GEOPHYSICAL RESEARCH LETTERS, VOL. 31, LXXXXX, doi:10.1029/2004GL020197, 2004

Restoring and flux adjustment in simulating variability of an idealized ocean

Harper Simmons[1,2] and Igor V. Polyakov[1]

Received 7 April 2004; accepted 8 July 2004; published XX Month 2004.

[1] Multi-century model runs are used to investigate the impact of restoring and flux adjustment on the variability of an idealized Atlantic Ocean forced by net atmospheric heat flux possessing decadal and multi-decadal variability. Restoring suppresses simulated modes of variability, causes phase shifts, and modifies nonlinear relations in the model. Flux adjustment has little effect on the water temperature variability, however it suppresses low-frequency variability of the meridional overturning circulation and causes a phase shift of multi-decadal mode of the meridional heat transport. An important effect of flux adjustment is that it may misrepresent physical mechanisms substituting, for example, dynamically-driven meridional heat transport by equivalent amount of heat supplied locally, though surface heat fluxes. We conclude that restoring provides a poor framework for simulation of climate variability. Flux adjustment is less damaging, however, it modulates internal modes of variability in ways not fully understood. *INDEX TERMS:* 3210 Mathematical Geophysics: Modeling; 3230 Mathematical Geophysics: Numerical solutions; 4255 Oceanography: General: Numerical modeling. **Citation:** Simmons, H., and I. V. Polyakov (2004), Restoring and flux adjustment in simulating variability of an idealized ocean, *Geophys. Res. Lett.*, *31*, LXXXXX, doi:10.1029/2004GL020197.

1. Introduction

[2] Models play substantial role in assessment of global climate change and variability, and understanding of their limitations is of key importance to the general community. Cumulative effects of model errors may result in drift of a numerical solution and may dominate simulations of climate trends and variability. Artificial terms added to the equation for scalar quantities (like water temperature and salinity) are often employed to drag the numerical solution toward observations. Restoring constrains temperature and/or salinity towards climatological values over some time-scale, whereas flux adjustment balances surface fluxes at the ocean-atmosphere interface to maintain integral quantities of scalar variables. The latter seems to have less restrictions on variability of the model parameters. These two methods are commonly used in global and regional climate simulations. For example, four out of six models used in Arctic Ocean model inter-comparison use restoring [*Steele et al.*, 2001]. Using an Arctic Ocean model, *Zhang et al.* [1998] demonstrated that restoring may be a useful tool in simulating mean state and short-term variations in the Arctic Ocean. Additionally, retrospective analysis of model simulations from the past where use of flux adjustment was a common practice requires a clear understanding of possible limitations of the method. There are suggestions that flux adjustment may suppress variability in climate models [*Pierce et al.*, 1995]. However, *Duffy et al.* [2000], comparing variability of surface air temperature derived from 17 simulations with and without flux adjustment, argued that there is no evidence that flux adjustment suppresses variability. The goal of this study is to show potential impacts of restoring and flux adjustment on the simulated variability of an idealized ocean. A simplest possible model configuration (ocean-alone model with simplified model domain and forcing) allows us, via spectral analysis, to show clearly the differences of frequency variability in experiments with/without restoring and flux adjustment. However, in this paper we will not attempt to divine the causes for these differences evident in the spectra.

2. Model and Design of Numerical Experiments

[3] The model consists of the free surface MOM4.0 z-coordinate ocean model [*Griffies et al.*, 2003]. The domain (Figure 1) is an idealized channel, extending zonally from 300°E to 350°E and latitudinally from 70°S to 80°N, with 2 degrees meridional and 1.5 degrees latitudinal resolution. There are 18 vertical levels, with vertical spacing increasing from 5 m at the uppermost level to 560 m at depth. The basin has sloping walls along the boundaries. The minimum and maximum depths are 500 m and 3500 m. A meridional mid-basin 15°-wide sloping-wall ridge has a minimum depth of 1000 m. This domain configuration aims to imitate basic features of the Atlantic Ocean, but note that a Southern circumpolar channel is not present.

