



Freshwater transport in the coupled ocean-atmosphere system: a passive ocean

Article

Accepted Version

Ferreira, D. and Marshall, J. (2015) Freshwater transport in the coupled ocean-atmosphere system: a passive ocean. *Ocean Dynamics*, 65 (7). pp. 1029-1036. ISSN 1616-7341 doi: <https://doi.org/10.1007/s10236-015-0846-6> Available at <http://centaur.reading.ac.uk/40161/>

It is advisable to refer to the publisher's version if you intend to cite from the work.

To link to this article DOI: <http://dx.doi.org/10.1007/s10236-015-0846-6>

Publisher: Springer Verlag

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1 **Freshwater transport in the coupled**
2 **ocean-atmosphere system: a passive ocean**

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5 Received: date / Accepted: date

6 **Abstract** Conservation of water demands that meridional ocean and atmo-
7 sphere freshwater transports (FWT) are of equal magnitude but opposite in
8 direction. This suggests that the atmospheric FWT and its associated latent
9 heat (LH) transport could be thought of as a “coupled ocean/atmosphere
10 mode”. But what is the true nature of this coupling? Is the ocean passive or
11 active?

12 Here we analyze a series of simulations with a coupled ocean-atmosphere-
13 sea ice model employing highly idealized geometries but with markedly differ-

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ent coupled climates and patterns of ocean circulation. Exploiting streamfunc-
tions in specific humidity coordinates for the atmosphere and salt coordinates
for the ocean to represent FWT in their respective medium, we find that at-
mospheric FWT/LH transport is essentially independent of the ocean state.
Ocean circulation and salinity distribution adjust to achieve a return freshwa-
ter pathway demanded of them by the atmosphere. So, although ocean and
atmosphere FWTs are indeed coupled by mass conservation, the ocean is a
passive component acting as a reservoir of freshwater.

Keywords Freshwater transport · latent heat transport · hydrological cycle

1 Introduction

A large fraction of the Equator to Pole energy transport is achieved by the at-
mosphere through latent heat (LH) transport — see for example the summary
in Marshall and Plumb (2008). In mid- to high-latitudes where the atmo-
sphere dominates the total meridional energy transport, the LH contribution
is as large or larger than the dry static energy transport (DSE, the sum of
sensible and potential energy fluxes). The LH flux is associated with a trans-
port of moisture from regions of net evaporation in the subtropics to regions
of net precipitation in the tropics and high-latitudes. In steady state, the
atmospheric moisture transport must be balanced by an equal but opposite
freshwater transport (FWT) by the oceans (neglecting small storage terms
and transport by rivers and sea ice).

35 This constraint is the reason why the atmospheric LH transport is some-
36 times thought of as a “coupled mode of transport” (Rhines et al, 2008)¹ or a
37 “joint atmosphere-ocean process” (Bryden and Imawaki, 2001). For the same
38 reasons, Wijffels (2001) describes the oceanic FWT as a fundamental part of
39 the planetary energy budget. Such characterizations put the ocean, through
40 its FWT, at the heart of the coupled ocean-atmosphere energy cycle. Is this
41 justified? There is no doubt that the atmospheric LH transport and ocean
42 FWT are related. It is the *nature* of this coupling between atmospheric LH
43 transport and ocean circulation that is the focus of attention of the present
44 study. Does the oceanic FWT constrain the working of atmospheric LH trans-
45 port ? Or is the ocean passive, i.e. is the coupling one-way? At a time when the
46 hydrological cycle is predicted to intensify (Held and Soden, 2006) and salinity
47 is already observed to be changing at the surface and at depth (Durack and
48 Wijffels, 2010), it is important to clarify the nature of the relationship between
49 the two major components of the global hydrological cycle.

50 To the atmospheric scientist, the answer to the above question is obvious:
51 the atmospheric water cycle is driven by atmospheric processes and the ocean
52 is a passive agent, providing the reservoir of water but little more. To many
53 oceanographers, the answer is less clear: the FWT is a truly coupled prob-
54 lem with the ocean supplying both the freshwater and the heat required for
55 evaporation (among other arguments advanced below).

¹ From Rhines et al (2008): “Latent heat is fresh water (2.4 PW per Sverdrup), and its transport is an intrinsically coupled ocean/atmosphere mode”

56 This contribution is an attempt to bring some clarity to this discussion and
57 bridge the gap between the different perspectives. We hope it is a particular fit
58 to this Special Issue on “Atmosphere and Ocean dynamics” in honor of Richard
59 Greatbatch, whose work over the years has made such important contributions
60 to our understanding of the two fluids and their interaction.

61 To address this question, we analyze a series of idealized simulations with
62 a coupled ocean-atmosphere-sea ice General Circulation Model (GCM). We
63 employ highly idealized geometries in which continents are reduced to narrow
64 barriers. The sequence from Aqua to Ridge to Drake to Double-Drake can
65 be regarded as a “cartoon” that increases the level of geometrical complexity
66 (Fig. 1): from the pure Aquaplanet (where there are no topographic constraints
67 on ocean circulation) to the Double-Drake (in which interhemispheric and
68 zonal asymmetries are present) — see Marshall et al (2007); Enderton and
69 Marshall (2009). By changing the geometrical constraints, the sequence of
70 simulations switches in, one by one, key components of the ocean circulation
71 (subtropical cells, gyres, zonal jets, inter-hemispheric meridional overturning
72 circulation (MOC), etc). Our simulations range across a wide spectrum of
73 climates, as illustrated by the variety of forms of the MOC shown in Fig. 2
74 (right). As shown by Ferreira et al (2010), when viewed through the lens of heat
75 and freshwater transport, the climate of Double-Drake exhibits an uncanny
76 resemblance to the real world, with, notably, the localization of deep water
77 formation in the Small basin only, in analogy with the contrasting circulations
78 of the Atlantic and Pacific basins of today’s climate.

