

Ocean gyres driven by surface buoyancy forcing

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Key Points:

- Simulations show that large-scale midlatitude ocean gyres can be driven by buoyancy forcing alone.
- Scaling balance to estimate the relative contributions of wind and buoyancy forcing to gyre circulation is evaluated.
- Estimates of gyre transport should take surface buoyancy forcing into account.

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Abstract

Midlatitude gyres in the ocean are large scale horizontal circulations that are intensified on the western boundary of the ocean, giving rise to currents such as the Gulf Stream. The physical mechanism underlying gyres is widely recognised to involve the curl of the wind stress, which injects potential vorticity into the upper ocean. However, model results have highlighted the role of surface buoyancy fluxes (principally heating and cooling of the ocean surface) in driving circulation and enhancing gyre variability. Here we present two numerical simulations – one in the fully turbulent regime and the second an eddy-permitting ocean model – which show that gyre-like circulation can be driven by surface buoyancy fluxes alone. We explore this phenomenon through a combination of modelling and linear theory to highlight that the transport of ocean gyres depends upon surface buoyancy fluxes as well as wind stress.

Plain Language Summary

Ocean gyres are persistent, large-scale circulation features that give rise to important ocean currents such as the Gulf Stream in the North Atlantic and the Kuroshio current off the east coast of Japan. These gyres are critical in transporting heat from the tropics to the poles. Standard oceanographic theory suggests that these gyres are driven by wind stress, however, the simple theory that predicts the strength of these gyres fails in many parts of the ocean. In this manuscript we demonstrate that ocean gyres (complete with a rich eddy field and strong western boundary current) occur even in the absence of wind forcing. Thus, we contend that a significant component of gyre circulation, particularly in the subpolar regions, is due to temperature-driven buoyancy fluxes. This result represents a profound change to our understanding of one of the most fundamental aspects of the oceans large-scale circulation.

1 Introduction

Ocean circulation is primarily driven by a combination of mechanical forcing due to wind stress and buoyancy forcing (heating/cooling and freshening/salinification) at the ocean surface (Ferrari & Wunsch, 2009; Hughes, Hogg, & Griffiths, 2009; Stewart, Ferrari, & Thompson, 2014; Tailleux, 2009). The relative influence of buoyancy and mechanical forcing on individual current systems is a major challenge of oceanographic research that aims to predict the response of the ocean circulation to climate change. Traditional oceanographic theory predicts that horizontal, near-surface currents are driven by wind (W. H. Munk, 1950), with the deep overturning circulation (involving vertical transport) driven by differences in buoyancy (Schmitz & McCartney, 1993). This viewpoint has been revised in recent decades. Firstly, it has been recognised that an equilibrated ocean overturning circulation must balance the generation of buoyancy differences at the surface with turbulent mixing throughout the ocean (W. Munk & Wunsch, 1998). This oceanic turbulence is powered by mechanical forces, including wind stress and the bottom stress induced by tidal flow (Ferrari & Wunsch, 2009; Wunsch & Ferrari, 2004). Secondly, the Antarctic Circumpolar Current (ACC) in the Southern Ocean is a surface-intensified current traditionally considered to be driven by wind stress (Johnson & Bryden, 1989; W. H. Munk & Palm en, 1951). However, recent research suggests a strong role for buoyancy forcing in driving the mean ACC (Gent, Large, & Bryan, 2001; Hogg, 2010), with wind stress primarily controlling the strength of the mesoscale eddy field (Hogg et al., 2015). Thus, the traditional delineation between near-surface wind-driven flow and deep buoyancy-driven circulation no longer appears to hold. These advances motivate us to re-examine the cornerstone oceanographic theory of wind-driven gyres.

The oceanic wind-driven circulation is typically estimated by assuming a linear balance between the input of potential vorticity at the surface and meridional transport across

62 gradients in planetary vorticity ($\beta = \frac{\partial f}{\partial y}$, where f is the Coriolis parameter and y the
63 meridional coordinate). This argument can be derived from Sverdrup balance for the geostrophic
64 velocity, v_g :

$$65 \quad \beta v_g = f \frac{\partial w}{\partial z}. \quad (1)$$

66 If transport is integrated over the ocean depth H , then to first order, it can be shown
67 (Wunsch, 2011) that the meridional transport is given by

$$68 \quad \beta v H = (\nabla \times \tau) \cdot \hat{\mathbf{z}} \quad (2)$$

69 where v is the vertically averaged meridional velocity, $\hat{\mathbf{z}}$ is the vertical unit vector and
70 τ the wind stress vector (where stress is normalised by reference density). This is the
71 Sverdrup transport relation (Sverdrup, 1947) commonly used to estimate barotropic merid-
72 ional transport in the interior of ocean gyres, closed by an inertial return flow along the
73 western boundary. The meridional variation in the wind stress thus governs the direc-
74 tion of meridional flow in these large-scale gyres: combined with the westward intensi-
75 fication of the return flow (Stommel, 1948) the subtropical gyre (equatorward of the max-
76 imum wind stress) spins in an anticyclonic sense (clockwise in the northern hemisphere),
77 while the subpolar gyre (poleward of the maximum wind stress) rotates in a cyclonic di-
78 rection. At the boundary between these gyres, which is predicted to occur at the loca-
79 tion where wind stress curl is zero, western boundary currents separate from the coast
80 to close the gyres.

81 The theory of wind-driven gyres has developed considerably over the succeeding
82 decades, but the dependence of the circulation on wind stress curl has remained a key
83 element of any theory. The ventilated thermocline theory (Luyten, Pedlosky, & Stom-
84 mel, 1982) uses a layered framework to demonstrate that the vertical structure of the
85 solution depends upon stratification. However, under this model, only isopycnal layers
86 that outcrop at the surface have a circulation; integration over the active layers recov-
87 ers the Sverdrup relation. Furthermore, it should be stressed that the linear balance in
88 Eq. (2) is complicated by nonlinear processes such as eddies; this can result in home-
89 genisation of potential vorticity (Rhines & Young, 1982) and the creation of strong in-
90 ertial recirculation gyres (Cessi, 1988; Cessi, Ierley, & Young, 1987) overprinting the lin-
91 ear circulation.

92 Nonlinearity and stratification complicate the evaluation of Sverdrup dynamics in
93 the ocean. Combining Argo float trajectories and density profiles, Gray and Riser (2014)
94 argue that Sverdrup balance (Eq. 1) is in agreement with observations in the subtrop-
95 ical gyres, consistent with Wunsch (2011). Quantitatively, the zonally averaged Sverdrup
96 balance in ocean reanalysis products and models is accurate to within 20% (Thomas, de
97 Boer, Johnson, & Stevens, 2014). Sverdrup balance is less accurate in the subpolar gyres
98 (Gray & Riser, 2014) implying that bottom topography steers the current (Bower et al.,
99 2002; Yeager, 2015). Colin de Verdière and Ollitrault (2016) further apply boundary con-
100 straints on the known ocean density field and Argo float trajectories to infer that the Sver-
101 drup transport relation (Eq. 2) underestimates the barotropic circulation by approxi-
102 mately a factor of two in subtropical gyres.

103 An alternative hypothesis for the generation of large scale horizontal circulation
104 in ocean basins is that it may be driven by surface buoyancy fluxes. This notion was first
105 proposed by Goldsborough (1933), who predicted a weak circulation to arise through the
106 sea level variations owing to freshwater fluxes. Comparisons of wind-driven with buoy-
107 ancy driven flow were also investigated by Luyten and Stommel (1986), but their sim-
108 ple model parameterised the buoyancy contribution as a diabatic interior flux, rather than
109 buoyancy flux through the surface of the ocean. Colin de Verdière (1988) outlined the-
110 oretical and numerical predictions to demonstrate large-scale gyre circulations could be
111 driven by buoyancy forcing alone. In particular, surface buoyancy forcing may contribute
112 to potential vorticity entry or exit (Czaja & Hausmann, 2009), hence creating horizon-
113 tal circulation on par with wind-driven gyres (Colin de Verdière, 1989). More recently,

114 coarse-resolution global ocean models run in the absence of wind stress forcing have helped
 115 to reinitiate the discussion about the relative roles of wind stress and buoyancy fluxes
 116 in driving large scale circulation (Gjermundsen, LaCasce, & Denstad, 2018). However,
 117 these previous studies produced a single large gyre and were unable to recreate the dou-
 118 ble gyre structure observed in the ocean.

119 In this paper we investigate the role of surface buoyancy forcing as an additional
 120 forcing of gyres using turbulence resolving and eddy-permitting numerical simulations.
 121 In our simulations, we define the gyres to be horizontal, recirculating, basin-scale mean
 122 flows, that are balanced by inertial western boundary currents. We compare both barotropic
 123 and baroclinic measures of the gyre strength with predictions from linear theory, to eval-
 124 uate whether gyres forced solely by surface buoyancy fluxes may be significant contrib-
 125 utors to the large-scale ocean circulation.

