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## Fast Warming of the Surface Ocean Under a Climatological Scenario

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1                   **Fast warming of the surface ocean under a**  
2                   **climatological scenario**

3                   **Q. Jamet<sup>1\*</sup>, W. K. Dewar<sup>1</sup>, N. Wienders<sup>1</sup> and B. Deremble<sup>2</sup>**

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6                   **Key Points:**

- 7                   • Weakly varying climatological winds reduce upper ocean vertical mixing, affect-
- 8                   ing the redistribution of air-sea fluxes
- 9                   • Coupled to an atmospheric boundary layer, the modeled ocean response to clima-
- 10                  tological winds is to warm up considerably at the surface
- 11                  • Results illustrate the pivotal improvements in air-sea interactions achieved by driv-
- 12                  ing an ocean model with an atmospheric boundary layer

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## Abstract

We examine various strategies for forcing ocean-only models, including an atmospheric boundary layer model. This surface forcing allows air-sea exchanges to affect atmospheric temperature and relative humidity, thus removing the assumption of an infinite atmospheric heat capacity associated with the prescription of these variables. When exposed to climatological winds, the simulated North Atlantic oceanic temperature warms considerably at the surface as compared to a model with full atmospheric variability. This warming is mainly explained by a weakened upper ocean vertical mixing in response to the weakly varying climatological winds. Specifying the atmospheric temperatures inhibits this warming, but depends on the unrealistic large atmospheric heat capacity. We thus interpret the simulated warmer ocean as a more physically consistent ocean response. We conclude the use of an atmospheric boundary layer model provides many benefits for ocean only modeling, although a 'normal' year strategy is required for maintaining high frequency winds.

## 1 Introduction

The ocean is one of the most important components of the climate system due to its large heat capacity. Understanding its dynamics remains challenging due to interactions with the other components of the climate system, particularly the atmosphere. It is common to decouple the climate system to reduce its complexity: A well known strategy is to use ocean-only numerical simulations for which air-sea interactions are control variables that one can adjust to study the ocean responses. However this approach strongly relies on parameterizations used to represent these interactions. We study here ocean-only models driven by different surface forcing strategies with the view toward assessing strengths and weaknesses of each.

Modeling a variable ocean under a specified but variable atmosphere is a useful and efficient idealization. However, some studies have highlighted an important caveat of this approach. Huck and Vallis (2001) have found that large scale modes of variability (Colin de Verdière and Huck, 1999; Huck et al., 1999) only appear if the ocean model was forced by prescribed fluxes rather than a prescribed atmosphere. In the first case (prescribed air-sea fluxes), the ocean is not as constrained as in the second case where the atmospheric conditions maintain the ocean in a state close to the forcing conditions. Additionally, Rahmstorf and Willebrand (1995) have shown that large scale ocean feedbacks on the

45 atmospheric temperature have significant impacts on the rate of the overturning circu-  
46 lation in the North Atlantic. Such feedbacks are not considered when the ocean surface  
47 is restored toward prescribed atmospheric conditions. These results illustrate the lim-  
48 itations associated with a prescribed atmospheric forcing, where the assumption of an  
49 infinite heat capacity for the atmosphere interferes with the development of internal ocean  
50 dynamics.

51 Following these ideas, we wish to assess if similar limitations would be at work for  
52 the development of an oceanic state of equilibrium under climatological winds. This ques-  
53 tion arises from the recognized contribution of the fast varying winds associated with  
54 synoptic weather systems to maintain realistic turbulent air-sea fluxes (Gulev, 1994; Hughes  
55 et al., 2012; Jung et al., 2014; Ponte and Rosen, 2004; Wu et al., 2016; Zhai et al., 2012)  
56 and upper ocean vertical mixing (Condrón and Renfrew, 2013; Holdsworth and Myers,  
57 2015; Wu et al., 2016). **This high frequency atmospheric variability is** indeed fil-  
58 tered out by the averaging used to construct climatological products, such that the high  
59 frequency wind variance strongly reduces at mid- and high latitudes (Fig. 1, top pan-  
60 els). Forcing an ocean model with climatological winds is therefore expected to signif-  
61 icantly modify air-sea fluxes and upper ocean vertical mixing. It is however less clear how  
62 the ocean will adjust to such a forcing if thermodynamic ocean feedbacks on the atmo-  
63 sphere are considered.

64 To identify the oceanic state that develops under climatological winds, we propose  
65 to work with an atmospheric boundary layer (CheapAML; Deremble et al., 2013). With  
66 this strategy, the assumption of an infinite atmospheric heat capacity is relaxed and at-  
67 mospheric temperature and humidity can respond to ocean surface structures. A detailed  
68 description of the model strategy is given in Section 2. The ocean response to climato-  
69 logical winds in this framework is described in Section 3. We compare these results with  
70 those obtained with a more traditional representation of air-sea fluxes, i.e. when the at-  
71 mosphere is prescribed, in Section 4, and assess the relevance of a 'normal' year to main-  
72 tain high frequency winds in Section 5. Finally, we conclude and discuss the results in  
73 Section 6.