[4] The *K*-profile parameterization (KPP) mixed layer scheme of *Large et al.* [1994] is employed. An additional background vertical diffusivity of 0.05 cm^2/sec in the upper ocean transitions to 1.0 cm^2/sec background in the abyss [*Bryan and Lewis*, 1979]. The *Redi* [1982] neutral diffusion and *Gent and McWilliams* [1990] (GM) skew-diffusion were both set to 1×10^3 m^2/sec. Horizontal friction is a traditional isotropic friction with a grid-space dependent background viscosity.

[5] Potential temperature is our active tracer, initially set to be horizontally uniform with exponential stratification. One thousand years with no restoring or flux adjustment are used for the model spin-up. During this period the model is

---

[1] International Arctic Research Center, University of Alaska Fairbanks, Fairbanks, Alaska, USA.

[2] Also at Geophysical Fluid Dynamics Laboratory, NOAA, Princeton, New Jersey, USA.

0094-8276/04/2004GL020197$05.00

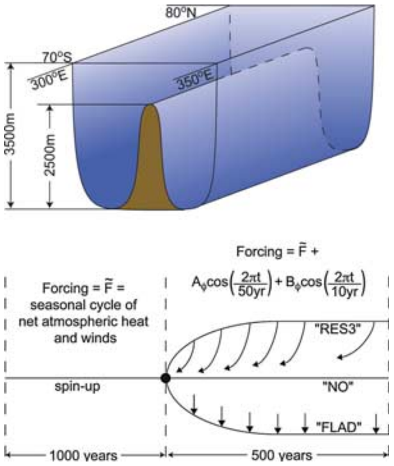

Figure 1. Schematic showing the model domain and forcing. Amplitudes of decadal and multi-decadal modes are shown by A_ϕ and B_ϕ, respectively. Arrows for "RES3" experiment show restoring of the simulated temperature to the initial condition whereas arrows for "FLAD" experiment show correction of the solution by spatially varying but constant in time heat flux.

forced with an annually repeating NCEP-based climatology of daily winds and net atmospheric heat flux (marked as $\widetilde{F}$ in Figure 1). The heat flux is computed from a zonal average of the Atlantic sector ocean. Heat flux has been adjusted so that there is zero net heat content change when integrated over the full model domain and over each year. Three 500-year experiments follow the spin-up using the final state of the spin-up as initial conditions. Note that by using an idealized domain we are not constrained by the fact that the "climatology" we obtained as a result of spin-up has substantial differences compared with realistic climatology. In the case of realistic basin configuration we would need to use some method (restoring) to drag the numerical solution toward observations. Otherwise, because of cumulative effects of model errors, the simulated climatology could have a little in common compared with realistic climatology.

[6] In these three experiments, forcing $\widetilde{F}$ used to spin-up the model is modulated by decadal and multi-decadal modes with amplitudes A_ϕ and B_ϕ, respectively. This addition imitates the major modes of variability found in the North Atlantic region [*Enfield et al.*, 2001; *Deser and Blackmon*, 1993]. The zonal amplitudes of multi-decadal mode are obtained averaging the NCEP data over 1960–1969 (negative phase of multi-decadal mode) and 1990–1999 (positive phase) and taking their difference. For the decadal mode, the amplitudes are obtained averaging over 1953, 1962, 1966, 1971, 1975, 1980, and 1986 (negative phase) and 1949, 1957, 1964, 1969, 1972, 1976, 1982, and 1990 (positive phase). Ten- and fifty-year period sine functions with the above amplitudes are used for time integration. The first experiment (denoted as "NO") uses no restoring or flux adjustment. In the second experiment (denoted as "RES3") surface temperature is restored to values from the end of the spin-up with a 3-month restoring constant. The last experiment (denoted as "FLAD") uses flux adjustment for the atmospheric heat based on time-averaged daily heat flux diagnosed from the "restoring" part of the RES3 surface net heat flux.

