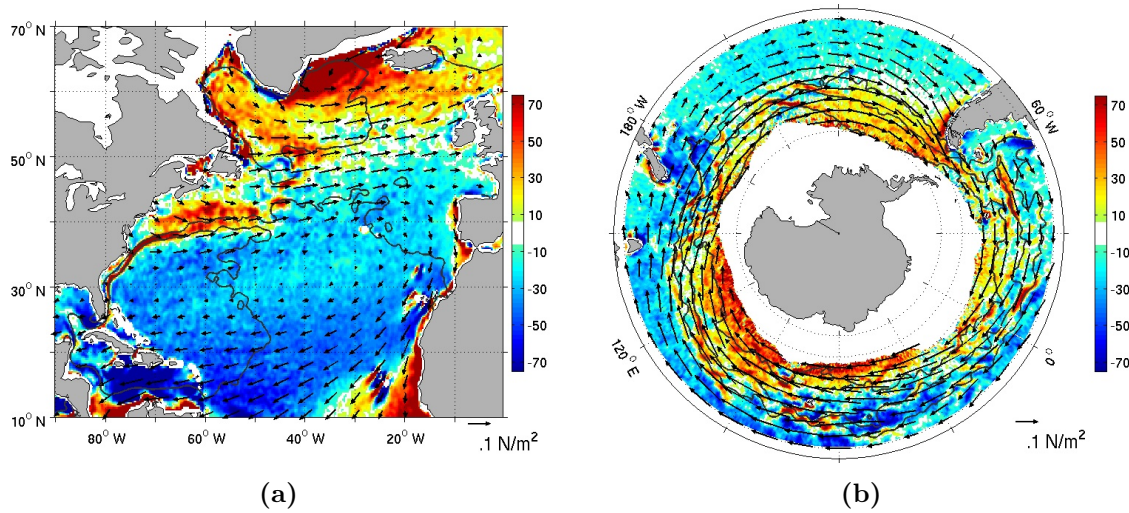


Fig. 2.4: Wind stress τ (black arrows) and Ekman “pumping” w_{ek} (coloured, in m/year and positive upward!), derived from the QuickSCAT-based SCOW climatology (cf. Appendix B.3). Characteristic SSH contours (-45, 5 and 45 cm in the NA, -55, 0 and 40 cm in the SO) are repeated from Fig. 2.1.

APE, but very special conditions need to be met for it to occur. As in Fig. 2.3b light fluid still everywhere overlies dense fluid, convection cannot do the job. However if random disturbances do not lift light parcels vertically, but slantwise, and specifically if they do so, as depicted in Fig. 2.3c, at a shallower angle with respect to the horizontal than that formed by background isopycnals, light parcels will rise into denser fluid and thus continue to rise. Conversely, at the same angle, dense parcels will be lowered into lighter fluid and continue to sink. Random disturbances that meet this specific condition, expressed as:

$$0 < \frac{w'}{v'} < \frac{\delta h}{L} \quad (2.14)$$

where w denotes vertical, v poleward flow and primes disturbances, will thus grow by extracting their energy from the APE stored in the sloping background isopycnals. They thus act to flatten isopycnals towards a state of reduced background energy by transporting “heat” poleward and upward⁴. Clearly, the window for their growth is opened by the isopycnal slopes, the larger these are the more they are subject to this baroclinic instability (a barotropic flow ($\delta h = 0$) is stable with respect to this type of instability).

Here it is important to remember that disturbances are constrained by the gov-

⁴ Heat transport, if potential density layering is temperature controlled, buoyancy ($b = -g\rho/\rho_0$) transport if salinity matters.

erning equations. In particular, in the absence of non-conservative processes, they are required to conserve potential vorticity (PV), which – as shown e.g. by Ertel (1942) – is given by the absolute vorticity (including planetary and relative contributions) on isopycnal surfaces weighted by their thickness $H_\rho = \Delta\rho/|\nabla\rho|$, in which $\Delta\rho$ is the density contrast between neighbouring isopycnals.

A simple model of baroclinic instability

Perhaps the simplest set-up prone to this baroclinic instability has been put forward by (Eady, 1949), and consists of a constant zonal shear $\bar{u}_z = U/H$ in thermal wind balance with a thus constant meridional density gradient ρ_y that is embedded in constant stratification ρ_z between two rigid lids at $z = -H$ and 0 and in constant planetary vorticity $f = f_0$. Eady moreover assumed incompressible, hydrostatic flow on scales L small compared to the radius of the Earth and characterized by $Ro \ll 1$. In this quasi-geostrophic framework, the perturbation PV is given by

$$q' = \frac{1}{\rho_0 f_0} (\nabla^2 p + \frac{H^2}{L_d^2} p_{zz}), \quad (2.15)$$

as shown e.g. by Cushman-Roisin (1994). As in this specific set-up the background flow has zero relative vorticity and features constant isopycnal slopes, the background PV = f_0/H_ρ is constant. The initially zero q' of disturbances thus has to remain 0 throughout their evolution. This requires that relative vorticity and stretching contributions to q' (1st and 2nd terms in (2.15)) have comparable magnitudes, which can only be achieved if horizontal and vertical scales of disturbances, L_e and H_e , scale as:

$$\frac{L_e}{L_d} \sim \frac{H_e}{H}. \quad (2.16)$$

We can now derive a scaling for $\frac{w'}{v'}$ considering the vorticity balance, which in this set-up simplifies to: $(\partial_t + U\partial_x)\zeta' = -f_0\partial_z w'$, where ζ is relative vorticity. The two sides are comparable only if $U\frac{v'}{L_e^2} \sim f_0\frac{w'}{H_e}$. Using (2.16) and noting that $U \sim f_0\frac{L_d^2}{H}\frac{\delta h}{L}$, we obtain $\frac{w'}{v'} = \frac{UH_e}{f_0L_e^2} = \left(\frac{L_d}{L}\right)^2 \frac{\delta h}{L}$. The condition for instability (2.14) therefore reads:

$$0 < \left(\frac{L_d}{L_e}\right) < 1, \quad (2.17)$$

that is disturbances will only grow if they are larger, but not much larger than L_d . Disturbances smaller than L_d would be constrained to move steeper than isopycnals and are thus stable, while those with $L \gg L_d$ move nearly horizontally and therefore hardly lift any light (or lower dense) fluid.

This scaling analysis closely reflects the analytic solution of the instability (e.g. Eady, 1949; Gill, 1982). This shows that the fastest growth occurs indeed for $L_e \geq L_d$, and does so at a rate $\omega_{\text{eady}} = 0.31 \frac{U}{L_d}$, which clearly highlights that the background baroclinic shear with its associated isopycnal slopes and available potential energy is at the origin of the instability.

A crucial role in this solution is played by the boundary conditions $w(z = -H, 0) = 0$. In adiabatic conditions, the buoyancy $b = -g\rho/\rho_0$ of a water parcel is conserved $d_t b = 0$. As $w = 0$, at top and bottom boundaries buoyancy is conserved following the (horizontal) geostrophic flow only. In the chosen set-up this implies:

$$(\partial_t + \mathbf{u}_g \cdot \nabla)b = (\partial_t + \bar{u}\partial_x)b' + v'_g \bar{b}_y = 0, \quad (2.18)$$

supporting the propagation of boundary waves along $\bar{b}$ -contours. As illustrated in the rhs panels of Fig. 2.5, temperature ($\sim -b$) and pressure of such perturbations, which propagate westward (eastward) with respect to the mean flow, are in phase (out-of-phase) at the top (bottom) boundaries. Considering e.g. the top boundary, the heat transported poleward on the western side of pressure cores is thus returned equatorward on their eastern side, resulting in no net heat transport. However, once $L_e \geq L_d$, from (2.16), the pressure perturbations induced by T' decay away so slowly from the respective boundaries that their associated v'_g perturbations contribute to the advection of T' at the opposing boundary. As illustrated in the lhs panels of Fig. 2.5, this induces small shifts in phase in warm anomalies, to the west with respect to surface highs, and to the east with respect to bottom lows (their magnitude is typically $\leq$ half the radius of pressure cores, given by a quarter perturbation wavelength $r = \lambda/4$). Contrary to the case of stable waves (rhs panels), v'_g and T' are thus no longer in quadrature, so that unstable waves transport heat towards colder water (poleward in the chosen set up) and release the potential energy stored in the background shear.

Baroclinic instability in the ocean

In the ocean, the stratification and shears are strongly surface intensified, so that wave motions on temperature contours near the sea floor are not available for the instability. However in the interior ocean, contrary to the Eady model, the PV of the background state is not constant as both planetary vorticity f ($df/dy = \beta$) and isopycnal thickness H_ρ vary, supporting planetary wave motions on contours of f/H_ρ (denoted planetary PV (PPV) in the following).

As shown by a recent observational study (Tulloch et al., 2011), in the extratropics

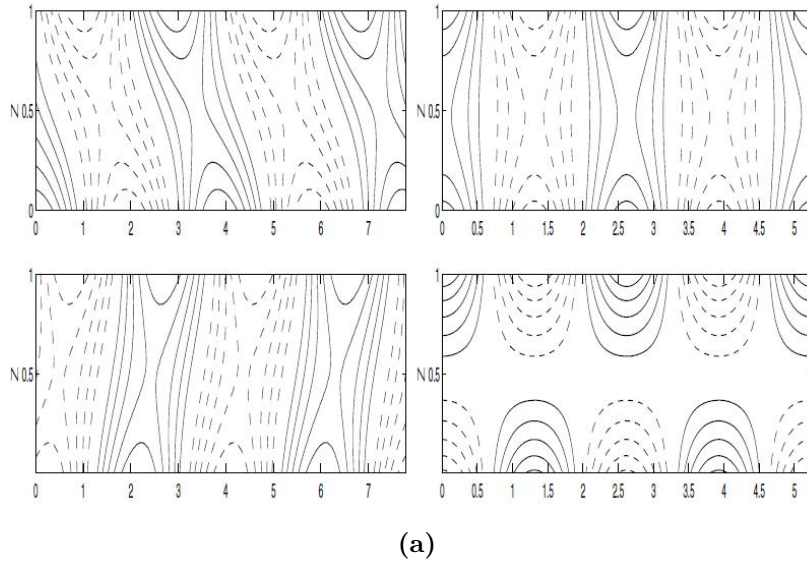


Fig. 2.5: Pressure (top panels) and, hydrostatically related, temperature perturbations (bottom panels) of stable (lhs panels) versus unstable disturbances (rhs panels) in the Eady model. Solid contours denote positive, dashed contours negative values, the abscissa is along the mean shear, and the ordinate points upward, units are normalized. From Vallis (2006).

we observe two prototype set-ups of opposing PPV (or surface density) gradients that lead to wave interaction and instability. As schematized in Fig. 2.6a, in regions of strong surface intensified eastward currents, such as GS and ACC, the large associated isopycnal slopes lead to a reversal of the PPV gradient below the thermocline, where isopycnals thin out rapidly towards the equator. Here the deformation scale relevant to the instability is thus the first baroclinic Rossby radius (of which a dataset, together with its observed spatial distribution, is introduced in Appendix B.6). Most unstable disturbances are therefore deep-reaching and have $L \geq L_d$ (Fig. 2.7b, red shades). The second prototype set-up corresponds to the westward return flows of subtropical gyres. As illustrated in Fig. 2.6b, here shears $\partial_z \mathbf{u}_g$ are westward throughout the thermocline, but reverse to eastward in the top ~ 200 m to support the observed poleward temperature gradient in the seasonal thermocline. Surface isopycnals sloping downward and thermocline isopycnals sloping upward toward the equator thus sandwich between them a rapidly equatorward-thinning isopycnal layer at the level of peak westward currents, on which PPV increases equatorward. Here waves propagating eastward (with respect to mean flow) on the equatorward PPV gradient (at ~ 200 m) interact with westward propagating waves on the poleward SST gradient, and therefore lead to shallow disturbances with $L < L_d$ (Fig. 2.7b, blue shades), as here the relevant

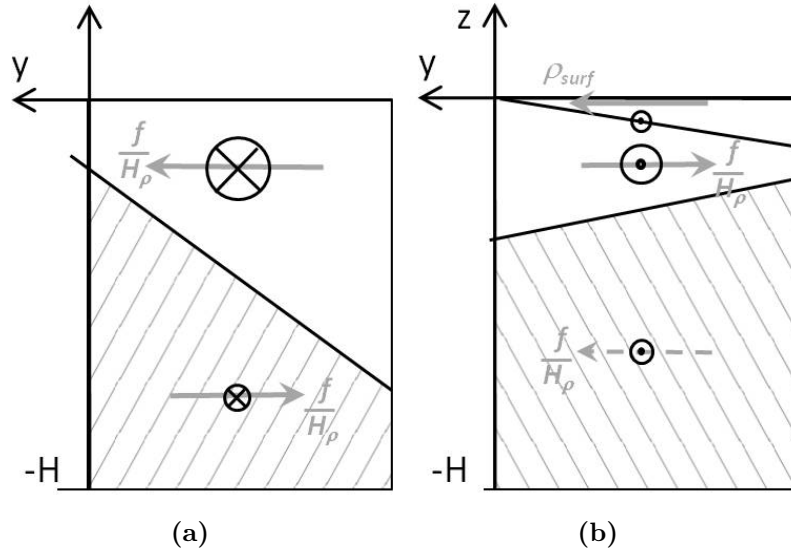


Fig. 2.6: Schematic (appropriate to the NH) of observed planetary potential vorticity gradients (grey arrows), supporting planetary wave propagation to their left, in **(a)** eastward flows such as GS and ACC (crosses) and **(b)** subtropical gyre westward return flows (dots). Lines denote interfaces between potential density layers of thickness H_ρ . (Notation as in text.)

deformation scale corresponds to a higher vertical mode (cf. Appendix B.6).

Conditions for instability are thus met throughout the different typical regimes observed in the ocean, and stability analysis indeed shows the entire ocean to be baroclinically unstable (Tulloch et al., 2011). As suggested by Eady's model, observed growth rates, displayed in Fig. 2.7a, are indeed largest in regions of vigorous shears, in which η bunch together in Fig. 2.1 and isopycnals slope steeply. Here disturbances can e-fold within less than 20 days, whereas in more quiescent gyre interiors they typically take ~ 100 days.

The transfer of the (potential) energy stored in the ocean thermocline to energy of eddies via baroclinic instability is thus indeed expected to be a key process to balance the energy input by the wind. If they balance exactly, and following the argument of GGS, within the spin-up time of wind-driven circulation (on the order of 3 years, section 2.1.2), the eddy energy would approach the $\overline{APE}$ of the background flow⁵. For eddies with $L_e \geq L_d$, nearly half of this is in the form of eddy kinetic energy (EKE). In this case: $EKE \leq \frac{1}{2}\overline{APE} \approx \frac{1}{2} \left(\frac{L}{L_d} \right)^2 \overline{KE}$. The kinetic energy of eddies could

⁵ Obviously the rate at which eddies are dissipated would also need to be considered in a full treatment.

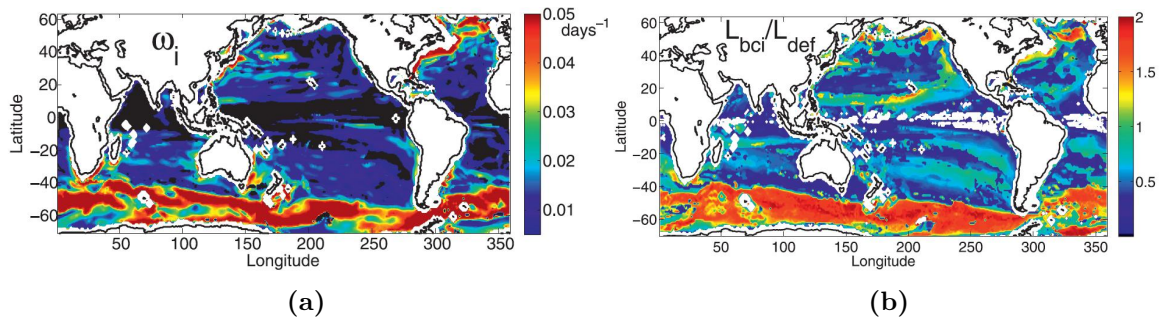


Fig. 2.7: Global distribution of (a) observed maximum growth rate and (b) length scale L , compared to L_d , of large scale ($L > L_d/5$) baroclinic instability, from Tulloch et al. (2011).

thus be $\mathcal{O}(100)$ times larger than that of the baroclinic background flow, providing an explanation for the dominance of oceanic measurement by transient eddies and the observed KE peak at the mesoscale (cf. chapter 1).

2.2 Dynamics of SST anomalies

The temperature of the sea surface (SST) is a key variable controlling the interaction between ocean and atmosphere. In the tropics, SST is well-known to be a key ingredient in energetic ocean-atmosphere interactions (El Niño), but also in mid-latitudes, SST provides the thermal boundary condition for the atmospheric circulation, which as shown above (section 2.1.3, Eady, 1949) can have a decisive impact on atmospheric heat transport. Particularly in regions of narrow fronts (≤ 100 km), midlatitude SSTs are indeed observed to shape the (turbulent or stable) character of the atmospheric boundary layer (Sweet et al., 1981; Small et al., 2008) and exert an impact on the atmospheric circulation up to the tropopause (Minobe et al., 2008; Czaja and Blunt, 2011).

After an overview of the oceanic surface layer for whose temperature SST is representative (section 2.2.1), section 2.2.2 derives the mechanisms by which ocean and atmosphere mutually set its evolution, and examines how ocean mesoscale eddies could make a contribution therein.

2.2.1 The oceanic surface mixed-layer

Exchanges of momentum, heat and fresh water across the sea surface generate turbulent boundary layers at the interface between ocean and atmosphere, in which intense

turbulent mixing homogenizes properties with depth. In the atmosphere, the marine atmospheric boundary layer (MABL) has a typical thickness of ~ 1 km, whereas in the ocean, the surface mixed-layer, throughout which temperature and salinity are close to depth-independent and oxygen content is high, reaches a typical depth of ~ 100 m. This depth varies on diurnal, seasonal and inter-annual time-scales, and is not only controlled by the surface wind and buoyancy forcing, but also by the oceanic flow, especially on meso- and submesoscales (e.g. Dewar, 1986; Boccaletti et al., 2007).

Figure 2.8 displays the typical characteristics of the mixed-layer depth distribution in NA and SO, as observed in a large-scale monthly climatology (developed by de Boyer Montégut et al. (2004) and described in Appendix B.1).

Maximum annual ranges (displayed in the rhs panels) can exceed 300 m and generally coincide with the regions of deepest annual-average mixed-layer depths (lhs panels). These are found in Labrador and Nordic Seas, as well as along and on the equatorward flanks of the ACC. In these regions, characterized by intense winter-time cooling, the strong seasonal cycle of autumn-to-spring mixed-layer deepening and summer-time restratification is driven by air-sea buoyancy fluxes. Relatively deep winter mixed-layers are also found in the mode-water formation region to the south of the GS. Apart from continental shelves, minimum annual ranges of mixed-layer depth are found in the subtropical regions, which are characterized by a year-round supply of wind-driven mixing by the trade winds. Interestingly, this results in deeper minimum climatological mixed-layers (middle panels) at low compared to high-latitudes in the NA. Deep minimum mixed-layer depths are also found along Florida current and GS extension in the NA, while in the SO, the minimum map is quite noisy and highlights the reduced SO data content in this climatology (cf. Appendix B.1).

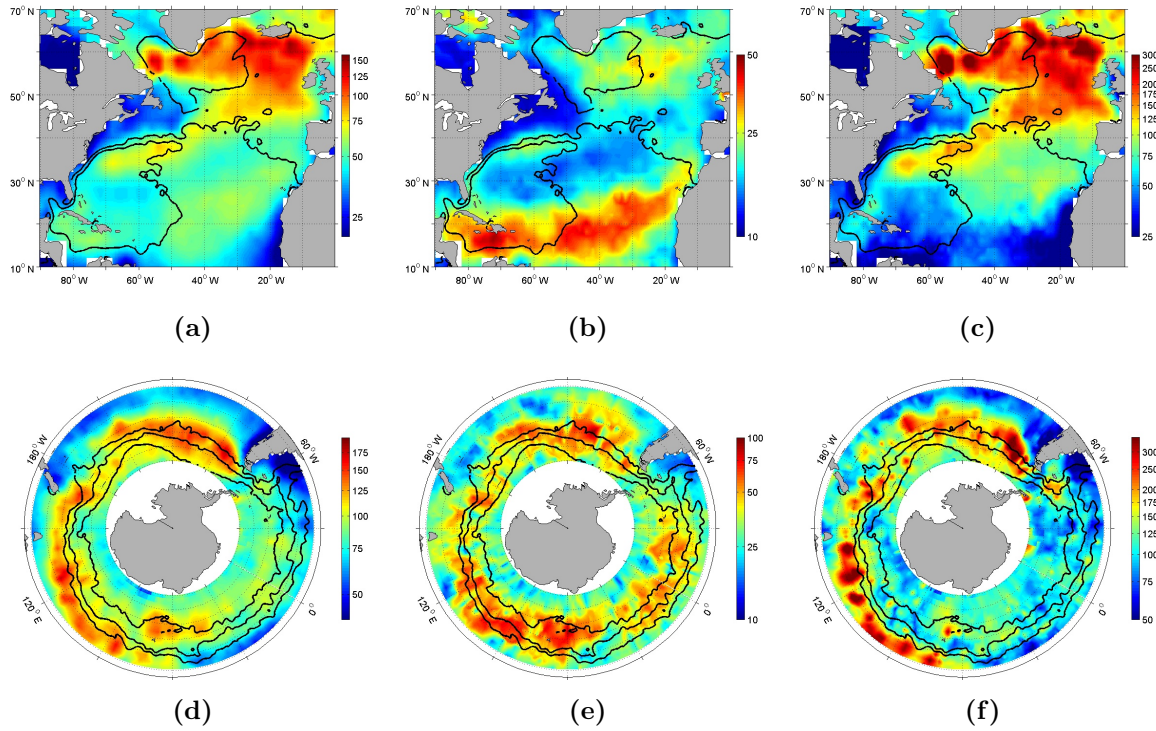


Fig. 2.8: The observed large-scale (de Boyer Montégut et al., 2004) mixed-layer depth climatology. Lhs panels display the annual mean, middle panels the minimum and rhs panels the annual range (maximum minus minimum) of the mixed-layer depth climatology, in m. Characteristic contours of the 5½-year (June 2002–December 2007) average sea surface height $\bar{\eta}$ are overlaid: -45, 5 and 45 cm in the NA (top panels), -55, 0 and 40 cm in the SO (bottom panels).

2.2.2 SST dynamics in a bulk mixed-layer view

In the surface layers of the ocean, where the flow is nearly incompressible ($\nabla \cdot \mathbf{u} + \partial_z w = 0$) and the specific heat capacity of sea water c_p (at constant pressure) can be taken as constant, the first law of thermodynamics states that the temperature T of a parcel (small unit of volume) can only rise if it gains heat:

$$\rho_0 c_p (\partial_t + \mathbf{u} \cdot \nabla + w \partial_z) T = -d_z q - \rho_0 c_p (\nabla \cdot (\mathbf{u}^t T^t) + \partial_z (w^t T^t)). \quad (2.19)$$

Apart from the net turbulent (and ultimately molecular) heat exchange between the parcel and its surroundings (2nd term on rhs), near the ocean surface heat gain also results from the convergence of the penetrating flux of solar radiation q (in $W m^{-2}$, and always downward, i.e. < 0 , as z is upward).

As the mixed-layer is constantly homogenized with depth, the evolution of SST, on time and space scales larger than that of mixed-layer turbulence, is largely given by that of the (depth-average) mixed-layer temperature, simply denoted T (and typically used interchangeably with SST) in the following. The evolution equation for T is thus obtained by vertically integrating (2.19) from the base of the mixed-layer ($z = -h_{ml}$) to the surface ($z = 0$), which, as shown by Kraus and Turner (1967) and summarized by Frankignoul (1985), yields:

$$h_{ml} d_t T = -\frac{Q - Q_{sub}}{\rho_0 c_p} - w_{entr} (T - T_{sub}) - \int_{-h_{ml}}^0 \nabla \cdot (\mathbf{u}^t T^t) dz. \quad (2.20)$$

Here $d_t \equiv \partial_t + \mathbf{u} \cdot \nabla$ denotes the rate of change following the (depth-average) mixed-layer flow on resolved scales (or large and mesoscales), which shows little vertical shear in the mixed-layer (and is simply denoted $\mathbf{u}$ in the following), whereas the last term on the right-hand side denotes the mixed-layer temperature flux convergence due to unresolved motions (or typically submesoscales, cf. Boccaletti et al. (2007)) and can be cast into a heat flux $Q_{mix} = -\rho_0 c_p \int_{-h}^0 \nabla \cdot (\mathbf{u}^t T^t) dz$. In (2.20), Q denotes the net upward ($Q > 0$) heat flux at the sea surface, including radiative (solar and infrared) and turbulent (latent and sensible) heat fluxes, and the (typically small) fraction of solar radiation that penetrates across the mixed-layer base is denoted Q_{sub} (always downward, i.e. < 0). We have moreover introduced an entrainment velocity:

$$w_{entr} \equiv \Lambda (\partial_t h_{ml} + \nabla \cdot (h_{ml} \mathbf{u})) = \Lambda (\partial_t h_{ml} + w(-h_{ml}) + \mathbf{u}_{sub} \cdot \nabla h_{ml}), \quad (2.21)$$

in which $\Lambda() = 1$, if its argument > 0 , and zero otherwise. This reflects that a net influx of water across the (moving) mixed-layer base has an impact on the evolution of T ,

but a net outflux has not⁶. The second equality in (2.21) is obtained by integrating the incompressibility equation across the mixed layer $\int_{-h_{ml}}^0 \nabla \cdot \mathbf{u} dz + [w]_{-h_{ml}}^0 = \nabla \cdot (h_{ml} \mathbf{u}) - \mathbf{u}_{sub} \cdot \nabla h_{ml} - w(-h_{ml}) = 0$ (where we have used Leibniz's rule), and clearly shows that entrainment can result from time-variations in the mixed-layer depth, vertical inflow across the mixed-layer base or horizontal flow from regions of shallow to deep mixed layers (termed lateral “induction”). From this, the (upward) entrainment heat flux $Q_{entr} = -\rho_0 c_p w_{entr} (T - T_{sub})$ is determined by the temperature “jump” $(T - T_{sub})$ across the mixed-layer base, which relies on the observation that the mixed-layer is typically well separated from the subsurface ocean by a transition zone in which properties change abruptly with depth. We can now rewrite (2.20) as:

$$h_{ml} d_t T = -\frac{Q_{net}}{\rho_0 c_p}, \quad (2.22)$$

in which $Q_{net} = Q - Q_{sub} - Q_{entr} - Q_{mix}$ is the net mixed-layer heat flux divergence.

The evolution of both T and h_{ml} is characterized by a large seasonal cycle (described in chapter 3.1) that often dominates their variability in midlatitudes. To facilitate the study of (non-seasonal) SST variability and the contribution of transient mesoscale eddies to it, it is therefore beneficial to consider the evolution of non-seasonal SST anomalies (SSTA or T') separately from that of the seasonal climatology $\bar{T}$, where the latter is obtained as long-term average of T evaluated separately at each given time of the year (e.g. each January, February, etc., if T is given as a monthly time series). After decomposition of each variable's variability into a seasonal mean (denoted by overbars) and a departure from it (denoted by primes, as generally in the remainder of this study), seasonal averaging of (2.22) then yields the evolution equation for $\bar{T}$ following the climatological mean path of water parcels ($d_{\bar{t}} \equiv \partial_t + \bar{\mathbf{u}} \cdot \nabla$):

$$\overline{h_{ml} d_t T} = \bar{h}_{ml} d_{\bar{t}} \bar{T} + \bar{\mathcal{E}} = -\frac{\bar{Q}_{net}}{\rho_0 c_p}. \quad (2.23)$$

Here $\bar{\mathcal{E}}$ gives the eddy rectification of the climatological mixed-layer heat budget, i.e. $\bar{\mathcal{E}} = \bar{h}_{ml} \overline{\mathbf{u}' \cdot \nabla T'} + \overline{h'_{ml} d_t T'} + \overline{h'_{ml} \mathbf{u}' \cdot \nabla \bar{T}}$.

Subtracting (2.23) from (2.22), noting that $d_t = d_{\bar{t}} + \mathbf{u}' \cdot \nabla$ and dividing by $\bar{h}_{ml}$,

⁶ Note that the evolution, following the flow, of the mixed-layer depth itself ($d_t h_{ml}$) is obtained from the mixed-layer turbulent kinetic energy budget (e.g. Kraus and Turner, 1967; Frankignoul, 1985).

yields the evolution of a parcel's SSTA:

$$\begin{aligned} d_t T' &= -\frac{Q'_{net}}{\rho_0 c_p \bar{h}_{ml}} - \mathbf{u}' \cdot \nabla \bar{T} - \frac{h'_{ml}}{\bar{h}_{ml}} d_t T + \frac{\bar{\mathcal{E}}}{\bar{h}_{ml}} \\ &= -\frac{Q'_{net}}{\rho_0 c_p \bar{h}_{ml}} - \mathbf{u}' \cdot \nabla \bar{T} - \frac{h'_{ml}}{\bar{h}_{ml}} d_t \bar{T} - \frac{(h'_{ml} \mathbf{u}')'}{\bar{h}_{ml}} \cdot \nabla \bar{T} - (h'_{ml} d_t T')'. \end{aligned} \quad (2.24)$$

This shows that in a homogeneous environment ($\bar{T} = \text{cste}$ in time and space), departures of the parcel's temperature from $\bar{T}$ are controlled by the anomalous mixed-layer heat convergence $-Q'_{net}$ (and by the 3rd-order term $-(h'_{ml} d_t T')'$). In the presence of spatial and seasonal variations in the background SST, SSTA can moreover be generated by deviations of the parcel from its climatological mean path ($\mathbf{u}' \neq 0$) and by fluctuations of the parcel's mixed-layer around its mean depth. If we do not follow a parcel, but instead measure SSTA from a fixed location or following the climatological mean flow, the relevant T' evolution equation is

$$d_t T' \equiv (\partial_t + \bar{\mathbf{u}} \cdot \nabla) T' = -\frac{Q'_{net}}{\rho_0 c_p \bar{h}_{ml}} - \mathbf{u}' \cdot \nabla \bar{T} - \frac{h'_{ml}}{\bar{h}_{ml}} d_t T - \frac{\mathcal{E}'}{\bar{h}_{ml}}. \quad (2.25)$$

Here it is noteworthy to mention that for large-scale, low Rossby number flows, the second term on the left-hand sides of (2.24) and (2.25) mainly includes geostrophic and Ekman contributions $\mathbf{u}' = \mathbf{u}'_g + \mathbf{u}'_{ek}$, the first given by anomalies in (2.4), the second from (2.10) by anomalies in

$$\mathbf{u}_{ek} = -\mathbf{k} \wedge \frac{\boldsymbol{\tau}}{\rho_0 f \bar{h}_{ml}}, \quad (2.26)$$

where we have assumed that Ekman layer and mixed-layer approximately coincide.

2.2.3 Mechanisms

Many of the terms that control the T' evolution, as given by the right-hand sides of (2.24) or (2.25), are themselves, either directly or indirectly, dependant on T' . They therefore do not only force departures of T from $\bar{T}$, but also feed back on pre-existing T' , acting to reinforce (positive feedback, $d_t T \propto T'$) or damp them (negative feedback, $d_t T \propto -T'$). As reviewed in chapter 5.1.1, the perhaps climatologically most important of these feedbacks results from the dependence of air-sea heat flux anomalies Q' on T' , and is shown from observations to be almost always negative in the extra-tropics.

Here we briefly consider the various forcing mechanisms of SSTA, as given in (2.25). The atmosphere principally forces SSTA on the large spatial scales of atmospheric synoptic disturbances (whose deformation radius is typically ~ 1000 km). The surface

temperature, moisture and wind speed anomalies associated with the latter induce perturbations in air-sea heat fluxes Q' and these are the dominant mechanism of the atmospheric forcing (Cayan, 1992a). Wind speed anomalies also force T' via anomalous Ekman currents, that advect background SST $-\mathbf{u}'_{ek} \cdot \nabla \bar{T}$. This Ekman forcing is typically observed to reinforce Q' (Frankignoul, 1985; Kushnir et al., 2002) and can be of comparable magnitude, particularly in the SO (Verdy et al., 2006). Although the energy of the atmospheric disturbances largely resides on synoptic time scales (of a few days), they are observed to generate persistent large-scale T' (~ 3 months). As shown by Frankignoul and Hasselmann (1977), this reflects the fact that T' are the integrated (red noise) response to the “random” high-frequency (white noise) atmospheric forcing (in a manner analogous to Brownian motion generated by the molecular random walk). On longer time-scales (from months to decades), the principal modes of atmospheric variability, the North Atlantic Oscillation (NAO) in the NA and the Southern Annular Mode (SAM) in the SO, also play an observed role in forcing persistent basin-scale patterns of SSTA in NA (e.g. Cayan, 1992a; Czaja and Frankignoul, 2002) and SO (Verdy et al., 2006; Ciaso and Thompson, 2008), where remote forcing by the principal mode of tropical Pacific coupled ocean-atmosphere variability, the El Niño-Southern Oscillation (ENSO) is also important. As atmospherically forced Ekman pumping anomalies are typically of minor importance (Frankignoul, 1985), entrainment mostly acts as a damping on large-scale atmospherically forced SSTA. Occasionally it however re-entrains the previous year’s SSTA that have been shielded from air-sea interactions in the seasonal thermocline during summer and thereby enhances SSTA persistence (e.g. Deser et al., 2003).

Large-scale (geostrophic) ocean currents do not only play an important role in modulating atmospherically forced SST variability through advection ($-\overline{\mathbf{u}_g} \cdot \nabla T'$), as particularly evident along the ACC (e.g. Verdy et al., 2006), but also force SSTA of relatively large spatial extents through their own (intrinsic or atmospherically forced) low-frequency variability ($\mathbf{u}'_g$), expressed as large-scale meandering or shifts in current paths (e.g. Frankignoul et al., 2001; Sokolov and Rintoul, 2003; Dong et al., 2006b; Sokolov and Rintoul, 2009).

Mesoscale eddies and SST

As they represent a large fraction of the ocean’s kinetic energy (chapters 1 & 2.1), deep transient mesoscale eddies are also likely to contribute to the observed variability in SSTA, and the observational quantification of this contribution is the subject of this

thesis. The eddy contribution is expected to typically occur on smaller scales (~ 100 km) than the atmospheric forcing, and from (2.24), the relevant mechanisms are likely to include the following:

- Deep geostrophic eddy motions, if baroclinic, are in a near thermal wind balance with subsurface temperature anomalies T'_{sub} , and in the presence of a background entrainment velocity $\bar{w}_{entr}$ these will be mapped to the surface mixed-layer (e.g. Klein and Hua, 1990; Halliwell et al., 1991a). Via this mechanism, we may thus expect to primarily see eddy T' from autumn to spring.
- Eddy contributions to Q'_{entr} are also expected via anomalies in the entrainment velocity w'_{entr} that are induced by eddy geostrophic flow across the sloping mixed-layer base $-\mathbf{u}'_g \cdot \nabla \bar{h}_{ml}$, and to a lesser extent via eddy-induced $w'(-\bar{h}_{ml})$.
- The above likely result in eddy mixed-layer depth anomalies h'_{ml} , which will in turn drive T' via the 3rd term on the rhs of (2.24).
- As eddying flows are expected to extend up to the surface, in the regions of significant horizontal mixed-layer temperature gradients, they will also force T'' via $-\mathbf{u}'_g \cdot \nabla \bar{T}$.
- Once eddy T' have been generated, they are expected to tend via T'_{sub} , reducing the entrainment forcing, and, moreover, to be quickly damped via their feedback on air-sea heat fluxes $Q'(T')$ (chapter 5.1.1). Eddy-induced mixed-layer temperature gradients are moreover expected to be eroded by submesoscale mixed-layer instabilities, leading to an additional negative feedback $Q'_{mix}(T')$.

Whether the mesoscale forcing contributes importantly to observed SST variability, and through which mechanisms, are key questions addressed in this thesis. As the set of observations used here (chapter 3), and in general presently available data, do not (yet) permit basin-wide observational estimates of all of the above-mentioned contributions, our approach will be to estimate those accessible and assess how well they may or not explain the observed characteristics of SST variability.

2.3 Summary

This chapter has provided an introduction to the dynamics of mesoscale eddies and SSTA. In the following chapter we will review methods to observe both and describe two datasets that will be used in the remainder of this study to observationally establish the signature of the former in the latter.

Chapter 3

OBSERVATIONS AND PRELIMINARY DIAGNOSTICS

In this chapter we introduce the two satellite datasets that form the core of this thesis – through-cloud microwave observations of SST (section 3.1) and multi-altimeter observations of SSH (section 3.2), assess their capacity to resolve transient mesoscale processes and describe an algorithm that automatically identifies and tracks these propagating eddies in the SSH field throughout their lifetime (section 3.2.4).

As preliminary for the analysis of the relationship between SST and SSH to be presented in the following chapters, section 3.3.1 introduces the SSH variability observed in NA and SO basins and examines its association with eddy circulations. Two contrasting oceanic regimes, which will often be the focus of the following analyses, are identified in section 3.3.2, with a first overview of their characteristics provided in section 3.3.3.

(The specifications of complementary, climatological, datasets referred to at several places within this thesis are provided in Appendix B).

3.1 Observing sea surface temperature variability

The analysis of observed mesoscale eddy SST signatures presented in this study is based on a dataset created by Remote Sensing Systems (REMSS) through optimal-interpolation (OI) of measurements by the Advanced Microwave Scanning Radiometer operating on the National Aeronautics and Space Administration (NASA) Earth Observing System satellite Aqua (AMSR-E). Since its launch in May 2002, AMSR-E scans the Earth’s microwave (MW) radiation at six frequencies ranging from 6.9 to 89 GHz, from which SST and other important climatological variables such as wind speed, atmospheric water vapour content, cloud liquid water and rain rate are retrieved (Wentz and Meissner, 2000).

After providing some background information on SST observing systems (section 3.1.1), here we assess what AMSR-E OI SST data represent (section 3.1.2) and give some detail on the further processing applied to them in this study (sections 3.1.3 and 3.1.4).

3.1.1 How we observe SST

Earliest observations of oceanic sea surface temperature involved dropping a bucket into the water and measuring its temperature once back on board. Today SST observations from research ships are more sophisticated and the development of simple measurement devices, such as ship engine room intake or hull sensors and, with a much better accuracy, expendable bathythermographs (XBT), allow routine on-the-move measurements of ocean temperatures by merchant ships that are coordinated in the Voluntary Observing Ship program by the World Meteorological Organization. Ship measurements, characterized by their highly inhomogeneous coverage (of not more than 10% of the world's oceans)¹, have been the principal source of SST observations until the late seventies and eighties when the first satellites available for climate research opened the way to essentially two new methods of observing SST.

Firstly, they enabled the deployment of moored and drifting buoys that transmit their data (and position) via satellites. Whereas moored buoys are concentrated in the tropics and close to coasts, drifting buoys, coordinated under the Global Drifter Program (www.aoml.noaa.gov/phod/dac/gdp.html), are much more evenly distributed. Today they acquire the bulk of in-situ measurements of SST, covering every month a slightly larger area than declining ship observations. With 4% (see footnote¹), this however remains only a small fraction of the global ocean (Xu and Ignatov, 2010).

The second innovation of the satellite-era is based on a different method for observing temperature that exploits the known temperature-dependence of the long-wave radiation emitted by water-containing bodies and had already been used for SST measurements from air planes. The development of a satellite-based radiometer, the Advanced Very High Resolution Radiometer (AVHRR) that scans infra-red (IR) emissions from the Earth's surface with a footprint of only 1 km and has been operating on different the National oceanographic and atmospheric administration (NOAA) satellites from 1978 until today (with added channels), made it possible for the first time to construct global synoptic maps of SST at a very high resolution (~ 4 km).

¹ This has been quantified by Xu and Ignatov (2010) by defining coverage as the monthly median of the daily percentage of 1° -by- 1° boxes covered by observations.

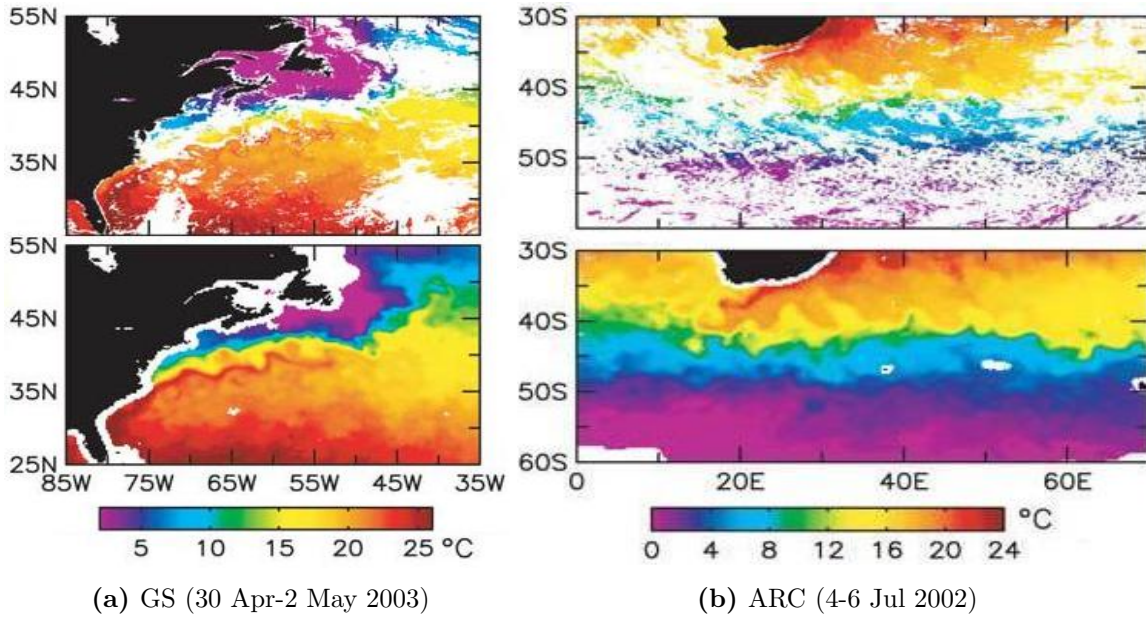


Fig. 3.1: IR (top, AVHRR, 8-km grid) versus MW (bottom, AMSR-E, 1/4° grid) observations of SST (3-day average maps). Missing values are indicated in white. Figure adapted from Chelton and Wentz (2005).

The major drawback of IR SST measurements is their limitation to clear-sky conditions as otherwise the sensed radiation is primarily that emitted from the top of clouds. This leads to severe reductions in data coverage, especially in regions of persistent cloud-cover, and basically precludes studies focussed on high-latitude phenomena such as deep-convection or the Southern Ocean circulation (Emery et al., 2006; Chelton and Wentz, 2005). In order to create continuous gap-free SST fields for climate research, optimum-interpolation (OI) techniques were devised that, however, lead to a highly degraded spatio-temporal resolution OI SST compared to original observations (Reynolds and Smith, 1994; Reynolds et al., 2002).

To circumvent this limitation a new technique was devised exploiting the weak SST dependence of microwave radiation emitted from the sea surface traversing clouds and aerosols nearly unattenuated. After the launch, in late 1997, of the Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (TMI) whose measurements were limited to low latitudes and high SST, since June 2002 AMSR-E measurements provide near-daily (89%) global observations of SST through clouds making detailed studies of extratropical WBC extension and the Southern Ocean SST possible for the first time. The improved coverage of MW compared to IR observations is illustrated in Figure 3.1 that shows 3-day average snapshots of SST for GS and ARC regions.

Whereas persistent cloud-cover prevents IR observations throughout the entire three day period, MW observations are only unavailable within ~ 75 km of land and over sea-ice. Not present in these examples, rain, high wind speeds (> 20 m/s) and sun glint also preclude MW measurements of SST and need to be identified and removed.

Although AMSR-E measurements have a far larger footprint and therefore effective resolution (~ 50 km) than raw AVHRR data, their through-cloud observing capacity greatly reduces the spatio-temporal smoothing required to construct gap-free SST datasets (Chelton and Wentz, 2005), which makes them the best choice for an observational study of mesoscale eddy SST imprints.

Currently, new insights gained from the study of MW SST are transformed into efforts to advance data processing techniques compared to traditional conservative OI methods in order to make gap-free IR and merged MW plus IR SST datasets available at a higher resolution without the drawback of decreased accuracy that is still inherent in available merged datasets (e.g. REMSS; Reynolds and Chelton, 2010; Donlon et al., 2012). Further work should compare analyses based on these higher-resolution SST datasets, once they become available, to the results presented in this study that are based on the MW-only OI SST dataset, which is described in some detail in the following section.

3.1.2 The AMSR-E OI SST dataset

The algorithm used to retrieve SST from AMSR-E measurements of MW (at 6.9 GHz) brightness temperature, that are weakly dependant on SST but also on surface wind-speed, is described in detail by Wentz and Meissner (2000). From these, REMSS provides a daily optimally-interpolated MW SST dataset on a $\frac{1}{4}^\circ$ grid. This study uses version 2 of this MW OI SST dataset². Here we quickly describe those processing steps applied by REMSS that need to be kept in mind when assessing what the final SST data represent³.

The OI process (Reynolds et al., 2004) transforms irregularly-spaced AMSR-E observations into a continuous SST field without missing data due to orbital gaps and the presence of sun glitter, rain and high wind speeds (> 20 m/s). This is achieved by adding a weighted sum of all available raw data as increments to the previous

² AMSR-E data, available at www.remss.com, are produced by REMSS and sponsored by the NASA Earth Science MEaSUREs DISCOVER Project and the AMSR-E Science Team.

³ refer to www.remss.com/sst/microwave-oi-sst-data-description.html for a full, referenced documentation

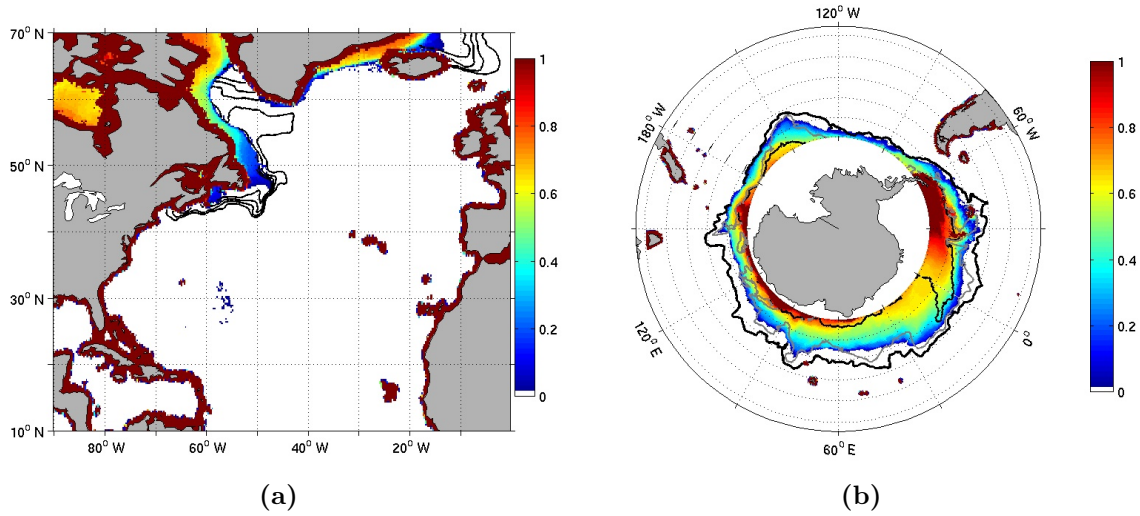


Fig. 3.2: Fraction of days during the study period with missing values of OI AMSR-E SST (colour, *). The contours show for the NA (a) the southernmost extent of the 1°, 2°, 3° and 4° C surface isotherms at the end of winter (March) and for the SO (b) the extreme (black, end of summer = February, end of winter = September) and annual mean (grey) positions of the 1° C surface isotherm. * areas south of 65°S and on continents are not included in the dataset.

day's analysis. In case of instrument outages, that occurred on a total of 28 days during the 5½ year period analysed here, the OI SST field is therefore frozen to the last day with available observations. As illustrated in Figure 3.2, the resulting OI coverage is continuous away from land (> 75 km) and sea-ice. The latter is at the origin of temporal variations in the OI SST coverage which are approximately limited to regions poleward of the end-of-winter outcrop position of the 1°C isotherm. Remaining problems in the OI dataset, of which some have been cured in its newly available version 3, include data contamination due to undetected sea ice, usually during its advancing or retreating phases, and due to undetected precipitation, often occurring near the edges of rain cells. Its adverse effects, especially on resulting SST gradients, have been clearly shown by Reynolds et al. (2007).

Whereas in-situ measurements of SST are taken at a depth of one to several meters from ships and at roughly half a meter by drifting and moored buoys (Reynolds et al., 2002), radiative measurements represent the temperature of a thin surface layer. IR radiation is emitted from the sub-millimetre depth oceanic skin layer, whose temperature is usually ~ 0.2 K cooler than the sub-skin just below (at a depth of ~ 1 mm), from where the longer microwaves penetrate (Donlon et al., 2002). MW radiometry is therefore unaffected by the cool-skin effect, it however measures SST in the layer

of maximum diurnal warming. This can reach several K especially in conditions of light wind, but, as it decays strongly with depth, it has little impact on SST measured in situ (i.e. deeper in the water column). From a comparison to TMI SST and its error statistics with respect to buoy observations Chelton and Wentz (2005) estimate an accuracy of raw AMSR-E SST of 0.3 to 0.4 K. An accuracy assessment for higher latitudes than those covered by TMI is provided by Dong et al. (2006a) for two repeat XBT transects through Drake passage and south of Australia. It uncovers warm biases (~ 0.2 K) of AMSR-E subskin SST with respect to bulk SST in spring and summer, cold biases in fall and winter, that closely average out in the annual mean. To facilitate calibration and validation of MW against in-situ observations, a modelled diurnal warming cycle (Gentemann et al., 2003) is removed from the raw AMSR-E subskin SST that represent mid-day and mid-night conditions as observed during the ascending and descending passes of Aqua’s sun-synchronous orbit. The final OI dataset therefore represents daily-minimum SST as it would be measured at a depth of roughly 1 m. This likely enhances MW OI SST accuracy assessed with respect to bulk in-situ SST, however residual subskin-bulk differences may remain at high latitudes as the diurnal warming model has been calibrated with low-latitude TMI observations.

AMSR-E measurement error statistics are estimated from retrieval errors, depending mainly on SST and wind speed, and from a collocation with near-real time in-situ SST provided by the Global Ocean Data Assimilation Experiment (GODAE). The weights employed by the OI scheme are determined based on these error statistics and on assumed correlation scales of 4 days and 100 km. Applied to raw AMSR-E data that contain information on time scales limited by the sampling rate of 1 day over 89% of the globe and on space scales limited by the ~ 56 km footprint, these correlation scales represent the lower bound for the time- and space scales resolved in the MW OI SST dataset. Whereas the temporal resolution is more than sufficient to resolve typical mesoscale processes, the spatial resolution is mid-way through the mesoscale spectrum and represents a clear limitation at high latitudes. The adequacy of this dataset to resolve mesoscale eddies will be discussed in more detail together with the presentation of results.

3.1.3 The seasonal climatology and SST anomalies (SSTA)

The study of the mesoscale contribution to SST variability and transient eddy SST signatures is facilitated by separating SST anomalies (SSTA) from the long-term mean SST and its seasonal cycle that otherwise dominates observed SST variability, as in-

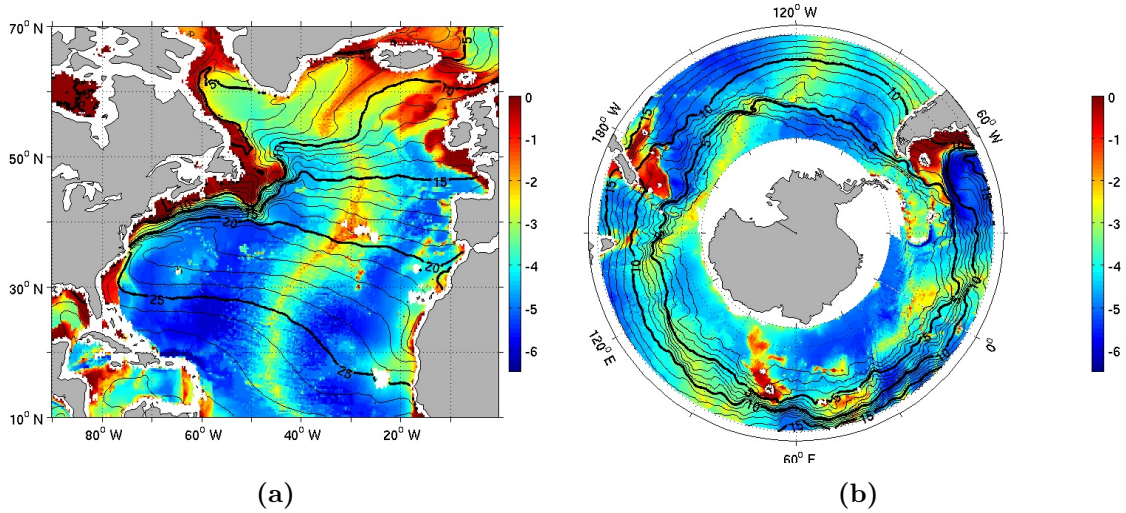


Fig. 3.3: Climatological mean SST from the June 2002 to December 2007 AMSR-E OI dataset (contoured in thin every 1°C, thick every 5°C). SST contours are overlaid onto the ocean basin’s bathymetry (in km) from the GEBCO dataset, introduced in section B.7.

troduced in chapter 2.2.2.

SSTA are obtained by removing the observed seasonal MW OI SST climatology, that is estimated by averaging all observations available at a given location and day of year over the June 2002 to December 2007 period studied. The long-term mean and the seasonal cycle of the removed SST climatology are illustrated in Figures 3.3 and 3.4, respectively. Interesting characteristics of the mean-state include a narrow front associated with the GS that hugs the edge of the coastal shelf and many small-scale structures coinciding with the complex bathymetric features along the ACC, as observed by many previous studies (e.g. Hughes and Ash, 2001). The amplitude of the seasonal cycle, as measured by the rms of monthly climatological SST deviations from the mean state, is highest on continental shelves, especially north of the GS where the end-of-winter outcrop of 10°C isotherm meets the end-of-summer outcrop of the 25°C isotherm. A relatively strong seasonal SST cycle with a monthly rms of nearly 3 K is found in the NA subtropical mode water region that is characterized by the winter-time spreading of 18°C waters. At low and high latitudes the seasonal SST cycle is much weaker, especially along the course of the ACC, that sees hardly any seasonal isotherm displacements in its Indian sector.

As shown in Figure 3.5, the broad features of the seasonal MW SST climatology, although it is based only on a 5½ year period, agree well with a 25-year cli-

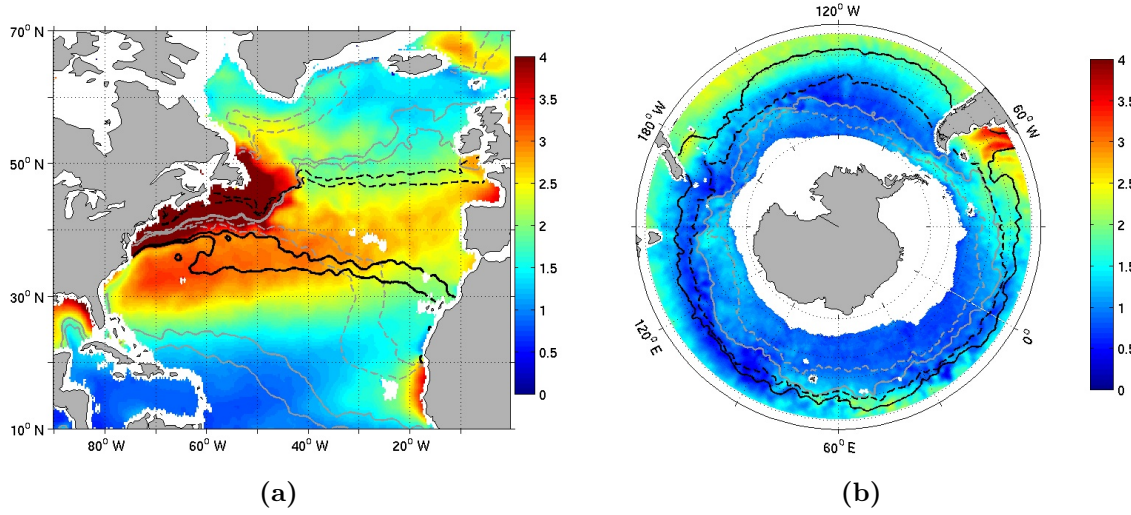


Fig. 3.4: Amplitude of the seasonal cycle in $\overline{\text{SST}}$. Colours show the inter-month rms of $\overline{\text{SST}}$ (in K), contours show the extreme positions of surface isotherms at the end of winter (lines, NA: March, SO: September) and summer (dashes, NA: August, SO: February) for a) the NA ($10^\circ \pm 0.5^\circ\text{C}$ and $25^\circ \pm 0.5^\circ\text{C}$ in grey, $18^\circ \pm 0.5^\circ\text{C}$ in black) and b) the SO (4°C in grey, 9°C in black).

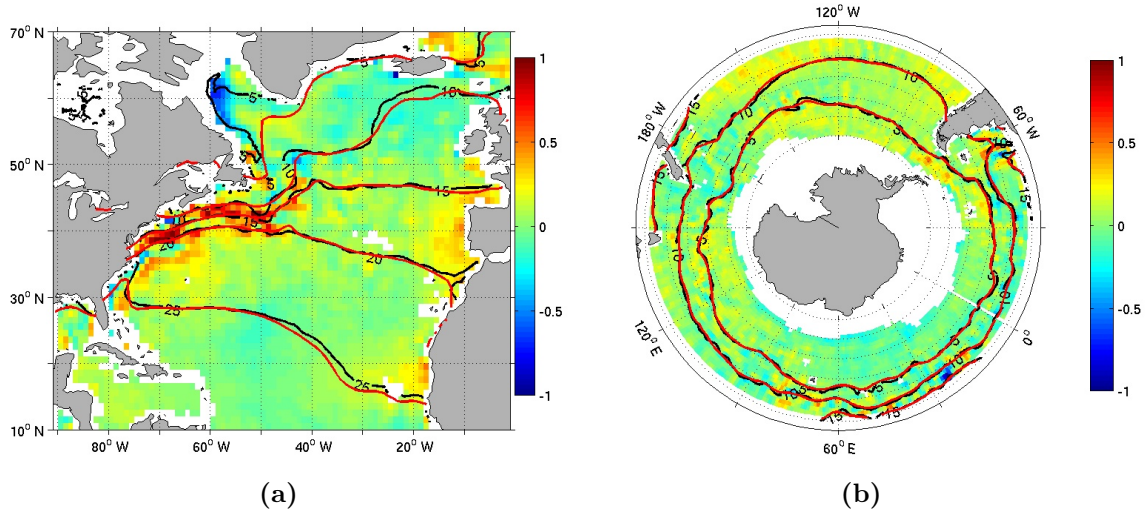


Fig. 3.5: Comparison of the 5½ year MW OI SST climatology to a 25-year (1982-2008) low-resolution (1°) IR SST climatology based on optimal interpolation of AVHRR data (Reynolds et al., 2002). The long-term mean MW and IR SST (in $^\circ\text{C}$) are contoured in black and red, respectively. The difference in the amplitude of the seasonal cycles is displayed in colour (as the difference, in K, of the inter-month $\overline{\text{SST}}$ rms of MW data, linearly interpolated on the low-resolution grid, minus that of IR data).

matology constructed from optimum-interpolation of AVHRR and in-situ SST, the Reynolds et al. (2002) OI.v2 IR SST, described in section 3.1.4. Some notable differences include slightly warmer MW SST in the NA, especially at high latitudes, an enhanced (reduced) MW seasonal cycle in regions of strong (weak) seasonal cycles in both datasets, and more eddying, fine-scale features in the higher resolution MW dataset.

To what extent these quasi-steady mesoscale features in the MW SST climatology are an observed physical feature (that would remain even in infinite time series) or an artefact of the short averaging time, remains to be tested as longer-time high-resolution SST datasets become available. Whereas the former cannot a priori be excluded, especially as an analysis of the global frequency-wavenumber spectrum of eddy sea surface height variability presented recently by Wunsch (2010) already provides some arguments in its favour, the latter is certainly true to some extent so that the amplitude of the strongest transient SSTA is likely underestimated in this study. The 5½ year averaging period moreover induces a potential bias of SSTA statistics towards NH fall with respect to NH spring conditions. Appendix A provides further discussion of the biases induced by the short time-series length in both climatology and anomalies, highlighting the need for further research to fully assess the sensitivity of results to the method of defining SSTA by examining how well an unbiased full spectrum of non-seasonal SST variability is retained.

Snapshots and a detailed discussion of the resulting SSTA dataset are presented in chapter 4.

3.1.4 A proxy for large-scale SSTA

In order to strengthen conclusions on the observed imprint of mesoscale eddies in SSTA, it is desirable to test observational results against the null-hypothesis of SST variability being purely forced by large-scale processes. This requires a SSTA dataset that contains no information on mesoscale eddies, but fully retains the observed large-scale SST variability.

Based on the discussion in the previous chapter of the scale separation between typical synoptic disturbances in the ocean (~ 100 km) and the atmosphere (~ 1000 km), the latter being a principal contributor to the large-scale forcing of SST variability, in the following we define the “large-scale” as variability occurring on spatial scales larger than 5° . The desired large-scale SSTA dataset is then obtained by binning and averaging available $\frac{1}{4}^\circ$ MW-SSTA, constructed as introduced above, onto a predefined

5°-by-5° grid. At places different coarse grids, as a 10°-by-10° version, will also be used for the bin-averaging of SSTA, but, unless explicitly mentioned, the notion of “large-scale SSTA” will generally refer to this 5°-by-5° dataset.

Bearing in mind its different statistics, occasionally the Reynolds et al. (2002) OI.v2 IR SST dataset will also be consulted as a large-scale SST dataset, since the large correlation scales of ~ 900 km zonally and ~ 600 km meridionally, used for its OI analysis (Chelton and Wentz, 2005), effectively prevent the resolution of mesoscale processes. The dataset, provided weekly on a 1° grid, is obtained from *ftp.emc.ncep.noaa.gov/cmb/sst/oisst.v2/* for the 25-year period 1981 to 2008. Large-scale IR SSTA are constructed in the same way as described above for MW SSTA by subtraction of the long-term seasonal climatology. Figure 3.5 compares it to the AMSR-E SST climatology.

3.2 Observing transient eddies in sea surface height

This section introduces (SSH) observations and their relation to eddying currents on the oceanic mesoscale. We revisit what SSH represents (section 3.2.1), how we observe it (section 3.2.2) and which processes, time and space scales are captured by the observational dataset used in this study (section 3.2.3). A dataset of propagating SSH anomalies representing the tracks of nonlinear eddies is then introduced in section 3.2.4.

3.2.1 Why we observe SSH

SSH (η) measures deviations of the ocean surface from the geoid (at $z=0$). As these correspond to changes in ocean surface pressure (2.2), SSH observations allow us to compute the surface geostrophic flow (2.4), and this makes them very appealing for studies of the ocean surface circulation on large and meso-scales.

Observations of SSH moreover take a special place in the spectrum of radiation-based measurements from space. Whereas the latter are usually restricted to observing properties at the sea surface as the ocean is very nearly opaque to electromagnetic radiation, SSH is intimately linked to properties of the underlying flow. This can be seen by examining the expression for the mass (per unit area) of an oceanic fluid column $m_{oc} = \int_{-H}^{\eta} \rho dz$. Splitting the integral into an interior (from the sea floor at $z=-H$ to $z=0$) and a surface (from $z=0$ to η) component, where density can be regarded

as a constant ρ_0 , yields an expression for SSH

$$\eta = m_{oc}/\rho_0 + \int_{-H}^0 -\rho/\rho_0 dz \quad (3.1)$$

that reveals its direct relationship to changes in the mass and the relative volume of the entire oceanic fluid column. As discussed in chapter 2.1, spatial gradients in these two components respectively drive depth-independent barotropic and vertically-sheared baroclinic flows.

Although this illustrates the principal connection of η variability to deep oceanic flows, it does not constrain how observed kinetic energy is actually partitioned into the various vertical modes, which is currently still a matter of debate. Theory and idealized modelling (Smith and Vallis, 2001) suggest that in the presence of a sharp ocean-like thermocline η variability principally reflects deflections of this thermocline (i.e. the first baroclinic mode), a finding that is consistent with observational analyses of η variability (Stammer, 1997a) and scattered current-meter data (Wunsch, 1997) in midlatitudes. A recent OGCM analysis (Lapeyre, 2009) suggests that a surface-trapped mode might be equally important and observations (Isern-Fontanet et al., 2006) support this for flows with horizontal scales on the small end and smaller than the oceanic mesoscale.

3.2.2 How we observe SSH

As illustrated in Figure 3.6, the principle of altimeter observations of SSH from space relies on the determination, to centimetric accuracy, of the difference between three distances, that are all 10^4 to 10^8 times larger than this required accuracy:

$$\eta = (O - R) - h_g = h - h_g. \quad (3.2)$$

The height h of the sea surface above an arbitrary reference ellipsoid is given by the difference between the height of the spacecraft orbit (O) above this ellipsoid, determined precisely from orbit models constrained to satellite tracking observations (especially from the ground-based Doppler orbitography system DORIS developed by CNES, cf. Chelton et al. (2001, section 4) for details), and the range (R) between the spacecraft and the sea surface, measured by active microwave (radar) altimetry. Natural microwave emission from the sea surface, used amongst others for the passive microwave SST measurements described above, is much less energetic and therefore

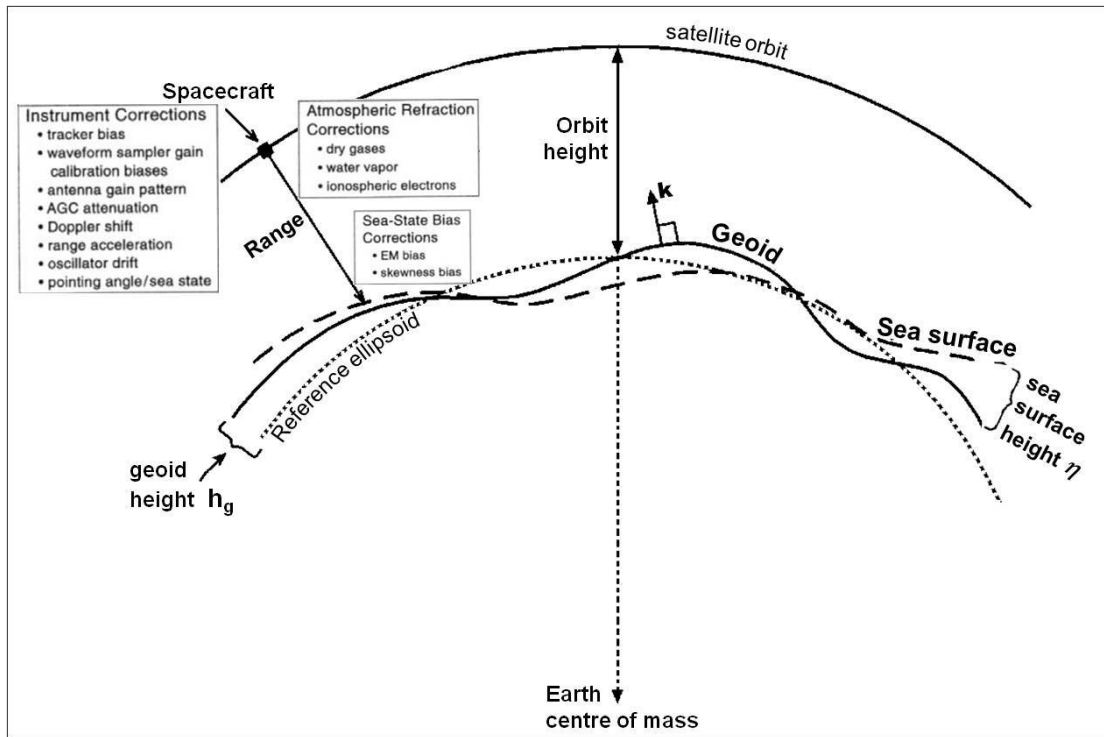


Fig. 3.6: A schematic of the SSH observing system (not to scale). Adapted from Wunsch and Stammer (1998) and Chelton et al. (2001).

easily distinguished from the reflected signal. In order to achieve an accurate measurement of R , the round-trip travel time of the radar-pulse that is emitted by the altimeter on board the orbiting spacecraft and reflected by the ocean surface, needs to be corrected for delays due to the interaction of the pulse with the ionosphere, the dry and the wet troposphere and for biases arising from the way the pulse is reflected from the rough and wavy sea surface. The development of these and a range of altimeter instrument-specific corrections, listed in Figure 3.6 and explained in detail e.g. in the review by Chelton et al. (2001), led to a reduction in measurement errors by more than an order of magnitude during the early stages of satellite altimetry. The required centimetric accuracy was first achieved with the Topex/Poseidon (T/P) mission, launched in 1992.

The third, and up to now the most accuracy limiting component of η , is the estimation of the Earth's geoid, to which any dynamically significant SSH observations need to be referenced. The geoid, that reflects the distribution of masses in the Earth's mantle and crust, is only to the lowest approximation a sphere flattened at the poles. In reality, it shows a complex topography h_g with respect to this reference ellipsoid that features amplitudes of up to 100 m and highly energetic slopes on all spatial

scales. This topography is revealed in unprecedented accuracy (1–2 cm) and detail (spatial resolution of 100 km) by the current GOCE (Gravity field and steady-state Ocean Circulation Explorer) mission. However, its data have not yet been incorporated into available SSH datasets. These circumvent the problem of an accurate geoid estimation by subtracting the time-average from altimeter observations to obtain dynamically significant SSH $\eta = h - \bar{h}$, and are therefore limited to studies of SSH and ocean time variability. Studies of the time-mean circulation rely on mean dynamic topographies estimated from in-situ observations. These were traditionally derived from hydrographic observations via the thermal wind relation, and therefore limited to information on the current shear with respect to motions at a chosen reference depth. More recently, in-situ estimates of the absolute time-mean dynamic topography have become available using observations by deep floats (Gille, 2003a) and surface drifters (Niiler et al., 2003).

Apart from reflecting geostrophic surface currents, SSH variability also reflects “external” forcing by ocean tides and atmospheric pressure loads. Satellite altimetry greatly enhanced the understanding and accurate modelling of oceanic tides, so that today tidal SSH variability η_T , which reaches amplitudes of ~ 1 m in open oceans, is predicted globally with centimetric accuracy. Atmospheric pressure fluctuations δp_a (defined as deviation from the global mean) induce SSH fluctuations with typical amplitudes of ~ 10 cm in the extratropics (~ 2 cm in the tropics). As reviewed by Wunsch and Stammer (1997), these are (at periods longer than a few days and outside the tropics) to a good approximation isostatic $\eta_a = -\delta p_a / \rho_0 g$, acting to cancel out atmospherically forced pressure gradients just below the sea surface. It is only the residual SSH variability $\eta_d = \eta - (\eta_T + \eta_a)$ that directly represents non-tidal ocean surface pressure variability and therefore, following (2.4), the streamlines for the surface geostrophic flow. As this dynamic SSH η_d is the desired variable for ocean circulation studies, observed SSH is corrected for the “externally” forced variability using tidal models and atmospheric assimilations. (In the following, η_d will often be simply referred to as SSH or η). High-frequency variability that is not resolved by the altimeter measurements (for the near 10-day repeat orbit of T/P this corresponds to periods below 20 days) aliases to signals at lower frequencies. Apart from ocean tides, it mainly reflects fast wind-forced high-latitude barotropic motions. For an accurate study of the resolved lower-frequency variabilities, these need to be corrected for as well.

The relatively high temporal sampling by T/P in its near 10-day repeat orbit comes

at the cost of a reduced spatial sampling by its large ground-track spacing reaching 315 km at the equator. This severely limits its effective wavelength resolution capacity to only 6° in both zonal and meridional directions in every two repeat cycles (Chelton and Schlax, 2003). T/P observations alone are therefore clearly too coarse for studies of the oceanic mesoscale.

This limitation is addressed by merging observations from T/P with that of other, simultaneously operating, altimeter missions. For the first time, this became possible with the successful recalibration (Le Traon and Ogor, 1998) of altimeter observations from the European Research satellites (ERS 1 and 2), which were originally characterized by very high orbit errors. Collected since 1991 on a lower frequency, 35-day, repeat orbit, they feature a much smaller ground-track spacing of only 90 km at the equator. Their merging with T/P observations, using the mapping method developed by Le Traon et al. (1998), greatly reduces mapping errors (Ducet et al., 2000). Compared to the gridded dataset based on T/P observations alone, this doubles the effective wavelength resolution to about 3° every 20 days (Chelton and Schlax, 2003). Figure 3.7 displays a typical snapshot of global SSH variability as seen by the single-altimeter (T/P, top panel) and by the merged two-altimeter dataset (T/P + ERS, bottom panel). It strikingly illustrates the enhanced capacity of the merged dataset to resolve narrow current meanders and isolated eddies. As shown by Chelton et al. (2011b), the two-altimeter configuration adequately resolves eddying features with radius scales larger than about 0.4° . With Envisat replacing ERS satellites on their 35-day repeat orbit in 2002 and Jason-1 stepping into the 10-day T/P orbit in 2001 (followed by Jason-2 in 2009), this configuration has been providing continuous mesoscale resolving SSH coverage from 1992 until today.