79 A central result of our study is immediately apparent in Fig. 2. Although
80 the ocean circulations of each coupled climate are very different — they have to
81 be because the geometry of the ocean basins differs so markedly between them
82 — the meridional FW transport hardly changes across the climate states. To
83 probe meridional transports further we diagnose overturning streamfunctions
84 in specific humidity q and in salt S coordinates for the atmosphere and ocean
85 respectively. Such approaches have been widely used in the atmosphere and
86 ocean, notably in studies of energy transports (e.g. Karoly et al (1997), Held
87 and Schneider (1999), Pauluis et al (2008) for the atmosphere; Saenko and
88 Merryfield (2006), Lumpkin and Speer (2007), Ferrari and Ferreira (2011) for
89 the ocean; Czaja and Marshall (2006) for the coupled system). Their appeal
90 lies in the fact that they naturally include transports associated with standing
91 and transient eddies and directly relate to the net meridional transport. Here,
92 q - and S -coordinate streamfunctions reveal atmospheric and oceanic FWTs,
93 respectively, and elegantly illustrate the symmetry between the two FWTs
94 and their connection through Evaporation minus Precipitation (E-P).

95 We find that the dynamics of the oceanic FWT takes very different forms
96 under rather similar E-P patterns and argue that the atmospheric FWT and
97 LH transports are largely independent of the ocean. That is, the “coupling” of
98 the atmospheric LH ”mode” is primarily one-way with the ocean responding
99 passively to atmospheric dynamics.

100 Our paper is organized as follows. Section 2 briefly describes our coupled
101 GCM and the computation of tracer-based streamfunctions. In section 3, we

investigate the S - and q -coordinate streamfunction and show how it informs us about the dynamics of FWT and the symmetry between ocean and atmosphere FWT in the climate system. Conclusions are given in section 4.

2 Model and methods

2.1 The coupled GCM

We use the MITgcm in a coupled ocean-atmosphere-sea ice set-up (Marshall et al, 1997a,b). All components use the same cubed-sphere grid at C24 resolution (3.75° at the equator) (Adcroft et al, 2004). Both ocean and atmosphere are primitive equation models and are generated from the same dynamical core exploiting an isomorphism between ocean and atmosphere dynamics (Marshall et al, 2004).

The atmospheric physics is based on the SPEEDY scheme (Molteni, 2003) at low vertical resolution (5 levels). It comprises a 4-band radiation scheme, a parametrization of moist convection, diagnostic clouds and a boundary layer scheme. The 3-km deep, flat-bottomed ocean model has 15 vertical levels. Effects of mesoscale eddies are parametrized as an advective process (Gent and McWilliams, 1990) together with an isopycnal diffusion (Redi, 1982), both with a transfer coefficient of $1200 \text{ m}^2\text{s}^{-1}$. Convective adjustment, implemented as an enhanced vertical mixing of potential temperature and salinity, is used to represent ocean convection (Klinger et al, 1996). The background vertical diffusion is uniform and set to $3 \times 10^{-5} \text{ m}^2\text{s}^{-1}$.

123 The sea-ice component is based on the Winton (2000) thermodynamic
124 model (two layers of ice plus surface snow cover). The prognostic variables are
125 ice fraction, snow and ice thickness, and a two-level enthalpy representation
126 which accounts for brine pockets, employing an energy conserving formulation.
127 The land model is a simple 2-layer model with prognostic temperature, soil
128 moisture, run-off, and snow height. The atmospheric CO₂ level is prescribed
129 at present day values. The seasonal cycle of insolation is represented (using an
130 obliquity of 23.5°, and zero eccentricity) but there is no diurnal cycle.

131 Finally, as discussed by Campin et al (2008), our coupled model achieves
132 perfect (machine-accuracy) conservation of freshwater, heat and salt during
133 extended climate simulations, a property which is crucial to the fidelity and
134 integrity of the coupled system. This is made possible by the use of the rescaled
135 height coordinate z^* in the ocean (Adcroft and Campin, 2004). Importantly
136 here, this coordinate permits the use of real freshwater boundary conditions
137 everywhere, including at the sea ice ocean interface. The set-up is identical
138 to that used in Ferreira et al (2010, 2011), to which the reader is referred for
139 further details.

140 All simulations used in this study were integrated for 5000 years or more
141 and reached a statistical equilibrium. Diagnostics are based on 50-year aver-
142 ages.

 143 2.2 Tracer-based overturning

144 For each set of contemporaneous 3d flow field and tracer field C (whether
 145 they are instantaneous or time-averaged fields), all meridional mass/volume
 146 fluxes² at a given latitude ϕ are first binned according to the value of C . One
 147 thus obtains a 2d field $M(\phi, C)$ which contains the accumulated mass fluxes
 148 advecting tracer value between C and $C+dC$ at latitude ϕ . The streamfunction
 149 in the C -coordinate is then computed as:

$$150 \quad \Psi(\phi, C) = - \int_{C_{min}}^C M(\phi, C) dC, \quad (1)$$

151 where C_{min} is the minimum value of the tracer. The unit of Ψ is the Sver-
 152 drup which is equal to $10^6 \text{ m}^3 \text{ s}^{-1}$ in the ocean and, as in Czaja and Marshall
 153 (2006), to 10^9 kg s^{-1} in the atmosphere (equivalent to the mass transport of an
 154 oceanic Sv, $\sim 10^3 \text{ kg m}^{-3} \times 10^6 \text{ m}^3 \text{ s}^{-1}$). The integral of the tracer-coordinate
 155 streamfunction over the full tracer range is equal to the net *advective* merid-
 156 ional transport of the tracer. [In our ocean model, the total transport also has
 157 a small diffusive contribution due to the horizontal component of the isopycnal
 158 diffusion.] In the following, we denote the C -coordinate overturning by $\Psi(C)$.