3. Simulated Variability With and Without Restoring and Flux Adjustment

[7] Mean distribution of the water temperature from the three experiments is very similar (not shown). Time series of the water temperature anomalies (WTA) averaged over the entire basin are shown in Figure 2 (top). As is expected, WTA from experiments "NO" and "FLAD" are practically identical. WTA variability at periods >10 years is strongly suppressed by restoring (experiment "RES3", green line in Figure 2, top). Spectral analysis of these three time series supports this conclusion showing that the effect of restoring is similar to that of high-pass filter (in Figure 3, top, green line ("RES3") is lower than blue ("NO") and red ("FLAD") lines at periods >20 years. Depending upon the choice of restoring time constant one can get stronger/weaker damping effects (for this purposes, a model run with a 6-months restoring constant was carried out, not shown). There is also a 5–6 year delay of the WTA response to atmospheric forcing in "RES3" experiment compared with WTA from "NO" and "FLAD" experiments (Figure 2).

[8] Intensity of the meridional overturning circulation (MOC) is a key climatic parameter, and next we show how restoring and flux adjustment affect its variability. Time series of anomalies of minimum MOC (analogous to North Atlantic Deep Water production) are shown in Figure 2 (middle). MOC variability is much more complex than the WTA variability, with stronger decadal mode in all time series and pronounced differences between MOC

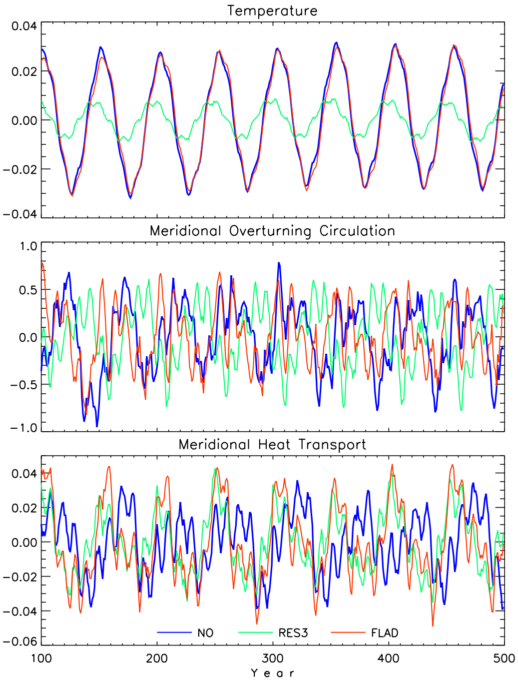

Figure 2. Time series of detrended basin-averaged water temperature (top, °C), meridional overturning circulation (middle, Sv = $10^6 m^3/s$), and meridional heat transport (bottom, PW) anomalies relative to the last 400 years of the model integration.

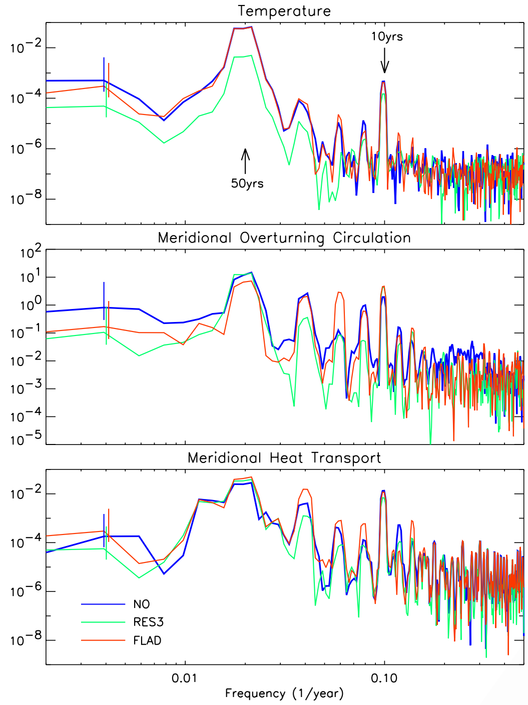

Figure 3. Power spectra of the water temperature (top), meridional overturning circulation (middle), and meridional heat transport (bottom). Vertical bars show 95% confidence intervals.