From mid-2002 onwards, there have been most of the time three, sometimes even four, altimeters in orbit simultaneously. The T/P orbit has been nearly continuously flown by two satellites in a tandem-configuration and GEOSAT Follow On (GFO), which belongs to a third family of altimeters that operates in an intermediate 17-day repeat orbit at a 160 km equatorial ground-track spacing, has been providing altimeter observations for the 10-year period 1998-2008. As shown by Pascual et al. (2006), merging observations from four instead of two altimeters leads to further improvements in the representation of eddies and fronts and to a significant increase in the resolved eddy kinetic energy (of up to $400 \text{ cm}^2/\text{s}^2$ in the Gulf Stream region). Although the degree to which current multi-mission SSH datasets sample the oceanic mesoscale circulation remains somewhat open, they clearly represent a significant improvement over early

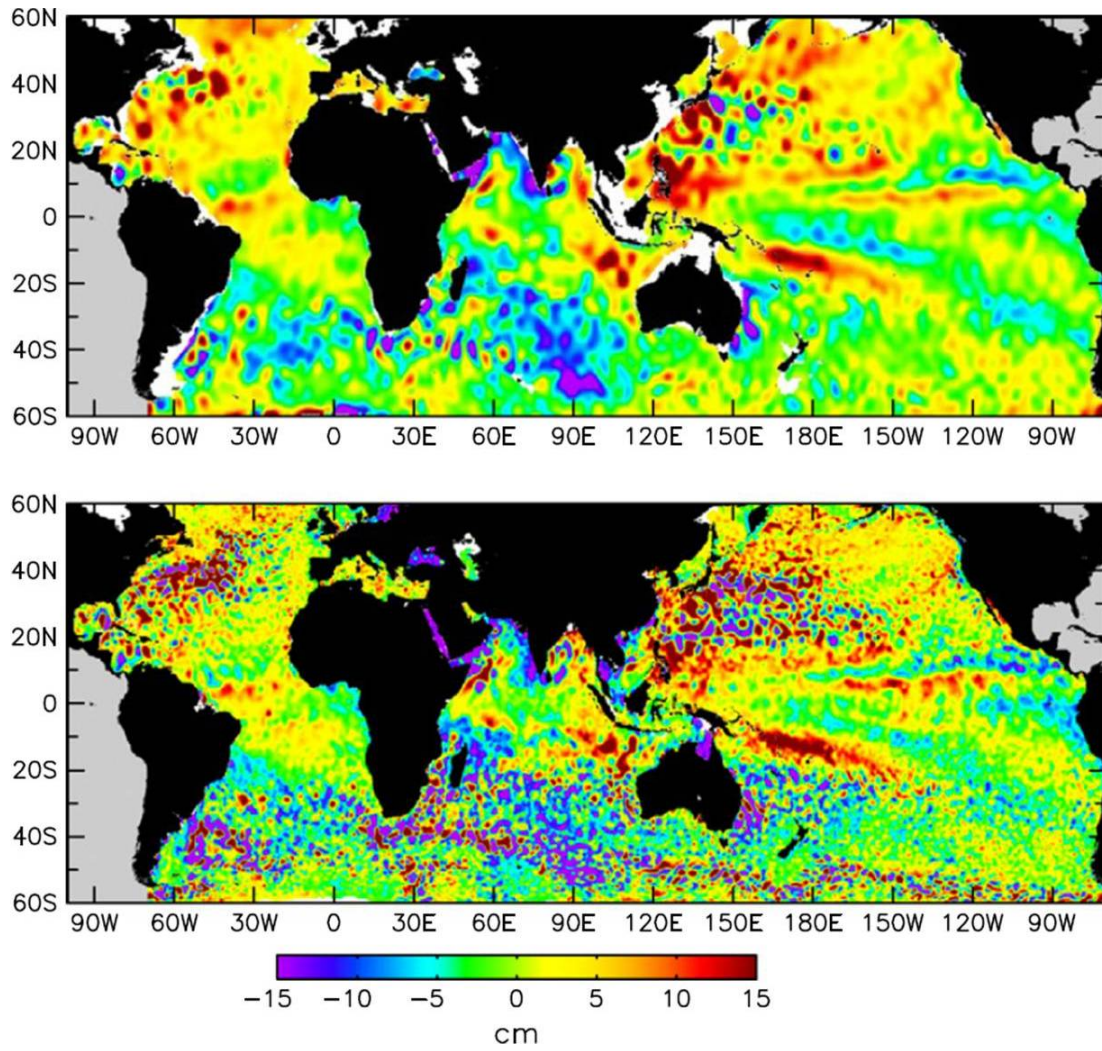


Fig. 3.7: A snapshot of global SSH variability on 28 August 1996 (with respect to a long-term mean), as seen by T/P alone (top panel) and a dataset merging observations from both T/P and ERS-1 altimeters (bottom panel). From Chelton et al. (2011b).

T/P-only observations and will therefore form the basis for the present analysis.

3.2.3 This study's multi-altimeter SSH dataset

We use a multi-altimeter dataset provided by AVISO (Archiving, Validation and Interpretation of Satellite Oceanic Data)⁴, specifically the merged near-real time dynamic topography.

In the June 2002 to December 2007 period analysed here, it is based on measure-

⁴ The altimeter products were produced by Ssalto/Duacs and distributed by AVISO with support from Cnes.

ments from at least three altimeters, JASON-1 in the 10-day repeat orbit, GFO in the 17-day repeat orbit and ERS-2/Envisat in the 35-day repeat orbit. Four altimeters contribute in the three year period from September 2002 to the end of T/P's life in October 2005, during which data from T/P on its tandem orbit with JASON-1 are also incorporated.

Low-pass filtered along-track data from all available altimeters are, after the removal of a consistent mean profile, merged into a global map through optimal interpolation (Ducet et al., 2000). The details of the data processing, merging and mapping process applied by AVISO are described on its website (www.aviso.oceanobs.com) and in the SSALTO/DUACS user handbook⁵. The resulting merged dataset is provided every 3-7 days on a $1/3^\circ$ Mercator grid, which corresponds to a grid spacing in both zonal and meridional directions of ~ 30 km in midlatitudes, ~ 25 km at 50° and ~ 15 km at 65° .

The effective *temporal* resolution of the merged dataset is limited both by the covariance function used in the optimum interpolation, whose temporal e-folding scale is set to a uniform 15 days in the extra-tropics, and by the temporal sampling period of the altimeter repeat orbits of at best 20 days. Higher frequency barotropic ocean variability forced by atmospheric pressure and wind fluctuations is removed by a high-resolution Dynamic Atmospheric Correction (DAC). Together with corrections for altimeter-specific measurement errors and tidal variability, this correction is applied to individual along-track data before the merging process. Whereas at low frequencies (> 20 days) it only removes the static, inverted barometer, response of the ocean to atmospheric pressure forcing, at high-frequencies the correction is dynamic. Estimated from an ECMWF-forced (6-hour, $1/4^\circ$ surface pressure and wind speed) barotropic model with enhanced resolution over continental shelves (Carrere and Lyard, 2003), it leads to a significant reduction of the aliased high-frequency variability, especially close to coasts and at high latitudes. Residuals of high-frequency variability, together with residuals in satellite orbit errors that remain after the multi-mission cross-over minimization (Le Traon and Ogor, 1998), lead to long wavelength errors which are minimized in the multi-mission optimum interpolation process (Le Traon et al., 1998).

The effective *spatial* resolution of the merged dataset is set by the spatial sampling by the 3 to 4 contributing altimeters, as well as by the processing and mapping specifics

⁵ The dataset used in this study has been obtained from AVISO in January 2009. The corresponding user handbook is Dibarboure et al. (2009). The current version of the handbook can be obtained at www.aviso.oceanobs.com/fileadmin/documents/data/tools/hdbk_duacs.pdf

described in Ducet et al. (2000). The low-pass filtering of along-track data prior to the merging process, designed to reduce measurement noise, employs cut-off wavelengths of 100 km in mid-latitudes and 70 km poleward of 50°. The optimum-interpolation uses isotropic covariance functions whose zero-crossing scale is set to decrease with latitude, from ~ 150 km in mid-latitude WBC systems to just below 100 km on the southern edge of the ACC. The mapping process therefore effectively eliminates the short-wavelength variability present in the along-track data (Ducet et al., 2000). When merging data from only two altimeters this corresponds to wavelength scales below 3° (Chelton et al., 2011b), or about 150 to 200 km in the Gulf Stream region (Ducet et al., 2000). By applying the mapping procedure to high-resolution model outputs that are re-sampled along altimeter ground tracks, Le Traon et al. (2001) have shown that incorporating data from 4 instead of 2 altimeters leads to a significant reduction of mapping errors, of up to a factor two in regions of high mesoscale variability such as the Gulf Stream system. The implied enhancement of accurately resolved spatial scales has been demonstrated by Pascual et al. (2006).

In the near-real time version of the merged dataset, which is used here (based on v8.1 of the Duacs-NRT software), only half the information (past, not future) is available for the mapping process. Together with a lower quality orbit estimate available in near-real time, this reduces the accuracy (and therefore effective resolution capacity) of the near-real time dataset compared to the delayed-time version. Future work should therefore test the robustness of the conclusions presented here against an analysis based on the delayed-time dataset. Such a reanalysis will be especially valuable with the 2010 completely reprocessed version of the delayed-time dataset, that is provided daily (instead of weekly) and is presently available until January 2011, resulting in almost 9 years of homogeneous-quality multi-mission SSH data contemporary with global through-cloud SST observations.

This analysis is based on a dataset of absolute dynamic topography η_d , which is obtained by AVISO by adding a mean dynamic topography, based on in-situ and GRACE data (Rio et al., 2005), to along-track data before the merging process. Although, in principle, this makes studies of the nature of eddies possible, due to possible inaccuracies in η_d related to an uncertain estimate of the geoid, the mean dynamic topography component of the dataset is only considered for illustrative purposes. (It is further discussed in comparison to other such datasets in Appendix B.4). The key parts of the analysis presented hereafter are restricted to anomalies of SSH, denoted as SSHA or η' . For consistency, these are constructed in the same way as SSTA by the removal

of the seasonal climatology $\bar{\eta}$, which is obtained by averaging merged altimeter data over the study period after their linear interpolation to a daily time-grid. (For correlation analyses with SSTA presented below, a small remaining temporal mean and a linear trend are additionally removed from both anomaly datasets, and SSHA are linearly interpolated to the $\frac{1}{4}^\circ$ SST grid.) A discussion of the resulting SSH seasonal climatology and the definition of SSHA as deviations from the latter is provided in Appendix A.2. The observed SSHA variability itself and the processes that set its distribution are introduced in section 3.3.1.

3.2.4 A global dataset of nonlinear mesoscale eddy tracks

A global dataset of the observed tracks of mesoscale eddies, defined as propagating SSHA, is established by an automated two-step, eddy identification and tracking algorithm that has been developed and described in detail by Chelton et al. (2011b, referred to as CSS11 in the following).

The algorithm identifies eddies as closed contours in spatially high-pass filtered SSH (retaining scales smaller than 20° longitude by 10° latitude). It thus prevents the noise-amplifying differentiation required by previous vorticity-based eddy identification methods (Isern-Fontanet et al., 2003; Chelton et al., 2007). The closed contour criterion moreover allows to identify eddies even when they are embedded in a surrounding perturbation in SSH of larger scale and amplitude. Since the algorithm is not applied to a full SSH field, identified eddies do not necessarily correspond to closed streamlines in the absolute surface flow, but may in principle also retain wave-like or meandering features. Inspection of tracked eddies however reveals that in $>98\%$ of observations of eddies, with lifetimes of at least 16 weeks and poleward of $\pm 20^\circ$, the ratio of eddy rotational to translational speeds U/c is greater than unity (CSS11, their Fig. 16). The bulk of long-lived extratropical tracked eddies is thus nonlinear, able to trap and carry water parcels with them. Tracks with lifetimes shorter than the 4-month threshold may however well be less nonlinear. Further discussion of the nature of tracked eddies is provided in chapters 5 and 6.

The algorithm tracks eddies by connecting closest identified eddies between weekly maps, requiring them to be not too different in size and amplitude and to lie within a circle of 150 km radius that is centred on the previous week's eddy centre location⁶. In

⁶ In low-latitude regions (approximately equatorward of 15°N and S), in which $1.75\Delta s > 150$ km (where Δs gives the weekly propagation distance of non-dispersive Rossby waves – cf. Appendix B.6), the search region is extended westward into a zonally-elongated westward pointing ellipse, whose

the regions focussed on in this study (cf. section 3.3.2), the search radius is very large compared to the weekly propagation distance of long first-mode baroclinic Rossby waves ($\approx 1\text{-}5$ km in the ACC, $5\text{-}10$ km in the GS and $10\text{-}50$ km in the NA subtropical gyre interior – cf. Appendix B.6). The tracking algorithm is therefore expected to realistically capture a broad range of eddy propagation statistics, including poleward, equatorward and eastward propagation. Only in the cores of the narrowest GS and ACC jets, whose speeds can exceed $25\text{ cm/s} \approx 150\text{ km/week}$, eddies may be advected eastward fast enough to be lost by the algorithm, which may result in artificially short tracked eddy lifetimes in these regions. Whenever results rely on eddy lifetime or propagation statistics, it is thus important to keep in mind their sensitivity to the specifics of the eddy tracking algorithm. The bulk of the analysis presented in this thesis, however, primarily relies on the eddy identification part of the algorithm.

Note that the algorithm only retains eddies with amplitudes of at least 1 cm and lifetimes of at least 4 weeks. Note moreover that, to generate the eddy track dataset described in CSS11, the tracking algorithm is applied to a different SSHA field (the $1/4^\circ$ AVISO reference series, which merges observations from two altimeters, one in the 10-day and the other in the 35-day repeat orbit, and from which SSHA are obtained by spatial high-pass filtering) compared to the one used in this study for further analysis of the SSH variability (as described in section 3.2.3 above, the $1/3^\circ$ Mercator near real-time AVISO updated series, from which SSHA are obtained by removal of a seasonal background state). This is an inconsistency in the analysis and further work will need to test the robustness of results with respect to the specifics of the SSHA dataset used.

The resulting eddy track dataset, version 091029 of which is used here, provides eddy coordinates $s(x,y,t)$, as well as measures for the eddy amplitude and radius. Since the beginning of 2011, an updated version of the dataset is publicly available at <http://cioss.coas.oregonstate.edu/eddies/>.

Tracked eddies are observed to originate nearly everywhere in the world’s ocean and partition into roughly equal numbers of cyclones and anticyclones (characterized respectively by lows and highs in sea level) with average amplitudes of $\sim 10\text{ cm}$, radii of $\sim 100\text{ km}$ and lifetimes of ~ 6 months. The subset of tracked eddies analysed in this study is introduced in terms of eddy amplitude, radius, rotational velocity, lifetime, propagation speed and direction in section 3.3.3. Further analysis is provided throughout the later chapters.

western extremum lies at a distance of $1.75\Delta s$ from the previous week’s eddy centre.

3.3 Preliminary diagnostics of observed SSH and eddy variability

Here we introduce and examine the nature of the observed SSH variability in NA and SO basins (section 3.3.1), which leads to the identification of two contrasting oceanic regimes (section 3.3.2), whose propagating eddy variability is then introduced (section 3.3.3). Further analysis of fixed-location, as well as track-following SSH variability is provided in the following chapters.

3.3.1 Overview of observed NA and SO SSH variability

This section provides an overview of the variability observed in the AVISO multi-altimeter SSHA dataset, introduced in section 3.2.3, for the NA and SO basins.

Figure 3.8 displays the rms amplitude of the fluctuations in SSH, $\sigma(\eta') = \sqrt{\overline{\eta'^2}}$, observed over the 5½-year study period. SSHA amplitudes are likely underestimated in this dataset, not only due to its imperfect mesoscale resolution capacity, but also, and specifically in the regions of highest SSHA rms, due to a fraction of the non-seasonal SSH variability retained in the estimate of the (seasonal) background state $\bar{\eta}$, with respect to which the observed SSH fluctuations are defined. (Further discussion of this is provided in Appendices A.2 & A.3). As seen in the Figure, over much of the tropical, subtropical and subpolar interiors of ocean basins, and specifically their eastern parts, SSH variability is observed to be weak, typically limited to less than 5 cm. The variability gradually increases on approaching the western parts of subtropical gyres, and the ACC in the SO, to typically ~ 10 cm. Apart from over the Mexican Gulf’s loop-current, peaks in SSHA rms that can exceed 20 cm are found in the world’s major WBC systems and their extensions, specifically along the separated GS and, in the SO, along the Agulhas return current (ARC) and in the Brazil-Malvinas confluence (BMC). Along the ACC itself, SSHA rms rarely exceeds 15 cm.

To get an idea of corresponding surface geostrophic eddy kinetic energy (EKE) and flow speeds ($V = \sqrt{2 \text{ KE}}$), where eddy denotes time variability here, Fig. 3.9a,d display the time-average magnitude of the observed SSHA gradients $|\overline{\nabla \eta'}|$. For a typical mid-latitude eddy, $|\nabla \eta'| = 10 \text{ cm}/100 \text{ km}$ and $V = \frac{g}{f} |\nabla \eta'| \sim 10 \text{ cm/s}$. Observed $|\nabla \eta'|$ magnitudes vary by nearly an order of magnitude to both sides of this typical value, and show a spatial distribution that closely coincides with the observed SSHA rms. Observed SSHA therefore reflect time-varying surface flows with typical speeds of only 1 cm/s in eastern gyre interiors, but more than 1 m/s in the regions of the

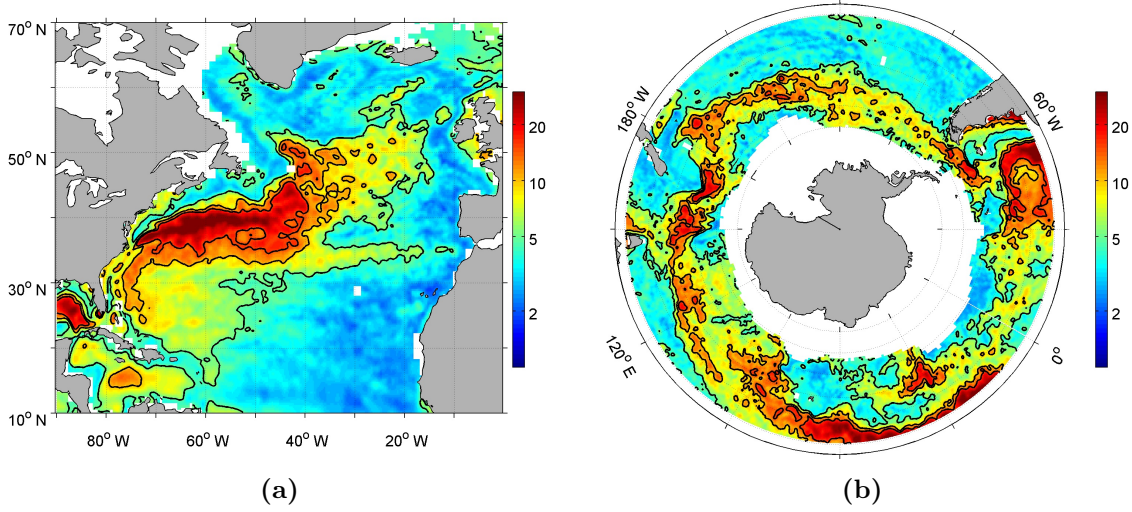


Fig. 3.8: SSHA rms $\sigma(\eta')$, mapped on a logarithmic colour-scale. Contours highlight the 5, 10 and 15 cm levels for a) NA, b) SO.

largest SSHA rms. The ratio of the time-varying to the $5\frac{1}{2}$ -year mean SSH gradient magnitudes is displayed in the rhs panels of Fig. 3.9. It reveals that the kinetic energy of the time-varying flow nearly everywhere exceeds that of the time-mean flow by more than one order of magnitude ($EKE/\overline{KE} = (|\nabla\eta'|/|\nabla\bar{\eta}|)^2$). Apart from some high-latitude and continental shelf regions, this is true everywhere the mean-flow occurs on large, basin scales. However, where the $5\frac{1}{2}$ -year mean flow occurs itself on mesoscales and forms narrow meandering jets, which are clearly apparent in $|\nabla\bar{\eta}|$ displayed in the middle panels of Fig. 3.9, its flow speeds become comparable to or exceed those of the time-varying flow. Although mean flows are expected to contain energy on mesoscales even in infinitely long time series, we may expect a different partition, biased towards an even larger contribution of EKE with respect to $\overline{KE}$, in longer time series than that used here.

As illustrated in Figure 3.10, the distribution of SSHA rms is strongly tied to that of the background mean flow and the observed planetary vorticity field. The regions of most intense SSH variability, indicated by the areas speckled in white, are observed to coincide with regions of tightly spaced contours of the $5\frac{1}{2}$ -year average SSH, denoted $\bar{\eta}$ in the following. As discussed in chapter 2.1, these typically correspond to regions of steeply sloping isopycnals and large amounts of APE, in which, from comparison to Fig. 2.7a, baroclinic disturbances grow fastest.

However, especially at high-latitude, where the stratification is weak, the mean flow is not purely baroclinic, but extends all the way down to the sea floor, where its interaction with the bathymetry (of which a dataset is described in Appendix B.7) also plays

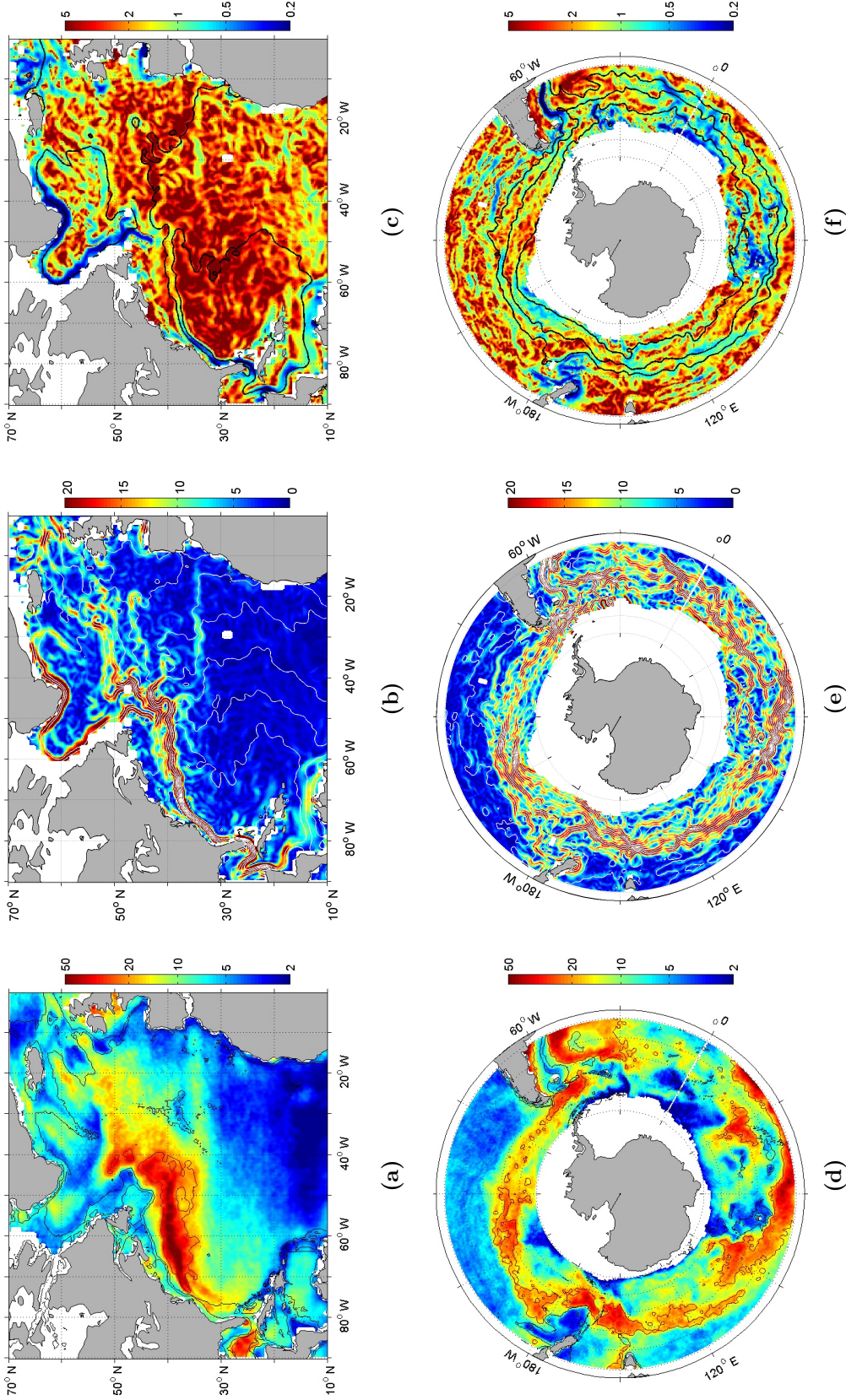


Fig. 3.9: Lhs panels display $|\nabla\eta'|$, in $\text{cm}/100\text{km}$, which is a direct proxy for typical surface geostrophic speeds $V = \frac{g}{f} |\nabla\eta'|$ and surface kinetic energy $EKE = V^2/2$, so that in midlatitudes a gradient of $10 \text{ cm}/100 \text{ km}$ corresponds to $V \sim 10 \text{ cm/s}$ and $EKE \sim 50 \text{ cm}^2/\text{s}^2$. **Middle panels** display the corresponding quantity for the 5½-year mean flow $|\nabla\bar{\eta}|$ (also in $\text{cm}/100\text{km}$, but on a different colour scale), and **rhs panels** display their (dimensionless) ratio $|\nabla\eta'|/|\nabla\bar{\eta}|$. Moreover, in (a,d) 200 and 2000 m isobaths are contoured in black, and NA 15 cm and SO 10 cm SSHA rms levels in red. In (b,e) $\bar{\eta}$ is contoured every 10 cm in white, and in (c,f) a few $\bar{\eta}$ contours are displayed in thick black (-45, 5 and 45 cm in the NA, -55, 0 and 40 cm in the SO).

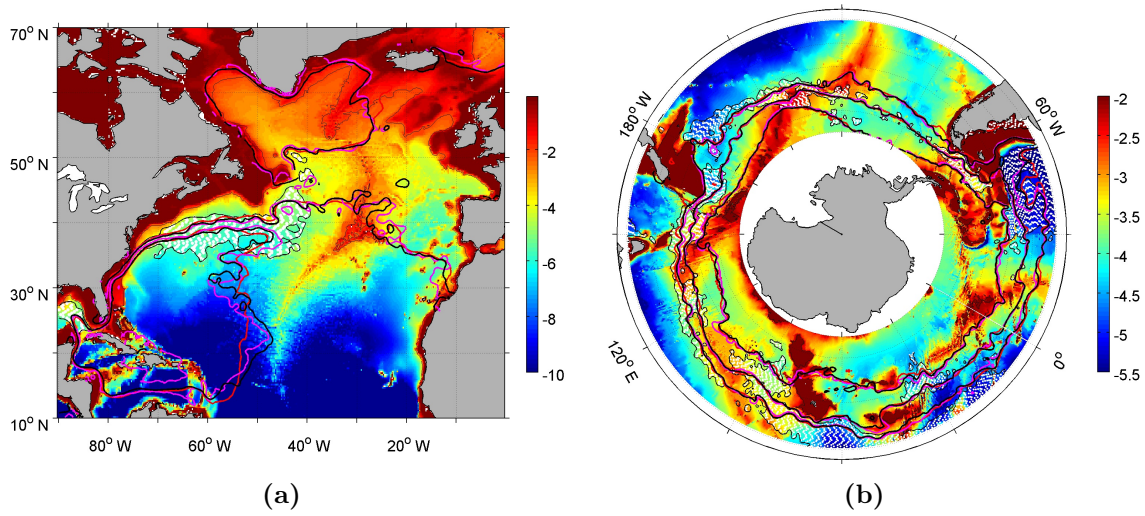


Fig. 3.10: White-speckling highlights regions of most intense observed SSH variability, in which SSHA rms exceeds 15 cm in the NA (a), and 10 cm in the SO (b). These are placed in the context of the observed planetary vorticity field (colours display its inverse H/f , in $km \times 10^4 s$) and the observed surface geostrophic mean flow (represented by contours of $\bar{\eta}$ (black): -45, 5 and 45 cm in the NA, -55, 0 and 40 cm in the SO. For comparison, corresponding contours of two other mean flow datasets, $\bar{\eta}_{CLS}$ (magenta) and $\bar{\eta}_{MAX}$ (red), described in Appendix B.4, are also shown.)

a role in the generation of SSH variability. Comparison between $\bar{\eta}$ and the background planetary potential vorticity field, whose inverse H/f is displayed by the colour-map in Figure 3.10, indeed reveals that the strongest gradients of $\bar{\eta}$ coincide with regions of intense bathymetric constraints on the flow. These are observed over the GS before its separation from the coast, and along the path of the ACC where it encounters, and is forced around, major SO bathymetric features, such as the Kerguelen plateau (75°E), the South-east Indian Ridge (south of Tasmania), the narrow Macquarie Ridge (south-west of New Zealand) or the Pacific Antarctic Rise. Whereas these regions of peak jet strength themselves are not associated with particularly elevated levels of SSH variability, SSHA rms is observed to systematically peak downstream, where the H/f field ceases to be such a strong constraint and jets are observed to widen again. This pattern can be rationalized in terms of baroclinic disturbances growing as they are advected downstream of the regions of highest instability and strongest $\bar{\eta}$ gradients, in which they are generated. Alternatively, the generation of SSH variability could also be thought of as a way to dissipate the relative vorticity acquired by the flow in order to permit its crossing of background H/f contours (Moore et al., 1999).

Apart from time-varying flows of the oceanic mesoscale, which encompass meandering currents, growing waves and fully-developed non-linear eddies, SSH variability can also reflect direct atmospheric forcing through fluctuations in air-sea buoyancy fluxes and wind-stress (curl). (Tidal and atmospheric pressure forcing are removed from the dataset.) The excited response in SSH can be local, through mainly steric adjustment, or non-local, through forced barotropic and baroclinic Rossby waves (Stammer, 1997b). These can lead to basin-scale adjustments of the observed SSH field to atmospheric variability on time scales longer than the waves take to cross the basin. Whereas the fast barotropic adjustment is also, to some extent, removed from the dataset (section 3.2.3), the $\rho/\Delta\rho \sim 300$ times slower propagation of baroclinic waves limits baroclinic basin-scale adjustments, examples of which are visible in several-year to decadal trends in the observed SSH field (Stammer, 1997a; Haekkinen and Rhines, 2004; Wunsch et al., 2007), to interannual time scales. This atmospherically-forced SSH variability is expected to occur on scales that are large compared to the oceanic mesoscale, but comparable to those of atmospheric variability, that is on scales of ~ 1000 km for forcing by synoptic fluctuations, and on scales comparable to ocean basins for the oceanic adjustment to the principal modes of interannual atmospheric variability.

As a given SSH anomaly of amplitude $\sigma(\eta')$ and spatial scale $L_{\eta'}$ is associated with a surface geostrophic speed anomaly proportional to $|\nabla\eta'| \sim \frac{\sigma(\eta')}{L_{\eta'}}$, we can estimate the typical spatial scales of the observed surface geostrophic flow variability as $L_{\eta'} = \sigma(\eta')/|\nabla\eta'|$. As the gradient operator involved in its calculation is a spatial high-pass filter, $L_{\eta'}$ cannot be used to assess the respective contributions of (atmospherically-forced) large-scale versus mesoscale signals to the variability in SSH (this will be assessed more carefully in the next chapter), but only to the variability in the SSHA slopes/surface currents. Displayed Fig. 3.11a,b, $L_{\eta'}$ reveals that the geostrophic flow variability predominately occurs on spatial scales close to 100 km (~ 50 – 150 km), and thus on the oceanic mesoscale.

$L_{\eta'}$ reveals several interesting regional features. They include a line of relatively small scales (blue in Fig. 3.11a and b) along the continental shelf edge (indicated by the thin black 2000 m isobath contour), that is especially pronounced off the eastern coasts of North America, Argentina and the Campbell plateau and separates the variability on continental shelves from that in deep basin interiors. Moreover, at relatively high latitude, red spots in $L_{\eta'}$, found along the edges of the NA subpolar gyre, in the middle of the SP, the BMC and over SO bathymetric plateaus (Kerguelen, Campbell, Mid-

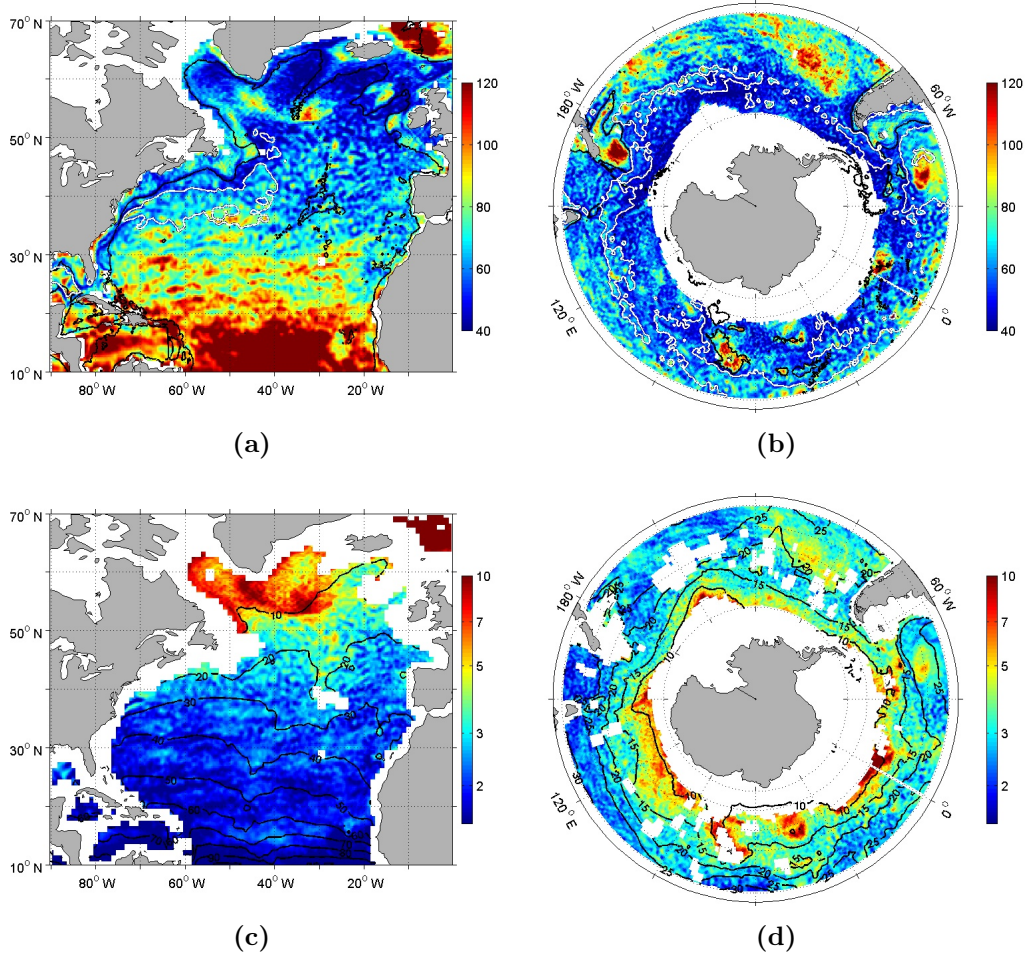


Fig. 3.11: Top panels map the predominant spatial scale of observed surface geostrophic flow variability $L_{\eta'} = \sigma(\eta')/|\nabla\eta'|$, in km (with the 15 and 10 cm SSHA rms contoured white in NA and S0, and the 2-km isobath contoured black). Bottom panels map $L_{\eta'}/R_d$, the ratio of $L_{\eta'}$ and the deformation radius R_d , and contour R_d (in km).

Atlantic Ridge), point out isolated regions predominated by much larger spatial scales than their surroundings. Comparison with Fig. 3.10 reveals that they approximately coincide with closed $\frac{1}{2}$ contours, in which wind-stress curl forcing is efficient in forcing high-frequency and large-scale barotropic SSH variability (cf. also Hughes et al., 2010).

On top of these, $L_{\eta'}$ shows a systematic decrease in the scales of geostrophic flow variability from low to high latitudes. The ratio of $L_{\eta'}$ to the first baroclinic deformation radius R_d (Appendix B.6), displayed in the lower panels of Fig. 3.11, shows that in low- and mid-latitudes the latitudinal dependence of $L_{\eta'}$ approximately follows, but is slightly weaker than the $\sim \frac{1}{2}$ dependence of the deformation radius. A relatively sudden change of this relationship is observed at very high-latitude, where

$L_{\eta'} \sim 10 \times R_d$. Unless the nature of observed SSH variability changes rather abruptly in these regions, which roughly coincide with the subpolar gyres, this might indicate an insufficient capacity of the SSH dataset to resolve mesoscale processes in the NA subpolar gyre and south of the ACC, where Rossby radii drop below ~ 10 km.

Throughout low and mid-latitudes, $L_{\eta'}$ varies between $\sim 1 - 5 \times R_d$, and is thus close to the deformation scale $L_d = \pi R_d$ (cf. Appendix B.6). In particular, subtropics display ratios ≤ 2 (or $L_{\eta'}/L_d < 1$), whereas GS and ACC show ratios between $\approx 3-5$ (or $L_{\eta'}/L_d \geq 1$). This indicates that $L_{\eta'}$ is not far off from the scale of maximum linear growth through baroclinic instability (Fig. 2.7b), and follows the latter's increase from subtropical return flows to strong eastward currents, but with a by far less clear-cut separation. In agreement with the analysis of SSH spectra by Tulloch et al. (2011), but subject to confirmation with better and longer datasets, this suggests that there is not much room for a growth in the energy containing scale through nonlinear eddy interactions, whose importance is a topic of active research (e.g. Scott and Wang, 2005).

In summary, observed $L_{\eta'}$ suggest that mesoscale processes are adequately resolved in the SSH dataset across mid-latitudes, the major WBC regions and the ACC, and that in the deep ocean, apart from a few regions of large-scale dominance, the observed geostrophic flow variability primarily occurs on oceanic mesoscales, and is consistent with its generation through baroclinic instability.

3.3.2 Introducing two contrasting oceanic regimes

Based on the above discussion of observed fixed-location SSH and mesoscale eddy variability, it is of use for further analyses to introduce two contrasting oceanic regimes: On the one hand the regions of intense observed mesoscale eddy variability associated with the major current systems, on the other hand regions of weak eddy variability, for which we choose to concentrate on the quiescent eastern parts of the observed subtropical gyres.

Here, the former, “energetic” regions of GS and ACC, are defined by imposing a lower bound on SSHA rms of 15 cm in the NA, and 10 cm in SO. The choice of a lower value for ACC regions reflects its generally lower-amplitude SSH fluctuations compared to WBC systems. The latter, “quiescent” eastern subtropics of NA and South Pacific (SP), are defined as the regions east of 60°W , $20^\circ\text{--}32^\circ\text{N}$ in the NA, and east of 150°W , $40^\circ\text{--}48^\circ\text{S}$, in the SP. The regimes' boundaries, such defined, are mapped by the black contours in Figure 3.12.

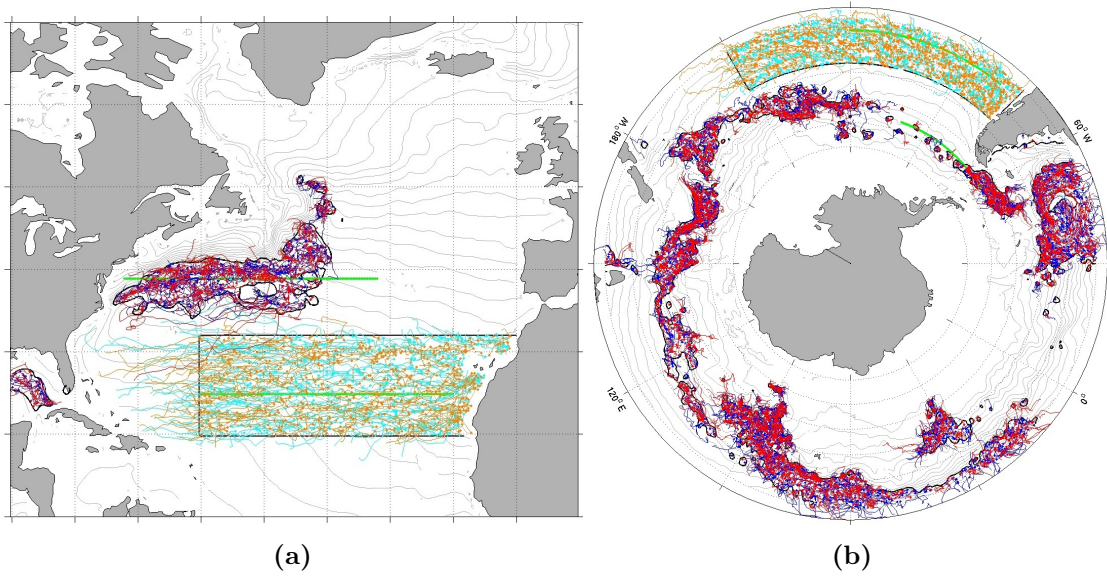


Fig. 3.12: The tracks of propagating eddies with origins in energetic and quiet regimes of NA (a) and SO (b), displayed over the background SST distribution (thin grey contours, every 1K). The energetic regime is defined by an SSHA rms criterion (≥ 15 cm in NA, ≥ 10 cm in SO), the quiet regime as the region east of 60°W , 20° - 32°N in the NA, and east of 150°W , 40° - 48°S in the SP. The tracks of anticyclones and cyclones with origins (dots) in the energetic regime are shown in red and blue lines, those with origins in the quiet regime in orange and cyan lines. (Green zonal sections indicate the location of the longitude-time plots shown in Figures 4.8 and 4.11.)

3.3.3 Overview of the regimes' observed eddy variability

Figure 3.12 also maps the tracks of propagating mesoscale eddies in the SSHA field that are observed to originate in these two regimes. They are detected and tracked by the automated tracking algorithm, as developed by CSS11 and described in section 3.2.4. With 99% of tracked extratropical eddies characterized by $Ro < 0.1$, identified eddies are typically in a near-geostrophic balance at the surface (CSS11, their Fig. 14). During the 5½-year study period, the algorithm identifies a roughly equal number of anticyclones and cyclones in each regime. Slightly more than 500 eddies are observed to originate in the energetic GS system. With typical eddying geostrophic flow speeds of 25 - 40 cm/s , these are amongst the highest amplitude eddies of the world oceans. Consistent with the fixed-location distribution of SSHA rms and eddy kinetic energy (Figs. 3.8 & 3.9a,b), the nearly 4000 eddies with origins along the high-variability regions of the ACC have slightly lower amplitudes, rotating typically at 15 - 20 cm/s . Each

year, on average ~ 100 eddies of much weaker amplitude ($\sim 5 \text{ cm/s}$) are detected to be generated in the quiescent subtropics of the NA, slightly more (~ 150) in that of the SP.

The above estimates of the typical eddy surface geostrophic flow speed, V_e , are obtained from observations of the eddy amplitude A_e and scale L_e , as $V_e = gA_e/|f|L_e$. Here, the eddy radius-scale, L_e , approximates the radius of maximum eddy rotation (i.e. it gives the radius of a circle whose area is equal to that enclosed by the SSHA contour along which the geostrophic speed anomaly is observed to peak). L_e is observed to differ little between anticyclones and cyclones ($\leq 5 \text{ km}$), and amounts to $\sim 90 \text{ km}$ over NA eddies, and to slightly less, $\sim 80 \text{ km}$, over the on average higher latitude SO eddies. Over GS and SP eddies (at $\sim 40^\circ\text{N,S}$), L_e is thereby about equal to the regions' average deformation scale $L_d = \pi R_d$ (cf. Appendix B.6). Over the on average higher latitude ACC eddies L_e is slightly larger than L_d ($L_e/L_d \approx 1.25$ and $L_d/\pi \approx 20 \text{ km}$), whereas over the lower latitude eddies with origins in the NA subtropics it is slightly smaller ($L_e/L_d \approx 0.7$ and $L_d/\pi \approx 45 \text{ km}$). Eddy scales are thus typically π times larger than R_d , but vary less strongly with latitude, in agreement with recent results of Tulloch et al. (2011).

The eddy amplitude, A_e , is estimated as the difference between the SSHA peak within the eddy interior and the large-scale background SSHA averaged around the eddy's outermost closed SSHA contour, as detected by the tracking algorithm. Eddy amplitude A_e and speed V_e are thus by definition positive over anticyclones and negative over cyclones. The distribution of eddies' V_e , observed in each weekly snapshot along their tracks (i.e. >50000 snapshots of energetic, >30000 of quiet regions' eddies), is displayed in Figure 3.13. For eddies in both regimes, the observed distribution shows a pronounced skew, positive over anticyclones and negative over cyclones. It indicates that, for eddies of each polarity, observed average V_e amplitudes are dominated by relatively rare high-amplitude eddies, whereas weak eddy events are more frequent. Whereas in the quiet regime, anticyclones and cyclones are observed to “rotate” on average at similar, slow speeds, the intense eddies observed in the energetic regime show a clear asymmetry between on average faster rotating cyclones (22 cm/s) and slower anticyclones (16 cm/s). CSS11 attribute this to ageostrophic effects (gradient wind) that become important only over high-amplitude eddies and work to strengthen cyclonic lows and weaken anticyclonic highs.

Irrespective of the regime in which eddies originate, tracked eddies are observed to live on average between 3 and 4 months. The frequency of occurrence of different

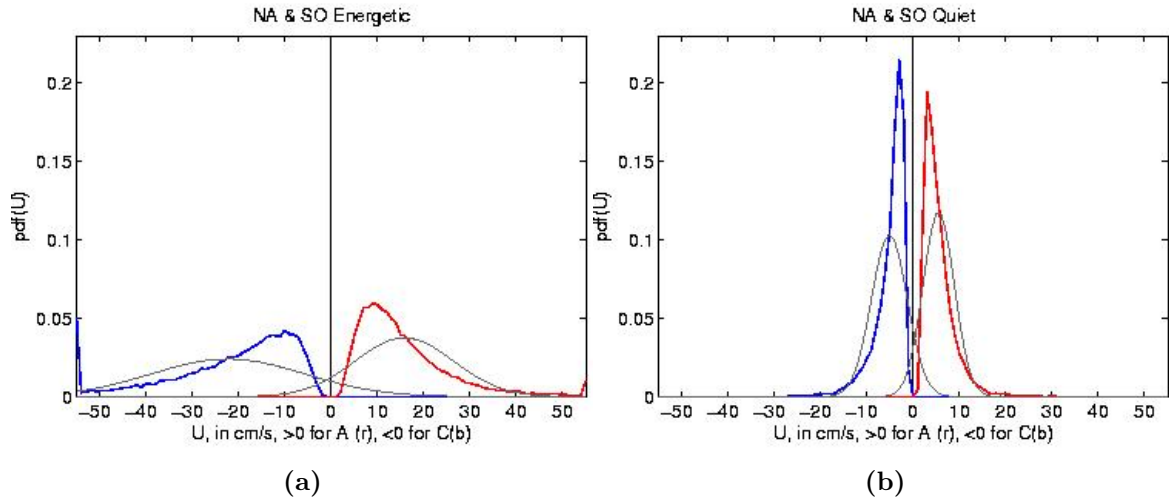


Fig. 3.13: Probability distribution of the eddy geostrophic flow speed V_e , observed over weekly snapshots of anticyclones (red) and cyclones (blue) with origins in the energetic (a) and quiet regime (b) of NA and SO. (The tracks of these eddies are displayed in Fig. 3.12). The thin grey curves show the normal distributions of equal mean and standard deviation as observed V_e distributions.

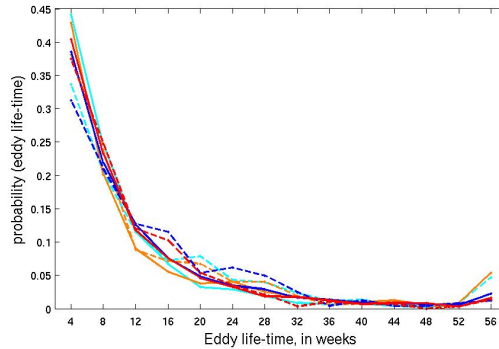


Fig. 3.14: Observed occurrence of different tracked eddy lifetimes, displayed separately for anticyclones & cyclones with origins in energetic (red & blue) and quiet (orange & cyan) regions of NA (dashed) and SO (solid).

tracked eddy lifetimes, displayed in Figure 3.14, also shows a substantial (positive) skew. Thus, the bulk of tracked eddies lives shorter than average, and is trackable over only a few weeks. (Only eddies tracked for at least 4 weeks are retained by the algorithm.) Still, in both regimes and basins considered, a substantial number of eddies is observed to live longer than half a year, and a few eddies are tracked for as long as 2-3 years.

A closer inspection of Figure 3.12 reveals insights into the nature of the eddies'

propagation during their lifetime. In quiescent interiors, tracked eddies are observed to propagate nearly due west. As shown by CSS11 (their Fig. 22), here eddy propagation speeds differ little, by less than 20%, from the local long baroclinic Rossby wave phase speed c_{lrw} (Appendix B.6), which is also apparent from the longitude-time plots of SSHA through these regions (taken along the green dotted lines in Fig. 3.12, and displayed in Fig. 4.11 in the next chapter). The movements of eddies with origins in energetic regions are more complicated. They show both meridional and zonal displacements, and, instead of pure westward propagation, they are often characterized by episodes of eastward displacements when advected by the strong background flow of the GS or the ACC. This leads to the complex patterns observed in the SSHA longitude-time plots across the GS and Drake passage, displayed in Fig. 4.8. A more detailed analysis of the nature of observed eddy propagation is provided in chapter 5.3.

3.4 Summary

This chapter has introduced observational datasets of SSTA and SSHA, as well as a dataset of propagating SSHA tracks, whose mutual relationships will be analysed in the following. These datasets are based on multi-mission altimetry and through-cloud MW radiometry, which have both revolutionized our observing capacities of mesoscale signals at the sea surface in the recent years. Nevertheless their resolution remains restricted to scales of roughly ≤ 50 km and 20 days. Specifically as the SSH dataset is not the best currently available, this calls for future sensitivity tests of the results to be presented in this thesis. As available time series lengthen, this will also allow for an improved separation between time-variability and seasonal background states.

The observed SSH variability has been shown to be associated with vigorous mesoscale surface currents, whose energy levels drop off rapidly from major current systems to gyre interiors, and these two extremes, GS and ACC energetic regions versus quiescent subtropical interiors, will form the focus of the analysis to follow.

Chapter 4

THE OBSERVED MESOSCALE CONTRIBUTION TO SST VARIABILITY

This chapter addresses whether the transient oceanic mesoscale circulation systematically contributes to the observed variability in SST. Section 4.1 provides an illustration of the observed SST variability in NA and SO and sets out the key questions of the chapter as well as the strategies for their solution. While revisiting the association of the observed SSH variability with eddying circulations, the observed SST-SSH relationship is analysed, both at fixed-location (section 4.2) and following eddy tracks (section 4.3). Section 4.4 presents evidence for a seasonal modulation of the mesoscale contribution to SST variability, and analyses it as a tool towards a better understanding of the governing mechanisms. Results are discussed in comparison to previous observational studies and summarized in section 4.5.

4.1 Introduction to the observed SST variability

As introduced in chapter 2.2, sea surface temperatures shape the conditions at the atmosphere's lower thermal boundary and carry imprints of a large variety of processes on all scales. The time-mean SST field features distinct zones of intensified gradients, narrow fronts in SST. As discussed in chapter 1, these time-mean thermal surface expressions of the oceanic mesoscale provide a recently discovered pathway to ocean-atmosphere coupling in midlatitudes, where on large-scales the ocean is believed to respond primarily passively to atmospheric forcing.

Could such a coupling also occur over the meanders, waves and propagating vortices of the time-varying oceanic mesoscale circulation? A prerequisite to being able to address this question is to establish whether the transient mesoscale circulation has

the potential to become visible to the atmospheric boundary layer or whether the atmosphere merely sees a distorted mirror-image of itself through the signatures, its own fluctuations, on synoptic to interannual time-scales (cf. chapter 2.2.3), imprint on the ocean surface. Here the question of visibility does not ask what the atmosphere actually does or can react to, but which are the dominant processes that shape the variability at the atmosphere's lower thermal boundary. Does the transient oceanic mesoscale contribute significantly to observed SST variability?

Figure 4.1 displays a snapshot of the SST variability observed on a given day, the 31.12.2005 on this occasion, for NA and SO. Surface isotherms, indicated by the contours, are observed to undulate around their climatological mean position (which is displayed in Figure 3.3). They thereby create SST anomalies (SSTA) with respect to the background seasonal climatology, which are displayed in colour.

The amplitude of these SST fluctuations is typically ~ 1 K. Their rms amplitude for the whole 5½-year study period, $\sigma(T')$, displayed in Figure 4.2, ranges from less than 0.5 K to more than 2.5 K. As shown by the thin white contour that indicates the 0.4 K rms level, which is the conservative estimate for the measurement error on individual AMSR-E SST observations (cf. section 3.1.2), typical SSTA amplitudes nearly everywhere stand out from measurement noise. The distribution of SSTA rms reveals a systematic spatial pattern. Amplitudes are weak throughout the bulk of ocean basins, and apart from several coastal regions, only exceed the 1 K level in the vicinity of major western-boundary current (WBC) systems, the Gulf-Stream (GS) in the NA, the Brazil-Malvinas confluence (BMC) and the Agulhas Return Current (ARC) in the SO. The ACC is however the only place where the observed SSTA rms exceeds that of seasonal SST fluctuations $\sigma(\overline{\text{SST}})$ (which are displayed in Figure 3.4). In the GS region, the SSTA rms is only slightly larger than 50% of $\sigma(\overline{\text{SST}})$, and in basin interiors it is limited to $\sim 1/3$ of the seasonal SST variability.

The snapshot of SST variability in Figure 4.1 illustrates what the atmosphere sees on a given day, in departure from the quasi-steady background SST field that drives its mean circulation. It suggests that the nature of the observed SST variability and of the atmosphere's thermal boundary is fuzzy, carrying the simultaneous imprints of processes acting from meso- to basin scales everywhere. In many regions the snapshot reveals isotherm displacements that are coherent over large areas. The associated SSTA show large spatial patterns, at wavelengths comparable to the size of the basins. An example of such a large-scale wavelike SSTA pattern is very pronounced in the South Pacific in this snapshot. Although of weaker amplitude, a large-scale,

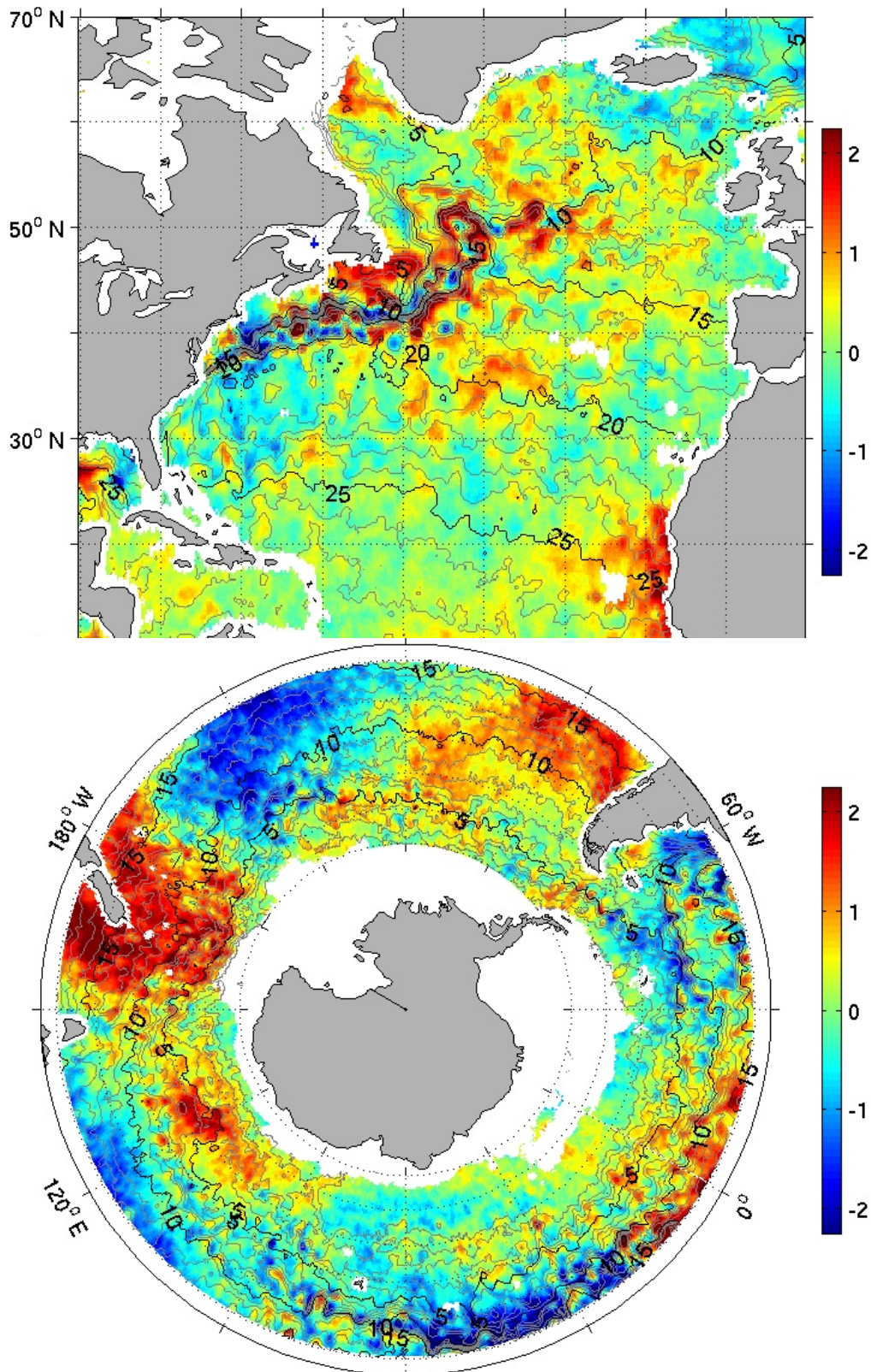


Fig. 4.1: A snapshot of NA (top panel) and SO (lower panel) SST variability observed on 31.12.2005. Contours display the full AMSR-E SST field (every 1°C) and colours the SST anomalies (SSTA) from the background seasonal climatology.

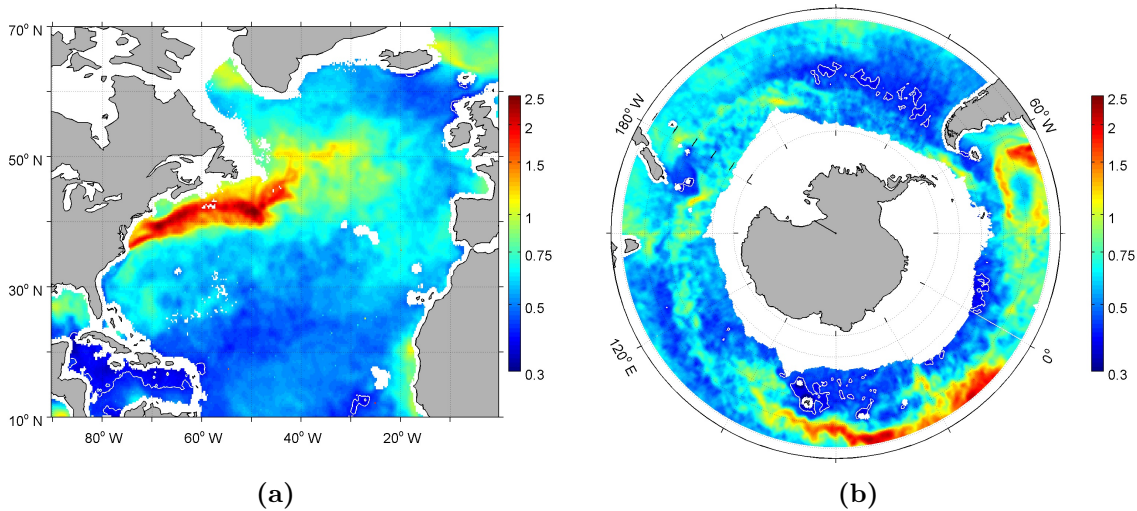


Fig. 4.2: The observed SSTA rms $\sigma(T')$ over the 5½-year study period, mapped on a logarithmic colour-scale, for NA (a) and SO (b). Typical SSTA amplitudes almost everywhere exceed the 0.4 K level, contoured in white, which corresponds to the estimated AMSR-E single-observation measurement error.

quadrupole-like pattern is also visible in the NA. Superimposed on these large-scale fluctuations in SST, is a smaller-scale fuzziness in the displacement of isotherms. From a visual examination, the associated smaller scale SSTA extend typically across ~ 100 km. Regionally, these smaller scale undulations in isotherms can reach very large amplitudes, until sometimes pinching off into isolated patches of closed SST contours, examples of which are found mainly in the GS system and along the ARC, but also occasionally in basin interiors.

This suggests that these small-scale SSTA, in both their wave-like and vortex-like forms, are the signatures of the transient oceanic mesoscale. However, in contrast to the well-documented time-mean picture, the contribution of transient mesoscale eddies to SST variability is so far less systematically established from observations (cf. section 4.5), and propagating eddies could instead have little impact on the observed SST field or, alternatively, their SST anomalies could be quickly eroded by turbulent air-sea heat exchanges, reflecting the proposed enhancement of the negative feedback between heat fluxes and SST towards smaller scales, which will be introduced in chapter 5.1.1.

In order to be able to make a more general, quantitative, statement on this, the two key goals of this chapter are to develop observational measures that allow to establish: i) whether the association of small-scale SSTA with mesoscale eddies is systematic, and

ii) whether the mesoscale contributes significantly (to zero-order) to the full observed SST variability.

Such observational estimates are key for evaluating the potential for eddy-atmosphere interactions in midlatitudes, and they could, moreover, provide a tool to test the realism of both, eddy-resolving ocean-only models and coupled models moving towards mesoscale resolution.

Establishing these measures requires an observational proxy for eddying circulations and their intensity. Following the discussion in the previous chapter (section 3.3.1), multi-altimeter SSH can, in general, provide such a proxy. The association of SSH with eddying circulations will be revisited in more detail throughout the following sections, which analyse the observed SST-SSH relationship, both at fixed-location (section 4.2) and following eddies tracks (section 4.3), as a tool to examine the above questions on the nature of the observed SST variability and its association with the oceanic mesoscale.

4.2 Diagnostics at fixed-location

This section provides an analysis of the relationship between SST and SSH from observations at fixed-location. Re-examining the extent datasets allow to do so, sections 4.2.1 and 4.2.2 establish an observational measure for the mesoscale contribution to SST variability. Mechanisms that may set the strength and spatial distribution of the measured contribution are proposed in section 4.2.3.

4.2.1 The observed SST-SSH relationship at fixed location

Figure 4.3 displays snapshots of SSTA (colours), similar to those in Figure 4.1, observed on two different dates over the subtropical North Atlantic and the Pacific-Atlantic Drake passage connection in the Southern Ocean. The SSH field, observed on the same dates, is characterized by high and low anomalies (solid and dashed contours) at typical radius scales of ~ 100 km. They are present throughout the regions and intensify in amplitude (~ 10 cm) towards major current systems and the associated regions of high SSHA rms (cf. Fig. 3.8 and the discussion in the chapter 3.3.1). In this snapshot, eddies in the SSH field are often found to coincide with anomalies in the SST field. This association of warmer SST with highs in SSH, and cooler SST with lows, seems to be especially pronounced for, but not limited to, the strongest eddies, and signs of it can be found throughout the basins. It is however not clear from this

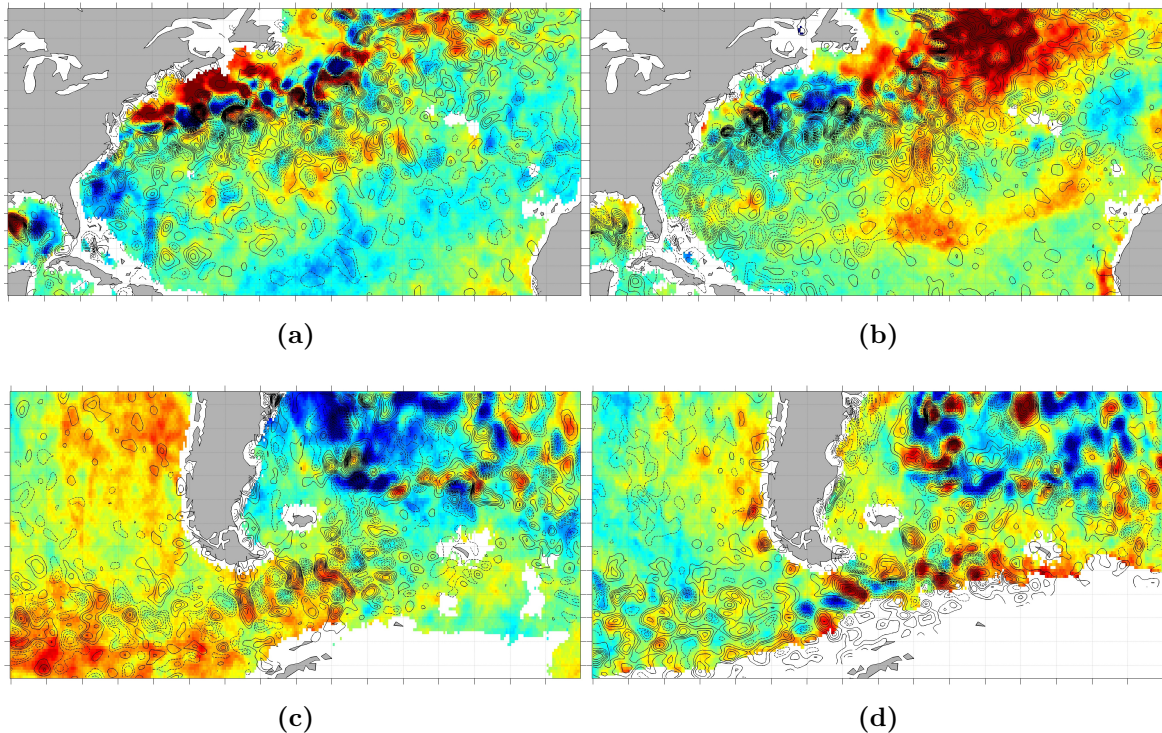


Fig. 4.3: Snapshots of observed SSTA and SSHA, as observed on 15.2.2006 (lhs panels) and 15.8.2006 (rhs panels). SSTA is coloured (on the same ± 2.25 K colour-scale as in Fig. 4.1), highs & lows in SSHA are contoured (every 4 cm, solid & dashed). Top panels show the NA subtropical gyre ($\sim 20^\circ$ - 50° N, 90° - 10° W), bottom panels the SO Drake passage region ($\sim 40^\circ$ - 65° S, 100° - 20° W), with longitude ticked every 5° , latitude every 2° . (For reference, the spatial correlation between SSTA and SSHA in the snapshots displayed here is approximately 0.5, 0.3, 0.4 & 0.6 in panels a, b, c & d, respectively.)

snapshot whether this association of SSH with SST anomalies is systematic. Indeed, just as many observed SSH anomalies can be found devoid of a specific associated (mesoscale) pattern in SSTA, and are instead observed embedded in SST anomalies that extend over much larger spatial scales, comparable to the size of ocean basins.

A way to generalize the information in these snapshots for the whole 5½-year study period is to examine the cross-correlation $\langle T'(t)\eta'(t + \tau) \rangle$ between time series of SSTA and SSHA observed at each fixed-location. Here, and throughout this study for correlation analyses, $\langle \rangle$ denotes the time-average of the enclosed product with each factor normalized by its rms, and, beforehand, the removal of a small remaining temporal mean and a linear trend from anomaly time series. The cross-correlation observed

at zero time-lag ($\tau = 0$), simply denoted by $\langle T'\eta' \rangle$ in the following, is displayed in Figure 4.4 for the entire NA and SO basins. (The lag-aspect of the observed cross-correlation function will be explored in more detail in chapter 6).

If both SSTA and SSHA were uncorrelated Gaussian white noise with daily sampling, the probability to observe a correlation below 0.05 in magnitude would be 95%. Since especially SSH observations do not truly resolve daily time scales and signals are correlated in time, the degrees of freedom are effectively reduced and the probability to observe non-zero correlations in the case of uncorrelated time series rises. The white shaded areas in the figures give an idea (exact only for Gaussian time series) where correlations are not significantly different from zero at the 95% confidence level if only one observation every 15 days of the 2040-day long time series is independent. (Here 15 days is chosen as it corresponds to the mid-latitude e-folding time scale applied by AVISO in the processing of SSH data).

Although maps are a bit noisy, they reveal a systematic pattern. Apart from a few isolated spots, observed zero-lag correlations between SSTA and SSHA are positive everywhere. The two fields are only weakly correlated throughout basin interiors, with correlations rising towards the regions of elevated SSHA rms, where they locally reach values of 0.7 in the GS region and can exceed 0.8 along the ACC.

The observed joint increase of SSTA-SSHA correlations and the intensity of SSH variability suggests that observed correlations can be interpreted as a measure for the strength of the mesoscale eddy contribution to observed SST variability and that the latter peaks in the world's major current systems where observed intense mesoscale eddies are systematically associated with signatures in SST, in the form of warm-top anticyclones and cold-top cyclones similar to those visible in the snapshots in Fig. 4.3. Relatively high correlations are also observed in a few locations characterized by low SSH variability. Do these also reflect an increased mesoscale eddy contribution to the SST field? The following paragraph serves to assess the confidence we can place in this interpretation of the observed SSTA-SSHA correlation field.

As established in chapter 3.3.1, mesoscale eddying circulations dominate the surface geostrophic current variability derived from the observed SSHA field throughout midlatitude basin interiors and ACC regions. However, the SSHA field itself might also contain variability on larger scales, reflecting mostly steric processes. Mixed-layer thermohaline SSHA signals are bound to be highly correlated to observed SSTA, irrespective of the scales on which they occur. Therefore, the observed relationship between SST and SSH fields, uncovered here by examining $\langle T'\eta' \rangle$, can only inform on the contribu-

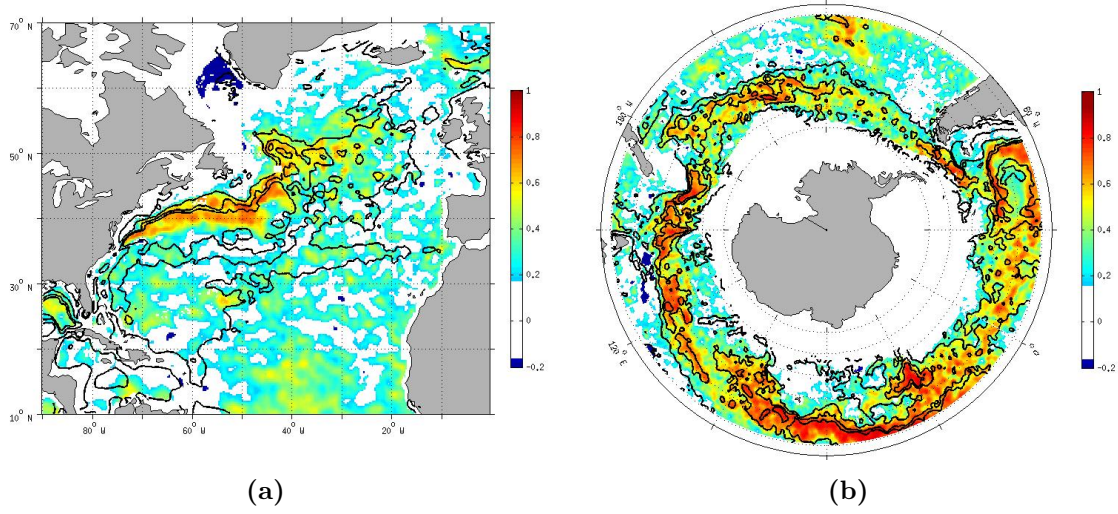


Fig. 4.4: The observed cross-correlation between η' and T' at zero time-lag. The intensity of observed SSH variability is indicated by the 5, 10 and 15 cm levels of $\sigma(\eta')$ in the NA (a), by the 5 and 10 cm levels in the SO (b). In white areas observed cross-correlations are not significantly different from zero at the 95% confidence level, taking 1 observation every 15 days to be independent.

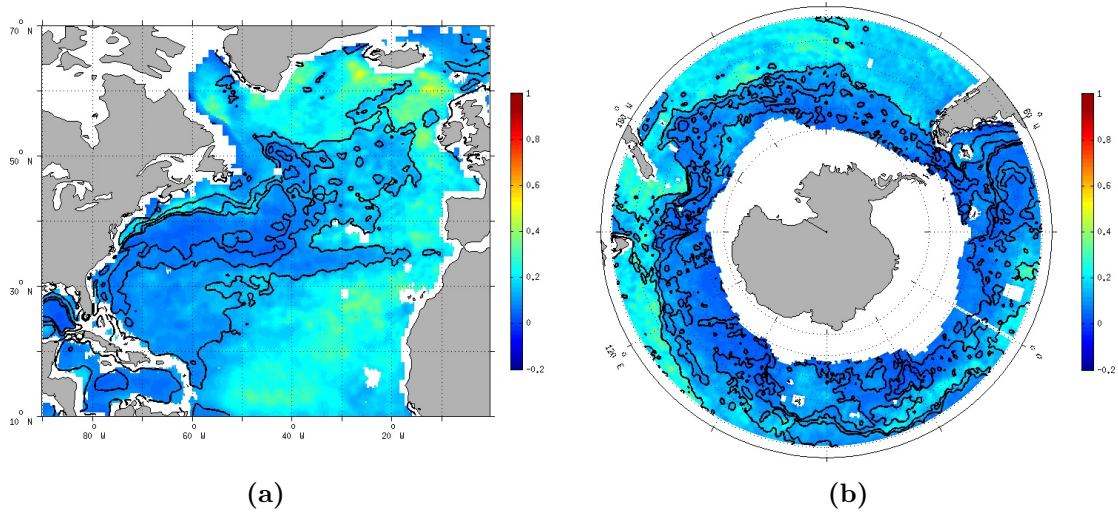


Fig. 4.5: Observed mixed-layer thermosteric contribution to SSH variability $\sigma(\eta'_s)/\sigma(\eta')$ for (a) NA and (b) SO. Since this ratio can be viewed as proxy for $\langle T'\eta' \rangle$ in the case of absolutely no correlation between deep, non-steric SSHA signals and SSTA, it is mapped on the same colour-scale (apart from the white shading) as observed $\langle T'\eta' \rangle$ (Fig. 4.4). The 5, 10 and 15 cm contours of $\sigma(\eta')$ are overlaid.

tion of subsurface mesoscale circulations to the SST field, if mixed-layer thermosteric contributions to the observed SSHA variability are small. If the mixed-layer is vertically homogeneous, mixed-layer thermosteric height anomalies $\eta'_s = (\int_{-h_{ml}}^0 -\rho/\rho_0 dz)'$ can be estimated as $\eta'_s \sim \overline{\alpha_T h_{ml}} T'$ (neglecting terms associated with non-seasonal fluctuations in the thermal expansion coefficient α_T and the mixed-layer depth h_{ml}). Thus, in typical midlatitude conditions ($\overline{h_{ml}} \sim 50$ m, $\overline{\alpha_T} \sim 0.2 \times 10^{-3} \text{K}^{-1}$), a 1 K anomaly in SST is associated with a surface steric height anomaly $\eta'_s \sim 1$ cm. Throughout large parts of western ocean basins, and certainly in major current systems, this is typically an order of magnitude smaller than the full observed SSHA variability, so that here, from this quick scaling, η' anomalies predominately reflect subsurface eddying circulations. This is contrary to seasonal SSH variability that is largely dominated by steric signals associated with the much larger seasonal fluctuations in SST (cf. Appendix A.2).

A more thorough investigation of the mixed-layer thermosteric contribution to SSH variability is given in Figure 4.5, that displays the observed rms ratio $\sigma(\eta'_s)/\sigma(\eta')$ of the two fields, where η'_s have been estimated from MW SSTA combined with a seasonal $\overline{h_{ml}}$ climatology (Appendix B.1) and the climatological-mean surface distribution of α_T (Appendix B.5), thereby neglecting its seasonal variation. The Figure reveals a clear-cut distribution. Typical amplitudes of η'_s are more than an order of magnitude smaller than that of full SSHA, everywhere the observed SSH variability is large, so that here the contribution of mixed-layer thermosteric signals to the full observed SSH variability is negligible. As soon as observed SSHA rms drops below ~ 5 cm, this is no longer the case and mixed-layer thermosteric effects start to represent a significant fraction of the full observed SSH variability ($\geq 20\%$). There are two exceptions to this simple picture. On the one hand, the regions of low SSH variability in the very high-latitude cold-water sphere poleward of the ACC, where, despite the rather deep mixed-layers, a reduction of the thermal expansion coefficient to less than half of its mid-latitude value prevents important mixed-layer steric height signals. On the other hand, the regions of peak SST variability in the vicinity of the GS and the ARC that lead to a significant mixed-layer thermosteric contribution ($\sim 25\%$) despite the relative high levels of observed SSH variability.

In general this indicates that the low levels of SSTA-SSHA zero-lag correlations observed throughout eastern ocean interiors cannot inform on the mesoscale contribution to SST variability due to the relatively weak deep subsurface contribution to the observed SSHA field. But that the high SSTA-SSHA zero-lag correlations observed in

the regions characterized by a predominant deep mesoscale eddy contribution to the SSH field can be interpreted as a firm indicator of a systematic association of mesoscale eddies with signatures in the SST field.