## 2 Numerical Experiments

We use the MIT general circulation model (MITgcm; Marshall et al., 1997) in a regional configuration of the North Atlantic: the domain extends from 20°S to 55°N with a horizontal resolution of  $\frac{1}{4}^\circ$  (see Supporting Information). The mixed layer depth computed by the non-local K-Profile Parametrization (KPP) scheme of Large et al. (1994) is used in the heat budget of Section 3.2.

At the surface, different strategies are used to force the ocean model and look at their impact on the ocean dynamics. In a first set of experiments, we couple the ocean model to the atmospheric boundary layer model CheapAML (Deremble et al., 2013). With this approach, we better represent the air-sea exchanges, and we also let the ocean develop its internal dynamics (not necessarily correlated to a prescribed atmospheric state). In CheapAML, winds are assumed to be the least sensitive atmospheric variable to ocean surface structure, and are thus prescribed. The remaining atmospheric variables, i.e. temperature and humidity, are advected by these prescribed winds and are modified by the air-sea fluxes. Over the ocean, the temporal evolution of these atmospheric variables is computed using a forced advection-diffusion equation. Over land, temperature and humidity are strongly relaxed toward the reanalysis prescribed values. The atmospheric variables prescribed in CheapAML are applied every 6 hours and derived from the Drakkar forcing set (cf Supporting Information).

This configuration is run for 10 years in three different experiments. In the first one, referred to as AML\_FULL, we use the full range of atmospheric time scales, from sub-daily (6-hourly) to interannual, over the period 1958-1967. In the second one, referred to as AML\_CLIM, we use a yearly repeated climatological atmospheric seasonal cycle. To consistently filter the year-to-year atmospheric variability, the climatology has been computed as an ensemble average of all the years between 1958-1977. This second experiment is run for 10 years, i.e. for the same duration as the first experiment. In the third experiment, referred to as AML\_NY, we use a yearly repeated 'normal' year forcing (Large and Yeager, 2004), constructed as the realistic atmospheric forcing from August 2003 to July 2004. This experiment is run for 5 years only. Initial conditions are common to all experiments, and are derived from a  $\frac{1}{12}^\circ$  global ocean-only simulation (cf Supporting Information).

105 To understand how the atmospheric temperature and humidity in CheapAML re-  
 106 sponds to the ocean surface dynamics, two additional experiments are conducted where  
 107 all atmospheric variables (wind, **atmospheric surface air** temperature and humidity)  
 108 are prescribed. This strategy is commonly used in the ocean modeling community and  
 109 it will serve as a reference test case to which we will compare our experiments. With these  
 110 prescribed atmospheric variables, we compute the air-sea fluxes the same way as the pre-  
 111 vious cases but there is no feedback on the atmospheric temperature and humidity. As  
 112 in AML\_FULL and AML\_CLIM we run two experiments with either fully varying or cli-  
 113 matological winds. We name these two experiments FORC\_FULL and FORC\_CLIM, re-  
 114 spectively.

### 115 **3 Fully Varying vs Climatological Wind Experiments**

116 Because the Sea Surface Temperature (SST) is an oceanic variable sensitive to air-  
 117 sea exchange, we first compare the model SST for the two experiments AML\_CLIM and  
 118 AML\_FULL (Fig. 2, top panels) after 10 years of simulation. The yearly averaged SST  
 119 differences between the two experiments are very large in amplitude, reaching more than  
 120  $8^{\circ}\text{C}$  in the subtropical gyre, and spreading over the North Atlantic, north of  $20^{\circ}\text{N}$ . At  
 121 the center of the subtropical gyre where the largest SST differences are observed, the time  
 122 evolution of SST over the course of the 10 years of simulation reveals that such large dif-  
 123 ferences are reached quickly, after 5 months, suggesting a fast dynamic response of the  
 124 ocean. The mechanism that drives the warming of the subtropical gyre in the AML\_CLIM  
 125 experiment is described in the two following sections. We focus on the first two years of  
 126 simulation where most of the differences build up.

#### 127 **3.1 Heat Fluxes**

128 In our configuration, the components of the net heat fluxes which vary from ex-  
 129 periment to experiment are the latent and sensible fluxes, as well as the outgoing long-  
 130 wave radiation associated with the SST (the other components are prescribed). We dis-  
 131 cuss their respective contribution for the net heat fluxes in the following.