126 2 Scaling

127 The buoyancy budget of the upper ocean can be approximated by integrating the
 128 advection-diffusion equation for density over a layer of constant depth h , assuming a lin-
 129 ear equation of state. We further assume that advective contributions are dominated by
 130 the depth-averaged velocity (u, v) with negligible vertical velocity, while diffusive con-
 131 tributions are dominated by vertical fluxes to give

$$132 \quad \frac{\partial b}{\partial t} + u \frac{\partial b}{\partial x} + v \frac{\partial b}{\partial y} = \frac{F_S}{h} - \frac{F_h}{h}, \quad (3)$$

133 where b is the depth-averaged buoyancy (related to the depth-averaged ocean density,
 134 ρ , by $b = g(\rho_0 - \rho)/\rho_0$; g being the acceleration due to gravity). Thus, the vertical buoy-
 135 ancy flux at the surface (F_S) and at depth h (F_h) balances the lateral advection of buoy-
 136 ancy. A Reynolds decomposition of the flow, in which we divide the flow in to mean ($\bar{b}(y)$)
 137 and transient (b') components, gives

$$138 \quad \bar{v} \frac{\partial \bar{b}}{\partial y} + \overline{\mathbf{u}' \cdot \nabla b'} \approx \frac{\overline{F_S}}{h}, \quad (4)$$

139 if F_h is assumed negligible. By analogy with the Sverdrup transport relation, in which
 140 the potential vorticity budget is linearised about a background state, Eq. (3) is linearised
 141 to eliminate the eddy buoyancy fluxes ($\overline{\mathbf{u}' \cdot \nabla b'}$), leaving a zeroth order balance between
 142 meridional advection of buoyancy and the surface forcing,

$$143 \quad \bar{v} \sim \frac{F_S}{h \frac{\partial \bar{b}}{\partial y}}. \quad (5)$$

144 Equation (5) describes a type of Walin balance (Walin, 1982) in which meridional flow
 145 in the ocean scales with surface buoyancy flux. It cannot be used to predict the circula-
 146 tion, but can be used in diagnostic mode. As with Sverdrup balance, it assumes that
 147 nonlinearity and eddy fluxes are secondary effects, and suggests equatorward flow in the
 148 subtropical ocean (assuming that F_S is generally positive there) and poleward flow in
 149 subpolar regions. This relationship may break down when vertical fluxes are significant,
 150 and is difficult to apply to real-world scenarios where neither \bar{b} nor F_S is independent
 151 of longitude. Importantly, Eq. (5) does not replace Sverdrup balance; instead, vortic-
 152 ity and buoyancy must both be balanced, implying that gyre strength should depend upon
 153 buoyancy fluxes, as well as wind stress curl.

154 3 Models

155 Two different methods are used to understand the dynamics of buoyancy-forced
 156 gyres. The first method is a Direct Numerical Simulation in which all scales of turbu-
 157 lence in a fluid domain are resolved but the scales and aspect ratio differ from the ocean.

The second method is to use an eddy-permitting ocean model that has parameterised mixing and convection but can resolve the ocean aspect ratio and additionally incorporate the relative effects of wind stress. These two methods are used to highlight, firstly, the fundamental nature of buoyancy-driven circulation in a rotating fluid and, secondly, to compare the magnitude of buoyancy-driven circulation with that driven by wind; each are outlined in detail in the following subsections.

3.1 Direct Numerical Simulation

Three-dimensional Navier-Stokes, advection-diffusion and continuity equations were solved in the incompressible, nonhydrostatic, Boussinesq approximation in a rotating coordinate frame using Direct Numerical Simulation (DNS). The flow solution assumes a linear equation of state with density solely dependent on temperature, and is calculated in dimensionless form. The algorithm and approach has previously been described and tested for large Rayleigh number convection (Gayen, Griffiths, & Hughes, 2014; Gayen, Hughes, & Griffiths, 2013; Vreugdenhil, Gayen, & Griffiths, 2016). We consider a rectangular basin geometry of length and width L and water depth H , with an aspect ratio $A = H/L = 0.2$. No-normal flow and no-slip boundaries for velocity and no flux temperature condition were used at the side-walls and bottom surface. A stress-free condition for velocity was imposed at the upper surface, and upper surface boundary temperature is prescribed as $T = \Delta T |\sin(\pi y/2L)|$, where (x, y) are horizontal coordinates and ΔT is the maximum temperature difference; flux into the fluid domain occurs through molecular diffusion. The thermal forcing is expressed in terms of the dimensionless Rayleigh number $Ra = g\alpha\Delta TL^3/\kappa\nu$, with gravitational acceleration g , applied temperature difference ΔT across the basin length, and the fluids thermal expansion coefficient α , molecular diffusivity of heat κ and molecular kinematic viscosity ν . We use variable Coriolis frequency represented by $f = f_0 \sin(\pi y/2L)$; we omit the non-traditional Coriolis terms despite simulating in the non-hydrostatic regime. The strength of rotation is represented by the Rossby number $Ro = u_g/Lf_0$ where $u_g = g\alpha\Delta T/f_0$ is geostrophic velocity scale. Simulations are performed at $Ra = 8 \times 10^{11}$, $Ro = 0.037$ and Prandtl number $Pr = 5$. In dimensional terms, assuming molecular values of viscosity, the domain of this simulation is on the laboratory scale – approximately 1 m horizontal and 20 cm deep, and the reduced gravity, $g\alpha\Delta T \approx 0.15 \text{ m s}^{-2}$.

The solution grid for the DNS is $1536 \times 1536 \times 512$ cells, clustered to resolve the upper thermal boundary layer and bottom Ekman layer, side boundary layers on the vertical walls and active convection areas. The grid resolution $[\Delta x, \Delta y, \Delta z]$ is comparable to the Batchelor microscale $\eta_b = (\nu^3/\varepsilon^*)^{1/4} Pr^{-1/2}$ (where ε^* is the local dissipation rate) to ensure the resolution criterion $\eta_b \geq [\Delta x, \Delta y, \Delta z]_{\max}$ is satisfied and the smallest scales of motion are accurately resolved everywhere in this domain (Gayen et al., 2014). Most importantly, adequate resolution was confirmed by accurate closure of the mechanical energy budget (Gayen et al., 2014; Vreugdenhil, Gayen, & Griffiths, 2016). Thermal equilibrium was achieved when secular change in the energetic quantities was not detectable, with net heat input continuing to fluctuate about zero with amplitude of the order of 1.5% of the heat throughput.

3.2 Ocean Model Simulations

The Direct Numerical Simulation is designed to resolve all scales of turbulence and convection, but has a larger aspect ratio and smaller Rayleigh number than the ocean. Hence, DNS techniques cannot be used to estimate the relative magnitude of gyres driven by buoyancy and wind stress. These questions are addressed using an eddy-permitting (0.25°) ocean circulation model in an idealised domain (Fig. 1). We use MOM6 (Adcroft et al., 2019) in a rectangular domain spanning 40° in longitude and from the Equator to 70°N in latitude, with 0.25° resolution on a Mercator grid and 75 vertical levels (using the z^* vertical coordinate) in a 4000 m deep basin. The Equatorial boundary includes

209 a thin (3° wide) region in which vertical diffusivity is enhanced by a factor of 500 to close
 210 the overturning circulation. Other side boundaries are insulating, with a vertical wall
 211 in the north and sloping sidewalls to the east and west. The surface forcing includes steady
 212 zonal wind stress, τ , and temperature restoring to a meridionally symmetric profile with
 213 piston velocity F_C , equivalent to a restoring flux, $\lambda = \rho C_p F_C$ in $\text{W m}^{-2} \text{K}^{-1}$ (see Haney,
 214 1971). The equation of state depends linearly on temperature. The upper ocean has an
 215 amplified diffusivity (up to $K_\rho = 2 \times 10^{-3} \text{ m}^2/\text{s}$ in the reference case, compared with
 216 $K_\rho = 2 \times 10^{-5} \text{ m}^2/\text{s}$ in the interior) as a crude parameterization for mixed layer tur-
 217 bulance (Vreugdenhil, Hogg, Griffiths, & Hughes, 2016). Convection is handled via CVMix
 218 whereby diffusivity is enhanced at statically unstable interfaces. The surface forcing per-
 219 mits variation of the wind stress (τ) and buoyancy forcing (F_C) that can be varied in-
 220 dependently, noting the difficulty of unambiguously separating the forcing components
 221 in such a nonlinear system.

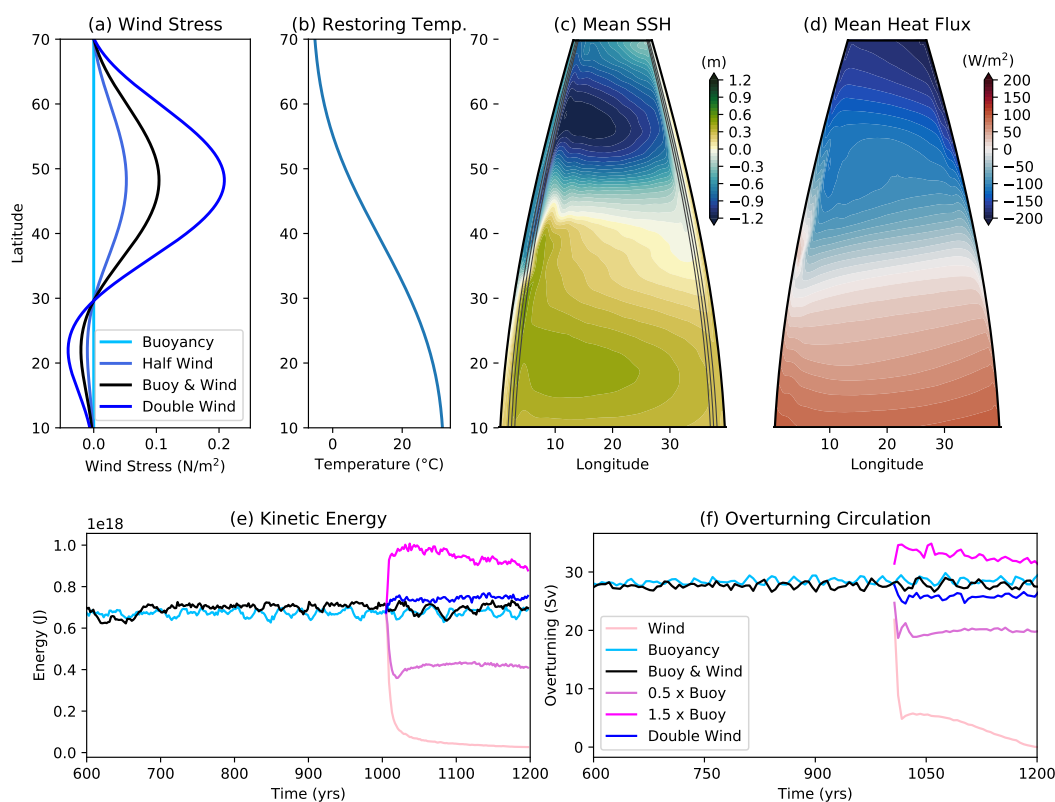


Figure 1. Ocean model configuration. (a) Wind stress; (b) Surface restoring temperature field; (c) Mean sea surface height in reference case, with topography indicated by contours; (d) Mean surface heat flux for reference case; (e) Time series of domain-wide kinetic energy; and (f) Time series of the residual overturning circulation calculated in density space and evaluated at $20\text{--}30^\circ\text{N}$ for 6 selected cases.