159 In the atmosphere, the DSE and LH are given by $C_{pa}T + gz$ and L_vq
 160 where T is the absolute temperature, C_{pa} the specific heat capacity at constant
 161 pressure, g the gravitational acceleration, z the height, L_v the latent heat
 162 of vaporization and q the specific humidity. Their sum is the moist static
 163 energy (MSE). The overturnings are computed from 5 years of daily snap-shots

² Mass fluxes in the atmosphere and volume fluxes in the (Boussinesq) ocean

164 (atmosphere) and 5-day averages (ocean). Note finally that $\Psi(S)$ includes both
165 Eulerian and (parameterized) eddy-induced transports.

166 **3 The coupled hydrological cycle**

167 **3.1 The atmospheric branch**

168 In all of our aqua-planet configurations $\Psi(MSE)$ is broadly similar in shape
169 and magnitude, and similar to those previously discussed in the literature from
170 both observations and atmospheric GCMs — e.g. Karoly et al (1997), Held and
171 Schneider (1999), Czaja and Marshall (2006), and Pauluis et al (2008). More-
172 over, the decomposition of $\Psi(MSE)$ in DSE and LH/ q components closely
173 resembles that seen in the ERA-Interim re-analysis (see Döös and Nilsson,
174 2011). Here we focus on the atmospheric contribution to the hydrological cy-
175 cle as encapsulated in $\Psi(q)$.

176 As can be seen in Fig. 3 (top), $\Psi(q)$ comprises two counter-rotating cells
177 in each hemisphere. Note that high q values correspond to the bottom of
178 the troposphere. The dominant (~ 180 Sv) mid-to-high latitude cells represent
179 wet (dry) air parcels moving poleward (equatorward), and thus are associated
180 with poleward moisture/LH transports from 20-30° into the high-latitudes.
181 These cells largely result from synoptic scale eddies (Döös and Nilsson, 2011).
182 Poleward of 40°, the (surface) poleward branch of $\Psi(q)$ (typically between
183 ~ 1000 -650 mb) is tilted upward, showing that air parcels gradually dry out as
184 they move toward colder temperature (consistent with the Clausius-Clapeyron

185 relationship). These streamlines are more tilted than the mean surface q (thick
 186 solid). This suggests that air parcels are lifted off the ground and thus expe-
 187 rience a more rapid cooling and drying than if they were moving along the
 188 surface. In our GCM, drying is mainly achieved through large-scale conden-
 189 sation in synoptic-scale weather systems. The return flow is nearly horizontal
 190 with dry ($\leq 2 \text{ g kg}^{-1}$) air parcels flowing equatorward in the upper tropo-
 191 sphere. The mid-to-high latitude cell is closed between 20 and 30° where air
 192 parcels “recharge” with moisture (downward arrow in $\Psi(q)$, Fig. 3 top). This
 193 moistening occurs through turbulent interaction with the boundary layer at
 194 latitudes where evaporation dominates over precipitation. Warm/moist out-
 195 breaks from lower latitudes play a primary role with poleward flow occurring
 196 at higher moisture than the mean surface value. In the tropics (20°S - 20°N),
 197 $\Psi(q)$ comprises 2 weak cells converging moisture on the Equator. These are
 198 mainly due to the time-mean Hadley circulation, with moist air flowing at low
 199 levels toward the Equator and dryer air flowing poleward aloft.

200 The atmospheric FWT scales as $\Delta q \Psi(q)$ with Δq the moisture change
 201 between the poleward and equatorward branch (often $\Delta q \sim q_s$, the mean
 202 surface specific humidity). The pattern of $\Psi(q)$ evidently reflects the large-
 203 scale E-P pattern (Fig. 3, middle). The fluxes of moisture into high latitudes
 204 and the deep tropics are matched by net precipitation while the moistening
 205 (downward branch of $\Psi(q)$) corresponds to net evaporation from the ocean
 206 surface.

3.2 The oceanic branch

The streamfunctions in S -coordinates for our coupled simulations are shown in Figs. 3 (bottom) and 4. At the scaling level, $\Psi(S)$ is related to the ocean FWT, F_w , through $S_o F_w \simeq \Psi(S) \Delta S$ where ΔS is the salinity difference between the northward and southward flowing branch of the streamfunction. This is the freshwater analog of the relation between the (potential) temperature-coordinate streamfunction $\Psi(T)$ and the Ocean Heat Transport H : $H \sim \Psi(T) \Delta T$ (e.g. Czaja and Marshall, 2006). We now summarize key properties of our solutions (Figs. 3 bottom and 4)

The S -space circulation, $\Psi(S)$, has two counter-rotating cells in each hemisphere. A narrow cell is found at high salinity between 0 and 30°, transporting FW poleward from the deep tropics into the subtropics. In Aqua, where gyres are absent, these cells result from Ekman-driven subtropical overturning cells. They capture the poleward Ekman flow at the surface becoming saltier and the associated interior return flow. These cells are rather similar across all configurations, suggesting that they are mainly due to the vertical component of the wind-driven circulation even in the presence of gyral circulations.