The fact that correlations are independent of amplitude might cast doubt on the above statements. Here we show they are correct through a different way of reasoning. Observed SSH variability can be expressed as the sum, $\eta' = \eta'_o + \eta'_s$, of a subsurface contribution η'_o , reflecting the signature of deep mesoscale eddies, and a surface steric contribution η'_s . We want to test whether mesoscale eddies are systematically associated with SST signals against the null hypothesis (H_0) that they are not. If H_0 is true, observed η'_o would be, apart from through chance, completely uncorrelated to T' , so that, in this case, the expected SSTA-SSHA zero-lag correlation can be expressed as:

$$\langle T' \eta' \rangle |_{H_0} = \frac{\overline{T' \eta'_o} + \overline{T' \eta'_s}}{\sigma(T') \sigma(\eta')} \sim \frac{0 + f \overline{T'^2}}{\sigma(T') \sigma(\eta')} = \frac{f \sigma(T')}{\sigma(\eta')} \sim \frac{\sigma(\eta'_s)}{\sigma(\eta')}, \quad (4.1)$$

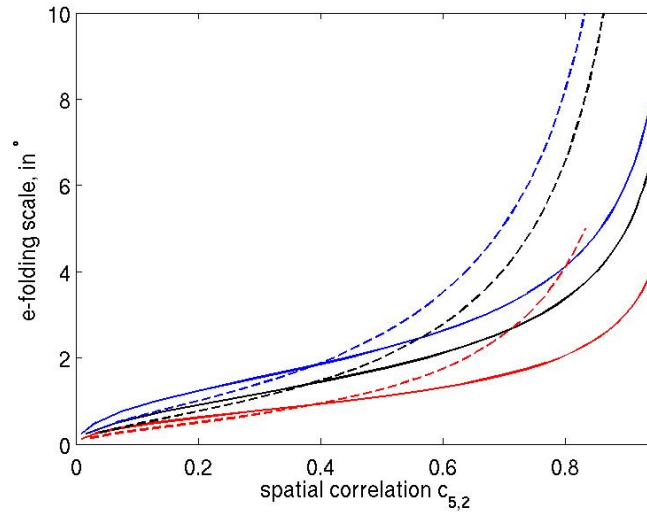
where thermosteric SSHA have been approximated as $\eta'_s \sim f T'$ with $f = f(x, y) = \frac{\overline{\alpha_T h_{ml}}}{\sigma(\eta')}$, thereby neglecting the seasonal variations in f . To this level of approximation, the ratio of mixed-layer thermosteric to full SSHA rms, displayed in Figure 4.5, therefore represents a proxy for $\langle T' \eta' \rangle$ expected if H_0 were true. A comparison to observed $\langle T' \eta' \rangle$, displayed in Figure 4.4 on the same colour-scale, reveals that in regions of weak eddy (SSHA) variability H_0 does a (relatively) good job, whereas in regions of large eddy variability we clearly need to evoke departures from H_0 , namely systematic mesoscale eddy SSTA signatures, in order to explain observations.

4.2.2 Dominant spatial scales of variability

In this section, we aim to assess whether the SSTA signatures, found to be systematically associated with mesoscale SSHA in major current systems, contribute significantly to the (full) SST variability observed in these regions, i.e. to establish an observational measure for question (ii) introduced above (section 4.1).

As synoptic to interannual atmospheric fluctuations predominately force SST variability on spatial scales larger than the oceanic mesoscale, comparable to the atmospheric deformation radius up to oceanic basin scales (chapter 2.2; Frankignoul, 1985), a way to quantify the observed mesoscale contribution to SST variability from observations, is to examine on which spatial scales the latter predominately occurs.

Information on the spatial scales of variability is contained in the auto-correlation matrix of observed SSTA, specifically in its cross-section $\langle T'(x, y) T'(x + \delta x, y + \delta y) \rangle$ at zero time-lag. In order to condense it into a quick, visual measure, the spatial SSTA



(a)

Fig. 4.6: Relationship between physical scales of variability (ordinate, in degree longitude or latitude) and observed values of $c_{5,2}$ (abscissa), as evaluated from theoretical, Gaussian (solid) or exponential (dashed), spatial auto-correlation matrices with prescribed e-folding scales. Both an isotropic ($X=Y$: black) and a zonally elongated case (Y : red, $X = 2 \times Y$: blue) is shown.

auto-correlation matrix, estimated at each grid point (x, y) , is integrated (averaged) over spatial lags up to a specified maximum zonal and meridional lag, $\pm X/2$ and $\pm Y/2$, respectively:

$$c_{X,Y}(x, y) = \frac{1}{\frac{X}{\delta x} \frac{Y}{\delta y}} \sum_{\delta x=-X/2}^{X/2} \sum_{\delta y=-Y/2}^{Y/2} \langle T'(x, y) T'(x + \delta x, y + \delta y) \rangle \quad (4.2)$$

The resulting spatial correlation measure $c_{X,Y}$ can take values in the interval $[-1, 1]$. Values close to 1 indicate that the variability predominately occurs on scales much larger than the size of the box X - Y , small values indicate e-folding scales of variability smaller or comparable to the box-size, and a value of 0 (or less) indicates zero-crossing scales of variability close to (or less than) half of the box-size.

The choice of the box-size X - Y is a trade-off between the spatial scales of variability that can be easily distinguished by $c_{X,Y}$ and the spatial detail retained in the resulting map. Having examined box-sizes from 0.5° to 20° , a box-size of 5° -longitude by 2° -latitude is found to be a good compromise, as it is one of the smallest possible box-sizes that is still larger than the oceanic mesoscale, but conclusions, presented here at the

example of $c_{5^\circ, 2^\circ}$, are insensitive to variations in box size or shape in the vicinity of this choice. To get an idea of the actual physical scales corresponding to this correlation measure, Figure 4.6 displays the values of $c_{5^\circ, 2^\circ}$ obtained, following (4.2), from a range of different theoretical spatial auto-correlation matrices, with prescribed Gaussian or exponential shapes and e-folding scales varying, in an isotropic or a zonally elongated manner, from $\frac{1}{4}^\circ$ to 20° . It shows that values of $c_{5^\circ, 2^\circ}$ close to 1 correspond to e-folding scales larger than 5° for Gaussian and larger than 10° for exponential correlation functions. As $c_{5^\circ, 2^\circ}$ drop below 1, associated scales decrease very quickly at first. This decrease levels off once $c_{5^\circ, 2^\circ}$ reaches values ≤ 0.6 in the case of Gaussian, ≤ 0.5 in the case of exponential correlation functions and associated scales lie below $\sim 2^\circ$. For the correlation functions examined here, typical mesoscales (~ 100 km) correspond to $c_{5^\circ, 2^\circ} \sim 0.3$.

The spatial correlation measure applied to the observed SSTA auto-correlation function, $c_{5^\circ, 2^\circ}(T')$, is displayed in the top panels of Figure 4.7 for every location in NA and SO. High spatial correlation values ($c_{5^\circ, 2^\circ}(T') \geq 0.8$, red shades) are observed to prevail in the tropics and throughout most parts of ocean gyres. They indicate that here the observed SST variability is shaped by coherent SSTA signals on spatial scales comparable to, or larger than, 5° by 2° . On approaching major current systems, observed spatial SSTA correlations drop nearly continuously, until, over the ACC and the western boundary current extensions of GS, BMC, and ARC, observed $c_{5^\circ, 2^\circ}(T')$ reaches values close to ~ 0.3 (blue shades). These low values reveal that here observed SSTA predominately occur on spatial scales much smaller than 5° by 2° , i.e. on scales comparable to the oceanic mesoscale.

In quiescent basin interiors, where large-scale variability in the SSH field is not negligible, the $\langle T'\eta' \rangle$ measure, examined above, was found to carry little information on the eddy SST signal, but was shown to be consistent with an absence of systematic eddy signals in the SST field. The high values of $c_{5^\circ, 2^\circ}(T')$, observed in these regions, confirm a predominance of large-scale SSTA. They are interpreted to reflect a controlling influence of large-scale atmospheric forcing on the observed SST variability away from major current systems.

SSTA spatial correlation scales (Fig. 4.7a, b) moreover allow to distinguish where relatively high observed $\langle T'\eta' \rangle$ (Fig. 4.4) are a measure of systematic eddy SSTA and where they merely reflect an association between large-scale SSTA and mixed-layer thermocline height anomalies. Thus, the weakly enhanced, positive $\langle T'\eta' \rangle$ observed near the subtropical convergence zone in the western NA ($\sim 28^\circ\text{N}$) and over the Azores

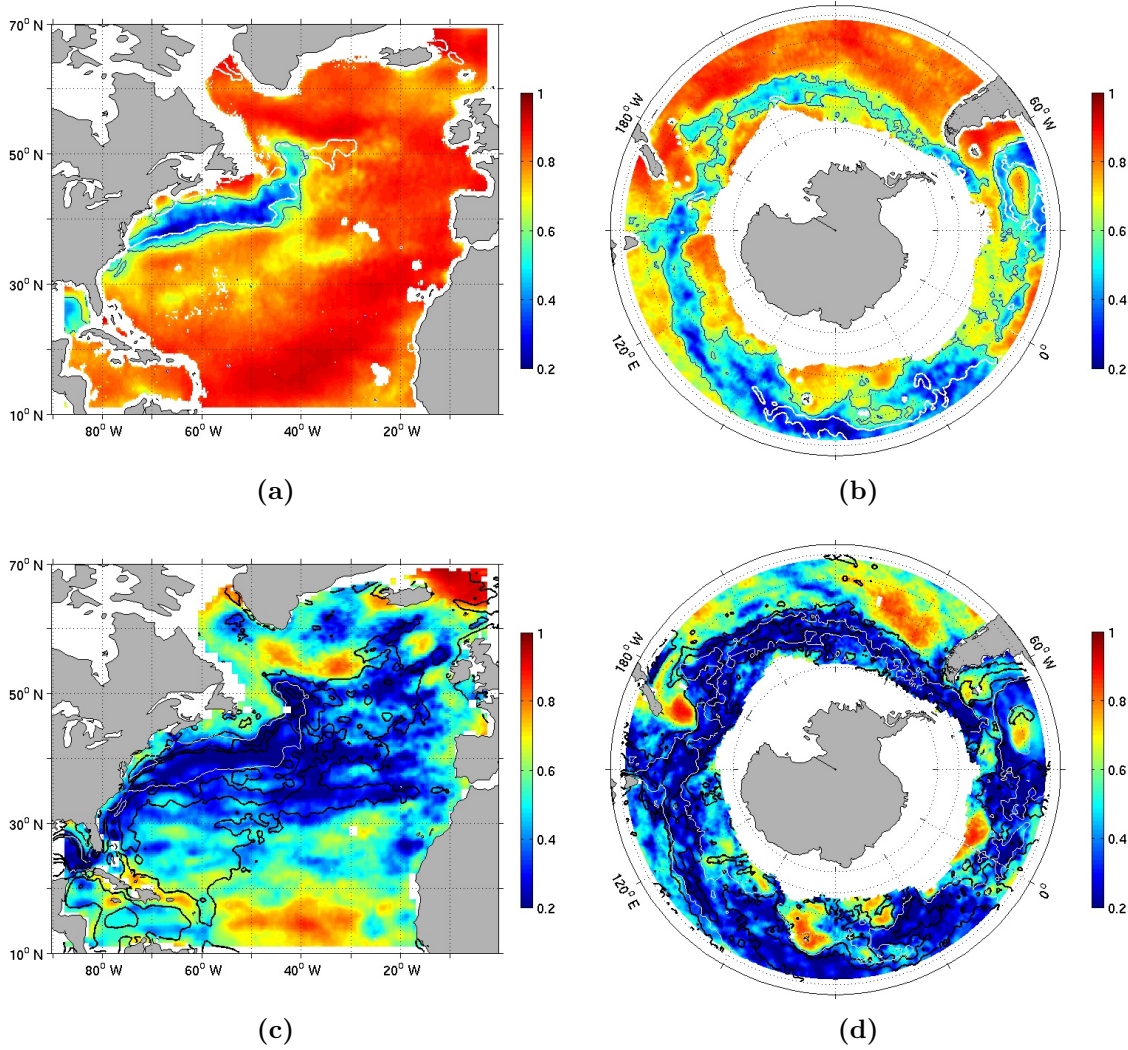


Fig. 4.7: A correlation measure for the characteristic spatial scales of observed SSTA variability, $c_{5^\circ, 2^\circ}(T')$, established following (4.2), is displayed in the top panels, the corresponding measure for SSHA, $c_{5^\circ, 2^\circ}(\eta')$, in the bottom panels (lhs: NA, rhs: SO). An approximate cut-off between regions of predominately large and small scale SSTA variability is indicated by the thin contour of $c_{5^\circ, 2^\circ}(T') = 0.6$ (blue in top, white in bottom panels). Regions of high SST variability are indicated by the white $\sigma(T')=1K$ contour, in the top panels, regions of high SSH variability by the black $\sigma(\eta') = 5$ and 10 cm contours, in the bottom panels (the 15 cm contour is also shown for (c) the NA).

Current in the eastern NA ($\sim 36^\circ\text{N}$) are indeed associated with reduced spatial scales in SSTA, suggestive of an enhanced mesoscale influence on SST variability in these regions. However the elevated $\langle T'\eta' \rangle$ levels observed in NA tropics and the SP subtropical gyre just reflect large-scale co-variability in the two fields.

The combination of high $\langle T'\eta' \rangle$ and low $c_{5^\circ,2^\circ}(T')$ therefore provides a robust observational measure of a dominant role of the oceanic mesoscale in shaping observed SST variability. This combination is observed to be limited to the vicinity of WBCs and the ACC. In these regions the observed SST variability reaches its highest observed amplitudes, with SSTA rms typically exceeding the 1 K level (shown by the white contour in Fig. 4.7a, b). Therefore, whereas large-scale (atmospheric) forcing controls observed SST variability nearly everywhere, the oceanic mesoscale forcing dominates where its amplitudes are largest.

The bottom panels of Figure 4.7 present the same spatial correlation measure, $c_{5^\circ,2^\circ}(\eta')$, for observed SSHA. In contrast to $L_{\eta'}$ (introduced in chapter 3.3.1 and displayed in Fig. 3.11a, b), that reflects the scales of dynamic SSH variability, i.e. of the associated geostrophic flow anomalies, $c_{5^\circ,2^\circ}(\eta')$ does not involve the gradient operator and therefore retains the full spectrum of SSHA variability. The low values of $c_{5^\circ,2^\circ}(\eta')$ observed over vast extents of NA and SO (blue shades in Fig. 4.7c, d) reveal that here also the full SSHA variability, not only $L_{\eta'}$, occurs predominately on the mesoscale. In the SO, the few regions of large-scale SSHA (red shades in Fig. 4.7c, d) correspond well to the regions, locked mostly to bathymetric features as discussed above, in which also the corresponding flow occurs on large spatial scales. However in the eastern NA, even though $L_{\eta'}$ indicates mesoscales, $c_{5^\circ,2^\circ}(\eta')$ takes larger values that indicate an important large-scale, dynamically passive contribution to the observed SSHA variability. This coincides with non-negligible mixed-layer thermosteric height signals (cf. Fig. 4.5) and both are limited to regions of low SSHA variability, at rms amplitudes below ~ 5 cm.

A comparison between top and bottom panels of Figure 4.7 reveals a clear discrepancy between the characteristic spatial scales of SSTA and SSHA. Over the bulk of the regions with a clear presence of mesoscale eddy signatures in SSHA (blue shades, in c, d), characteristic SSTA scales are much larger than “mesoscale” (red shades, in a, b). It is only in a tiny window of the vast regions of mesoscale SSHA, in which SSTA and SSHA scales become comparable and eddies make a significant contribution to the observed SST variability, and, thus, to the variability of the atmosphere’s lower thermal boundary. The mechanisms that set the location and strength of this mesoscale

window “to the atmosphere” and generate the striking discrepancy between large-scale SSTA and “mesoscale” SSHa, observed elsewhere, are examined in the next section.

4.2.3 Mechanisms

The observed discrepancy between the predominant scales of SSHa and SSTA shows that the geographic distribution of mesoscale eddies, alone, which are observed to predominate the SSHa field nearly everywhere, cannot explain the localization of the observed strong mesoscale contribution to SST variability.

The next most obvious assumption is a relation of the mesoscale SSTA signal to the amplitude, in η' , of observed mesoscale eddies. For example, from thermal wind arguments we expect a simple relationship between eddy η' and interior eddy T' amplitudes, if eddies are baroclinic and the observed density-stratification is mainly temperature-controlled. Thus, if eddy-induced pressure perturbations are compensated at the depth H_{comp} , the thermal wind relation reads $f \frac{\delta U_g}{\delta z} \sim f \frac{U_g^{surf}}{H_{comp}} \sim \frac{g \delta_y \rho}{\rho_0} \sim \frac{g \alpha_T T'}{L}$ and substituting U_g^{surf} with its value $\frac{g \eta'}{f L}$ yields $T' \sim \frac{\eta'}{\alpha_T H_{comp}}$, or an interior eddy-induced temperature anomaly of ~ 0.5 K for every 10 cm perturbation in SSH, if compensated over the depth of the main thermocline (~ 1 km).

In the surface mixed-layer, where other processes force, primarily large-scale, variability in SST, the mesoscale contribution to SST variability will then be substantial if the amplitudes of eddy-induced SSTA stand out from or are at least comparable to that of the former. If atmospheric forcing generates a continuum of large-scale SST variability with little variation in amplitude across an oceanic basin, the spatial distribution of the mesoscale contribution to SST variability should then be simply related to that of the eddy amplitude.

Within the vast regions in which low values of $c_{5^\circ, 2^\circ}(\eta')$ indicate the predominance of mesoscale eddies in the SSHa field (blue shades in Fig. 4.7c, d), typical eddy amplitudes, as measured by the observed SSHa rms (black contours), indeed show a large spread. This is also the case for the mesoscale signal in SST variability. Hand in hand with increasing SSHa rms across these regions, the mesoscale signal in SSTA is observed to intensify, as indicated by the increase of $\langle T' \eta' \rangle$ (Fig. 4.4) together with the reduction in the characteristic spatial scales of SST variability (Fig. 4.7a, b), suggesting that the intensity of mesoscale eddies provides a good zero-order explanation, following the above reasoning, for the distribution of the eddy contribution to SST variability. This is illustrated in Figure 4.8 that displays observed SSHa and SSTA along longitude-time cross-sections through the GS after its separation from the coast

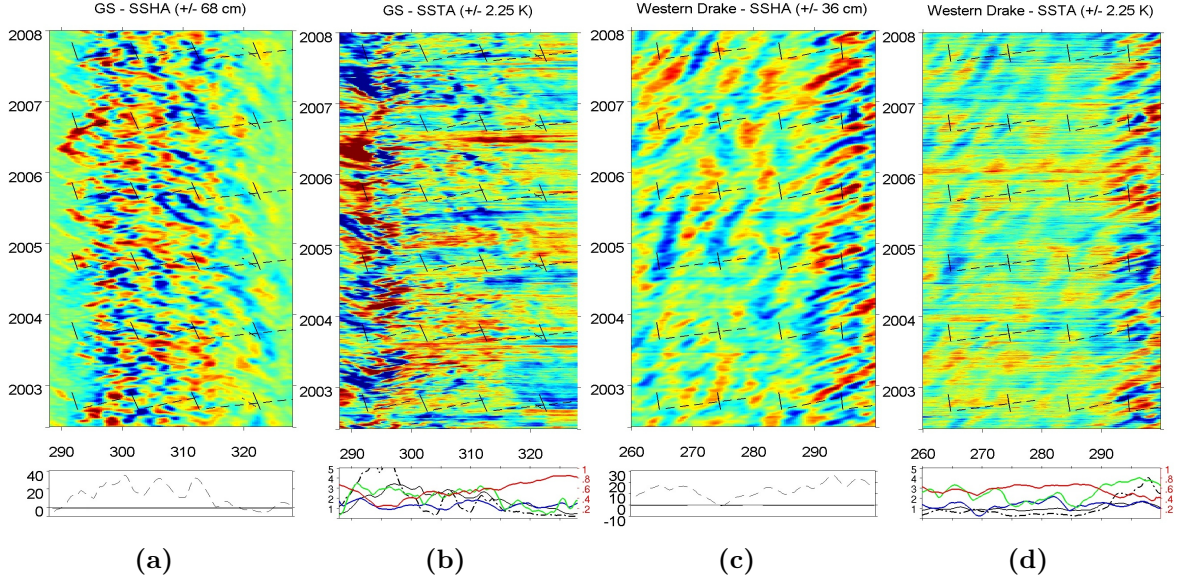


Fig. 4.8: Time-longitude maps of η' and T' for cross-sections through the separated GS at 39°N (a & b) and the entrance to Drake passage at 58°S (c & d). The location of these sections is mapped in Figure 3.12. The linear colour-scale, with cold/warm colours for negative/positive anomalies, varies between the panels, as indicated at the top, to adapt to the varying levels of variability. Solid lines propagate with the time-mean zonal long baroclinic Rossby wave phase-speed c_{lrw} observed at their origin, dashed lines with that of the zonal time-mean flow as estimated from $\bar{\eta}$. Cross-sections of both are shown in the bottom panels of a & c. Bottom panels of b & d display, on the lhs y-axis, cross-sections of $\sigma(\eta')$ (thin black, in 10 cm) and $\sigma(\mathbf{u}'_g \cdot \nabla \bar{T})$ (dash-dot, in $\text{K}/_{10 \text{ days}}$), and on the rhs y-axis, that of $\langle T'\eta' \rangle$ (green), $c_{5^\circ, 2^\circ}(T')$ (red) and $c_{5^\circ, 2^\circ}(\eta')$ (blue).

and the northern branches of the ACC as it enters Drake passage.

SSHA plots clearly show intense mesoscale signals in a competition between westward propagation at a speed comparable to $c_{lrw} = -\beta R_d^2$, the zonal phase-speed of long first-mode baroclinic Rossby waves (indicated by the solid lines), and eastward advection by the strong zonal mean-flow (dashed lines). Corresponding SSTA plots show associated propagating warm and cold mesoscale anomalies, as well as both higher- and lower frequency zonally banded anomalies, indicative of observed large-scale SST variability at these locations. Along the Drake-passage section, the observed mesoscale control on the SSTA field clearly becomes more pronounced towards the East as eddy SSHA amplitudes intensify upon the entrance to the passage, consistent with

the above hypothesis. Along the GS section, however, the strongest eddies that are observed within the core of the separated current (between 295° and 315°E) are not associated with the highest amplitude small-scale SSTA and large-scale variability in the SSTA field is observed to be at least as pronounced as that associated with the mesoscale. Instead highest amplitude mesoscale SSTA and a firm mesoscale control on the SST variability is observed slightly to the West, where eddies are present but less intense.

Re-examining the observed spatial distribution of a mesoscale SSTA signal, in more detail, reveals that in the GS region the highest levels of observed $\langle T'\eta' \rangle$ (red shades in Fig. 4.4 a) are indeed not centred over the region of the most intense SSHA rms (black contours in Fig. 4.4 a), but situated on its North-Western rim. This is also the case for observed $c_{5^\circ, 2^\circ}(T')$ (Fig. 4.7a). Although less pronounced, such a shift between SSHA rms and the mesoscale contribution to SSTA is also observed over the ARC. Over the ACC itself, the regions of a firm mesoscale signature in SST variability coincide generally rather well with the peaks in SSHA rms. Although it does a good job over the ACC, the intensity of mesoscale SSH variability alone is therefore, especially over western-boundary currents and their extensions, not sufficient to explain the observed spatial distribution of the mesoscale contribution to SST variability.

To refine the picture, we need to evoke, not simply the eddy strength, but the eddy efficiency in forcing SST anomalies. In the following, we show that considering only one of the mechanisms, through which, following (2.25), eddies can in principal force SST anomalies (cf. chapter 2.2.3), yields a satisfactory explanation for the location of the strongest observed mesoscale contribution to SST variability.

The mechanism in question is the stirring by geostrophic eddying circulations of the climatological isotherms in the surface mixed-layer. Given by $\mathbf{u}'_g \cdot \nabla \bar{T}$, it can be directly estimated from the SST and SSH data and its rms amplitude, observed over the study-period, is mapped, for NA and SO, in Figure 4.9. The Figure shows that the observed amplitudes of eddy surface isotherm stirring vary strongly across an oceanic basin, by nearly two orders of magnitude around a typical value of 0.5 K/10 days. The latter can be achieved, for example, by an eddying flow of 10 cm/s stirring a background SST gradient of 0.5 K/100 km. Observed amplitudes are weakest, below 0.1 K/10 days, in quiescent, especially Eastern, basin interiors, and increase to well above 1 K/10 days in some areas associated with the major current systems. Thus, especially over the ACC, the broad features of the spatial distribution of eddy stirring resemble closely to those of $\sigma(\eta')$. However over the GS system, the intensity of the eddy isotherm

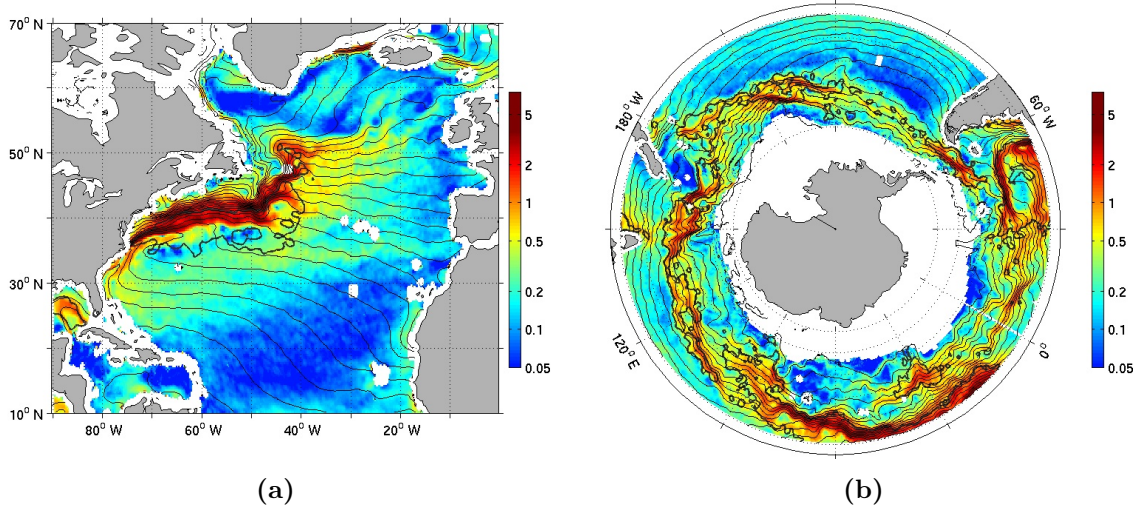


Fig. 4.9: The intensity of eddy stirring of surface isotherms, measured by the rms of anomalous geostrophic background SST advection $\sigma(\mathbf{u}'_g \cdot \nabla \bar{T})$, is mapped on a logarithmic colour-scale, in $\text{K}/10 \text{ days}$. Background surface isotherms are overlaid, with a contour interval of 1°C . The thick black contour delimits regions of intense SSH variability, in which $\sigma(\eta')$ exceeds 15 cm in the NA (a), 10 cm in the SO (b).

stirring is characterized by the same spatial shift to the North-West with respect to the peak SSHA rms located over the GS current core (cf. Fig. 3.10), as is the strongest mesoscale contribution to SST variability, revealed by the highest $\langle T'\eta' \rangle$ (Fig. 4.4 a) and the smallest $c_{5^\circ, 2^\circ}(T')$ (Fig. 4.7 a). This joint observed spatial distribution is consistent with surface eddy stirring of mixed-layer isotherms being a predominant mechanism through which the mesoscale generates SST variability.

Apart from the eddy strength measured by $\sigma(\eta')$, the spatial distribution of the eddy isotherm stirring is set by $\nabla \bar{T}$, the strength and position of climatological SST gradients¹. The fact that SST gradients and $\sigma(\eta')$ are closely tied to each other, and therefore show the same broad increase from eastern gyre interiors towards major current systems, explains that the amplitude variations of observed eddy stirring are an order of magnitude larger than that of the eddy strength itself. Along the ACC, the details of the two fields' spatial distribution match well, too, as here fronts in SST closely coincide with the cores of observed ACC jets (e.g. Hughes and Ash, 2001). This

¹ The eddy isotherm stirring scales as $\eta' \times \frac{g}{fL_{\eta'}} \times \nabla \bar{T}$, but $fL_{\eta'}$, the eddy length scales after accounting for their basic meridional dependence, are observed to vary little across (deep) basin interiors (cf. Fig. 3.11).

is not the case over intense WBCs. Supported by meridional continental side walls, these distort the mainly zonal SST distribution, expected in their absence, by advecting heat, poleward in the case of the GS and the Agulhas Current and its retroflected branch as it joins the ACC. They thereby create a warm surface tongue over their cores such that isotherms are pushed aside where they bunch together to intense SST fronts, which are especially intense on the poleward (shoreward) side of jet cores, and are well-known in the case of the GS as its North wall. The GS south wall is much less pronounced in SST. Over the GS, the mesoscale SST contribution through eddy surface isotherm stirring therefore peaks over the North wall, and not over the current core where eddying circulations are most intense, but cannot generated variability in SST through mixed-layer isotherm stirring due to the near-zero gradient in SST. Although less pronounced, observed SST fronts in regions of weaker eddy variability, such as the NA subtropical convergence zone, are also associated with enhanced eddy stirring and reduced spatial scales of SST variability.

To test (a bit more quantitatively) how well the intensity of eddy mixed-layer isotherm stirring explains the observed distribution of the eddy SST impact, Figure 4.10 displays $\sigma(\mathbf{u}'_g \cdot \nabla \overline{T})$, mapped, separately for NA and SO, not in physical space, but as function of the two variables that jointly characterize the strength of the mesoscale contribution to SST variability, $c_{5^\circ, 2^\circ}(T')$ (x-axis) and $\langle T'\eta' \rangle$ (y-axis). The Figure shows that increases in the intensity of eddy isotherm stirring are indeed quite systematically associated with decreasing characteristic spatial scales of SSTA and increasing SSTA-SSHA zero-lag correlations, reflecting enhanced mesoscale control on SST variability. For low values in eddy stirring, a rise in eddy stirring is mainly reflected by a decrease in the spatial scales of SSTA. In this regime of relative large spatial SSTA scales, $\langle T'\eta' \rangle$ is no good indicator for eddy contribution to SST variability, as due to steric effects high $\langle T'\eta' \rangle$ can also represent correlated signals on large spatial scales, an effect that is especially pronounced in the low-latitude NA and in the region of low SSH variability situated over the bathymetric ridge in the mid-SP. Once eddy stirring exceeds a value of $\sim 0.5 \text{ K/}_{10 \text{ days}}$, a further rise in the stirring intensity is now also systematically associated with increasing values of $\langle T'\eta' \rangle$. Interestingly, in this regime, we observe a nearly linear increase of the minimum observed $\langle T'\eta' \rangle$ with decreasing spatial scales (white areas in the figure indicate regions of the $c_{5^\circ, 2^\circ}(T') - \langle T'\eta' \rangle$ parameter space that are not realized in the observations), i.e. SSTA on small spatial scales are always highly correlated to η' . For the highest observed values of eddy stirring (exceeding $\sim 2 \text{ K/}_{10 \text{ days}}$), both criteria for mesoscale control on

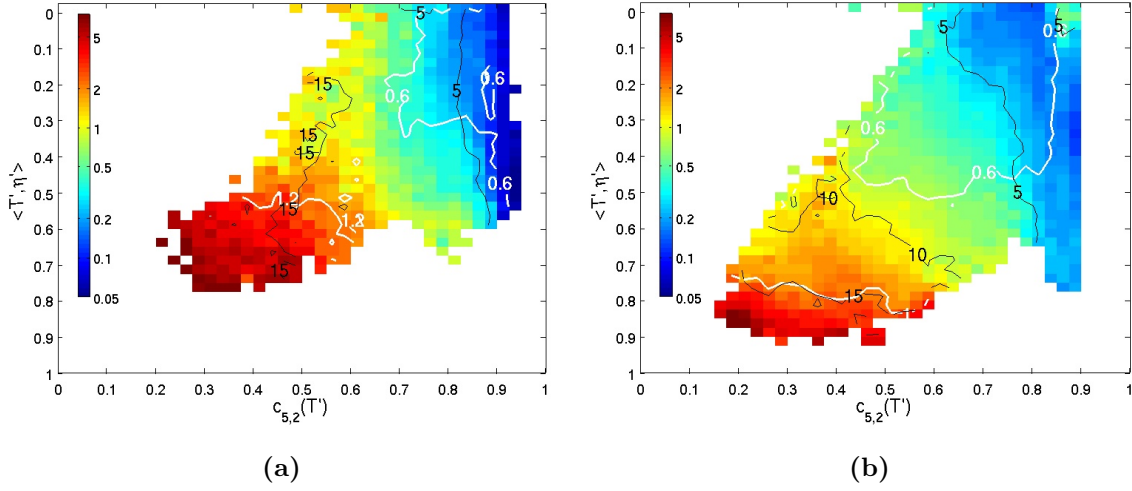


Fig. 4.10: The intensity of eddy stirring of surface isotherms $\sigma(\mathbf{u}'_g \cdot \nabla \overline{T})$, in K/10 days, on the same logarithmic colour-scale as in Fig. 4.9, mapped however not in geographic coordinates but as function of $c_{5^\circ, 2^\circ}(T')$, the SSTA spatial correlation, and of $\langle T'\eta' \rangle$, the SSTA-SSHA cross-correlation at zero time-lag. For a) the NA and b) the SO. Contours of $\sigma(T')$ (in K, white) and $\sigma(\eta')$ (in cm, black) indicate the intensity of the observed SST and SSH variability, mapped in the same space.

SST variability are always fulfilled.

Whereas in the SO $\sigma(\eta')$, contoured in black, and $\sigma(\mathbf{u}'_g \cdot \nabla \overline{T})$ have nearly identical distributions in the $c_{5^\circ, 2^\circ}(T') - \langle T'\eta' \rangle$ space and are therefore as good indicators of an eddy signal in SST variability, in the NA eddy stirring is a much better indicator. Here, due to the observed reversal in the sign of $\partial_y \overline{T}$ over the GS core, highest $\sigma(\eta')$ (≥ 15 cm) can be associated with $\langle T'\eta' \rangle$ as low as 0.2 and to $c_{5^\circ, 2^\circ}(T')$ higher than 0.5.

In both NA and SO, the rms amplitude of observed SST variability, contoured in white, is observed to jointly increase with the intensity of eddy stirring and peaks once observed SST variability is firmly controlled by the mesoscale.

4.2.4 Summary

The analysis of SSTA and SSHA observations at fixed-location indicates that *only* in narrow regions in the vicinity of the world's major current systems, in which elevated eddy variability coincides with strong gradients in the time-mean SST field, mesoscale eddies are i) systematically associated with signatures in SSTA, in the form of warm-top anticyclones and cold-top cyclones, and ii) thereby make a zero-order contribution

to the observed SST variability. *Elsewhere*, likely atmospherically forced, large spatial scales are observed to dominate the variability in SST. The regions of intense eddy stirring can therefore be seen as a window of communication between the oceanic mesoscale and the atmosphere (or at least its surface boundary layer).

The observational measure for the mesoscale contribution to SST variability, developed here, partly relies on the assumption of a separation between the spatial scales of atmospherically and ocean mesoscale eddy forced SST variability. Although, compared to oceanic mesoscales, the scales of synoptic atmospheric disturbances are indeed typically an order of magnitude larger, one of the mechanisms through which they force SSTA, advection by anomalous Ekman currents, directly depends on the oceanic SST field, and can therefore in principal also generate SST variability on the smaller spatial scales of the latter, in particular in regions of strong fronts or sign reversals in the SST gradient. However in combination with the high-levels of $\langle T'\eta' \rangle$, observed in regions in which η' has been established to principally reflect subsurface mesoscale signals, the observed predominance of small spatial scales of SST variability unequivocally points to its firm control by the oceanic mesoscale.

The variance of atmospherically forced SST variability, also for other reasons than its dependence on the oceanic frontal structure, varies across oceanic basins. To zero-order, it is however not necessary to evoke such variations, as, in order to explain the observed time-mean distribution of the mesoscale contribution to SST variability, it suffices to account for only one of the mesoscale T' forcing terms, the stirring of mixed-layer isotherms by geostrophic eddying circulations. This points towards the importance of this mechanism in generating the observed eddy signal in SST variability. The importance of horizontal stirring ($-\mathbf{u}' \cdot \nabla \bar{T}$) is likely a peculiarity of the (destratified) surface mixed-layer, as in the stratified subsurface ocean below the mixed-layer vertical eddy stirring of the background stratification ($-w' \partial_z \bar{T}$) is expected to play an at least as important role in the generation of mesoscale temperature anomalies.

Observations show mesoscale circulations throughout oceanic basins, also away from major SST fronts (cf. Fig. 4.7c, d or Fig. 3.11). Although these are not observed to leave a dominant contribution to SST variability, are they indeed not associated with systematic SST signals? Figure 4.11 presents a longitude-time plot of η' through the quiescent eastern subtropical gyres of NA (a) and SP (c). It clearly shows the predominance of propagating, Rossby-wave like features in the *SSHA* field. The corresponding SSTA fields, displayed in panels c and d, show a zonally banded structure that illustrates the important contribution of large-scale processes to the SST variabil-

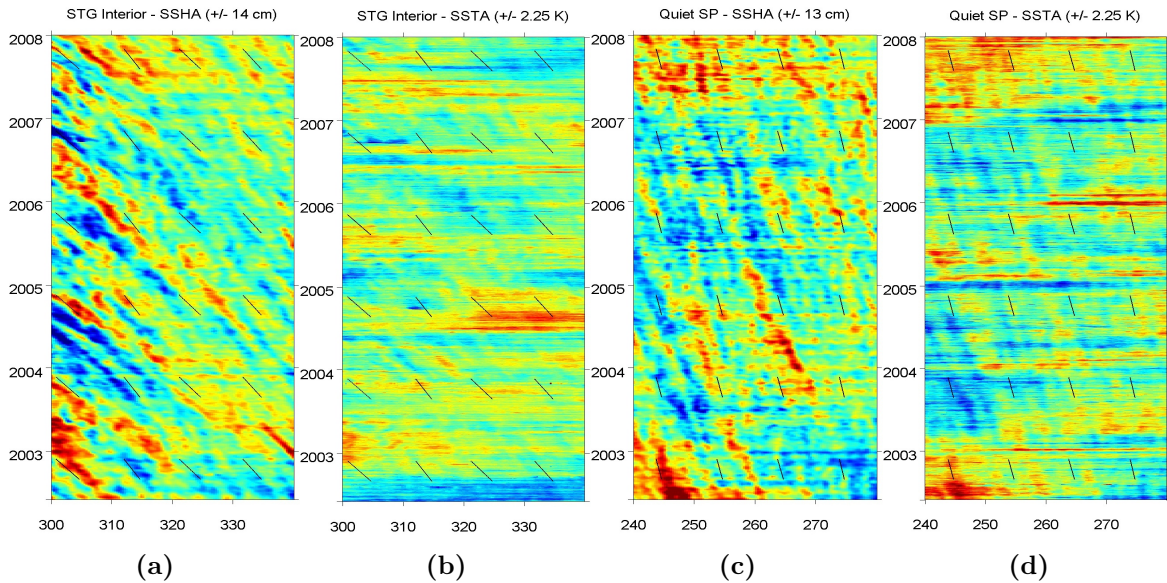


Fig. 4.11: Time-longitude maps of η' and T' in the eastern subtropical interiors of the NA (a & b) and the SP (c & d).

ity in these regions. Upon closer inspection, however, weaker amplitude, propagating signals reminiscent of those in the SSHA field become discernible in the SSTA field, too. This suggests the weak mesoscale circulations in basin interiors are also associated with systematic signatures in the SST field. Although they are not of predominant importance for the full SST variability observed in these regions, they merit further investigation. This requires some kind of filtering, to separate out the dominant large-scale variability. In the next section (4.3), we will do so by using observed eddy tracks as a natural filter.

4.3 Diagnostics following eddy tracks

This section provides an analysis of the relationship between SST and SSH following the tracks of propagating eddies, as observed in the CSS11 dataset described in chapter 3.2.4. Section 4.3.1 establishes the mesoscale SST signal following the eddy tracks and thereby complements the analysis at fixed-location in addressing the key questions of this chapter outlined in section 4.1. Section 4.3.2 provides a synthesis of the observed SST-SSH relationship with fixed-location results and assesses possible mechanisms.

Motivated by the finding of the fixed-location analysis, presented above, of a very spatially inhomogeneous nature of the observed SST variability and of the oceanic

mesoscale contribution to it, the analysis will focus on propagating eddies with origins in two contrasting oceanic regimes: the regions of intense observed eddy variability associated with GS and ACC (“energetic regions”), for which fixed-location results revealed an important mesoscale contribution to the observed SST variability, contrasted to the low-variability eastern subtropics of NA and SP (“quiescent regions”), which have been characterized by the predominance of large-scale SST variability. Eddy tracks observed to originate within these two regimes have been mapped (Fig. 3.12) and introduced in chapter 3.3.3. The regimes’ definitions, given in chapter 3.3.2, independently of SST observations, allow further investigation of the nature of their SST variability, as presented below.

4.3.1 Observed SSTA following eddy tracks

Following propagating eddies in the SSHA field provides a different perspective on SST variability, compared to observing from a fixed-location. Specifically, it allows direct examination of the signatures propagating eddies leave in the observed SST field. To do so, MW observations of SSTA are evaluated within the cores of propagating eddies, following the eddy position in time and space. Thus, an eddy core-average T' is obtained for each weekly eddy snapshot, as average of the daily, $\frac{1}{4}^\circ$ -by- $\frac{1}{4}^\circ$ MW-SSTA, over the 7-day period centred on the eddy position in time, and over all grid points whose centres lie within the instantaneous eddy core². Eddy cores are observed to extend across roughly 100 km, and therefore contain on average about twenty $\frac{1}{4}^\circ$ -by- $\frac{1}{4}^\circ$ grid boxes (25 in the GS, 22 in the ACC, 20 in the quiet NA, 15 in the SP). Each snapshot’s eddy-core T' is thus an average over typically $7 \times 20 = 140$ gridded SSTA data.

Eddy-core T' are estimated weekly along all eddy tracks with origins in the energetic and quiet regimes of NA and SO, i.e. all tracks displayed in Fig. 3.12. The resulting probability distributions of eddy-core T' , observed over energetic and quiet regions’ cyclones and anticyclones, are displayed in Figure 4.12. These pdfs are a powerful diagnostic. As outlined in following two paragraphs, they provide the answers to questions i) and ii), posed above (in section 4.1).

² Here eddy cores are defined to be delimited by the instantaneous eddy e-folding radius $r_{1/e}$. The latter defines a circle that encloses the same area as the SSHA contour along which the eddy amplitude e-folds, i.e. along which SSHA differs by A_e/e from background SSHA (cf. CSS11, their appendix B.3). At an rms between individual eddy snapshots of nearly 20 km, $r_{1/e}$ is observed to average to ~ 65 km over NA, ~ 60 km over ACC and ~ 50 km over subtropical SP eddies, and is therefore systematically

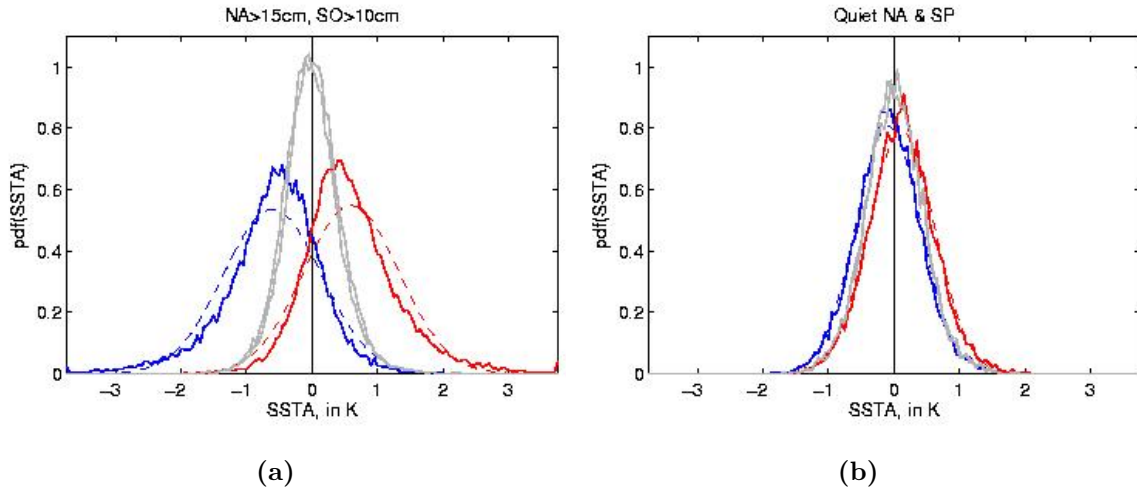


Fig. 4.12: Probability distribution of eddy-core SSTA observed over individual, weekly snapshots of propagating anticyclones (red) & cyclones (blue) with origins in (a) energetic and (b) quiet regimes of NA and SO. This results in 25680 (27060) snapshots of anticyclones (cyclones) in the energetic regime, 17589 (16840) in the quiet regime. Thin dashed curves show the corresponding normal distributions (mean and standard deviation equal to observations). The probability distribution of large-scale (5° -by- 5°) SSTA, evaluated at the eddy positions, is displayed in grey (separately for cyclones and anticyclones).

i) Systematic eddy SSTA? The first visual impression from Figure 4.12, is that the distributions of eddy core-average T' , observed over anticyclonic (red) and cyclonic (blue) eddies with origins in the energetic regime (panel a), are clearly distinct, shifted respectively towards warm and cold SSTA, with all-eddy average $\langle T' \rangle$ of $+0.6$ K for anticyclones, -0.6 K for cyclones³. In contrast, over eddies with origins in the quiescent subtropics, cyclonic and anticyclonic eddy-core T' distributions nearly coincide. This first impression is consistent with the fixed-location analysis, and suggests that the association between warm-top anticyclones and cold-top cyclones is systematic only in energetic regions.

However, formally, all-snapshot averages of eddy-core T' are non-zero, also over eddies in the quiet regime. Here $\langle T' \rangle$ amount to 0.12 K over anticyclones, and, at -

smaller than L_e (roughly $\frac{2}{3}$).

³ Here and in the following, $\langle \rangle$ denotes the average over all individual, weekly snapshots of eddies with origins in a specified bin, estimated separately for eddies of each polarity, i.e. anticyclones and cyclones.

0.09 K, to slightly less over cyclones. As shown by the observed T' pdf, these averages are the result of the cancellation between large negative and positive SST anomalies observed over individual eddy snapshots, whose rms amplitude by far exceeds the all-eddy average T' (by a factor of ~ 5). In the following, we test whether the quiet regime's eddy-core $\langle T' \rangle$ might nevertheless reflect a possibly systematic eddy signature in SSTA in ocean basin interiors.

To test whether, in the light of the large observed T' fluctuations between eddy snapshots, the observed $\langle T' \rangle$ are just averaging artefacts, i.e. whether they are non-zero just by chance, a t-test is performed. (As shown by Fig. 4.12b, the condition of normality is well satisfied by observed eddy-core T' in quiet regions). It attests $\langle T' \rangle$ to be significantly different from zero, positive over anticyclones, negative over cyclones, at the 99% confidence level. Even after reducing the number of degrees of freedom from the number of individual weekly eddy snapshots to the number of eddy tracks, i.e. by more than an order of magnitude, to account for correlations, the observed quiet regions eddy-core $\langle T' \rangle$ passes the test of statistical significance (at the same confidence level).

Although the quiet regime's average eddy-core SSTA lies well below the measurement accuracy of *individual* AMSR-E SST measurements, estimated at approximately 0.4 K by Chelton and Wentz (2005), the averaging over the ~ 17000 individual snapshots (for each eddy polarity) reduces the measurement-induced error on $\langle T' \rangle$ by approximately two orders of magnitude, to ~ 3 mK, which corresponds to only about 3% of the eddy signal.

To test whether observed $\langle T' \rangle$ is indeed a mesoscale signal and cannot be explained by residuals of non-mesoscale sources of SST variability (e.g. atmospheric forcing), the track-following analysis is re-performed with a large-scale SSTA dataset, T'_{LS} , obtained by bin-averaging MW-SSTA over a predefined 5° -by- 5° grid, as explained in chapter 3.1.4. The average of large-scale SSTA along the eddies' paths, $\langle T'_{LS} \rangle = 0.01$ K, is observed to be an order of magnitude smaller than the full eddy-core T' amplitude, $\langle T' \rangle = 0.11$ K. The number of the quiet regime's eddies is therefore large enough to effectively average out large-scale SSTA variability. This also holds when considering NA and SP quiet regions' eddies separately (not shown). Observed $\langle T' \rangle$ can thus safely be interpreted as the robust mesoscale SSTA signature associated with propagating eddies in quiescent subtropical basin interiors.

The above tests evidently also hold for the observed, much larger amplitude, energetic regions' eddy T' signal. The track-following analysis of SST variability therefore

Tab. 4.1: Mesoscale versus large-scale control on observed SST variability.

	following eddies' tracks				at fixed-location
	$\langle T' \rangle$	$\pm \langle T'_{LS} \rangle$	$\sigma(T')$	$\sigma(T'_{LS})$	$\sigma(T')$
Energetic regions	0.60	± 0.02	0.74	0.44	0.90
Quiescent regions	0.11	± 0.01	0.51	0.45	0.56

Mean $\langle \rangle$ and standard-deviation $\sigma(\cdot)$, over all snapshots of eddies with origins in the energetic and quiet regimes of NA and SO, of eddy-core SSTA (T') and large-scale SSTA evaluated at the eddy position (T'_{LS}) - contrasted to observed fixed-location SSTA rms averaged over the regime. All in K. Observations over cyclonic eddies are multiplied by -1 before averaging.

shows that propagating eddies are, on average, associated with systematic signatures in the SST field, in the form of warm-top anticyclones and cold-top cyclones, in both energetic and quiescent regimes of oceanic basins. They are observed to be statistically significant, unaffected by measurement errors and cannot be explained by residuals of large-scale SST variability along the eddy paths.

The following paragraph assesses how this result can be combined with the seemingly contradictory result, obtained by the analysis at fixed-location, of a predominant large-scale control on SST variability in quiet regions.

ii) Significant mesoscale contribution to observed SST variability? After having established the statistically robust average eddy T' signature in the previous paragraph, here we focus on the spread of the observed eddy-core T' distribution around this robust average. To examine the processes at its origin, further analysis of the large-scale (5° -by- 5°) SSTA following eddy tracks is provided.

The observed probability distribution of T'_{LS} , evaluated over individual eddy snapshots, is displayed by the grey curves in Fig. 4.12. The distribution of T'_{LS} , and therefore the observed large-scale SSTA variability along eddy tracks, shows no difference between anticyclonic and cyclonic eddy snapshots (displayed by the two, virtually identical, grey curves in each panel of the Figure), nor between quiescent and energetic regimes. Over eddies in both regimes, T'_{LS} distributions are observed to be Gaussian to a very good approximation, have averages that are statistically indistinguishable from zero (as established above), and have similar typical rms spreads of ~ 0.5 K.

Over eddies with origins in the energetic regime (Fig. 4.12a), the observed eddy-core T' distributions (red for anticyclones, blue for cyclones) clearly differ from that of

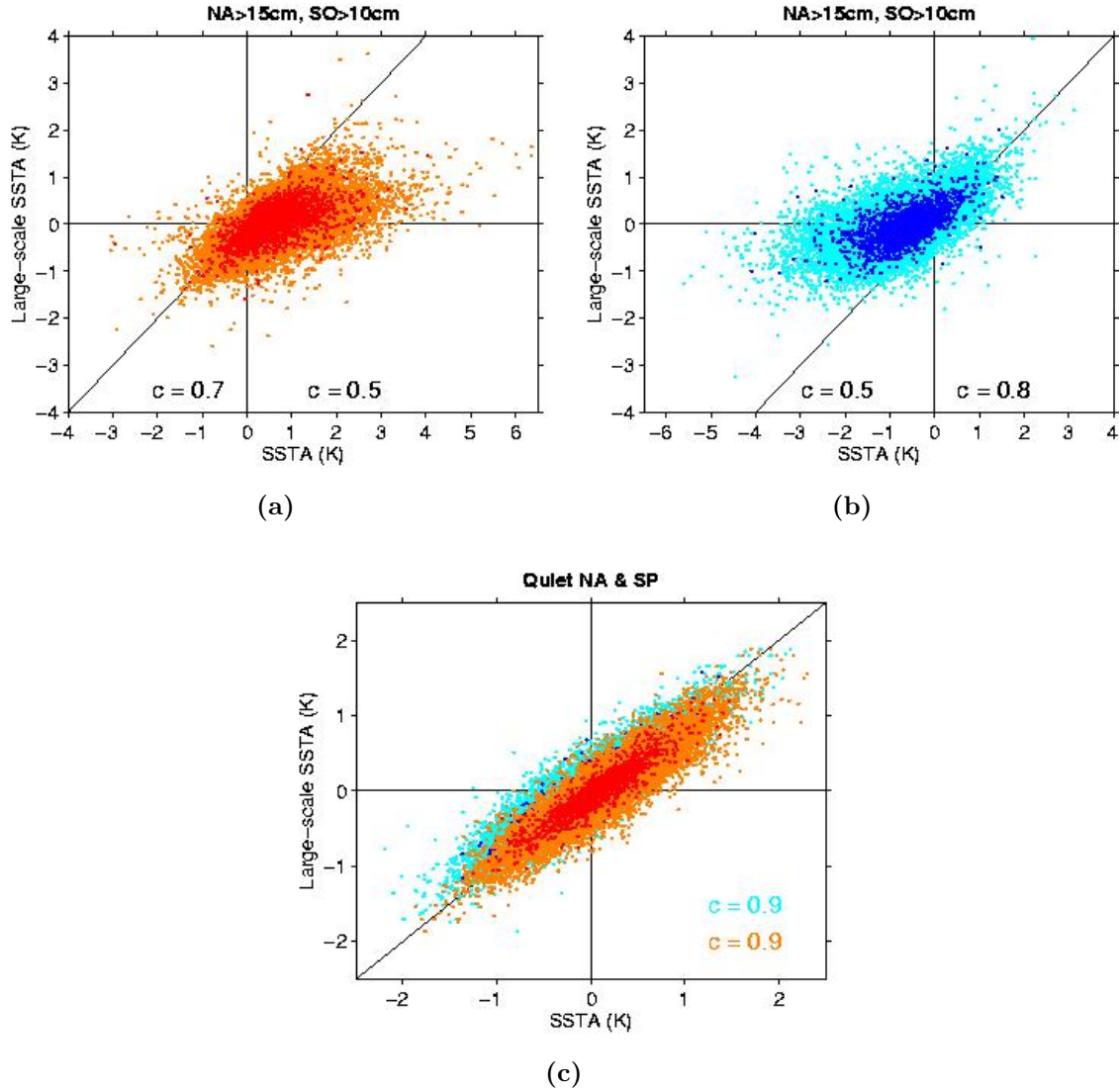


Fig. 4.13: Scatter between eddy-core T' and T'_{LS} (large-scale SSTA observed along the eddy paths), displayed in (a) & (b) for propagating anticyclones & cyclones with origins in the energetic regime, in (c) for eddies of both polarities with origins in the quiet regime. Orange & cyan dots show observations over individual, weekly snapshots of anticyclonic & cyclonic eddies, red & blue dots over each eddy's lifetime-average. Observed correlations (c) between single-snapshot T' and T'_{LS} are estimated separately for $T' < 0$ and $T' > 0$ in the energetic regime, together for the whole ensemble in the quiet regime.

large-scale SSTA variability (grey). Here the typical amplitude of the fluctuations of (full) eddy-core T' between individual eddy snapshots ($\sigma(T') \sim 0.74$ K) is nearly twice as large as that of large-scale SSTA fluctuations ($\sigma(T'_{LS}) \sim 0.44$ K).

In contrast, over quiet regions' eddies (Fig. 4.12b), the distributions of full and large-scale T' closely coincide. They only differ by a slight, but significant, shift towards warm T' for anticyclones and cold T' for cyclones (as established in the previous paragraph). Compared to energetic regions, the amplitude of SSTA fluctuations between eddy events is reduced to about $\frac{2}{3}$ in quiet interiors ($\sigma(T') \sim 0.51$ K). Here it is therefore comparable to that of large-scale SSTA fluctuations ($\sigma(T'_{LS}) \sim 0.45$ K), whose amplitude varies only little between oceanic regimes.

Figure 4.13c displays the scatter between the fluctuations in full and large-scale T' observed over quiet regions' eddy events. It is observed to be very weak, and, over both anticyclonic (orange) and cyclonic eddy snapshots (cyan), the two fields are highly correlated to each other ($c(T', T'_{LS}) \sim 0.9$). Such, over a given snapshot of an eddy core in the quiet regime, the observed T' is almost always determined by the large-scale SST anomaly prevailing at the eddy position. As inter-eddy SSTA fluctuations are a close proxy for the observed (fixed-location) SST variability⁴, the track-following analysis underlines the large-scale nature of quiet regions' SST variability established above by the fixed-location analysis.

The track-following analysis moreover explains why the weak, but systematic, quiet regions' eddy T' signal is invisible when observing at a fixed-location. It is masked by the predominant large-scale SST variability whose typical amplitudes are nearly 5 times as large as that of eddy T' (cf. Table 4.1). As shown by the red and blue dots in Figure 4.13c, even eddy lifetime-average T' show the same strong association with T'_{LS} , as observed over single snapshots. Averaging over each individual eddy's lifetime is thus not enough to isolate the systematic eddy T' signature, which is only obtained after substantial further filtering (i.e. averaging over a large number of eddies, as well as their lifetimes) of large-scale variability.

In energetic regions, the scatter between T' and T'_{LS} observed over eddy cores

⁴ As shown in Table 4.1, both variables have comparable rms. The only filter is the selective sampling over eddy cores, which are observed to be present at any given location in the quiet regime about a quarter of the time. If eddy tracking captures a significant fraction of the observed mesoscale variability (cf. CSS11 for a detailed assessment of the fraction of variability captured by the tracking procedure), in the remaining time, SST variability will be controlled by non-mesoscale processes. Therefore, if at all, fixed-location SST variability should show even more large-scale control than inter-eddy SSTA fluctuations.

conveys a different picture. It is displayed in Figure 4.13 a (for anticyclones) and b (for cyclones). For snapshots of anticyclones with warm SSTA (positive x-axis in Fig. 4.13a) and of cyclones with cold SSTA (negative x-axis in Fig. 4.13b), observed inter-eddy SSTA fluctuations are only weakly correlated to large-scale fluctuations ($c \sim 0.4$). They are therefore interpreted to principally reflect mesoscale processes, such as differences between eddies and their interaction with other eddies and the mean flow. However, over eddy snapshots that show cold (warm) anomalies over anticyclones (cyclones), large-scale and full SSTA are observed to be highly correlated ($c \sim 0.7$). This is interpreted as the distinct influence of large-scale atmospheric forcing, manifested, and in part masking the eddy T' signal, even in these regions of high SSHA rms. Together with the observed near doubling from large-scale to full inter-eddy SSTA rms amplitudes (cf. Table 4.1), this shows that, in the energetic regions of GS and ACC, *both* mesoscale and large-scale processes contribute to zero-order to the observed SST variability. This explains why, even in energetic regions, in a given eddy snapshot, as shown by the SSTA pdf, SSTA alone does not always reveal the eddy's presence.

The shapes of the observed eddy-core T' distributions (Fig. 4.12) provide further evidence for the above conclusions on the mesoscale contribution to observed SSTA variability. Large-scale forcing of SST variability, represented by T'_{LS} , shows an approximately normal distribution in both regimes⁵. Consistent with large-scale control on the observed variability in the quiet regime, here the observed distributions of eddy-core T' closely follow that of the large-scale forcing. In the energetic regime, however, eddy-core T' distributions are observed to differ from that of large-scale SSTA, not only in mean and spread, but also in their shape. For eddies of each polarity, they are skewed towards eddy events with smaller than average eddy-core T' amplitudes ($\text{skew}(T') = +0.94$ over anticyclones, -0.7 over cyclones). Although less pronounced, this skew is reminiscent of that of the eddy amplitude, or V_e , distributions themselves, which have been displayed in Fig. 3.13 and show a pronounced skew not only in the energetic ($\text{skew}(V_e) \sim \pm 2$), but equally so in the quiet regime. If V_e can be thought of as representing the mesoscale eddy forcing of SST variability, the observed energetic regions' eddy T' distributions can be interpreted as a mixture between the one-polarity and Gaussian large-scale forcing, and the two polarity and skewed mesoscale forcing of SST variability. To the contrary, observed quiet regions' eddy T' distributions would then show little signature of the eddy forcing.

⁵ T'_{LS} has 0 skewness in quiet regions, and a very weak, positive skewness in energetic regions, especially over the GS.

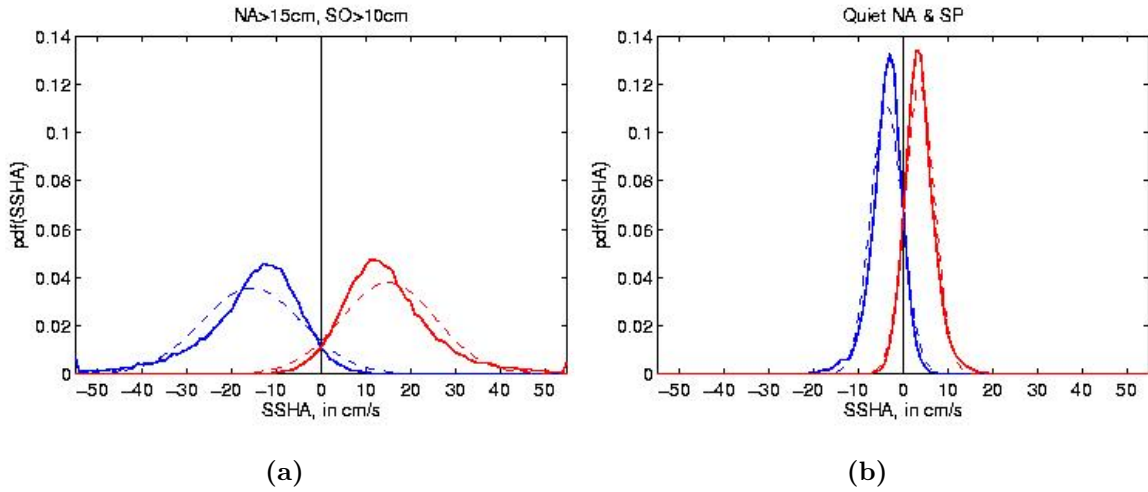


Fig. 4.14: Probability distribution of eddy-core SSHA observed over individual, weekly snapshots of propagating anticyclones (red) and cyclones (blue) with origins in (a) energetic and (b) quiet regimes of NA and SO. Dashed curves show the corresponding normal distributions (mean and standard deviation equal to observations).

The combination of track-following and fixed-location analysis is thus able to provide an answer to the question of the importance of the mesoscale contribution to SST variability (question (ii) above):

- over major current systems, both, scales smaller and larger than 5° -by- 5° , i.e. meso- and large-scales, together shape observed SSTA variability.
- in quiet basin interiors, observed SSTA variability occurs on scales larger than 5° -by- 5° , that masks the systematic eddy T' in individual eddy snapshots and when observing from a fixed-location. Therefore here, to zero order, the mesoscale contribution to SST variability is negligible.

4.3.2 Synthesis

Here we provide a synthesis of the track-following and fixed-location analyses of the observed SSH-SSH relationship provided so far, and examine possible mechanisms behind it. (Important further analysis of this relationship, its mechanisms as well as implications, is provided in chapter 6, as well as in chapter 5).

Synthesis with fixed-location observations To facilitate comparison with the analysis at fixed-location, observed SSHA are evaluated over eddy cores in the same

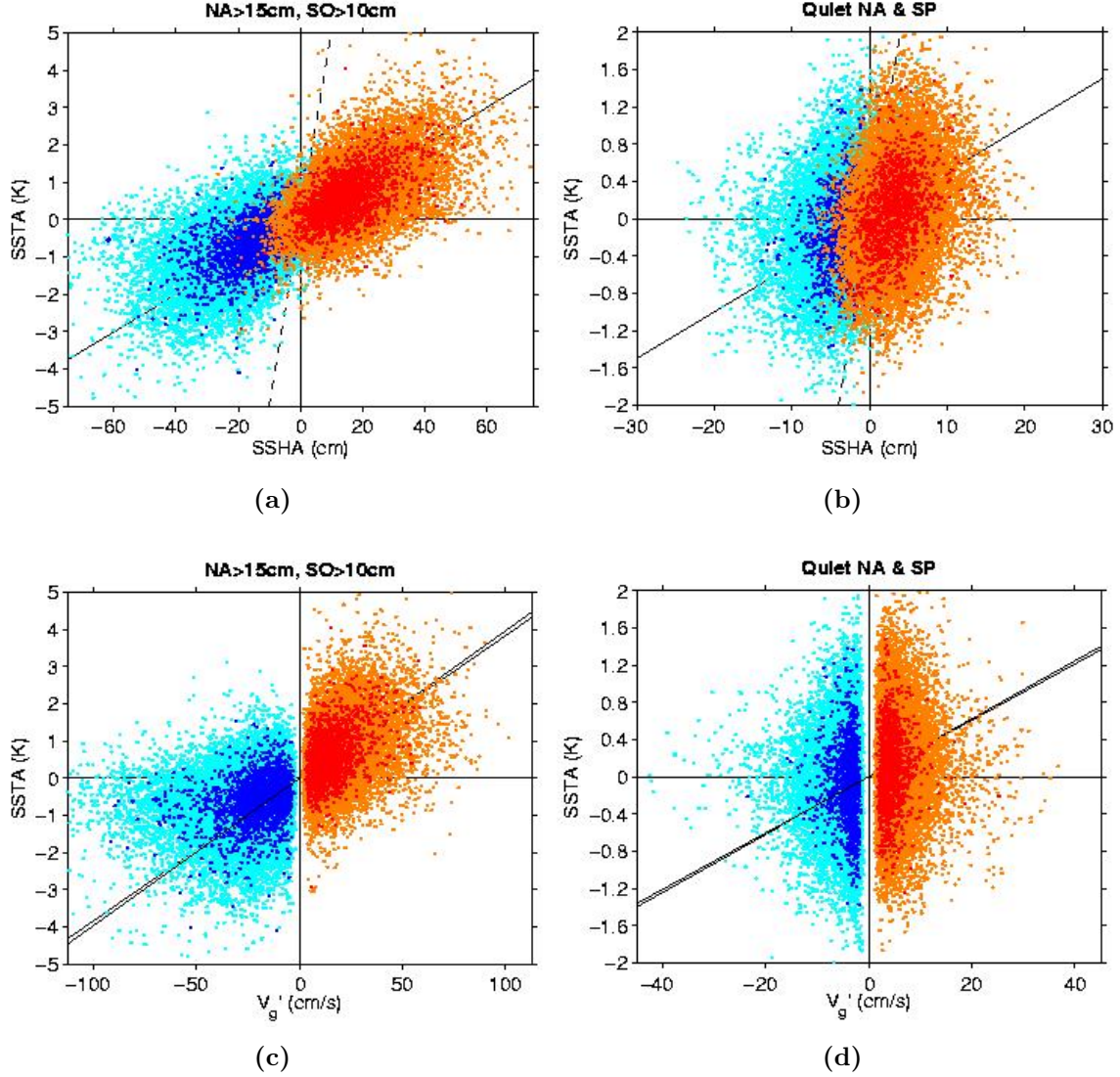


Fig. 4.15: The scatter of eddy-core T' versus eddy-core η' (top panels), and versus V_e (bottom panels), for propagating eddies with origins in energetic (lhs panels) and quiet regimes (rhs panels) of NA and SO. Observations over individual, weekly snapshots of anticyclonic & cyclonic eddies are scattered in orange & cyan, the averages over the lifetime of each eddy in red & blue. Solid lines correspond to the constant ratio of $0.5 K/10 cm$. (In bottom panels, $\langle g/|f| L_e \rangle$, estimated separately for anticyclones and cyclones, is used as conversion factor between eddy amplitude and speed). In the top panels, dashed lines additionally indicate a different constant ratio of $0.5 K/1 cm$.

way as described above for T' . Resulting pdfs are displayed in Figure 4.14. They show clear differences with respect to those of the eddy amplitude A_e (not shown), or similarly, the eddy flow speed V_e (shown in Fig. 3.13). Observed all-eddy average $\langle \eta' \rangle$ (cf. Table 4.2) are similar to $\langle A_e \rangle$, but don't show the latter's asymmetry towards more intense cyclones in the energetic regime.

The typical rms amplitude of the fluctuations in η' between individual eddy snapshots reaches, at ~ 10 cm, about $\frac{2}{3}$ of the all-eddy average $\langle \eta' \rangle$ in the energetic regime. In the quiet regime, the η' rms is nearly as large as $\langle \eta' \rangle$ (~ 3 -4 cm). Eddy-core η' is therefore not an unambiguous indicator of an eddy's presence, as high anomalies in sea level over cyclones, and low anomalies over anticyclones, are observed over a substantial number of eddy snapshots, $\sim 12\%$ in quiet regions, but also nearly 5% in the energetic regions of GS and ACC. Although they occur typically 4 times less often than warm SSTA over cyclones and cold SSTA over anticyclones, attributed above to a masking of the eddy SST signal by large-scale SST variability, the observed opposing polarity eddy-core η' highlight the important contribution of large-scale variability, through steric expansion, also to the observed SSHA field. Especially pronounced over the weak eddies in quiet regions, it is also distinguishable over energetic eddies.

These large-scale SSHA signals are filtered out in the estimation of eddy amplitude, and thus V_e , through their definition from the difference between the extremum of η' observed within the eddy interior and the average of observed surrounding η' . The different nature of eddy V_e and η' is also reflected in differences in shape of between their observed distributions. Compared to V_e pdfs (Fig. 3.13), η' pdfs (Fig. 4.14) are much more Gaussian, with skews reduced by a factor of 2 (from 2 to 1) in energetic regions, by more than factor of 4 in quiet regions. From the same arguments as discussed above for T' , this is expected if large-scale effects are important in shaping the observed η' variability between eddy cores.

Figure 4.15a, b display the observed scatter between T' and η' observed along eddy paths. For both energetic and quiet regimes, it shows a systematic association, in individual snapshots as well as over eddy lifetime averages. But whereas over energetic eddies (panel a) the scatter approximately follows a T'/η' ratio of $0.5 \text{ K}/10 \text{ cm}$, which is indicated by the solid lines, over quiet eddies it more closely follows a ratio of $\sim 0.5 \text{ K}/1 \text{ cm}$, indicated by the dashed lines. To facilitate the interpretation of these observations, it is useful to estimate the depth over which the observed T' anomaly would need to extend unmodified, in order to generate the observed η' anomaly through steric expansion or contraction. From $\eta' = \int_{-H_{obs}}^0 \alpha_T T' dz \sim \alpha_T T' H_{obs}$, this depth is

obtained as $H_{obs} = \frac{\eta'}{\alpha_T T'}$. The ratio of $\sim^{0.5} \text{ K}/_{10 \text{ cm}}$, observed over energetic eddies, thus corresponds to a compensation depth of $\sim 1 \text{ km}$. This is consistent with observed eddy T' reflecting the extension of deep eddy temperature anomalies, in thermal wind balance with observed eddy η' , into the surface mixed-layer. In contrast, the ratio of $\sim^{0.5} \text{ K}/_{1 \text{ cm}}$, observed over the quiet eddies, corresponds to $H_{obs} \sim 100 \text{ m}$, which is close to the depth of the surface mixed-layer. The strong association between η' and T' observed over quiet regions' eddy snapshots, as well as lifetime averages, is therefore consistent with the predominant influence of mixed-layer thermosteric signals on the variability of both.

For comparison, the bottom panels of Figure 4.15 present the observed scatter between eddy T' and V_e . Over energetic eddies, the association of T' with the unambiguous indicator of the eddy presence V_e (panel c), is nearly as strong as that to η' (panel a), prone to steric contamination, and follows a comparable slope. In contrast, over quiet eddies, the correlation is completely lost. Consistent with the shallow compensation depth required to explain observations in the quiet regime, here the observed strong correlation between T' and η' along eddies' paths therefore only reflects shallow, steric, and likely large-scale, variability.

The statistically significant positive zero-lag correlation between SSTA and SSHA, observed in the quiet regime at fixed-location (Fig. 4.4), is thus also identified as solely reflecting the correlated (large-scale) steric variability in the two fields. This confirms, what has already been hinted at by the fixed-location analysis itself. Without prior filtering, an analysis of fixed-location observations of SST and SSH cannot detect a mesoscale contribution to SST variability in quiescent gyre interiors.

Interpretation with respect to possible mechanisms In the quiet regime, a systematic eddy T' signal only appears after averaging over many eddy paths. Consistently, in the quiet regime, the all-eddy average $\langle T' \rangle / \langle \eta' \rangle$ ratio, which is unaffected by large-scale variability, clearly differs from that observed over individual eddies (as indicated by the slope of the scatter in Fig. 4.15 b). Instead, it is comparable to the ratio observed over energetic regions' eddies (be it in single snapshots or in the bin-average). Scaled by the eddy amplitude, the quiet regions' eddy T' signal is therefore nearly as intense as that of energetic eddies, and also broadly consistent with reflecting deep-reaching warm and cold eddy cores in thermal wind balance with eddy η' .

As summarized in Table 4.2, the observed quiet regime's ratio is actually slightly weaker than the $^{0.4} \text{ K}/_{10 \text{ cm}}$ $\langle T' \rangle / \langle \eta' \rangle$ ratio observed over energetic eddies. Specifically, and

Tab. 4.2: Observed eddy amplitudes.

	Energetic			Quiet		
	ALL	GS	ACC	ALL	NA	SP
$\langle \eta' \rangle$, in cm						
A	15.1	21.3	14.2	3.7	4.5	3.3
C	-15.5	-24.7	-14.3	-3.7	-4.6	-2.9
$\langle T' \rangle$, in K						
A	0.60	0.56	0.61	0.12	0.08	0.14
C	-0.60	-0.64	-0.60	-0.09	-0.05	-0.12
$\langle T' \rangle / \langle \eta' \rangle$, in $K/10 \text{ cm}$						
A	0.40	0.26	0.43	0.31	0.18	0.43
C	0.39	0.26	0.42	0.23	0.11	0.41
$\langle V_e \rangle$, in cm/s						
A	16.1	22.8	15.0	5.5	6.4	4.9
C	-22.6	-39.4	-20.3	-5.1	-6.9	-3.5
$\langle V_e \nabla \bar{T} \rangle$, in $K/10 \text{ days}$						
A	1.50	2.50	1.36	0.21	0.16	0.24
C	-1.87	-3.14	-1.69	-0.16	-0.16	-0.16
$\langle T' \rangle / \langle V_e \nabla \bar{T} \rangle$, in days						
A	4.0	2.2	4.5	5.6	4.9	5.9
C	3.2	2.0	3.5	5.3	3.1	7.2

Averages over eddy-cores and all snapshots of eddies with origins in the indicated bins. A: anticyclones, C: cyclones, NA: North Atlantic, SP: South Pacific.