222 Eight experiments are conducted. The reference case (referred to as Buoyancy &
 223 Wind) uses control values of wind stress (black line in Fig. 1a with $\tau_{\text{max}} = 0.1 \text{ N m}^{-2}$),
 224 with the buoyancy flux via the temperature restoring profile from Fig. 1b (with $F_C =$
 225 0.2 m/day , equivalent to $\lambda \approx 10 \text{ W m}^{-2} \text{K}^{-1}$). The reference case is run for 1200 years
 226 to demonstrate full equilibration. We vary wind stress, with a Double Wind case, a Half
 227 Wind case and an additional case with no wind stress forcing (referred to as Buoyancy;

228 Fig. 1a) with each simulation run for 200 years from the year 1000 state of the Refer-
 229 ence simulation as the initial condition. To further demonstrate equilibration, we run
 230 the Buoyancy case for the full 1200 years, and also as a 200 year perturbation, with re-
 231 sults indistinguishable (see Fig. S1, Supplementary Material). We vary buoyancy flux
 232 by varying F_C from 0.0 through to 0.3 m/day. Note that the Wind case, in which buoy-
 233 ancy flux is completely switched off ($F_C = 0.0\text{m/day}$) erodes the stratification to gener-
 234 ate a barotropic flow after ~ 100 years.

235 4 Results

236 4.1 Direct Numerical Simulation

237 The role of buoyancy forcing in driving ocean gyres is first tested using Direct Nu-
 238 merical Simulation (DNS). We apply differential thermal forcing to the surface of an ide-
 239 alised domain (heating in the south, cooling in the north) with low Rossby number ($Ro =$
 240 0.037) to simulate planetary rotation and large Rayleigh number ($Ra = 8 \times 10^{11}$) to
 241 maintain turbulent convection. This convection is active over the northern part of the
 242 domain where cooling destabilizes the water column, forming vertical plumes that ext-
 243 tend through the full depth of the box (vertical velocity plane in Fig. 2). A northward
 244 current forms along the eastern boundary where it is held against the boundary by Cori-
 245 olis force. This current initiates cyclonic gyre circulation filling the width of the box over
 246 the cooling region as shown by the depth-averaged streamfunction (blue contours, up-
 247 per plane of Fig. 2). Thus, even in the absence of wind forcing, a cyclonic gyre forms
 248 analogous to an oceanic subpolar gyre.

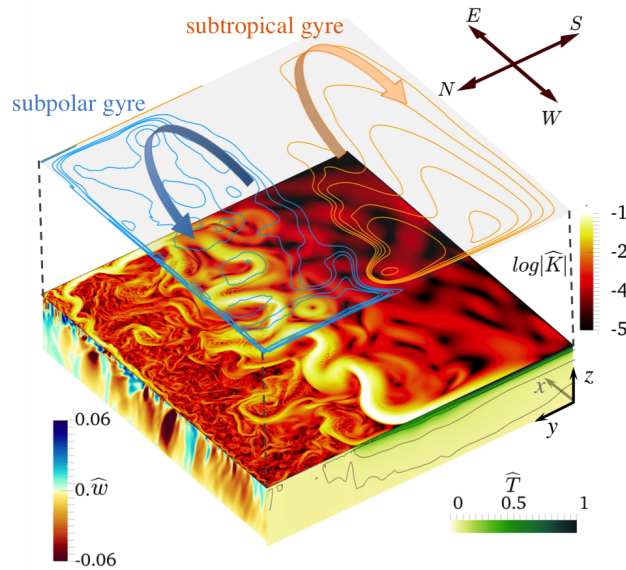


Figure 2. Snapshot of DNS illustrating the character of buoyancy forced ocean gyre circulation. The horizontal plane shows normalized kinetic energy $1/2[u^2 + v^2 + w^2]/(g'/f_0)^2$ (on a logarithmic scale) on the plane $z = H$. The panel on the northern boundary shows vertical velocity w normalized by g'/f_0 (upward in blue, downward in brown) at $y = 0.95L$. The panel on the western boundary shows the normalized temperature field $(T - T_C)/\Delta T$ at $x = 0.05L$. These two planes are within their respective lateral boundary layers. The elevated floating plane shows contours of mean depth-integrated streamfunction. Gravity is oriented vertically downwards, rotation is anticlockwise and solutions are obtained at a thermally equilibrated state.

249 Conversely, in the southern part of the domain, heating stabilizes the flow (see the
250 $y - z$ plane in Fig. 2). Here, the barotropic streamfunction indicates a southward in-
251 terior transport (consistent with Eq. 5), with a northward current of warm water along
252 the western boundary. This current is intensified by the presence of Rossby waves, which
253 propagate from east to west (see supporting movie). The current separates from the bound-
254 ary near the mid-point of the domain (also the location of the negligible heat input, anal-
255 ogous to the latitude of negligible wind stress curl in Sverdrup theory) and travels east-
256 ward across the basin, constrained by thermal wind balance (Vallis, 2006). This current
257 is baroclinically unstable and produces a rich mesoscale eddy field. The western bound-
258 ary current and eastward geostrophic basin-scale flow, along with weak southward in-
259 terior flow, form a large-scale anticyclonic gyre. The double-gyre structure found here
260 differs from the single dominant gyres found in previous simulations (Colin de Verdière,
261 1988; Gjermundsen et al., 2018), suggesting that resolving finer scales in the flow may
262 be important for generating realistic buoyancy-driven gyres. The circulation in this sim-
263 ulation is broadly consistent with oceanic midlatitude gyre circulation, despite the ab-
264 sence of any wind stress forcing in these simulations, demonstrating the principle that
265 gyres could be driven by buoyancy fluxes alone.

266 4.2 Ocean model simulations

267 The ocean model simulations constructed here use high resolution to explicitly rep-
268 resent eddy fluxes; simulations conducted at lower resolution (1° lateral resolution with
269 parameterised eddies) show a very different character in the circulation (see Fig. S2, Sup-
270 plementary Material). The model is forced by a combination of wind stress and surface
271 buoyancy fluxes (Fig. 1), which are varied independently in 8 different experiments. The
272 ocean model state is averaged over the last 100 years of a 200-year perturbation to es-
273 timate the circulation (Fig. 3).

274 The barotropic (depth-integrated) circulation in these experiments (Fig. 3a-e) em-
275 phasises the formation of gyre-like circulation in the subpolar region (north of 45°N) of
276 this idealised domain. When only wind stress forcing is applied (Fig. 3a), a subpolar gyre
277 arises that closely matches the prediction of Sverdrup balance. With only buoyancy forc-
278 ing (Fig. 3b) a gyre still exists, noting that the northward meridional circulation in this
279 case is dominated by transport along the Eastern boundary. Importantly, however, the
280 spatial extent of the gyre is similar to the Wind case, and incorporates an inertial west-
281 ern boundary current that separates from the coast at mid-latitudes. Buoyancy and wind
282 together (Fig. 3c) produce a stronger gyre in which elements of both interior and east-
283 ern boundary transport combine. Further strengthening either the wind or buoyancy forc-
284 ing acts to enhance the circulation in this gyre.

285 The barotropic (depth-averaged) subtropical gyre (Equatorward of 40°N) in these
286 experiments is relatively weak, but we show below that stronger horizontal circulation
287 occurs above the thermocline, similar to the circulation proposed in the ventilated ther-
288 mocline theory (Luyten et al., 1982). To highlight the nature of this upper layer circula-
289 tion, we follow Luyten et al. (1982) in selecting an isopycnal which outcrops close to
290 the zero wind stress curl line ($\rho = 1034 \text{ kg m}^{-3}$) and integrate transport above this isopy-
291 cnal. To connect the upper layer transport to the ventilated thermocline theory, we re-
292 move the contribution of the meridional overturning circulation which flows northwards
293 in the upper lower and southwards at depth; specifically, we calculate the basin- and layer-
294 averaged transport at 10°N and subtract this transport from the streamfunction at all
295 latitudes. The remainder gives a streamfunction which clearly delineates a subtropical
296 upper-layer circulation (Fig. 3 f-j). This gyre-like circulation is consistent with mean Sea
297 Surface Height in these experiments, which is shown in Fig. S3.