132 In AML\_FULL, the time mean and standard deviation of the latent and sensible  
 133 heat fluxes (computed over the 10 years of simulation) are  $117\pm 37$  and  $8\pm 8$   $\text{W m}^{-2}$ , re-  
 134 spectively. Added together, these fluxes are sufficiently strong to induce positive (up-

ward) net heat fluxes during the first two months of simulation (Fig. 3, top panel). They contribute to the cooling of the ocean surface at the beginning of the simulation which is consistent with winter time (January-February). In AML\_CLIM, the turbulent fluxes are reduced by more than 50% ( $57 \pm 19 \text{ W m}^{-2}$ ;  $1 \pm 1.4 \text{ W m}^{-2}$  for latent and sensible heat fluxes, respectively), consistent with earlier results (Hughes et al., 2012). As a consequence, the too weak turbulent heat fluxes lead to negative (downward) net heat fluxes, contributing to the warming of the surface ocean at the beginning of the simulation (top right panel of Fig. 2).

The mechanisms that drive the reduction in turbulent air-sea fluxes are further investigated by looking at the sensible heat flux amplitude  $S = \rho_A C_p^{(A)} |\mathbf{u}| (SST - T_a - \gamma h)$ , with  $\rho_A = 1.3 \text{ kg m}^{-3}$ ,  $C_p^{(A)} = 1004 \text{ J kg}^{-1} \text{ K}^{-1}$ ,  $\gamma = 0.0098 \text{ K m}^{-1}$  the dry atmospheric adiabatic lapse rate and  $h = 10 \text{ m}$  the height at which turbulent air-sea fluxes are computed. Fig. 3 (right panel) show the sensible heat as a function of the two main contributing factors, i.e. the **scalar** wind speed  $|\mathbf{u}|$  and the air-sea temperature difference ( $SST - T_a$ ). As a response to a weaker wind variance in AML\_CLIM, there are no winds stronger than  $5 \text{ m s}^{-1}$  (top right panel). However, for wind speed weaker than  $5 \text{ m s}^{-1}$ , the sensible heat fluxes in AML\_CLIM remain weaker than those obtained under fully varying winds, suggesting that the changes in air-sea fluxes are not only driven by the weaker climatological wind speed  $|\mathbf{u}|$ . The other parameter that contributes to the strength of the sensible heat fluxes is the air-sea temperature difference  $SST - T_a$ . In AML\_CLIM, the air-sea temperature differences do not exceed  $\pm 1^\circ\text{C}$  (Fig. 3, bottom right panel), while they range from about  $-2^\circ\text{C}$  to about  $+4^\circ\text{C}$  in AML\_FULL. Under fully varying winds, there are thus oceanic processes that take the ocean surface away from the overlying atmosphere and lead to larger air-sea temperature differences. We show in Section 3.2 that those processes are associated with upper ocean vertical mixing.

As the ocean surface quickly warms up at the beginning of the simulation when exposed to climatological winds, the outgoing longwave radiation increases accordingly (outgoing longwave radiation is proportional to  $SST^4$ ). The system reaches a new state after 5 months with a new SST about  $8^\circ\text{C}$  warmer than for the fully varying wind experiment. The upward longwave radiation is  $40\text{-}50 \text{ W m}^{-2}$  stronger in the climatological wind experiment, which balances about 80% of the  $-60 \text{ W m}^{-2}$  time mean difference in turbulent heat fluxes induced by the weakly varying climatological winds. The net air-sea heat fluxes are of comparable amplitude in the two experiments, preventing the SST

168 difference to be much larger than 8°C. After the 5 months of initial transition, the model  
 169 slowly drifts toward its new state of equilibrium with an SST trend of about +0.25°C/yr.  
 170 Note that since atmospheric downward longwave radiations are prescribed, the radia-  
 171 tive effects of a warmer atmosphere are not considered. Such an effect would positively  
 172 contribute to the upper ocean warming since increased downward longwave atmospheric  
 173 radiations would induce larger warming of the ocean surface (Fig. S3).

### 174 3.2 Oceanic Vertical Mixing

175 We now describe the differences between the two equilibria in terms of oceanic dy-  
 176 namics. We performed a heat budget following Peter et al. (2006) for the box at the cen-  
 177 ter of the subtropical gyre where the SST difference is the largest. The temperature ten-  
 178 dency  $\partial_t \langle T \rangle$  within the mixed layer  $h(x, y, t)$  (computed by the KPP parameterization)  
 179 is decomposed into advective terms, a flux term and dissipation terms (cf Supporting  
 180 Information). Comparing the results of this heat budget for the two experiments using  
 181 CheapAML (Fig. S5), the most important difference in the processes controlling the tem-  
 182 perature is found to be the upper ocean vertical mixing.