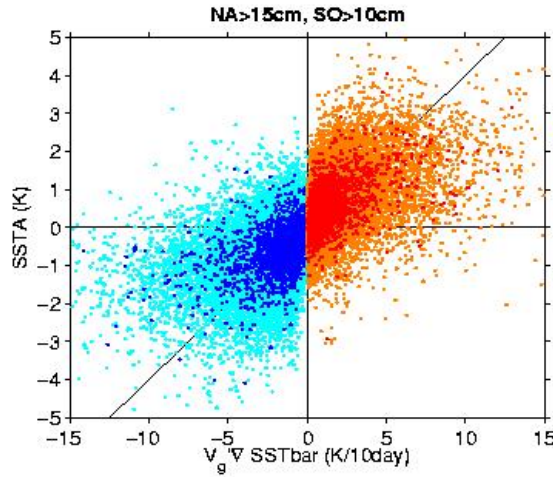


Fig. 4.16: The scatter between eddy-core T' and the intensity of eddy isotherm stirring, estimated as $V_e |\nabla \bar{T}|$, with the latter factor averaged over eddy cores, observed over individual, weekly snapshots (orange & cyan) and lifetime averages (red & blue) of anticyclonic & cyclonic eddies with origins in the energetic regime of GS and ACC. The solid lines correspond to the constant ratio $T' / V_e |\nabla \bar{T}| = 4$ days.

in contrast to energetic eddies, it is weaker over cyclones compared to anticyclones, and, as in the energetic regime, weaker over NA compared to the SO eddies. Repeating the above calculation of the required depth, over which eddy (potential) temperature anomalies would need to extend, if vertically constant, to annul the eddy pressure perturbation, with bin-average values, yields ~ 1 km for SO eddies in both ACC and quiet SP, ~ 2 km for GS eddies and ~ 3 -4 km over eddies with origins in the NA subtropical gyre. Deeper reaching baroclinic eddy circulations over GS than ACC, as well as extremely deep-reaching baroclinic eddy cores in the quiescent NA, seem to be unlikely, and rather point to the importance of neglected physics in this simplistic calculation. These include barotropic contributions to eddy η' , as well as variation of the eddy temperature perturbation with depth, either in the form of a continuous depth-decline or as a discontinuity at the mixed layer base.

In energetic regions, where large-scale contamination is relatively weak, the $T'-V_e$ relationship observed over individual eddies (Fig. 4.15c) can provide further insight. It reveals an asymmetry between the two polarities, such that a given V_e is associated with slightly larger T' over anticyclones than over cyclones. Specifically, the scatter shows pronounced outliers, present only over cyclones, for which very high eddy velocities (~ 1 m/s and higher) are associated with very weak negative or even slightly positive eddy-core T' . As shown in Figure 4.16, these disappear, in this pronounced form, if eddy-core T' are instead scattered against the eddy stirring intensity, estimated as $V_e \left| \nabla \overline{T} \right|$ evaluated along eddy tracks. Consistent with the fixed-location analysis, this hints towards the importance of horizontal eddy stirring in the mixed-layer in generating the observed eddy T' . Lower V_e - T' correlations over cyclones are especially pronounced over GS eddies (not shown), and can thus be understood from the preferential location of cyclones over warm current cores, where they cannot impact SSTA through stirring.

The observed scatter in Fig. 4.16 closely follows a $T'/V_e \left| \nabla \overline{T} \right|$ ratio of $0.6 \text{ K}/1.5 \text{ K}/10 \text{ days}$, comparable to that obtained from energetic regions' all-eddy averages (Tab. 4.2). If stirring is the only process through which eddies generate mixed-layer T' , this yields an estimate for the observed equilibration time scale $\tau = T'/V_e \left| \nabla \overline{T} \right|$ of ~ 4 days, which is extremely fast. In the quiet regime all-eddy averages yield comparable equilibration times (between 5 and 6 days, cf. Tab. 4.2). These short observed equilibration time scales suggest that a local, near-instantaneous generation (i.e. over each weekly snapshot) of observed eddy T' by eddy isotherm stirring is possible. Whether a possible damping of eddy T' to the atmosphere can provide the necessary equilibration of the observed eddy stirring, will be further discussed in chapters 5 and 6. In the

mixed-layer, both dynamics (2-D versus 3-D) and thermodynamics (interaction with the atmosphere) are thus different from the interior, and might in this way be at the origin of a possible disconnect in the eddy temperature perturbation at the mixed-layer base, as hinted at above.

In summary, the track-following analysis is consistent with the importance, demonstrated by the analysis of the observed fixed-location SSTA-SSHA relationship, of eddy stirring in setting the mesoscale window “to the atmosphere”. In regions of weak eddy stirring, the systematic eddy T' are masked variance-wise by (likely atmospherically-forced) large-scales that dominate the observed SST variability. The next section will shed further light on how deep into the ocean the atmosphere might see through this mesoscale window, and on the processes that set its extent in space and time.

4.4 The seasonal modulation of a mesoscale contribution to SST variability

This section examines the possibility of temporal variations in the observed mesoscale contribution to SST variability, and thus in the “mesoscale window to the atmosphere” (cf. section 4.2.4), focussing on seasonal time scales. As SSTA persistence is typically on the order of a season on large spatial scales (Frankignoul, 1985), possibly shorter on mesoscales (chapter 5), the impact of possible seasonal modulations in SSTA forcing mechanisms is expected to be seen. After a quick note on the method (section 4.4.1), section 4.4.2 establishes that the spatial scales of SST variability indeed undergo a substantial seasonal modulation. In section 4.4.3 the observed seasonality is used as tool to better understand the mechanisms that control whether, where and when the mesoscale significantly contributes to SST variability.

4.4.1 Method

In section 4.2, 5° -by- 2° box-averages of the spatial-lag SSTA auto-correlation matrix, denoted $c_{5^\circ,2^\circ}(T')$ and estimated following (4.2), have been established as a good observational measure of the mesoscale contribution to SST variability. To examine their possible seasonality, $5\frac{1}{2}$ -year SSTA time-series are divided into 22 equal-length 90-day chunks, representing 6 realisations of JJA and SON, 5 of DJF and MAM. $c_{5^\circ,2^\circ}(T')$ are estimated for each of these chunks in the same way as above (with the only difference of no prior linear de-trending, due to the short time series length), and subsequently averaged over all realizations of a given season. The analysis to be presented below

(sections 4.4.2 & 4.4.3) thus only applies to intraseasonal and higher-frequency SST variability.

Before examining the variations in $c_{5^\circ, 2^\circ}(T')$ between individual seasons, here we first assess the impact of the differing statistics (5½-year versus 90-days) on the estimates, by averaging $c_{5^\circ, 2^\circ}(T')$ over all available 90-day chunks. The spatial distribution of the resulting quantity (not shown) shows very similar features to $c_{5^\circ, 2^\circ}(T')$ estimated directly from the full 5½-year time series, which has been displayed in Fig. 4.7a, b above. Figure 4.17 displays the difference between the two estimates. It shows that the estimate from the full time series results in typically +0.05 to +0.2 larger box-average correlations throughout ocean basins. The 90-day based estimate is thus systematically biased towards smaller scales. This bias is observed to be especially large in the equatorial North Atlantic, and reverses sign only in the regions of high SSHA rms (indicated by the black contour), associated with GS, ACC and Mexican Loop Current. Here multi-year SSTA time series therefore decorrelate quicker in space than 3-month time series.

The observed differences can be understood from the different time scales of SST variability captured by the two estimates. Throughout basin interiors and especially in the tropics, interannual SST variability, not present in the 3-month estimate, is observed to occur on larger spatial scales than the variability on synoptic to intraseasonal time scales. In contrast, in regions of intense eddy variability, the contribution to SST variability of processes on spatial scales smaller than 5°-by-2° is observed to become even more pronounced as time scales longer than ~100 days start to be captured. Here, the mesoscale therefore likely contributes to SST variability, not only on time scales comparable to the ~100 days, traditionally thought of as characterizing the mesoscale, but also on interannual time scales. This is consistent with other recent discoveries of low frequency variability on the oceanic mesoscale (Qiu et al., 2007; Wunsch, 2010).

4.4.2 Observed seasonality

Figures 4.18 and 4.19 display the SSTA spatial correlations, $c_{5^\circ, 2^\circ}(T')$, observed during the 4 different seasons, for NA and SO, respectively. All seasons display the same contrast, as observed in the annual mean (Fig. 4.7a, b), between large scales of SST variability (red shades) prevailing in basin interiors, and small scales (blue shades) close to the western boundary and along the path of the ACC. As approximate cut-off between these two modes of variability, the 0.6 $c_{5^\circ, 2^\circ}(T')$ contour is indicated in thin blue. The smallest observed correlations (dark blue shades), comparable to ~0.3 or

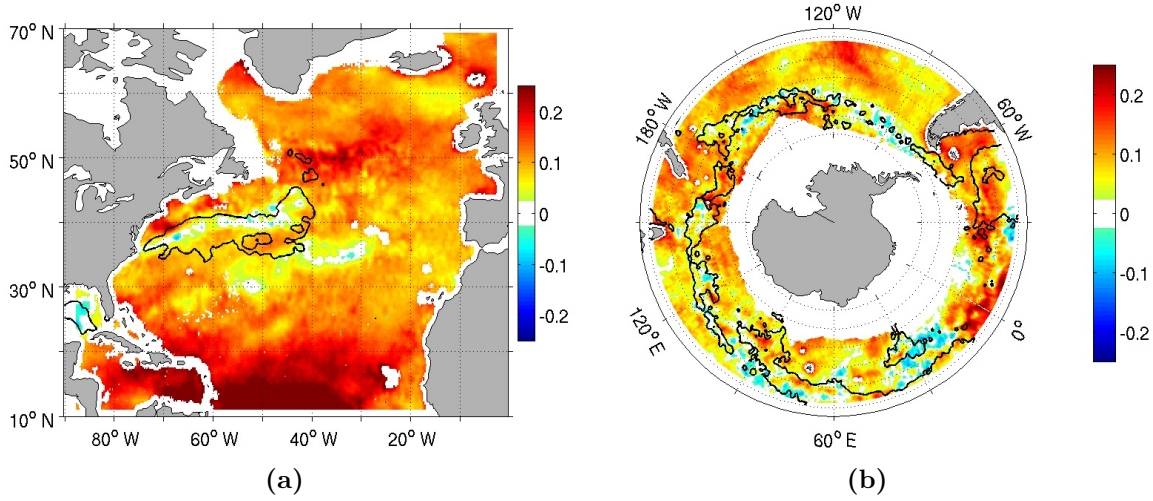


Fig. 4.17: The difference between SSTA spatial correlations, $c_{5^\circ, 2^\circ}(T')$, estimated from the full 5½-year time series, and the average of $c_{5^\circ, 2^\circ}(T')$ estimated from individual 90-day chunks, illustrating the dependence of the spatial scales of SST variability on time scale. The 15 and 10 cm $\sigma(\eta')$ levels are contoured in black, for NA (a) and SO (b), respectively.

smaller, have been established above to correspond to typical oceanic mesoscales (cf. Fig. 4.6 and associated text). Throughout the year, they are found to coincide with the regions of enhanced eddy isotherm stirring, marked by the black $1 \text{ K}/_{10 \text{ day}} \sigma(\mathbf{u}'_g \cdot \nabla \bar{T})$ contour estimated separately for each season.

On top of this broad time-mean distribution, visible in each individual season, lie pronounced seasonal differences. In the winter season (displayed in the top-left panels), the smallest correlations (comparable to ~ 0.3) cover large areas, approximately delimited by $1 \text{ K}/_{10 \text{ day}}$ winter stirring contour. Intermediate scales (correlations ≤ 0.6) are found to stretch into basin interiors far beyond the intense stirring regions (especially in the NA and Indian ocean subtropical gyres). Large scales (≥ 0.75), in turn, are observed to be limited to the far eastern subtropical gyres of NA and SP, as well as to the Labrador sea proximity of the NA subpolar gyre. In spring (top-right panels), very small scales of SST variability are still observed to prevail throughout the regions of intense stirring, but the observed large scales of variability become larger (reddening shades) and reach across larger parts of basin interiors. In summer (bottom-left panels), large-scale SST variability predominates in almost the entire basins, with low correlations (~ 0.3) receding to a small portion of the GS's intense summer-time eddy stirring region (black contour). Large summer-time SSTA spatial correlations, exceed-

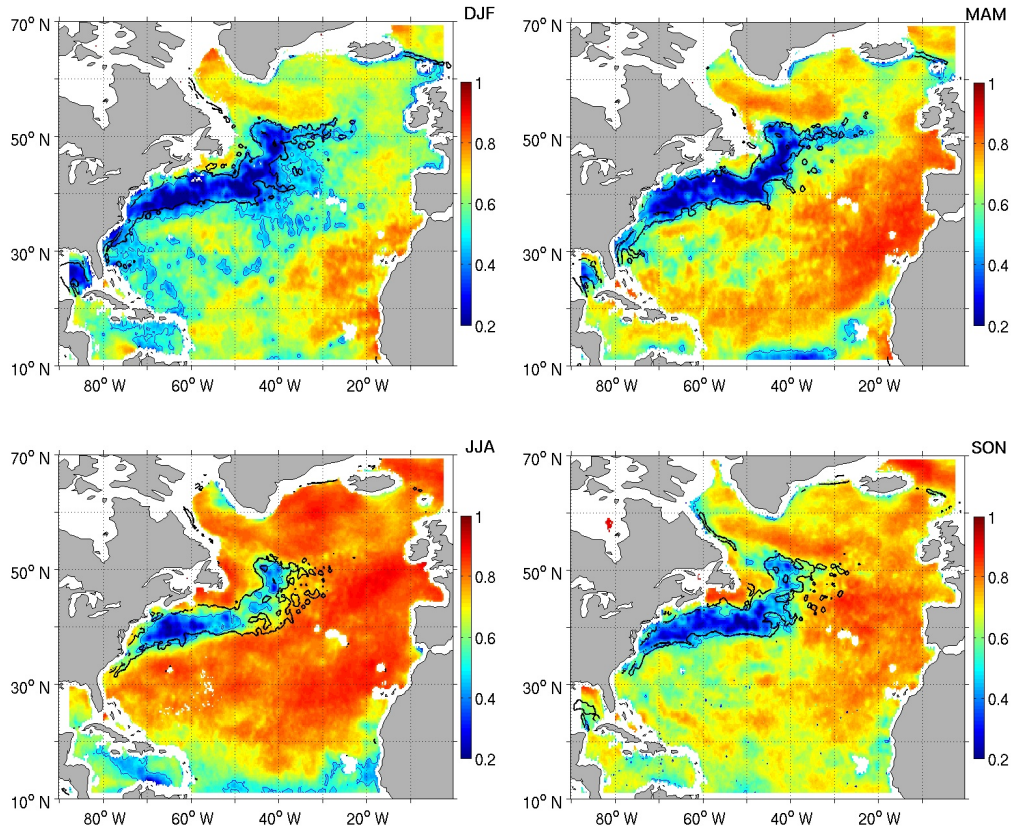


Fig. 4.18: The 4 seasons of observed SSTA spatial correlations, $c_{5^\circ, 2^\circ}(T')$, mapped for the NA ($c=0.6$ contoured in thin blue). Top-left: winter, top-right: spring, bottom-left: summer, bottom-right: autumn. The thick black contour marks an eddy stirring intensity of $1 \text{ K}/10 \text{ day}$, estimated separately for each season as $\sigma(\mathbf{u}'_g \cdot \nabla \overline{T})$.

ing 0.6 (thin blue contour), are also observed in parts of the high-stirring areas of ACC. In the autumn (bottom-right panels), regions of low SSTA correlations again become more extensive.

A significant contribution of the oceanic mesoscale to observed SST variability is thus observed to be most intense, as well as to cover the largest spatial extent, in the winter and spring season. The most pronounced exception to this observed general seasonal progression in the nature of SST variability, is formed by the NA tropics (equatorward of $\sim 15^\circ\text{N}$). Here, the smallest scales of variability are observed in spring and summer, instead of in winter (and spring) as in the extratropical NA and SO.

By displaying the difference of $c_{5^\circ, 2^\circ}(T')$ between individual seasons, Figure 4.20 reveals two main modes of the seasonal modulation in the nature of SST variabil-

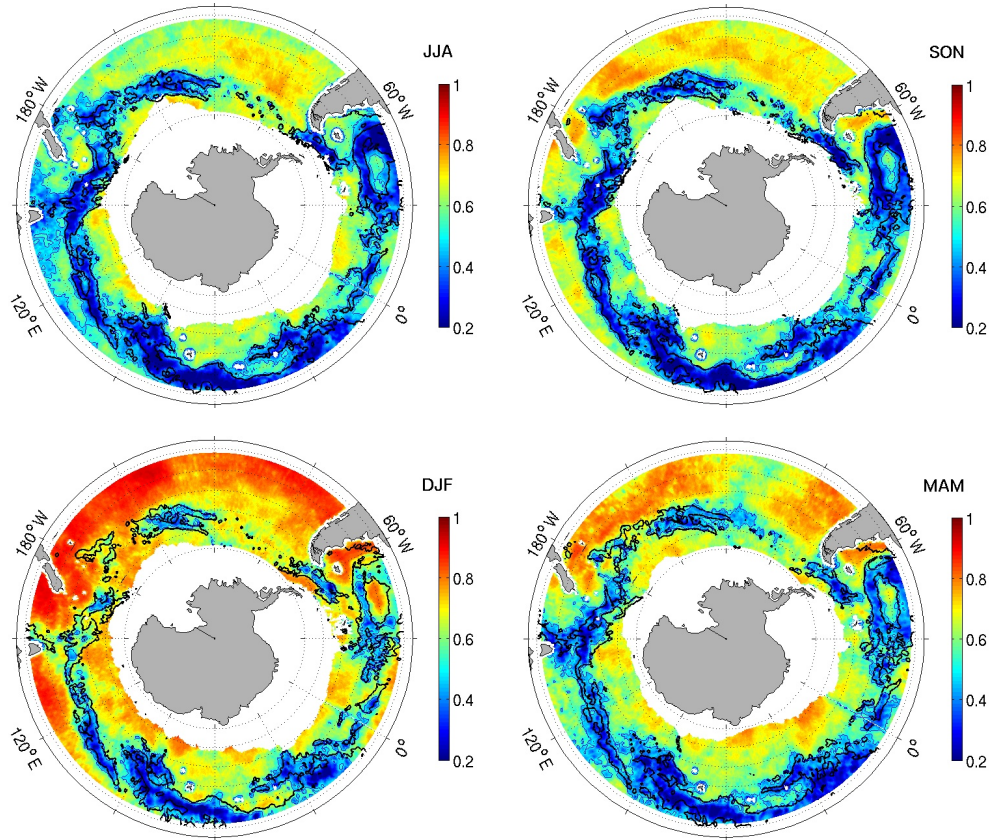


Fig. 4.19: Same as Fig. 4.18, but for the SO.

ity. Compared to the summer season, in the winter (lhs panels) SST variability shifts towards smaller scales over nearly the entire extratropical oceanic basins. The few exceptions to this observed basin-scale winter-time scale decrease include the tropics, very high latitudes (off Greenland, poleward of the ACC), as well as a few patches within GS and ACC. The observed summer to winter drop in $c_{5^\circ, 2^\circ}(T')$ has a typical amplitude of 0.25. It is less pronounced, both, in regions where large scales predominate the variability throughout the year (south of the ACC, in the eastern SP and the south-eastern NA STG), as well as in some parts of the intense eddy stirring regions of GS and ACC (indicated by the black contour), characterized by the smallest observed time-mean scales of SST variability.

Observed spring-autumn differences in $c_{5^\circ, 2^\circ}(T')$ (rhs panels of Fig. 4.20) reveal a more subtle, regionally tilting pattern. Compared to the autumn, in spring smaller scales of SST variability emerge along the GS and its extension, the NA subtropical convergence zone (STCZ), the Mexican Loop current, as well as most of the time-mean ACC jets, i.e. in the regions of strong time-mean eddy stirring (contoured). To the

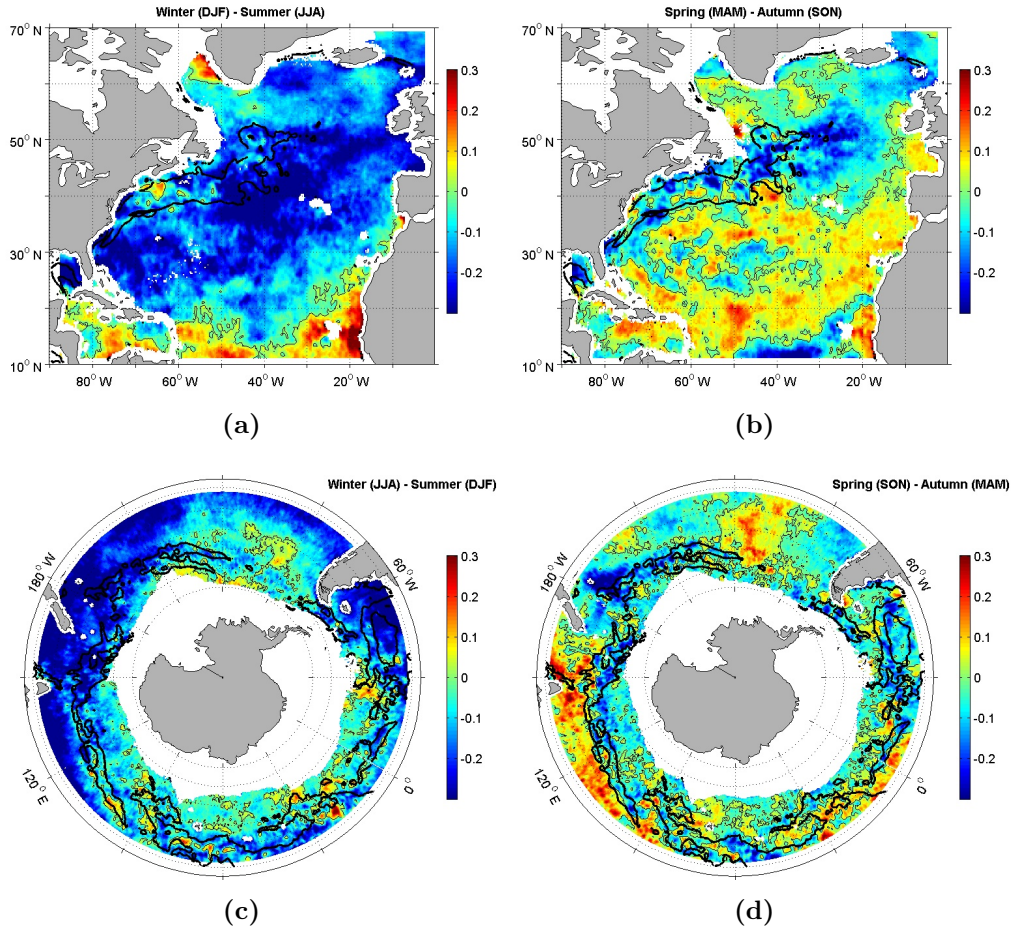


Fig. 4.20: Seasonal differences in $c_{5^\circ, 2^\circ}(T')$ (lhs panels: winter - summer, rhs: spring - autumn) observed in NA (top) and SO (bottom). Sign-change contoured.

contrary, SSTA scales are smaller in autumn across basin interiors, especially in the south-eastern subtropical gyres, as well as in the mode water regions just equatorward of major currents (GS and ACC, particularly in its Indian sector).

The characteristic features of these two main modes of seasonal modulation, a summer-to-winter basin-wide decrease in spatial SSTA scales and a spring-to-autumn tilt towards smaller scales in intense eddy stirring regions and larger scales elsewhere, are indeed observed when estimated from each individual year (i.e. from one, instead of 5-6 realizations of any given season, not shown)⁶. This confirms that the attribution of these modes to seasonal, rather than interannual changes is robust. The next section

⁶ When estimated for individual years, the diagnostic, displayed in Fig. 4.20 for the 5½-year mean, shows larger-amplitude $c_{5^\circ, 2^\circ}(T')$ differences of a more patchy nature, reflecting some degree of inter-annual variability in SSTA spatial scales. These average to the relatively large-scale, weaker amplitude seasonal scale changes observed in the 5-6 year mean.

examines possible mechanisms behind these observed seasonal changes.

4.4.3 Mechanisms

For each season, observed SSHA spatial correlations ($c_{5^\circ, 2^\circ}(\eta')$, not shown) confirm the annual mean pattern of a widespread presence of mesoscale eddies in the SSHA field (cf. Fig. 4.7c, d). The rms amplitudes of SSH variability also remain nearly unchanged throughout the seasons (cf. Fig. 3.8 for the time-mean). A systematic seasonal cycle is only observed in regions of weak time-mean SSH variability (≤ 5 cm), which show a winter-time increase in both spatial scales and amplitudes of SSHA. In agreement with the results of sections 4.2.1 and 4.3.2, this can be attributed to enhanced large-scale, steric SSH variability over the deep winter-time mixed layers. Seasonal variations in the presence, or in the intensity, of mesoscale eddies can thus not account for the observed seasonal dependence of SSTA spatial scales.

In the following we consider whether the intensity of mesoscale SSTA forcing varies with season, and might in this way provide a mechanism for the observed seasonal modulation in the small-scale contribution to SST variability.

Eddy stirring In each season, the intensity of the geostrophic eddy stirring of background mixed-layer isotherms, measured by its rms amplitude $\sigma(\mathbf{u}'_g \cdot \nabla \overline{T})$, shows the same, close to two order of magnitude increase from basin interiors to major current fronts, as in the annual mean (cf. Fig. 4.9). This has been found to account well for the observed time-mean distribution of SSTA spatial scales (cf. section 4.2.3), which is also apparent in each individual season (cf. Figures 4.19 and 4.19). The summer-to-winter and autumn-to-spring differences in $\sigma(\mathbf{u}'_g \cdot \nabla \overline{T})$ are displayed in Figure 4.21. They reveal a pronounced winter- and spring-time intensification of the eddy stirring intensity in the regions of strongest time-mean stirring, especially along the main ACC and GS fronts, but also in several other time-mean frontal regions, such as the NA subtropical convergence zone (STCZ). This intensification is predominately controlled by seasonal variations in the frontal strength $\nabla \overline{T}$, rather than in the eddy amplitude $\sigma(\|\mathbf{u}'_g\|)$ (not shown). This winter and spring time frontal intensification pattern is reminiscent of the autumn-to-spring differences in observed SSTA spatial scales (Fig. 4.20b, d), with reduced scales over intensified eddy stirring in frontal zones. Whereas the observed seasonal modulation in the eddy stirring intensity thus likely plays a role in the observed autumn-to-spring tilting mode of SSTA spatial scales, it can certainly not fully account for the observed basin-wide decrease of spatial scales from summer to winter.

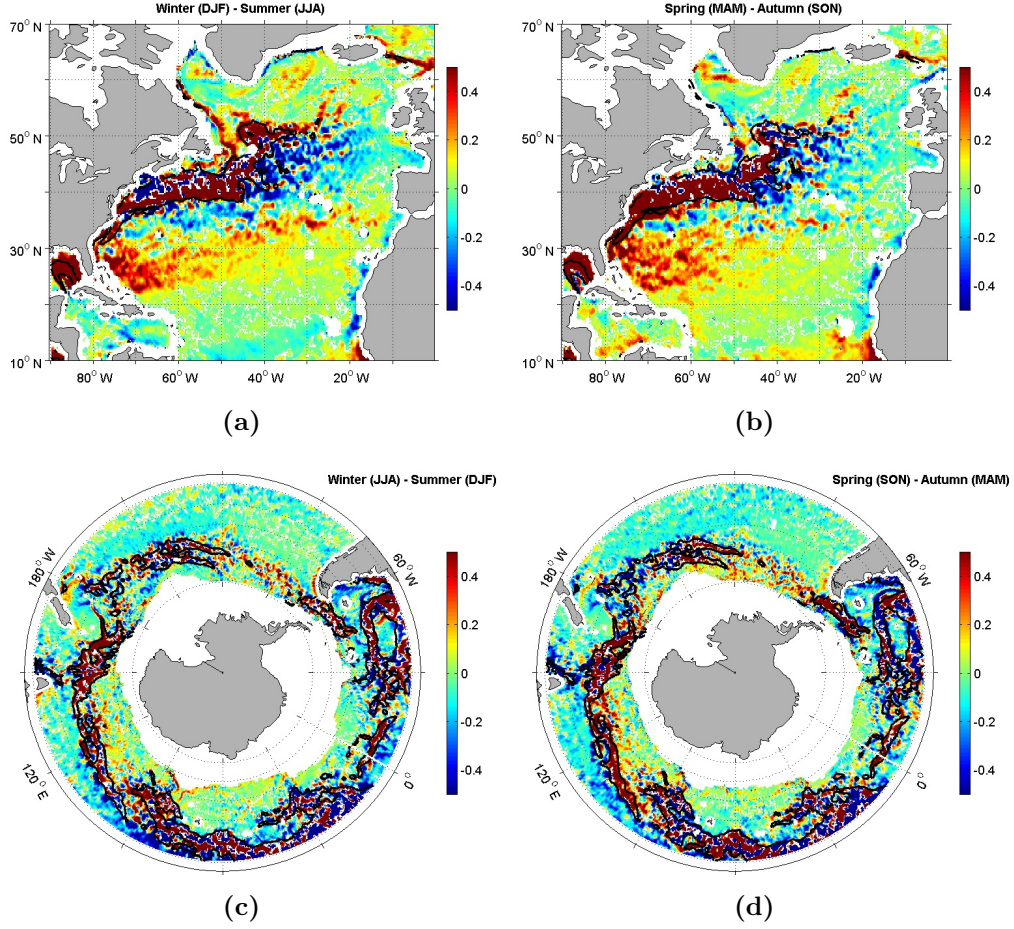


Fig. 4.21: Observed seasonal differences in the eddy stirring intensity $\sigma(\mathbf{u}'_g \cdot \nabla \bar{T})$, in $\text{K}/_{10 \text{ day}}$, mapped only if they exceed 10% of the annual mean. (lhs panels) winter - summer, (rhs) spring - autumn. Black contours delimit regions in which the time-mean stirring rms exceeds $1 \text{ K}/_{10 \text{ day}}$.

Entrainment The large observed seasonal variations in the depth of the surface mixed-layer provide an additional, only seasonally appearing SSTA forcing mechanism, the entrainment forcing $F'_{entr} = \frac{\bar{w}_{entr}}{h_{ml}} (T'_{sub} - T')$ due to one component of the climatological entrainment velocity $\bar{w}_{entr} = \Lambda(\partial_t \bar{h}_{ml})$ (where $\Lambda() = 0$ if its argument ≤ 0 , 1 otherwise), which is by nature seasonal as it only acts during the mixed-layer deepening season⁷. For each season, the rate of the entrainment forcing, $\frac{\bar{w}_{entr}}{h_{ml}}$ (i.e. the

⁷ The full entrainment velocity, $\bar{w}_{entr} = \Lambda(\partial_t \bar{h}_{ml} + \nabla \cdot (\bar{h}_{ml} \mathbf{u}))$, includes contributions from lateral induction and vertical flow across the mixed-layer base, which can both operate throughout the year (if mean currents flow towards deeper time-mean mixed-layers, and in regions characterized by time-mean Ekman suction, such as the NA SPG and the region poleward of the ACC core). The full entrainment forcing moreover includes a term that involves anomalies in the entrainment velocity,

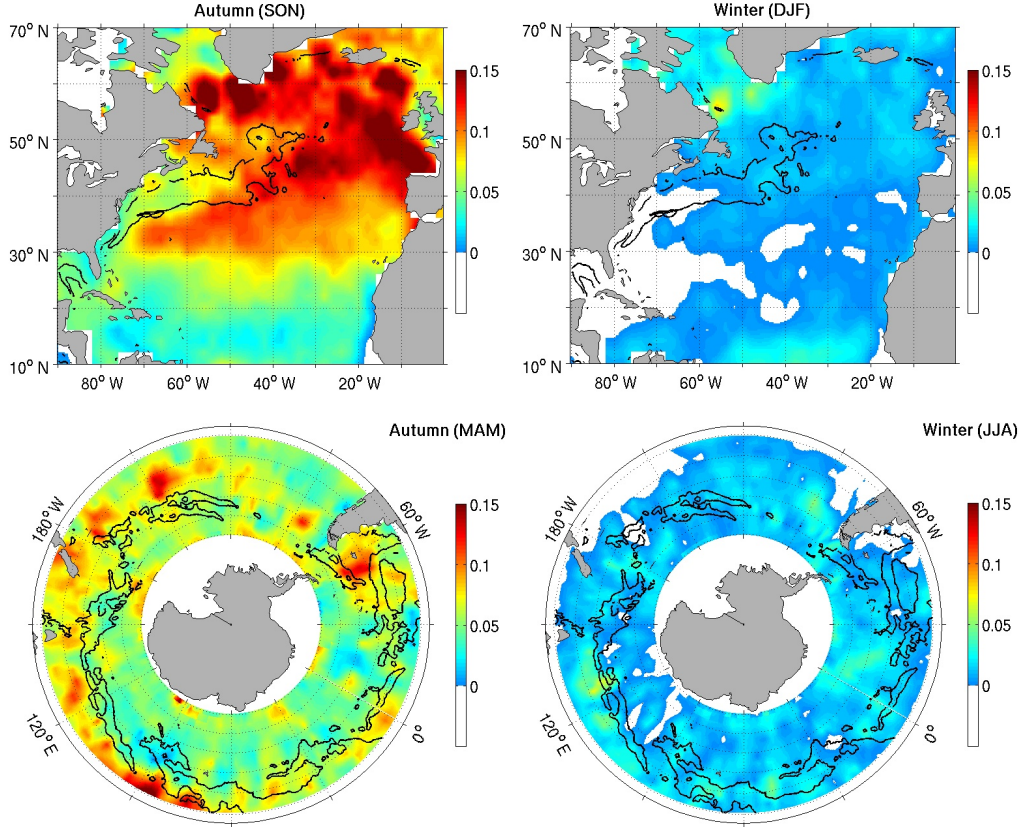


Fig. 4.22: The observed climatological entrainment rate $\bar{w}_{entr}/\bar{h}_{ml}$ is mapped in units of $1/10$ day for autumn (lhs) and winter (rhs panels). ($1/10$ days corresponds to a deepening of $\sim 100\%/100$ days.) Black contours indicate where time-mean $\sigma(\mathbf{u}'_g \cdot \nabla \bar{T})$ equals $1 \text{ K}/10 \text{ day}$.

inverse of the time it takes to restore a SSTA to a subsurface temperature anomaly), is estimated from the observed mixed-layer depth climatology (cf. section B.1) as $\frac{\Lambda(\Delta\bar{h}_{ml}/\Delta t)}{\bar{h}_{ml}}$, with Δ defined as the difference between following and preceding seasons. With a few exceptions (in very high latitudes and tropics), the resulting quantity is zero across oceanic basins in spring and summer, when the mixed layer shoals. As shown in Figure 4.22, it peaks in the autumn, throughout which the mixed-layer deepens by more than 50% in regions of moderate mixed-layer depth seasonality, and by more than 150% in regions of mode-water formation and deep convection. The entrainment rate is much weaker in winter (rhs panels), when mixed-layers are deep and, in most places, continue to deepen only moderately (by typically 25%) until the start of spring.

$\frac{w'_{entr}}{\bar{h}_{ml}}(\bar{T}_{sub} - \bar{T})$, as well as non-linear contributions, which, in principal, both also operate throughout the year, and are not considered here.

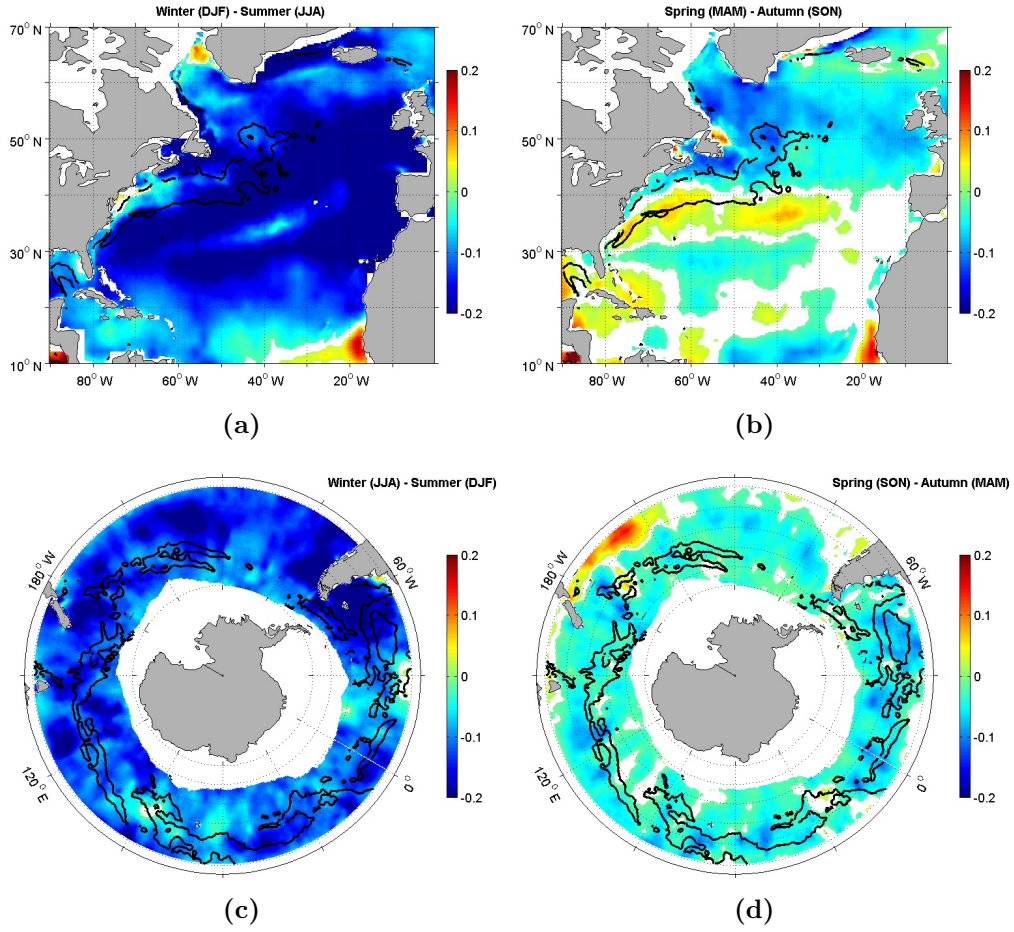


Fig. 4.23: Seasonal differences in $|\overline{\mathbf{u}_a - \mathbf{u}_o}|/\bar{h}_{ml}$ (in K/s), as discussed in the text, a proxy for (one component of) the atmospheric turbulent heat flux forcing of large-scale SSTA (in $\text{K}/10 \text{ day}$). (lhs panels) show winter - summer, (rhs) spring - autumn differences, mapped only if they exceed 10% of the annual mean. Time-mean $\sigma(\mathbf{u}'_g \cdot \nabla \bar{T}) = 1 \text{ K}/10 \text{ day}$ contoured.

If the temperature variability just below the base of the mixed layer is dominated by mesoscale fluctuations, which needs to be further investigated, this component of the entrainment flux would act to create small scales in SSTA throughout autumn and winter. If $T'_{sub} - T'$ is on the order of 1K, it does so with a comparable magnitude as the eddy stirring in basin interiors (upon multiplication by 1K Fig. 4.22 gives F'_{entr} in $\text{K}/10 \text{ days}$). An order of magnitude increase of $T'_{sub} - T'$ would be required to match the stirring intensity also in frontal regions.

Atmospheric forcing Seasonal variations in the depth of the mixed layer also affect the atmospheric forcing of large-scale SSTA. This inversely depends on $\bar{h}_{ml}$ and

is thus much more efficient over the shallow summer-time mixed layers. However, the generally stronger winter-time winds and air-sea contrasts oppose this mixed-layer depth effect. To quickly assess the net effect of these opposing variations on the seasonality of the large-scale atmospheric SSTA forcing, Figure 4.23 maps the winter-summer and spring-autumn differences in $\overline{\Delta u}/\overline{h_{ml}}$. Here the air-sea speed difference $\overline{\Delta u} = |\overline{\mathbf{u}_a} - \overline{\mathbf{u}_o}|$, which is operational in setting not only the Ekman forcing, but also the turbulent heat flux forcing (cf. chapter 2.2.3), is derived from QuickSCAT observations (cf. section B.3). The Figure shows that the large summer-to-winter mixed-layer depth increase by far outweighs the winter-time intensification in differential air-sea speed in both hemispheres. Whereas in the SO the slightly deeper spring-time mixed layers also control the observed spring-autumn differences, in the NA these are set by shifts in the intensity and position of the atmospheric wind systems.

The large-scale SSTA forcing by turbulent air-sea heat flux anomalies is driven by anomalies in the air-sea temperature and humidity difference through $\overline{\Delta u}(\Delta T' + \Delta q')$, by air-sea speed anomalies through $\Delta u'(\Delta \overline{T} + \Delta \overline{q})$, as well as by non-linear contributions. The first term, which dominates the extratropical turbulent heat flux forcing, at least in Northern Hemisphere winter conditions (Cayan, 1992b), can be expressed as $F'_{turb} = \frac{-(Q'_s + Q'_l)}{\rho_0 c_p \overline{h_{ml}}} = \left(\frac{\rho^a c_s c_p^a \Delta T'}{\rho_0 c_p} \left(1 + \frac{1}{Bo}\right) \right) \overline{\Delta u}/\overline{h_{ml}}$. $\overline{\Delta u}/\overline{h_{ml}}$ is thus a direct proxy for its efficiency. For a typical midlatitude Bowen ratio $Bo = \frac{Q'_s}{Q'_l}$ of $1/3$ and $\Delta T' \sim 1$ K, the term within the brackets scales as $10^{-6} K \sim 1 \text{ K}/10 \text{ day s}^{-1}$. For these values, and if seasonal and spatial variations of $\Delta T'$ and Bo are relatively small, Fig. 4.23 reveals a basin-wide winter-time reduction in this contribution to the large-scale turbulent heat flux forcing by $\sim 0.1-0.2 \text{ K}/10 \text{ day}$, as well as more subtle regional spring-autumn differences. A complete assessment of the seasonal changes in atmospheric SSTA forcing would need to test the above assumptions from observations, take into account seasonality in air-sea temperature contrasts and, especially in the Southern Ocean, include Ekman forcing anomalies (which have a different spatial and seasonal modulation). This quick estimate suggests that seasonal changes in the intensity of large-scale atmospheric forcing are of (at least) comparable magnitude to those in mesoscale eddy stirring throughout basin interiors (i.e. away from major fronts).

Attribution to the seasonal modulation of SSTA spatial scales Seasonal change patterns of SSTA scales (Fig. 4.20) strikingly resemble to those of $\overline{\Delta u}/\overline{h_{ml}}$, the proxy for the atmospheric forcing intensity (Fig. 4.23), such that smaller scales coincide, in time and space, with weakened large-scale forcing. This suggests an important

role of atmospheric forcing in creating the seasonal modulation in the nature of SSTA. Indeed, Figure 4.24, shows that, in each individual season, minima in SSTA scales (colour mapped) emerge in the NA basin at the seasonally shifting position of zero wind stress observed in the subtropical transition from trades to westerlies (as revealed by the $(\overline{\mathbf{u}_a - \mathbf{u}_o})/\bar{h}_{ml}$ vectors, whose magnitude is a proxy for one component of the atmospheric forcing strength, as discussed above). Conversely, the largest-magnitude forcing vectors, observed over the shallow summer-time mixed-layers, as well as locally over each season's peaks in wind stress, coincide with the largest SSTA spatial scales.

Seasonal changes in the (rms) amplitude of SST variability, displayed in Figure 4.25, provide further insight into the respective roles of large-scale and mesoscale forcing processes in setting the observed seasonal modulation of SST variability. Where large-scale forcing plays a predominant role, seasonally enhanced (weakened) SSTA amplitudes are expected to coincide with a seasonal increase (reduction) in SSTA spatial scales. Regions, in which this is observed, are hatched in the Figure, and cover vast parts of ocean basins. They are characterized by a predominant winter-time reduction in amplitude and scale of SSTA (blue shades in Figs. 4.25 and 4.20), which is indeed observed to coincide, as of the $\overline{\Delta q}/\bar{h}_{ml}$ proxy, with a reduction in the winter-time large-scale turbulent heat flux forcing (blue shades in Fig. 4.23). In the remaining regions (un-hatched), the seasonal modulation of SST variability carries the signature of small-scale forcing (i.e. enhanced amplitudes over smaller spatial scales of SSTA, and vice versa). They are preferentially located over the main oceanic frontal zones and characterized by a winter- and spring-time intensification of SSTA amplitudes (red shades in Fig. 4.25), which coincides with the observed winter- and spring-time frontal intensification pattern of eddy stirring (red shades in Fig. 4.21).

The seasonal modulation, in both amplitude and scale, of the observed SST variability (on intra-seasonal and higher frequencies – cf. estimation from 90-day chunks) is thus well explained by a combination of seasonal changes in the intensity of eddy stirring and large-scale atmospheric forcing. In general the former predominate the observed modulation in frontal regions, the latter in basin interiors. (One pronounced exception, worth mentioning here, is the eddy stirring control observed in the relatively quiescent region equatorward of the ACC in its Indian sector, characterized by enhanced eddy stirring in summer and autumn when ACC fronts relax and spread into the quiescent interior.)

Further work and mesoscale resolving datasets of subsurface temperature variability

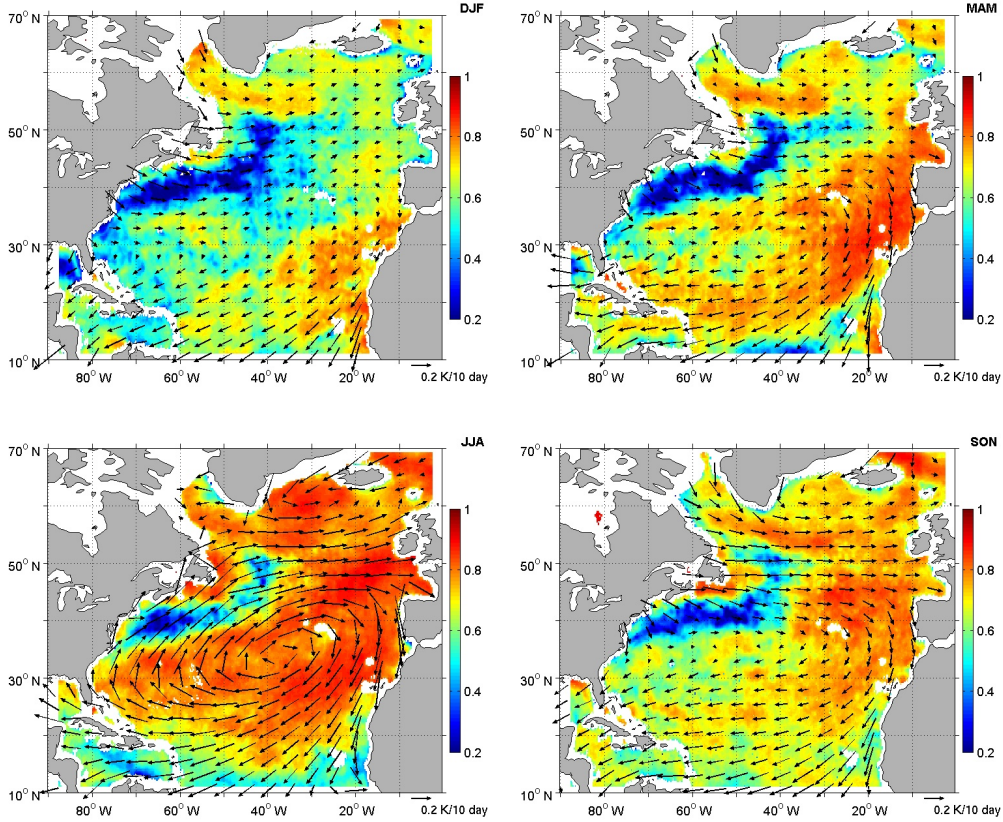


Fig. 4.24: The 4 seasons of observed SSTA spatial correlations, as in Fig. 4.18, mapped, for the NA, over $(\overline{u_a - u_o})/\bar{h}_{ml}$ vectors. The scale vector indicates an air-sea speed difference of 10 m/s acting over a mixed-layer depth of 50 m. As discussed in the text, this corresponds roughly to a turbulent heat flux forcing of 0.2 K/10 day .
(For reference, the NA average of the correlation between seasonal anomalies (i.e. deviations from the annual mean) in $|\overline{u_a - u_o}|/\bar{h}_{ml}$ and $c_{5^\circ, 2^\circ}(T')$, evaluated at each grid point, is ≈ 0.6 .)

(and mixed-layer depth) are required to assess whether, indeed and additionally to the above two mechanisms, climatological (and eddy-driven) entrainment forcing play an important role. Whether climatological entrainment, operational in autumn and winter (Fig. 4.22) acts as large or mesoscale SSTA forcing, depends on the predominant scales of subsurface temperature variability. For a substantial baroclinic contribution to observed mesoscale SSHA, their geographic distribution is expected to resemble to those of SSHA (Fig. 4.7c,d). Climatological entrainment would thus act to liken winter (and spring) SSTA scales (top panels of Figs. 4.18 and 4.19) to those of SSHA, which

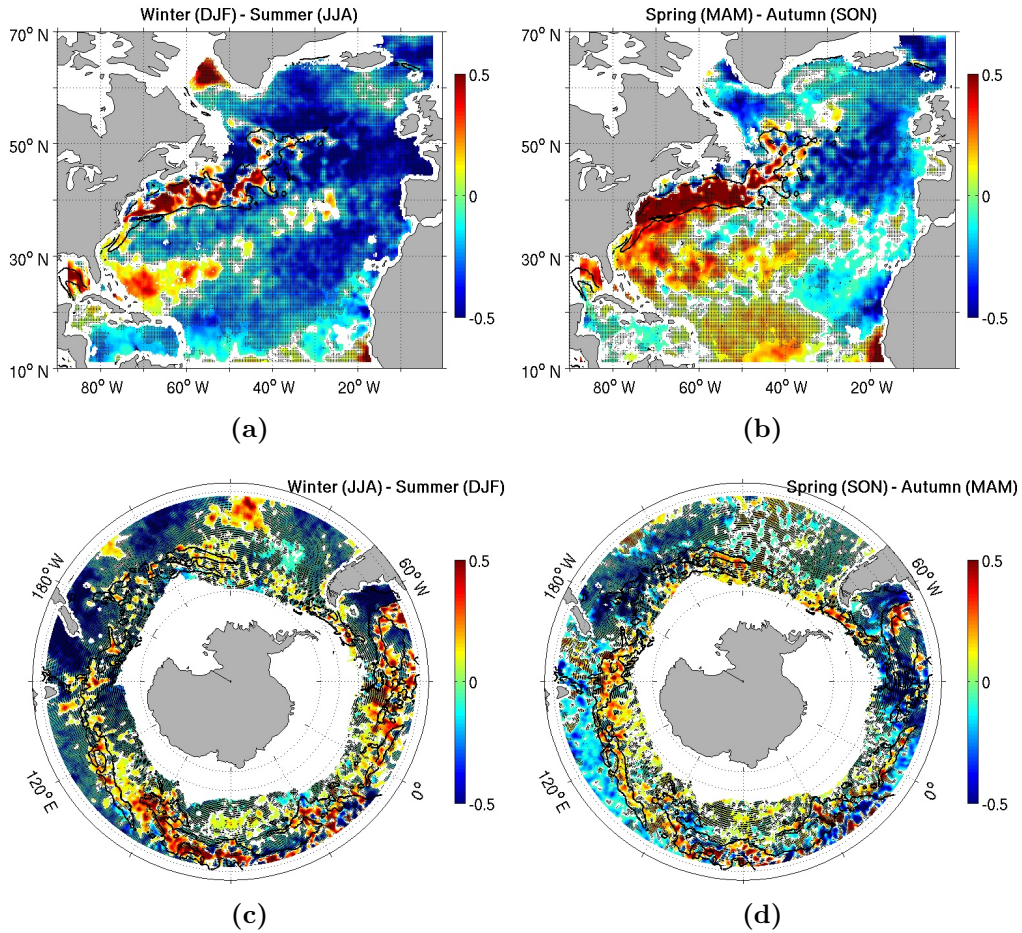


Fig. 4.25: Observed seasonal differences in SSTA rms (in K), mapped if they exceed 10% of the annual mean. In hatched areas, reduced (increased) rms coincide with reduced (increased) spatial scales of SSTA. For winter - summer (lhs), and spring - autumn (rhs panels). Time-mean $\sigma(\mathbf{u}'_g \cdot \nabla \bar{T}) = 1 \text{ K}/_{10 \text{ day}}$ contoured.

is indeed observed.

The small and intermediate scales that appear in SSTA in winter and spring could in principal also be forced by the atmosphere (through Ekman suction). However the fact that they mostly coincide with reduced (instead of enhanced) winds and, at the same time, with enhanced eddy stirring, supports their mesoscale nature⁸. This observed coincidence between reduced large-scale and enhanced mesoscale forcing is particularly apparent over the NA subtropical convergence zone in spring, and illustrates how the atmospheric circulation is operational in generating the observed (spring-time)

⁸ This would be best confirmed by examining seasonal changes in $\langle T'\zeta' \rangle$, which in contrast to $\langle T'\eta' \rangle$ are substantially less affected by large-scale steric effects.

intensification of oceanic SST fronts, and thus the enhanced eddy stirring intensities at the origin of the appearing small scales in SSTA.

The snapshots in Fig. 4.3 contrast intense small-scale SSTA observed over mesoscale SSHA features at the end of the entraining season (panels a, d) to large-scale SSTA prevailing even over the most intense GS and ACC eddies at the end of the shallowing season (panels b, c). This contrast is thus not a mere coincidence, but a typical illustration of the observed seasonal SSTA modulation. It likely reflects both, the reduced atmospheric forcing over the deepest mixed-layers, as well as the enhanced mesoscale forcing through the stirring of intensified end-of-winter-time SST fronts.

4.5 Discussion and Conclusions

Thanks to numerous dedicated field programs and continuously improving altimetric observation capacities, the presence of, often densely-packed, mesoscale current variability throughout almost the entire world oceans (Fig. 4.7c, d; McWilliams, 1976; Richman et al., 1977; Smith, 2007), its predominant contribution to the ocean's kinetic energy, as well as the pronounced increase in energy levels towards western boundary currents, and too a lesser extent, towards the path of the ACC (Fig. 3.9; Schmitz, 1984; Stammer, 1997a; Scott and Wang, 2005), are today well established.

Perturbations in SST on spatial scales much smaller than that of ocean basins and atmospheric synoptic disturbances, but instead comparable to oceanic mesoscales, have also been described by a vast number of studies, based on in-situ observations, beginning at least as early as 1940 (Iselin, 1940), as well as on IR radiometry, newly available in the 1970s and 1980s. Whereas a particular emphasis lay on the characterization from IR charts of propagating SSTA in the vicinity of the Gulf Stream, often identified as energetic “rings” spawned by the current, (e.g. Halliwell and Mooers, 1979; Richardson et al., 1979; Ring Group, 1981; Brown et al., 1986; Auer, 1987; Cornillon et al., 1989; Park et al., 2006), these ring-type SSTA have also been observed over energetic WBCs and their extensions in the Southern Ocean, in particular in association with the Agulhas current and its retroflexion (Lutjeharms and Valentine, 1988; van Aken et al., 2003) and the Brazil-current (Lentini et al., 2002). Also along the ACC itself mesoscale SSTA have been characterized, both in the form of frontal meanders (Moore et al., 1999) and isolated eddies (Joyce et al., 1981; Ansorge and Lutjeharms, 2005; Swart et al., 2008). Away from these major current systems, a range of studies characterized small scales in SSTA in regions of intermediate eddy variability, in the

Northern Hemisphere’s subtropical convergence zones (Voorhis et al., 1976; Halliwell et al., 1991a; Roemmich and Gilson, 2001) and in the central NA along the NA drift (Williams, 1988), but also in the quiescent eastern parts of ocean basins (Businger and Shaw, 1984; Bourras et al., 2004; Cipollini et al., 1997; Hill et al., 2000).

The advent of through-cloud microwave radiometry, available globally since the launch of AMSR-E in June 2002, renewed the interest in mesoscale SST signals, and led to the observational establishment of the presence of *quasi-permanent* mesoscale features in SST in all the major frontal zones across the globe (Small et al., 2008, for a review), including Kuroshio, GS, BMC and ARC (O’Neill et al., 2005; Liu et al., 2007). Apart from a notable exception from observations by the earlier TRMM MW radiometer limited to relatively low latitudes in the North Pacific (Qiu and Chen, 2005), the observational characterization of *transient* mesoscale features in MW-SSTA has so far received less attention.

However, due to the sparseness of in-situ observations, the connection to an eddy flow field of the observed transient SST anomalies, which are the focus of this study, could not always be made explicit. Indeed, the properties of ocean “rings” have thus often been studied solely from observed mixed-layer thermal anomalies assumed connected to subsurface eddying flows. Although observational coverage of surface mesoscale circulations dramatically increased with the advent of satellite altimetry, early attempts to link satellite-based SSH and SST variability have necessarily been limited by the low resolution capacities ($\geq 1^\circ$) of mono-mission altimetry and homogeneous infra-red datasets (e.g. Halliwell et al., 1991b; Cipollini et al., 1997; Leeuwenburgh and Stammer, 2001). The merged multi-altimeter datasets available today (cf. 3.2.3) start to adequately resolve the mesoscale. Their combination with MW-SSTA thus provides independent homogeneous-quality mesoscale-resolving observations of SST and SSH, the basic requirement to establish, as attempted in this chapter, observational measures that address two so far relatively poorly characterized aspects of transient eddy SSTA, namely **i)** their systematic association with eddying circulations, and **ii)** their contribution to the full variability in SST. Key results on these aspects are summarized and discussed in the next two paragraphs, which are followed by a brief conclusion.

i) Systematic transient eddy SSTA In-situ observations (and their combination to satellite SST) can traditionally only establish a connection between mesoscale SST and pressure anomalies for a few individual eddies. Whereas some in-situ sampled

eddies clearly show warm/cold cores extending through the thermocline up into the mixed-layer (van Ballegooyen et al., 1994; Gouretski and Danilov, 1994; Ansorge and Lutjeharms, 2005; Roemmich and Gilson, 2001; Bourras et al., 2004, and many GS ring studies), other studies report a quick loss of the eddy SST signal, especially for cyclones in summer (Auer, 1987; Morrow et al., 2004), or observe no clear/systematic SST expression over the eddies sampled (e.g. Joyce and Stalcup (1985) reviews warm or cold caps over GS anticyclones, Parker (1971) surveys 62 GS cold-core rings, from which not all had systematically cold SSTA, and in the eastern Sargasso sea McCartney et al. (1978) observes cyclones without any clear mixed-layer T' , but whose cold dome sharpens the thermocline at the mixed-layer base). Eddies with subsurface velocity maxima are also reported in the literature (e.g., using repeat ADCP transects from the Oleander project, Wei et al. (2008) sample two GS anticyclones sub-surface velocity maxima, and in a “Megapolygon-87” CTD section Maximenko et al. (2001) describe a Subarctic Front–Kuroshio Extension anticyclone (their Fig. 17), whose distinct warm-core extends through the thermocline to only 125 m below the surface), and these are observed in association, with either no significantly non-zero mixed-layer buoyancy anomaly, with opposing polarity T' signals, i.e. cold-top anticyclones, warm-top cyclones (reported e.g. by Simpson et al. (1984) for the California current, or by McGillicuddy et al. (2007) and Li and Hansell (2008) over cyclones and anticyclones (termed mode-water eddies) in the Sargasso Sea), or with salinity-controlled mixed-layer buoyancy anomalies (Joyce and Stalcup (1985) samples a warm-salty top GS anticyclone, Joyce et al. (1981) a cold-fresh top Drake passage cyclone). Williams (1988) linked their appearance to the air-sea damping of the eddy T' signal.

In contrast, eddy-identification from altimetry makes it possible (e.g. Roemmich and Gilson, 2001) to meet the additional observational requirement of a large sampling rate, which is necessary to isolate a statistically robust eddy signal in the mixed-layer. Here we take advantage of the large number of eddy tracks identified by the automated eddy tracking algorithm (chapter 3.2.4) in the regions considered here (≥ 1200 *per year*). By averaging SST fluctuations along them, a statistically robust and systematically non-zero signature of propagating mesoscale eddies in the SST field is revealed. It is characterized by warm-tops over anticyclones and cold-tops over cyclones (section 4.3.1,i).

This result is not limited to eddies originating in “energetic” regions, where high SSHa rms characterizes (cf. chapter 3.3.1) the enhanced levels of mesoscale variability associated with major current systems (GS, ARC, BMC and ACC). It equally applies

to “quiescent” oceanic interiors, studied here at the example of two regions in NA and SP eastern subtropical gyres (cf. chapters 3.3.2 & 3.3.3).

In energetic regions, the eddy SST signature, characterized by an observed amplitude of just above $\frac{1}{2}$ K, is easy to detect. It clearly shows up already in raw Eulerian statistics of SSTA and SSHA, which in these regions reveal elevated and positive simultaneous correlations between the two variables, only to be explained by the presence of mesoscale correlated signals (cf. section 4.2.1), and observed to be concentrated in relatively narrow bands along major currents’ SST fronts in which the eddy stirring of background isotherms is intense (cf. Figs. 4.4 and 4.9). Consistent with significant correlations found in comparable-resolution observations in the vicinity of the Kuroshio by Qiu and Chen (2005)⁹, such a localized occurrence of intense mesoscale eddy SST amplitudes could, unsurprisingly, not be detected in a previous global study from lower resolution observations (Leeuwenburgh and Stammer, 2001), which, instead, reported weaker correlations with a much broader geographical distribution. As shown in this chapter, the latter are partly attributable to Rossby-wave like features visible throughout ocean basins in both SSTA and SSHA (Figs. 4.8 & 4.11; Cipollini et al., 1997; Hill et al., 2000; Leeuwenburgh and Stammer, 2001), but also to a substantial part to large-scale steric effects.

Detecting the weak SST signature (≈ 0.1 K) of mesoscale circulations in quiescent parts of the ocean is more challenging. Here it is often masked by more intense (≈ 0.5 K) large-scale atmospherically forced SST variability, which through correlated thermosteric SSHA dominates Eulerian statistics (Figs. 4.4 & 4.5) and the signal in each individual eddy snapshot, and still to a large degree in the average over the lifetime of each eddy track (Fig. 4.15b). Thus, away from major frontal regions, a substantial sampling (i.e. following many eddies for a long time, important to take into account for in-situ studies) is needed to isolated a robust eddy SST signal, unless a form of spatial filtering is applied.

In energetic regions differences between eddies or eddy time variability are observed to contribute to zero-order to the spread in SSTA between eddy snapshots (and will be examined in more detail in chapter 5). In quiescent regions, to the contrary, such contributions are not discernible (Figs. 4.12 & 4.13).

Additional spread in SSTA between eddy snapshots is induced by observation and processing errors. In particular, due to the short time-series ($n = 5.5$ years), a sub-

⁹ Also reported by White and Annis (2003) for several current systems, but from an analysis based on extremely short time-series and spatial filtering on a pre-set mesoscale window.

stantial contribution derives from random errors in the removed, empirically estimated seasonal background state $\overline{T}$, as discussed in Appendix A.3. There it is moreover shown that, following an eddy, a fraction $\frac{1}{n}$ of its SST perturbation is systematically retained and removed with $\overline{T}$, which induces a systematic *low* bias in estimated average eddy anomaly (at $\frac{n-1}{n}$ of its true observed value). The statistically significant average eddy SST signature is thus a robust result of the observations analysed here.

Whether the result that in both energetic and quiescent regions the average satellite-tracked eddy is a warm-top anticyclone, or a cold-top cyclone, implies an average dynamical nature of eddies, or whether it is the result of a mixture of differing eddy dynamics and mechanisms of SSTA generation remains to be assessed. For example, dynamical constraints on the eddy T' signal are unclear if salinity anomalies play an important role in shaping the eddy mixed-layer density perturbation, and thus specifically at high latitudes. In midlatitudes, no systematic association between eddy pressure and SST anomalies is a priori expected for barotropic eddies, and an opposing association compared to that observed for eddies with subsurface velocity maxima. Warm/cold perturbations in SST over anticyclones/cyclones are only expected as dynamic part of surface intensified baroclinic eddies in midlatitudes if their deep warm/cold cores in thermal wind balance with the geostrophic eddy flow extend up to the surface. In this sense, the observed amplitude ratio of eddy perturbations in SST and SSH of $\approx 0.3 - 0.4K/10cm$ (Table 4.2) is consistent, from a broad scaling, with baroclinic eddy T' cores extending throughout the thermocline. More specifically, a systematic association between eddy SST and pressure anomalies is only expected, if the subsurface baroclinic shear extends to the surface, but not if momentum is homogenized in the mixed-layer. If the latter were to be the case, salinity anomalies would have to compensate the temperature-induced eddy buoyancy anomalies, and it would be fascinating to assess this using in-situ observations collected by the growing network of ARGO floats.

ii) The transient mesoscale contribution to SST variability Whereas, by averaging SSTA over a large-number of propagating eddies tracked by CSS11's algorithm, a robust eddy SST signal is detected throughout ocean basins, Eulerian statistics of SST and SSH observations (section 4.2) reveal a stark contrast in the nature of SST variability between basin interiors and major frontal zones.

In basin interiors, large-scale processes are found to predominate the variability in SST, thereby masking the weak eddy SST signal. A zero-order contribution of transient

mesoscale motions to SST variability is therefore limited to energetic regions, in which eddies are intense and efficiently generate mixed-layer heat content anomalies through geostrophic stirring of the background SST fronts.

This supports eddy stirring as main generation mechanism of mesoscale SSTA¹⁰, which thus acts to localize the impact of the transient mesoscale to regions in which also time-mean SST changes occur on mesoscales. Such an important role of surface eddy stirring in creating a geographically inhomogeneous distribution of the “eddy noise” in SST has been previously suggested (Voorhis et al., 1976; Frankignoul, 1985). Although the striking localization of a mesoscale nature of SST variability, reported here, could, as mentioned above, not be diagnosed from previous lower resolution global studies, enhanced mesoscale SST variability near time-mean fronts is supported, amongst others, through the localization of observed rings in IR-SST in the vicinity of the Gulf Stream, and locally by Halliwell and Mooers (1979) for the shelf-slope front coast-ward of the GS, and by Voorhis et al. (1976) and Halliwell et al. (1991b) for the NA subtropical convergence zone (captured by reduced SSTA spatial scales in the observations analysed here (Fig. 4.7a, 60-70°W, 28°N), the latter zone is characterized by a much weaker mesoscale contribution to SST variability compared to e.g. the Gulf Stream fronts).

Precluded by the sparseness of in-situ data before satellite and ARGO eras, the SST variability of the Southern Ocean has only been started to be analysed fairly recently. Here, studies from low resolution datasets reported large-scale atmospherically forced SST variability (cf. chapter 2.2.3) even along the path of the ACC (Verdy et al., 2006; Vivier et al., 2010). Only analyses of higher-resolution MW-SST (Dong et al., 2006b) and in-situ XBT data south of Tasmania (Sokolov and Rintoul, 2003) additionally revealed an important role of oceanic dynamics, through frontal displacements (and meandering), on large-scale SSTA along the ACC path. The analysis of MW-SST and multi-altimeter SSH, presented here, suggests that large-scale processes and small scales dynamically associated with transient mesoscale eddy forcing jointly shape the variability in SST along the ACC path.

This result applies to all, wave-like, nonlinear as well as meandering, types of transient mesoscale circulation phenomena. It is based on a diagnostic of spatial cor-

¹⁰ Although this chapter only considered variations in SST forcing mechanisms, spatial and temporal variations in the intensity with which SST anomalies are damped may additionally contribute to the observed geography and seasonality of the mesoscale SST contribution, but, as further discussed in chapter 5, are challenging to assess from observations.

relations, developed in section 4.2.2, which has been revealed to be a very useful and straightforward tool to visualize the geographic distribution of the contribution of different spatial scales to the variability in a variable (see Fig. 4.7 for SST and SSH). By itself the diagnostic can only attest a zero-order contribution to the variability by “small” spatial scales (~ 100 km). Although these may in principle be atmospherically forced by storms and hurricanes (e.g. Frankignoul, 1985; Monaldo et al., 1997; Emanuel, 2001) and wind-island interactions (Sangrè et al., 2007), their dynamic association with transient mesoscale processes is confirmed by the analysis of the observed SSTA-SSHA relationship at fixed-location (e.g. Fig. 4.10 or Figs. 4.4 & 4.5), as well as following eddy tracks (Fig. 4.15a). As the relatively large footprint (~ 50 km) of the AMSR-E radiometer is likely to partly smooth out mesoscale variability, especially at high latitude, the application of this spatial-correlation diagnostic to higher-resolution global SST datasets, available from IR observations only (but with cloud gaps) and as gap-free blends of IR and MW observations (but with inhomogeneous effective resolution capacity), would be fascinating (but more challenging), and might potentially reveal higher amplitude mesoscale signals, whose contribution to the full variability in SST might be even more substantial in major current regions and might extend further into oceanic interiors than reported here.

The spatial-correlation diagnostic moreover reveals that the observed zero-order mesoscale contribution to SST variability in major frontal zones (blue shades in Fig. 4.7a,b) is not limited to time scales of ≈ 100 days, but remains important at lower frequencies (Fig. 4.17), where it is even more pronounced. It is present throughout the year (Fig. 4.18&4.19), slightly seasonally pulsing in intensity and spatial extent together with the intensity of background SST fronts and thus that of the eddy isotherm stirring (as suggested by Frankignoul, 1985 and shown in Fig. 4.21).

Across ocean basins the nature of SST variability shows a significant seasonal modulation, characterized by the summer-time predominance of large spatial scales and the appearance of smaller scales where atmospheric forcing efficiencies are reduced, i.e. over the deep winter-time mixed-layers and, moreover, following the seasonally migrating westerly-to-trade transition (zero wind stress) in the NA. The enhanced large-scale summer-time SST variability reported previously along the ACC (Ciasto and Thompson, 2008, and Verdy et al., 2006 specifically for the mid-Pacific) is shown here to represent only a fraction of the full SST variance, which along the ACC path (and specifically over the strong fronts in the ridge-system of the Pacific-Antarctic Rise) is dominated by the mesoscale contribution and thus peaks together with frontal

strength in winter and spring. The observed seasonal modulation of the nature of SST variability suggests that the atmosphere is instrumental, both through its direct impact on mixed-layer depths and its indirect impact on frontal intensity, in determining where mesoscale SST perturbations are large enough to significantly contribute to the variability in SST and thus to the variability at the atmosphere's lower thermal boundary.

Now that both observations of T' and η' start to accurately resolve the mesoscale, transient mesoscale-resolving datasets of subsurface temperature variability and mixed-layer depths are the missing pieces in the puzzle of closing the mesoscale mixed-layer heat budget and further examining the mechanisms that set the nature of SST variability (and, especially, to assess whether eddy-driven $w'\Delta\bar{T}$ (Stevenson, 1983) and climatological entrainment $\bar{w}_{entr}\Delta T'$ (Williams, 1988), shown previously from scattered in-situ observations in the NA to importantly impact mesoscale T' signals, are important contributors to the observed seasonal modulation of SSTA spatial scales).

Summary of Conclusions In this chapter, to assess whether the mesoscale eddy field is associated with systematic signals in SST, the observed relationship between the temporal variability in through-cloud MW SST and multi-altimeter SSH has been analysed both at fixed-location and directly following nonlinear eddy tracks (established by CSS11). Key results are:

- Propagating eddies leave a systematic imprint on observed SST. This is not restricted to the regions of high SSHA rms associated with major current systems, as a robust eddy SST signature is also established for quiet ocean interiors, but detectable there only from large observational samples.
- In quiet interiors, observed SST variability is dominated by large-scale forcing, that masks the eddy SST signature at fixed-location. Large-scale forcing also contributes to zero-order in regions of high SSHA rms, where a leading mesoscale contribution to SST variability only peaks along narrow bands of intense eddy stirring.

This suggests that the potential for two-way eddy-atmosphere interactions in mid-latitudes is indeed primarily limited to the major frontal zones, where it has been previously observed over mesoscale SSTA mainly in the form of quasi-steady meanders (see Chelton and Xie (2010) for a review), but also in the form of isolated propagating features (Park et al., 2006). Only here the oceanic mesoscale is instrumental in shaping the variability in the atmosphere's lower thermal boundary, and does so throughout the year. Elsewhere the atmosphere predominately sees the result of its own variability.

Although too small to make a zero-order contribution to SST variability, the robust mesoscale SST signals detected in basin interiors might still be visible to the MABL and lead to coupling, as observed from isolated in-situ observations in the North-East NA (by Businger and Shaw, 1984 and in an exceptionally well documented sample by Bourras et al., 2004). This however would be very challenging to systematically detect in the midst of the much larger-amplitude large-scale passively forced variability.

The application of the spatial SSTA autocorrelation diagnostic (Eq. 4.2 and Fig. 4.7), which allows to simply visualize the contribution of different spatial scales to the variability in SST, to coupled general circulation models on their way to resolve eddies, would be a straightforward first tool to compare to observations and further the realism of the modelled representation of mesoscale processes in SST and ocean-MABL coupling.

Chapter 5

TOWARDS UNDERSTANDING THE OBSERVED PERSISTENCE OF MESOSCALE SSTA

This chapter characterizes the persistence in time of the systematic transient eddy SST signals uncovered in the previous chapter, providing an analysis both at fixed-location (section 5.2) and following eddy tracks (section 5.3). This is relevant as it determines the time scale, over which mesoscale SSTA can potentially interact with the atmosphere.

Having singled out geostrophic eddy stirring in the mixed-layer as key forcing mechanism of mesoscale SSTA in the previous chapter, here we examine the mechanisms that control their persistence once generated, and, importantly, elucidate the role played in this by air-sea interactions on the mesoscale. The latter determine the strength of mesoscale air-sea heat exchanges, which, as outlined below, is a key parameter in setting the role of the mesoscale in the climate system, but is nonetheless so far only poorly quantified from observations. Constraints on its strength are presented in section 5.1.

Section 5.4 concludes.

5.1 Constraints on the mesoscale air-sea feedback strength

5.1.1 Introduction to the air-sea feedback

Air-sea heat fluxes play a key role in the generation of SSTA (cf. section 2.2), especially on the large spatial scales of atmospheric synoptic disturbances (referred to as “large-scale” in the following). However, they also respond to and thereby feed back on already existing SSTA (at all spatial scales). The strength of this feedback between

anomalies in air-sea heat fluxes Q (positive upward) and SST is given by

$$\alpha = \left. \frac{\partial \langle Q' \rangle}{\partial T'} \right|_{\bar{T}}, \quad (5.1)$$

where primes indicate departures from the unperturbed (seasonal) background state (denoted by over-bars) and $\langle Q' \rangle$ the systematic heat flux response to a given perturbation in SST, which can be isolated from (essentially random) synoptic variability by composite averaging over a large range of different atmospheric conditions. Positive values of α (increased heat loss over warm SSTA/anomalous heat gain over cold SSTA) correspond to a negative heat flux-SST feedback, i.e. a damping of SSTA by air-sea heat exchanges.

Previous basin-wide observational estimates of α are based on gridded datasets of spatially scattered ship observations (such as the Comprehensive Ocean-Atmosphere Data Set, COADS) or atmospheric reanalyses (outputs of atmospheric weather prediction models constrained to past observations, such as the NCEP/NCAR reanalysis). They thus only inform on the feedback strength on spatial scales larger than $\mathcal{O}(1000 \text{ km})$, and largely exclude the Southern Ocean. Estimating α on these scales relies on the separation between feedback and forcing contributions to Q , which can be achieved by analysing the lagged covariance between time series of Q and T , a method, based on ideas from stochastic climate models, identified by Frankignoul et al. (1998). Resulting estimates indicate a net negative air-sea feedback¹, which is dominated by turbulent rather than radiative fluxes and amounts to 20–40 W/m²K in midlatitudes (Frankignoul and Kestenare, 2002a, denoted FK02 in the following). Latent heat flux anomalies, due to increased evaporation over warmer water, dominate this feedback and its increase in fall and winter, but especially at higher latitudes sensible heat flux anomalies also play an important role. The net longwave radiation emitted by the sea surface also depends on temperature, but only weakly so². Additionally, SST-induced changes in atmospheric water vapour content and cloud properties/coverage may indirectly contribute to the longwave feedback³, and, as incident shortwave radiation is

¹ although locally at high latitude and in the tropics, in individual seasons and in individual components of it, the feedback may be positive.