298 The isopycnal we have used to define the upper ocean transport is found at 500-
299 800 m depth in most experiments; the main exception being the Wind only case (Fig. 3f)

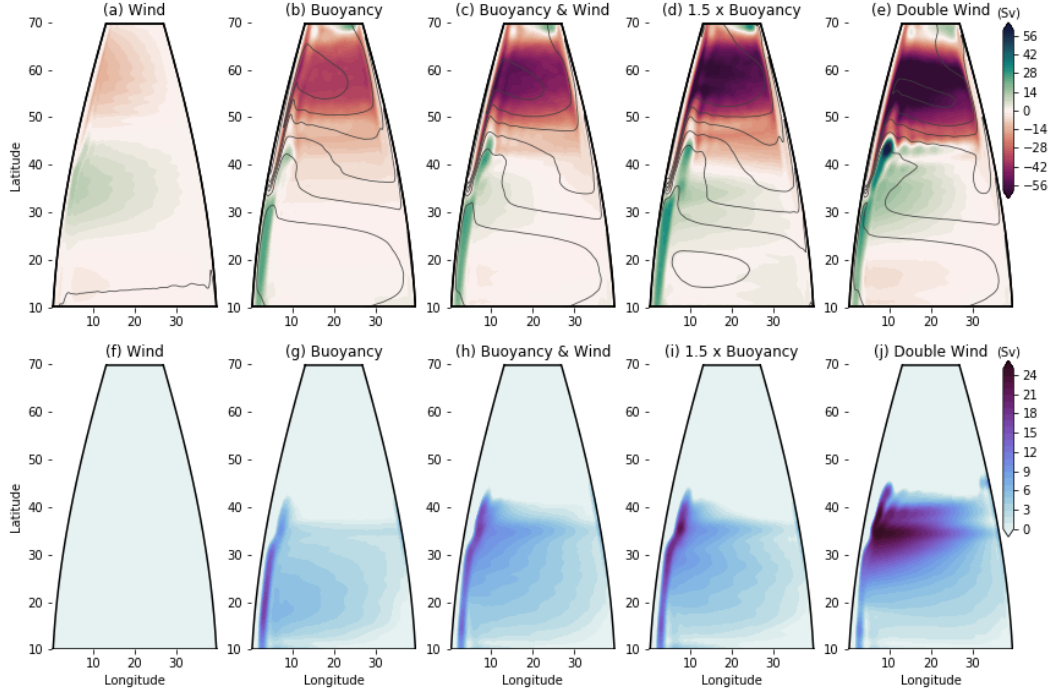


Figure 3. Mean circulation for selected ocean model simulations. (a-e) Barotropic mean streamfunction, overlain with contours of mean Sea Surface Temperature (Contour Interval 2°C); (f-j) Upper streamfunction, which selects the subtropical gyre above the thermocline. The cases shown are (a,f) Wind forcing only, noting that the upper ocean streamfunction in (f) is not defined; (b,g) Buoyancy forcing only; (c,h) Reference case with Buoyancy & Wind stress together; (d,i) Reference wind stress with increased buoyancy flux, and (e,j) Reference buoyancy with doubled wind stress.

300 where this isopycnal does not exist. However, in the reference case with Buoyancy and
 301 Wind forcing (Fig. 3h), the linear component of the gyre at $20\text{-}30^{\circ}\text{N}$ closely matches Luyten
 302 et al.'s (1982) ventilated thermocline prediction, noting the formation of a stronger, in-
 303 inertial gyre close to the latitude western boundary current separation. This nonlinear re-
 304 circulation is amplified by stronger wind forcing (Fig. 3j), but is relatively insensitive
 305 to additional buoyancy flux (Fig. 3i). Nonetheless, in the Buoyancy-only case (Fig. 3g)
 306 a linear gyre with transport of around 10 Sv is generated.

307 We now make a quantitative comparison of the gyre circulation in both the barotropic
 308 and upper ocean streamfunctions. We calculate the transport in each experiment at 10°E
 309 to avoid the contribution of nonlinear flow on the western boundary. We take the ab-
 310 solute value of the minimum barotropic streamfunction to represent the subpolar gyre
 311 transport (Fig. 4a,c,d), and the maximum value of the upper ocean circulation to rep-
 312 resent the subtropical gyre transport (Fig. 4b,e,f). The timeseries of selected cases (Fig.
 313 4a,b) confirm that the circulation is in equilibrium after $O(100)$ years of the perturba-
 314 tion simulations.

315 The barotropic subpolar gyre strength (averaged over the last 100 years) varies lin-
 316 early with heat flux (Fig. 4c) and is consistent with predictions from Eq. (5) (evaluated
 317 for $h = 1500\text{m}$ and from 45°N - 60°N , where the meridional temperature gradient is max-
 318 imum; see Fig. S4). The subpolar gyre also increases linearly with wind stress (Fig. 4d),
 319 however is $\sim 50\text{ Sv}$ larger than the linear Sverdrup transport prediction. We infer from

320 this result that the barotropic circulation of the subpolar gyre in this model is governed
 321 primarily by surface buoyancy flux, consistent with Eq. (5).

322 On the other hand, the upper layer circulation in the subtropical gyre is similar
 323 to the ventilated thermocline prediction (estimated by the blue dashed line in Fig. 4f),
 324 particularly at large values of wind stress. At lower values of wind stress a small (10 Sv)
 325 circulation persists, which we infer is driven by surface buoyancy forcing; however, the
 326 dependence of the circulation on surface buoyancy flux is over-predicted by Eq. (5) (see
 327 Fig. 4e), implying that the simplifications made in formulating that balance are not valid
 328 in the regime of these experiments. We assume that the most significant omission is that
 329 of eddy fluxes (Eq. 4), and infer from this result that linear theory is insufficient to pre-
 330 dict the formation of the buoyancy-forced subtropical gyre in these simulations. Nonethe-
 331 less, the existence of a finite baroclinic subtropical gyre (Fig. 3g and cyan dot in Fig.
 332 4f) – alongside a barotropic subtropical gyre (Fig. 3b and cyan dot in Fig. 4d) – in
 333 the absence of any wind stress forcing confirm the critical importance of surface buoy-
 334 ancy forcing in contributing to large scale ocean gyres.

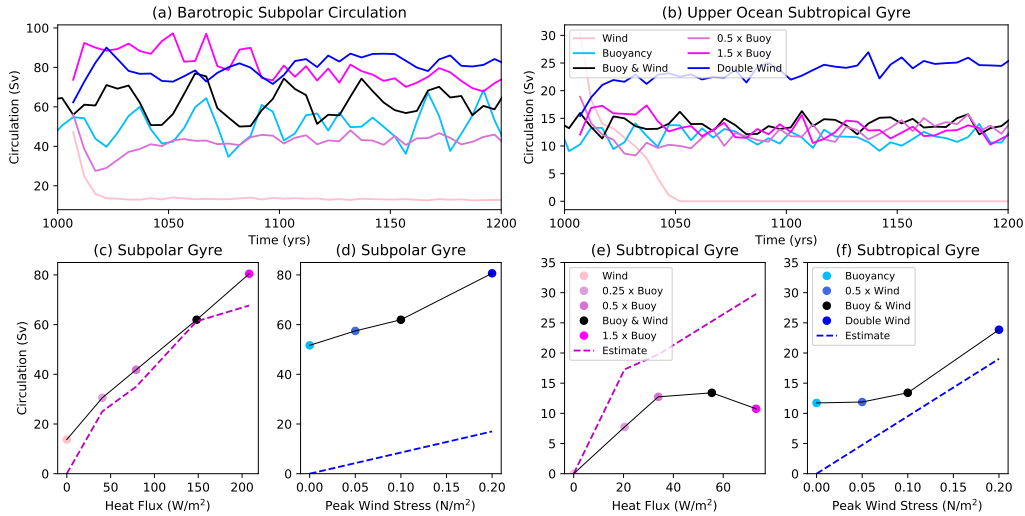


Figure 4. Integrated statistics for ocean model simulations and comparison with scaling. Time series of of: (a) Barotropic subpolar gyre transport, and (b) Subtropical upper ocean gyre transport. Subpolar gyre dependence on: (c) heat flux and (d) wind stress; subtropical gyre dependence on (e) heat flux and (f) wind stress. Dashed lines show theoretical scaling, estimated from (c) Eq. (5) using the upper 1500m depth-averaged meridional temperature gradient from 45°N-60°N; (d) The Sverdrup transport relation (Eq. 2); (e) Equation (5) using the upper 500m depth-averaged meridional temperature gradient 20°N-35°N, and (f) ventilated thermocline theory (Luyten et al., 1982).

5 Discussion & Conclusions

335 The theoretical framework and the numerical model results presented here suggest
 336 that the accepted moniker “wind-driven gyres” is an incomplete description of the mid-
 337 latitude ocean circulation. The notion that buoyancy forcing contributes to the trans-
 338 port of, particularly, the subpolar gyre is robust across several levels of idealisation, im-
 339 plying that these results may apply to the ocean. This notion is consistent with earlier
 340 studies on buoyancy driven gyres (e.g. Colin de Verdière, 1988), but is more significant
 341 when modelled in the eddying regime. In particular, previous attempts to model buoy-
 342

343 ancy driven gyres at lower resolution (Colin de Verdière, 1988; Gjermundsen et al., 2018)
 344 tended to create a single large-scale gyre in each basin; in the eddy permitting config-
 345 uration we find that a more realistic double gyre, with an inertial western boundary cur-
 346 rent, emerges naturally even in the absence of wind stress. We reiterate that these re-
 347 sults are not intended to replace the theory of wind-driven gyres, but merely to point
 348 out that a component of the gyre circulation, particularly in subpolar regions, can be
 349 forced by surface buoyancy flux.