The broader cell spans a large salinity and latitudinal range in each hemisphere and take on various forms and magnitude. Each transports FW from the high-latitudes into the subtropics. In Aqua, they are rather weak, especially at high-latitudes. This is the salt equivalent of the vanishing Deacon cell familiar in density coordinates (Döös and Webb, 1994), resulting from the cancellation between the wind and eddy-driven circulations. In Ridge, the presence of gyres

230 is evident, notably the subpolar gyre near 50° . Deep water formation is seen
231 in the northernmost part of the cell ($\sim 60^\circ$) associated with a small salinity
232 gradient and a horizontal (isohaline) equatorward flow at 34.5 psu.

233 To first order, Drake combines the SH of Aqua and the NH of Ridge al-
234 though the asymmetry of the climate results in an interhemispheric cell in
235 which northern deep waters are carried into the SH. In S -coordinates, the
236 equatorward flow of deep water (~ 20 - 30 Sv confined within in a narrow range
237 of salinity near 34.5 psu) manifests itself as an intense “horizontal” flow and
238 a sharp transition in the streamfunction. Note the contrast between the large
239 circulation at 60 - 70° N acting on a small salt contrast and the vanishingly small
240 circulation near 60 - 70° S acting on a very large gradient. The Double-Drake
241 set-up is characterized by the split of the northern clockwise cell into two cells,
242 one for each basin. The saltier one, found in the small basin, is associated with
243 deep water formation in this basin while the fresher one is dominated by con-
244 tributions from wind-driven circulation in the large basin. This latter cell is
245 more reminiscent of the form found in the zonally re-entrant southern ocean.
246 As discussed elsewhere (Ferreira et al, 2015), the structure seen in Double-
247 Drake is similar to that found in ocean state estimates (e.g. the new ECCOv4
248 ocean state estimate; Forget et al, 2015) and is also consistent with inferences
249 made from (sparse) hydrographic sections (see Talley, 2008).

3.3 The ocean-atmosphere symmetry

The fact that both $\Psi(S)$ and $\Psi(q)$ are tightly linked to the E-P pattern results in the symmetry clearly evident in Fig. 3: poleward atmospheric moisture transport (counterclockwise cell) is associated with net precipitation on its poleward flank and net evaporation on its equatorward flank, and is matched with a clockwise salinity cell which transports freshwater equatorwards (and vice-versa).

The ocean-atmosphere symmetry is striking in Aqua but holds even in more complex geometries. The pattern of $\Psi(q)$ hardly changes across all simulations although its magnitude varies by about $\pm 20\%$ (peak values are in the range 185 ± 35 Sv, not shown). The atmospheric FWT however varies by less than $\pm 7\%$ (typically 1.6 ± 0.1 Sv). Because the atmospheric FWT is mostly eddy-driven, larger $\Psi(q)$ are associated with larger equator-to-pole temperature gradients and colder/drier mid-to-high latitudes. That is, variations of $\Psi(q)$ are partially compensated by variations of q_s resulting in the FWT, which scales as $\Psi(q) \times q_s$, being relatively constant. On the ocean side, as seen previously in Figs. 3 and 4, $\Psi(S)$ for Drake, Ridge, and Double-Drake all exhibit 4 cells approximately sitting below the 4 atmospheric $\Psi(q)$ cells.

Our coupled simulations exhibit a wide range of climates and circulations: North-South symmetric states (Aqua and Ridge), a zonally re-entrant ocean with sea-ice (Drake-Double and Drake) or without sea-ice (Aqua), North-South asymmetry, and multiple basins with ocean gyres and deep connection. As such the ocean circulation and hence detailed FWT pathways, take on strikingly

273 different forms. This is most obvious at high-latitudes where the FWT falls
274 into two broad categories: a strong circulation $\Psi(S)$ with a small ΔS or a weak
275 $\Psi(S)$ with a large ΔS . The first regime is characteristic of zonally bounded
276 convective basins while the second is typical of zonally re-entrant wind and
277 eddy driven oceans. However, both regimes achieve the same meridional FWT.
278 This is illustrated in Fig. 5 where ΔS is plotted against $\Psi(S)$ at 60° for all
279 the configurations (see caption for details). Both $\Psi(S)$ and ΔS vary by nearly
280 one order of magnitude and yet the product $\Psi(S)\Delta S$ hardly varies at all. The
281 thin line plotted in Fig. 5 maps out the $\Psi(S)$ and ΔS whose product is exactly
282 constant and equal to the FWT demanded by the atmosphere at 60°N (0.25 Sv
283 here).

284 As noted earlier, the atmospheric and oceanic FWT and associated E-P
285 pattern remains nearly unchanged across our coupled solutions (Fig. 2). It
286 is clear that ocean dynamics has little influence on the atmospheric FW/LH
287 transport. The LH transport is primarily set by atmospheric dynamics: evap-
288 oration is large at the edge of the Hadley cell which transports moisture
289 equatorward while synoptic eddies developing in the mid-latitude baroclinic
290 zone transport moisture poleward where it is rained out following Clausius-
291 Clapeyron. The ocean responds passively to the atmospheric E-P pattern
292 transporting what is demanded of it and takes on different forms depend-
293 ing on geometrical constraints and the climate state. There is little to suggest
294 that the “coupled LH mode” is anything but a one way relationship.