183 While the upper ocean vertical mixing is relatively constant through the year in  
 184 AML\_CLIM with a mixed layer depth which does not exceed 15 m, it exceeds 40 m in  
 185 AML\_FULL during winter in response to atmospheric storms. Thus, even though the  
 186 net heat fluxes in AML\_CLIM are weaker (cf Section 3.1), their contribution for the heat  
 187 budget of the mixed layer quickly exceed 0.5 °C/day. In AML\_FULL by contrast, sur-  
 188 face heat fluxes warm a thicker mixed layer, and thus lead to a weaker temperature ten-  
 189 dency during the first 5 months of simulation (Fig. S5, top right panel). Also note that,  
 190 in AML\_FULL, the vertical diffusion at the bottom of the mixed layer  $\frac{1}{h} K_z \partial_z T|_{z=-h}$  con-  
 191 trols most of the balance between surface heating and internal ocean processes, i.e. it  
 192 explains a significant fraction of the residual  $\partial_t \langle T \rangle + \frac{Q_{net}}{\rho_0 C_p h}$  (Fig. S5, bottom panels).  
 193 The other terms (lateral diffusion, advective and entrainment terms; cf Eq. (4) in Sup-  
 194 porting Information) are at least one order of magnitude smaller. In AML\_CLIM by con-  
 195 trast, the vertical diffusion at the bottom of the mixed layer is weaker and explains a smaller  
 196 fraction of that residual. This suggests that the oceanic processes within the mixed layer  
 197 have changed. We suspect night time convection comes into play, but we cannot draw  
 198 firm conclusions with the 5-day averaged outputs used in this study.

199 The large reduction in mixed layer depth in AML\_CLIM is observed all over the  
 200 domain, where the maximum depth of the mixed layer computed by the KPP param-  
 201 eterization is about 3 to 4 times shallower North of 20°N (Fig. 1, bottom panels). This  
 202 spatial pattern resembles the wind variance (top panels). In fact, in AML\_FULL, the high  
 203 frequency wind variance induces a vertical velocity shear  $\partial_z \mathbf{u}$  in the upper layers, that  
 204 destabilizes the ocean: **the Richardson number  $R_i = \frac{N^2}{\partial_z \mathbf{u}}$  (with  $N^2$  the buoyancy  
 205 frequency) decreases, and ultimately falls below the critical value  $R_i^{crit.} = 0.3$   
 206 used in our configuration. The vertical structure of the ocean is thus unsta-  
 207 ble to Kelvin-Helmholz shear instability, the leading process driving ocean  
 208 vertical mixing for such low  $R_i$  (Mack and Schoeberlein, 2004). In AML\_CLIM  
 209 by contrast, the vertical velocity shear is much weaker in response to the weaker high  
 210 frequency variance of the climatological winds, and the ocean is more stable. If less mix-  
 211 ing occurs in the upper ocean, the surface heat fluxes induce a fast warming of the up-  
 212 per ocean.**

#### 213 4 A Prescribed Atmosphere

214 Most numerical studies that use climatological atmospheric fields do not use an at-  
 215 mosphere boundary layer model to compute the atmospheric temperature and humid-  
 216 ity (Grégorio et al., 2015; Penduff et al., 2011; Sérazin et al., 2015). In order to compare  
 217 our results with these kind of experiments, we perform two additional runs (FORC\_FULL  
 218 and FORC\_CLIM) for which all atmospheric fields (including temperature and humid-  
 219 ity) are prescribed. After 10 years, the simulated SST difference between FORC\_FULL  
 220 and FORC\_CLIM share a relatively similar spatial pattern with the AML experiments  
 221 (Fig. 2, left panels), but those differences are much weaker, and do not exceed 2.5°C in  
 222 the subtropical gyre. Note that, consistent with the temperature difference observed be-  
 223 tween the two AML experiments, the SST difference observed in the subtropical gyre  
 224 is also reached after only 5 months of simulation (Fig. 2, right panels).

225 From these comparisons, we conclude that prescribing the atmospheric state re-  
 226 duces the effects of climatological winds on the temperature of the upper ocean layers.  
 227 The underlying physical basis remains however questionable. Due to the weak high fre-  
 228 quency variance of the climatological winds, the vertical ocean mixing remains weak. The  
 229 difference in the mixed layer depth computed by the KPP scheme is very similar to what  
 230 is shown in Fig. 1 for the AML experiments. As a consequence, the upper ocean tends