² The longwave radiative feedback is given by $\alpha_{LW} = \partial \langle (\epsilon \sigma T^4 - \epsilon^a \sigma T^{a4})' \rangle / \partial T$, where $\sigma = 5.73 \cdot 10^{-8} \text{ W/m}^2 \text{ K}^{-4}$ is the Stefan-Boltzmann constant, $\epsilon \leq 1$ the emissivity of sea water, and T^a the temperature of the surface atmosphere, whose effective emissivity ϵ^a parametrizes the net effect of water vapour and clouds. α_{LW} reaches a maximum of $4\epsilon\sigma\bar{T}^3$ ($< 5 \text{ W/m}^2\text{K}$, at 15°C), when the atmospheric temperature does not adjust to SSTA ($\partial \langle T^{a'} \rangle / \partial T' = 0$).

³ via $-\sigma\bar{T}^a \frac{\partial \langle \epsilon^{a'} \rangle}{\partial T'}$

not directly dependent on SST, any shortwave radiative feedback purely relies on such indirect effects, which are currently not well understood. Observations of radiative fluxes usually rely on parametrizations, and whether (Park et al., 2005) or not (FK02) they incorporate satellite observations of clouds, they indeed yield a very weak net radiative feedback (of about ± 0 to $5 \text{ W/m}^2\text{K}$, changing sign from negative in winter to positive in summer and spring).

SSTA that span the globe or entire ocean basins are estimated to be weakly damped, at a rate of $\sim 3 \text{ W/m}^2\text{K}$ (Haney, 1971; Rahmstorf and Willebrand, 1995), with comparable contributions by radiative and turbulent fluxes. For the predominant basin-scale modes of SST variability (that show di-, tri- or quadrupoles within a basin), the air-sea feedback strength is estimated at $\mathcal{O}(10) \text{ W/m}^2\text{K}$ (by FK02), and is already, apart from equatorial regions, observed to be predominately provided by the turbulent heat flux response (which typically exceeds the radiative response by an order of magnitude). As the spatial scales of SSTA decrease further, the turbulent feedback is observed to systematically intensify, by typically $5\text{-}10 \text{ W/m}^2\text{K}$ from basin- to synoptic scales (FK02).

To understand this scale-dependence of the turbulent feedback, consider that a negative air-sea feedback restores the surface atmosphere to the underlying SSTA, thereby weakening the local air-sea temperature and humidity contrasts at its origin. The strength of the turbulent feedback thus depends on how quickly the atmosphere regenerates these contrasts by advecting away the induced temperature and moisture perturbations. Conceivably, the efficiency of this process should increase as (horizontal and vertical) spatial scales of the perturbations decrease. Over larger-scale SSTA, atmospheric temperatures will thus adjust more strongly to SSTA ($\partial \langle T^{a'} \rangle / \partial T' > 0$), leading to a reduced strength of the air-sea damping (Bretherton, 1982; Frankignoul, 1985; Barsugli and Battisti, 1998), with the strongest adjustment/weakest damping expected for basin-wide SSTA, over which only continents are available for the atmospheric heat export. Following this idea, Rahmstorf and Willebrand (1995) represent the atmospheric advective heat transport as a diffusive, i.e. inversely scale-dependent process in their simple atmospheric energy balance model, leading to a predicted increase in α from its global minimum to $\sim 50 \text{ W/m}^2\text{K}$ on scales of $\sim 400 \text{ km}$. From similar theoretical considerations, Bretherton (1982) expects a feedback strength of $\sim 100 \text{ W/m}^2\text{K}$ on scales of $\mathcal{O}(10) \text{ km}$.

Both the heating perturbations induced by the damping of underlying SSTA, complicated by the possible non-local release of latent heat, as well as the atmospheric

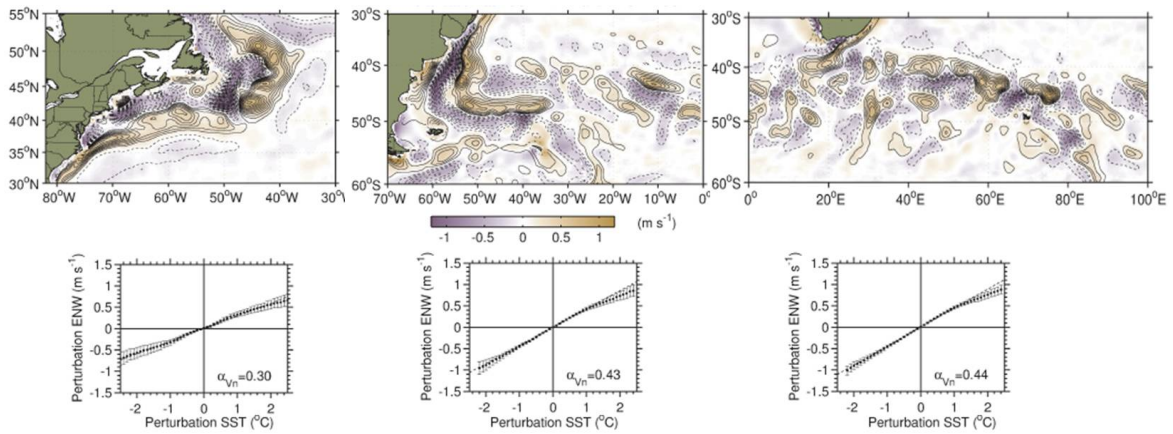


Fig. 5.1: The association between spatially high-pass filtered AMSR-E SST and QuikSCAT scatterometer-derived equivalent wind speed⁵, observed over GS (lhs), BMC (middle) and ARC (rhs). Top panels map observed 7-year averages (SSTA contoured every 0.25 K, negative dashed), bottom panels scatter observed monthly anomalies (as averages over bins of SSTA, with the indicated 95% confidence ranges based on a t-test, and the slopes on least-squares fits). From O'Neill et al. (2012).

response to them, have been and continue to be the focus of many modelling (reviewed by Frankignoul, 1985; Kushnir et al., 2002, and more recently e.g. Ferrari et al., 2008; Maze et al., 2011) and observational studies (e.g. Czaja and Frankignoul, 2002; Frankignoul and Sennéchaël, 2007). Large-scale extratropical SSTA are found to induce not only a thermodynamical ($\partial \langle T^{a'} \rangle / \partial T'$), but also a (weak) dynamical response ($\partial \langle u^{a'} \rangle / \partial T'$), which is however strongly dependent on the background atmospheric state and often masked by larger intrinsic atmospheric variability. Whether this weak dynamical atmospheric response to large-scale SSTA systematically modifies the strength of the observed negative air-sea feedback is unclear. Instead, pronounced negative correlations between large-scale u'_a and T' (observed simultaneously and when the atmosphere leads) are the signature of large-scale atmospheric heat flux (and Ekman) forcing of SSTA (e.g. Kushnir et al., 2002). They stand in stark contrast to recently discovered (e.g. Chelton et al., 2004; Small et al., 2008) systematic positive correlations between anomalies in wind stress magnitude and SST on the mesoscale, which are the signature of mesoscale SSTA forcing the MABL. Such small-scale anomalies,

⁵ Scatterometer-derived equivalent wind speed is effectively a measure for the air-sea speed difference if the atmosphere were neutrally stratified. It thus differs from the observed 10-m wind speed due to both departures from neutral stratification and the presence of ocean currents, but O'Neill

characterized by enhanced (reduced) wind speed⁵ over warm (cold) mesoscale SSTA, are clearly visible in the examples displayed in Fig. 5.1. They act to increase (reduce) turbulent oceanic heat loss, leading to a mesoscale modulation in air-sea heat fluxes that tends to damp the SSTA at their origin. Together with the expected reduced atmospheric thermal adjustment, this observed mesoscale dynamical coupling could thus lead to a significantly larger air-sea feedback on the oceanic mesoscale compared to that observed on the large scale (by FK02).

Including the observed scale-dependence of the air-sea feedback into the boundary condition of (large-scale) uncoupled ocean- or atmosphere-only models, has proved crucial for a realistic representation of the ocean's (thermo-haline) circulation (Rahmstorf and Willebrand, 1995) and the atmospheric variability (Barsugli and Battisti, 1998). On the mesoscale, however, the air-sea feedback strength is so far poorly constrained from observations. As it provides the surface thermal boundary condition for high-resolution ocean-only integrations, this represents an important limitation for the realism of mesoscale modelling. In particular, the mesoscale feedback strength is a key parameter, which not only determines (in part) the amplitude and persistence of eddy SST anomalies, and thus the scope for a significant transient ocean-atmosphere coupling in midlatitudes (with its influences on both ocean and atmosphere, see e.g. Spall, 2007a; Hogg et al., 2009; Small et al., 2008), but moreover controls eddy-driven water mass formation (Cerovecki and Marshall, 2008), as well as magnitude and nature of mesoscale poleward heat transports. The latter aspect is underlined by eddy-permitting models, in which, consistent with predictions from early theoretical tracer transport studies (Plumb and Mahlman, 1987; Plumb, 1979), fronts are observed to broaden (Zhai and Greatbatch, 2006b) and eddy heat fluxes become irreversible (Drijfhout, 1994a,b) as the air-sea damping is increased. The implied enhancement of effective eddy diffusivities in the presence of mesoscale mixed-layer heat flux-SST feedbacks is clearly demonstrated in Fig. 5.2: As illustrated in the rhs schematic in Fig. 5.2b, eddy stirring generates mesoscale anomalies and enhanced gradients in the SST field, which are efficiently diffused away by small-scale mixing, and eventually molecular diffusion, thus leading to much larger eddy diffusivities κ ($\mathcal{O}(1000\text{m}^2\text{s}^{-1})$) than in the absence of eddies. The middle schematic in Fig. 5.2b illustrates how the additional source of irreversibility provided by mesoscale air-sea damping leads to a faster dissipation of mesoscale SSTA than in the presence of molecular dispersion alone, and

(2012) show that its response to SST typically differs from that of the actual wind speed by only 10-30%. See also discussion in section 5.1.3 “Dynamical coupling” (and Appendix B.3).

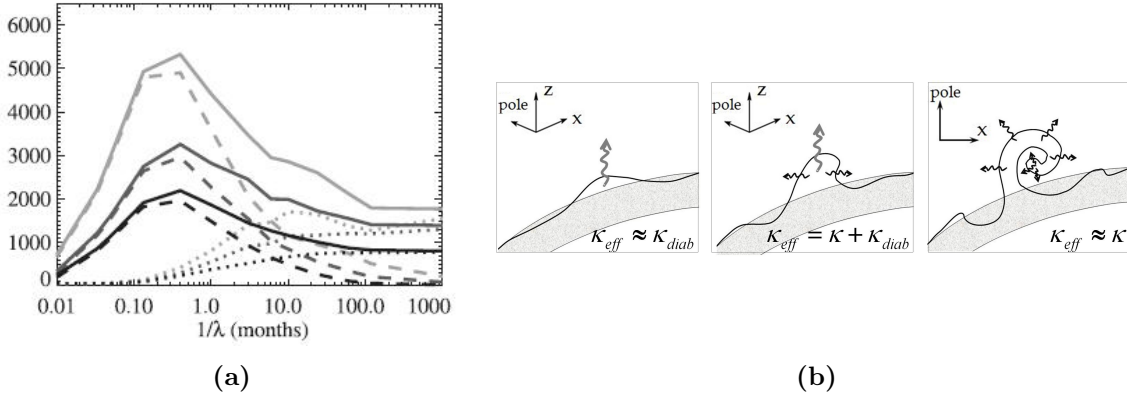


Fig. 5.2: Impact of the mesoscale mixed-layer heat flux damping on effective eddy diffusivities. **a)** κ_{eff} (solid) = κ (dotted) + κ_{diab} (dashed), in $m^2 s^{-1}$, as function of the damping time-scale $1/\lambda$, resulting from advecting a streamwise-average SST field with altimetry-derived velocities in a $1/20^\circ$ integration (Shuckburgh et al., 2011), for three different SO streamline bands (black: ACC core, dark: equatorward flank/ARC, light: ARC/Agulhas). **b)** Schematic illustrating how background isotherms (thin curve, grey shading) strained by eddy stirring (thick curve) result in mesoscale SSTA, which are damped by small-scale mixing/molecular diffusion (black arrows) and SST-induced mixed-layer heat loss (grey arrow), in very fast ($1/\lambda = \mathcal{O}(\text{days})$): lhs), intermediate ($1/\lambda = \mathcal{O}(\text{months})$): middle) and very weak damping regimes ($1/\lambda = \mathcal{O}(\text{years})$): rhs).

thus to an even larger effective eddy heat diffusion down the background mixed-layer SST gradient ($\kappa_{eff} = \kappa + \kappa_{diab} > \kappa$). Faster damping (decreasing abscissa in Fig. 5.2a) therefore enhances κ_{eff} , even though the contribution by κ alone decreases once (as illustrated in the lhs schematic in Fig. 5.2b) mesoscale SSTA and associated gradients are damped quicker than they are stirred. As seen in Fig. 5.2a, the increase in κ_{eff} induced by air-sea damping can reach more than a factor of 2, but is bounded by an upper threshold, beyond which diffusivities drop to zero and which is reached as the damping becomes too fast to allow eddies to generate SSTA.

From the above arguments it is clear that a robust observational estimate of the scale-dependence of the air-sea feedback and its strength on the mesoscale is crucial, to further our understanding and realistic modelling of the oceanic circulation, its response to external forcing, and the role of the oceanic mesoscale therein. In the remainder of this section, we use air-sea climatologies, as well as previous direct observations of mesoscale air-sea interactions, to develop constraints for the predicted

Chapter 5. Towards understanding the observed persistence of mesoscale SSTA

increase of the air-sea damping strength from large-scales (section 5.1.2) to mesoscales (section 5.1.3), which, after a discussion of the associated air-sea feedback time scales (section 5.1.4), are summarized in section 5.1.5.

5.1.2 Climatological bounds for the air-sea feedback on large scales

Given the above discussion, here we focus on the (dominant) turbulent contribution to the air-sea feedback. Turbulent air-sea heat fluxes (sensible Q_S and latent Q_L) can be expressed through bulk formulae, which parametrize the boundary layer turbulence giving rise to the transfer of heat and moisture from the air-sea interface through the MABL in terms of air-sea contrasts and the (larger scale) flow, via empirically estimated bulk coefficients (c_S for sensible, c_L for latent heat transfer):

$$\begin{aligned} Q_S &= \rho^a u^a c_S c_p^a (T - T^a) \\ Q_L &= \rho^a u^a c_L L (q_{sat}(T) - q(T^a)), \end{aligned} \quad (5.2)$$

in which ρ^a, c_p^a, T^a and q^a are density, specific heat capacity, temperature and specific humidity of the surface atmosphere (usually evaluated 10 m above sea-level), T and q_{sat} are temperature and specific humidity of the (saturated) ocean surface, L is the latent heat of evaporation, and $u^a \equiv |\mathbf{u}^a - \mathbf{u}|$ is the wind speed with respect to the moving ocean surface (with $\mathbf{u}^a$ & $\mathbf{u}$ denoting surface vector wind and current).

Using (5.2) in (5.1), allows evaluation of the turbulent heat flux feedback for a given air-sea state. It results partly from a dynamic adjustment of the boundary layer winds to SSTA (through $\partial \langle u^a \rangle / \partial T'$), partly from a purely thermodynamic feedback. By setting the former to zero, expressions for the latter are derived⁶. The sensible contribution to the thermodynamic feedback is given by:

$$\alpha_S \equiv \left. \frac{\partial \langle Q'_S \rangle}{\partial T'} \right|_{\bar{T}} = \rho^a \overline{u^a} c_S c_p^a \left. \frac{\partial \langle (T - T^a)' \rangle}{\partial T'} \right|_{\bar{T}} = \rho^a \overline{u^a} c_S c_p^a \left(1 - \frac{\partial \langle T^a \rangle}{\partial T'} \right). \quad (5.3)$$

Similarly, noting that $q(T^a) \approx r_H q_{sat}(T^a)$, where $r_H \equiv \frac{e}{e_{sat}} \approx \frac{q}{q_{sat}}$ is the relative humidity (and e the water vapour pressure), the latent thermodynamic feedback becomes:

$$\alpha_L \equiv \left. \frac{\partial \langle Q'_L \rangle}{\partial T'} \right|_{\bar{T}} = \rho^a \overline{u^a} c_L L \left. \frac{\partial \langle (q_{sat}(T) - q(T^a))' \rangle}{\partial T'} \right|_{\bar{T}} = \rho^a \overline{u^a} c_L L \left(\left. \frac{dq_{sat}}{dT} \right|_{\bar{T}} - \overline{r_H} \left. \frac{dq_{sat}}{dT} \right|_{\bar{T}^a} \frac{\partial \langle T^a \rangle}{\partial T'} \right), \quad (5.4)$$

in which $\partial \langle r'_H \rangle / \partial T' = 0$, i.e. the atmospheric moisture response $\langle q^a \rangle$ is assumed to be purely temperature controlled (and given by $\overline{r_H} \langle q_{sat}^a \rangle$)⁷.

⁶ For $\partial \langle T^a \rangle / \partial T' < 1$, the SST-induced modification of sea-air contrasts also modifies transfer coefficients via their empirically established (Large et al., 1994; Fairall et al., 2003) increase with decreasing MABL stability. This induces a weak additional contribution enhancing the negative feedback, which is neglected here.

⁷ A systematic adjustment of r_H would induce the additional term $-\partial \langle r'_H \rangle / \partial T' q_{sat}(\bar{T}^a)$ in (5.4).

From these expressions it is clear that the thermodynamic feedback crucially depends on $\partial \langle T^a \rangle / \partial T'$, i.e. on the degree of the atmospheric temperature adjustment to the underlying SSTA, which, as discussed above, is expected to increase with the spatial scale of the anomaly. By assuming in turn complete and no atmospheric thermal adjustment, it is thus possible to estimate lower and upper benchmarks on the thermodynamic feedback strength. Setting $\partial \langle T^a \rangle / \partial T'$ to 0 in (5.3) & (5.4), yields the upper bound (denoted α_{ref}):

$$\alpha_{ref} = \alpha_{Sref} + \alpha_{Lref} = \rho^a \overline{u^a} \left(c_S c_p^a + c_L L \frac{dq_{sat}}{dT} \Big|_{\overline{T}} \right). \quad (5.5)$$

Conversely, using $\partial \langle T^a \rangle / \partial T' = 1$ in (5.3) & (5.4), the lower bound is obtained:

$$\alpha_{lower} = \alpha_{Slower} + \alpha_{Llower} = 0 + \rho^a \overline{u^a} c_L L \left(\frac{dq_{sat}}{dT} \Big|_{\overline{T}} - \overline{r_H} \frac{dq_{sat}}{dT} \Big|_{\overline{T^a}} \right). \quad (5.6)$$

As, in the latter case, atmospheric and ocean temperature anomalies are fully adjusted, there is no sensible contribution to the feedback and a latent heat flux response purely arises from the desaturation of the atmosphere and, if present, the background air-sea temperature difference.

Estimation from air-sea climatology The thermodynamic benchmarks (5.5) & (5.6) can be considered as bounds on the observed turbulent feedback on the large scale, where, following the discussion in section 5.1.1, the dynamic adjustment is expected to be weak. They can be directly estimated from air-sea climatology (as they are independent of anomalies), which previously has only been done for zonal averages (Haney, 1971) or a single typical midlatitude condition (Frankignoul et al., 1998). Here we establish their observed strength and spatial distribution for NA and SO, using a large-scale objective analysis of the COADS dataset (da Silva et al., 1994, see Appendix B.2 for details). To do so, we set $c_p^a = 10^3 \text{ J K}^{-1} \text{ kg}^{-1}$, $L = 2.5 \cdot 10^6 \text{ J kg}^{-1}$, and use an integrated form of the Clausius-Clapeyron relation (cf. Appendix B.2) to compute the strongly temperature-dependent $q_{sat}(T, p)$. We furthermore take typical midlatitude values for air-density ρ^a as 1.22 kg m^{-3} , for sea-level pressure p as 1015 hPa, and for the wind speed and stability dependent c_S & c_L as $1.15 \cdot 10^{-3}$ (as recommended by Fairall et al., 2003), and approximate the background air-sea speed difference with the surface wind speed climatology. Whereas including climatological variations of the former does not strongly modify results⁸, it would be interesting to further examine how strongly a relaxation of the latter assumption would.

⁸ They act to weakly increase α over (i) high-latitude low pressure zones due to a weak increase in $\frac{dq_{sat}}{dT}(T, p)$, (ii) the strong air-sea contrast of equatorward WBCs due to higher c_S, c_L , where stability

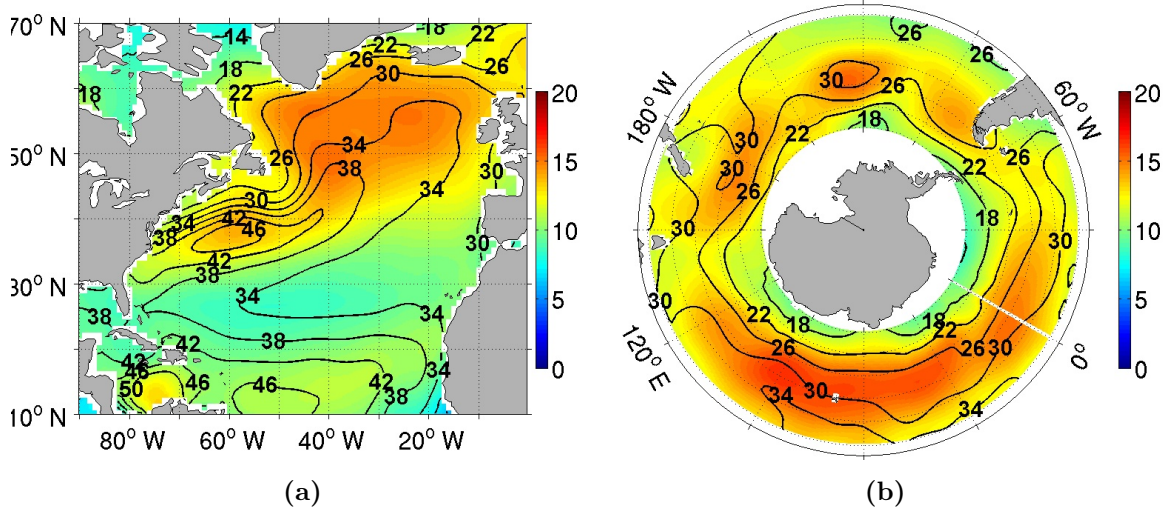


Fig. 5.3: Contours of the reference thermodynamic upper bound for the (turbulent = latent + sensible) air-sea feedback, from (5.5), in $\text{W/m}^2\text{K}$, for NA (a) and SO (b). Colours show its sensible contribution.

Figure 5.3 contours the reference upper bound for the turbulent air-sea feedback α_{ref} (5.5). Its magnitude is observed at typically $25\text{--}30 \text{ W/m}^2\text{K}$ over ACC and NA subpolar gyre, increasing towards $\geq 45 \text{ W/m}^2\text{K}$ in NA tropics and over the GS warm tongue. As further discussed in section 5.2, a feedback at this rate would lead to a relatively quick air-sea damping of SSTA in $\approx 2\text{--}3$ month in midlatitudes. At high latitude, where the feedback is weak and mixed-layers are deep, the reference damping time-scale is, at ≈ 5 months, much longer. Figure 5.4 contours the much weaker (typically $5 \text{ W/m}^2\text{K}$) lower bound for the large-scale turbulent feedback (5.6). Together, the contours in Figs. 5.4 & 5.3 show that, over the SO, thermodynamic processes damp SSTA with an intensity of at least $2\text{--}6 \text{ W/m}^2\text{K}$ and at most $20\text{--}35 \text{ W/m}^2\text{K}$, depending on the degree of the atmospheric thermal adjustment. In the NA, the damping strength is bound between $2\text{--}12 \text{ W/m}^2\text{K}$ on the lower end and $20\text{--}50 \text{ W/m}^2\text{K}$ on the upper end.

Whereas for $\partial \langle T^{a'} \rangle / \partial T' = 1$ (Fig. 5.4), the weak turbulent feedback is entirely provided by the latent contribution, once $\partial \langle T^{a'} \rangle / \partial T' = 0$ (Fig. 5.3), the sensible heat flux response to SSTA (given by $\rho^a \bar{u}^a c_S c_p^a$ and coloured in Fig. 5.3) provides about half the feedback at high latitude. Over the warm waters of NA subtropics and tropics, however, the (via $\left. \frac{dq_{sat}}{dT} \right|_{\bar{T}}$) strongly SST-dependent latent contribution still dominates.

is reduced and (iii) the low wind speed subtropical convergence zones due to higher c_S, c_L , where wind speed is low (cf. Fairall et al., 2003).

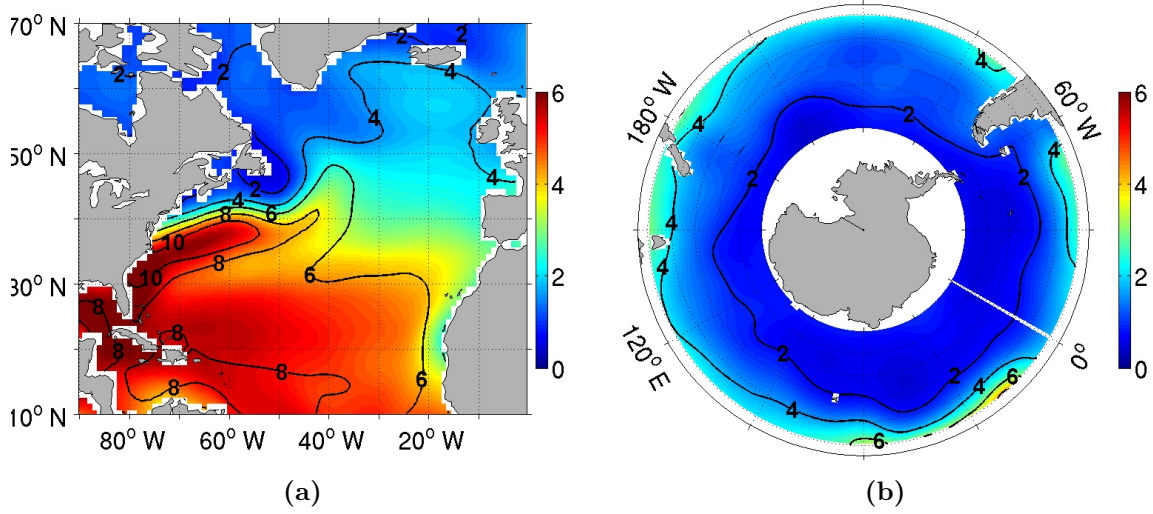


Fig. 5.4: Lower thermodynamic bound for the turbulent air-sea feedback, from (5.6), contoured in $\text{W/m}^2\text{K}$, for NA (a) and SO (b). Colours show $\overline{q_{sat}(T)} - q^a$ (in g/kg).

Both bounds show a relatively uniform spatial distribution in the SO, increasing slowly, together with the background SST field, from the cold pole equator- and warm-ward. The NA, which spans a broader range of latitudes and reaches warmer background SST, shows a larger variation and higher peaks of the air-sea feedback. A pronounced dip to this general latitudinal dependence of the upper bound is observed in the NA subtropical convergence. This is induced by the dependence of α_{ref} (5.5) on wind speed (whose spatial patterns are coloured in Fig. 5.3, as the displayed sensible bound is solely a function of u^a), which reaches a minimum between the peak westerlies and trade winds that enhance air-sea damping over the GS to the North and over the tropical gyre to the South. In the SO, peak westerlies over the ACC core are also visible as weak local maxima in α_{ref} (Fig. 5.3b). Additional spatial structure to the air-sea damping fields is provided by the oceanic mean flow distortion of the background SST field, which is weak for the ACC, but very pronounced for WBC's.

Most of the spatial structure of the lower bound (contoured in Fig. 5.4) is set by the background air-sea humidity contrast $\overline{q_{sat}(T)} - q^a$ (coloured in Fig. 5.4). The latter generally increases with SST, due to the exponential increase of q_{sat} with temperature and the weak observed variations in r_H over the oceans. The lower bound is thus weak over the equilibrated SO (weak thermal air-sea contrast, relatively high r_H), but peaks over warm poleward WBCs, such as the GS, ARC and Brazil current, locally increasing the lowest possible values of the thermal damping to $\sim 5\text{--}10 \text{ W/m}^2\text{K}$. Conversely, these

drop to only $\sim 1 \text{ W/m}^2\text{K}$ over cold equatorward WBCs such as NA and Malvinas current, characterized by local minima in air-sea contrasts.

This shows that complete thermal atmospheric adjustment is not enough to reduce the thermodynamic air-sea feedback to zero. To do so nearly everywhere, additionally, a non-zero, positive relative humidity adjustment $\partial \langle r'_H \rangle / \partial T'$ of $\sim +2\%/K$ is required (not shown). This, however, implies a larger moisture anomaly over warm SSTA than predicted by the Clausius-Clapeyron scaling ($\overline{r_H} \langle q_{sat}' \rangle$), and the reverse (i.e. $\partial \langle r'_H \rangle / \partial T' < 0$) is more likely (and observed in an analysis of individual ship-reports by Park et al. (2005)), confirming the above-zero lower bound on the (negative) thermodynamic air-sea feedback.

The maximum bound on the purely thermodynamic air-sea feedback is mainly set by background SST and wind-speed ((5.5) and Fig. 5.3). Its most intriguing aspects are the strong peak (at $\geq 40 \text{ W/m}^2\text{K}$) over the warm tongue of the separated GS and its rapid drop across the strong SST front of the latter's North Wall, both features, which are expected to be even more pronounced in higher resolution air-sea climatologies.

In summary, the regions associated with high mesoscale eddy activity, ACC and GS, show very different expected maximum thermodynamic damping climatologies: Whereas over the ACC, damping strength is fairly uniform and rarely exceeds $25\text{-}30 \text{ W/m}^2\text{K}$, it reaches a strong local maximum over the GS warm tongue, where air-sea damping can exceed $45 \text{ W/m}^2\text{K}$ without invoking any additional dynamical feedbacks.

Comparison to direct observational estimates of the large-scale turbulent feedback Due to the sparse data coverage, there are to my knowledge no direct estimates of α in the SO (i.e. south of 40°S). For the Northern Hemisphere, FK02 provide estimates from COADS and NCEP, which due to their low effective resolution are representative for the heat flux SST feedback on the large-scale. In the NA, the observed large-scale feedback (reproduced from FK02 in Fig. 5.5c, noting that NCEP turbulent fluxes are thought to be biased high) is indeed well delimited by the benchmarks estimated above (Figs. 5.4a & 5.3a). This is equally true for the total observed large-scale air-sea feedback (as its turbulent contribution dominates in midlatitudes, not shown). Over the GS system, the NA drift and the atmospheric storm-tracks approaching the British Isles, the observed large-scale feedback is close to the upper thermodynamic bound. In the NA subtropics and the north-western subpolar gyre, it lies somewhere between upper and lower bounds, and in the tropics it is close to the lower bound. This comparison suggests that the atmosphere adjusts

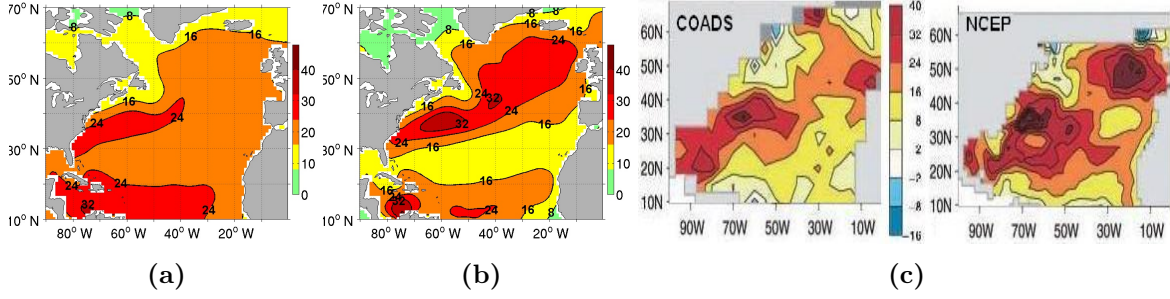


Fig. 5.5: Large-scale thermodynamic turbulent air-sea feedback in the NA: (a) estimated from air-sea climatology as (5.3) + (5.4) using $\partial \langle T^{a'} \rangle / \partial T' = 0.5$, (b) as (a) but using $\partial \langle T^{a'} \rangle / \partial T'$ linearly decreasing from 1 to 0.25 as $\overline{u^a}$ increases from 5 to 10 m/s (basin-average $\sim$ as in (a)), (c) observed applying lagged covariance analysis to NCEP and COADS reanalyses (from FK02, their Fig. 2, middle panels).

thermally to SSTA on the large-scale, but with geographically varying intensity.

Indeed, Figure 5.5a displays the large-scale feedback estimated from climatology assuming a constant atmospheric thermal adjustment of 50% (i.e. (5.3) + (5.4) using $\partial \langle T^{a'} \rangle / \partial T' = 0.5$). Whereas magnitudes (10–30 W/m²K) broadly agree with those observed (Fig. 5.5c), the resulting spatial distribution does so only poorly. If instead specifying a spatially-inhomogeneous $\partial \langle T^{a'} \rangle / \partial T'$, decreasing with increasing background wind speed⁹, the distribution of α (Fig. 5.5b) recovers many of the observed features, such as the peak over the storm track and the low values in the subtropical convergence. Obviously resemblance is still not perfect, with discrepancies especially over the Florida current, and a clearly different regime in the tropics, where the observed feedback (Fig. 5.5c) remains much weaker than that deduced from climatology (Fig. 5.5a,b), suggesting either $\partial \langle T^{a'} \rangle / \partial T'$ close to 1, or the importance of other mechanisms. Still, in the extratropics, a wind speed-dependent large-scale atmospheric thermal adjustment (of typically 50%) explains observations reasonably well, and seems physically plausible, with intense winds leaving less time for the “atmosphere” to adjust to underlying SSTA. (Adopting the same wind-speed dependent $\partial \langle T^{a'} \rangle / \partial T'$ in the SO, would yield a $\partial \langle T^{a'} \rangle / \partial T'$ increasing from ~ 0.25 over the ACC westerlies to ~ 0.5 on their flanks, and a large-scale turbulent feedback of ~ 20 W/m²K with a peak along the ACC, instead of a gradual equatorward increase as suggested by the upper bound.)

⁹ Arbitrarily, a linear decrease from 1 for wind-speeds below 5 m/s to 0.25 for wind-speeds exceeding 10 m/s is chosen. This yields typical values of $\partial \langle T^{a'} \rangle / \partial T' > 0.6$ in the subtropics, and < 0.2 in the westerlies, with a NA basin average of ~ 0.5 .

More recently, Park et al. (2005) estimate a similar feedback strength from monthly, 10° -by- 10° averages of individual COADS ship report-derived Q' and T' , interestingly detailing its thermodynamic and dynamical contributions. They also report a substantial atmospheric thermal adjustment, whose observed seasonal modulation (increasing in summer when background winds decrease, and reversely decreasing in winter) supports the idea of a wind-speed dependent $\partial \langle T^{a'} \rangle / \partial T'$. In their estimate, a slightly larger $\partial \langle T^{a'} \rangle / \partial T'$ ($\geq 75\%$, when averaged over the extratropical NA) compared to that deduced above would weaken the negative feedback, but is mediated both by a systematic negative large-scale humidity adjustment ($\partial \langle r'_H \rangle / \partial T' \sim -1\%/K$) and, puzzlingly, a weak positive “large-scale” dynamical coupling ($\partial \langle u^{a'} \rangle / \partial T' \sim 0.15 \text{ m s}^{-1}/K$). However it is unclear whether the latter does not reflect mesoscale residuals in his Q' calculated from individual ship reports. Their observation of a systematic, negative $\partial \langle u^{a'} \rangle / \partial T'$ in the tropics (observed to reduce the feedback by $10 \text{ W/m}^2 K$ in the tropical average), moreover provides a possible explanation for the weak observed feedback there.

The above discussion highlights that, even on the large-scale, the mechanisms of the turbulent air-sea feedback are still relatively poorly quantified from observations. Unsurprisingly then, coarse-grid coupled climate models still show a large spread in its modelled strength (Frankignoul and Kestenare, 2002b; Frankignoul et al., 2004), often significantly underestimating its magnitude, and not capturing observed features such as the peak along GS and storm track. This can be understood from the fact that a correct modelling of the air-sea feedback relies on the complicated problem of assessing the atmospheric response to SSTA.

In combination with the observed magnitude of the large-scale feedback strength (Park et al., 2005, FK02), the simple climatological constraints, presented here, show that, in the extratropics, the full potential of the purely thermodynamic turbulent air-sea coupling is realized only where background winds are strong, whereas elsewhere substantial large-scale thermal atmospheric adjustment ($\geq 50\%$) needs to be evoked to explain observations. Further observational and modelling studies are required to clarify whether the impact of dynamic coupling is indeed negligible on these large spatial scales.

5.1.3 Assessing the scope for an enhanced mesoscale feedback strength

Observationally estimating the mesoscale air-sea feedback strength is challenging. Whereas basin-wide mesoscale resolving MABL datasets only just become available (e.g. Bourassa et al., 2010; Dong et al., 2010), indirect observational estimates from

statistics of SST and SSH (e.g. Halliwell et al., 1991b; Zhai and Greatbatch, 2006b) have difficulties isolating air-sea from other feedback effects on mesoscale SSTA (cf. section 5.2). Even deriving α from the few dedicated, but scattered, field studies is not straightforward due to the requirement of isolating the SST-induced heat flux response from synoptic variability (i.e. estimating $\langle Q' \rangle$ in (5.1)).

Whereas modelling studies often focus on the effect of an existing mesoscale air-sea damping on the circulation (e.g. Drijfhout, 1994b; Zhai and Greatbatch, 2006b; Shuckburgh et al., 2011), few have focussed on estimating its strength, as this ideally requires mesoscale resolving coupled experiments with an explicitly resolved, or at least parametrized MABL. Without the latter, the atmospheric response to mesoscale SSTA cannot be assessed (i.e. $\partial \langle T^{a'} \rangle / \partial T' = \partial \langle u^{a'} \rangle / \partial T' = \dots = 0$), and the estimated feedback reduces to estimates of α_{ref} for the given modelled air-sea state. (Such an estimate is provided by Shuckburgh et al. (2011) for the SO, in which the higher resolution model SST field leads to finer spatial structures and locally larger values of α_{ref} .) Today, models adequate to study mesoscale air-sea interactions start to be designed, both in idealized (e.g. Spall, 2007b) and realistic configurations (Bryan et al., 2010). Whereas their analysis so far primarily focussed on understanding origins and impact of the recently discovered mesoscale dynamical coupling, diagnosing the intimately related mesoscale air-sea feedback strength from their outputs would be straightforward.

Due to the lack of direct observational estimates and to guide the interpretation of future modelling results, in the following we develop climatological constraints for the increase of the air-sea feedback from the large to the meso-scale, and shed light on the mechanisms underlying its spatial distribution.

Dynamical coupling As main contrast to the large-scale, a positive dynamical coupling ($\partial \langle u^{a'} \rangle / \partial T' > 0$) acts to enhance the feedback on the mesoscale. It primarily reflects a surface wind speed response to mesoscale SSTA gradients, which has been discovered by early air-craft and in-situ observations (e.g. Sweet et al., 1981). Its presence in all parts of the world ocean, in which the mesoscale contribution to SST variability (cf. chapter 4) is substantial, has been systematically established with the advent of high-resolution satellite-based scatterometer wind observations (QuickSCAT), by averaging observed small-scale perturbations in the wind stress magnitude τ over bins of SSTA (Chelton et al., 2004; with progress since reviewed by Small et al., 2008; Chelton and Xie, 2010). However, the observed coupling between wind stress and SST, given

by:

$$\frac{\partial \langle \tau' \rangle}{\partial T'} = \frac{\partial \langle \rho^a c_D u^{a2} \rangle}{\partial T'} \approx \rho^a \overline{c_D} 2\overline{u^a} \frac{\partial \langle u^{a'} \rangle}{\partial T'} + \rho^a \overline{u^{a2}} \frac{\partial \langle c'_D \rangle}{\partial T'}, \quad (5.7)$$

may not only reflect the SST-induced modification of MABL winds (with respect to the moving ocean), but also that of stability (via the strongly wind speed and stability dependent bulk momentum transfer coefficient c_D). Whereas Liu et al. (2007) have advocated the importance of the latter effect, recently, the dynamical nature of the coupling, i.e. a dominance of the former effect, could be confirmed from an analysis of buoy pairs across SST fronts (O'Neill, 2012, in agreement with previous results, e.g. O'Neill et al., 2005; Spall, 2007b). In a global analysis of 7-years of AMSR-E and QuickSCAT data, O'Neill et al. (2012), ON12a hereafter, establish that both $\partial \langle \tau' \rangle / \partial T'$ and $\partial \langle u^{a'} \rangle / \partial T'$ linearly depend on T' . Whereas the former strongly varies with background wind speed (as expected from the first term in (5.7)), so that, especially in the NH, it nearly disappears over the weak summer winds, $\partial \langle u^{a'} \rangle / \partial T'$ itself varies little geographically and is present year-round (consistent with the observed year-round presence of mesoscale SST variability in major frontal zones, cf. Figs. 4.18 & 4.19). It averages to 0.3 and 0.45 m/sK over GS and ARC/BMC, irrespective of analysis and dataset details (ON12a). Although local peaks may reach up to 0.6 and 0.8 m/sK in NH and SH, seasonal averages consistently lie below 0.5 m/sK.

The enhancement of the heat flux SST feedback due to this observed wind speed-SST coupling is given by¹⁰:

$$d\alpha_{dyn} = \rho^a (c_p^a \overline{c_S} (\overline{T} - \overline{T^a}) + L \overline{c_L} (\overline{q} - \overline{q^a})) \frac{\partial \langle u^{a'} \rangle}{\partial T'}, \quad (5.8)$$

defined such that $\alpha = \alpha_{ref} + d\alpha_{dyn}$. The lhs panels of Fig. 5.6 display $d\alpha_{dyn}$ estimated from air-sea climatology (cf. section 5.1.2), taking $\partial \langle u^{a'} \rangle / \partial T' = 0.5$ m/sK, on the high end of that observed by ON12a¹¹. For typical midlatitude conditions ($\overline{T} - \overline{T^a} = 0.5$ K, $\overline{q} - \overline{q^a} = 4$ g/kg) and a realistic coupling strength, (5.8) scales as $\approx 10^{-3}(10^3 0.5 + 2.5 10^6 4 10^{-3}) 0.5$ W/m²K. The enhancement of α through dynamical coupling is thus expected at *only* ≈ 5 W/m²K. Fig. 5.6a,c show that $d\alpha_{dyn}$ is observed about 5 times larger over GS and warm low latitudes (~ 5 -12 W/m²K) than over the equilibrated ACC and NA subpolar gyre (~ 2 -4 W/m²K), illustrating the key role of background air-sea contrasts in setting how strongly a given coupling strength enhances the turbulent

¹⁰ The weak wind-speed dependence of the transfer coefficients ($c_S \approx c_L \approx c$) induces an additional contribution of dynamical coupling (via $\frac{\partial c}{\partial u^a} \frac{\partial \langle u^{a'} \rangle}{\partial T'}$), which acts to reduce/enhance the feedback in weak/strong winds ($\frac{\partial c}{\partial u^a} < / > 0$) and is neglected.

¹¹ For reference, $\partial \langle u^{a'} \rangle / \partial T' = 0.5$ m/sK corresponds to $\partial \langle \tau' \rangle / \partial T' = 0.01$ - 0.02 N m⁻²K⁻¹ for $\overline{u^a}$ from 7.5-12.5 m/s, if $\rho^a = 1.22$ kg m⁻³ and $c_D = 1.3 \cdot 10^{-3}$.

heat flux feedback. In the SO, $d\alpha_{dyn}$ is likely to be important ($>5 \text{ W/m}^2\text{K}$) only over ARC and Brazil current.

Therefore the maximum mesoscale feedback strength, given by $d\alpha_{dyn}(5.8) + \alpha_{ref}(5.5)$, is only little enhanced with respect to the reference bound (Fig. 5.3). In this climatology, it reaches $\sim 30 \text{ W/m}^2\text{K}$ over the ACC, $35\text{--}40 \text{ W/m}^2\text{K}$ over BMC and ARC, and $>50 \text{ W/m}^2\text{K}$ over the GS, with the spatial patterns mainly set by α_{ref} . If $\partial\langle u^{a'} \rangle / \partial T'$ doubles to 1 m/sK (which is larger than any locally observed value shown by ON12a), the maximum mesoscale air-sea damping would increase by a further 5 (10) $\text{W/m}^2\text{K}$ over ARC (GS), with little effect over the equilibrated ACC itself (where $\alpha \approx \alpha_{ref} \leq 30 \text{ W/m}^2\text{K}$). For the mesoscale air-sea feedback to reach $100 \text{ W/m}^2\text{K}$, $\partial\langle u^{a'} \rangle / \partial T'$ would need to increase to about 5 m/sK in midlatitudes, and to more than 10 m/sK in high latitudes), i.e. by 1-2 orders of magnitude with respect to its observed value (of ON12a). Combined with recent satellite observations, climatology thus leaves little room for a substantial increase of the mesoscale air-sea feedback beyond its reference bound.

Thermal adjustment A realization of the above maximum mesoscale feedback requires the atmospheric thermal adjustment to indeed strongly reduce from its large-scale value of $\sim 50\%$ towards smaller scales. Here we explore hints from previous regional studies to assess whether $\partial\langle T^{a'} \rangle / \partial T'$ is indeed negligible on the mesoscale. Most early NA field programs examining the MABL impact of oceanic fronts (e.g. Sweet et al., 1981, Bane and Osgood, 1989; Atkinson et al., 1989 during the GALE experiment, Friehe et al., 1991 during FASINEX, and Kwon et al., 1998; Giordani et al., 1998 during SEMAPHORE) are biased towards a given meteorological condition, but all show increases not only in wind speed, but also in boundary-layer temperature and humidity between cold and warm sides of mesoscale SST fronts. In a composite of ship sections taken in 4 consecutive years Pezzi et al. (2009) observe $\partial\langle T^{a'} \rangle / \partial T' \sim 0.4\text{--}0.5$ across the meridional BMC front, in agreement with the climatological analysis by Tokinaga et al. (2005). (Larger $\partial\langle T^{a'} \rangle / \partial T' \sim 0.8$ are derived for a zonal mesoscale front to the west, which lies along instead of across the background wind direction). Substantial $\partial\langle T^{a'} \rangle / \partial T' (\sim 0.5)$ over narrow cross-wind midlatitude SST fronts are also suggested by the idealized coupled modelling study of Spall (2007b). Over isolated mesoscale SSTA (“eddies”), in contrast, Park et al. (2006) assume $\partial\langle T^{a'} \rangle / \partial T' \sim 0$ to interpret their observations of wind stress coupling. MABL simulations by Bourras et al. (2004), initialized to three months of in-situ observations, yield a non-zero, but

weak $\partial \langle T^{a'} \rangle / \partial T'$ ($\sim .25$) over a cold mesoscale SSTA in the north-eastern NA. In an analysis of AMSR-E SST and >2 years of air temperature measurements from the Atmospheric Infrared Sounder (AIRS), Liu et al. (2007) obtain $\partial \langle T^{a'} \rangle / \partial T' \sim 0.2-0.3$ over semi-permanent mesoscale SSTA associated with ARC meanders. The correlation analysis of monthly, BMC-average T' , $T^{a'}$ and $(T' - T^{a'})$ (obtained from a high-resolution reanalysis of COADS) by Tokinaga et al. (2005) supports substantial SST-induced variations in MABL stability due to a weak, but non-zero atmospheric thermal adjustment to transient mesoscale SSTA ($\partial \langle T^{a'} \rangle / \partial T' \sim .2$).

Observations thus suggest that on the mesoscale, too, atmospheric thermal adjustment acts to reduce the air-sea feedback. Denoted $d\alpha_{adj}$, its contribution to the feedback is given by the difference between the purely thermodynamic feedback (5.3)+(5.4) and its upper reference bound α_{ref} (5.5):

$$d\alpha_{adj} = -\rho^a \bar{u}^a \left(c_p^a \bar{c}_S + L \bar{c}_L \bar{r}_H \frac{dq_{sat}}{dT} \Big|_{\bar{T}^a} \right) \frac{\partial \langle T^{a'} \rangle}{\partial T'}. \quad (5.9)$$

The middle panels of Fig. 5.6 show $d\alpha_{adj}$, taking $\partial \langle T^{a'} \rangle / \partial T' = 0.25$, and clearly illustrate its increase with wind-speed and temperature. At $\bar{T}^a = 10$ K ($\frac{dq_{sat}}{dT} \Big|_{\bar{T}^a} \approx 0.5$ g/kg), latent and sensible adjustment contribute about equally and $d\alpha_{adj}$ scales as $-1.4 \bar{u}^a (1 + 1) \partial \langle T^{a'} \rangle / \partial T'$. Thus, in the westerlies ($\bar{u}^a \geq 10$ m/s), a relatively weak mesoscale thermal adjustment of 25% reduces the damping by at least ≈ 7 W/m²K, i.e. by an amount comparable to its increase due to the observed mesoscale dynamical coupling.

Net effect Some of the above quoted observational and modelling examples of mesoscale air-sea interactions over eddies and fronts are summarized in Table 5.1 (with parameters mostly visually estimated from published results and thus approximate). In all of them, both $\partial \langle u^{a'} \rangle / \partial T'$ (0.3-0.6 m/sK) and $\partial \langle T^{a'} \rangle / \partial T'$ (0.2-0.5) are operational. Their impact on the heat flux feedback depends on the climatological condition of the respective study, and is estimated using (5.5), (5.8) & (5.9), to yield the indicated values of α_{ref} , $d\alpha_{dyn}$ & $d\alpha_{adj}$. In all cases, dynamical coupling and thermal adjustment contributions to the feedback are weak individually (typically ± 5 W/m²K), and moreover substantially compensate, so that $d\alpha_{net} = d\alpha_{dyn} + d\alpha_{adj}$, i.e. the net increase of the damping from its reference bound α_{ref} towards the mesoscale¹², is even weaker. In these examples, the observed mesoscale air-sea feedback $\alpha = \alpha_{ref} + d\alpha_{net}$ thus lies

¹² This neglects second order terms, such as $\partial \langle u^{a'} T^{a'} \rangle / \partial T'$, as well as the effect of $\partial \langle r'_H \rangle / \partial T'$, which for relatively dry/cold air over cold/warm SSTA would add to the reduction of the feedback.

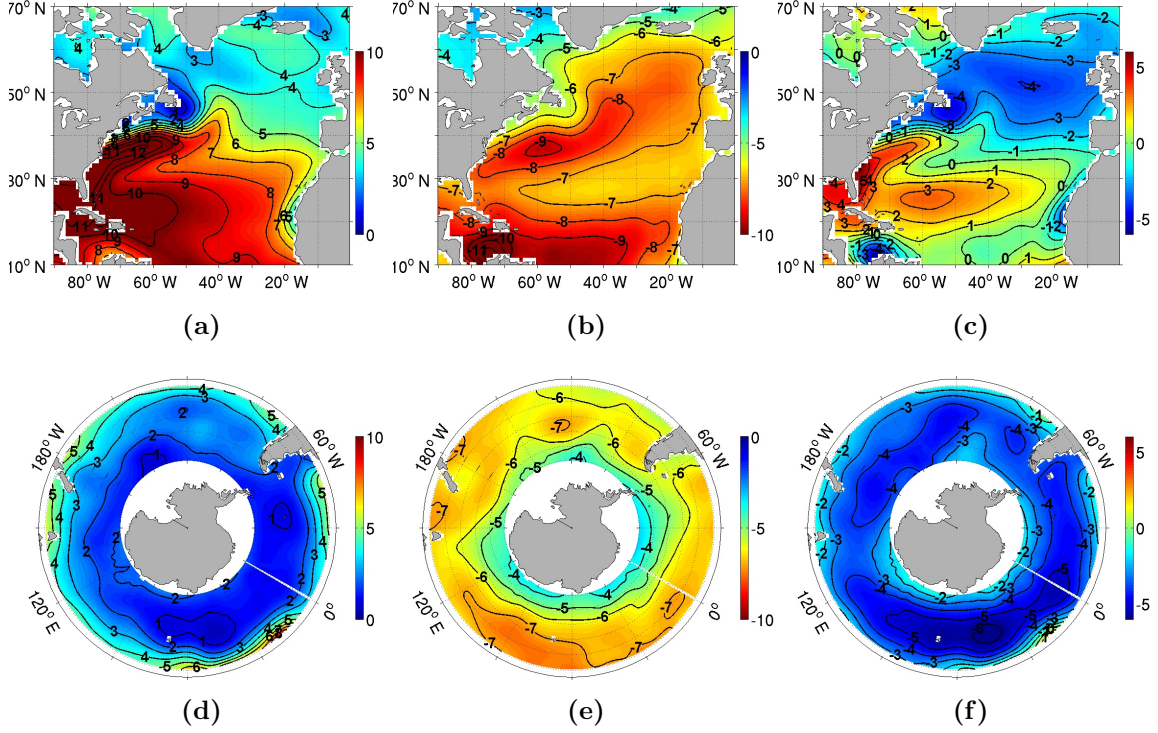


Fig. 5.6: Estimates of the increase in the mesoscale air-sea feedback with respect to α_{ref} (Fig. 5.3), induced by (a,d) dynamical coupling ((5.8), using $\partial \langle u^{a'} \rangle / \partial T' = 0.5 \text{ m/sK}$), (b,e) thermal adjustment ((5.9), using $\partial \langle T^{a'} \rangle / \partial T' = 0.25$) and (c,f) their sum, for NA (top) and SO (bottom). (Colour scale varies.)

within only a few $\text{W/m}^2\text{K}$ of the reference bound. Whereas Spall (2007b) (last column in Table 5.1) model a larger reduction (that likely reflects his rather untypically high choice of $\overline{u^a}$), a recent realistically configured mesoscale-resolving coupled simulation (Bryan et al., 2010) obtains simultaneous regression coefficients between small-scale heat flux and SST anomalies that are indeed close to α_{ref} over GS and ARC.

As illustrated by the map of $d\alpha_{net}$ in the rhs panels of Fig. 5.6, obtained as sum of its middle & lhs panels, a weak net contribution of dynamical coupling and thermal adjustment to the mesoscale feedback is also obtained from air-sea climatology. Fig. 5.6c,f illustrate how the balance is expected to be biased towards $d\alpha_{dyn}$ and thus an enhancement of α with respect to α_{ref} on the mesoscale, only where air-sea contrasts are strong (so that from (5.8) $\partial \langle u^{a'} \rangle / \partial T'$ is efficient) and winds are weak (so that from (5.9) $\partial \langle T^{a'} \rangle / \partial T'$ is inefficient).

For the chosen set of parameters ($\partial \langle u^{a'} \rangle / \partial T' = 0.5 \text{ m/sK}$, $\partial \langle T^{a'} \rangle / \partial T' = 0.25$), dynamical coupling wins in the warm NA subtropics/thermal adjustment over the cold, high wind regions of the NA subpolar gyre and the entire SO. The resulting imbalances are weak,

Tab. 5.1: Examples of previously observed mesoscale air-sea interactions, and estimates of the associated turbulent feedback.

	Bourras et al. (2004)	Liu et al. (2007)(AC)	Liu et al. (2007)(BD)	Pezzi et al. (2009)	Spall (2007b)
δT	-1	+1	+1	7	5
$\overline{u^a}$	8	9	9	7	15
$\overline{T}$	14.3	19.0	13.0	13	17.5
$\overline{q} - \overline{q^a}$	1.7	3.9	2.1	2.3	4.1
$\partial \langle T^a \rangle / \partial T'$	0.2	0.2	0.3	0.4	0.5
$\partial \langle u^a \rangle / \partial T'$	0.6	0.5	0.3	0.6	0.6
α_{ref}	20.4	27.4	22.5	17.5	44.2
$d\alpha_{dyn}$	+2.7 (+13%)	+5.1 (+18%)	+1.8 (+8%)	+3.5 (+20%)	+8.3 (+19%)
$d\alpha_{adj}$	-4.5 (-22%)	-4.4 (-16%)	-5.8 (-26%)	-5.3 (-30%)	-17.4 (-39%)
$d\alpha_{net}$	-1.8 (-9%)	+0.6 (+2%)	-4.0 (-18%)	-1.8 (-10%)	-9.1 (-20%)
α	18.6	28.1	18.5	15.7	35.1

in SI units, apart from $\overline{q} - \overline{q^a}$ in g/kg and % of α_{ref} , and notation as in text ($\alpha = \alpha_{ref} + d\alpha_{dyn} + d\alpha_{adj} = \alpha_{ref} + d\alpha_{net}$). Columns 1-3: “eddyding” features (1: 3-month observation in eastern NA, 2-3: quasipermanent SSTA features (denoted by letters) in ARC), 4-5: fronts (4: ship-observation composite across BMC, 5: idealized coupled modelling study representative of strong winds across narrow midlatitude fronts).

enhancing/reducing the mesoscale feedback by less than $5 \text{ W/m}^2 \text{ K}$ with respect to the reference bound. With the slightly weaker $\partial \langle u^a \rangle / \partial T'$ ($\sim 0.3 \text{ m/sK}$) observed in the NA (ON12a), here too thermal adjustment would win everywhere, leading to a net reduction of the mesoscale feedback with respect to α_{ref} by ~ 4 and $2 \text{ W/m}^2 \text{ K}$ in subpolar and subtropical gyres, with α_{ref} only just reached over the GS itself (not shown). The mesoscale feedback strength is thus mostly set by α_{ref} (contoured in Fig. 5.3). Indeed, from the considerations presented here, α_{ref} is a realistic, but also an upper, benchmark for the mesoscale feedback.

The above set of parameters is suggested to be representative for transient/ eddy-like mesoscale SSTA. Over mesoscale SSTA associated with fronts the few observations reviewed above suggest a stronger thermal adjustment ($\partial \langle T^a \rangle / \partial T' \sim 0.5$ instead of 0.25, cf. Table 5.1). This is physically plausible, as over the former winds will always be directed across an SST front, whereas over the latter most of the time an along-front component of the wind will allow the atmosphere to adjust (more strongly). The suggested weaker SST-induced modification of MABL stability over along-wind fronts is consistent with a systematically weaker observed mesoscale coupling of SST with wind stress curl than with wind stress divergence (Chelton et al., 2004, if interpreting curl/-divergence signatures as resulting from mainly along-/cross-front winds). The fact

that Tokinaga et al. (2005) deduce a strong $\partial \langle T^{a'} \rangle / \partial T' \sim 0.8$ and a weak $\partial \langle u^{a'} \rangle / \partial T' \sim 0.15$ for a zonal front to the west of the BMC aligned with background westerlies supports this interpretation. If a more systematic sampling attests the above values as representative, they suggest that both enhanced thermal adjustment and reduced dynamical coupling act to reduce the heat flux damping of frontal mesoscale SSTA (which are mesoscale in only one, large-scale in the other direction) compared to eddy-like features (mesoscale in two directions), introducing a subtle scale-dependence of the heat flux feedback.

5.1.4 The associated air-sea feedback time scale

The associated air-sea damping time scale of transient mesoscale SSTA $\tau_{air-sea} = \rho_o c_p h_{ml} / \alpha$, crucially depends on $\rho_o c_p h_{ml}$, the thermal inertia of the well-mixed surface layer that SST-induced air-sea heat fluxes act upon. Using a mixed-layer depth climatology (Appendix B.1) and the above result of a mesoscale feedback strength closely constrained to α_{ref} , function of mainly wind speed and SST, its climatology is estimated as $\tau_{ref} = \rho_o c_p \bar{h}_{ml} / \alpha_{ref}$ and displayed in Fig. 5.7. In the annual mean (lhs panels), it is observed to be typically 3 months in mid-latitudes, shorter (< 2 months) at low, and longer ($> 4-6$ months) at high latitudes. Its spatial patterns are primarily set by pronounced regional contrasts in mixed-layer depth (of nearly an order of magnitude between low-latitude/continental shelves and high-latitude/major current flanks, cf. Fig. 2.8), rather than by the relatively weak geographical variations in α_{ref} (Fig. 5.3). This also applies to the very pronounced seasonal variation of τ_{ref} , seen in the remaining panels of Fig. 5.7. τ_{ref} is much faster in summer (middle panels: typically 1-2 months in midlatitudes, 3-4 along the ACC) than in winter (rhs: 3-4 months in midlatitudes, ≥ 6 along the ACC), since shallow summer-time mixed layers by far outweigh the relatively weak summer-time reduction of α_{ref} (the latter is induced by weaker winds, but mediated by warmer SST).

Due to the feedback of SST-induced air-sea heat fluxes on mixed-layer depth, which remains to be systematically established from observations, warm/cold mesoscale SSTA are likely damped slower/faster than suggested by τ_{ref} (an aspect further examined in section 5.3). Large-scale SSTA are also damped to the atmosphere, but, especially away from atmospheric storm tracks and in the tropics, slightly slower than τ_{ref} (cf. section 5.1.2). Whether and how possible seasonal variations in atmospheric thermal adjustment ($\partial \langle T^{a'} \rangle / \partial T'$), mesoscale dynamical coupling ($\partial \langle u^{a'} \rangle / \partial T'$), as well as induced mixed-layer depth anomalies (h'_{ml}), modify the seasonal cycle of the air-sea damping

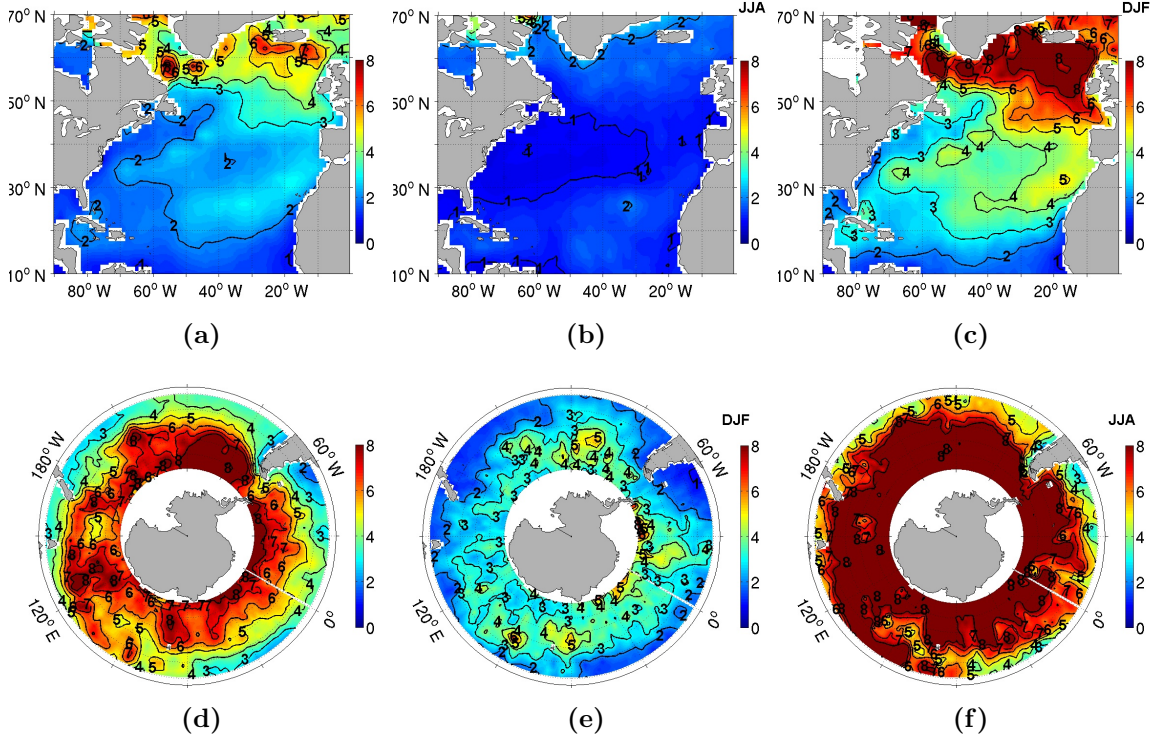


Fig. 5.7: Reference air-sea damping time scale $\tau_{ref} = \rho_o c_p \bar{h}_{ml} / \alpha_{ref}$ (in 30 days), for climatological mean (lhs), summer (middle) and winter seasons (rhs), in NA (top) and SO (bottom).

experienced by mesoscale and large-scale SSTA with respect to that given by τ_{ref} , is not clear from present data. (Further theoretical and observational work is thus required to assess whether the quicker air-sea damping of summer-time SSTA, suggested by τ_{ref} , may contribute to generating the observed reduced summer-time contribution of the mesoscale to SST variability (Fig. 4.20) established in section 4.4.3.)

5.1.5 Summary

This section has explored constraints on the strength of the negative air-sea heat flux-SST feedback on the mesoscale, imposed by air-sea climatology and previous observations of mesoscale air-sea interactions. The key results are:

- Over mesoscale SSTA, observed atmospheric dynamical coupling and thermal adjustment compete at enhancing/reducing the negative air-sea feedback with respect to its reference bound α_{ref} (positive values denote negative feedback). The latter is set by the background air-sea state as mainly an increasing function of differential air-sea speed and SST (given by (5.5) and contoured in Fig. 5.3).

- Mesoscale dynamical coupling, whose strength is relatively well constrained from recent QuickSCAT observational studies, can only impact the air-sea feedback in the presence of background air-sea contrasts. Its contribution to the feedback is thus close to negligible over the equilibrated ACC ($1\text{--}2\text{ W/m}^2\text{K}$). A substantial enhancement of the negative feedback is restricted to warm poleward currents (GS, ARC, Brazil current) and regions equatorward of 30° , but even there it hardly exceeds $\sim 5\text{ W/m}^2\text{K}$.
- Atmospheric thermal adjustment is systematically encountered in previous observational studies of mesoscale SST-MABL interaction. At a suggested $\sim 25\%$ over transient mesoscale SSTA, the induced reduction of the negative air-sea feedback is observed to everywhere (slightly more than) balance out the enhancement through coupling.
- α_{ref} is thus not only an upper bound for the large-scale feedback (observed to be lower by typically $10\text{--}20\text{ W/m}^2\text{K}$ in the extratropical NA), but moreover a realistic (upper) benchmark for the mesoscale air-sea feedback strength. The latter is expected to be weaker than α_{ref} by $\sim 0\text{--}5\text{ W/m}^2\text{K}$ and to amount to $20\text{--}35\text{ W/m}^2\text{K}$ in the SO and $30\text{--}45\text{ W/m}^2\text{K}$ in the NA, with the larger values characterizing the eddy-rich ARC, BMC and GS regions.
- A mesoscale feedback of $\sim 100\text{ W/m}^2\text{K}$ is inconsistent with present observations. They instead suggest that the air-sea feedback varies in space with the background air-sea state, at least as strongly as it does between atmospheric synoptic and oceanic mesoscale. Due to both larger thermal adjustment and weaker dynamical coupling, the thermal damping of the SST contrast across mesoscale fronts is suggested to lie somewhere between the large-scale and transient mesoscale feedback strengths.
- The associated air-sea damping time of transient mesoscale SSTA is shorter over the GS (~ 2 months) than along the ACC ($4\text{--}8$ months). The observed mesoscale air-sea damping thereby falls into a regime, in which, from previous modelling studies (cf. Fig. 5.2), it is expected to provide a large enough source of irreversibility to the mixed-layer to substantially enhance observed eddy diffusivities (more so over GS than ACC). Subject to further research, pronounced seasonal variations in the air-sea feedback time (faster in summer), induced by its direct dependence on mixed-layer depth, may induce a non-negligible seasonal modulation in the mixed-layer eddy heat diffusion (enhanced in summer).

The sensitivity of these constraints to the chosen mixed-layer depth (de Boyer Montégut et al., 2004) and air-sea climatology (da Silva et al., 1994) and their limitations (espe-

cially in the SO), as well as to effects not addressed here, specifically, to the neglected impact of ocean currents on the differential air-sea speed u^a (which likely leads to a slight overestimate of the feedback, especially over ACC westerlies), to the neglected contribution by SST-induced modifications of transfer coefficients ($\partial \langle c'_L \rangle / \partial T'$), and to deviations of the mesoscale MABL moisture adjustment from a Clausius-Clapeyron scaling (reduced r_H over warm SSTA would imply a weaker net atmospheric adjustment, and thus a larger feedback, than obtained here by setting $\partial \langle r'_H \rangle / \partial T' = 0$), remains to be assessed.

Recent observations of a systematic collocation between anomalies in cloud-liquid water content, cloud-top height and all-sky albedo with quasi-steady mesoscale SSTA in the ARC (O'Neill et al., 2005; Liu et al., 2007; Bryan et al., 2010) provide indication for a systematic radiative feedback specific to the oceanic mesoscale that merits further investigation. However, given its weak modelled strength (negative and of a few $\text{W/m}^2\text{K}$ —Bryan et al., 2010), the mesoscale air-sea feedback is indeed expected to be largely dominated by the turbulent contribution, considered here.

A better understanding of the suggested weak enhancement of the air-sea feedback as scales decrease below the atmospheric synoptic (“large”) scale requires a basin-wide assessment, from mesoscale resolving datasets of MABL temperature and humidity, of the atmospheric adjustment on both large- and mesoscales (and moreover an assessment of whether indeed a dynamical MABL response to large-scale SSTA is negligible). Early arguments favouring a strong scale-dependence of the damping strength (Bretherton, 1982; Rahmstorf and Willebrand, 1995) build on an inversely scale-dependent efficiency of atmospheric heat export, and also assume the vertical scales of induced atmospheric heating perturbations to strongly depend on the horizontal scale of underlying SSTA (with mesoscale SSTA only inducing changes in the relatively shallow MABL, contrary to large-scale SSTA over the whole troposphere). However recent observations reveal a deep-reaching impact on the atmosphere also over the mesoscale, in the form of fronts and quasi-permanent eddies (Minobe et al., 2010; Liu et al., 2007). This might explain why atmospheric thermal adjustment, and thus the air-sea feedback, is not as scale-dependent as previously assumed. A different/additional possible explanation might be that the inverse scale-dependence derived from diffusive arguments only applies to scales ≥ 1000 km, on which the atmospheric heat export is primarily driven by baroclinic waves, whereas different processes are likely to be relevant on oceanic mesoscales.

5.2 On the persistence of SSTA - at a given location

To gain further insight into the mesoscale contribution to SST variability, established from its space scales in chapter 4, this (and the following) section characterize the time scales of MW-SSTA observed at fixed locations (and following eddies), and establish how they link to the above result of a relatively weak mesoscale air-sea feedback strength.

The persistence of SSTA at a given location is characterized by their (temporal) auto-correlation function, obtained at each grid point as $r_{TT}(\tau) = \langle T'(t)T'(t + \tau) \rangle \equiv \overline{T'(t)T'(t + \tau)} / \overline{T'^2}$, where t denotes time and τ the temporal lag (hereafter given in days d). Figure 5.8 displays r_{TT} observed in energetic (a) and quiet regimes (b) of NA and SO (as defined in section 3.3.2). The median of both regions' r_{TT} (grey) decays nearly exponentially, and indicates a quick decorrelation of SSTA at any given location. The observed SSTA decay rate is on average faster in the energetic regime ($\sim e^{-t/20d}$ to $e^{-t/30d}$) than in the quiescent regime ($\sim e^{-t/40d}$)¹³. Indeed, the 5-95% range of each regime's r_{TT} (grey shading) shows that in quiet regions SSTA can remain correlated at $\geq 1/e$ for ~ 3 months, whereas in energetic regions, where the mesoscale contribution to SST variability is substantial, SSTA can e-fold within only ~ 10 days.

Figure 5.9 maps the underlying spatial patterns of SSTA persistence, here simply characterized by the time lag $\tau_{1/e}(T')$, at which r_{TT} first e-folds¹⁴. By this measure, SSTA persists longest (≥ 3 months) at fixed locations in quiescent eastern subtropics (and tropics), as well as at a few patches at high latitude. Approaching the regions of elevated eddy variability (indicated by the contours of SSHA rms in Fig. 5.9), observed fixed-location SSTA memory drops to a month or less. (However, short SSTA persistence (≤ 30 days) is also found in several regions of low eddy variability, such as in the eastern subpolar NA, and especially the very quick SSTA decay observed around Antarctica may be explained by artefacts in SSTA due to undetected sea ice - cf. section 3.1.2).

¹³ Here r_{TT} shows a fast initial drop, after which it decays slower than exponentially between lags of 1–6 weeks.

¹⁴ Qualitatively, patterns are insensitive to the use of different methods to characterize SSTA persistence, such as exponential fits (over the first two, three, four, etc. to 9 months) or the integral of r_{TT} to its first zero-crossing, but both local and basin-wide quantitative differences arise. (Such, exponential fits yield generally larger persistence with increasing maximum lag, and the zero-crossing time itself yields larger persistence than all other methods and very noisy results.)

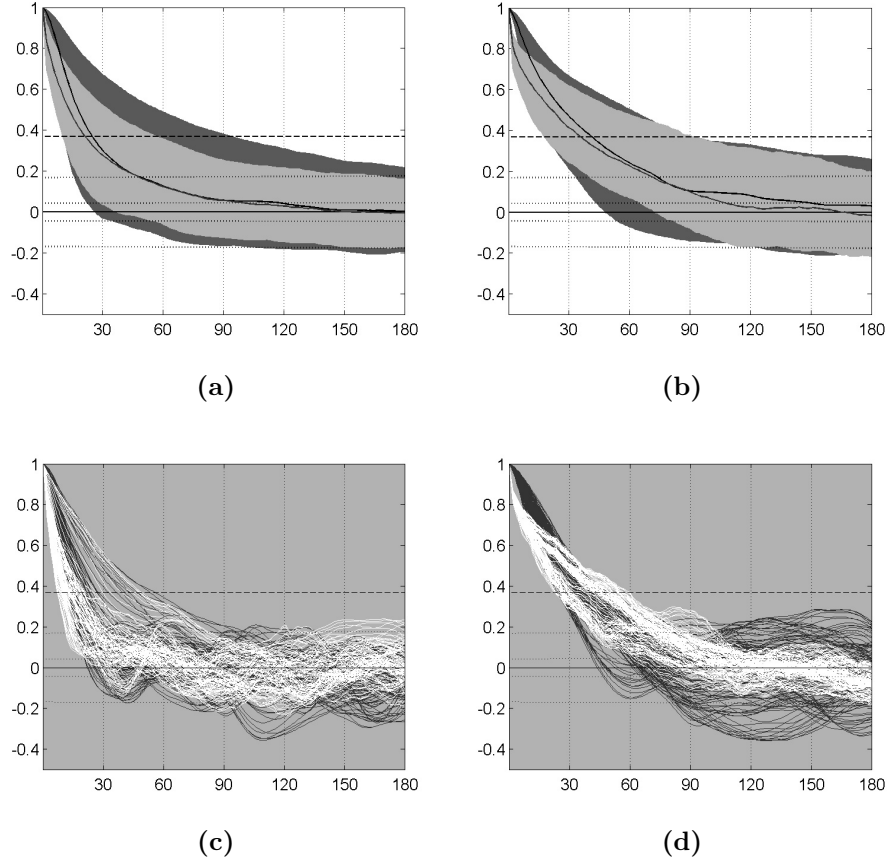


Fig. 5.8: Auto-correlation functions of SSTA (light) and SSHA (dark) observed in energetic (a) and quiet regions (b) of NA and SO (lines shows the median, shadings the 5-95 percentile range). Bottom panels illustrate auto-correlations obtained at individual grid points along zonal sections (dotted green in Fig. 3.12a) through the GS at 39°N (c) and through the eastern NA subtropical gyre at 25°N (d). (In all panels, dotted lines indicate where correlations are significantly different from zero at 95% confidence if one observation every day & every 15 days is independent.)