295 Before going on to our conclusions, we discuss a number of arguments that
296 might point to a more active oceanic FWT in the global energy cycle that
297 have been suggested to us:

- 298 – Oceans supply the heat required for evaporation. Our coupled simulations
299 span a wide spectrum of climates, with polar sea surface temperatures
300 varying from 10°C to freezing point (with sea ice cover). Similarly, the
301 meridional Ocean Heat Transport varies greatly between configurations
302 (for example from 0 to 1 PW at 50°N/S, see Ferreira et al (2010)). Despite
303 these very large ranges in available heat and supply by the ocean, the at-
304 mospheric LH/moisture transport varies little between configurations (as
305 implied by the similarity of the E-P patterns, Fig. 2, top left). It is worth
306 emphasizing that the evaporation patterns are very similar across the con-
307 figurations. A notable exception is where sea ice is present/absent. In this
308 case, differences in evaporation locally peak at 0.8 mm/day as evapora-
309 tion is severely limited by the cold temperatures and the capping effect of
310 sea ice. However, this effect is largely compensated by a reduction in pre-
311 cipitation and the E-P change (and thus the moisture transport) remains
312 relatively small. Even in this favorable limit where the ocean has a large
313 impact on E, negative feedbacks strongly limit its ability to influence the
314 atmospheric moisture/LH transport. This suggests that the heat supplied
315 by the ocean is not a critical factor in controlling the atmospheric LH
316 transport.

-
- 317 – Ocean processes such as salt barriers regulate the moisture flux to the at-
318 mosphere. Masson et al (2005) suggested that salt barriers (formed by pre-
319 cipitation) could have a strong impact on SSTs and precipitation, pointing
320 to an oceanic feedback on atmospheric moisture flux. These effects, how-
321 ever, appear to be very localized to the Equatorial region and unlikely to
322 have a large scale impact.
- 323 – Effects of the Goldsbrough circulation. The Goldsbrough circulation is
324 the barotropic ocean circulation induced by the surface mass flux (E-
325 P). Huang and Schmitt (1993) estimate the magnitude of this circulation
326 in the range 0.5-1.5 Sv. Assuming that this circulation acts on a East-
327 West temperature gradient ΔT of about 2°C , its meridional transport is
328 $\rho_o C_p \Psi_{Gold} \Delta T \sim 0.012\text{PW}$, a very small number unlikely to have a sizeable
329 climatic impact. It is worth emphasizing that the ocean component of our
330 coupled model uses real freshwater boundary conditions and thus includes
331 the physics associated with the Goldsbrough circulation.
- 332 – The coupled simulations are rather idealized, notably in their representa-
333 tion of the deep overturning circulation and internal mixing. Moreover, our
334 ocean model employs a constant vertical mixing coefficient. Observations
335 suggest that abyssal mixing varies greatly in space (i.e. Polzin et al, 1997)
336 although thermocline values appear uniformly low (Ledwell et al, 1993,
337 2011). The dependance of the FWT on mixing is unclear. To test the possi-
338 ble sensitivity of the FWT to mixing, we carried out a Double-Drake exper-
339 iment with increased diapycnal mixing at depth following Bryan and Lewis

(1979) (an arctangent profile with diffusivities increasing from 3×10^{-5} in the thermocline to 10^{-4} m s^{-2} at the bottom). Changes to FWT are very small, less than 0.04Sv. Ferrari and Ferreira (2011) showed that abyssal mixing had a small impact on the ocean heat transport although it did change the strength of the deep overturning cells. This is because the deep temperature gradients over which these cells act are weak. In respect of the FWT, the salinity gradients in the ocean compensate for changes in the circulation to ensure that the ocean FWT balances the pattern of E-P. Nonetheless the idealized nature of our coupled simulations is a caveat and warrants further investigations with more complex models.

4 Conclusions

We have explored the dynamics of FWT in the coupled ocean-atmosphere system using a series of idealized coupled simulations. To this end, we introduce streamfunctions in salt- and specific humidity-coordinates. Both present the dynamics of FWT in their respective realms.

The symmetry of the ocean and atmosphere FWT is clearly revealed in the symmetry of $\Psi(q)$ and $\Psi(S)$, encapsulating the transformation and exchange of freshwater in the coupled system. This is why the FWT and the atmospheric LH transport are sometimes described as “coupled ocean-atmosphere modes”. However $\Psi(S)$ reveals that fundamentally different modes of ocean FWT dynamics can exist under very similar E-P conditions. Two limit cases were identified: 1) a large circulation/small salt stratification mode typical of

362 regions of deep water formation and 2) a weak circulation/large stratification
363 mode found in zonally re-entrant regions. The ocean FWT dynamics does not
364 impact the atmospheric moisture transport: the ocean FWT is essentially pas-
365 sive in this “coupled mode”. Instead, the necessary ocean FWT is set to first
366 order by atmospheric dynamics. The ocean circulation and salinity stratifica-
367 tion adjust to this imposed boundary condition in different ways depending
368 on the geometrical constraints. In other words, the atmospheric freshwater is
369 returned “for free” with the ocean adjusting its $\Psi(S)$ and ΔS to match F_w :
370 slow ocean circulation leading to a large salinity contrast and vice-versa.

371 Our conclusions will not come as a surprise to atmospheric scientists. The
372 term “coupled mode” used to describe the atmospheric LH transport appears
373 in the oceanographic literature. We argue that this term puts an undeserved
374 emphasis on the ocean in a phenomenon which is essentially the result of
375 dynamics internal to the atmosphere. It is worth underscoring that our con-
376 clusions are limited to the steady state case. It is unclear whether the ocean
377 FWT takes a more active role in a transient climate change in which, for exam-
378 ple, the climate system is subject to a forcing perturbation. Another limitation
379 is that our simulations do not cover the full range of climate states suggested
380 by the paleoclimate record, which shows that Earth has experienced Snowball
381 and hothouse climates. Although it seems unlikely that the ocean would take
382 a more active role in the warm climate limit, this may not be true in very cold
383 climates with extensive sea ice cover.