350 The simulations in this study are compared with estimates based on linear theory.
 351 These comparisons highlight that the theories employed are unable to fully describe the
 352 system: in particular, linear buoyancy theory overpredicts the baroclinic subtropical gyre,
 353 while the Sverdrup transport equation underpredicts the barotropic subpolar gyre cir-
 354 culation in these simulations. However, the sensitivity of the subpolar gyre to buoyancy
 355 forcing, in combination with the existence of significant subpolar and subtropical gyre
 356 structures in the absence of wind stress, underscores the importance of incorporating sur-
 357 face buoyancy fluxes into our understanding of gyre circulation.

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Figure 1.

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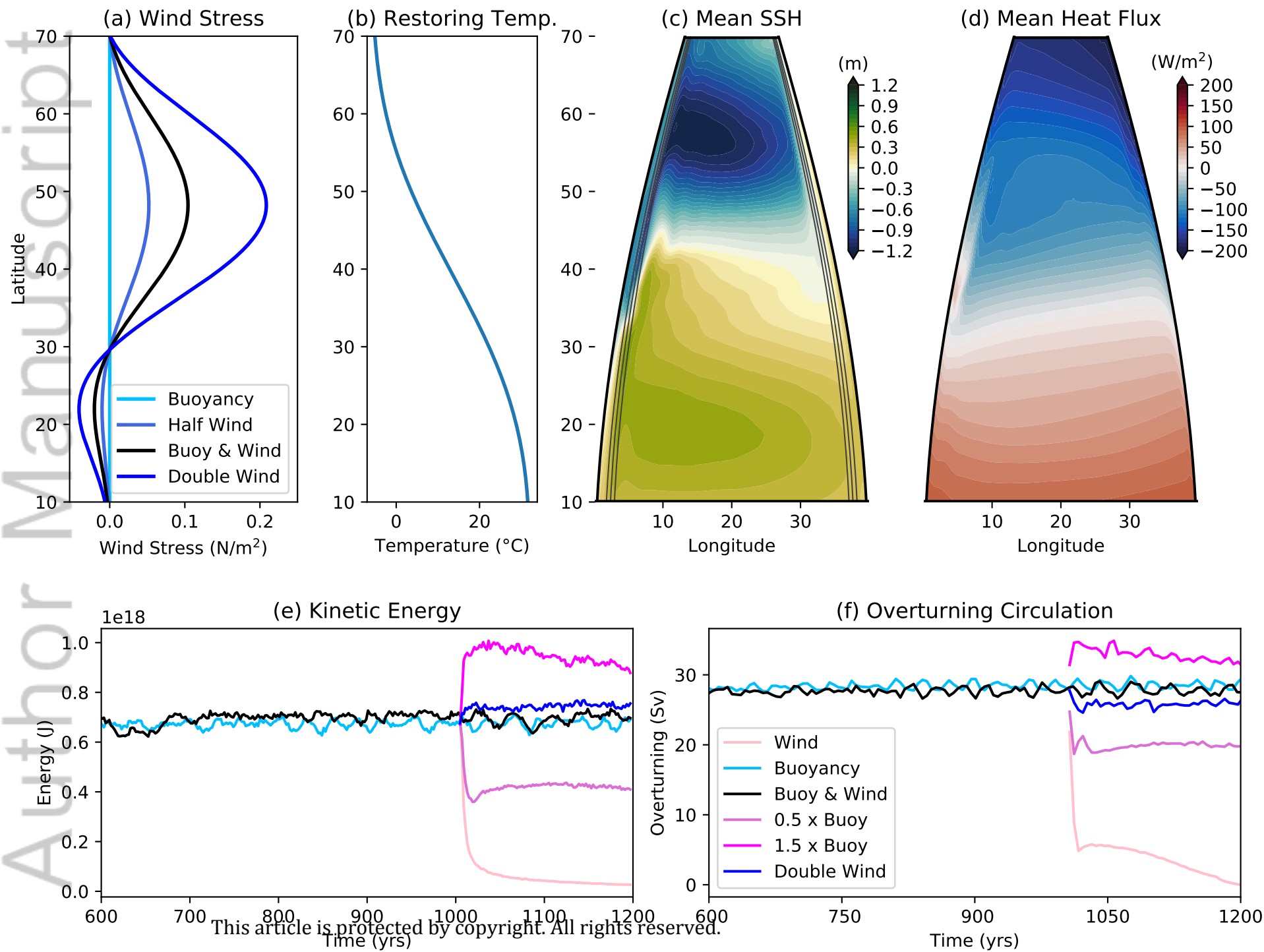


Figure 2.

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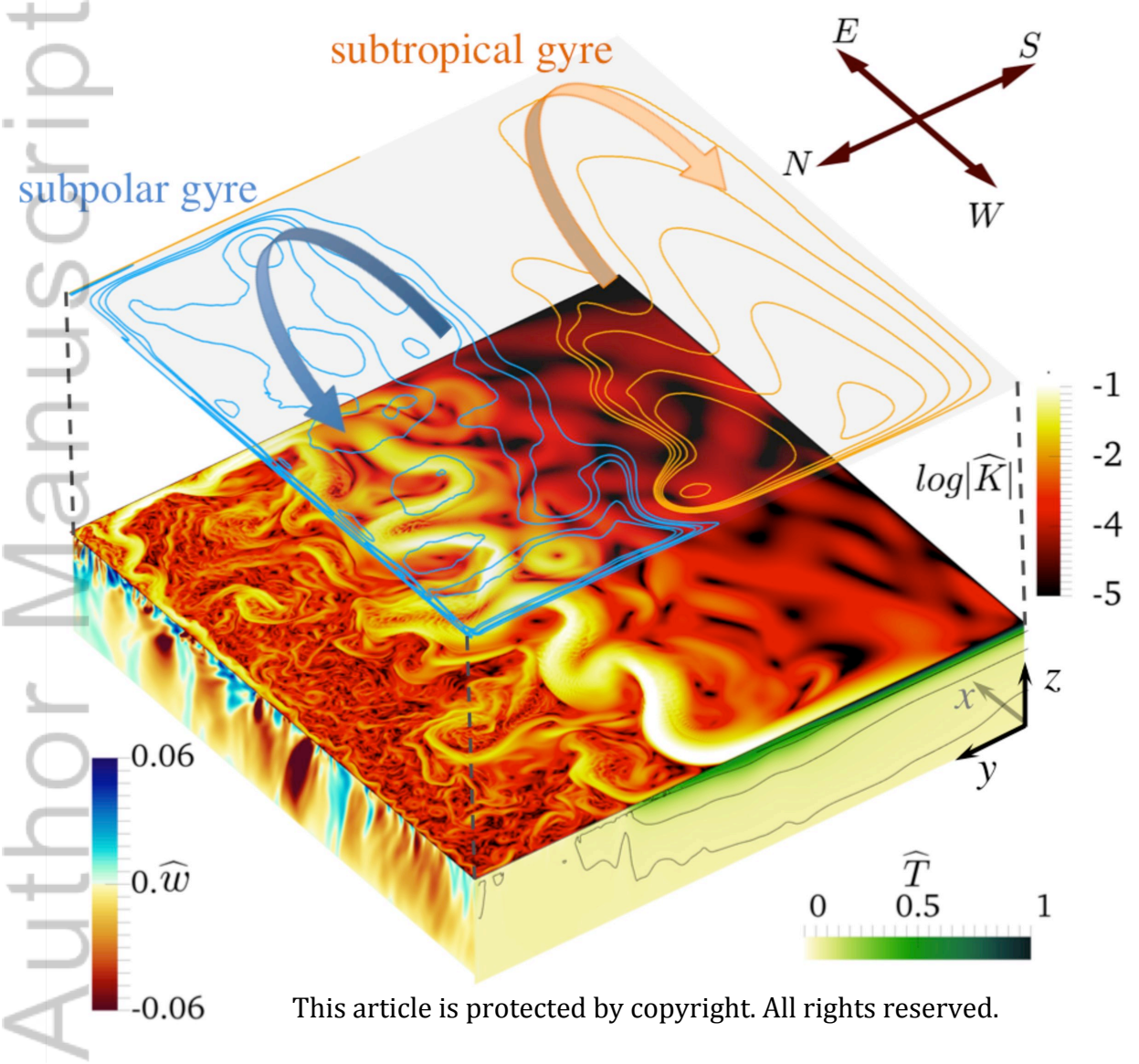


Figure 3.