Link to air-sea damping and previous observations The general patterns of $\tau_{1/e}(T')$ (Fig. 5.9) are not reminiscent of those seen in τ_{ref} (Fig. 5.7a,d), expected to well approximate the time it takes to damp mesoscale SSTA to the atmosphere, with large-scale SSTA typically damped slightly slower. Instead, throughout high-latitudes and mid-latitude current systems, SSTA persistence is much shorter than suggested by air-sea damping. Only the long $\tau_{1/e}(T')$ of ~ 2 – 3 months observed in the NA subtropics are consistent with air-sea damping being the predominant factor limiting SSTA persistence. Here, from comparison with Fig. 4.7a, SST variability typically

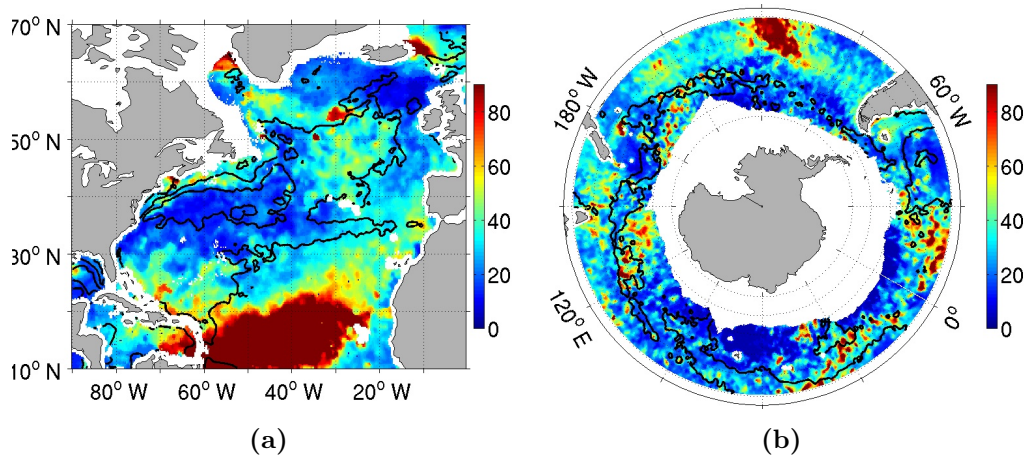


Fig. 5.9: The observed SSTA auto-correlation e-folding time (in days) for NA (a) and SO (b). Black contours delimit regions of elevated SSH variability, $\sigma(\eta') = 5$ & 15 cm in (a), 10 cm in (b).

occurs on large spatial scales and observed SSTA persistence compares well to results of previous studies in the NH, which are all based on datasets of large-scale SST variability (Frankignoul, 1985; Kushnir et al., 2002, with particularly good agreement to the estimate provided for the eastern NA from a 5°-by-5° monthly SSTA dataset by Frankignoul et al. (1998)). Extending the result of chapter 4, this indicates that in NA quiet regions observed MW-SSTA evolve on large scales in both space and in time, and do so in a manner predominately controlled by basin-scale air-sea interactions. (In the tropical NA, SSTA persist even longer than indicated by τ_{ref} , as there the *observed* large-scale air-sea damping (shown in Fig. 5.5c) is substantially weaker than its reference bound.)

Whereas the observed fixed-location persistence of MW-SSTA is comparable to that of lower resolution (1°, weekly) IR-SSTA (Leeuwenburgh and Stammer, 2001) in the NA, it is typically shorter in the SO. There, reference air-sea damping (Fig. 5.7d) would allow for a much longer SSTA persistence of on average 4-5 months, which is not realized in SSTA observations. The few bands of elevated SSTA persistence observed along the ACC (Fig. 5.9b) indeed do not coincide with the regions of largest SSTA spatial scales, but instead closely follow pronounced bathymetric ridges (such as the Mid-Atlantic ridge and its extension towards the south of Africa, as well as, to the south of New Zealand, the arc formed by South-East Indian and Pacific-Antarctic ridges - cf. Fig. 3.3), which suggests different (smaller-scale) dynamics at work.

The very short SSTA persistence of only 10–30 days observed typically in NA and

SO energetic regions implies, for typical mixed-layer depths of 50–100 m, a damping rate of 200–400 W/m²K. This can clearly not be accounted for by observed air-sea heat exchanges. Although noise in the data¹⁵, due to measurement errors, undetected precipitation and sea-ice (section 3.1.2), as well as day-to-day noise present in the removed SST seasonal background state due to its estimation from only 5–6 years of data¹⁶, may artificially reduce SSTA memory, it is likely that other processes need to be evoked to account for the regionally observed short fixed-location persistence of MW-SSTA.

Mechanisms From the (linearized) slab mixed-layer heat budget (section 2.2), the evolution of SSTA at a fixed location is given by

$$\partial_t T' = F' - \frac{\alpha_D T'}{\rho_o c_p h_{ml}} - \bar{\mathbf{u}} \cdot \nabla T'. \quad (5.10)$$

Once generated by anomalous forcing F' (mainly through large-scale air-sea heat and Ekman fluxes, and mesoscale isotherm stirring $\mathbf{u}'_g \cdot \nabla \bar{T}$), SSTA are diabatically damped by the SST-induced mixed-layer heat convergence $-\alpha_D T'$, which includes contributions from submesoscale mixing and entrainment that may modify the persistence of SSTA with respect to that expected from air-sea damping alone¹⁷. At a given location, however, from the 3rd term in (5.10), advection by the background flow provides an additional “forcing” or “damping” mechanism. The associated time scale is given by $\tau_{adv} = \sigma(T') / \sigma(\bar{\mathbf{u}} \cdot \nabla T')$. Advection by a typical mean-flow of 10 cm/s thus acts to damp a mesoscale SSTA of 1K within only 10 days, which is comparable to the observed fixed-location SSTA decay time in major current systems. To the contrary, if the SSTA is large-scale (~ 1000 km), this would take about 10 times longer (i.e. several months). On the large-scale, the apparent damping induced by background advection $\rho_o c_p h_{ml} / \tau_{adv}$ is thus, at ~ 20 W/m²K (for the above parameters and a 50 m mixed-layer), comparable to the damping induced by air-sea heat fluxes. On the mesoscale, however “damping” by advection is typically an order of magnitude larger (~ 200

¹⁵ Note that apart from the removal of a small linear trend no filtering has been applied to the daily SSTA time series before calculation of r_{TT} .

¹⁶ As further discussed in Appendix A.1, this aspect induces systematic artefacts in SSTA statistics, highlighting the need for further research into fully assessing the sensitivity of results to these biases, as well as on devising methods of minimizing their effect in observational studies based on short time series.

¹⁷ Entrainment is expected to enhance rather than reduce the persistence of SSTA on the large-scale (due to winter-time re-emergence, e.g. Deser et al., 2003; de Coetlogon G. and Frankignoul, 2003), and likely, following the discussion in the next section, also on the mesoscale.

$\text{W/m}^2\text{K}$). This shows that, as previously recognized (Frankignoul and Reynolds, 1983; Frankignoul, 1985), advection is an important factor limiting SSTA persistence at a given location, the more so the smaller the characteristic SSTA spatial scale.

An estimate of τ_{adv} , obtained from observed SSTA and a seasonal climatology of the background mean-flow¹⁸, is mapped in the top panels of Fig. 5.10, and reveals a very close agreement in overall magnitude and spatial distribution to the observed SSTA decay time scale (Fig. 5.9). For comparison $\tau_{stir} = \sigma(T')/\sigma(\mathbf{u}'_g \cdot \nabla \bar{T})$, which gives the time scale of dissipation required to explain observed SSTA magnitudes if mixed-layer eddy stirring were the only SSTA forcing mechanism¹⁹, is mapped in the bottom panels of Fig. 5.10. It is observed to be comparable in magnitude to τ_{adv} (top panels) throughout ocean basins, whereas τ_{ref} (Fig. 5.7a,c) is typically more than 5 times slower. Thus, it is predominately advection by the background flow that provides the apparent (i.e. fixed-location) dissipation required to balance the generation of SSTA through eddy stirring.

Fig. 5.10a,b show that in basin interiors, where predominant SSTA spatial scales are large and geostrophic flows so weak that Ekman currents (typically 1–2 cm/s) are of comparable importance, background advection allows SSTA to persist for longer than 2–3 months at a given location. Here τ_{ref} (Fig. 5.7a,c) becomes comparable to τ_{adv} , so that air-sea damping accelerates the fixed-location SSTA decay with respect to that expected from advection alone. This explains why here (apart from the regions of extremely long persistence) SSTA often decays (Fig. 5.9) slightly faster than suggested by τ_{adv} (and is better explained by the net (advection + air-sea damping) time scale $1/(1/\tau_{adv} + 1/\tau_{ref})$, not shown), which underlines the important role of air-sea damping in the fixed-location mixed-layer heat budget in quiet interiors.

In energetic regions of GS and ACC, to the contrary, background advection alone would limit the fixed-location SSTA persistence to less than a month, in current cores often to only ~ 10 days. Here observed SSTA persistence is indeed short, but not as

¹⁸ To derive a monthly climatology of the Ekman flow, here we use the Risien and Chelton (2008) Scatterometer Climatology of Ocean Winds (SCOW), representative for the 1999–2009 period (cf. Appendix B.3), and take the climatological mixed-layer depth (cf. Appendix B.1) as proxy for the climatological Ekman layer depth. The background geostrophic flow is derived from the CNES-CLS09 mean dynamic topography recently developed by Rio et al. (2011) (cf. Appendix B.4), from which a monthly MDT climatology adjusted to the 1999–2009 period is obtained by adding the monthly sea-level anomaly climatology of that period (obtained from the 2-satellite AVISO reference series).

¹⁹ τ_{stir} thus underestimates the required fixed-location dissipation particularly in quiet regions, where other SSTA forcing mechanisms are important.

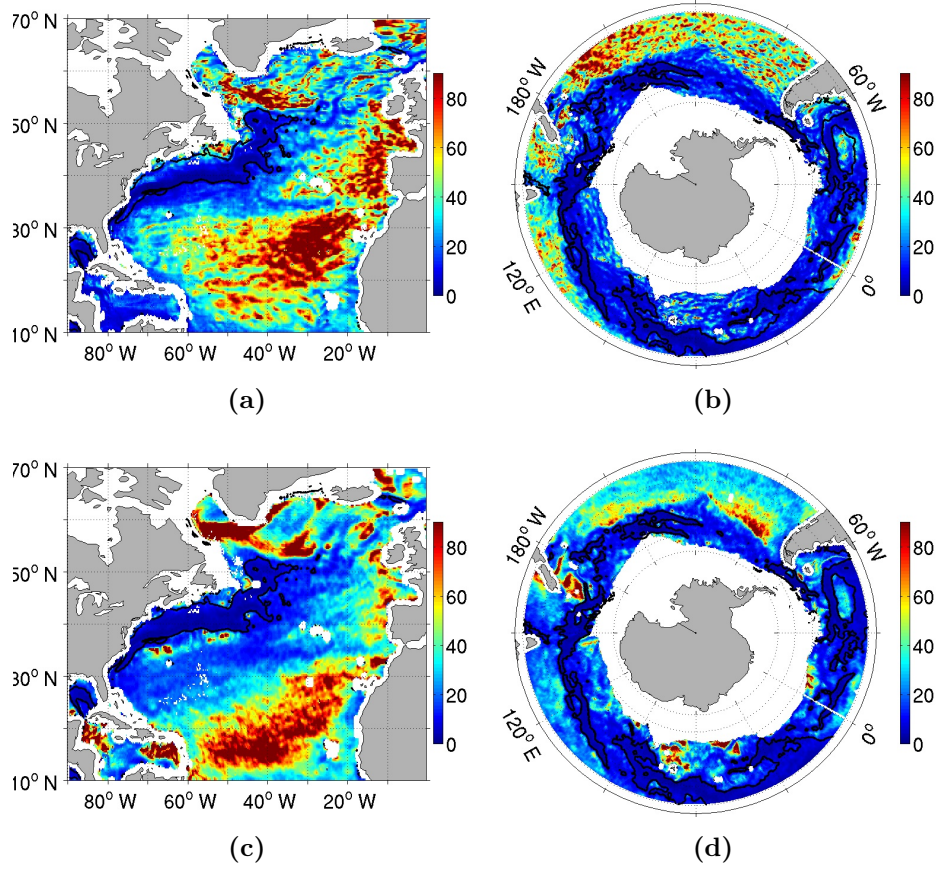


Fig. 5.10: Time scales characterizing mean-flow advection (top: $\tau_{adv} = \sigma(T')/\sigma(\bar{\mathbf{u}} \cdot \nabla T')$) and eddy stirring (bottom: $\tau_{stir} = \sigma(T')/\sigma(\mathbf{u}'_g \cdot \nabla \bar{T})$) of SSTA, coloured in days for NA (lhs) and SO (rhs). Contours indicate $\sigma(\mathbf{u}'_g \cdot \nabla \bar{T}) = 1 \text{ K}/_{10 \text{ day}}$.

short as indicated by τ_{adv} . If in these regions of a large mesoscale contribution to SST variability, the observed SSTA decay simply reflects how quickly mesoscale SSTA pass a given location, this could be explained by the observed competition of westward eddy propagation against typically eastward mean flow advection. If such a strong impact of the mesoscale on SSTA persistence is indeed observed, we would expect SSTA statistics to closely follow those of SSHA.

Fig. 5.8a,b compare auto-correlation functions of observed SSHA $r_{\eta\eta}$ (dark curves and shading) to those of SSTA (light curves and shading). On average the two decorrelate comparably fast in both energetic and quiescent regimes, but the shapes of SSTA and SSHA auto-correlation functions are only comparable in energetic regions. This becomes more clearly apparent in the bottom panels of Fig. 5.8, which display r_{TT} (white) and $r_{\eta\eta}$ (black) observed at several individual locations within the two regional

bins: In the energetic regime (Fig. 5.8c), the quick SSTA decay at a given location indeed follows that seen in SSHA. Here individual r_{TT} show pronounced low-frequency oscillations, with periods comparable to ~ 100 days, which are clearly reminiscent of those in the corresponding $r_{\eta\eta}$. They provide an explanation for the significantly lower than zero²⁰ bin-average r_{TT} observed at lags larger than ~ 2 months in Fig. 5.8a (grey shading), and show that, in energetic regions, observed r_{TT} are inconsistent with the purely exponential SSTA decay, expected for predominately atmospherically-forced SST variability (Frankignoul and Hasselmann, 1977). (The near-exponential decay seen in Fig. 5.8a arises only as average over locally varying eddy recurrence characteristics). In agreement with previous results on SSTA statistics from near-mesoscale resolving datasets in eddy-rich frontal zones (Halliwell et al., 1991a), this provides a clear indication of the impact of the oceanic mesoscale on SSTA statistics in energetic regions, confirming the results of chapter 4. In quiet regions (Fig. 5.8d), to the contrary, the low-frequency oscillations in $r_{\eta\eta}$ find an only weak correspondence in r_{TT} , which instead decays more smoothly and close to exponentially, confirming the more large-scale nature of SST variability in basin interiors.

Fig. 5.11 provides comparison of the spatial distribution of temporal SSTA and SSHA statistics. The SSHA e-folding decay time $\tau_{1/e}(\eta')$ is mapped in Fig. 5.11a,b. Although its range from 20 days to ≥ 3 months is comparable to that of $\tau_{1/e}(T')$ (Fig. 5.9), their spatial patterns differ. Inspection of SSHA longitude-time plots (not shown) reveals that, especially over the bathymetric features along the ACC and the mid-NA ridge, much of the apparent noisy/small-scale structure of $\tau_{1/e}(\eta')$ is set by regionally varying net SSHA propagation speeds. The ratio of fixed-location SSTA to SSHA persistence, displayed in Fig. 5.11c,d, highlights the different processes captured by the two fields. For example, SSTA is observed to persist much longer than SSHA over continental shelves/shallow seas (where short SSHA persistence likely reflects residuals of barotropic high-frequency variability, not entirely removed during the AVISO processing, cf. sections 3.2.2 & 3.2.3) and in the tropics. In regions of intermediate eddy variability (mostly between the two black contours), SSTA typically decays quicker than SSHA (as here large scales dominate the variability of the former, but mesoscale eddy propagation that of the latter, cf. Fig. 4.7). Importantly, however, Fig. 5.11c,d strikingly confirms that throughout the regions of intense eddy stirring (inner black contour), in which SSTA and SSHA evolve on similar spatial scales, the

²⁰ The dotted lines indicates where r differs from zero at 95% confidence if only one observations per 2 weeks is independent, which is much less then suggested by the near-daily MW-SST resolution.

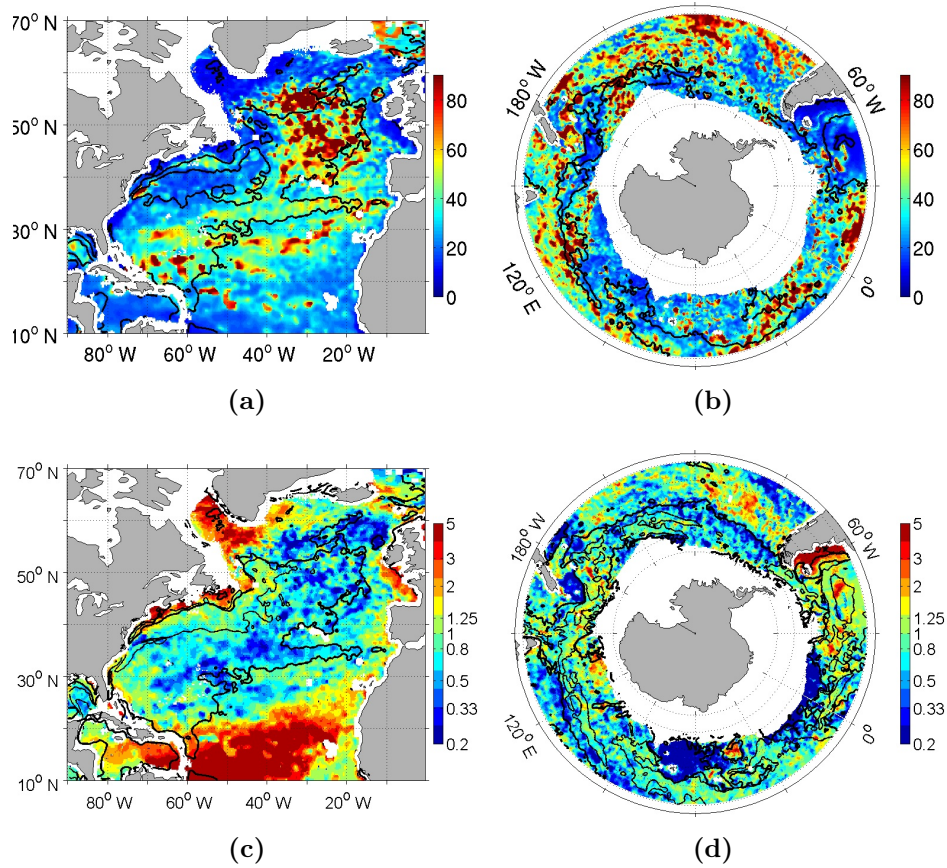


Fig. 5.11: Observed SSHA auto-correlation e-folding time (top, in days) and the ratio of SSTA to SSHA e-folding times (bottom), for NA (lhs) and SO (rhs). Contours in (a) indicate $\sigma(\eta') = 5$ & 15 cm, in (b) 10 cm. Inner and outer contours in c,d indicate $\sigma(\mathbf{u}'_g \cdot \nabla T) = 1 \text{ K}/10 \text{ day}$ and $\sigma(\eta') = 5$ cm.

observed quick fixed-location SSTA decay is also comparable to that of SSHA. Here the oceanic mesoscale therefore has a profound impact not only on spatial but also on temporal SSTA statistics.

Summary Differing temporal resolution capacities of SST and SSH datasets, as well as the imperfect separation of non-seasonal from seasonal variability estimated here from short observational time series (Appendix A), limit the realism of present observational results, which are therefore subject to future confirmation. Within these limitations, the analysis of fixed-location temporal SSTA and SSHA statistics corroborates the findings of chapter 4 on the nature of observed SST variability:

- In quiescent basin interiors, SST variability is predominately shaped by large-scale, atmospherically-forced anomalies that slowly decay (during several months) as ex-

pected from their feedback on turbulent air-sea heat exchanges.

- Only in regions of vigorous eddy stirring, observed along GS and ACC SST fronts, (deep-reaching) mesoscale SSHA control MW-SSTA scales in both space and time. This reflects the only weak damping of the latter to the atmosphere (section 5.1), which here has little impact on the fixed-location persistence of SSTA, the latter being instead controlled by the self-induced propagation and mean-flow advection of eddying features.

To shed more light on the evolution of the SST-SSH relationship of mesoscale eddies through time, and the role diabatic processes such as air-sea damping or entrainment may play therein, we now turn to an analysis of the persistence of SSTA following the tracks of propagating eddies.

5.3 On the persistence of SSTA - following eddy tracks

This section extends the analysis of eddy-core SSTA provided in chapter 4.3 for propagating eddies with origins in energetic and quiet oceanic regimes, by characterizing their evolution throughout the eddy lifetime (section 5.3.1). A preliminary analysis of the mechanisms likely to be relevant is provided in section 5.3.2.

5.3.1 The evolution of SSTA throughout the eddy lifetime

Figure 5.12 displays the average evolution of eddy-core SSTA $\langle T'(s) \rangle$ (red for anticyclones, blue for cyclones) following SSHA tracks in time and space (as indicated by the coordinate $s = s(x, y, t)$). It is isolated by averaging SSTA over eddy cores and over many eddies. Instead of including all available eddy snapshots (as in chapter 4.3.1i), here composites are evaluated separately for each week within the eddy lifetime, which drastically reduces the number of observations that enter each average. Nevertheless $\langle T'(s) \rangle$ clearly stand out from residuals of large-scale SST variability (thin stems in Fig. 5.12, obtained by repeating the analysis with a large-scale, 10° -by- 10° box-averaged SSTA dataset) throughout the lifetime of energetic regions' eddies (Fig. 5.12a). Following quiet regions' eddies (Fig. 5.12b), this is only observed during the first months. After ~ 4 months for cyclones (9 for anticyclones) large-scale SSTA residuals become comparable in magnitude, indicating that eddy numbers are not sufficient any more (≤ 400 cyclones, and 75 anticyclones) to robustly isolate the weak eddy signal.

In stark contrast to the short persistence of mesoscale SSTA at fixed locations (≤ 1 month where SST variability occurs on mesoscales), Fig. 5.12 reveals that SSTA persist

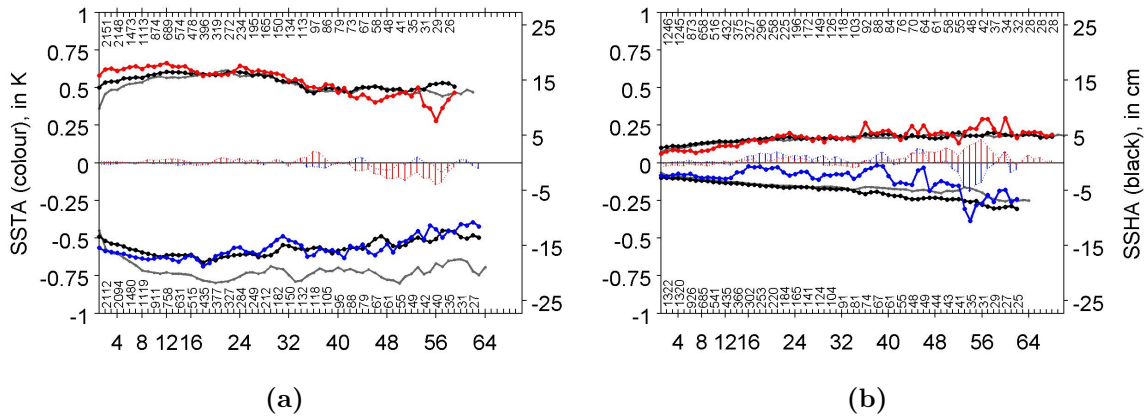


Fig. 5.12: Evolution through time (abscissa, in weeks) of eddy-core SSTA composite-averaged over propagating anticyclones (red) & cyclones (blue) with origins in (a) energetic and (b) quiet regimes of NA and SO. Black and grey curves show the corresponding composite evolution of eddy-core SSHA and eddy amplitude, thin stems that of large-scale (10° -by- 10°) SSTA, evaluated at the eddy positions. As indicated at the top & bottom of the Figure, respectively, decreasing eddy numbers are available for compositing with increasing eddy lifetime (only composites over at least 25 eddies are displayed).

almost unchanged throughout the eddy lifetime when following their paths, i.e. for several months and longer. The persistence of eddy SSTA is observed to be strongly tied to that of core-average SSHA $\langle \eta' \rangle$ (black), or similarly the eddy amplitude $\langle A_e \rangle$ (grey - cf. 3.3.3)²¹. As both eddy SSTA and SSHA amplitudes are expected to be biased low (at typically 80% of the true observed eddy-core amplitudes, as shown in Appendix A.3), their ratio should be unaffected. Approximately constant in time, it is observed at typically 0.4 K/10 cm over both energetic (0.6K/15 cm) and quiet eddies (0.2K/5 cm), and thereby clearly cannot be explained from mixed-layer steric effects (cf. chapter 4.3.2). Fig. 5.12 thus extends the result of chapter 4, and shows that along GS and ACC, but also in quiescent interiors, deep-reaching mesoscale circulations are systematically associated with surface thermal anomalies that persist throughout their lifetime.

²¹ Although they do not filter large-scale SSH signals as A_e and are based on a different SSH dataset (section 3.2.3), $\langle \eta' \rangle$ are seen to be an overall surprisingly good proxy for the eddy amplitude, but display a weaker initial growth and do not capture the asymmetry of A_e towards more intense cyclones in energetic regions (cf. chapter 4.3.2).

Figure 5.13 composite-averages all individual eddy-core time series $T'(s)$ and $\eta'(s)$ while they are aligned, instead of at the start of the eddy life as in Fig. 5.12, at the week in their life that corresponds to the end of winter (denoted by time zero), which gives the solid curves. (The dashed curves are obtained similarly by aligning eddy time series at the end of summer). Fig. 5.13 thereby reveals that the composite-average eddy life-time evolution shown in Fig. 5.12 conceals substantial seasonal variations, which are observed in eddy-core SSTA, but not (or much less) in the eddy (SSHA) intensity itself. In particular, SSTA are observed to be substantially weakened at the end of summer (dashed curves around time zero in Fig. 5.13) compared to the end of winter (solid curves near time zero), over both energetic (a) and quiet eddies (b). Quiet regions' eddies are seen to nearly completely lose their SST expressions in late summer, but given the observed large T' fluctuations and comparable amplitude large-scale SSTA residuals seen in Fig. 5.13b, this result needs to be treated with caution. In energetic regions the seasonal loss of the tight eddy SSTA-SSHA connection is robust, and, contrary to previous reports of a preferential summer-time loss of cyclones' SST signals (Auer, 1987; Morrow et al., 2004), observed to be on average as strong over anticyclones as over cyclones.

Apart from the striking (annual-average) SSTA persistence, Fig. 5.12 also shows small time-variations in composite eddy SSTA. On the one hand, $\langle T'(s) \rangle$ displays erratic low-amplitude (typically $\leq 0.05\text{K}$) week-to-week fluctuations, which increase in amplitude throughout the eddy lifetime (as eddy numbers decrease), and, from comparison to the thin stems in the Figure, merely reflect fluctuating residuals of the non eddy-induced, large-scale SST variability at the eddy positions. On the other hand, $\langle T'(s) \rangle$ shows a slow evolution, which is representative of the eddy-induced signal (not seen in large-scale residuals) and is paralleled by the eddy amplitude evolution. In the quiet regime, is characterized by a continuous growth, in the energetic regime, by an initial growth followed by a decay.

To assess how the varying composition of $\langle T'(s) \rangle$ with time affects the latter evolution, Fig. 5.14 shows SSTA and SSHA composite-averaged separately over eddies with lifetimes of 4, 5–8, 9–16 and >16 weeks. This has the advantage that a constant number of eddies enters the respective composites throughout the first weeks/months of the eddy life-time (indicated by solid curves, which change to dashes once numbers decrease). Fig. 5.14a confirms that initial growth and following decay represent a systematic lifetime evolution of energetic eddy amplitudes and SSTA signatures. It moreover reveals that the initial growth is typically stronger and lasts longer over

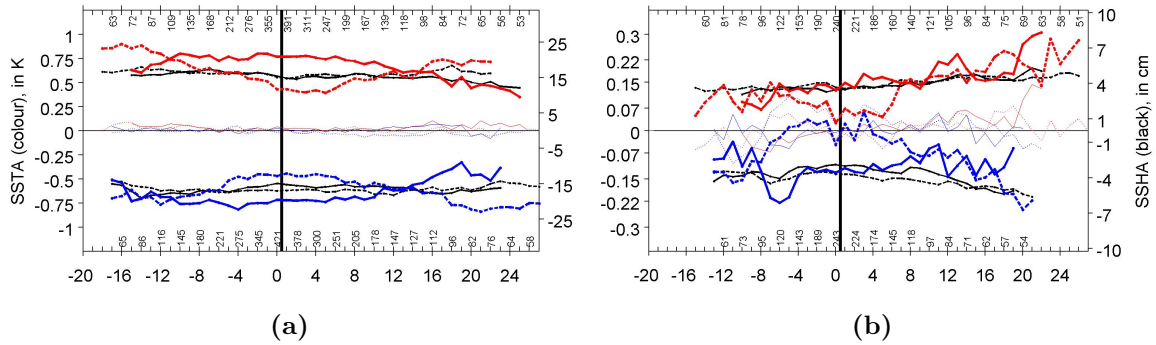


Fig. 5.13: Evolution through time (abscissa, in weeks) of eddy-core SSTA (red for anticyclones/blue for cyclones) and SSHA (black). Before composite-averaging, here individual eddy-core time series are not aligned at the start of their lifetime as in Fig. 5.12, but for the solid curves at the time they transit from the “mixed-layer deepening” to the “mixed-layer shallowing” season, and for the dashed curves at the time they transit from the shallowing to the deepening season. Guided by mixed-layer depth observations in energetic regions, here the deepening season is simply defined as the August-January period in the NA, and as the February-July period in the SO; the shallowing season is given by the remaining months. Negative time on the abscissa thus counts back the weeks from the end of the deepening season (“end of winter”) for the solid curves, and the weeks from the end of the shallowing season (“end of summer”) for the dashed curves. Only composites over at least 50 eddies are displayed, and large-scale SSTA composites are shown by the thin curves (without stems), otherwise same as Fig. 5.12.

longer-lived eddies, so that longer-lived eddies have larger amplitudes, in agreement with results of CSS11, and these longer-lived, larger-amplitude eddies are on average associated with more intense SSTA signals (which is consistent with the positive correlation between eddy lifetime-average amplitude and SSTA seen in Fig. 4.15). Fig. 5.14b indicates that similar features are also characteristic for quiet regions’ eddies, but clearly shows that a larger sampling size is required to robustly isolate these. Here the observed growth seen in eddy amplitude and anticyclones’ SSTA during almost a year likely, at least partly, reflects the increasing contribution to the average of more intense longer-lived eddies, rather than a characteristic intensification of eddy SSTA during the eddy life-time.

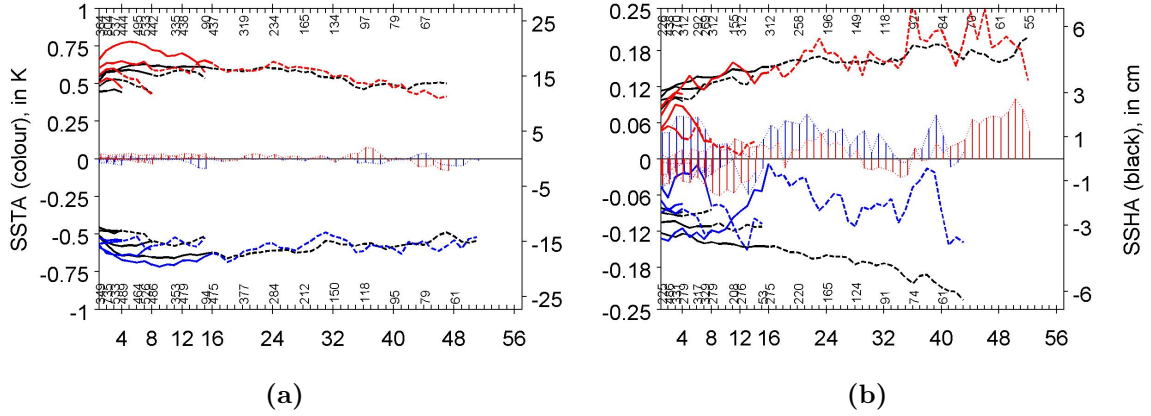


Fig. 5.14: Composite lifetime-evolution of eddy-core SSTA and SSHA displayed separately for eddies with lifetimes of 4, 5–8, 9–16 and >16 weeks. Curves change from solid to dashed once the number of eddies composited starts to decrease (only composites over at least 50 eddies shown). Note the different scale in panel b). Otherwise same as Fig. 5.12.

The typical life-time evolution of eddy SSTA can be better isolated from effects induced by eddy composition changes, by composite-averaging, instead of $T'(s)$ itself, the change of SSTA observed along individual eddy paths $d_s T'(s)$ (in which we have defined d_s to refer to the absolute derivative of a quantity along the eddy trajectory s). Since SSTA cannot change independently from the temperature of the underlying, vertically almost homogeneous mixed-layer, here we directly examine the associated eddy mixed-layer “heat content” variations, to first-order given by $\langle \rho_0 c_p h_{ml} d_s T' \rangle$. In their estimate from observations, we set the heat capacity of sea surface water $\rho_0 c_p$ to $4.1 \cdot 10^6 J/Km^3$, approximate the depth of the mixed-layer by its large-scale monthly climatology $\bar{h}_{ml}$ (Appendix B.1) averaged along the eddy paths, and, as before, obtain T' by averaging over eddy cores. Presented by the black lines in Fig. 5.15, $\langle \rho_0 c_p \bar{h}_{ml} d_s T' \rangle$ act like a magnifying glass for the small systematic evolution of eddy SSTA. Following quiet regions’ eddies (Fig. 5.15c,d), they reach typically only $5 W/m^2$ and can be almost entirely explained by corresponding large-scale (10° -by- 10°) heat content change residuals (indicated by the grey shading). Here, where large spatial scales predominate fixed-location SST variability, the time-evolution of the systematic, but weak eddy SSTA signatures can therefore not be further studied from the present set of observations. Following energetic eddies (Fig. 5.15a,b), to the contrary, $\langle \rho_0 c_p \bar{h}_{ml} d_s T' \rangle$ oscillate between typically $\pm 20 W/m^2$ and clearly stand out from the large-scale residuals, at least for the first 3–4 months, during which more than 500 eddies are available for aver-

aging. Using energetic eddy $\langle \rho_0 c_p \bar{h}_{ml} d_s T' \rangle$, the next section presents some preliminary attempts to shed light on the processes that sustain the observed marked persistence of eddy SSTA. (Further analysis of quiet regions' eddy SSTA is provided in chapter 6.)

5.3.2 Mechanisms

The black curves in Fig. 5.15a,b indicate that during the first month of the energetic regions' eddy life, warm anticyclones on average gain heat and cold cyclones lose heat, corresponding to a growth of their mixed layer heat content anomalies at a rate of $\sim 20 \text{ W/m}^2$. After this initial growth, which is observed to be slightly more pronounced over anticyclones, most of the time anticyclones lose heat and cyclones anomalously gain heat. Such a decay of eddy mixed-layer heat content anomalies is expected from a turbulent heat exchange with the atmospheric boundary layer. Indeed, the reference turbulent heat flux damping (cf. section 5.1) of eddy SSTA $\langle -\alpha_{ref} T' \rangle$, averaged along eddy paths (Fig. 5.15a,b, light grey curves²²), is seen to constantly remove $\sim 20 \text{ W/m}^2$ from the eddies' mixed-layer heat anomalies and thus constitutes an upper bound for their observed decay.

The fact that mixed-layer heat anomaly changes along eddy tracks have, at $\sim 20 \text{ W/m}^2$, physically reasonable orders of magnitude, and that their decay can be explained mechanistically by a damping to the atmosphere, suggests that observed eddy tracks are not inconsistent with representing moving parcels of water. Such an interpretation of tracks has been put forward, at least for long-lived, westward propagating eddies, by CSS11, given the observed high non-linearity of tracked eddies. Indeed in energetic regions, the ratio of eddy rotational to translational speeds U/c is observed to typically exceed 10, i.e. it is certainly $\gg \pi$ as required for eddy cores to trap and carry water (cf. chapter 3.2.4). Moreover quasi-geostrophic simulations (Early et al., 2011) only accurately reproduce observed eddy and SSHA statistics (such as the observed power close to non-dispersive line in the SSHA frequency-wavenumber spectrum) when including non-linear terms, in which case eddy cores are indeed modelled as “perfect” transporters of fluid (i.e. as coherent structures not exchanging parcels with their surroundings). Although an exact closure of the eddy mixed-layer heat budget is not expected from present observations, mainly since the eddy amplitude e-folding radius is not a very good delimiter for the actual water-mass transporting eddy core (expected

²² Note that here the 2002-2007 AMSR-E SST climatology is used instead of the COADS dataset to evaluate α_{ref} from (5.5), but resulting differences are minor.

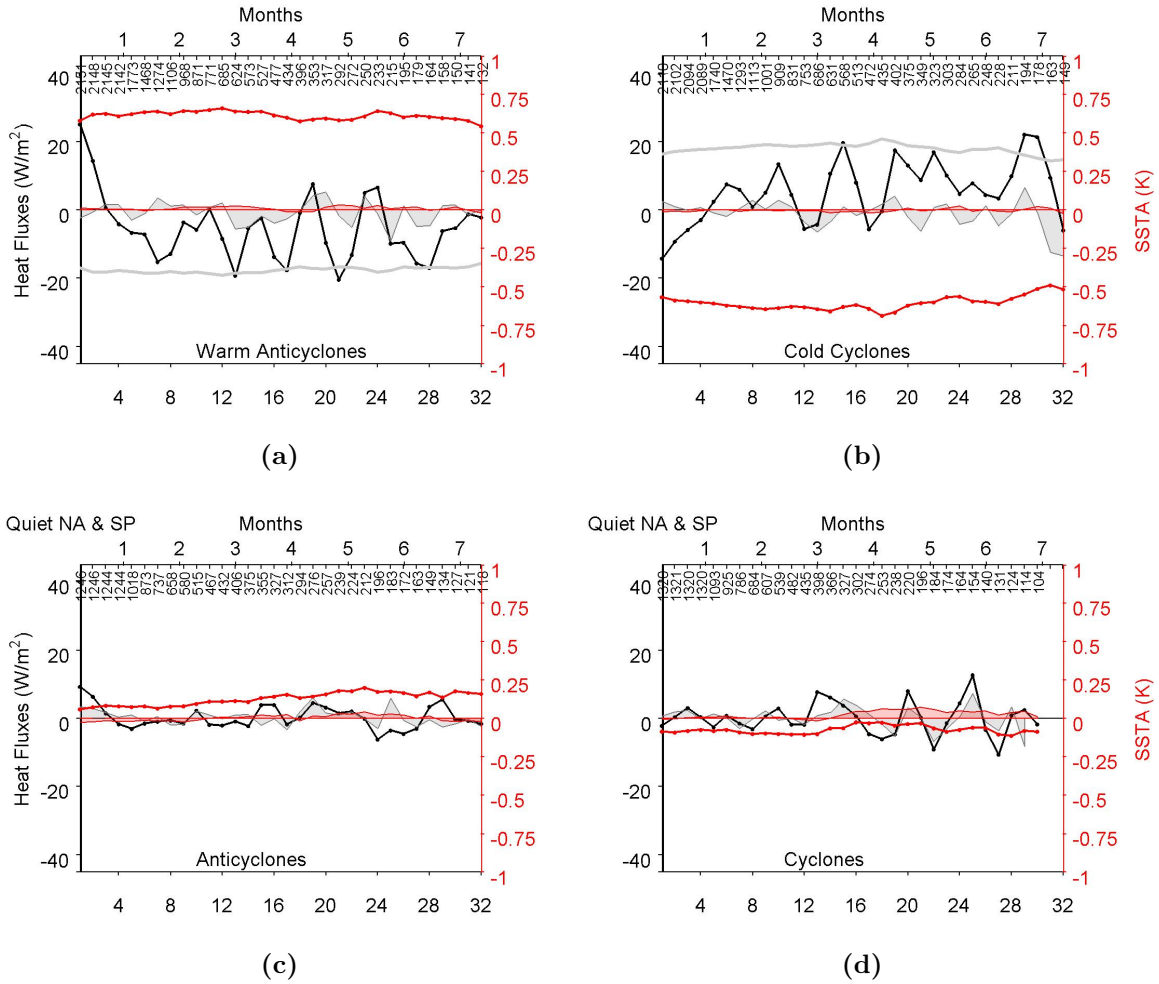


Fig. 5.15: Eddy-core SSTA (red) and associated mixed-layer heat content changes (black) averaged over all anticyclones (left) and cyclones (right) with origins in energetic (top) and quiet regions (bottom), during the first 7 months of their lifetime. Associated large-scale (10° -by- 10°) SSTA and heat content changes indicated in red and grey shading. The light grey lines in a,b) moreover show the reference air-sea heat flux damping $Q'_{ref} = -\alpha_{ref}T'$ averaged along eddy tracks. (Only composites over at least 100 eddies displayed, numbers at top.)

to be anisotropic, and better described by the instantaneous zero relative vorticity contour, Early et al., 2011), it is interesting to ask whether the observed initial growth of the eddy mixed-layer heat anomalies can also be physically explained in this paradigm of tracks?

Insight on this question is provided by taking a closer look on observed eddy propagation statistics, that have been touched upon in chapter 3.3.3. Figure 5.16a re-displays energetic and quiescent regions eddy tracks from Fig. 3.12, but with their initial position removed. In energetic regions (Fig. 5.16a), tracked eddy movements are thus revealed to be erratic, and near-isotropic in zonal and meridional displacements. They thereby stand in stark contrast to the clear westward propagation of tracks with origins in quiet interiors (Fig. 5.16b). Closer inspection of Fig. 5.16a reveals a slight preference of energetic regions' anticyclones (black dashes) to propagate poleward, and of cyclones (grey lines) to propagate equatorward. This is confirmed to be systematic by the dashed curves in Fig. 5.16c that display the composite-average meridional track velocity, obtained directly from observed tracks as $v = dy/dt$. $\langle v \rangle$ is indeed observed poleward over anticyclones (black) and equatorward over cyclones (grey), and typically decreases from only 1 cm/s (0.5 cm/s over cyclones) to near-zero during the first 4 months of the energetic eddy lifetime²³. This poleward/equatorward drift $\langle v \rangle$ (dashed curves) is associated with a systematic composite cross-isotherm track velocity $\langle v_T \rangle$ (thick solid curves in Fig. 5.16c, obtained by projecting each eddy's track velocity on the background SST gradient, i.e. as $v_T = -\mathbf{u} \cdot \nabla \bar{T} / |\nabla \bar{T}|$), where the near-identical $\langle v \rangle$ and $\langle v_T \rangle$ simply reflect that in the average over all GS and ACC eddies background isotherms are near-zonal²⁴. The modification of the background potential vorticity field by intense currents likely plays a role in setting these particular propagation characteristics of energetic eddies, that are reminiscent of the shedding of eddies by meandering jets. In contrast to quiet basin interiors (Fig. 5.16d) where both polarity eddy tracks move on average equatorward (dashes) and towards warmer water (solid curves), in energetic regions, and in their early life, warm anticyclones move poleward and towards colder water, cold cyclones equatorward and towards warmer

²³ Although an energetic track typically covers a net distance $\langle \sqrt{dx^2 + dy^2} \rangle$ of $\approx 20\text{--}30$ km within a week, cancellation of equatorward and poleward, as well as east- and westward eddy movements leads to much shorter *average* zonal and meridional weekly track displacements of $|\langle dy \rangle| \sim |\langle dx \rangle| \sim 5$ km, and thus weak composite zonal and meridional track velocities of ~ 5 km/week ≤ 1 km/day ~ 1 cm/s.

²⁴ Also the seasonality in $\bar{T}$ has hardly any effect, as seen by the near-identical thin curves that display the composite track velocity across the all-time mean surface isotherms $-\mathbf{u} \cdot \nabla \bar{\bar{T}} / |\nabla \bar{\bar{T}}|$

water, and thus on average down the background SST gradient.

From the SSTA equation derived in chapter 2.2, the evolution of $\langle T' \rangle$ (within the climatological mixed-layer $\bar{h}_{ml}$) following coherent eddy cores (for which changes along their path d/ds correspond to changes following the full surface velocity d/dt , is given by:

$$\langle \rho_0 c_p \bar{h}_{ml} d_s T' \rangle = - \langle Q'_{net} \rangle - \langle \rho_0 c_p \bar{h}_{ml} \mathbf{u}' \cdot \nabla \bar{T} \rangle - \langle \rho_0 c_p h'_{ml} d_t T \rangle, \quad (5.11)$$

where the “ $\langle \rangle$ ” highlight that a characteristic eddy signal is only isolated from non-eddy variability after composite-averaging.

Akin to expectations from down-gradient diffusion, the observed down-gradient propagation of tracks is thus, from the second term on the rhs of (5.11), expected to lead to a growth of eddy mixed-layer heat content anomalies, if energetic eddy tracks indeed correspond to the movement of parcels. Particularly, at $v_T \approx 0.5$ cm/s acting on average background SST gradients of 1K/100 km and a mixed-layer of typically 100 m, the growth induced by $-\langle \rho_0 c_p \bar{h}_{ml} \mathbf{u}' \cdot \nabla \bar{T} \rangle$ is expected at a rate of ≈ 20 W/m², so that stirring by the down-gradient eddy drift could support the observed initial growth of energetic eddy SSTA seen in Fig. 5.15a,b.

Complications arise however as, from (5.11), only that component of $\langle v_T \rangle$ (thick curves in Fig. 5.16c) that reflects departures from any systematic time-mean cross-isotherm flow averaged over eddy positions $\langle \bar{v}_T \rangle$ acts on the eddy mixed-layer heat anomaly budget. Given the uncertainty in present mean-flow estimates and their relatively coarse resolution, here we propose to indirectly estimate $\langle \bar{v}_T \rangle$, by simply assuming that any composite cross-isotherm track propagation, averaged over all observed tracks, including *both* anticyclones and cyclones, denoted $[v_T]$ in the following, reflects this mean-flow effect. The anomalous cross-isotherm propagation of tracks $\langle v'_T \rangle$ is then given by departures of anticyclones' and cyclones' $\langle v_T \rangle$ from this all-eddy average.

The resulting indirect estimate of the stirring term $-\langle \rho_0 c_p \bar{h}_{ml} (\langle v_T \rangle - [v_T]) |\nabla \bar{T}| \rangle$ is displayed by the dashed curves in Fig. 5.17. It is seen to indeed lead to a systematic anomalous mixed-layer heat gain/loss over anticyclones/cyclones of ≥ 20 W/m² at the start of the eddy life that decays to near zero within typically 10 weeks. This decay results from a decrease, during the eddy lifetime, of both, the anomalous cross-frontal eddy propagation $v_T - [v_T]$ (by construction antisymmetric between anticyclones and cyclones) and the composite SST gradient (this weakens from ~ 1 to 0.75 K/100 km over eddies of both polarities, which indicates that, in the energetic region average, eddies move away from fronts as they age). Comparison to $\langle \rho_0 c_p \bar{h}_{ml} d_s T' \rangle$ (black curves

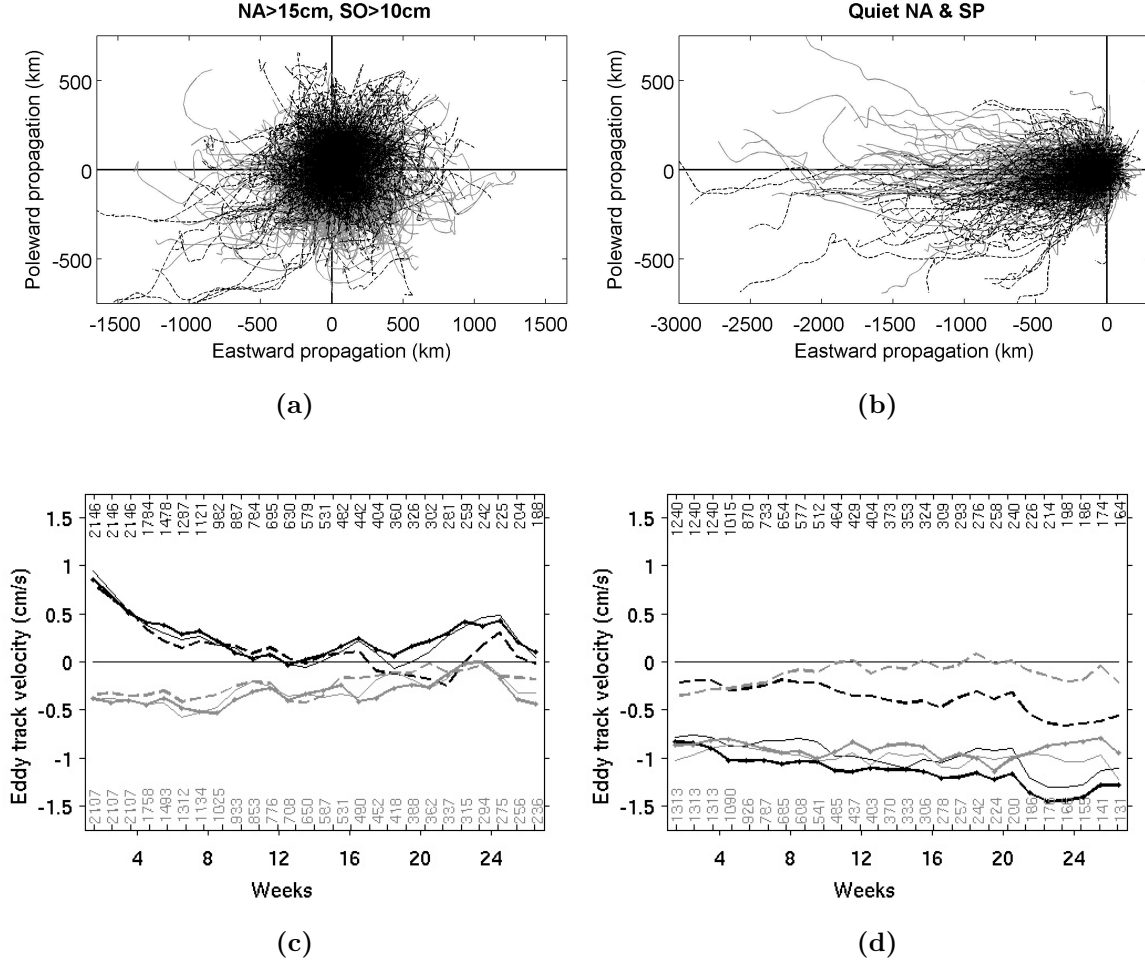


Fig. 5.16: Observed propagation statistics of tracks with origins in energetic (lhs) and quiet regions (rhs) of NA and SO. **Top panels** display $\mathbf{x}(t) - \mathbf{x}(t_0)$ of anticyclonic (black dashes) and cyclonic eddy tracks (grey lines). **Bottom panels** show the meridional (dashed) and cross-isotherm (solid) track velocity composite-averaged throughout the life-time of anticyclones (black) and cyclones (grey), so that positive values indicate propagation towards colder surface waters or higher latitudes. (Here, thin solid lines compare track propagation across time-mean to that across seasonally varying background isotherms (thick), and the numbers of anticyclones & cyclones are indicated at top & bottom of the panels.)

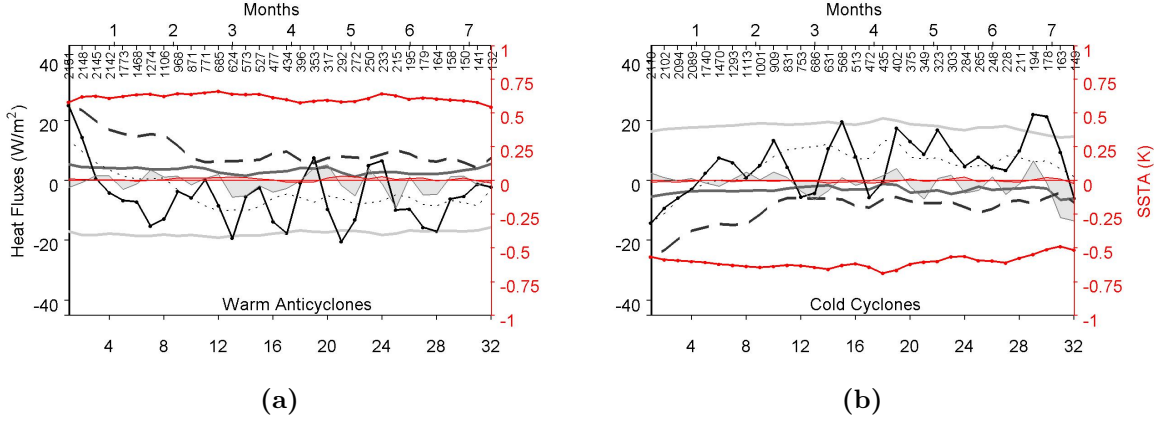


Fig. 5.17: Composite energetic eddy mixed-layer heat anomaly budget: Observed $\langle \rho_0 c_p \bar{h}_{ml} d_s T' \rangle$ (black) compared to anomalous stirring by the eddy drift (dashed, estimated as $-\langle \rho_0 c_p \bar{h}_{ml} (v_T - [v_T]) |\nabla \bar{T}| \rangle$), air-sea heat flux damping Q'_{ref} (light grey) and entrainment feedback Q'_{entr} (dark grey). Otherwise as Fig. 5.15a,b.

in Fig. 5.17) shows that stirring is of sufficient magnitude to support the initial growth of observed eddy mixed-layer heat anomalies. Indeed, it provides an upper bound for their observed growth. As seen by the dotted curves, together, non-local stirring by the anomalous eddy drift and air-sea damping (light grey) give a good account for observed $\langle \rho_0 c_p \bar{h}_{ml} d_s T' \rangle$, only smoothing out their week-to-week fluctuations (which are also seen in the residual large-scale $\langle \rho_0 c_p \bar{h}_{ml} d_s T' \rangle$ and thus likely averaging artefacts): During the first ~ 2 months of the eddy life, stirring more than balances the damping to the atmosphere, thereby generating the observed initial growth of energetic eddy SSTA. Thereafter eddies propagate more passively along isotherms, so that air-sea damping wins over stirring, which leads to the observed decay of eddy heat contents. Bearing in mind the large error bars of the analysis, a competition between stirring and air-sea damping thus provides a possible explanation for the evolution of SSTA following energetic regions' tracks, which is consistent with the latter reflecting the movement of parcels (below the Ekman layer).

In the above estimate $[v_T]$ is weak, suggesting a weak cross-isotherm component of the mean flow. Whereas the geostrophic mean-flow might indeed be approximately aligned with background isotherms in the energetic region average, the observed time-mean westerlies certainly induce a systematic Ekman drift advecting cold water equatorward across the near-zonal background isotherms. Using the SCOW climatology (App. B.3), $\langle \bar{\mathbf{u}}_{ek} \rangle$ and $\langle \bar{\mathbf{u}}_{ek} \cdot \nabla \bar{T} / |\nabla \bar{T}| \rangle$ are estimated at ≈ 1.25 cm/s warm- and equa-

torward. Observed cross-isotherm eddy velocities (Fig. 5.16c) however do not show a net warm-ward offset. The idealized modelling results of Morel and Thomas (2009) may provide some insight in this puzzle. They state that over vortices the Ekman drift is not the only effect of the wind, but that wind-induced potential vorticity anomalies generated in the vortex's sloping isopycnals induce a drift that compensates the Ekman-induced flow for vortices extending up to the surface (as is the case for the tracked energetic eddies). It is thus not unlikely that our indirect estimate of $\langle \overline{v_T} \rangle$ captures the effect of both the direct (Ekman) and the vortex-mediated drift induced by the time-mean westerlies. This compensation effect may be directly connected to the observed eddy-induced transports, whose importance in compensating the wind-induced overturning especially in the SO has been laid out in chapters 1 and 2, and it will be a while until observations become adequate to estimate their combined effect on time-mean transports.

As a further test of the suggested heat budget, we subdivide the track dataset into eddies with lifetimes of $<$ and ≥ 16 weeks. Fig. 5.18 displays the eddy mixed-layer heat anomaly budget for the two groups, which both contain $\approx 50\%$ of energetic regions' tracks. Variations of black and dashed curves in the Figure follow each other, and as these estimates do not directly depend on each other, this provides support for an impact of stirring by the anomalous eddy drift on the eddy heat budget. Specifically, Fig. 5.18 reveals that a near-doubling in magnitude of the anomalous eddy stirring (dashed curves) over long-lived (rhs) compared to short-lived eddies (lhs) could provide an explanation for the larger and longer-lasting growth of eddy mixed-layer heat anomalies (black curves) observed over long-lived eddies. The more vigorous down-gradient propagation of longer-lived eddy tracks thus likely accounts for their larger eddy-core SSTA signatures (Fig. 5.14).

The dotted curves in Figs. 5.17 & 5.18 also include a weak time-mean contribution of an estimate for one component of the anomalous entrainment flux $Q'_{entr} = \rho_0 c_p \overline{w}_{entr} (T'_{sub} - T')$, which is shown by the dark grey curves. In this estimate $\overline{w}_{entr}$ is obtained from the change of the background mixed-layer depth following eddy paths as $\overline{w}_{entr} \approx \Lambda(d_s \overline{h}_{ml})$. T'_{sub} is assumed to be constant, since shielded from air-sea interactions, and, as Q'_{entr} restores eddy SSTA to T'_{sub} while mixed-layers deep, taken to be 0.8 K, which is close to $\langle T' \rangle$ at the end of the deepening season. (This is more physical than relating T'_{sub} to SSTA observed over individual eddies, as in that case large-scale SSTA present at the eddy position cannot be damped out by Q'_{entr} .)

This estimate was originally motivated by the search for an explanation of the

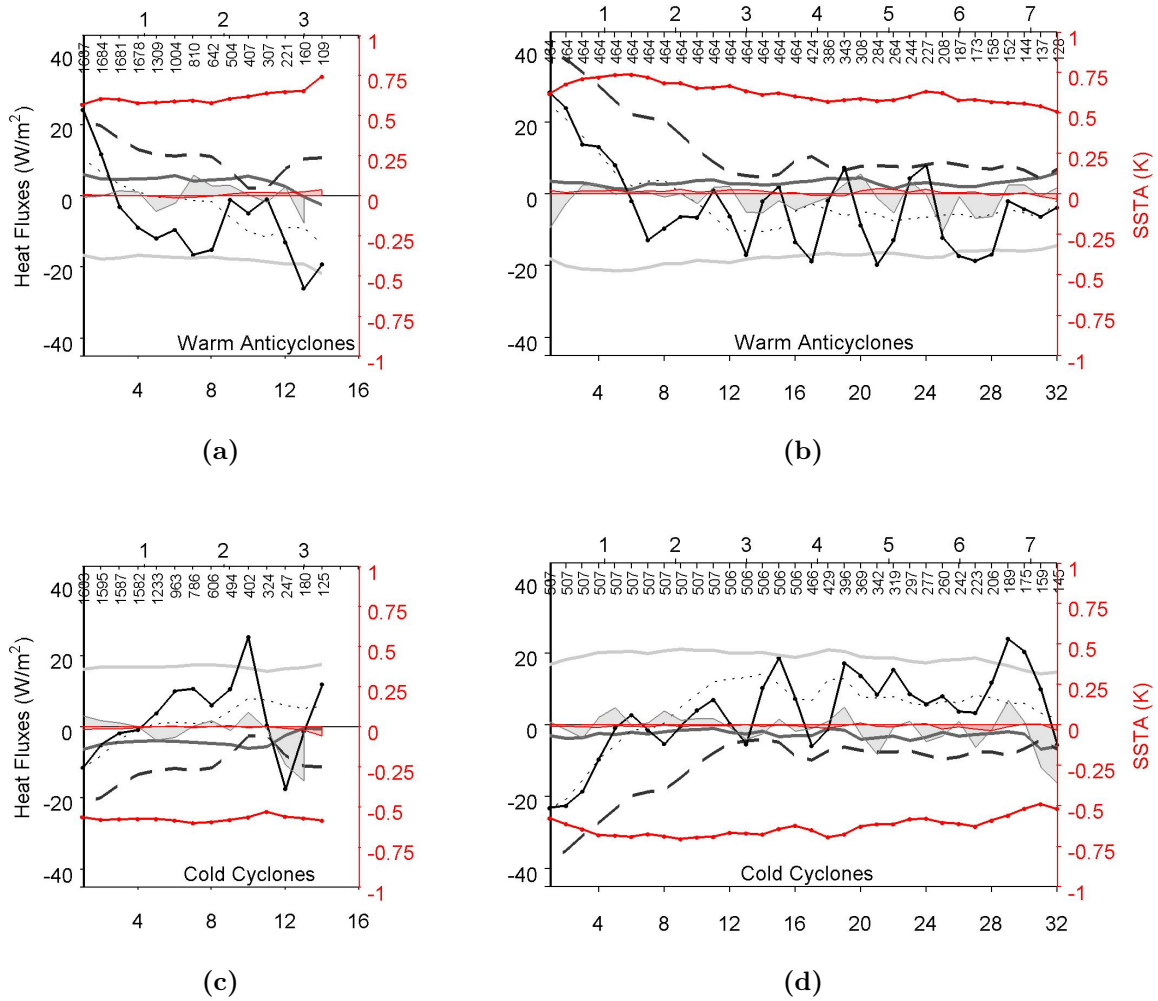


Fig. 5.18: Observed energetic eddy mixed-layer heat anomaly budget. As in Fig. 5.17, but composite-averaged separately over short-lived (< 16 weeks, lhs) and long-lived eddies (≥ 16 weeks, rhs).

observed seasonality in eddy T' (Fig. 5.13). However, as revealed in Fig. 5.19, the observed summer-time loss of the tight SSTA–SSHA relationship (Fig. 5.13) can also be explained by the eddy stirring, rather than solely by the absence of entrainment in the shallowing season. Indeed, the dashed curves in Fig. 5.19 (estimated indirectly as above) reveal a more vigorous and longer-lasting stirring by the cross-isotherm drift in winter (while mixed-layers deepen and background SST fronts are intensified) that sustains the observed growth of eddy mixed-layer heat anomalies (black) against air-sea damping for ≥ 2 months (Fig. 5.19a,b), compared to only ~ 2 –3 weeks in summer (Fig. 5.19c,d). The observed seasonal pulsing of background SST fronts (intensified in winter) is thus suggested to make eddy stirring a key driver of the seasonal modulation

of mesoscale SSTA, observed both along eddy tracks, and in the intensity of their contribution to (fixed-location) SST variability (chapter 4.4.3): Along tracks stirring sustains eddy T' against their damping to the atmosphere, at fixed-location it competes against their masking by large-scale atmospheric forcing. Subject to confirmation by further subsurface analysis, entrainment, and other seasonal processes²⁵ do not play the leading role in this seasonal modulation.

The suggested impact of down-gradient propagation on eddy mixed-layer heat anomalies is consistent with tracks being coherent eddies shed by the stream, such as the intense isolated “rings” described by a multitude of in-situ studies along the GS. As confirmed by a recent study (Sokolov and Rintoul, 2009) frequent shedding of coherent vortices is also observed along the energetic regions of the ACC, specifically in regions down-stream of major bathymetric features, where vortices can be present $\geq 5\%$ of the time in their estimate. However, particularly lower-amplitude, and thus typically shorter-lived (Fig. 5.14a), energetic eddy tracks likely also capture meanders with open circulation cells that contribute significantly to the SSHA variability along GS and ACC. For them an impact of diffusion and damping (the anomalous mixed-layer heat convergence) on the track-following evolution of $T'(s)$ is at most expected during the initial formation period of the meander-like eddy (at a given position along the stream), as thereafter meanders, unlike parcels, merely follow the up- or down-stream phase-propagation of the wave/meander along the jet. The nature of energetic eddy tracks thus remains somewhat open, and assessing and separating the respective contributions of the two groups of eddies to GS and ACC eddy and SST variability remains a subject for further studies. Importantly, such further analysis of energetic regions’ eddy heat anomaly budgets should also carefully investigate the sensitivity of eddy statistics in these regions, to the parameter choices adopted by the eddy identification and tracking algorithm at the basis of the present study (cf. chapter 3.2.4).

²⁵ If eddies induce significant mixed-layer depth anomalies, the term $-\langle \rho_0 c_p h'_{ml} \partial_t \bar{T} \rangle$ in (5.11) provides additional seasonality to mixed-layer eddy heat anomalies. For $h'_{ml} > 0$ over anticyclones, < 0 over cyclones, as expected from the feedback of eddy SSTA on turbulent air-sea heat fluxes, this term leads to a decay of the eddy heat anomaly in spring/summer while mixed layers shoal ($\partial_t \bar{T} > 0$), and to a growth while they deepen, as required to explain observed $\langle \rho_0 c_p \bar{h}_{ml} d_s T' \rangle$. For $|h'_{ml}| \sim 10\text{m}$ and a seasonal SST cycle between 1 and 4 K ($\partial_t \bar{T} \sim 1\text{--}4\text{K}/90\text{d} \sim 0.1\text{--}0.4\text{K}/10\text{day}$), its magnitude is estimated at $4\text{--}15\text{ W/m}^2$. In regions of pronounced seasonal SST variability, this mechanism might thus reveal to be important.

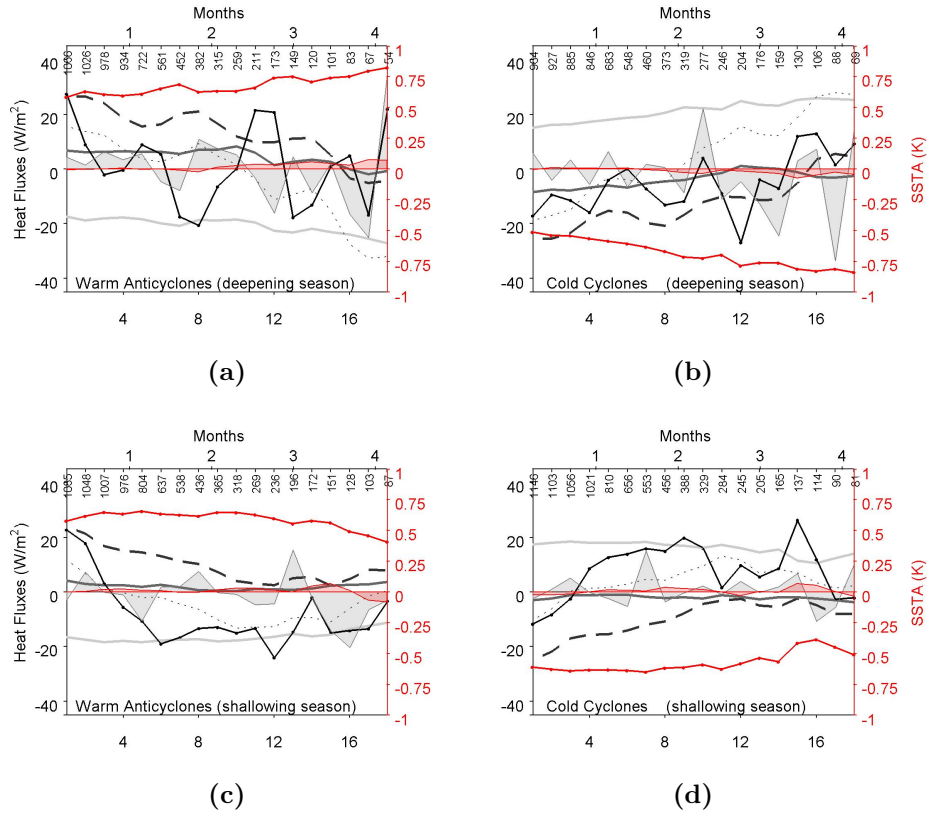


Fig. 5.19: As Fig. 5.17, but displayed separately for energetic eddies that originate in the average mixed-layer deepening season (top, February-July/August-January in SO/NA) and shallowing season (bottom, remaining time of year). Eddies are only composite-averaged as long as they remain within the respective season.

5.4 Conclusions

Following eddy tracks in the SSHA field, reveals that transient mesoscale eddies are systematically associated with SST signatures, which are not quickly erased by air-sea interactions, but instead persist throughout the eddy lifetime, of typically several months. Only composite averaging over many eddy tracks allows isolation of these eddy signals from observational errors (such as those induced by the imperfect separation of non-seasonal from seasonal SST variability as estimated from short time series, cf. Appendix A) and large-scale SST variability at the eddy position. In this composite-average sense, persistent eddy SST signals are a robust observational result throughout ocean basins.

In quiescent eastern subtropics, eddies and their SSTA signals are so weak that any possible systematic evolution along eddy tracks cannot be detected. Instead week-to-

week changes in composite eddy SSTA simply reflect varying residuals of large-scale atmospherically-forced SSTA. The latter predominate fixed-location SST variability in these regions, where therefore at any given location SSTA persistence is simply set by the slow exponential decay (within 2–3 months) expected from large-scale air-sea interactions.

This stands in stark contrast to the regions of elevated eddy variability observed in the vicinity of GS and ACC. Here, at a given location, SSTA memory is quickly erased (within 10–30 days) by self-propagation and mean-flow advection of mesoscale eddies and their associated persistent SST signals. The weak air-sea feedback therefore plays hardly any role in limiting the fixed-location persistence of SSTA. Its time-space statistics are instead set by those of SSHA, highlighting the important role of the oceanic mesoscale in shaping SST variability in major current systems.

Unlike in basin interiors, in energetic regions eddy propagation statistics are reminiscent of diffusion: here anticyclones drift on average poleward, cyclones equatorward, and thus down the background SST gradient associated with GS and ACC jets.

Although these results are preliminary, observations suggest that the anomalous stirring of background isotherms, associated with this diffusion-like propagation observed during the first months of the eddy life, induces a growth of eddy SSTA and associated mixed-layer heat anomalies, the basic mechanism through which ocean “rings” shed by GS and ACC jets appear as anomalous water masses. Here its strength is estimated at $\sim 20 \text{ W/m}^2$ throughout the first months of the eddy life. Estimated eddy-core mixed-layer heat content anomalies, although subject to large error bars (due to the coarse MW-SST resolution, an only approximate definition of eddy cores and missing mesoscale subsurface information), are indeed characterized by an initial growth of $\sim 20 \text{ W/m}^2$. Moreover, tracks that drift more vigorously across SST fronts are characterized by a larger and longer-lasting initial SSTA growth, and larger life-time average SSTA signatures. This suggests that tracked eddies indeed carry water within their cores, and supports a non-local generation of their SSTA signals through stirring by the cross-isotherm eddy drift. Although it does not substantially limit SSTA persistence at fixed-location, air-sea damping, to which T' are subject to after their generation, plays an important role in the life-time evolution of SSTA following their tracks: the competition between eddy stirring and air-sea damping can explain why, after typically only one month into the tracked eddy life, observed eddy mixed-layer heat anomalies start to slowly decay (at $\sim 10 \text{ W/m}^2$). Given the seasonal pulsing of background SST fronts (weakened in summer), stirring and air-sea damping can also account for the

substantial seasonal variation, that is revealed upon closer inspection to characterize eddy T' signatures (weakened in summer) but not eddy (SSHA) amplitudes.

The fast initial growth of eddy SST and SSH perturbations is however also expected from observed linear baroclinic instability, so that energetic regions' SSHA tracks might also capture observed growing and subsequently decaying meanders on GS and ACC jets, which do not carry fluid with them at least in their later-life wave-like propagation along the stream. Dedicated further research into disentangling the contribution of meanders and coherent vortices to energetic regions' SSHA tracks would lead to interesting results on their respective roles in shaping the oceanic mesoscale contribution to SST variability.

Whether these mesoscale SSTA primarily stem from wave-like or coherent transient eddies, they must be associated with a significant modification of the thermal structure in the marine atmospheric boundary layer (or possibly, even in the free tropospheric atmosphere). This thermal modification (adjustment) of the MABL over the oceanic mesoscale is at the origin of the observed dynamic mesoscale ocean-atmosphere coupling (without it the MABL would not know about the mesoscale signal, apart from pure current effects), and in its absence eddy T' signals would be too quickly damped to significantly contribute to SST variability. As established in the first part of this chapter, such a quick air-sea damping, although previously suggested, is indeed inconsistent with air-sea climatology and observed mesoscale air-sea interaction. Instead, and in agreement with recent high-resolution coupled modelling experiments (Bryan et al., 2010) and the observed mesoscale contribution to SST variability (chapter 4), air-sea damping is observed to only moderately increase from the large to the mesoscale, where it is expected to regionally vary from $\sim 25 \text{ W/m}^2$ over the equilibrated ACC, to $\sim 45 \text{ W/m}^2$ over unstable poleward WBCs (cf. section 5.1.5 for a detailed summary). Thereby, the mesoscale air-sea damping lies in a regime in which it is expected to have a key impact on eddy heat transports and permits a substantial impact of the oceanic mesoscale on SST and ocean-atmosphere coupling.

Chapter 6

THE EDDY SIGNATURE IN SST AND THE ASSOCIATED HEAT TRANSPORT

As shown in chapter 4 and 5 following and averaging over many eddy tracks makes it possible to isolate a statistically robust and time-persistent signal of transient mesoscale motions in SST, even away from major current systems where SST variability primarily occurs on large spatial scales. Building on this result, this chapter characterizes the mesoscale eddy SST signature in more detail (section 6.1) and assesses the associated poleward heat transport in the mixed layer (section 6.2). After a discussion of the underlying mechanisms (section 6.3), section 6.4 summarizes conclusions. (The material presented in this chapter is a slight extension to a publication in Deep Sea Research Part I – available on-line at <http://dx.doi.org/10.1016/j.dsr.2012.08.005>).

6.1 The observed SST-SSH relationship following eddy tracks

Chapter 4.3 and 5.3 have provided an analysis of SSTA within the cores of propagating eddies in SSHA (chapter 3.2.4 & CSS11). Following the same eddy tracks originating in the “energetic” and “quiescent” regions of NA and SO (as defined in chapters 3.3.2 & 3.3.3), this section estimates the spatial structure of the eddy-induced SST perturbations. Comparison to fixed-location statistics and previous observational results is provided in sections 6.1.3 and 6.1.4.

6.1.1 “Energetic” regions

As discussed in the previous chapters, a large number of mesoscale eddies with origins in the energetic GS and ACC regions are tracked over the period studied (532 for the GS and 3782 for the ACC). With average tracked lifetime between 3–4 months, this

results in over 25000 weekly snapshots of both anticyclones and cyclones, for which weekly mean SSTA and SSHA data (obtained as described in section 2), centred on the eddy position in time and space, are evaluated. To establish a characteristic SSTA signature of energetic regions' eddies, track-following composite maps of eddy SSTA and SSHA are constructed by averaging over all of the weekly snapshots, along all eddy tracks ¹

The resulting composites are presented in Fig. 6.1a,b, and reveal a very clear signature of GS and ACC eddies in sea surface temperature and sea surface height. Near-isotropic rotating cores are seen in this average, with typical SSHA amplitudes of ≈ 15 cm (in contours). Within these cores, anticyclones are associated with warm SSTA, cyclones with cold SSTA, both with typical amplitudes of ≈ 0.75 K (in colour). This in-phase relationship between SSHA and SSTA is consistent with the strong correlations seen in Fig. 4.4. Warm/cold eddy cores are surrounded by weaker amplitude, negative lobes in both height and temperature anomalies. These are reminiscent of wave-like (meander) structures or may be indicative of densely packed eddies in these energetic regions.