384 Finally, we would like to emphasize that our conclusions do not imply that
385 there is little interest in studying the oceanic FWT and salinity distribution.
386 In fact, the passive nature of the ocean in this respect makes it a particular
387 efficient “tape recorder” of changes in the hydrological cycle.

388 **Acknowledgements** The authors would like to thank Richard Greatbatch for advice and
389 encouragements over the years. His openness, wide interest and enthusiasm for all matters
390 Atmosphere and Ocean is an example to all of us. We thank the Physical Oceanography
391 program of NSF.

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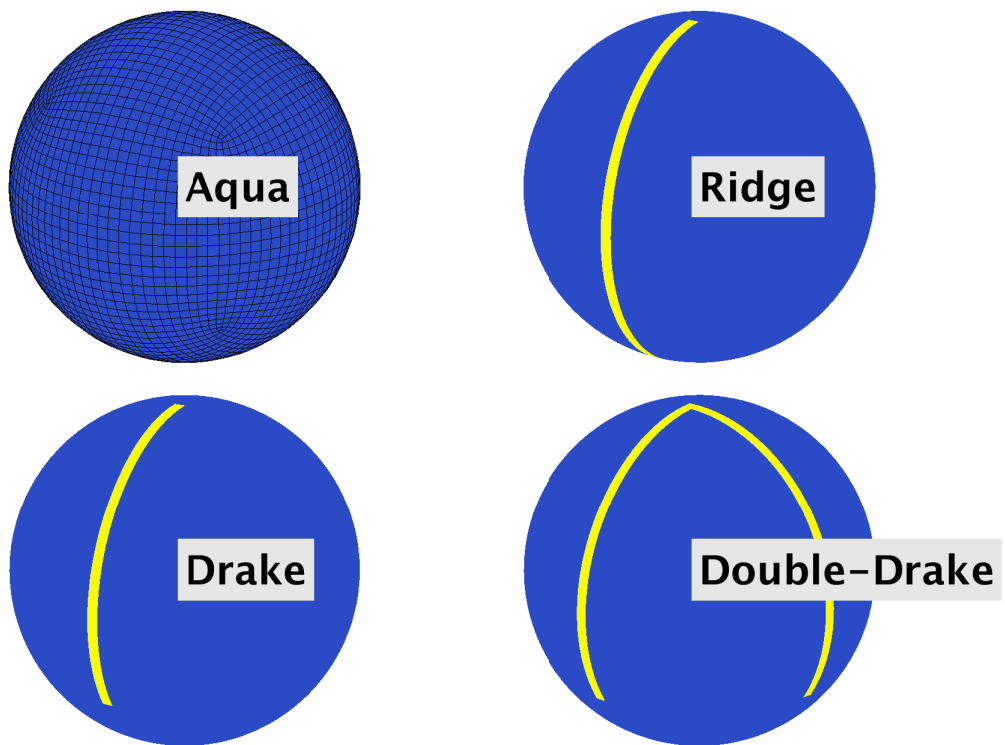


Fig. 1 The four continental configurations used in the present study. Yellow shading denotes the continental barriers extending from the flat bottomed ocean at 3000 m depth to the ocean surface. The cube-sphere mesh is indicated in the pure Aquaplanet configuration (top left).

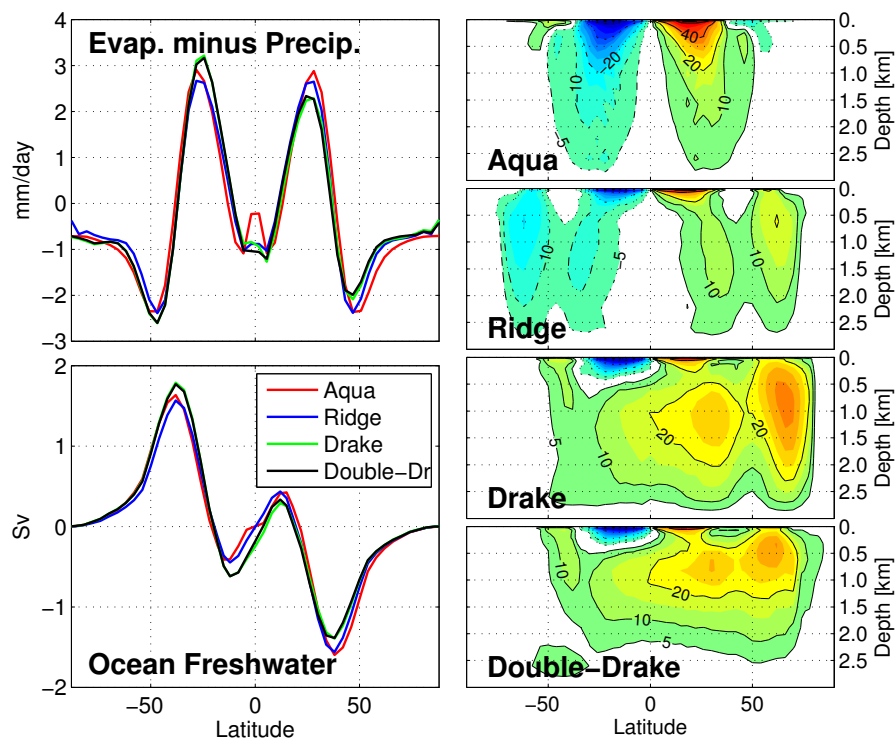


Fig. 2 Left: (top) Zonal- and time-average of Evaporation minus Precipitation (in mm/day) and (bottom) time-average ocean FWT (in Sv) for Aqua, Ridge, Drake and Double-Drake. Right: Residual-mean MOC (in Sv), the sum of the Eulerian and (parameterized) eddy overturnings. Clockwise and counterclockwise circulations are denoted by red and blue shadings, respectively.