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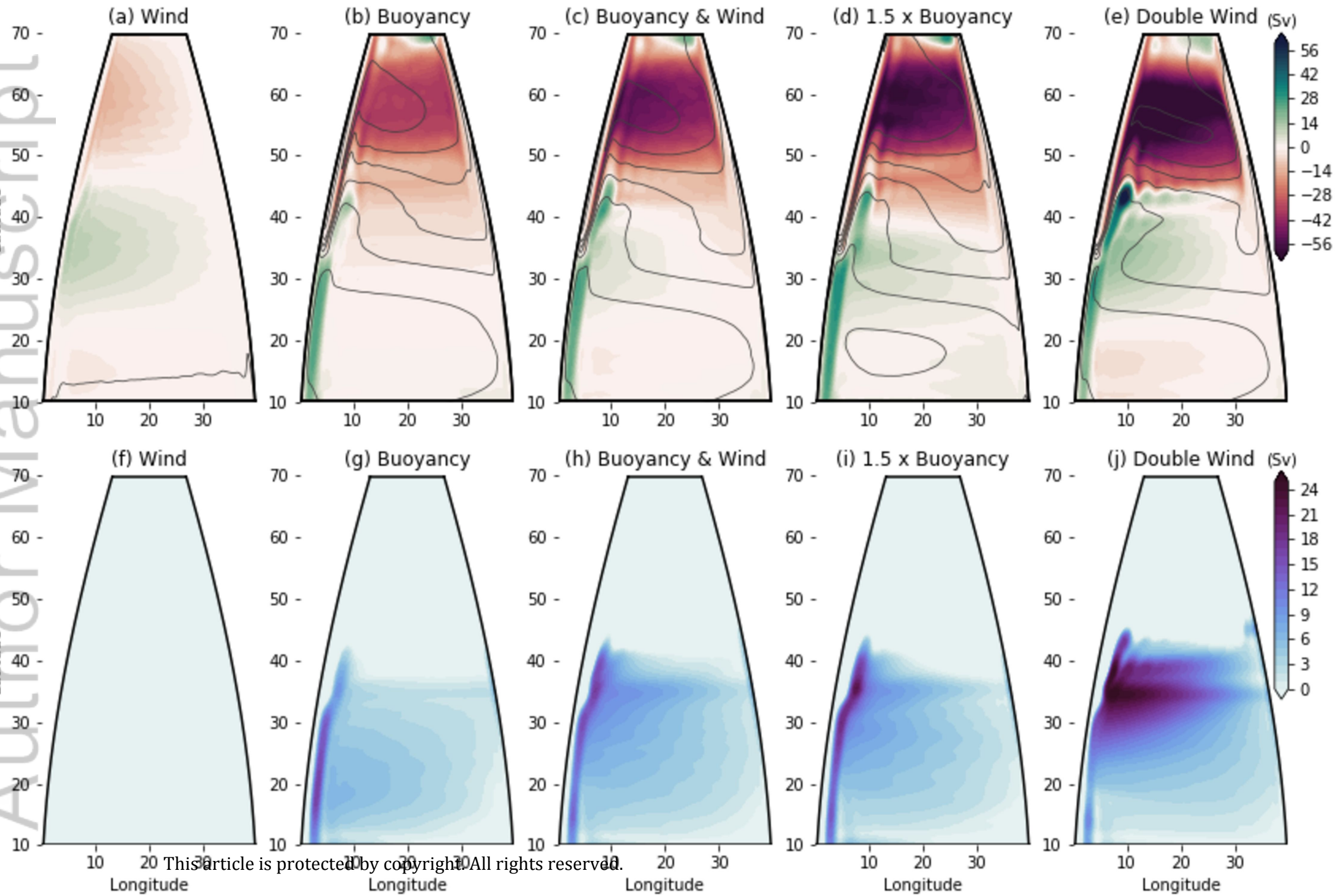
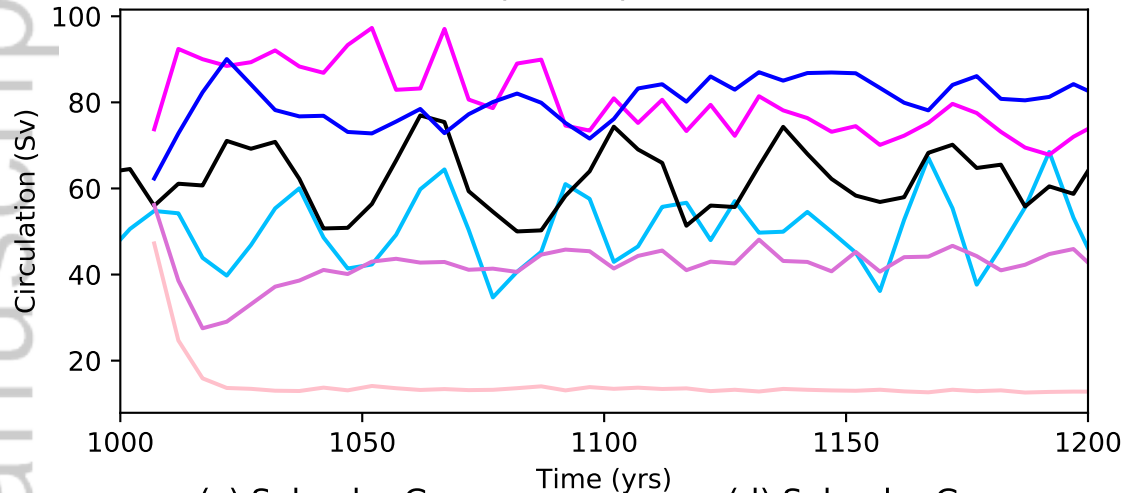


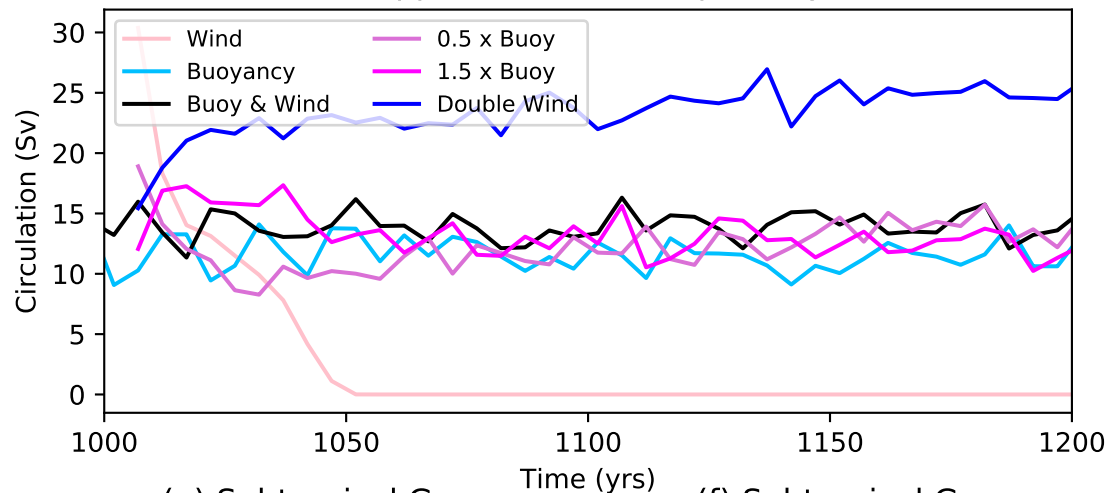
Figure 4.

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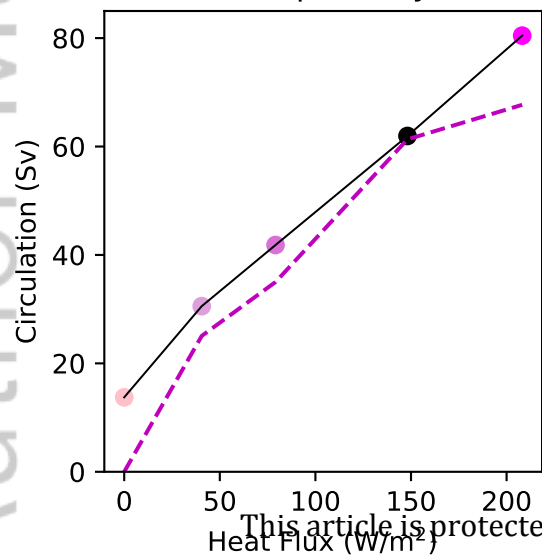
(a) Barotropic Subpolar Circulation