The averaging involved in the construction of eddy composites conceals important SSTA fluctuations between individual eddy events. As discussed in chapter 4.3.1, standard deviations of SSTA between eddy snapshots have the same order of magnitude as the average signal in Fig. 6.1a,b, implying that a given SSTA snapshot does not always reveal the presence of an eddy – a fact highlighted in previous studies (e.g. Halliwell and Mooers, 1979). In the light of these large fluctuations, a t-test provides an indication of the statistical significance of the observed composite SSTA signature. Even after reducing the number of degrees of freedom from the number of eddy events to the number of tracks, to account for correlations, observed composite SSTA are shown to be significantly different from zero at the 99% confidence level throughout the eddy cores and lobes. (The significance is only found to fall below the 99% confidence level (dotted in Fig. 6.1) in the transition region between positive and negative values.)

A closer inspection of Fig. 6.1 reveals that eddy SSTA and SSHA are not exactly in phase. For both anticyclones and cyclones, maximum SSTA occur slightly to the west of eddy maxima in SSHA. Over anticyclones, maximum SSTA are also shifted slightly

¹ Before averaging, individual snapshots of SSTA and SSHA are normalized by the instantaneous eddy radius, which is defined by CSS11 as the radius of maximum rotational eddy speed and is typically on the order of 80 km. A normalized grid spacing of $\frac{1}{2}$ eddy radii is chosen, which is comparable or slightly smaller than the $\frac{1}{4}^\circ$ AMSR-E SST resolution.

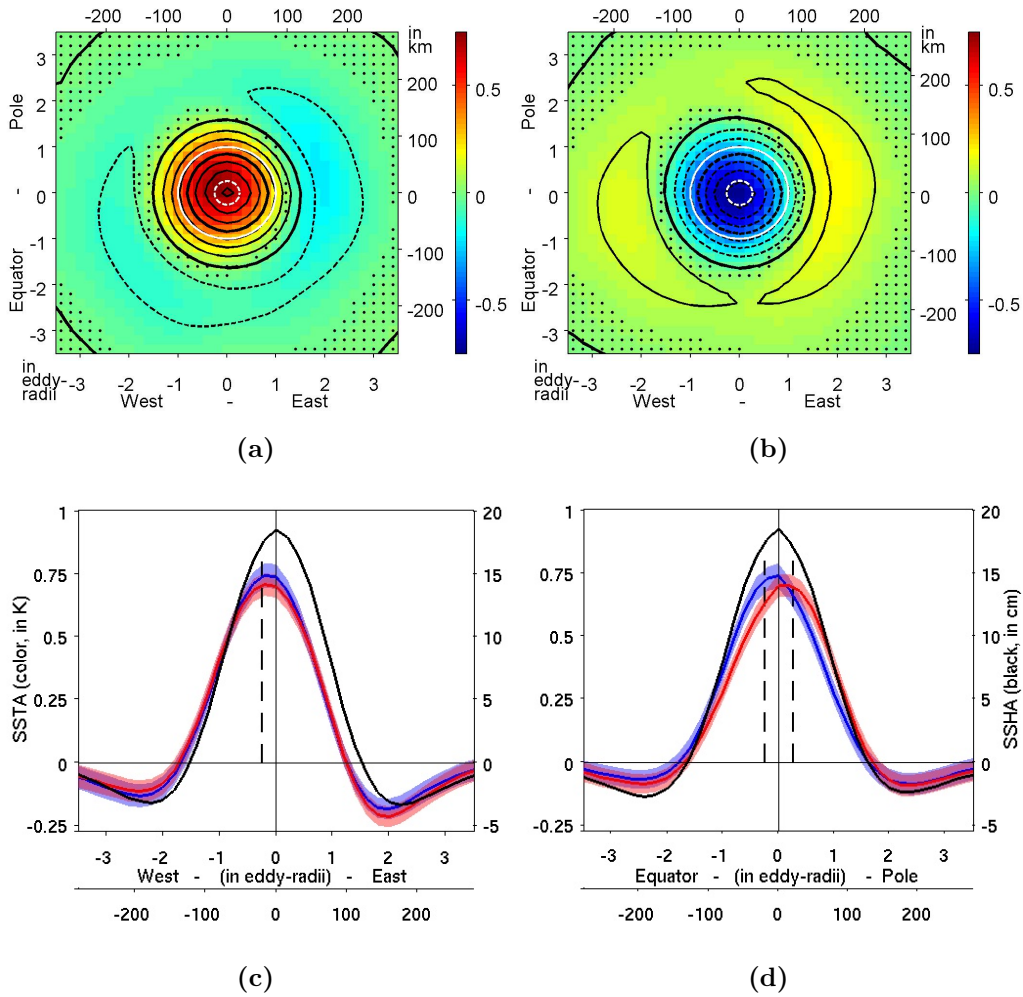


Fig. 6.1: Track-following composites of energetic regions' eddy SSTA and SSHA. Maps for anticyclones (a) and cyclones (b) colour SSTA and contour SSHA (every 2.5 cm with 0 & 10 cm in thick, negative values dotted). Zonal (c) and meridional (d) sections through the eddy centre highlight phase shifts of SSTA (red/blue for anticyclones/cyclones, cyclonic sections multiply by -1) compared to SSHA (black, for weighted both-polarity average). A composite t-test (see text for details of the statistics) indicates where SSTA are significantly different from zero at the 99% confidence level (non-dotted areas in a-b), the corresponding confidence intervals are shaded in the eddy SSTA sections (in c,d). Composites are normalized onto an eddy-radius grid (unit-circle in white), but composite average deformation radius (dashed in all panels) and absolute scales (km) are shown for reference.

Tab. 6.1: Variability of the observed westward phase shift.

	Energetic			Quiet		
	ALL	GS	ACC	ALL	NA	SP
ALL	0.73/ -0.2	0.76/ -0.2	0.72/ -0.2	0.15/ -0.6	0.10/ -0.6	0.19/ -0.6
A	0.71/ -0.2	0.65/ -0.4	0.72/ -0.2	0.16/ -0.6	0.10/ -0.8	0.20/ -0.6
C	0.74/ -0.2	0.88/ -0.2	0.72/ -0.2	0.14/ -0.6	0.10/ -0.6	0.18/ -0.6
short	0.69/ -0.2	0.69/ -0.2	0.69/ -0.2	0.11/ -0.6	0.05/ -0.6	0.14/ -0.6
long	0.76/ -0.2	0.82/ -0.2	0.75/ -0.2	0.18/ -0.6	0.13/ -0.8	0.23/ -0.6

Value/position of the maximum SSTA (in K/in eddy radii) along the east-west section through the centre of the composite over all eddies within the indicated groups.

A: anticyclones, C: cyclones, short/long: eddies with lifetimes $\leq/> 16$ weeks

poleward, and over cyclones, slightly equatorward. Zonal and meridional sections through the centre of the eddy composites (Fig. 6.1c-d) clearly show these phase shifts, and indicate that they are typically on the order of the local deformation radius ($R_d \approx 20$ km, as indicated by the dashed lines in Fig. 6.1a-d, and obtained by averaging the deformation radii calculated by Chelton et al. (1998) over the energetic regions).

To test whether this shift is statistically significant, we have repeated the composite analysis over several subsets of the original data (GS, SO, short and long lived eddies as indicated in Table 6.1). Despite resulting differences in SSTA amplitudes, subset-based composites all reveal a systematic westward shift, in the same direction as observed in the all-eddy composite, on the order of the local deformation radius, or $\approx 1/5$ of the eddy radius. In addition, we have also tested whether SSTA on the east side of the eddy ($x > 0$) were significantly different from SSTA on the west side ($x < 0$) of the eddy² by producing, rather than a composite for SSTA(x) as in Fig. 6.1, a composite for SSTA(x)-SSTA($-x$). As can be seen in Fig. 6.2, western and eastern eddy SSTA are statistically different from each other at the 99% confidence level everywhere within the eddy cores (dotted elsewhere, calculated using a t-test as in Fig. 6.1a-b) for both anticyclones (Fig. 6.2a) and cyclones (Fig. 6.2b). The slight westward phase shift between the surface temperature and pressure of an eddy is thus a robust feature of the observations used here.

² The eddy centre is defined by the eddy position in CSS11's dataset and observed to coincide with the maximum in composite SSA. East and west refer to distances x along a constant latitude from that centre.

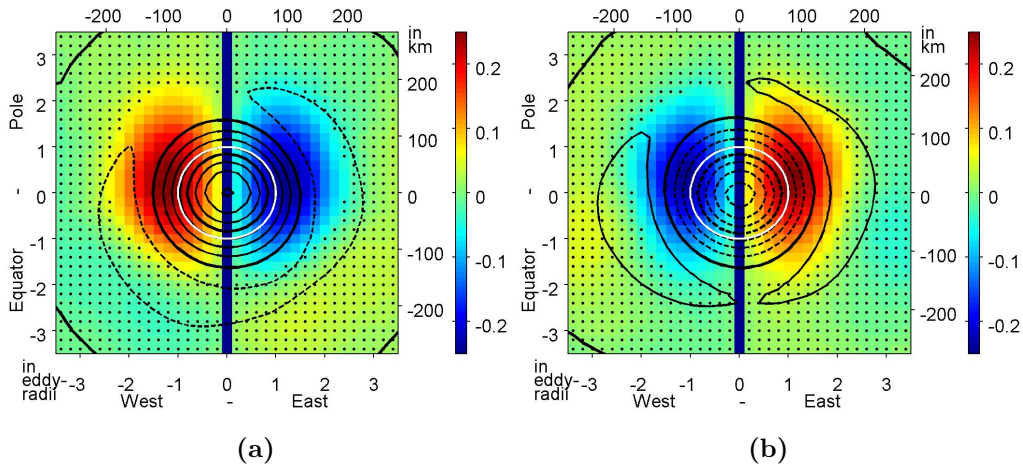


Fig. 6.2: Zonal asymmetry of energetic regions' eddy SSTA composite: $SSTA(x) - SSTA(-x)$. Non-dotted areas indicate where eastern and western composite SSTA are significantly different from each other at the 99% confidence level. Otherwise, apart from a different colour scale for the zonal SSTA difference, same as Fig. 6.1.

6.1.2 “Quiescent” regions

In quiescent regions composites of SSTA and SSHA are produced in the same way as described above for energetic eddies – more than 1500 and 1000 eddies originating in the interior of the South Pacific (SP) and North Atlantic (NA), respectively, were followed for this purpose. Composites are displayed separately for anticyclones and cyclones in Fig. 6.3a-b. The composite of SSHA is similar in shape and radius to that of the energetic eddies (Fig. 6.1a-b), but its amplitude (< 5 cm) is reduced by more than a factor of 4. The SSTA composite reveals a systematic but rather weak signature (reduced by a factor of 5 compared to the eddies' in energetic regions), with warm anomalies stronger over anticyclones and cold anomalies stronger over cyclones. Compared to Fig. 6.1a-b though, the SSTA signature is markedly more bipolar, and almost in quadrature with SSHA, rather than primarily in-phase. Although the anomalies in Fig. 6.3a-b are smaller than those in Fig. 6.1a-b, a t-test asserts that these features are statistically significant at the 99% level.

The phase shifts in both zonal and meridional directions are much larger in Fig. 6.3a-b than in Fig. 6.1a-b. This is clearly seen in the zonal and meridional sections through the composites provided in panels c-d of the Figures. Indeed, the systematic westward zonal phase shift is observed to reach more than half of the eddy radius r in quiet regions (or 1.5 times the local deformation radius). Table 1 indicates that this result

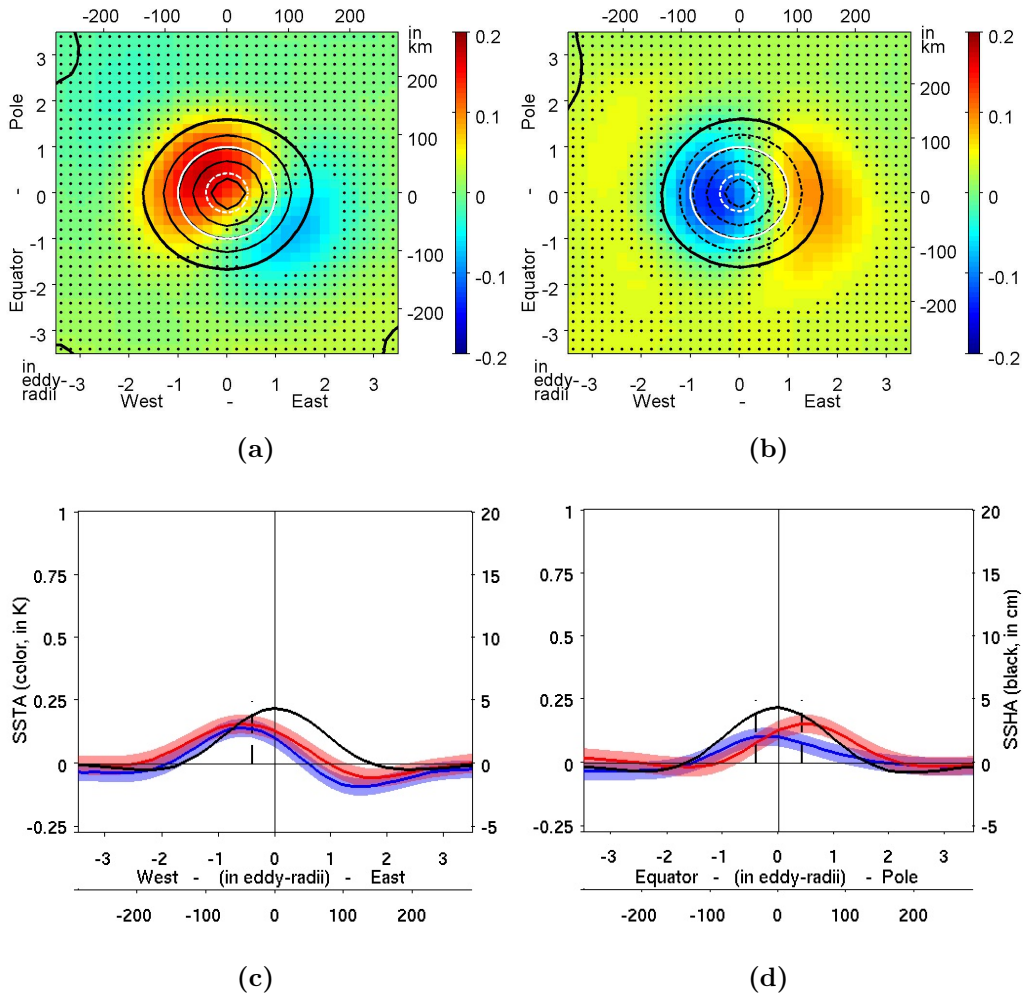


Fig. 6.3: Track-following composites of quiet regions' eddy SSTA and SSHA. Note the different SSTA colour-scale (± 0.2 K) and SSHA contour-intervals (1 cm, with 0 & 5 cm in thick) compared to the composite of energetic regions' eddies, otherwise same as Fig. 6.1.

holds for all subsets of the eddies considered, with maximum SSTA consistently shifted to the west of eddy centres by about $0.6 r$.

6.1.3 Link to fixed-location statistics

Figure 6.4 displays time-lagged SSHA-SSTA cross-correlations, observed at fixed locations in energetic (solid curve) and quiescent (dashed) regions. In contrast to the robust westward phase shifts between eddy SSTA and SSHA observed in the moving frame of eddy tracks, fixed-location cross-correlations consistently peak at zero time-lag. In energetic regions they are symmetric about a strong peak at zero lag,

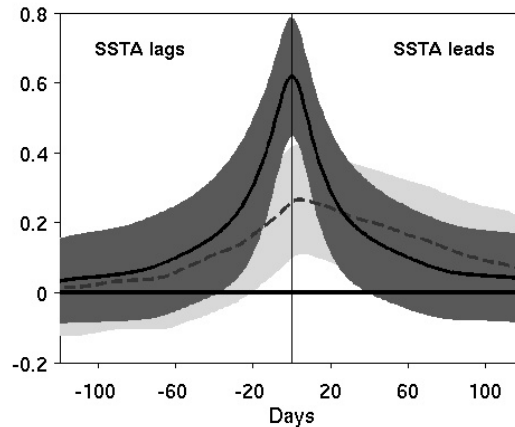


Fig. 6.4: Cross-correlation between SSHA and SSTA observed at fixed locations and bin-averaged separately over all data points in the energetic (solid line) and quiet regions (dashed line) of NA and SO. Shadings indicate the standard deviation of cross-correlations within these regional bins (dark for energetic, light for quiet regions).

in quiet regions the zero-lag peak is weaker and correlations are higher when SSTA leads (positive lags) than when it lags SSHA. To understand this discrepancy between fixed-location and track-following observations, the signature propagating eddies leave at a fixed-location needs to be revisited.

In quiet regions tracked eddies propagate, with little meridional deflection, nearly monotonically westward (Fig. 5.16b). From the eddy signal alone, SSTA-SSHA cross-correlations are therefore expected to peak when SSTA leads, at the positive time-lag $\tau = -\delta_x/c_{lrw}$ determined by the observed westward phase shift $\delta_x \approx -50km$ (Fig. 6.3) and the eddy propagation speed, which is observed to differ on average by less than 20% from the local long baroclinic Rossby wave phase speed c_{lrw} (CSR11, their Fig. 22). Cross-correlations (Fig. 6.4, dashed curve) are indeed higher over the expected range for τ ($\sim +16$ days in the NA, $+50$ days in the higher latitude SP quiet regions) than over the corresponding negative lags. However, as shown in chapter 4, in quiet regions SST variability is dominated by large-scale processes, and steric contributions to the weak SSH variability are important, so that eddy signals can only be isolated after substantial averaging/filtering. It is therefore expected that here, rather than the eddy signal, cross-correlations primarily capture large-scale signals in SST and the associated time-delayed steric response in SSH, reflecting the slow propagation of warming/cooling signals down through the water column.

In energetic regions, mesoscale variability dominates SSHA (for mixed layer depths

of 50 m, typical SSTA of 1K result in steric SSHA of less than 1 cm, or an order of magnitude smaller than observed SSHA rms). The strong in-phase relationship of surface temperature and height perturbations, observed here at fixed-locations (Fig. 6.4, solid curve), therefore reflects the passing of warm-top anticyclones and cold-top cyclones as seen in Fig. 6.1. As energetic regions' eddy movements are complicated, nearly isotropic in zonal and meridional displacements (Fig. 5.16a), here eddies propagate both westward and, when advected by GS and ACC, eastward across a given location. This causes westward shifted eddy SSTA to at times lead, at times lag eddy SSHA in time at a given location, and leads to the observed zero time-lag peak of SSHA-SSTA cross-correlations (Fig. 6.4, solid curve). The systematic small phase shift observed following energetic eddies (Fig. 6.1) therefore cannot be captured by the fixed-location analysis, which highlights the advantages of using a track-following analysis framework to detect systematic eddy signatures.

6.1.4 Link to previous studies

Apart from studies following rings in infra-red SST observations, which all implicitly assume an in-phase relationship between temperature and pressure fluctuations (Brown et al., 1986; Auer, 1987; Lentini et al., 2002), there exists, to our knowledge, no systematic track-following analysis of eddy SSTA signatures close to the world's major current systems and associated vigorous eddy fields. We thus compare our results with those published previously in regions of weak and moderate eddy activity.

Systematic phase shifts between eddy surface and subsurface temperature fields were perhaps first observed by Voorhis et al. (1976) during the MODE experiment conducted in the NA subtropical convergence zone, a region of intermediate eddy activity. Later, several analyses of fixed-location SSTA and SSHA observations (exploiting, instead of fixed-location time-lagged correlations as discussed above, spatio-temporal lag correlations and coherence analysis of wavenumber-frequency spectra), have, with the exception of Cipollini et al. (1997), all detected westward phase shifts of mesoscale SSTA centres with respect to pressure centres: Lower-resolution observations show relatively large phase shifts at larger scales (wavelengths ≈ 800 km) than those studied here (composite eddy wavelengths vary between 400 and 600 km), both in the NA STCZ (Halliwell et al., 1991b) and in a global analysis (Leeuwenburgh and Stammer, 2001). The latter study furthermore finds evidence of decreasing shifts with increasing latitude. For the region between 20°N and 40°N in the Pacific and from observations of comparable resolution to those analysed here, Qiu and Chen (2005) reveal phase

shifts between SSH and SST fluctuations on the order of the deformation radius, which in terms of magnitude, compare well with our observations in GS and ACC.

The combination of satellite altimetry and in-situ data has provided invaluable insight into the temperature and velocity distribution of mesoscale motions. Qiu and Chen (2005) use multi-mission altimetry to identify two eddies just south of the Kuroshio, whose temperature and salinity structure through depth is sampled by Argo floats and shows near-surface westward phase shifts of temperature with respect to pressure that are larger than the eddy radius r . The latter are consistent in direction, but not in magnitude, to those observed here, but it is important to emphasize the very small sample size of Qiu and Chen (two eddies) compared to ours (≥ 2500 in quiet and ≥ 4200 in energetic regions). It is possible that this discrepancy reflects a contamination of the mesoscale SST signal in Qiu and Chen (2005) by large scale processes, while our much larger sampling allows us to average out such effects. Finally, in a comprehensive study of over 410 intense eddy events detected in a repeat XBT transect through regions of weak to moderate eddy activity in the subtropical North Pacific, Roemmich and Gilson (2001) observe a systematic westward phase tilt of eddy temperature cores with decreasing depth. At the surface the westward phase shift reaches 0.4° longitude for eddy radii r of 1° longitude, or $\frac{2}{3} r$. This observation in regions of moderate eddy activity lies between our findings of shifts larger than $\frac{1}{2} r$ (typically $\frac{3}{5} r$) in quiet regions, and smaller than $\frac{1}{2} r$ (typically $\frac{1}{5} r$) in energetic regions.

6.2 Eddy mixed-layer heat transports

In this section, we examine the information the eddy signatures in SST, established above, carry on the mixed-layer eddy heat transport and its mechanisms, focussing purely on its poleward component. After introducing a framework for estimation the transport of heat by propagating eddies (section 6.2.1), sections 6.2.2, 6.2.3 and 6.2.4 compute it from observations. Finally, sections 6.2.5 and 6.2.6 estimate the associated mixed-layer eddy diffusivities for heat and provide insight into their spatial variability.

6.2.1 Framework for estimating heat transports

At a given latitude y , the poleward heat transport H throughout the mixed layer across a zonal extent L_x is given by

$$H(y) = \int_0^{L_x} \rho_0 c_p h_{ml} v T dx, \quad (6.1)$$

where $\rho_0 = 1025 \text{ kg m}^{-3}$ is an average density, $c_p = 4000 \text{ J kg}^{-1} \text{ K}^{-1}$ is the specific heat capacity of seawater, v is the meridional velocity and T is the temperature averaged over the mixed layer of depth h_{ml} .

Propagating mesoscale features contribute to H through correlations between the perturbations they induce in temperature, δT , and meridional velocity, δv . These occur if coherent meridional eddy movements (with velocities v_d) are accompanied by a systematic difference in δT between poleward and equatorward drifting eddies. This gives rise to a component of H referred to in the following as “drift heat transport”. Whereas this “drift” transport is typically focussed upon by observational studies of ocean “rings” (e.g. Ring Group, 1981; Joyce et al., 1981; Peterson et al., 1982; Morrow et al., 2004; Swart et al., 2008), it is only one component of the eddy heat transport. Heat is also transported by correlations between δT and $(\delta v - v_d)$. These occur if there is a temperature difference between the poleward and equatorward moving sides of each individual eddy, and give rise to a component of H which we will refer to as “swirl heat transport”.

To quantify these two components we consider what (6.1) would be if a single eddy contributed to the heat transport. To do so we decompose the temperature and velocity field into a background, or environmental component, and a perturbation associated with the eddy:

$$T = T_{env} + \delta T \quad \text{and} \quad v = v_{env} + \delta v. \quad (6.2)$$

Since we are considering the case where only the eddy contributes to the heat transport, we write $v_{env} = 0$ and $H = \delta H$. Further introducing the meridional drift velocity v_d of the eddy crossing the latitude in question by writing $\delta v = v_d + (\delta v - v_d)$, the contribution of a single eddy to (6.1) becomes:

$$\begin{aligned} \delta H &= \int_0^{L_x} \rho_0 c_p h_{ml} (\delta v - v_d) T dx & + \int_0^{L_x} \rho_0 c_p h_{ml} v_d T dx \\ &= \delta H_{swirl} & + \delta H_{drift}. \end{aligned} \quad (6.3)$$

The time-averaged net contribution of propagating mesoscale eddies to $H(y)$ is then obtained by summing (6.3) over all eddies $n(t)$ crossing L_x at a given time, and averaging in time:

$$H(y) = \overline{\sum_{i=1}^{n(t)} \delta H(i)} = \bar{n} \langle \delta H \rangle, \quad (6.4)$$

where $\langle \rangle$ denotes the composite average over all eddies at all times, and the overline the time-average. Similarly, the two components of heat transport can be written as:

$$H_{swirl}(y) = \bar{n} \langle \delta H_{swirl} \rangle \quad \text{and} \quad H_{drift}(y) = \bar{n} \langle \delta H_{drift} \rangle. \quad (6.5)$$

The requirement of zero time-average perturbation flow across L_x , $\langle \delta v \rangle = 0$, by noting that the swirl component of the flow ($\delta v - v_d$) is non-divergent across each eddy, implies that $\langle v_d \rangle = 0$ (which could simply reflect the same number of eddies drifting poleward and equatorward across L_x on average). This allows us to rewrite (6.5) as:

$$H_{swirl}(y) = \bar{n} \left\langle \int_0^{L_x} \rho_0 c_p h_{ml} (\delta v - v_d) \delta T dx \right\rangle + \epsilon \quad \text{and} \quad H_{drift}(y) = \bar{n} \left\langle \int_0^{L_x} \rho_0 c_p h_{ml} v_d \delta T dx \right\rangle + \epsilon, \quad (6.6)$$

as terms involving T_{env} vanish (to a residual ϵ due to correlations between δv and differences in T_{env} between individual eddies, which is neglected in the following).

We now apply (6.6) to observations, using the composites produced in the previous section (6.1) to isolate the eddy perturbations δT and δv , and using the eddy tracks as a proxy for the eddy drift velocity v_d . Note that several further approximations need to be made before doing so:

- Firstly, contributions to H by correlated inter-eddy fluctuations around all-eddy averages will not be computed (e.g., $\langle \delta v \delta T \rangle$ will be replaced by $\langle \delta v \rangle \langle \delta T \rangle$).
- Secondly, the mixed-layer depth in the above equations will be approximated by its large-scale seasonal climatology $\bar{h}_{ml}$ (Appendix B.1), which is evaluated and composite-averaged along eddy tracks. From the feedback of eddy δT on turbulent air-sea heat fluxes and thus convection (Dewar, 1986; Williams, 1988), eddy perturbations in mixed-layer depth are expected to be of opposing sign over anticyclones (enhanced air-sea heat loss and convective deepening) and cyclones (anomalous heat gain and restratification), so that it is not unreasonable that the averaged effect is captured by our calculation.
- Furthermore, we note that the regional bins, over which eddy composites have been estimated, span a finite meridional extent L_y (cf. chapter 3.3.2). In the following, we therefore provide an estimate for the contribution of propagating eddies to the

meridional-average poleward heat transport across these regions. Denoted H , this is simply obtained from (6.1) as:

$$H = \int_0^{L_y} H(y) dy / \int_0^{L_y} dy. \quad (6.7)$$

We now introduce $n_y \equiv L_y / \langle \lambda_y \rangle$, in which $\langle \lambda_y \rangle$ is the meridional extent of a composite-average eddy. With this, to the level of approximation and using the simplifications outlined above, (6.7) becomes:

$$H = \bar{n} \left\langle \int_0^{L_y} H(y) dy / \int_0^{L_y} dy \right\rangle = \frac{\bar{n}}{n_y} \left\langle \frac{1}{\lambda_y} \int_{\lambda_y} H(y) dy \right\rangle \equiv \frac{\bar{n}}{n_y} \langle \delta H \rangle. \quad (6.8)$$

$\bar{n}$ now refers to the time-average number of coexisting eddies in the regional bin considered, and, for brevity and use in the following, we have defined $\langle \delta H \rangle$ to denote the meridional-average heat transport carried by a composite eddy within $\langle \lambda_y \rangle$.

Combining all the above, we obtain:

$$H = H_{swirl} + H_{drift} = \frac{\bar{n}}{n_y} (\langle \delta H_{swirl} \rangle + \langle \delta H_{drift} \rangle), \quad (6.9)$$

with

$$\langle \delta H_{swirl} \rangle = \frac{\rho_0 c_p \langle \bar{h}_{ml} \rangle}{\langle \lambda_y \rangle} \int_{\langle \lambda_y \rangle} \int_{\langle \lambda_x \rangle} \langle \delta v - v_d \rangle \langle \delta T \rangle dx dy \quad (6.10)$$

and

$$\langle \delta H_{drift} \rangle = \frac{\rho_0 c_p \langle \bar{h}_{ml} \rangle}{\langle \lambda_y \rangle} \int_{\langle \lambda_y \rangle} \int_{\langle \lambda_x \rangle} \langle v_d \rangle \langle \delta T \rangle dx dy, \quad (6.11)$$

to be estimated in turn from observations in the following sections.

6.2.2 Swirl heat transport

The observed systematic meridional and zonal phase shifts of eddy SST signatures with respect to eddy pressure cores, established above (Figs. 6.1 & 6.3), imply correlations between SSTA and eddy geostrophic velocity anomalies. Integrated across closed eddy streamlines (to assure zero mass-transport), the composite anticyclone (characterized by a westward and poleward shift) therefore transports heat poleward and eastward, the cyclone (westward and equatorward shift) poleward and westward. Whereas zonal heat transports could cancel, poleward heat transports of cyclones and anticyclones add up.

To quantify the mixed-layer poleward heat transport thus implied by the observed westward phase shift between eddy SSTA and SSHA, the swirl meridional velocity

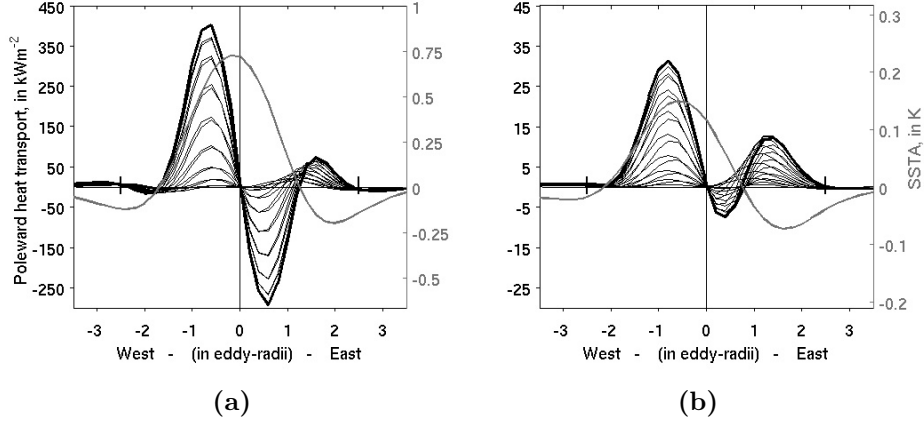


Fig. 6.5: Poleward heat transport at the surface due to eddy swirl, $\rho c_p \langle v'_g \rangle \langle T' \rangle$, observed along the zonal section through the eddy centre (thick black) and at other eddy latitudes (thin black), in energetic (a) and quiescent regions (b). Note the different scales. Composite eddy SSTA along the zonal section through the eddy centre is also shown (as both-polarity average, after multiplying cyclonic sections by -1).

field in (6.10) is estimated, using geostrophy, from the distribution of SSHA in the composites displayed in Figs. 6.1a-b (energetic regions) and 6.3a-b (quiescent regions). That is, we use $\langle \delta v - v_d \rangle = \langle v'_g \rangle \equiv \left\langle \frac{g}{f_0} \frac{\partial \text{SSHA}}{\partial x} \right\rangle$, where $g = 9.81 \text{ m}^2 \text{ s}^{-2}$ is gravity and f_0 is the local value of the Coriolis parameter. The temperature perturbation $\langle \delta T \rangle$ in (6.10) is likewise estimated from the SSTA composites $\langle T' \rangle$ shown in the same Figures. Using these composites and the assumptions discussed in section 5.1, the swirl heat transport due to an average eddy (6.10) is estimated as

$$\langle \delta H_{\text{swirl}} \rangle = \frac{\rho_0 c_p \langle \bar{h}_{ml} \rangle}{\langle \lambda_y \rangle} \int_{\langle \lambda_y \rangle} \int_{\langle \lambda_x \rangle} \langle v'_g \rangle \langle T' \rangle dx dy. \quad (6.12)$$

Local surface transports $\rho_0 c_p \langle v'_g \rangle \langle T' \rangle$ are displayed in Fig. 6.5 as function of longitude, and for different eddy latitudes, and after furthermore averaging transports obtained separately for the composite anticyclone (i.e. from panels a of Figures 6.1 & 6.3) and the composite cyclone (panels b). The thick black curves display transports observed along the zonal section through the eddy centre. They show that in both energetic (Fig. 6.5a) and quiescent regions (Fig. 6.5b), surface transports peak on the western side of the eddy centre. In energetic regions this peak in poleward heat transport reaches more than 300 kW m^{-2} , an order of magnitude more than in quiescent regions, where transport peaks at roughly 30 kW m^{-2} . Over the nearly in-phase

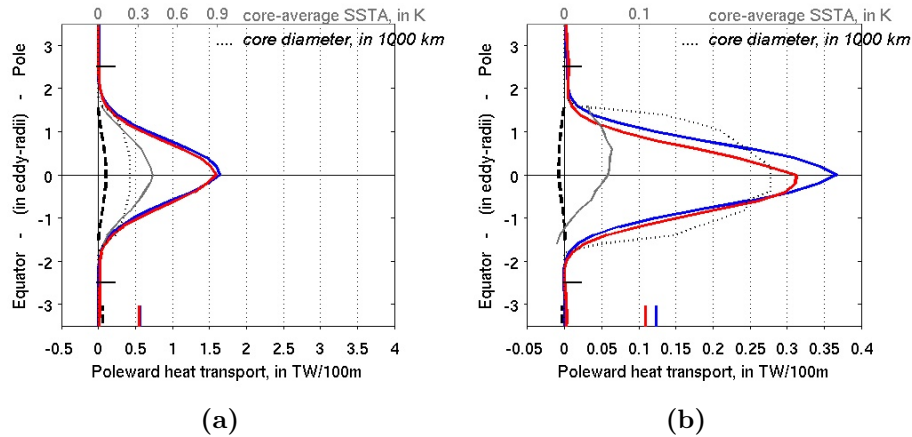


Fig. 6.6: Surface poleward heat transport by a single composite eddy in (a) energetic and (b) quiescent regions, zonally-integrated over the eddy wavelength λ_x , displayed in 10^{10} W/m, or equivalently in 1 TW/100 m, and as function of eddy latitude. The eddy swirl component $\int_{\lambda_x} \rho c_p \langle v'_g \rangle \langle T' \rangle dx$ is displayed in red for anticyclones, blue for cyclones, and the eddy drift component $\int_{\lambda_x} \rho c_p \langle v_d \rangle \langle T' \rangle dx$ in black dash as both-polarity average. Meridional averages of these curves over λ_y (indicated by short horizontal lines) are indicated by the short vertical lines at the bottom of the panels. As explained in the text, $\langle v_d \rangle$ is assumed uniform within, and zero without, the eddy core (defined to be delimited by the closed zero-contour of composite SSHA). The zonal diameter of this eddy core (dots), as well as the core-average composite SSTA (grey, as both-polarity average, after multiplying cyclonic sections by -1) are also displayed as function of eddy latitude in the Figure. Note the different scales in (a) and (b).

warm anticyclones and cold cyclones observed in energetic regions (Fig. 6.5a), the flow on the eastern eddy side returns a large part of this heat equatorward. In contrast, over quiescent regions' eddies (Fig. 6.5b), the eastern return flow mostly contributes to the poleward transport of heat, a consequence of the larger westward phase shift observed between their SSTA and SSHA. These differences emphasize the presence of larger rotational heat fluxes (e.g. Marshall and Shutts, 1981) in energetic compared to quiescent regions (see also section 6.2.6). Transports at other eddy latitudes are shown by the thin black curves, and are observed to decrease rapidly in amplitude with increasing meridional distance from the eddy centre latitude.

To obtain the net poleward heat transport due to an average eddy, the zonal

integral in (6.12) is computed over the “eddy wavelength” $\langle \lambda_x \rangle$, shown by the short vertical bars in Fig. 6.5 and taken to extend across 5 eddy radii, as this minimizes the mass transport across the eddy-centre latitude ($\int_{\langle \lambda_x \rangle} \langle v'_g(y=0) \rangle dx \approx 0$). The resulting $\rho_0 c_p \int_{\langle \lambda_x \rangle} \langle v'_g \rangle \langle T' \rangle dx$ are displayed by the red and blue curves in Figure 6.6 for the composite anticyclone and cyclone as function of eddy latitude. The curves show that zonally-integrated surface transports are poleward everywhere, and peak at the eddy centre latitude ($y = 0$), across which a single eddy found in energetic regions (Fig. 6.6a) transports on average $\approx 10^{10}$ W/m poleward at the surface, approximately an order of magnitude more than a single eddy in quiescent regions ($\approx 10^9$ W/m - Fig. 6.6b). Eddy transports are significant within about ± 1.5 eddy radii poleward and equatorward of the eddy centre. Beyond $|y| = 2.5$, heat transports are near zero, and these limits (indicated by the short horizontal bars in Fig. 6.6) are taken to define the composite-average meridional eddy extent $\langle \lambda_y \rangle$.

Averaging red and blue curves over $\langle \lambda_y \rangle$ yields the meridional-average surface poleward heat transport by a composite eddy due to the eddy swirl. This is indicated by the short vertical bars at the bottom of the panels, and reaches typically one third of the peak transport across the eddy centre latitude. After furthermore integrating over the climatological mixed-layer depth, we obtain the values of $\langle \delta H_{swirl} \rangle$ (6.12) given in Table 6.2 (row 1) for the composites of the different regions. As shown, a composite eddy in energetic regions transports $\mathcal{O}(10^{12}$ W) poleward in the mixed-layer, a composite eddy in quiescent regions $\mathcal{O}(10^{11}$ W).

6.2.3 Drift heat transport

The drift heat transport of an average eddy is estimated from (6.11) as:

$$\langle \delta H_{swirl} \rangle = \frac{\rho_0 c_p \langle \bar{h}_{ml} \rangle}{\langle \lambda_y \rangle} \int_{\langle \lambda_y \rangle} \int_{\langle \lambda_x \rangle} \langle v_d \rangle \langle T' \rangle dx dy, \quad (6.13)$$

in which $\langle \bar{h}_{ml} \rangle$, $\langle \lambda_x \rangle$, $\langle \lambda_y \rangle$ and $\langle T' \rangle$ are the same as in (6.12).

The average eddy drift velocity $\langle v_d \rangle$ in (6.13) is simply estimated from meridional track displacements. The composites of the meridional track velocity $\langle v \rangle$ have been displayed by the dashed lines in Fig. 5.16c,d for cyclones ($\langle v \rangle^C$, in grey) and anticyclones ($\langle v \rangle^A$, in black) as function of the eddy lifetime. To assure zero mass transport in (6.13), that is to eliminate a net meridional eddy transport or advection of tracks by the mean-flow, the eddy drift velocity is estimated from these as $\langle v_d \rangle = (\langle v \rangle^A - \langle v \rangle^C)/2$ (i.e. as half the difference between black and grey dashed curves in Fig. 5.16c,d), and its lifetime-average is used in (6.13).

Tab. 6.2: Composite single eddy poleward mixed-layer heat transport, and its swirl vs drift components.

			Energetic			Quiet		
			ALL	GS	ACC	ALL	NA	SP
1.	$\langle \delta H_{swirl} \rangle$	in TW	0.51	0.73	0.46	0.07	0.06	0.07
			0.76	1.09	0.69	0.11	0.09	0.10
2.	$\langle \delta H_{drift} \rangle$	in TW	0.05	0.00	0.06			
			0.06	0.00	0.07			
3.	$\langle \delta H \rangle$	in TW	0.56	0.73	0.52	0.07	0.06	0.07
			0.82	1.10	0.76	0.11	0.09	0.10
4.	$\langle \delta H_{swirl} \rangle / \langle \delta H \rangle$		0.91	1.00	0.89			
			0.93	1.00	0.91			
5.	$\langle \delta H_{drift} \rangle / \langle \delta H \rangle$		0.09	0.00	0.11			
			0.07	0.00	0.09			

Composite-averages over eddies with origins in energetic regions (GS and/or ACC) and quiescent regions (NA and/or SP). $\langle \delta H_{swirl} \rangle$ is given by (6.12), $\langle \delta H_{drift} \rangle$ by (6.13), and $\langle \delta H \rangle = \langle \delta H_{swirl} \rangle + \langle \delta H_{drift} \rangle$. See text for details.

Values in thin are not, values in **bold are corrected for a systematic low bias** (at $n^{-1/n} \simeq 0.8$, where n is the time-series length in years) in eddy composites of T' and v'_g , as detailed in Appendix A.3.

This estimate implies an interpretation of tracks as coherent vortices carrying water parcels within their cores, which is supported by the high non-linearity of eddies (CSS11, chapter 5.3.2), by the suggested impact of the down-gradient eddy drift on the initial growth of eddy mixed-layer heat anomalies (Fig. 5.17, chapter 5.3.2), and by previous observational (Sokolov and Rintoul, 2009), as well as idealized and realistic modelling studies (Early et al., 2011; Hallberg and Gnanadesikan, 2006). Nevertheless a coherent nature of tracks is not a priori the case, and tracks may as well capture wave-like SSHA propagation. In this light, the values of H_{drift} provided hereafter should be regarded as an upper bound to the transport of heat by the tracked eddy drift.

For the energetic regions of GS and ACC, Fig. 5.16c shows that warm anticyclones move poleward and towards colder waters (indicated by the composite cross-isotherm track velocity $\langle v_T \rangle$ in solid lines) at a velocity $|v| \approx |v_T| \approx 0.5 \text{ cm/s}$ in the first months of their lifetime, and conversely, that cold cyclones move equatorward and towards warmer water at a similar rate. Here the net eddy (mass) transport velocity ($\langle v \rangle^A + \langle v \rangle^C$) is indeed only a small residual of the recirculating eddy drift $\langle v_d \rangle$, both branches of which add up in terms of heat transport and lead to a net eddy transport of heat poleward.

In quiescent regions (Fig. 5.16d), in contrast, eddies move equatorward and “warm-

ward” irrespective of their temperature/height signal, and do so at different rates³. Although Fig. 5.16b,d also reveal weak meridional deflections of cyclones compared to anticyclones in quiet regions, here the net equatorward translation, possibly in the form of advection by the gyre mean-flow (see also CSS11), dominates meridional track displacements, with the drift being only a small residual. Therefore, we do not further quantify the associated drift heat transport⁴ here.

To evaluate (6.13) for energetic regions, $\langle v_d \rangle$ is assumed uniform within the eddy core, defined as the closed zero-contour in the SSHA composites⁵, and is set to zero outside. For reference, the zonal diameter of this eddy core is shown as function of eddy latitude by the thin dotted curves in Fig. 6.6. As $\langle v_d \rangle$ uses information from both anticyclones and cyclones, in the estimate of (6.13) we also use a both-polarity average eddy perturbation for $\langle T' \rangle$ (obtained as composite average over all snapshots of both anticyclones and cyclones, after multiplying the latter by -1). The grey curves in Fig. 6.6 display the eddy-core average of $\langle T' \rangle$ as function of eddy latitude. Further integrating meridionally and over the mixed-layer yields the mixed-layer heat anomaly within an eddy core $\delta q \equiv \rho_0 c_p \langle \bar{h}_{ml} \rangle \iint_{core} \langle T' \rangle dS$. It amounts to $\approx 0.34 \cdot 10^{19}$ J for a typical GS eddy. The composite ACC eddy heat anomaly ($\approx 0.45 \cdot 10^{19}$ J) is larger (primarily a result of deeper mixed-layers in the Southern Ocean) and at the lower end of in-situ estimates for the *depth-integrated* heat content of eddies previously observed along the ACC (ranging from 0.8 to $5.4 \cdot 10^{19}$ J – see Joyce et al., 1981; Peterson et al., 1982; Morrow et al., 2004; Swart et al., 2008).

Application of (6.13) provides the numbers for δH_{drift} given in Table 6.2, row 2. In the Gulf Stream region, where a systematic drift is only observed in the first month

³ Here $|v_T| > |v|$ reflects westward and warm-ward eddy propagation across meridionally tilted background isotherms.

⁴ In quiet regions, cyclones deflect poleward and anticyclones equatorward with respect to the all-eddy mean (Fig. 5.16b,d), as observed in the global average over long-lived, westward propagating tracks (CSS11), and in a way consistent with theoretical expectations for large decaying vortices on a beta-plane (McWilliams and Flierl, 1979; Nycander, 2001). If systematic and associated with the movement of water, this eddy drift would lead to a weak transport of heat equatorward in quiet regions. For reference, the value that would be obtained, following the same method as in energetic regions, is indicated by the black dashed curve in Fig. 6.6b, and, as shown, this would be more than compensated, in the surface mixed-layer, by the poleward eddy “swirl” heat transport (blue and red curves in Fig. 6.6b).

⁵ This likely overestimates the region, in which eddies carry water parcels and which is better approximated by the zero-contour in relative vorticity (Early et al., 2011). Results are however not very sensitive to this definition, as smaller cores are compensated by larger core-average $\langle T' \rangle$.

of the eddy lifetime (the lifetime-average of $\langle v_d \rangle \approx 0$), δH_{drift} is found to be near zero (only 0.25% of the total $\delta H = \delta H_{swirl} + \delta H_{drift}$, given in Table 6.2, row 3). Even in the ACC region, where $\langle v_d \rangle \approx 0.3 \text{ cm s}^{-1}$, δH_{drift} contributes only a fraction of the composite-average single eddy swirl heat transport ($\leq 10\%$). (For a comment on quiet regions' eddy drift heat transport see footnote⁴.)

The dominance of swirl over drift heat transports is also stunningly visualized in Fig. 6.6a, in which the thick black dashed curve displays the surface drift transport as function of eddy latitude, and the short black dash vertical bar at the bottom of the panel its $\langle \lambda_y \rangle$ average, in comparison to the order-of-magnitude larger swirl transports (blue and red curves and bars).

The implication of this calculation is that the eddies seen by the altimeter in energetic regions carry heat in the mixed layer primarily as a result of the small phase shift between their pressure and temperature centres, not as a result of their meridional drift. Put differently, even though the casual observation of Gulf Stream or ACC meanders pinching-off in satellite SST maps is suggestive of heat transport by eddy shedding, the associated drift velocity is not large enough to contribute significantly to the heat transport. Rather, it is the fact that the warm and cold cores are not actually centred on the pressure cores, but slightly displaced to the west, which predominately leads to eddy heat transport in the mixed-layer.

Although our focus is on the upper mixed layer, it is interesting to speculate on the relative contributions of drift and swirl heat transports to the column averaged heat transport. Indeed, while the swirl heat transport is likely primarily restricted to the mixed layer, consistent with the decrease of the phase shift with depth observed in the North Pacific (Roemmich and Gilson, 2001; Qiu and Chen, 2005), drifting warm and cold eddy cores extend through the thermocline ($\approx 1 \text{ km}$). As a result, it is possible that the two contributions to the depth-integrated eddy heat transport become more comparable over the ACC⁶.

6.2.4 Total mixed-layer eddy heat transport

Following the derivation in section 6.2.1 (eqs 6.7 to 6.11), the total contribution of propagating eddies to the meridional-average mixed-layer heat transport across the

⁶ For the GS this is only expected when considering cross-frontal, rather than poleward, transports, as here, partly due to geographical constraints on meridional eddy motions, but more importantly due its mechanistic link to the frontal instability, only the cross-frontal (but not the poleward) eddy drift velocity is systematically non-zero.

Tab. 6.3: Total eddy poleward mixed-layer heat transport.

		Energetic		Quiet	
		GS	ACC	NA	SP
1.	$\bar{n}/n_y$	17 ± 3	107 ± 8	18 ± 2	26 ± 2
		22 ± 3	140 ± 10	20 ± 2	29 ± 3
2.	H in PW	$.013 \pm .002$	$.056 \pm .004$	$.001 \pm .000$	$.002 \pm .000$
		$.019 \pm .003$	$.082 \pm .006$	$.002 \pm .000$	$.003 \pm .000$
		$.024 \pm .003$	$.107 \pm .007$	$.002 \pm .000$	$.003 \pm .000$
3.	$H/\Delta x$ in TW per 5° longitude	1.5 ± 0.2	0.9 ± 0.1	0.1 ± 0.0	0.1 ± 0.0
		2.2 ± 0.4	1.3 ± 0.1	0.2 ± 0.0	0.2 ± 0.0
		2.8 ± 0.4	1.7 ± 0.1	0.2 ± 0.0	0.2 ± 0.0
4.	κ in m^2/s	2000 ± 300	700 ± 100	500 ± 100	200 ± 20
		2900 ± 500	1100 ± 100	800 ± 100	300 ± 30
		3700 ± 500	1400 ± 100	900 ± 100	300 ± 30

For eddies with origins in energetic regions of GS and ACC, and quiescent regions of NA and SP. In energetic regions the relative contributions of swirl and drift to H , $H/\Delta x$ and κ are the same as those to δH (Table 6.2, rows 4-5); in quiet regions only the swirl heat transport is included. See text for details.

$\bar{n}$ gives the time-average number of coexisting eddies with origins in the respective regions:

- values in *italic print* use $\bar{n}$ obtained when counting eddies only as long as they remain in these regions,
- values in upright print use $\bar{n}$ obtained when including eddies that have left the regions to compensate for the entry into the regions of eddies with origins elsewhere.

As in Table 6.2, values in thin are not, values in **bold** are corrected for a systematic low bias in eddy composites of T' and v'_g .

different regions is obtained by multiplying δH (Table 6.2, row 3) by the meridionally scaled eddy number $\frac{\bar{n}}{n_y}$, given in Table 6.3, row 1.

The resulting values for H are given in Table 6.3, row 2. Propagating eddies tracked in SSHA induce a very modest net poleward mixed-layer heat transport across the quiescent regions of each North Atlantic and South Pacific (≈ 0.002 PW), which, as expected, contrasts with larger net poleward mixed-layer eddy heat transports observed across the GS (0.02 PW), and the ACC (0.1 PW). The relative contributions of swirl and drift components to H are the same as those to δH that have been indicated in Table 6.2, rows 4-5. Thus, in energetic regions $> 90\%$ of H are provided by the eddy swirl (- in quiescent regions, as discussed above, only the swirl heat transport is included).

Assuming that the principal source of uncertainty is the number of coexisting eddies, the latter's variability between different weeks provides an estimate of the uncertainty of the eddy heat transports, which is found to be on the order of 10% (and is indicated by the $\pm$ error margins in Table 6.3). To get a better idea of the sensitivity of H to the estimate of eddy numbers, we can moreover estimate upper and lower bounds for the number $\bar{n}$ of coexisting eddies that have originated in a regional

bin by in turn including and excluding eddies that have left the bin in the time-average. The latter (given by the values in italic print in Table 6.3, row 1) would give a realistic estimate of eddy numbers, and thus of associated heat transports, if eddies originating in the respective regional bin were the only eddies present in that region. However, if eddies with origins elsewhere pass through the region as well, it is an underestimate and the former, upper bound estimate (which includes eddies that leave the bin in the count, in compensation for eddies that enter the bin from elsewhere) may be more appropriate, and is thus also given in Table 6.3 (by the values in upright print). For the regions and time interval chosen, the former estimate typically lies within 75% and 90% of the latter.

Uncertainty moreover arises due to the short observational time series at the basis of the present estimate (with a number of years $n \approx 5^{1/2}$). As shown in Appendix A.3, this induces a systematic low bias in eddy composites of both T' and v'_g . Correcting for this leads to the numbers given in bold below the original numbers⁷ in Tables 6.2 and 6.3. They are enhanced by a factor $(n/n-1)^2 \simeq 1.5$ for the swirl heat transport, whereas the drift heat transport is only enhanced by $(n-1/n)^1 \simeq 1.2$, as the bias on eddy tracks, and thus v_d , is unknown.

It is worth emphasizing that the ACC dominates the numbers in Table 6.3, row 2, only due to its large zonal extent. Per longitude interval, ACC eddies transport on average only 60% of the heat GS eddies transport poleward in the mixed-layer (1.3 versus 2.2 TW per 5° longitude – Table 6.3, row 3), even though mixed-layers are typically twice as deep over ACC eddies. Section 6.2.6 sheds further light on this interesting discrepancy.

The meridional-average heat transport calculation, presented here, is based on observations of eddies originating from a broad range of latitudes⁸ and only applies to the surface mixed-layer. Apart from being sensitive to the exact eddy numbers, resulting values in Table 6.3 are also sensitive to the exact spatial dimensions of the chosen regional bins⁹, which underlines the approximate scaling nature of the present

⁷ For brevity, Table 6.3 shows only corrected values for the upper bound estimate - in bold upright print.

⁸ Note that the NA energetic region, referred to as GS in the Tables, also includes $\approx 8\%$ of tracks with origins in the Mexican Loop Current (see Fig. 3.12a).

⁹ For example, adding $\langle \lambda \rangle / 2$ to either end of the estimated (maximum) zonal and (average) meridional dimensions of the bins L_x and L_y , one obtains values of H that are only ≈ 0.5 - 0.75 times as large as those given in Table 6.3. Equivalent diffusivities (see section 6.2.5), via their dependence on the eddy density $\frac{\bar{n}}{n_y n_x}$ (see section 6.2.6), are also smaller by about the same factor.

estimate and urgently calls for a careful latitude-by-latitude or isotherm-by-isotherm assessment of the observed mixed-layer eddy heat transport in future work. As a result, a direct comparison of the present estimate to previous estimates of zonally and depth-integrated eddy heat transport is only tentative. Nevertheless, in the light of the observed strong surface intensification of eddy heat transports in both ACC and mid-latitudes (Wunsch, 1999; Phillips and Rintoul, 2000; Roemmich and Gilson, 2001; Qiu and Chen, 2005), it is interesting to note that the heat transported in the surface mixed-layer (≈ 100 m) by propagating eddies originating in the energetic ACC region, which is almost entirely provided by the eddy swirl component ($H_{swirl} + H_{drift} \approx H_{swirl} \approx 0.1$ PW), lies just at the lower end of the range of previous estimates for the total, depth-integrated, cross-ACC eddy heat transport (0.1 to 0.9 PW, as summarized by Gille 2003b). Given the sparseness of observations in the SO, the eddy heat transport required to close the oceanic heat-budget poleward of the ACC is uncertain, but according to the early 0.45 ± 0.3 PW estimate by deSzoek and Levine (1981, which is not inconsistent with the earth radiation budget and atmospheric reanalysis – Ganachaud and Wunsch, 2003; Trenberth and Caron, 2001), mixed-layer eddy heat fluxes are unlikely to do the job on their own.

Irrespective of this, a poleward mixed-layer eddy heat flux across major current systems is clearly established from present observations, and will thus (cf. chapter 1) play an important role in their surface buoyancy budget, water mass structure and overturning circulation, especially in the SO.

6.2.5 Implied diffusivities

For reference and comparison with other studies, a bulk diffusivity κ for the poleward mixed-layer heat transport by eddies in a given region can be estimated from the relation

$$\frac{(H_{swirl} + H_{drift})}{\rho c_p \langle \bar{h}_{ml} \rangle L_x} = \frac{\bar{n}}{n_y n_x \langle \lambda_y \rangle \langle \lambda_x \rangle} \int_{\langle \lambda_y \rangle} \int_{\langle \lambda_x \rangle} (\langle v'_g \rangle + \langle v_d \rangle) \langle T' \rangle dx dy = -\kappa \left\langle \frac{\partial \bar{T}}{\partial y} \right\rangle, \quad (6.14)$$

in which we have defined $n_x \equiv L_x / \langle \lambda_x \rangle$, and used (6.7), (6.12) and (6.13). In this equation, $\langle \partial \bar{T} / \partial y \rangle$ is the meridional gradient of the monthly SST climatology, evaluated and averaged along eddy tracks. (Note that for the quiescent regions, again, the drift heat transport is not included.)

Given in Table 6.3, row 4, the resulting meridional diffusivities are found to be positive everywhere, indicating that eddies act to remove the available potential energy

stored in the eastward mean surface shears (equatorward mixed-layer temperature, and buoyancy, gradients) observed both in energetic and in quiet regions.

In the South Pacific, where $\langle \partial \bar{T} / \partial y \rangle = -0.5$ K/100 km (associated with the eastward-flowing poleward branch of the subtropical gyre), one obtains $\kappa \approx 300$ m² s⁻¹. Larger diffusivities are found in the NA subtropics ($\kappa \approx 800$ m² s⁻¹), where the north-south tilting of isotherms leads to a meridional temperature gradient of only -0.2 K/100 km (associated with the decrease towards the surface in the upper thermocline of the mean westward flow observed there).

The SST gradient averaged over ACC and GS eddies is of the same order as in the quiescent regions (-0.8 K/100 km), but because of much larger values of H_{swirl} in energetic regions, diffusivities are larger (≈ 1100 m² s⁻¹ for the ACC and 2900 m² s⁻¹ for the GS – or, before bias correction, ≈ 700 and 2000 m² s⁻¹, respectively).

Note that numbers are again sensitive to the dimensions of the regional bin (see also footnote⁹) and to the exact estimate of eddy numbers – the values cited above are representative for the lower bound estimate, those obtained with the upper bound estimate are ≈ 1.1 to 1.3 times larger and given in the third sub-row of Table 6.3, row 4.

Diffusivities obtained here are broadly consistent with the spatial distribution of κ estimated from altimetry (Stammer, 1998). In particular, they closely agree with the larger values of κ observed over the GS core by Zhai and Greatbatch (2006a). The weaker values observed here in the energetic ACC region likely reflect a combination of weak diffusivities in the ACC core and larger diffusivities in the western boundary current extensions (Agulhas return current and Brazil-Malvinas confluence) on its equatorward flank (as estimated e.g. by Marshall et al., 2006).

Apart from a ≈ 100 m² s⁻¹ contribution by the drift mechanism in the ACC, mixed-layer eddy diffusivities are entirely provided by the eddy swirl.

6.2.6 Understanding the spatial variability of eddy diffusivities

To gain insight into the mechanisms that set the observed regional differences in the predominant swirl contribution the meridional mixed-layer eddy diffusivity for heat κ (given by the first term in (6.14)), we introduce a typical scale T_0 for the observed eddy SST signature (set to the average of positive and negative $\langle T' \rangle$ extrema observed along the zonal section through the eddy centre in Figs. 6.1a-b & 6.3a-b: $T_0 = (\langle T' \rangle_{max} - \langle T' \rangle_{min})/2$), a typical scale v_0 for the eddy “swirl” velocity (similarly set to the maximum $\langle v'_g \rangle$), and a typical (temperature mixing) length scale

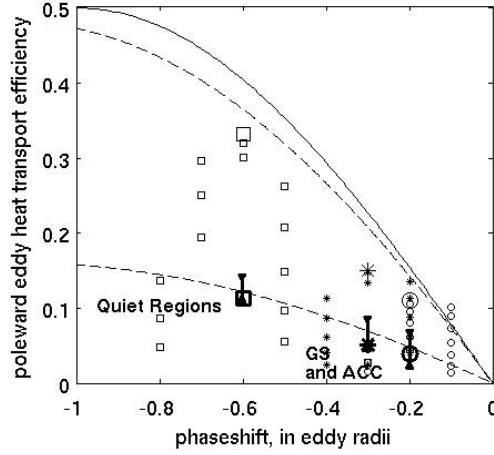


Fig. 6.7: The efficiency of the eddy poleward mixed-layer heat transport c_e (6.16), observed over eddies in quiescent regions $\square$, Gulf Stream ($*$) and ACC ($\circ$), is a simple function of the observed phase shift δ_x between eddy SSTA and SSHA. This function is shown for a plane wave (solid line) and for a localized wave-packet (two dashed lines, one for $c_e(y = 0)$, the other for the meridional average $c_e = c_e(y = 0)/3$). The small thin symbols show observed $[\delta_x(y \neq 0), c_e(y \neq 0)]$ pairs, the larger thin symbols show $[\delta_x(y = 0), c_e(y = 0)]$ pairs, and the large bold symbols indicate the observed meridional-average efficiency c_e as function of $\delta_x(y = 0)$. The difference in c_e , observed between quiescent and energetic oceanic regimes, clearly stands out from the error on c_e (indicated by the thick vertical bars), which is associated with the observed accuracy in δ_x of ± 0.1 eddy radii and estimated from the slope of the lower dashed curve.

$L_0 \equiv -T_0 / \langle \partial \bar{T} / \partial y \rangle$. With these new variables, the swirl diffusivity can be expressed as

$$\kappa = \frac{\bar{n}}{n_y n_x} c_e v_0 L_0. \quad (6.15)$$

Here $\frac{\bar{n}}{n_y n_x}$ provides a measure for the time-average eddy density, and the non-dimensional parameter

$$c_e \equiv \frac{1}{\langle \lambda_y \rangle} \int_{\langle \lambda_y \rangle} c_e(y) dy = \frac{1}{\langle \lambda_y \rangle \langle \lambda_x \rangle} \int_{\langle \lambda_y \rangle} \int_{\langle \lambda_x \rangle} \frac{\langle v'_g \rangle \langle T' \rangle}{v_0 T_0} dx dy \quad (6.16)$$

measures the efficiency of the eddy heat transport. c_e is zero when eddy perturbations in temperature and velocity are in quadrature at a given eddy latitude, and reaches its maximum value when they are in phase (or, in terms of pressure, $c_e = 0$ when

temperature and pressure fluctuations are in phase, and c_e is maximum when they are in quadrature). This shows that c_e is a simple function of the zonal phase shift between temperature and pressure (δ_x).

The shape of this function is displayed by the continuous curve in Fig. 6.7 for a simple sinusoidal plane wave signal¹⁰, for which c_e reaches a maximum of $\frac{1}{2\pi} \int_{-\pi}^{\pi} \sin^2 x dx = 1/2$ (for a phase shift $\delta_x = -1r_0$). Observed values of c_e and SSTA/SSHA phase shifts, obtained from eddy composites in the different regions, are superimposed on this curve by the symbols in Fig. 6.7, and, at the eddy centre latitude (large thin symbols show $\delta_x(y = 0)$, $c_e(y = 0)$), they closely follow the functional relationship obtained for the linear wave, or for a localized wave-packet (upper dashed line)¹¹.

Since eddy $\langle T' \rangle$ and $\langle v'_g \rangle$ amplitudes, $T_0(y)$ and $v_0(y)$, decrease away from the eddy centre latitude (cf. Figs. 6.1 & 6.3), the meridionally averaged eddy heat transport efficiency c_e (shown by the large bold symbols) is reduced compared to $c_e(y = 0)$ (specifically by the factor $\frac{1}{\langle \lambda_y \rangle} \int_{\langle \lambda_y \rangle} v_0(y) T_0(y) dy / v_0 T_0$, observed to be about $1/3$), but still follows the same simple functional relationship to $\delta_x(y = 0)$. As can be seen, c_e is about three times larger in quiescent than in energetic regions, and this regional variation thus reflects differences in the observed eddy SST signatures (via their phase shift). Note that, if this were the sole effect (i.e. if it were not compensated by an increase in v_0 or L_0), it would lead to larger diffusivities in the quiescent regions away from major current systems.

The variability of v_0 and L_0 amongst the regions is as follows. In quiet regions, $L_0 \approx 30km$, which is close to the local deformation scale R_d or about a third of the observed eddy radius (≈ 80 km). In energetic regions of GS and ACC, surface L_0 is enhanced to ≈ 60 km, which is more than twice as large as R_d , or more than $2/3$ of the eddy radius. The velocity scale v_0 is found to be more than three times larger in energetic regions compared to quiet regions.

The eddy density $\frac{\bar{n}}{n_y n_x}$ shows little variation between the regions, and typically amounts to slightly less (more) than 2, when choosing the lower (upper) bound estimates for $\bar{n}$ (cf. section 6.2.4). This indicates a dense packing of eddies in the time-

¹⁰ defined by height and temperature anomalies $\eta' = \eta_0 \cos k(x - ct)$ and $T' = T_0 \cos k(x - ct - \delta_x)$, where $k = 1/r_0 = 2\pi/L_x$ and, by geostrophy, $v'_g = -v_0 \sin k(x - ct)$, with $v_0 = g\eta_0 / f r_0$

¹¹ The compared to the linear wave slightly smaller observed $c_e(y = 0)$ reflect the eddy nonlinearity. The observed zonal shape of $\langle \eta' \rangle$ and $\langle T' \rangle$ more closely resembles a localized wave-packet, such as represented by a sinusoidal wave damped with a Gaussian centred at $x = 0$. In this case, differentiation yields a narrower structure of $\langle v'_g \rangle$ compared to $\langle \eta' \rangle$ and $\langle T' \rangle$, so that c_e of the damped wave (indicated by the dashed line) never reaches that of the linear wave.

average, but the exact value is highly sensitive to the exact estimate of eddy numbers and dimensions of the chosen regional bins (see also footnote⁹).

Combining these results reveals that the larger diffusivities observed over the GS compared to the ACC (see section 6.2.5 & Table 6.3) result mainly from more vigorous eddies (larger v_0) observed over the GS. The reduction in κ from energetic to quiescent regimes is driven by weaker eddies (v_0) with shorter mixing lengths (L_0), but, interestingly, this reduction in κ is importantly limited by the regional variations in the heat transport efficiency c_e : values of c_e observed here (≈ 0.05 in energetic, 0.11 in quiescent regions, cf. Fig. 6.7) are within the range proposed previously (0.01 to 0.4), but if c_e were a constant (as is usually assumed in the literature, e.g. Holloway and Kristmannsson, 1984; Visbeck et al., 1997; Stammer, 1998; Wunsch, 1999; Karsten and Marshall, 2002), the spread in κ between quiescent and energetic regimes would be three times larger.

6.3 Discussion and mechanisms

Although a detailed analysis of the mechanisms that set the eddy T' signature, its phase shift and thus eddy mixed-layer heat transports (via the efficiency c_e) is beyond the scope of this study, this section provides some discussion of the important mechanisms to consider.

Both interacting baroclinic vortices (Hogg and Stommel, 1985; Spall and Chapman, 1998) and growing/unstable baroclinic waves (e.g. Eady, 1949) can exhibit a shift in phase between their surface temperature and pressure perturbations and transport heat. Whether the phase shift observed here between eddy SSTA and SSHA is indeed the surface expression of a tilt in eddy thermal cores that slowly decays throughout the thermocline, as seen e.g. in observations in the subtropical North Pacific (Roemmich and Gilson, 2001), remains to be tested from systematic studies of subsurface data. Instead the observed phase shift might also be a phenomenon specific to the surface mixed layer, in which diabatic interactions of eddies with the subsurface (through entrainment) and with the atmosphere (through turbulent air-sea heat flux damping) could play an important role in its generation. As reviewed in chapter 5.1.1, such a key role of mesoscale air-sea interactions in setting eddy heat transports would be consistent with ideas from studies like Plumb and Mahlman (1987), Drijfhout (1994b), Zhai and Greatbatch (2006b), Greatbatch et al. (2007) and Shuckburgh et al. (2011).

Whereas especially the latter studies typically think of the eddy “drift”, when

rationalizing how air-sea damping of eddy T' imparts irreversibility to mixed-layer eddy heat transports, the same applies to the eddy “swirl”, found here to dominate heat transport. To see this, consider a simple linear plane wave travelling westward in an equatorward SST gradient. Without any air-sea damping (or vertical exchanges), the associated T' is controlled by $\partial_t T' = -v'_g \partial_y \bar{T}$, and will thus be exactly in phase with the travelling pressure signal (in quadrature with v'_g), resulting in purely rotational heat transports ($c_e = 0$). The presence of damping acts to modify phase and amplitude of T' , via $\partial_t T' = -v'_g \partial_y \bar{T} - \gamma T'$, where γ is an inverse damping time-scale. If damping and stirring time scales match, T' is expected to be shifted westward with respect to the eddy pressure core by $\approx -45^\circ$ (which corresponds to $\delta_x \approx -0.5$ eddy radii in the framework used above and in Fig. 6.7), and this drives a net poleward eddy mixed-layer heat transport ($c_e > 0$, and specifically $c_e \approx c_{emax}/2$, where c_{emax} is the maximum value that can be approached by c_e for a phase shift of -90°). As damping weakens below this value, phase shift, c_e and net heat transport decrease to zero. As damping increases, phase shift and c_e increase, but the eddy T' amplitude weakens, eventually also leading to a reduction in the net heat transport. κ associated with the (poleward) swirl heat transport of this simple damped wave thus follows a qualitatively similar functional dependence on the air-sea damping strength as that derived by Shuckburgh et al. (2011) (cf. Fig. 5.2).

It is interesting to ask whether this mechanism is also relevant for the heat transport by the nonlinear eddies, studied here. The T' signatures of the weakly non-linear eddies observed in quiescent regions (Fig. 6.3) indeed closely resemble eddy tracer anomalies obtained in nonlinear QG simulations (Fig. 6.8 lhs), in which only the above two mechanisms, stirring by the eddy “swirl” and damping, act: both show a bipolar anomaly signature whose predominant western pole¹² is shifted westward with respect to eddy SSHA. Isotherm-stirring by the eddy “swirl” thus plays an important role in the generation of quiet regions’ eddy T' signatures, but further modelling studies like those of Early et al. (2011) are required to assess whether air-sea damping alone, whose strength has been constrained in chapter 5.1, is indeed sufficient to provide the restoring required to generate the observed SSTA-SSHA phase-relationship¹³.

¹² Opposing anomaly signs are accounted for by the opposite background SST (equatorward) and model-tracer gradients (poleward).

¹³ The model composites of Chelton et al. (2011a) are obtained in a very strong damping regime (for typical 50 m mixed layers, a 30-day restoring corresponds to a damping strength of $100 \text{ W/m}^2 \text{ K}$) and display larger phase shifts than observed eddy T' signatures. This is expected if model tracer and observed SST gradients, and therefore stirring time scales, are comparable, but their relationship

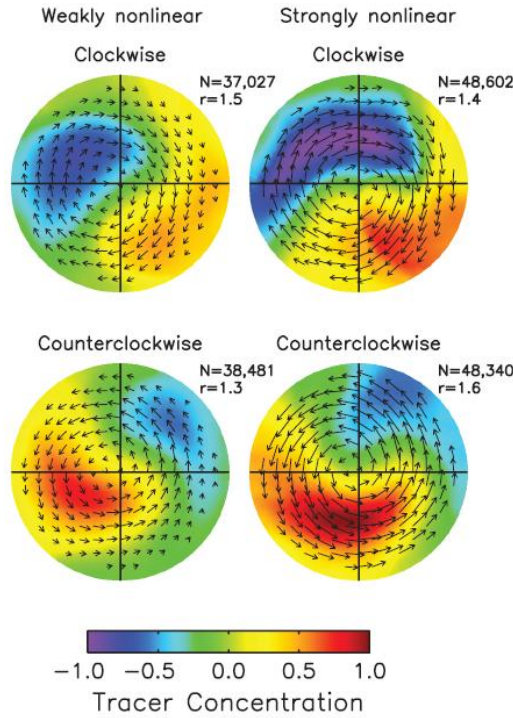


Fig. 6.8: Tracer anomaly composite-averaged over (N) random Gaussian eddies seeded in a nonlinear quasi-geostrophic simulation (Early et al., 2011) within a constant northward tracer gradient that is restored at a time-scale of 30 days. Initial eddy length scale and amplitude match observations in the weakly nonlinear case, in the strongly nonlinear simulation amplitudes are multiplied by a factor of 3. (Composites, normalized by the eddy radius, are displayed within 2 eddy radii, north is upward, vectors show the gradient of composite-average SSHA, and r indicates the magnitude ratio of the western to eastern tracer anomaly extrema.) Reproduced from Chelton et al. (2011a), their Fig. 3B,C.

The T' signature observed following energetic eddies (Fig. 6.1) is to the contrary very dissimilar to that obtained in the strongly nonlinear model simulations (Fig. 6.8 rhs). Here mechanism other than stirring by the eddy “swirl” clearly need to be evoked to account for their main characterizing feature, which is a near in-phase SSTA-SSHA relationship in the form of warm-top anticyclones and cold-top cyclones¹⁴. Here we is unclear.

¹⁴ In energetic regions, air-sea damping is nearly two orders of magnitude weaker than the stirring of the observed narrow fronts by the intense eddy “swirl” (chapter 5.2). Given the high eddy nonlinearity, in such a weak damping regime, we would expect temperature to become homogenized within eddy cores after some time. In this case, eddy T' would simply be a mirror image of the background

propose that this is generated through isotherm-stirring by the systematic cold-ward drift of anticyclones/warm-ward drift of cyclones (Fig. 5.16c), through which, in a manner expected from diffusive arguments, coherent eddies conserving their temperature will appear as anomalies in their new environment. This is consistent with the impact of energetic eddies' down-gradient drift on the initial growth of their T' , suggested in chapter 5.3. However, the generation, in this scenario, of the observed systematic small phase shift between eddy SSTA and SSHA, remains an open question.

Figures 6.9 & 6.10 provide further support for the important role of mixed-layer isotherm stirring by the eddy swirl in quiet, and by the eddy drift (diffusion) in energetic regions. They display the seasonal modulation occurring from winter (top panels) to summer (bottom panels) in background SST gradients (middle panels) and composite eddy T' signatures (remaining panels) in NA & SO, respectively. In the NA quiet regions, interestingly, background isotherms not only widen, but also tilt meridionally in summer (Fig. 6.9h) compared to winter (Fig. 6.9c). Consistently, eddy T' signatures not only weaken, but also rotate from a zonal to a NW-SE orientation from winter (Fig. 6.9d,e) to summer (Fig. 6.9i,j), which provides strong support for their generation through stirring of the background SST field by the eddy swirl.¹⁵ In energetic regions, the summer-time weakening of the SST fronts across which eddies drift (Fig. 6.9 & Fig. 6.10 h) is consistent with the weaker observed eddy T' signatures (Fig. 6.9 & Fig. 6.10 f,g), and these in turn are consistent with the weakened summer-time contribution of the oceanic mesoscale to T' variability observed in major current systems (chapter 4.4.3).

6.4 Conclusions

This chapter has provided a robust observational estimate of the signature of transient mesoscale motions in SST, and evaluated the associated mixed-layer heat transport, for energetic and quiet regimes of NA and SO by analysing the relationship between MW-SSTA and multi-altimeter SSHA following propagating eddy tracks.

$\bar{T}$ gradient across the eddy, i.e. we would expect to see positive T' on the poleward, and negative T' on the equatorward side of eddies. Although such meridional shifts are observed (Fig. 6.1d), they are not the main feature of energetic eddy T' .

¹⁵ In the SP quiet regions, the seasonality of $\bar{T}$ gradients, but not that of eddy T' , is reversed. Here, in winter, when background $\bar{T}$ gradients are weak (Fig. 6.10c), eddy T' only become less bipolar/more asymmetric, but not less intense (Fig. 6.10d,e), which may point to a role of entrainment in their generation.