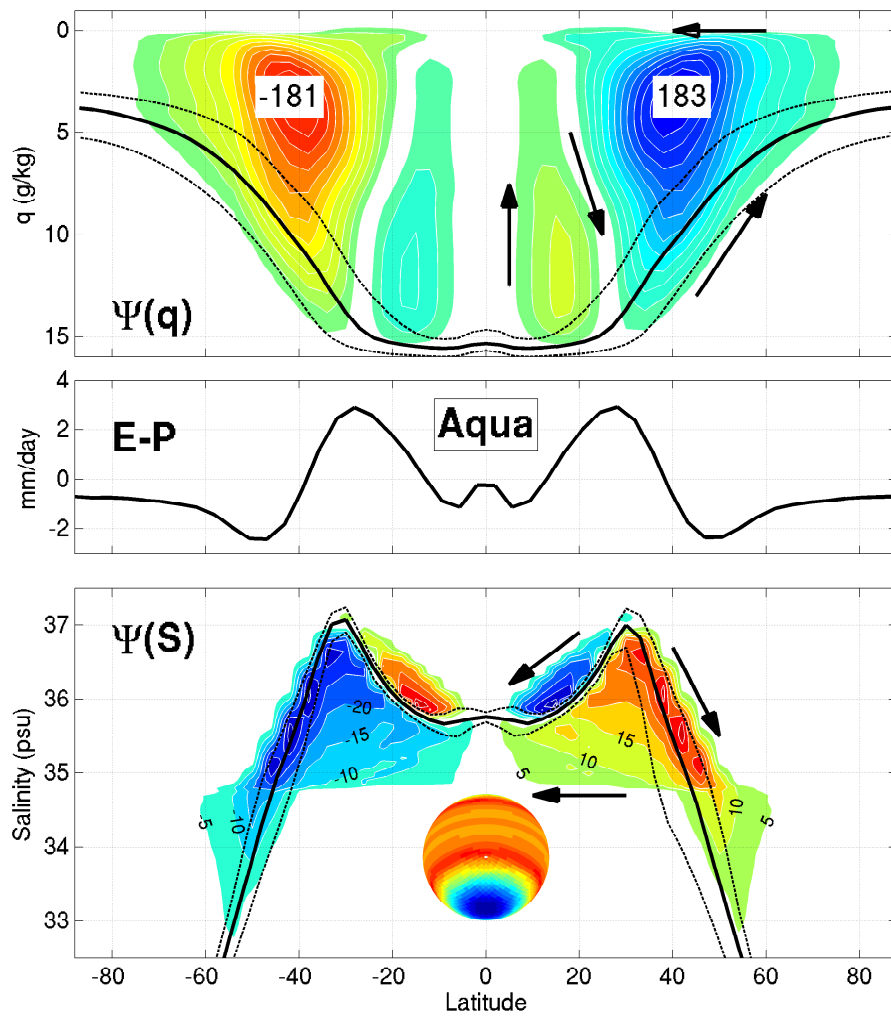


Fig. 3 The coupled hydrological cycle in Aqua: (top) $\Psi(q)$ for the atmosphere, (middle) E-P (mm/day) and (bottom) $\Psi(S)$ for the ocean. Clockwise and counter-clockwise circulations are shown in red and blue shadings, respectively. The median value (thick solid) and the 90% and 10% percentiles (dashed black) of the surface specific humidity (top) and sea surface salinity (bottom) are plotted. The SSS distribution of Aqua is shown in the bottom panel.