231 to warm up in FORC\_CLIM, but the atmosphere does not. In fact, because the atmo-  
232 spheric temperature is prescribed in this experiment, the ocean-atmosphere temperature  
233 differences increase, as shown in Fig. 3 (bottom right panel) for the subtropical gyre. In  
234 FORC\_CLIM, the  $SST - T_a$  difference is always positive and roughly 2-4°C. This il-  
235 lustrates the damping role of the atmosphere on the surface ocean temperatures, con-  
236 straining the upper ocean warming tendency. This increased  $SST - T_a$  difference coun-  
237 teracts the effect of climatological winds on the turbulent air-sea fluxes, such that for  
238 the same wind speed amplitude, the sensible heat fluxes are much larger in FORC\_CLIM  
239 than in AML\_CLIM and always positive (Fig. 3, top right panel). A similar scenario hap-  
240 pens for the latent heat fluxes, which results in turbulent heat fluxes in FORC\_CLIM  
241 which are of comparable amplitude than those found in FORC\_FULL (Fig. 3, bottom  
242 left panel). This is not consistent with previous studies (Gulev, 1994, 1997; Hughes et  
243 al., 2012), where the lack of high frequency wind variance is expected to significantly re-  
244 duce the magnitude of turbulent air-sea fluxes. In the AML experiments by contrast, since  
245 the atmospheric temperature follows the surface ocean warming we have shown that the  
246 reduced turbulent heat fluxes under climatological winds are consistently captured and  
247 balanced by increased outgoing longwave radiations. Since this latter scenario has bet-  
248 ter physical consistency, we argue that the dynamically consistent ocean response to an  
249 artificial climatological atmosphere is to warm considerably at the surface.

## 250 **5 A view toward ocean climate studies**

251 To study the oceanic variability and to distinguish between the atmospherically forced  
252 and the internally generated low-frequency oceanic variability, it is common to drive ocean  
253 models with climatological winds. When using an atmospheric boundary layer, our re-  
254 sults reveal that fast varying atmospheric winds have to be accounted for in order to main-  
255 tain a realistic oceanic state. Results of the AML\_NY experiment driven by a 'normal'  
256 year forcing (Large and Yeager, 2004) are promising in this respect. Both air-sea tur-  
257 bulent fluxes and the depth of the mixed layer compare well to those diagnosed in AML\_FULL  
258 (not shown). This results in smaller differences in SST between the two experiments (Fig. S2  
259 and S3) compared to those obtained with climatological winds. The oceanic state that  
260 develops under a 'normal' year forcing is thus more consistent with the oceanic state that  
261 develops under a fully varying forcing. This is likely to make the comparison between

262 the two experiments more relevant for the attribution of the origin of low-frequency oceanic  
263 variability.

264 For long time scales, where the ocean is expected to control air-sea exchanges (Bjerk-  
265 nes, 1964; Gulev et al., 2013; Jungclaus and Koenigk, 2010; Latif et al., 2006; Shaffrey  
266 and Sutton, 2006), there are potential benefits in using an atmospheric boundary layer  
267 model to drive ocean-only models. The simulations presented here are however too short  
268 to make any statement about low-frequency variability with good accuracy. To illustrate  
269 this idea, we have performed an exploratory 50 years, high resolution ( $1/12^\circ$ ) run in a  
270 regional configuration (same ocean model formulation coupled to CheapAML and driven  
271 by the same 'normal' year forcing). In this high resolution experiment, atmospheric tem-  
272 peratures computed by the atmospheric boundary layer model exhibit variability at time  
273 scales longer than one year (Fig. S4, left panel). This low-frequency atmospheric vari-  
274 ability is particularly strong downstream of Cape Hatteras and along the North Atlantic  
275 current (Fig. S4, right panel). This is indicative of a low-frequency atmospheric variabil-  
276 ity driven by ocean processes which is obviously missing when atmospheric temperatures  
277 are prescribed with a seasonally repeating cycle. **Such a low-frequency atmospheric**  
278 **variability feeds back onto the ocean, enhancing SST variability at low fre-**  
279 **quency. This is illustrated on Fig. S4 (right panel), where yearly averaged**  
280 **SST standard deviation is compared between this high resolution simulation**  
281 **and a companion simulation driven by prescribed 'normal' year atmospheric**  
282 **forcing. Although this feedback weakens SST variability in localized regions,**  
283 **the low-frequency SST variability is larger almost all over the domain with**  
284 **a pattern that largely replicates the pattern indicated by atmospheric tem-**  
285 **perature variations. Along the North Atlantic current, this increase is of about**  
286 **+0.3-0.4°C and contributes to 20-30% of the  $\sim 1.5^\circ\text{C}$  total SST low-frequency**  
287 **variability of this region.** These preliminary results are very encouraging and high-  
288 light the benefits of using an atmospheric boundary layer model to drive ocean-only mod-  
289 els in the context of climate studies. It could for instance bring new insights to the on-  
290 going debate on the origin of North Atlantic climate variability (Clement et al., 2015;  
291 Farneti, 2017; Zhang et al., 2016).