from "NO", "FLAD", and "RES3" experiments. Spectral analysis shows that restoring damps MOC variability at almost all frequencies (green line in Figure 3, middle which is lower relative to other curves). Flux adjustment is more selective, suppressing low frequencies only (>50 years). The MOC phase shift from the "FLAD" experiment is modest relative to "NO", with the multi-decadal mode from "RES3" experiment lagging those from "NO" and "FLAD" experiments (Figure 2, middle). MOC and meridional heat transport (MHT) are positively correlated at zero phase lag (not shown) for "NO" and "FLAD" experiments (correlation coefficient $R = 0.8$). Restoring on the other hand, alters this relationship, with MOC and MHT anti-correlated at zero phase lag ($R = -0.5$).

[9] Meridional heat transport associated with the MOC is an important element of the planetary heat balance [e.g., *Trenberth and Caron*, 2001]. Simulated time series of maximum MHT is shown in Figure 2, bottom. MHT is calculated using the zonally and vertically integrated heat transport including advective and GM contributions. Surprisingly, the MHT from "FLAD" and "RES3" (not "NO") experiments show coordinated set of changes with concerted envelopes of multi-decadal mode. These envelopes delay those from "NO" experiment by approximately 5 years. At basin-wide scale, restoring and flux adjustment do not suppress intensity of decadal and multi-decadal modes of variability as shown by spectral analysis in Figure 3, bottom, however, restoring damps variability at periods >50 years. Flux adjustment amplifies almost every frequency peak.

[10] Figure 4, top, shows anomalies of MHT and vertically-averaged circulation averaged for a negative phase of the MHT multi-decadal mode (years 435–460, see Figure 2, bottom). The anomalies are calculated relative to a mean over a complete 50-year cycle (435–485). Figure 4 also shows MHT and circulation anomaly differences between "RES3" and "NO" experiments (middle) and "FLAD" and "NO" experiments (bottom). Because of shift of phase between MHT in these experiments compared with "NO" experiment, MHC and circulation anomalies and means for "RES3"and "FLAD" experiments are calculated instead over model years 418–443 and 418–468. Figure 4, top, shows that the multi-decadal MHT and circulation variability is strong in the Northern Hemisphere with intensive variability of hemisphere-wide cyclonic gyre supplying the north-east part of the basin with heat (red color in the upper right corner of the upper panel). Both restoring (middle panel) and flux adjustment (bottom panel) partially suppress

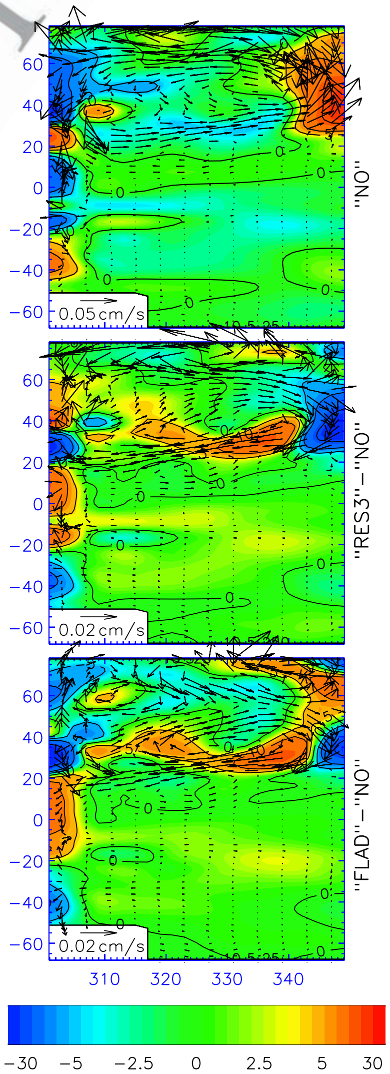

Figure 4. (Top) Meridional heat transport (MHT (TW), color) and vertically-averaged circulation anomalies ("NO" experiment) for the negative phase of the MHT multi-decadal variability. (middle and bottom) The MHT and circulation anomaly differences between "RES3" and "NO" experiments and "FLAD" and "NO" experiments, respectively (see text for details).