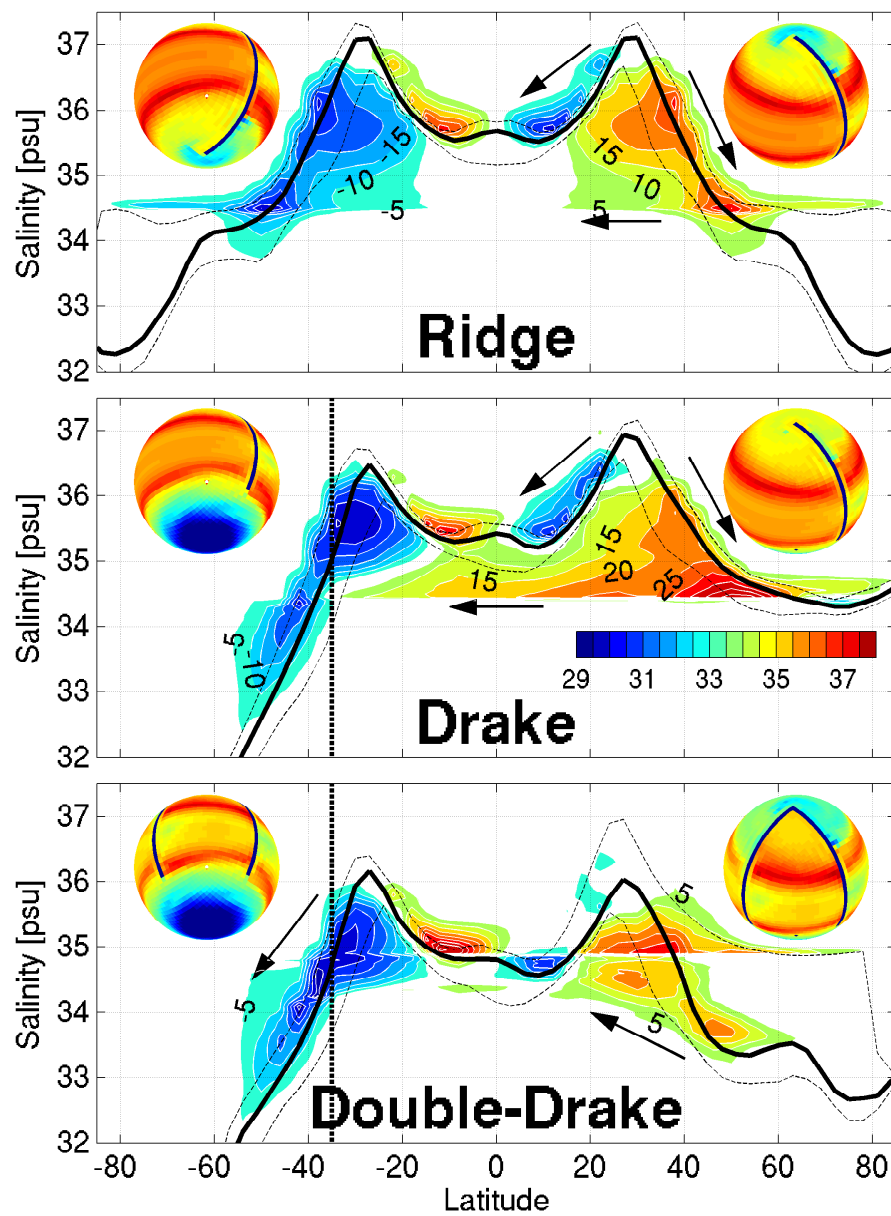


Fig. 4 $\Psi(S)$ for (top) Ridge, (middle) Drake and (bottom) Drake-Double (in Sv). See bottom panel of Fig. 3 for details. The thick dashed line in the middle and bottom panel indicates the southern limit of the land barrier (35°S). Globes in each panel show the annual mean SSS distributions (the associated colorbar is found in the middle panel).

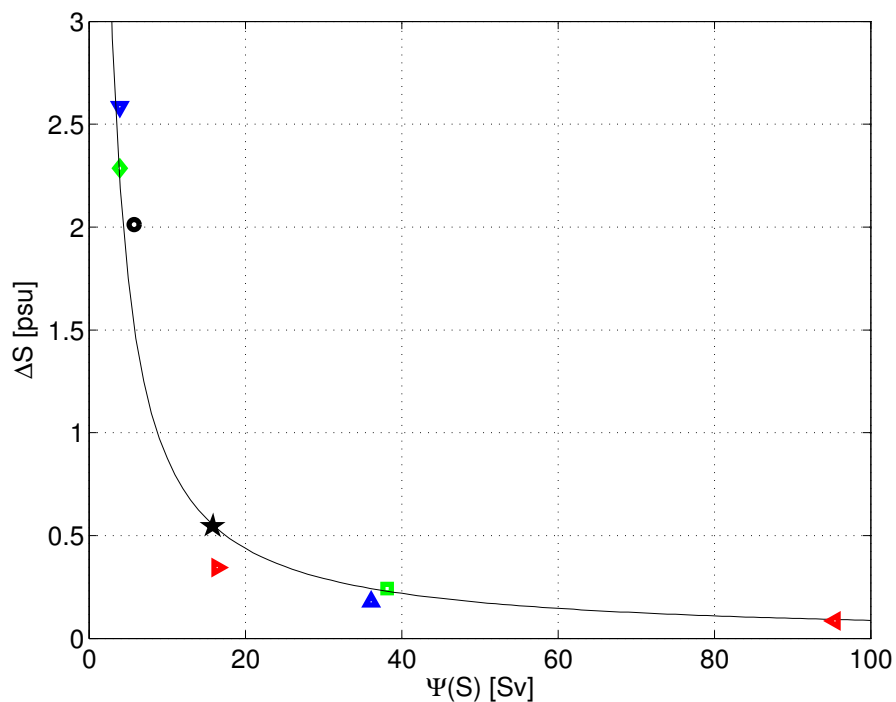


Fig. 5 Plot of the ΔS (psu) as a function of magnitude of $\Psi(S)$ (Sv) at 60° of latitude for Aqua (black circle), Ridge (black star), Northern (green square) and Southern (green diamond) hemispheres of Drake, the Northern (blue triangle up) and Southern (blue triangle down) hemispheres of Double-Drake, the Northern Large (red triangle right) and Small (red triangle left) basins of Double-Drake. The magnitude of $\Psi(S)$ is the sum of all poleward volume transports while ΔS is the averaged salinity of poleward flows minus the averaged salinity of equatorward flows. To account for the width of the Large (270°) and Small (90°) basins of Double-Drake, their Ψ are rescaled by factors $4/3$ and 4 , respectively. The thin solid line is a plot of $\Psi(S)\Delta S = F_w S_o$ a constant, given by $S_o = 35$ psu and $F_w = 0.25$ Sv, a typical value of FWT at 60°N in all configurations (see Fig. 2, bottom left).