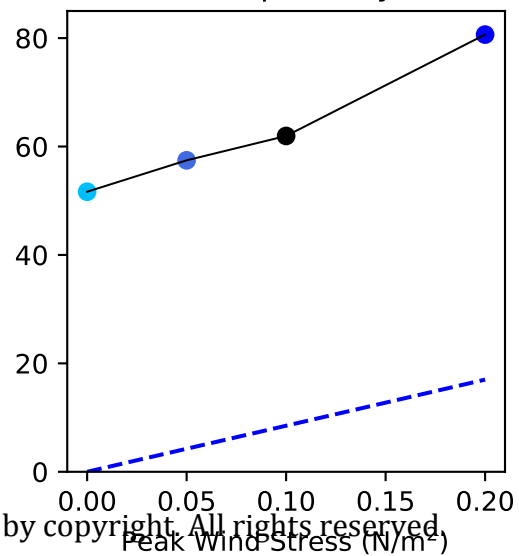
(b) Upper Ocean Subtropical Gyre



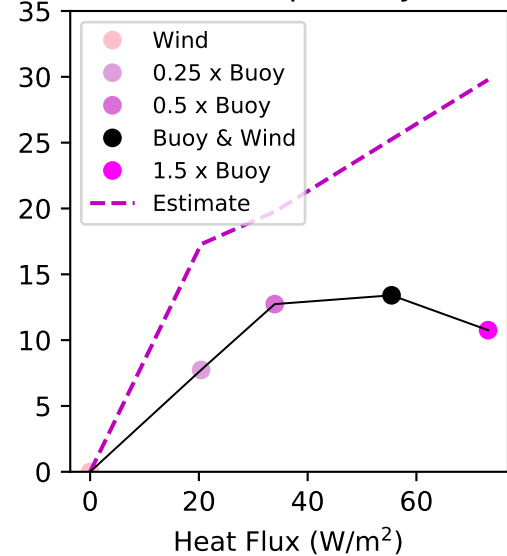
(c) Subpolar Gyre



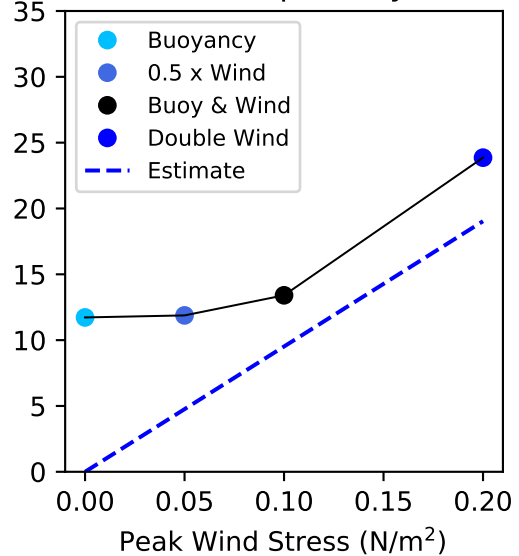
(d) Subpolar Gyre

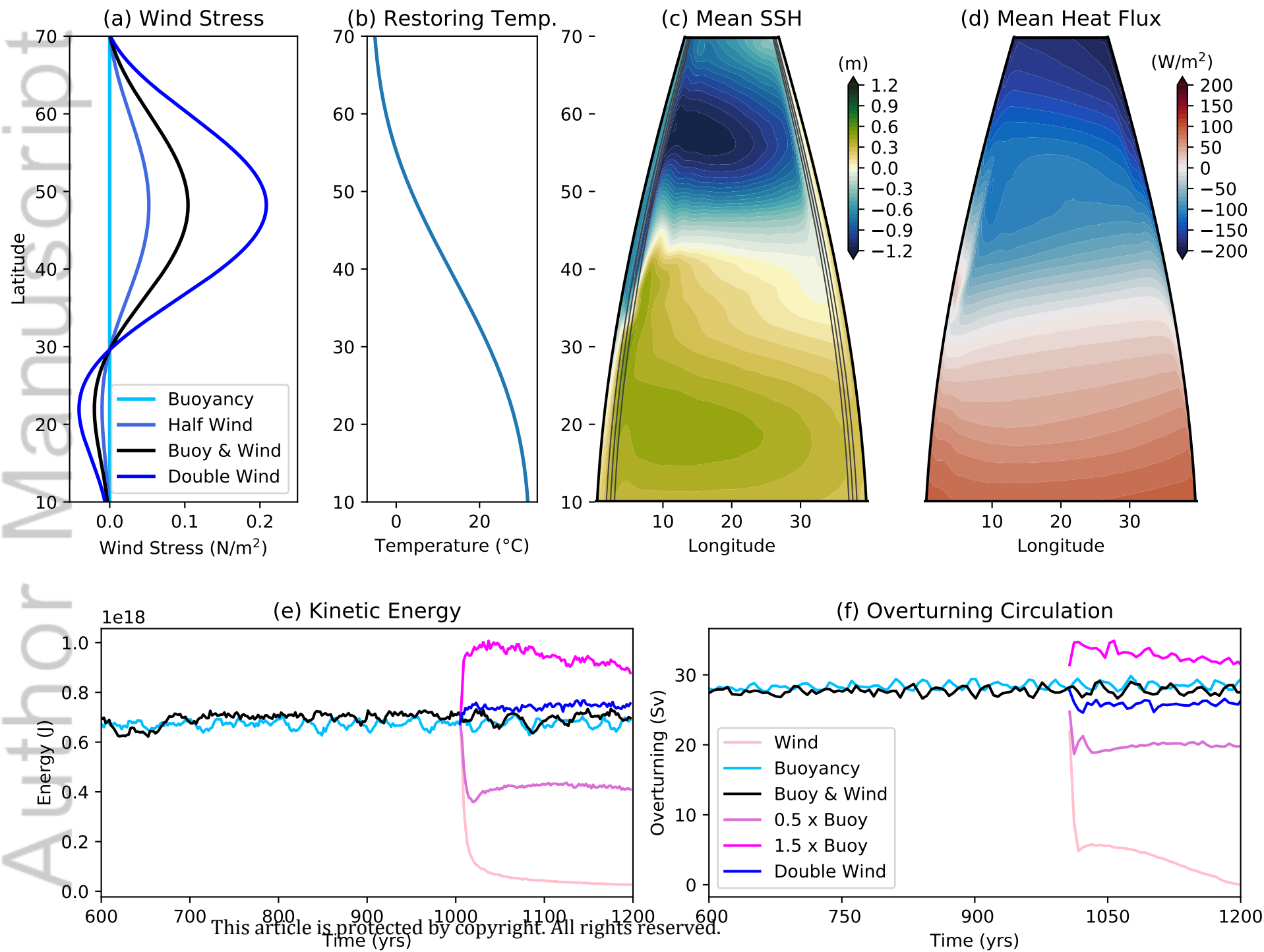


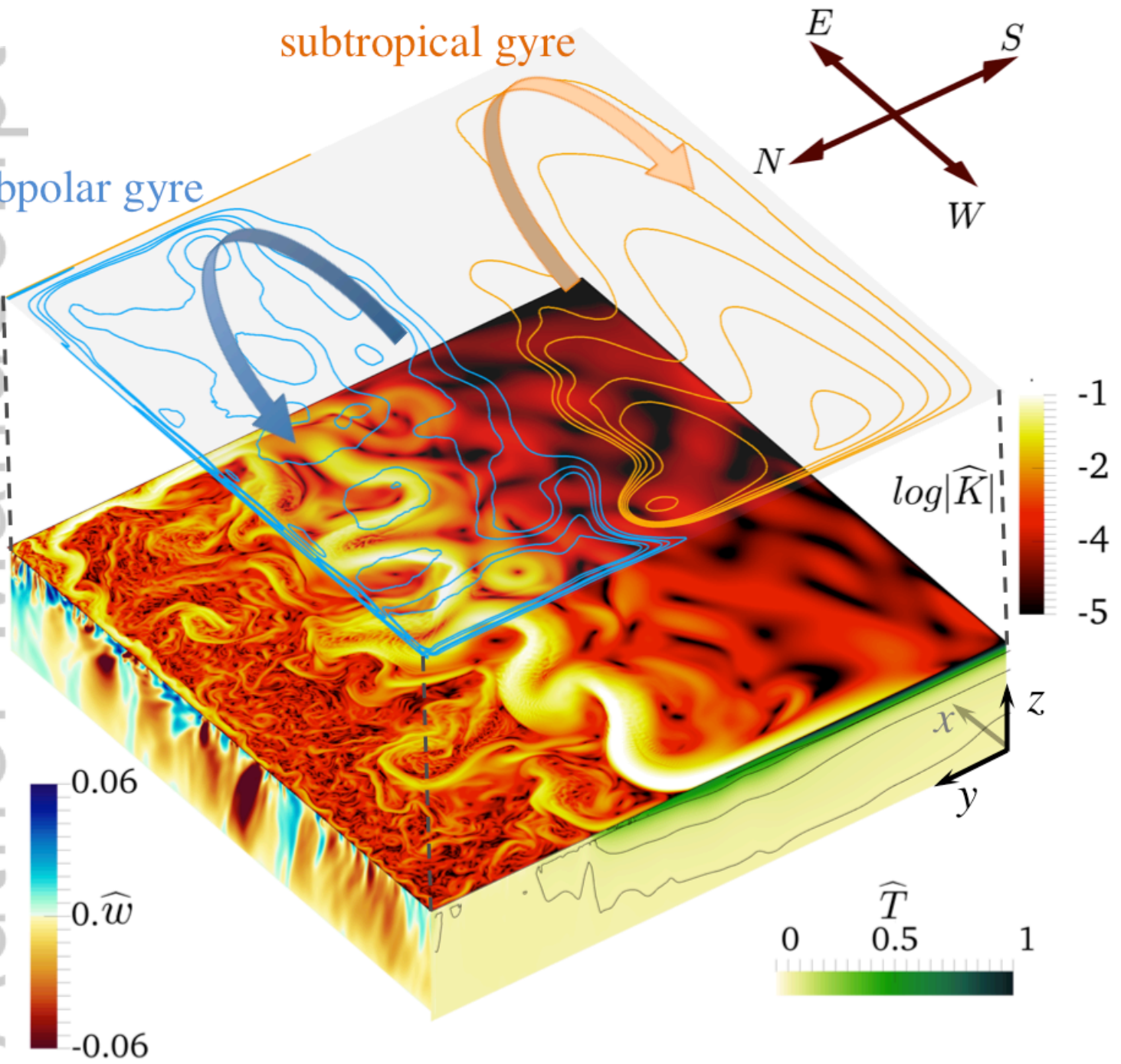
(e) Subtropical Gyre



(f) Subtropical Gyre

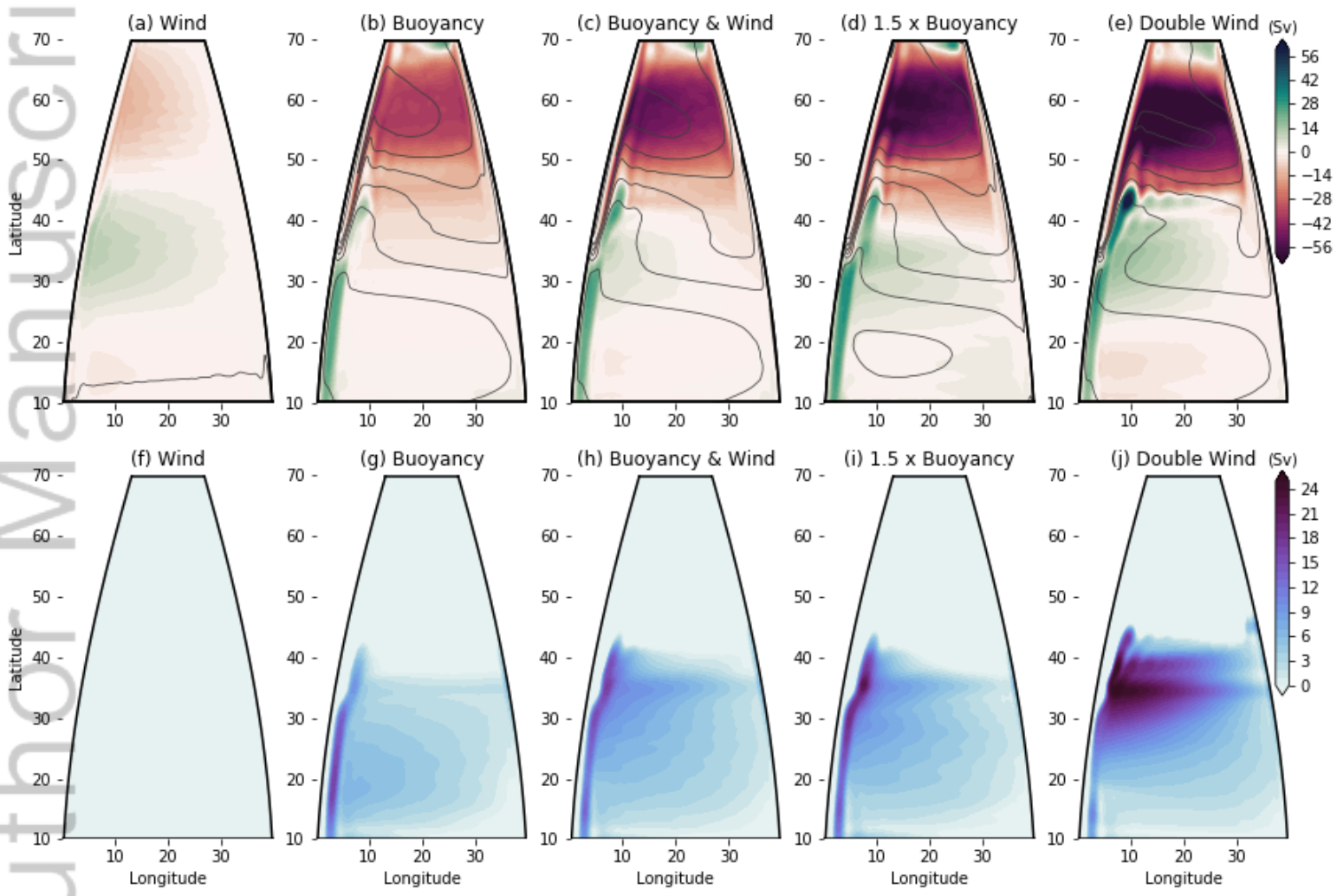






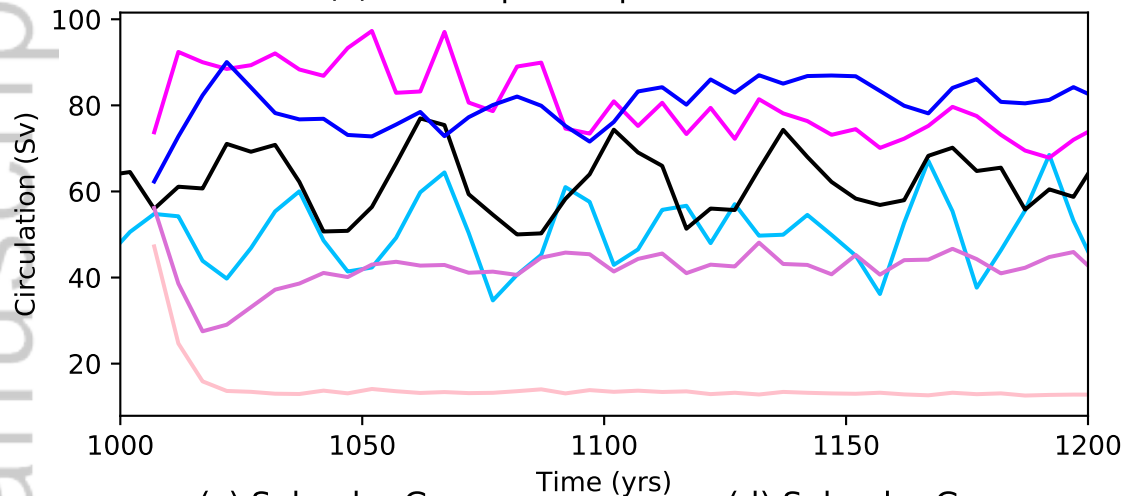