## 6 Conclusion

We have revisited the surface forcing strategy used to drive ocean-only models. The analysis of an ocean model in a regional North Atlantic configuration coupled to an atmospheric boundary layer model shows that the oceanic state that develops under climatological winds is much warmer at the surface than the one that develops in a companion experiment driven by fully varying winds. It is up to 8°C warmer in the North Atlantic subtropical gyre after only 5 months of simulation. Although significantly different than realistic conditions, we argue that those changes are physically consistent, and interpret this new oceanic state as likely when exposed to an artificial climatological atmosphere. This surface warming is the result of a side effect of climatological averaging on winds where fast varying synoptic weather systems are filtered out, strongly reducing upper ocean vertical mixing and turbulent air-sea fluxes. These results illustrate the key role played by thermodynamical ocean feedbacks on the atmosphere for the development of an oceanic state of equilibrium.

In the climatological scenario, the system reaches a new balance for which the warmer ocean surface induced by weak ocean vertical mixing is balanced by increased outgoing longwave radiation. This balance is quite different from the equilibrium reached in the traditional approach (where the atmospheric state is prescribed). In the latter case, the ocean vertical mixing remains weak, but the effects of the climatological winds on the turbulent air-sea fluxes are balanced by an increased contribution of the difference between the warming ocean and the prescribed atmosphere. The turbulent air-sea fluxes are strengthened and the atmosphere controls the surface ocean dynamics by damping the surface warming tendency. However, this 'traditional' approach relies on the unrealistic assumption of an infinite heat capacity for the atmosphere, whereas the ocean is more appropriately approximated as the slow climate component since its heat capacity is much larger than that of the atmosphere. Those results suggest that the use of an atmospheric boundary layer model rather than a prescribed atmosphere when decoupling an ocean model from the atmosphere is a more suitable strategy to better represent the physics of the air-sea interactions.

To isolate the oceanic dynamics from the low-frequency atmospheric forcing when an ocean model is coupled to an atmospheric boundary layer model, one thus needs a wind product that does not contain any interannual and longer variability but which ac-

324 counts for the fast varying winds. Last, a preliminary experiment driven by a 'normal'  
 325 year strategy (Large and Yeager, 2004) exhibits promising results: The oceanic state that  
 326 develops under such a forcing stays in better agreement with the fully varying experi-  
 327 ment, and atmospheric temperatures exhibit low-frequency variability **which feeds back**  
 328 **onto ocean surface dynamics**. We conclude that, combined with a normal year forc-  
 329 ing strategy, the use of an atmospheric boundary layer to drive ocean-only models for  
 330 climate oriented studies is very attractive.

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 336 [\\_GRL2019/](http://ocean.fsu.edu/~qjemet/share/data/clim_wind_GRL2019/).