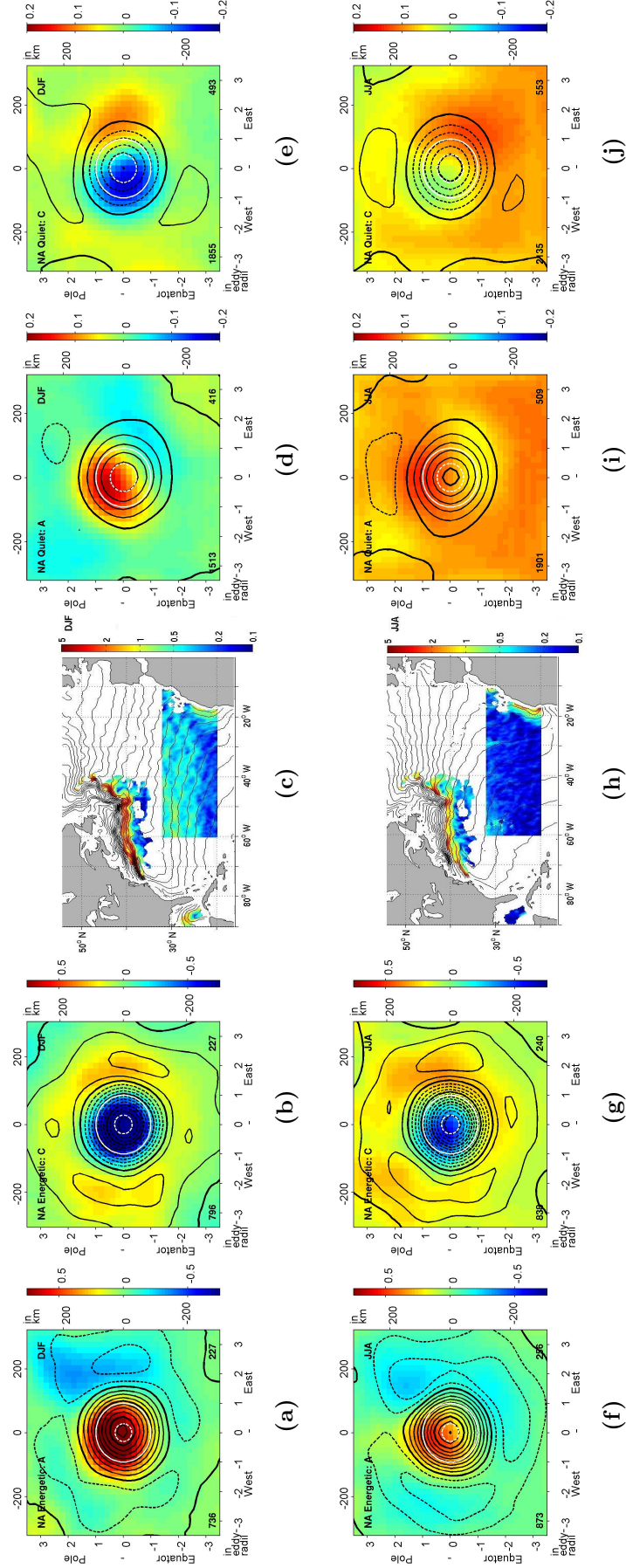


Fig. 6.9: Composite SSTA and SSHA of NA energetic (a,b,f,g – as Fig. 6.1) and quiet eddies (d,e,i,j – as Fig. 6.3) evaluated separately for winter (top) and summer season (bottom). Middle panels display the corresponding fields of $\bar{T}$ (contoured every $^{\circ}\text{C}$) and $|\nabla\bar{T}|$ (in $\text{K}/100\text{ km}$, coloured only in energetic and quiet regions).

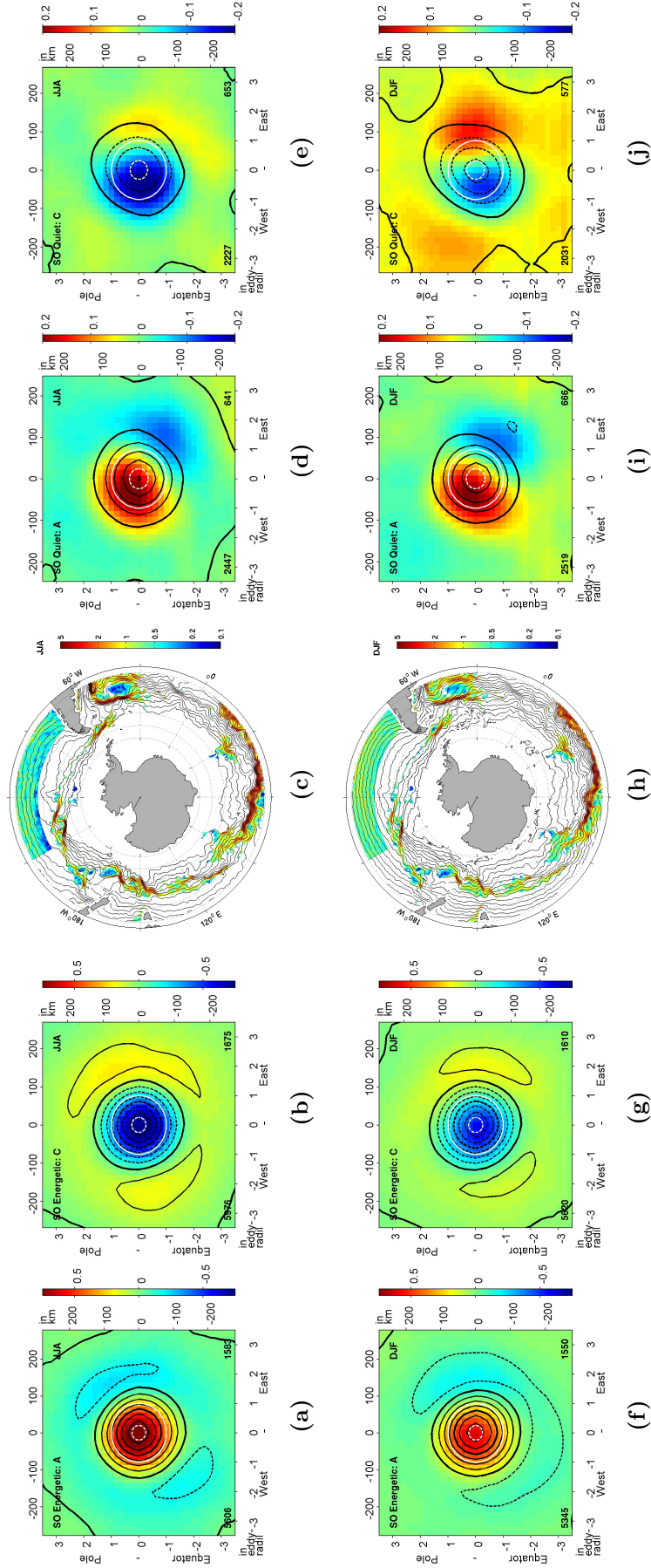


Fig. 6.10: As Fig. 6.9, but for the SO.

Key results are summarized as follows:

- Propagating eddies are found to have a clear signature in SST that is characterized by warm tops over anticyclones and cold tops over cyclones, but also by a systematic westward phase shift of SSTA with respect to rotating eddy cores.
- The eddy SST signature displays marked regional differences between basin interiors and major current systems: Over the relatively weak eddies originating in quiescent interiors the phase shift is pronounced and SST signatures are bipolar, showing an only weak asymmetry towards stronger warm poles over anticyclones/cold poles over cyclones. The intense eddies in GS and ACC, to the contrary, have strongly asymmetric SST signatures, whose dominant warm/cold poles are nearly in phase with high/low eddy pressure cores, but also display a systematic, though small, westward phase shift.
- The spatial phase shift between temperature and pressure signals of an eddy is such that it implies a systematic poleward eddy heat transport in the mixed layer, in both energetic and quiescent oceanic regimes.
- This mechanism of heat transport (“swirl”) is found to dominate over the “drift” heat transport associated with the poleward (equatorward) motion of warm anticyclones (cyclones) observed in ACC and GS regions. Mixed-layer eddy diffusivities, modest in quiescent interiors ($\kappa \approx 500 \text{ m}^2 \text{ s}^{-1}$), intermediate in the ACC ($\approx 1000 \text{ m}^2 \text{ s}^{-1}$) and large in the GS ($\approx 3000 \text{ m}^2 \text{ s}^{-1}$), thus primarily reflect the eddy swirl heat transport (Table 6.3).
- Scaling results suggest that the implied poleward mixed-layer eddy heat transport across the ACC region is significant ($\approx 0.1 \text{ PW}$). To confirm this a careful latitude-by-latitude, or front-by-front eddy composite analysis is required, where the challenge lies in systematically isolating the eddy signal from large-scale SST variability and retain its statistical significance while eddy numbers decrease.
- The regional variation of the phase shift characterizing eddy SST signatures drives a larger contribution of rotational fluxes to eddy heat transports in major current systems, and thereby limits the increase of eddy heat diffusivities from quiescent to energetic regions that is expected from the intensification of eddies and mixing lengths alone.

It is striking to observe westward phase shifts, reminiscent of linear unstable baroclinic disturbances, over tracked eddies whose dynamics have been shown to be nonlinear (CSS11) and whose radii are larger than the observed most baroclinically unstable scale (Smith, 2007; Tulloch et al., 2011). Since eddies are most likely generated by baroclinic

instability, they must evolve through an inverse energy cascade while conserving their SSTA-SSHA phase relationship. As such they appear to transport heat poleward in two stages. First, by extracting available potential energy from the environment during the initial stage of their growth (on the order of 10 days for energetic regions according to Tulloch et al. 2011), a behaviour captured by classical models of baroclinic instability (e.g. Eady, 1949). Second, by subsequently cascading to larger scale and continuing to transport heat poleward over much longer time scales (possibly during the entire eddy lifetime of several months/years), a behaviour that is reminiscent of the archetypal “heton” heat transport model introduced by Hogg (1985).

Whereas in quiet regions phase shifts have been detected in previous observational studies (e.g. Voorhis et al., 1976; Roemmich and Gilson, 2001), their systematic presence also over eddies in major current systems is a somewhat surprising result. Once full SSH observations attain sufficient accuracy and resolution, it would be fascinating to assess whether phase shifts are indeed observed over fully developed coherent eddies (“rings”), or whether the present result reflects a mixture of in-phase SST signatures observed over shed eddies and shifted signatures observed over growing current meanders.

The observed westward phase shift between temperature and pressure has been shown here to be the key mechanism by which mesoscale motions transport heat poleward in the mixed layer. Although different mechanisms clearly need to be evoked in the generation of quiescent and energetic regions’ eddy SST signatures (horizontal stirring by the eddy “swirl” plays a key role over the former, stirring by the eddy “drift” over the latter), it is unclear at this stage whether there is a single dominant mechanism to set their phase shift. As discussed above heat exchanges with the atmosphere on mesoscales, whose important role in the general circulation is starting to be widely appreciated (Greatbatch et al., 2007; Cerovecky and Marshall, 2008; Shuckburgh et al., 2011), may well play a crucial role in setting their magnitude, and further observational and modelling studies will hopefully shed light on this fascinating issue.

Chapter 7

CONCLUSIONS AND OUTLOOK

This thesis has provided an observational estimate of the signature of the ocean's transient mesoscale eddy field in sea surface temperature (SST).

As sketched in chapter 1, the primary aim was thereby to complement the previously relatively scattered observational knowledge of this variable that not only determines the scope for transient (mesoscale) ocean-atmosphere coupling in midlatitudes, but is also the prerequisite for any systematic eddy fluxes of heat through the surface mixed-layer. The latter are not only expected to supply an important fraction of total eddy heat transport (e.g. Wunsch, 1999), but theory (Radko and Marshall, 2003) and idealized eddy-resolving modelling studies (Hallberg and Gnanadesikan, 2006) suggest that they potentially play a key role in the ocean's general circulation and its distribution of water masses.

Observationally estimating the eddy SST signatures has only recently become possible globally, with the advances in the accuracy and resolution of sea surface height (SSH) observations opened by multi-mission altimetry, and the development of through-cloud microwave (MW) radiometry that for the first time allows the systematic study of SST variability at high latitudes in all weather conditions. Here we have taken advantage of these new observing techniques and estimated the eddy signal by analysing the relationship between transient fluctuations in through-cloud MW-SST (chapter 3.1) and multi-altimeter SSH (chapter 3.2). In comparison, the few earlier attempts to link satellite-based SST and SSH variability have, as reviewed in chapter 4 and with the exception of the recent study by Qiu and Chen (2005) in the North Pacific, been necessarily limited by the lower resolution observations available at the time.

The combination of automated eddy-identification from satellite altimetry and in-situ data has provided invaluable insight into eddy temperature distributions in the North Pacific (Roemmich and Gilson, 2001). Nonetheless, studies of SSTA systemati-

cally following eddy tracks remain rare. Note that the many previous studies tracking mesoscale anomalies in IR-SST alone, especially in the vicinity of the Gulf Stream (GS, cf. Park et al., 2006, for a recent example), have not explicitly addressed their relationship to eddying circulations. In this thesis, we have systematically evaluated the SST signature of propagating mesoscale eddies by following an indeed very large number of eddy tracks that are identified in the SSH field by the automated eddy tracking algorithm recently developed by CSS11 (chapter 3.2.4).

Switching between the complementary fixed-location and track-following perspectives on SST and SSH variability has made it possible to address each of the three questions posed in chapter 1: By averaging SST fluctuations along eddy tracks, we were able to characterize a statistically robust signature of transient mesoscale motions in the SST field, quantify the ensuing mesoscale contribution to SST variability, and estimate the associated eddy heat transport through the surface mixed-layer. The mechanisms underlying the observed mesoscale contribution to SST variability and poleward heat transport have also been touched upon, and importantly, climatological constraints on the strength of mesoscale air-sea heat exchanges have been provided.

The analysis has focussed on two contrasting oceanic regimes (chapter 3.3) observed in the North Atlantic (NA) and the Southern Ocean (SO): the “energetic” regions of peak eddy variability (typically ≥ 15 cm/s in the NA, and ≥ 10 cm in the SO) observed over the cores of the major NA and SO currents systems, and the “quiescent” regions of weak mesoscale eddy variability (typically ≤ 5 cm/s) observed in the eastern parts of subtropical gyres in NA and South Pacific (SP).

In the following sections, we briefly summarize our main observational results on each of the above topics, while emphasizing the important open questions and avenues for future research. (Detailed observational results can be found at the end of the respective chapters.) We commence, in section 7.1, by recalling the sensitivity of our results to the specific datasets and methods used, and propose some guidelines for future tests and reanalyses.

7.1 Sensitivity of results to dataset and method

To facilitate the study of transient processes, in this thesis, the variability in SST and SSH has been separated into the seasonal variability of climatological background state and non-seasonal departures from it, denoted SSTA and SSHA. However, as noted at several places throughout this study and systematically analysed in Appendix A, es-

timating this separation from the short observational time-series length available at the beginning of the study (5½-years) induces both noise terms and systematic biases in estimated temporal anomalies. It has been possible to evaluate the systematic bias and this has been accounted for in the discussion of observational results. Appendix A suggests that the noise terms may also lead to a systematic modification of the estimated space-time characteristics of SST and SSH variability, and these need to be carefully assessed, either in a future reanalyses with longer time series or by designing better methods to estimate the separation of seasonal and non-seasonal variability from short time series.

The spatial and temporal resolution of SST and SSH datasets has been discussed in detail in chapter 3 and is just about sufficient to resolve mesoscale processes. Whereas the trade-off between resolution and data coverage between available IR and MW observations of SST, makes the chosen SST dataset one of the currently best apt to study mesoscale processes in the SO, higher-resolution and more homogeneous-quality datasets of SSH are available, and the sensitivity of present results to the short-comings of the chosen SSH dataset (chapter 3.2) can thus be assessed through further analysis.

In summary, in the interpretation of below results it is important to remember that they pertain to the specific observational datasets of SSTA and SSHA, and to the specific eddy dataset, analysed in this thesis (chapter 3) and that their sensitivity to the short-comings of these datasets needs to be carefully assessed in future study.

7.2 Eddy SST signatures and their interaction with the atmosphere

- On average, propagating mesoscale eddies are observed to be associated with systematic SST signatures, that strikingly persist throughout the eddy life time, of months to years (Fig. 5.12). This is a key and observationally robust result of this thesis, and applies not only to the high-amplitude eddies with origins along GS and ACC, but also to those originating in quiescent subtropical interiors.
- Eddy SSTA have much larger amplitude ($\sim 0.6\text{K}$) over energetic compared to quiescent regions' eddies ($\sim 0.2\text{K}$). In quiescent regions, substantial averaging is therefore required to reveal the systematic eddy SST signal, which is masked by more intense large-scale variability in individual eddy snapshots. Although eddy signals are much easier to detect in energetic regions, even here they are often masked by large-scale variability in a given snapshot.

- The eddy SST signature is characterized by warm anticyclones and cold cyclones, but also by a westward phase shift of SSTA with respect to rotating eddy cores. This phase shift is more pronounced for eddies originating in quiescent regions of the subtropical gyres (Fig. 6.3), but is also clearly established for the GS and ACC regions (Fig. 6.1).
- The observed systematic and persistent eddy SST signature stand in striking contrast to the theory-based expectation (Bretherton, 1982) of a quick erosion of mesoscale SST signals by air-sea heat exchanges within only a few weeks (the suggested heat flux feedback of $100 \text{ W/m}^2 \text{ K}$ acting on a mesoscale thermal anomaly within a mixed-layer of 50 m depth yields a persistence of 20 days).

To resolve this paradox, chapter 5.1 has used in-situ observations and climatology to derive constraints on the observed mesoscale air-sea feedback. These revealed (cf. 5.1 for a detailed conclusion) that contrary to theoretical predictions air-sea damping of mesoscale SSTA is comparable to the damping on the large-scale. As suggested by in-situ case studies, and subject to confirmation from mesoscale resolving datasets of air temperature, this reflects the fact that mesoscale SSTA induced systematic temperature anomalies in the MABL (and above), which are not quickly exported by lateral atmospheric heat transport on these small scales.

Systematic mesoscale wind stress SST coupling enhances the damping of mesoscale SSTA, but the relatively weak observed air-sea contrasts, on whose magnitude this mechanism depends, limit its scope to less than $\sim 10 \text{ W/m}^2 \text{ K}$.

- We have also studied variations in eddy T' with time and between different groups of eddies. Longer-lived eddies are observed to be more intense and show larger-amplitude SSTA.
- Both cyclones' and anticyclones' SSTA, but not their SSHA, are characterized by a systematic seasonal modulation in their amplitude (Fig. 5.13), as well as in their spatial structure (Figs. 6.9 and 6.10).
- Preliminary analysis and qualitative arguments suggest that eddy stirring of background fronts is the main driver of the above lifetime and seasonal variations of eddy SST signals.

The main point that hampers, at present, a systematic analysis of the generation mechanisms of eddy SSTA, is the somewhat unclear nature of eddy tracks (wave-like or vortex-like). Composite analyses using full SSH observations, once of sufficient accuracy and resolution, will hopefully shed light on this, making it possible to unambiguously investigate the eddy heat anomaly budget, as well as the eddy contributions

to mixed-layer heat transports and SST variability, results on which are summarized below.

7.3 Implied mixed-layer eddy heat transports

- We have established two key additive mechanisms for mixed-layer eddy heat transport. A “drift” mechanism associated with eddy-induced mixed-layer recirculations, and a “swirl” mechanism that results from phase shifts between eddy SST and SSH fluctuations, in reminiscence of linear baroclinic instability.
- In quiet regions, no systematic or a very weak eddy-induced drift is seen in track displacements, so that mixed-layer heat transport is only driven by the swirl component.
- In major current systems, observations do reveal a systematic poleward/equatorward drift of warm anticyclones/cold cyclones, reminiscent of eddy shedding. Although eddy shedding from major currents is appealing as a mechanism for heat transport, the associated drift velocity is too weak to drive substantial drift heat transports in the mixed layer. Here, too, heat is transported poleward primarily as a result of the observed westward phase shift between SST and SSH through the swirl mechanism.
- Scalings based on observations (Table 6.3) suggest that the swirl mixed-layer eddy heat transport may indeed make a substantial contribution to the poleward heat transport in the ACC and ARC region (≈ 0.1 PW), but subsurface eddy fluxes are likely important in closing the heat budget of the Southern Ocean.
- Associated mixed-layer eddy diffusivities for heat (Table 6.3) thus reflect the eddy swirl heat transport, and the observed eddy SST signature plays a key role in setting their magnitude and spatial variations.

Further observational analysis is required to confirm the observed phase shift in higher resolution SST observations, which will be challenging due to their cloud contamination. Moreover analysis of longer observational time series will yield larger sampling sizes and thus make it possible to derive eddy composites clustered by latitude or streamline as to refine the observational estimate of mixed-layer heat transport presented here.

Subsurface information would not only provide further insight into the nature of eddies, and quantify the estimate of the drift heat transport, but more importantly also to allow us to assess both, the respective contribution of mixed-layer to the total eddy heat transport, and the partition of the total between its swirl and drift components.

Are the observed phase shift, and thus the predominance of swirl heat transports, phenomena specific to the surface mixed-layer, or does the phase shift instead slowly decay through the thermocline?

Clearly observational studies need to be considered jointly to idealized modelling studies to assess the generation mechanisms of the observed phase shift, and the role of air-sea interactions in setting the eddy SST signature, and thus mixed-layer eddy heat transports.

7.4 Nature of the observed MW-SST variability

The fixed-location analysis has revealed a striking discrepancy between the spatial distributions of SST and SSH variability:

- The observed SSH variability reveals the presence of transient eddies throughout almost the entirety of NA and SO ocean basins (Fig. 3.11) and over vast extents spreading from energetic currents into basin interiors the variability is dominated by mesoscales (lower panels of Fig. 4.7). Large-scale steric signals only become important in shaping SSH variability in the far-eastern gyre interiors (Fig. 4.5).
- In contrast, SST variability occurs on large spatial scales throughout the bulk of these regions (upper panels of Fig. 4.7). Even though a vigorous eddy field is observed it therefore does not to zero-order contribute to the variability at the atmosphere's lower thermal boundary. Instead, throughout the bulk of ocean basins, this variability is primarily shaped by large spatial scales, interpreted as reflecting the forcing by the atmosphere's own synoptic fluctuations. In these regions, the main characterizing features of SST variability are thus well described by the classical paradigm of large-scale midlatitude SSTA: SSTA persistence at a given location is controlled by the weak observed large-scale air-sea interactions, resulting in a near exponential decay of its fixed-location auto-correlations and a fixed-location persistence of typically 3 months.
- The mesoscale frontal zones observed in the time-mean SST field in association with major current systems and convergence zones open a "window" through this predominant large-scale variability, in which transient mesoscale eddies are observed to make a zero-order contribution to SST variability (upper panels of Fig. 4.7).
- The spatial extent of this mesoscale window to the atmosphere is given by the regions in which intense eddy variability overlaps with strong background SST gradients, which thus points to mixed-layer eddy isotherm stirring ($-\mathbf{u}'_g \cdot \nabla \overline{T}$) as predominant

generation mechanism of the observed mesoscale contribution to SST variability. Indeed, in these regions its time-space characteristics suggest a fundamentally different nature of SST variability compared to basin interiors: Here, at any given location, forcing by the eddy stirring is balanced through a “damping” by mean advection ($-\overline{\mathbf{u}}_g \cdot \nabla T'$), with a minor role of air-sea interactions in affecting fixed-location SSTA persistence. (Following tracks, however, air-sea damping is suggested to be felt by eddy SSTA). Along the ACC path, the SST front of the GS North wall and to a minor degree the NA subtropical convergence zone, (atmospherically-forced) large scales and (mesoscale eddy-forced) small scales thus jointly shape the observed variability in SST.

- We have revealed a systematic seasonal modulation in the nature of observed SST variability (chapter 4.4), characterized by an intensification of the mesoscale contribution and a basin-wide shift towards smaller predominant scales of SST variability in winter with respect to summer. This has been shown to reflect the joint effect of an increase in the efficiency of the mesoscale forcing and a reduction in the efficiency of the large-scale atmospheric forcing of SST variability in winter. The mechanism of the former being the observed winter-time intensification of background SST fronts, that of the latter the observed deeper winter mixed-layer layers. (The chosen SP quiet regions form an exception to this rule, as here the summer-time widening of ACC fronts to the south leads to an intensification of background SST gradients in summer.)
- Although of varying amplitude, a mesoscale contribution to SST variability is thus anchored by the main SST fronts year-round.

There thus appears to be a simple zero-order explanation for both the spatial and temporal variations in the predominant nature of SST variability. As large-scale atmospheric forcing recedes (over deeper mixed-layers or zero wind-stress lines), it reveals a larger mesoscale contribution to the variability, as long as both deep and surface oceanic fronts are available to interplay to generate mesoscale forcing: the former are at the origin of the eddying circulations, but in the absence of the latter (over the GS warm tongue, or in summer) the most vigorous eddy stirring does not generate SST variability. In basin interiors, deep and surface fronts are typically available (e.g. GGS), but both are too weak, so that mesoscale variability is masked by much larger amplitude atmospheric forcing. Anomalous exchanges between the mixed-layer and the subsurface ocean do not need to be evoked in this zero-order picture, but it would be interesting to confirm their role in a systematic track-following analysis using data

from the Argo network.

The spatial scale diagnostic developed here from the SSTA autocorrelation function has been revealed to be a simple tool to visualize the contribution of different spatial scales to the variability in SST. It would be fascinating to use this simple tool to visualize the large-scale and mesoscale midlatitude ocean-atmosphere interactions modelled in coupled simulations on their way to resolve the mesoscale. Other, observational results of this thesis, also provide important observational benchmarks for modelling studies, specifically the climatological benchmark estimate for air-sea exchanges on the mesoscale, and, although more challenging to diagnose, the observed eddy SST signatures themselves.

Here we have clearly demonstrated a mesoscale contribution to SST variability, and via the weak thermal air-sea damping, also to temperature variability in the MABL. Observations show that this leads to a systematic transient coupling between SSTA and the MABL, inducing small-scale surface convergence and vertical velocities, and regional case studies suggest the induced perturbations might indeed extend far higher into the troposphere (cf. chapter 5.1; Liu et al., 2007). This suggests two key questions to be resolved to further our understanding and realistic modelling of midlatitude ocean-atmosphere interactions:

Do we need to account for the presence of fully nonlinear oceanic vortices to accurately capture this mesoscale variability, or does the bulk of it simply reflect wave-like perturbations on the oceanic mean flow?

Does the fuzziness and bumpiness induced by transient ocean mesoscale eddies in the MABL and higher up, which is characterized by small scales and lower frequencies compared to the scales of atmospheric fluctuations, represent a simple noise, or may it instead have a systematic impact on the larger-scale atmospheric circulation?

It will be fascinating to progress on these questions from joint observational analysis of the exciting new near-mesoscale resolving datasets that currently become available for both the subsurface ocean and the atmosphere.

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APPENDIX

A Sensitivity of results to the seasonal background state

This appendix provides a detailed discussion of the definition of SST and SSH variability with respect to a seasonal background state estimated empirically from short observational time series (section A.1). In particular, it examines the partition of SSH variability into seasonal and non-seasonal components (section A.2), and concludes with an assessment of the biases in observed composite anomalies following eddy tracks (section A.3).

A.1 Variability as deviation from an observed seasonal mean

Variability in SST (and SSH) features both seasonal fluctuations in the background state ($\bar{T}$) as well as non-seasonal variability (T'):

$$T = \bar{T} + T'. \quad (\text{A.1.1})$$

To study the partitioning of the latter into oceanically and atmospherically forced components, it is crucial to separate it from the - at least in the case of SST predominant - contribution of the former.

To do so, following a standard practice in the literature (e.g. Frankignoul, 1985; Cayan, 1992a), the observed seasonal background state ($\bar{T}_{obs}$), is empirically estimated at each given location as:

$$\bar{T}_{obs} \equiv \frac{1}{n} \sum_{yr=1}^n T(yr) = \bar{T} + \frac{1}{n} \sum_{yr=1}^n T'(yr), \quad (\text{A.1.2})$$

where (A.1.1) is used to obtain the second expression and yr denotes the time-coordinate in years. Removing (A.1.2) from (A.1.1), the observed (non-seasonal) temperature anomaly T'_{obs} is then obtained as:

$$T'_{obs} \equiv T - \bar{T}_{obs} = (\bar{T} + T') - \bar{T}_{obs} = T' - \frac{1}{n} \sum_{yr=1}^n T'(yr). \quad (\text{A.1.3})$$

(A.1.2) and (A.1.3) clearly show that, as introduced in chapter 3.1.3, for finite time series length (of n years) this method induces biases, which are the more important the smaller n .

Specifically, the estimated seasonal cycle will retain a component of non-seasonal variability (from the 2nd term in (A.1.2)), which therefore is “missing” in observed anomalies (from the 2nd term in (A.1.3)). As seen from (A.1.3), observed anomalies will be moreover by construction anti-correlated to the anomalies observed at the same location and time in the previous and following years. This is illustrated in Fig. A.1, which displays the auto-correlation functions of MW-SSTA and multi-altimeter SSHA (obtained in the same way, refer to the next section for a discussion) observed during the June 2002– December 2007 period ($n \sim 5\frac{1}{2}$) in energetic and quiet regimes of NA and SO. Observed anomalies typically decorrelate within a few months (see also chapter 5.2), so that the artificially induced negative correlations at lags of 1 and 2 years clearly stand out from the otherwise likely not significantly non-zero correlations observed at long time lags. The Figure moreover illustrates how, due to the finite temporal resolution capacity of the data and, more importantly, the observed fixed-location anomaly persistence (of several months), the induced negative correlations are not limited to lags corresponding to exact multiples of 365 days, but induce slowly decaying negative biases in observed auto-correlations within time-windows of a few months. (Consistent with the observed slower fixed-location anomaly decay, these are slightly larger in quiet than in energetic regions.)

These artificial features in observed anomaly statistics are clearly unsatisfactory and highlight the need for further research into i) the sensitivity of results presented in this thesis with respect to biases in the seasonal background state, and ii) devising methods to minimize these biases for questions, on which only short data records are available.

Aspect ii) is specifically important for studies of mesoscale processes, as high-resolution satellite based observations are today typically limited to ~ 10 years, whereas studies of large-scale variability, which are moreover, due to the use of monthly averages, substantially less affected by noise, can usually draw on records of over half a century (e.g. Cayan, 1992a; Frankignoul et al., 1998). Recent studies of mesoscale signals in SSH, SST and wind stress thus often resort to spatial high-pass filtering to obtain anomalies (e.g. Chelton et al., 2004, 2011b), but this approach does not straightforwardly allow to study transient processes, as in regions of steep frontal or bathymetric gradients substantial time-mean and seasonal components are retained

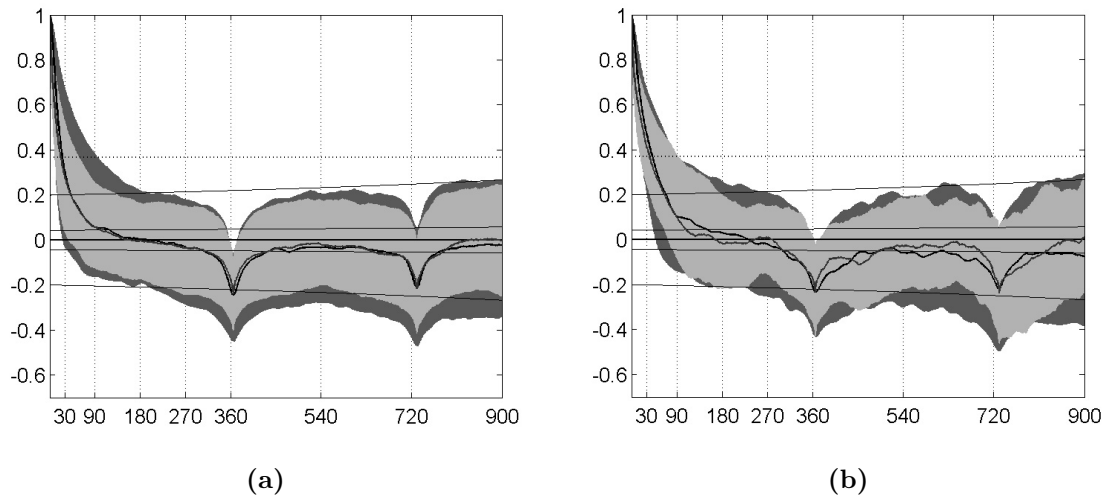


Fig. A.1: Autocorrelation functions of SSTA (grey) and SSHA (black) observed in energetic (a) and quiet regions (b) of NA and SO (lines shows the median, shadings the 5-95 percentile range). (Thin lines indicate where correlations are significantly different from zero at 95% confidence if one observation every day/every 21 days is independent.)

in the resulting small-scale anomalies. In this light, a, perhaps cleaner, approximate approach may be to obtain $\bar{T}_{obs}$ from sinusoidal fits to the data (i.e. as annual/semian-annual harmonic), a method also regularly used in previous studies (e.g. Leeuwenburgh and Stammer, 2001). Alternatively, day to day noise in $\bar{T}_{obs}$, obtained using (A.1.2), may be reduced through temporal smoothing (e.g. by using monthly running means) before its removal to obtain anomalies.

Further analysis on aspect i) is provided in the remaining two sections of this Appendix. Specifically, section A.2 assesses possible biases in the estimated separation of SSH variability into seasonal and non-seasonal components, and, importantly, section A.3 shows that the biases in observed eddy composite anomalies, upon which most key results of this thesis on eddy SST signatures and mixed-layer heat transports are based, are relatively well constrained.

A.2 Seasonal versus non-seasonal variability in SSH

The separation of SSH variability into seasonal and non-seasonal components is not as clear-cut as in the case of SST. Whereas amplitudes of SSTA are typically almost an order of magnitude smaller than the amplitude of the seasonal SST cycle, departures of SSH from its climatology are comparable in magnitude to the time-variations of the

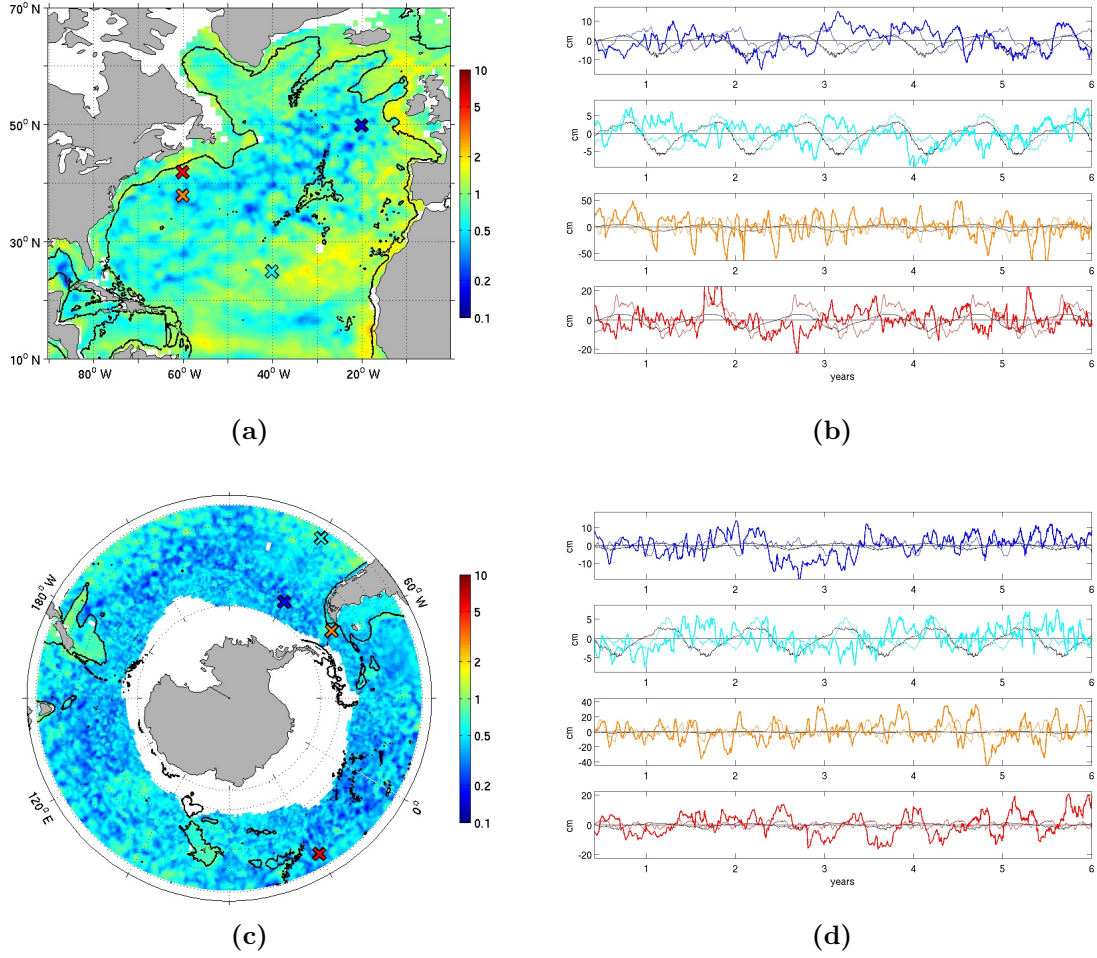


Fig. A.2: The colour panels display $\sigma(\bar{\eta})/\sigma(\eta')$, the ratio of seasonal to non-seasonal SSH variability, for NA (a) and SO (c). Examples of the associated time-series of η' (thick) and $\bar{\eta}$ (thin) are shown in (b, d), in the same colour as the crosses marking their location in (a, c). Thin black curves show the seasonal cycle of mixed-layer thermosteric height $\bar{\eta}_s = \alpha_T \bar{T}' \bar{h}_{ml}$. Y-axes ranges in (b, d) vary corresponding to local SSHA r.m.s.

SSH climatology itself. (Here the SSH climatology is obtained in the same way as its SST counterpart, through use of (A.1.2), and as described in chapters 3.1.3&3.2.3).

This is illustrated in Figure A.2. Panels A.2a and A.2c show the ratio $\sigma(\bar{\eta})/\sigma(\eta')$ for NA and SO, respectively. The amplitudes of the two fields differ generally by less than a factor 2. The ratio is largest and seasonal scales dominate the variance of SSH in parts of the tropical and subtropical Atlantic and over continental shelves and especially their edges, which are indicated by the 2 km bathymetry contour. Smallest values are found in the central deep parts of ocean basins, along the extended Gulf

Stream and the ACC. Here SSH variance is largest (not shown) and resides primarily on non-seasonal time scales. The generally larger ratios found in the NA compared to the SO are part of an inter-hemispheric asymmetry in the strength of the seasonal SSH cycle, which has been previously noted by a series of studies (e.g. Stammer, 1997b; Vinogradov et al., 2008). Examples of time-series of η' (thick) and $\bar{\eta}$ (thin) are displayed in panels A.2b and A.2d for the locations indicated by the crosses in the respective maps. Where the contribution of the seasonal variance is large, $\bar{\eta}$ has a smooth, near sinusoidal time-evolution, as expected for a seasonal cycle. However in regions of large SSH variances and small seasonal contributions, best illustrated by the orange curves, $\bar{\eta}$ varies on relatively high frequencies. These are likely an artefact of the short averaging time-span and the “true” seasonal cycle is expected to be smoother and of smaller amplitude. In these regions a substantial amount of non-seasonal SSH variability is therefore retained in $\bar{\eta}$ instead of η' . Scaling, observational and modelling studies (Gill and Niller, 1973; Vinogradov et al., 2008) have established that the seasonal SSH cycle, after removal of the response to seasonal atmospheric pressure loads, primarily reflects the storage of heat in the seasonal thermocline in response to local air-sea heat-fluxes. In comparison, local freshwater forcing is negligible in the regions considered, but seasonal variations in wind-stress and its curl are important in the tropical Atlantic and in the Southern Ocean, where they excite a primarily barotropic SSH response (which is in large parts removed by the dynamic atmospheric correction applied by AVISO). The seasonal air-sea heat flux variance peaks off the North-American coast where midlatitude westerlies bring continental air over the ocean and the GS extension and its distribution explains the broad regional pattern and amplitude of the seasonal SSH cycle including the observed inter-hemispheric asymmetry. The thin black curves in Fig. A.2b,d show the seasonal cycle of the mixed-layer thermosteric height $\bar{\eta}_s = \alpha_T \bar{T}' \bar{h}_{ml}$, obtained by integrating the expansion/contraction associated with seasonal SST anomalies over the mixed-layer depth $\bar{h}_{ml}$. Here $\bar{h}_{ml}$ is evaluated from a climatology, introduced in Appendix B.1, and the thermal expansion coefficient of sea water $\alpha_T = \frac{-1}{\rho_0} \frac{\delta \rho}{\delta T}$ is set to its climatological surface distribution, estimated as described in Appendix B.5. $\bar{\eta}_s$ agrees generally well in amplitude with the full $\bar{\eta}$. Especially in the regions of weak SSH variance, $\bar{\eta}$ and $\bar{\eta}_s$ correspond very well also in phase. Bearing in mind that $\bar{\eta}_s$ reflects only a part of the heat stored in the seasonal thermocline, as it neglects variations in α_T (which become particularly important at high latitude) and the heat stored below the shallowing summer-time mixed-layer (which could explain the observed phase differences), this broad agreement shows that the

SSH seasonal cycle indeed mainly reflects surface thermosteric effects and gives some confidence in its estimate from the 5½-year average everywhere apart from the regions of highest SSH variance.

Amplitudes of the seasonal air-sea heat flux forcing vary spatially and therefore induce seasonally varying SSH gradients. However as induced SSH are mostly large-scale, these gradients and associated surface flows are weak and the full SSH gradient shows little seasonal variability. Pressure gradients are moreover compensated below the shallow seasonal thermocline ($\sim$ the top 200 m) and therefore associated with small seasonal transport variations, so that seasonal SSH variability can be regarded as largely dynamically passive. (The seasonal wind-forced barotropic variability however, although associated with usually small SSH amplitudes, can lead to substantial transport variations in deep basins (Gill and Niller, 1973), but is mostly not represented in this dataset.) Consistently the transport and position of major currents does not show any significant variability on seasonal time scales (Frankignoul et al., 2001; Kelly et al., 2010), although this had been postulated by early studies (Kelly et al., 1999) based on short time-series that did not allow to distinguish interannual from seasonal variations. This is not true for the small spatial scales retained in $\bar{\eta}$ in this dataset that are associated with significant surface currents on the mesoscale. Since conclusions on the large-scale, dynamically passive nature of the seasonal SSH cycle are based on non-mesoscale resolving observational and reanalysis studies (Gill and Niller, 1973; Stammer, 1997b; Vinogradov et al., 2008), the possibility that mesoscale features in $\bar{\eta}$ are not entirely an averaging artefact, but an observed feature of the SSH seasonal cycle cannot be excluded. An analysis of longer time-series is necessary to assess this, but the fact that long-term (12 to 15 year) linear trends in multi-altimeter SSH still reveal a significant amount of mesoscale structure (Sokolov and Rintoul, 2009; Hughes et al., 2010) favours this possibility.

A.3 Systematic and random biases in eddy composite anomalies

Estimating the characteristic SST perturbation of propagating eddies by composite averaging is only approximate when a finite size of the records is available. Variability in SST associated with processes other than the mesoscale circulation of interest here (e.g. atmospheric forcing via Ekman advection and surface heat exchange), as well as AMSR-E measurement errors (cf. chapter 3.1.2), induce essentially random errors, which can be greatly reduced by averaging over many tracks and measurements. The fact that SST perturbations are evaluated with respect to an imperfect estimate of

the seasonal background state (cf. section A.1) induces additional biases. As shown below, these are not purely random, but also contain a systematic component, which is not reduced by the use of a large number of tracks.

To see this, consider the hypothetical case in which we have n years of observations at a given location and only during the last of those a warm eddy, characterized by a SST perturbation δT_o , passes. The empirical estimate of the seasonal cycle $\bar{T}_{obs}$, obtained from the observations using (A.1.2) and as described in chapter 3.1.3, then yields

$$\bar{T}_{obs} = \bar{T} + \frac{0 + \dots + 0 + \delta T_o}{n} = \bar{T} + \frac{\delta T_o}{n}, \quad (\text{A.3.1})$$

which differs from $\bar{T}$, the true seasonal cycle¹. Thus, whereas the true temperature anomaly T' at the given location along the track by definition equals to δT_o

$$T' \equiv T - \bar{T} = \delta T_o, \quad (\text{A.3.2})$$

the empirical estimate, obtained using (A.1.3), is only

$$T'_{obs} \equiv T - \bar{T}_{obs} = (\bar{T} + \delta T_o) - \bar{T}_{obs} = \frac{n-1}{n} \delta T_o < \delta T_o, \quad (\text{A.3.3})$$

after use of (A.3.1) and (A.3.2). For finite observational time series length n (in years), we thus expect to underestimate the true eddy signature in SST by the factor $(n-1)/n$.

To account for more realistic situations, in which eddies can pass a given location in all years and SST variability includes the above mentioned noise due to non-eddy variability, we generate synthetic data of T along the tracks of propagating eddies and estimate the composite average $\langle T'_{obs} \rangle$ in the same way as in observations (e.g. chapter 4.3.1). All tracks are chosen to be of 16-week duration (a relatively short duration, which gives a more stringent test), and are described by the coordinate $s(x,y,t)$, which follows each track in time and space. In this more general case, the empirical estimate of the seasonal cycle along a track, from (A.3.1), becomes:

$$\bar{T}_{obs}(s) = \bar{T}(s) + \frac{1}{n} \sum_{yr=1}^n T'(s + (0, 0, yr)), \quad (\text{A.3.4})$$

where the notation $s + (0, 0, yr)$ indicates the same location on a given track s at a given year yr . The observed temperature anomaly along the track, introduced by (A.3.3), is thus given by:

$$T'_{obs}(s) \equiv T(s) - \bar{T}(s) = T'(s) - \frac{1}{n} \sum_{yr=1}^n T'(s + (0, 0, yr)), \quad (\text{A.3.5})$$

¹ Note that here we consider only transient eddy perturbations, which by definition are deviations from the true seasonal background state.

in which both the passing of mesoscale eddies (T'_o), as well as atmospheric forcing (T'_a) contribute to the true temperature anomaly along the track ($T'(s) = T'_o(s) + T'_a(s)$).

As in the simple example above, we assume that an eddy's temperature perturbation $T'_o(s) = \delta T_o$ is constant along its track and the same for all tracked eddies. The passing of eddies along the eddy track in the other $n-1$ years (i.e. at $s + (0, 0, yr)$), also enters the estimate of T'_{obs} (from the second term in A.3.5), and is simulated as

$$T'_o(s + (0, 0, yr)) = \delta T_o \sum_{i=1}^3 \cos(\Lambda(yr)ks + \Phi(i)). \quad (\text{A.3.6})$$

Here the summation over 3 “waves” with random phase Φ reflects variation in the eddy polarity (warm or cold eddies) and amplitude (0 to $3 \delta T_o$), and for each year (yr) it is randomly chosen whether eddies propagate along the track ($\Lambda = 0$) or across ($\Lambda = 1$). In the latter case, the along-track wavelength $2\pi/k$ is set to 6 weeks (larger k are unrealistic given the observed eddy widths and propagation speeds, and for smaller k we essentially recover the case $\Lambda = 0$).

The atmospherically forced SST variability $T'_a(s)$ is assumed, for a harder test, to be correlated in space along each entire eddy track. It thus reduces to the temporal anomaly $T'_a(t)$ in the area traversed by the eddy. As is typical in such problems (e.g. Frankignoul, 1985), we simulate the associated time series as a 1st order autoregressive process (AR-1), with a decorrelation time of 4 months. The standard deviation of this process is denoted by $\sigma(T'_a)$ and, together with the number of tracks considered, is a controlling parameter in these simulations.

The AR-1 model and (A.3.6) determine the anomalies in (A.3.5) along an arbitrary number of tracks. Their weekly snapshots are then composite-averaged to obtain $\langle T'_{obs} \rangle$, as was done in the track-following analysis presented throughout chapters 4,5 & 6 for the SST and SSH data.

Figure A.3 displays the mean, upper and lower 5 percentiles, of $\langle T'_{obs} \rangle$ as a function of the number of tracks considered, based on a thousand independent realizations of the random processes described above for T'_o and T'_a . Dashed, black and grey curves display three different experiments, for which the ratio $\sigma(T'_a)/\delta T_o$ is set to 0, 1 and 5, respectively. The case $\sigma(T'_a)/\delta T_o = 0$ represents the ideal situation, in which no atmospheric forcing is present. As seen in Fig. A.3 (dashed curves), the normalized composite $\langle T'_{obs} \rangle / \delta T_o$ then rapidly approaches the limit $\frac{n-1}{n} \approx 0.82$ for $n = 5.5$ (half the tracks with 5 years, half with 6 years, composited in random order). Indeed, in 90% of realizations, the normalized composite is bounded by $\pm 2\%$ of this value as soon as more than 1000 tracks are composited.

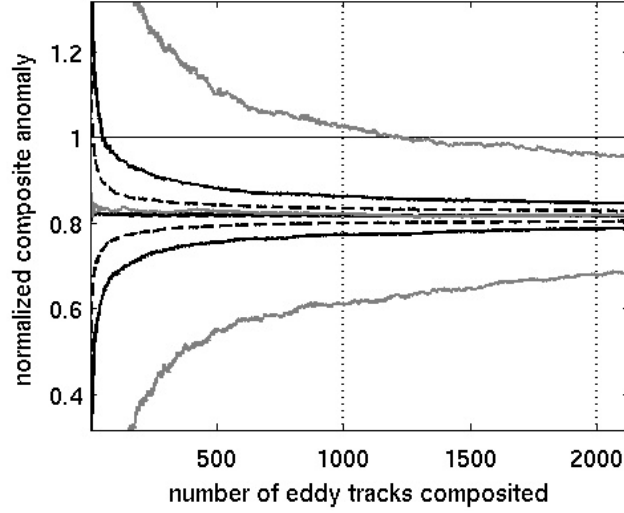


Fig. A.3: 1000 realization average, and upper/lower 5 percentiles, of the observed composite $\langle T'_{obs} \rangle$, normalized by δT_o , for three different experiments: $\sigma(T'_a)/\delta T_o = 0$ (dashed), 1 (black) and 5 (grey). As the number of tracks composited increases (abscissa), $\langle T'_{obs} \rangle/\delta T_o$ converges towards its expected value $\frac{n-1}{n}$ (≈ 0.8), where n is the observational time series length in years. The vertical dashed lines indicate the typical number of tracks considered in our study.

The enhanced spread in the black and grey lines in Fig. A.3 reflects masking of the eddy signal by atmospherically forced variability (term 1 in (A.3.4)), which also enhances the random bias in the estimated background seasonal state (term 2 in (A.3.4)). Here, in 90% of cases, the normalized eddy composite $\langle T'_{obs} \rangle/\delta T_o$ lies within $\pm 5\%$ (for $\sigma(T'_a)/\delta T_o = 1$, black) and 25% (for $\sigma(T'_a)/\delta T_o = 5$, grey) of $\frac{n-1}{n}$ once more than a thousand tracks are considered. (Note that in both cases the atmospherically forced variability present in individual eddy snapshots is reduced by more than 95% in the eddy composite, i.e. $\langle T'_a \rangle/\sigma(T'_a) < 0.05$.)

In summary, the observed composites systematically underestimate the true eddy perturbation by the factor $\frac{n-1}{n}$. At the 90% confidence interval, they lie within a narrow error margin around this value, both for SSHA ($< 2\%$, likened to the $\sigma(T'_a)/\delta T_o = 0$ experiment (dashed) since the large-scale steric to eddy SSHA variance ratio is observed to be small) and for SSTA in energetic regions ($< 5\%$ for the 2000 tracks typically composited in these regions (chapter 3.3.3), taking the $\sigma(T'_a)/\delta T_o = 1$ experiment (black) to be the relevant regime since observed SSTA have comparable large-scale and mesoscale variance in energetic regions). The $\sigma(T'_a)/\delta T_o = 5$ experiment (grey) is

the relevant regime for SSTA composites in quiescent regions. Their larger error margins ($<25\%$ for >1000 tracks composited, and the hard set of parameters chosen here) reflect the non-perfect averaging out of large-scale atmospherically forced variability predominating the SSTA variance in these regions.

B Climatological and other datasets

This Appendix provides, for reference, the technical details on a series of climatologies and related datasets that are introduced and used throughout this thesis to provide the context for the analysis of mesoscale SST and SSH observations. Before a description of datasets related to the oceanic time-mean flow (B.3, B.4 and B.5) and potential vorticity dynamics (B.6 and B.7), we introduce datasets characterizing ocean-atmosphere interactions through the depth of the ocean surface mixed-layer (B.1) and the climatological conditions at the air-sea interface (B.2).

Additionally, in-situ observations from a variety of field programs dedicated to the study of air-sea interaction and mesoscale circulations are consulted and introduced in chapter 5.

B.1 Mixed-layer depth

As discussed in section 2.2, the depth of the oceanic surface mixed-layer in direct contact with the atmospheric boundary layer through exchanges of buoyancy and momentum is crucial for the dynamics of SSTA. Controlled not only by the surface forcing, but also by the oceanic flow, especially on meso- and submesoscales (e.g. Dewar, 1986; Boccaletti et al., 2007), its depth varies on diurnal, seasonal and inter-annual time-scales.

Here we only capture its large-scale seasonal variations, as we use of a large-scale (2°) monthly mixed-layer depth climatology. It has been derived by de Boyer Montégut et al. (2004) from all vertically high-resolved, in-situ hydrographic profiles collected between 1941 and 2002. The climatology is obtained as the median of the mixed-layer depth estimated from each individual profile available within a given month and grid-box. This represents a significant improvement in accuracy compared to traditional climatologies derived from gridded hydrographic data. The mixed-layer depth is defined as the depth at which temperature differs by more than 0.2 K with respect to the temperature at 10-meter depth. The magnitude of the finite-difference criterion is chosen empirically such that it stands out from measurement noise, but does not yield deeper (larger) temperature discontinuities such as the seasonal thermocline. It is not defined directly with respect to the surface, but with respect to a depth of 10-meter, in order to avoid both near-surface measurement noise and diurnal mixed-layer variability.

The use of a temperature instead of a density based mixed-layer depth climatology

is most intuitive and adequate for our study as it indicates the depth to which in-situ calibrated MW SST are, in a climatological sense, a good proxy for temperature. The temperature-difference based definition generally yields the physically relevant mixed-layer depth. Only in low-latitude barrier layer regions, in which surface fresh-water layers created by intense seasonally varying precipitation define the layer directly in contact with the atmosphere, and in regions of high-latitude deep convection, where density is only weakly dependent on temperature and can be homogenized over a much deeper layer than temperature, the temperature-based definition is less adequate. Most importantly however, the spatial resolution and accuracy of a temperature-only mixed-layer depth climatology is by far superior to a density-based climatology that relies on extremely sparse salinity observations. Over vast extents of the Southern Hemisphere there are virtually no salinity profiles apart for the austral summer season. The 2-10 temperature profiles available throughout the seasons in most SO 2° -by- 2° boxes form a much better data base. Given the ~ 50 -1000 profiles available per grid-box in the NA, the resulting climatology is however still characterized by a strong inter-hemispheric asymmetry in data content.

The resulting spatial and seasonal distribution of mixed-layer depth in NA and SO have been displayed in Fig. 2.8 and discussed in chapter 2.2.

In this thesis, mixed-layer depth observations are shown to provide important insights into the nature of observed SST variability and its seasonal modulation. However, the coarse resolution of the mixed-layer depth climatology is one of the major sources of uncertainty in the estimates of eddy mixed-layer heat budgets and transports that may be highly sensitive to potential eddy-induced mixed-layer depth variations. Future analyses will thus greatly benefit from current ARGO float observations which start to build up a dataset that resolves the temporal and spatial variability of the mixed-layer down to mesoscales.

B.2 Air-sea climatology

To characterize climatological conditions at the air-sea interface we refer to the da Silva et al. (1994) Atlas of Surface Marine Data, which is available at the Lamont Doherty Earth Observatory (LDEO) Climate Data Library. It provides 1° monthly climatologies of a large range of variables, from which surface wind speed, the air-sea temperature difference, relative humidity and, for consistency, SST itself are consulted in this

study. Obtained as objective analysis² of the Comprehensive Ocean-Atmosphere Data Set (COADS), these are truly large-scale climatologies whose resolution is restricted to wavelengths greater than ~ 770 km. With COADS being mainly based on merchant ship observations, their underlying data content is not homogeneous in space and time, featuring severe limitations in data coverage in the Southern Ocean.

The saturation specific humidity q_{sat} , which gives the ratio of mass of water vapour (ρ_v) to moist air (ρ_a) at saturation, and is such a function of both temperature and (total) pressure p , is obtained as:

$$q_{sat}(T, p) = \frac{\rho_v}{\rho_a} = \frac{\epsilon e_{sat}(T)}{p - e_{sat}(1 - \epsilon)}, \quad (\text{B.2.1})$$

where $\epsilon = \frac{R_d}{R_v} \approx 0.62$ is the ratio of the gas constants of dry air and water vapour. In this, the saturation vapour pressure $e_{sat}(T)$ is a function of temperature T only, determined by the Clausius-Clapeyron relation:

$$\frac{de_{sat}}{dT} = \frac{1}{T} \frac{L}{1/\rho_v - 1/\rho_c}. \quad (\text{B.2.2})$$

Here ρ_v and ρ_c denote the densities of vapour and condensed phases. For $\rho_c \gg \rho_v$, it can be approximated as $Le_{sat}/R_v T^2$, using the ideal gas law for water vapour. Here we use the integrated form of (B.2.2), in the version of the MIT GCM package: $e_{sat}(T) = 6.112 e^{17.67T/(243.5+T)}$, which gives $e_{sat}(T)$ in hPa as function of T in $^{\circ}\text{C}$.

B.3 Wind stress

As discussed in chapter 2.1.2, the stress (2.9) exerted on the ocean surface by atmospheric surface winds (with respect to the oceanic flow) is a fundamental parameter in the oceanic circulation.

In this study we use a monthly wind stress climatology, representative for the 1999-2009 period, which is derived from near-mesoscale resolving QuickSCAT scatterometer winds and is provided on a $1/4^{\circ}$ grid as Scatterometer Climatology of Ocean Winds (SCOW) by Risien and Chelton (2008)³. Together with the resulting Ekman pumping

² A comprehensive documentation of the analysis method and applied bias corrections can be found at <http://iridl.ldeo.columbia.edu/SOURCES/.DASILVA/.SMD94/.dataset/.SMD94/>

³ Although QuickSCAT measurements of the active microwave radiation scattered back from the rough sea surface are a function of wind stress, they are calibrated to 10-m neutral-stability wind observations due to a lack of in-situ stress measurements. In this climatology, individual QuickSCAT measurements have been reconverted to wind stress using bulk aerodynamic formula with the Large et al. (1994) formulation for the (neutral-stability) drag coefficient before the mapping process which uses a 70 km smoothing to fill in gaps due to rain. SCOW is provided at <http://numbat.coas.oregonstate.edu/scow/>.

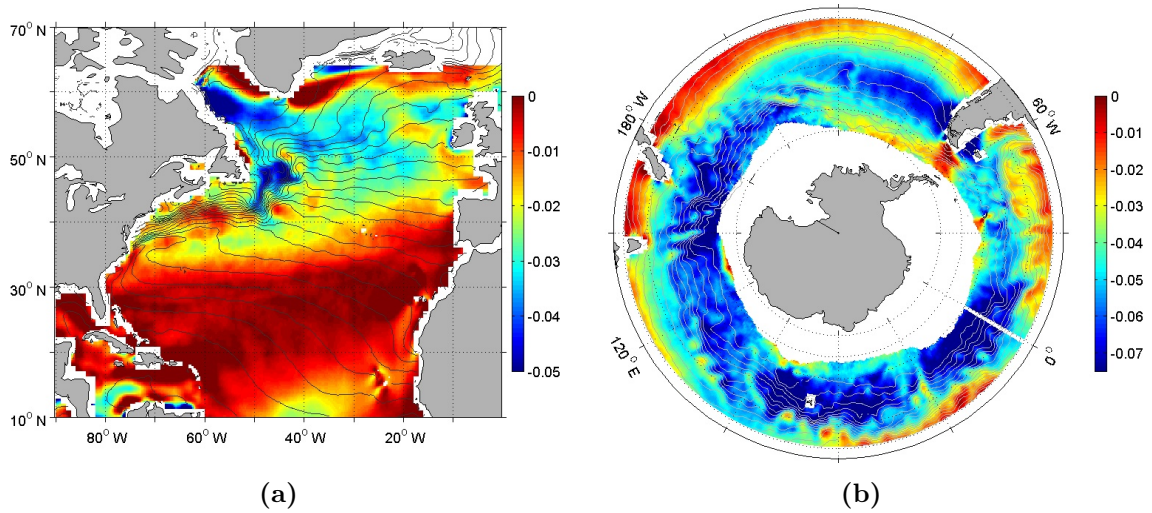


Fig. B.1: The difference in the magnitude of the zonal wind stress between the QuickSCAT and the NCEP derived climatologies, in Nm^{-2} , overlaid over climatological MW SST contours, highlights the superior resolution capacity of QuickSCAT compared to NCEP observations, and the SST dependence of small-scale features in the observed wind-stress.

w_{ek} , the climatological distribution of τ has been displayed in Fig. 2.4.

Here we provide a quick comparison of the SCOW climatology to a climatology obtained by seasonal-averaging of 10 years (also 1999-2009) of monthly wind stress data from the National Center for Environmental Prediction (NCEP) reanalysis 2 (Kanamitsu et al., 2002)⁴. The T62 spectral resolution of the NCEP reanalysis model limits the features that are resolved in this coarse-grid climatology (1° -longitude by $\frac{1}{3}^\circ$ -latitude) to scales larger than at best ~ 300 km.

The difference between QuickSCAT and the NCEP-derived zonal wind stress magnitude is mapped in Fig. B.1 on top of time-mean SST contours. The Figure shows that, apart from generally weaker wind stress in the SCOW climatology, the large-scale wind stress distribution of the two climatologies agrees relatively well. However, the Figure reveals pronounced small-scale differences between the two climatologies. Not resolved in the coarse NCEP climatology, these are physical features, that are directly related to observed intense oceanic flows and narrow mesoscale structures in the time-mean SST field.

⁴ This wind-stress reanalysis is used to force NCEP's Global Ocean Data Assimilation System (GODAS) and is provided together with GODAS outputs by the LDEO climate data library at <http://iridl.ldeo.columbia.edu/SOURCES/.NOAA/.NCEP/.EMC/.CMB/.GODAS>.

B.4 Mean Dynamic Topography

An estimate from altimetry of the time-mean ocean surface flow in geostrophic balance with a mean dynamic topography is challenging as it requires an accurate model of the geoid. Using the geoid model derived from the Gravity Recovery and Climate Experiment (GRACE) mission, it is presently only available for spatial scales larger than 200-300 km (Rio et al., 2011). Several techniques have therefore been developed that reconstruct the missing small scales of the mean dynamic topography using in-situ observations.

One of the first such estimates, the Maximenko and Niiler dynamic topography uses surface velocity observations from satellite-tracked drifters (from the Global Drifter Program). After removal of the Ekman flow field derived from NCEP reanalysis winds and the time-varying geostrophic surface flow derived from altimetry, these yield an estimate of the time-mean geostrophic surface circulation, which is merged with a large-scale mean dynamic topography, using the method described in Maximenko and Niiler (2005). The resulting mean dynamic topography, representative for the 1992-2002 period, is provided globally on a $\frac{1}{2}^\circ$ -grid⁵.

The mean dynamic topography component of the SSH dataset used in this thesis (cf. section 3.2.3) is obtained as the 5 $\frac{1}{2}$ -year time-average of η , and will be denoted $\bar{\bar{\eta}}$ in the following. $\bar{\bar{\eta}}$ is displayed in Fig. 2.1, and is based on the June 2002–December 2007 average of the merged multi-altimeter sea level anomaly (referenced to 1993-1999) and on a mean dynamic topography (for the 1993-1999 period). The latter has been developed by Rio et al. (2005) merging geoid and altimeter observations not only with surface drifter velocities, but also with dynamic height calculated from in-situ hydrographic temperature and salinity profiles (adjusted for the flow at the reference level). $\bar{\bar{\eta}}$ therefore represents an improvement compared to the Maximenko and Niiler mean dynamic topography $\bar{\bar{\eta}}_{MAX}$.

Recently, Rio et al. (2011) have developed an even higher-resolution mean dynamic topography (CNES-CLS09), referred to here as $\bar{\bar{\eta}}_{CLS}$. It includes a better Ekman model and better wind-data (ERA INTERIM instead of NCEP) compared to the Rio et al. (2005) estimate, and most importantly 6 years of hydrographic measurements from ARGO floats, which dramatically enhances its in-situ data content, especially in the Southern Ocean.

Figure B.2 shows characteristic $\bar{\bar{\eta}}$ contours of all three datasets superimposed on

⁵ The 1992-2002 mean ocean dynamic topography (version April 2006) has been obtained from Nikolai Maximenko (IPRC) and Peter Niiler (SIO).

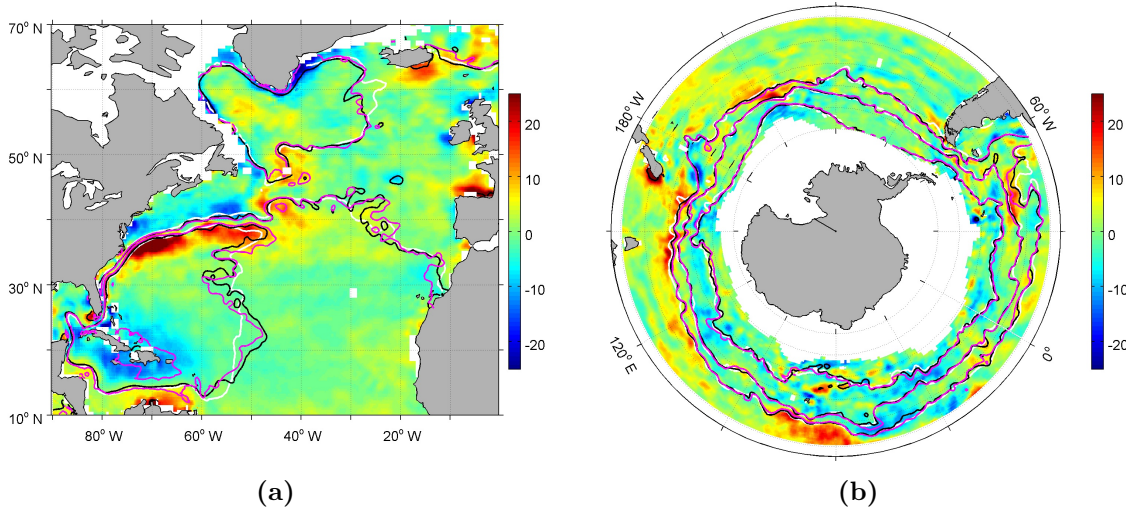


Fig. B.2: Comparison between the mean-dynamic topographies for the 5½-year study period. Colours display the difference between $\bar{\eta}_{CLS}$, based on CNES-CLS09, and $\bar{\eta}$, based on Rio2005, in cm. Contours show characteristic values, namely -45, 5 and 45 cm in the NA (a) and -55, 0 and 40 cm levels in the SO (b), of $\bar{\eta}_{CLS}$ (magenta) and $\bar{\eta}$ (black). Corresponding $\bar{\eta}_{MAX}$ contours for the 1992-2002 period are also shown (white). The basin-average SSH has been removed from all datasets.

each other for NA and SO. The general features of $\bar{\eta}$, including a nearly 2 m drop from subtropical to subpolar gyres, the intensified gradients associated with GS and ACC and bathymetric control, are seen in all three datasets. However, the details of current structures and velocities show important differences. This is illustrated by the colour-maps in Figure B.2, which display the difference between $\bar{\eta}_{CLS}$ and $\bar{\eta}$. Apart from big discrepancies between the two datasets over continental shelves, the Figure reveals the much steeper time-mean SSH gradients captured by the $\bar{\eta}_{CLS}$ dataset, which features stronger and narrower ACC jets, an intensified GS recirculation gyres, and stronger zonal jet-like features in the South Pacific. Although uncertainties remain, $\bar{\eta}_{CLS}$ thus provides the most realistic, highest-resolution estimate.

B.5 Hydrography

To estimate the time-mean climatological surface distribution of the thermal expansion coefficient of sea water $\alpha_T = \frac{-1}{\rho_0} \frac{\delta \rho}{\delta T}$, we apply the algorithm of McDougall (1987) to the time-mean MW SST and the World Ocean Atlas 2005's sea surface salinity field (Antonov et al., 2006, obtained from the LDEO climate library). Predominately

constrained by SST, α_T is observed to vary around a midlatitude value of $.2 \cdot 10^{-3} \text{ K}^{-1}$ (at SST $\sim 15 \text{ K}$), from $\sim .1$ to $.3 \cdot 10^{-3} \text{ K}^{-1}$ for SST of ~ 5 and 25°C , respectively.

The map of the annual-average dissolved oxygen (in ml/l) at 400 m depth, displayed in Fig. 1.1, is also taken from this atlas (Garcia et al., 2006).

B.6 Rossby Radius

As evident from the discussion in section 2.1, the Rossby radius of deformation is a key variable in geophysical fluid dynamics that characterizes the length scale at which the constraints imposed on fluid motions by stratification and rotation are of equal importance. Motions on larger scales store most of their energy in its potential form and are controlled by the vortex stretching component to potential vorticity, contrarily to motions on smaller scales which mainly contain energy in its kinetic form and are controlled by the relative vorticity contribution to potential vorticity.

The deformation radius is given by the distance long gravity waves travel in half an inertial period $T_i/2 = \pi/f$. This sets the length scale of linear geostrophic adjustment, which is different for motions with different vertical scales. For motions on large horizontal scales, the horizontal and vertical dependence of the flow can be separated into two sets of independent equations, where the solution to the vertical equation gives the vertical mode structure of the flow field. The gravest of this infinite set of orthogonal modes is the depth-independent barotropic mode, followed by a set of depth-dependent baroclinic modes, which feature an increasing number of sign changes in the vertical, one for the first baroclinic mode, two for the second, and so forth (e.g. Gill, 1982). The barotropic mode is associated with the fastest long gravity waves (for a typical ocean depth $H = 4 \text{ km}$, their horizontal phase-speed $c_0 = \sqrt{gH}$ is $\sim 200 \text{ m/s}$) and therefore the largest deformation radius $r_0 = c_0/f$ (2000 km in the above example at mid-latitudes). Baroclinic gravity waves, whose magnitude depends on the vertical density structure of the fluid, are much slower (typically 1-3 m/s for the first-mode) and associated deformation radii much smaller than their barotropic counterparts. They can be approximated as $r_n = \frac{1}{n} \int_{-H}^0 \frac{N}{\pi f}$, where $n \geq 1$ and $N = \sqrt{\frac{-g d\rho}{\rho_0 dz}}$ is the stratification frequency (Chelton et al., 1998).

In this study we use a dataset of the first baroclinic Rossby radius of deformation r_1 developed by Chelton et al. (1998) and provided globally on a 1° grid at www-po.coas.oregonstate.edu/research/po/research/rossby_radius. It is obtained as solution to the equation for the vertical mode structure applied to observed neutral density evaluated from the Levitus World Ocean Atlas 1994 hydrographic climatology.

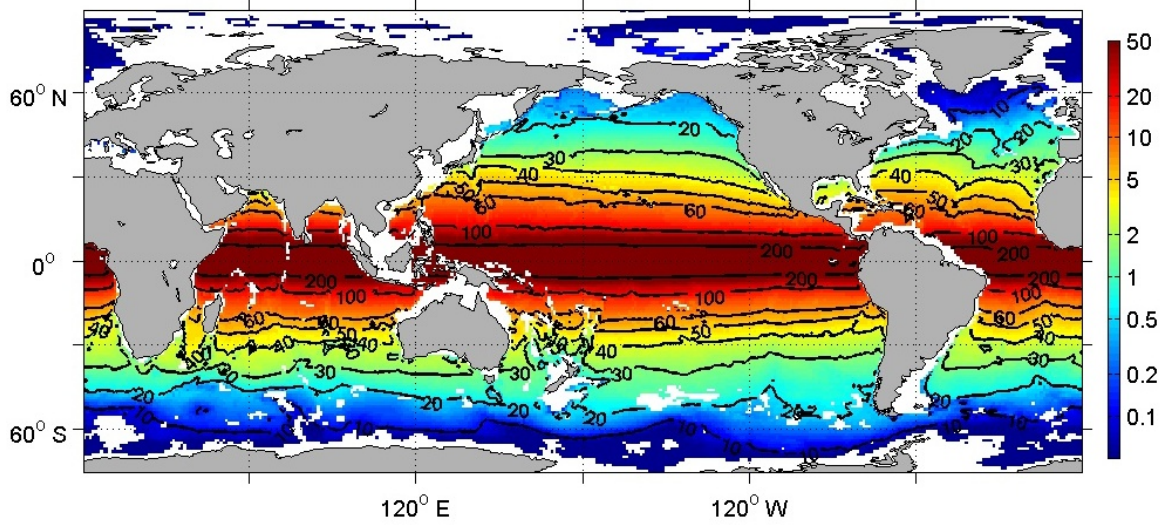


Fig. B.3: The global distribution of the deformation radius R_d (contours, in km), as provided by (Chelton et al., 1998), and the associated westward phase-speed of long first-mode baroclinic Rossby waves c_{trw} (colour-map, in cm/s, positive westward).

$r_1 = c_1/|f|$ will often be simply referred to as deformation or Rossby radius R_d . For vertically uniform stratification $R_d = \frac{NH}{\pi f}$, and the quantity $L_d = \frac{NH}{f} = \pi R_d$ will be referred to as deformation scale L_d .

The contours in Figure B.3 display the observed distribution of R_d for the global ocean. Its spatial structure is dominated by the latitudinal dependence of $1/|f|$ that leads to the observed steep decrease in R_d from over 200 km in the tropics to less than 10 km at high latitude. More subtle features of the R_d distribution include a strong control by bathymetric features particularly at high latitudes and an increase, at constant latitude, towards the western parts of subtropical gyres. These features are set by the spatial variations of c_1 that increases towards deeper waters and highly stratified regions, where baroclinic gravity waves are fastest.

The Rossby radii also set the dispersion relationship $\omega(k, l)$, between frequency ω and zonal and meridional wave-numbers k and l , of propagating linear wave solutions to the potential vorticity equation of a rotating-stratified fluid in the absence of forcing, dissipation and mean flow, referred to as planetary or Rossby waves. For a flat-bottomed ocean, their zonal phase speed is given by:

$$c_{xn} = \omega/k = \frac{-\beta}{k^2 + l^2 + (1/r_n)^2} \quad (\text{B.6.1})$$

Lines of constant phase of planetary waves thus always propagate westward, and for

first-mode long waves ($1/k, 1/l \gg R_d$) they do so at a rate $c_{lrw} = -\beta R_d^2$. c_{lrw} is coloured in Fig. B.3 and typically reaches a few cm/s in midlatitudes, but less than 1 cm/s in the ACC.

B.7 Bathymetry

A dataset of the observed topography of the sea floor, the oceanic bathymetry provided by GEBCO (General Bathymetric Chart of the Oceans group) at www.gebco.net, is used to evaluate the bathymetric constraints on the flow and for computation of the observed (no-flow) planetary potential vorticity distribution. Originally provided on a $1/60^\circ$ grid, Figure 3.3 displays its $1/4$ -grid linearly interpolated version, which is often used in this study.