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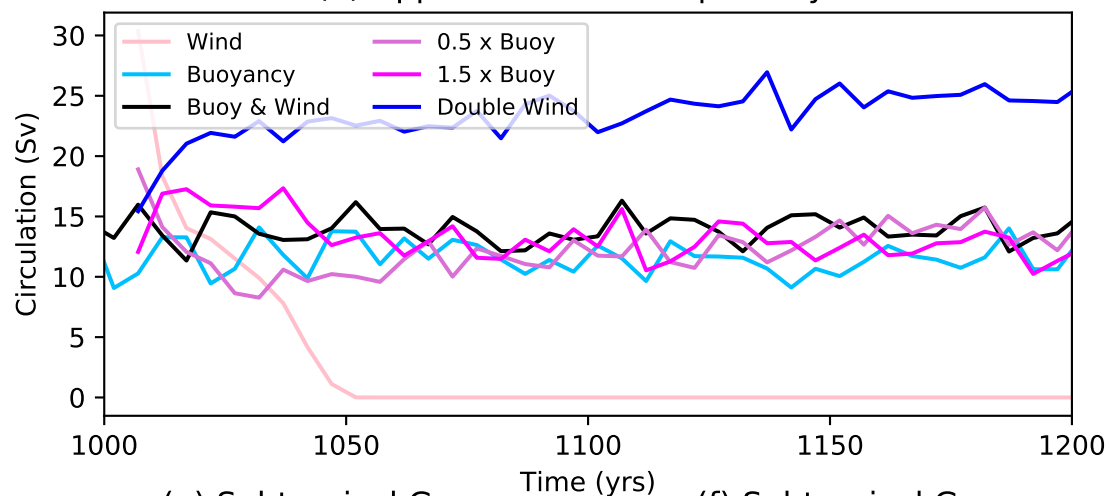


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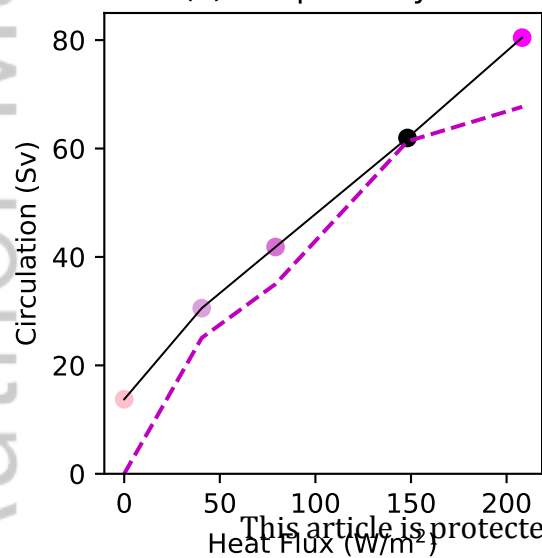
(a) Barotropic Subpolar Circulation



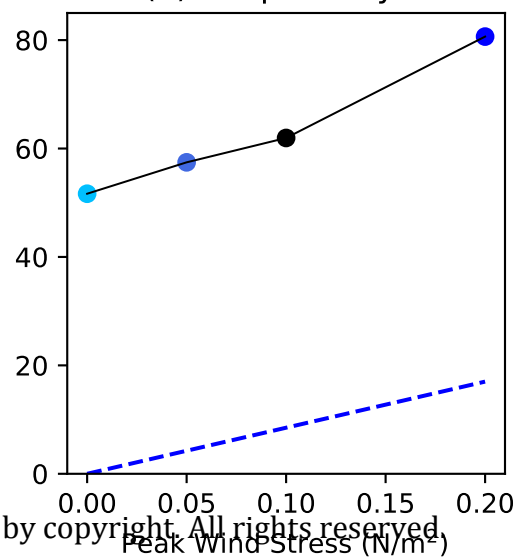
(b) Upper Ocean Subtropical Gyre



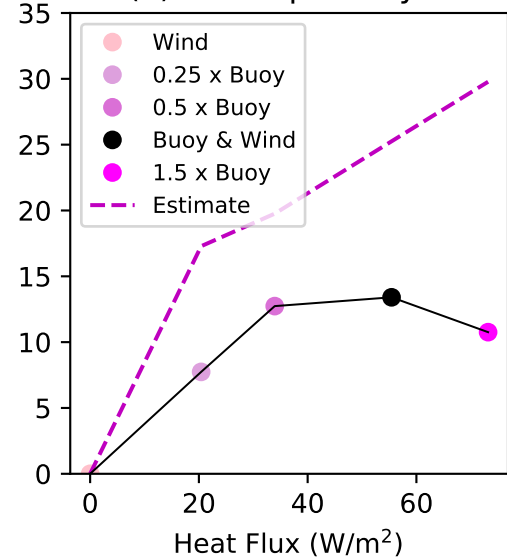
(c) Subpolar Gyre



(d) Subpolar Gyre



(e) Subtropical Gyre



(f) Subtropical Gyre