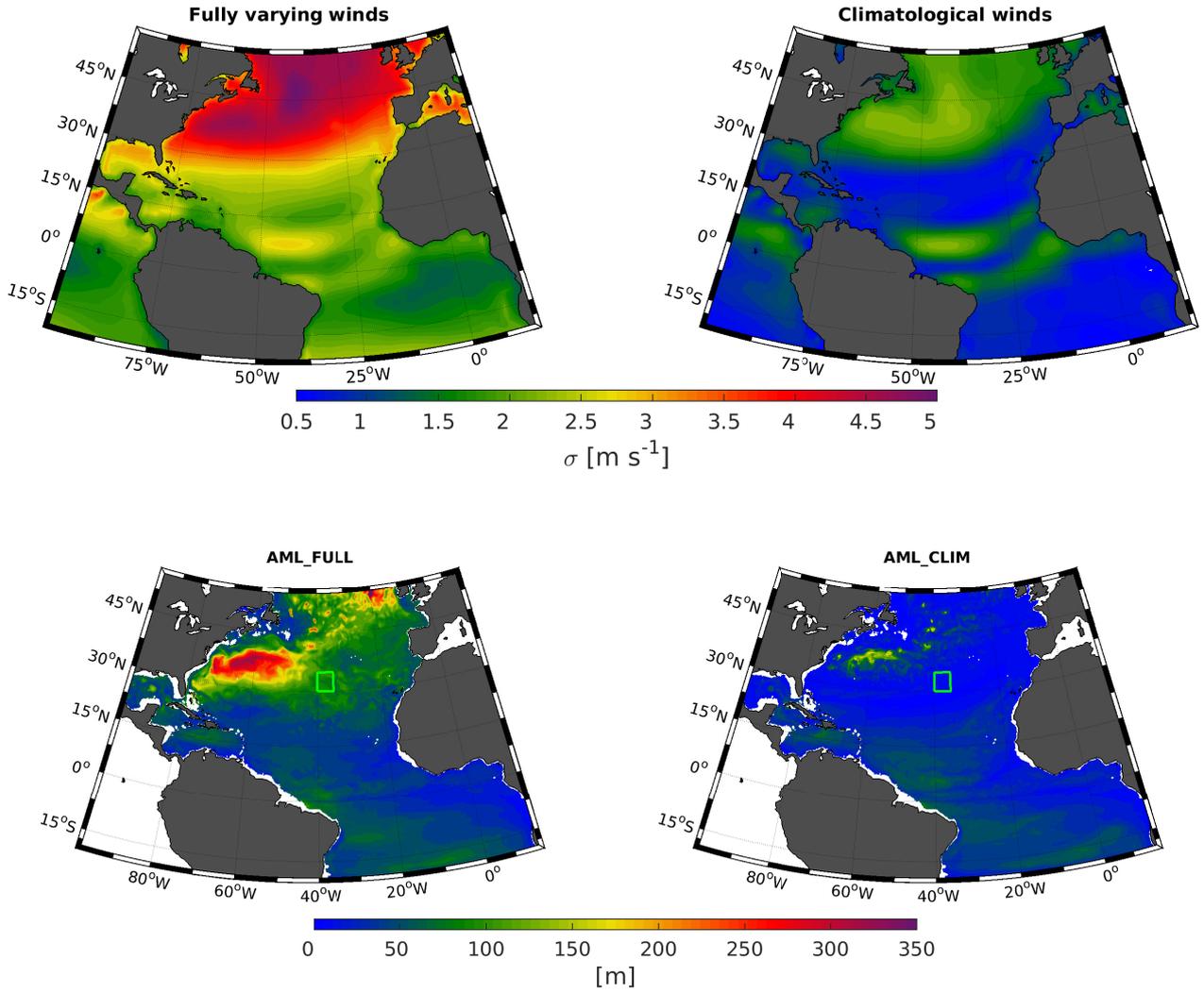
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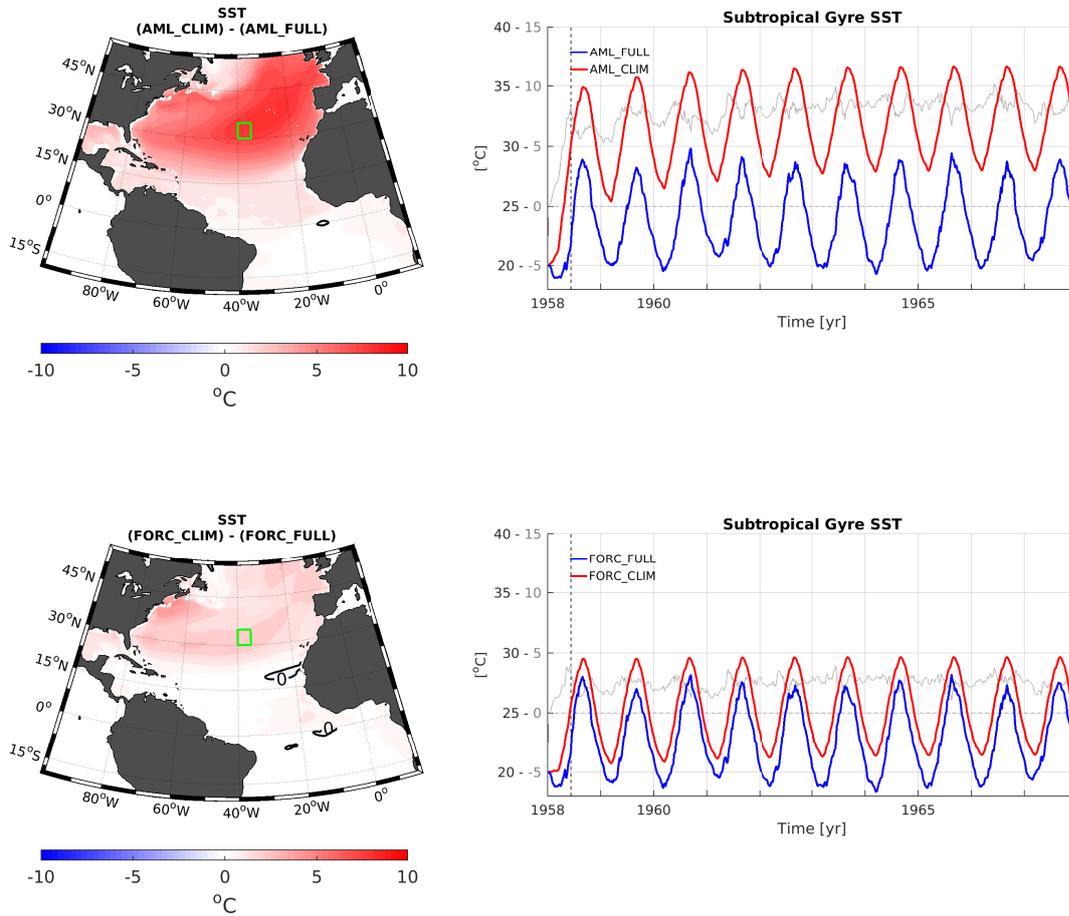
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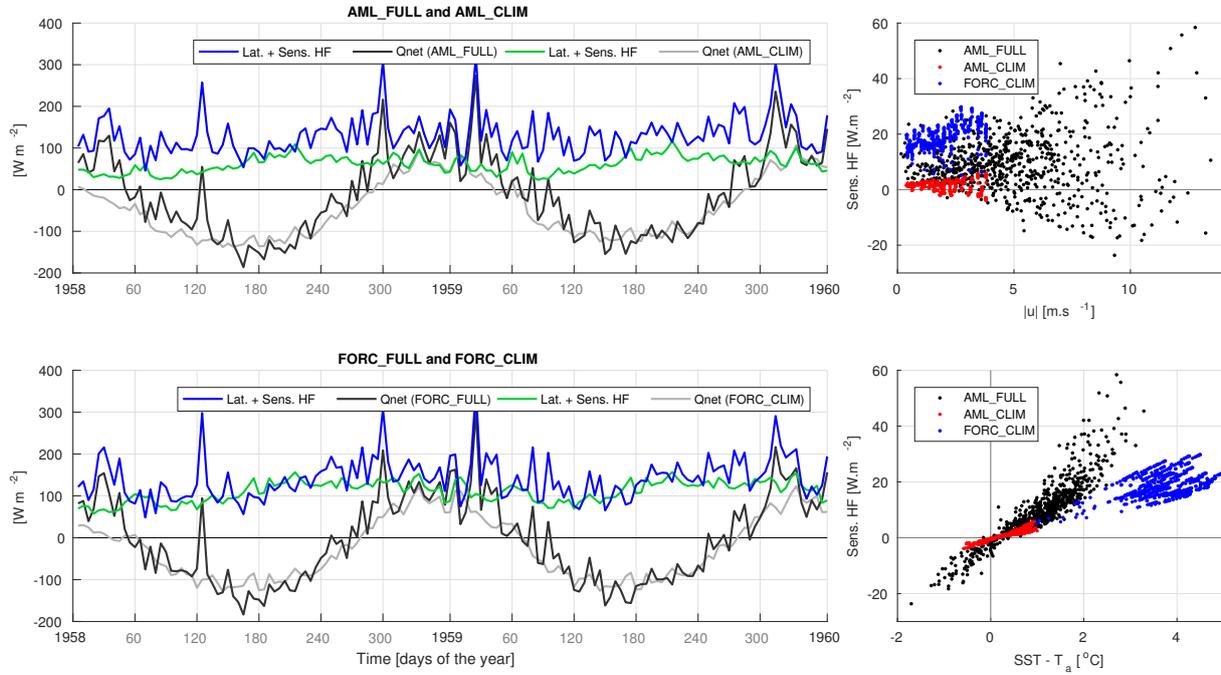
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**Figure 1.** (Top) Seasonal standard deviation  $\sigma$  [ $\text{m s}^{-1}$ ] of the wind speed  $\|u\| = \sqrt{u^2 + v^2}$  for the fully varying (left) and the climatological (right) winds. (Bottom) Maximum depth of the mixed layer [m] computed by the KPP parameterization during the first year of simulation for the AML\_FULL (left) and the AML\_CLIM (right) experiments.



**Figure 2.** (Top) Yearly averaged SST difference between the AML\_CLIM and the AML\_FULL experiment for the last year of simulation, i.e. 1967 (left), and (right) associated time series of the spatially averaged SST in the subtropical gyre ( $[40-35W;30-35N]$ , green box on the left panel). The gray line is the difference between the two experiments, with the associated scale on the left in gray. (Bottom) Same as top panels, but for the FORC experiments.



**Figure 3.** (Top left) Spatially averaged net heat fluxes  $Q_{net}$  (positive upward, [ $W m^{-2}$ ]) at the center of the subtropical gyre (green box of Fig. 2) for the AML\_FULL (dark gray line) and the AML\_CLIM (light gray line) experiments, and the associated contribution of the latent plus sensible heat fluxes (blue and green lines, respectively). (Bottom left) Same as top left panel but for the FORC experiments. (Right) Scatter plots of the sensible heat fluxes as a function of the wind speed (top) and the air-sea temperature difference  $SST - T_a$  (bottom) for AML\_FULL (black), AML\_CLIM (red) and FORC\_CLIM (blue). Data correspond to the 10 years long time series at a point within the subtropical gyre (green box of Fig. 2) and are not sensitive to which point is chosen.