this variability. For example, multi-decadal variability of the idealized "Gulf Stream" (lower part of the cyclonic gyre) is weakened by restoring and flux adjustment with reduced MHT variations. Since the advective northeastward heat transport is suppressed in "FLAD" experiment, there should be other means by which the system maintains anappropriate level of low-frequency WTA in the idealized North Atlantic. Surface heat flux diagnosed from experiments with restoring and used for flux adjustment provides these means: this flux does not exceed 1–2 W/m^2, however, local values over the "Gulf Stream" area are as high as 10–15 W/m^2. This suggests that flux adjustment mis-represents physical mechanisms substituting, for example, dynamically-driven meridional heat transport by equivalent amount of heat supplied thermodynamically (i.e., locally, through surface heat fluxes). The goal of this analysis was not to reproduces realistic physical processes governing the Gulf Stream (among other reasons, our resolution is not adequate), but to point out that physical mechanisms which drive simulated variability are different in the three above cases.

[11] Decadal and multi-decadal forcing, through nonlinear interactions, generate additional oceanic modes of variability (overtones) at several periods within 10–50-year band and at longer periods (Figure 3). In this, each basic (forcing) mode with 10- and 50-year period, due to the model nonlinear terms, generates oscillations whose periods are linear combinations of the original periods. Many of these spectral peaks are suppressed by restoring. For example, in "NO" experiment, the model generates low-frequency variability of MOC at periods >50 years (clearly seen by tracing the negative tips of 50-year cycle envelopes shown by blue line in Figure 2, middle). MOC from "RES3" experiment (and also from "FLAD" experiment) does not show this behavior (Figure 2, middle). Surprisingly, in many cases flux adjustment leads to amplified overtones. Spectral analysis provides evidence that, for example, the MHT overtones are stronger in "FLAD" experiment than in "NO" experiment (Figure 3, bottom). The model also generates much stronger MOC peak at 0.06 year^{-1} frequency (period ≈17 years) in "FLAD" experiment (Figure 3, middle) which may explain differences between time series of the MOC from "NO" and "FLAD" experiments apparent in Figure 2, middle.

4. Conclusions

[12] Global and regional models often use restoring or flux adjustment to suppress drift of a numerical solution from a mean state. We used a multi-century model runs to investigate possible impacts of restoring and flux adjustment on simulated oceanic variability of an idealized ocean (roughly imitating the Atlantic Ocean). Our experiments show that restoring suppresses variability, causes lagging of phase, and misrepresents nonlinear relations in the model, suppressing overtones. Flux adjustment is less damaging for simulated variability. However, for some important climatic parameters flux adjustment distorts variability in a way similar to restoring. For example, it suppresses low-frequency variability of the meridional overturning circulation and causes a phase shift of multi-decadal mode of the meridional heat transport. Flux adjustment is also selective with regards to nonlinear effects, suppressing some overtones and amplifying others. An important negative effect of flux adjustment found in our simulations is that it may misrepresent physical mechanisms substituting, for example, dynamically-driven meridional heat transport by equivalent amount of heat supplied thermodynamically (i.e., through local surface heat fluxes). Our simple model suggests that restoring provides a poor framework for simulation of climate variability. Flux adjustment may be useful for simulation of some parameters, however there is a danger of suppressing or amplifying modes of variability, creating phase distortions and/or misrepresenting physical mechanisms hidden behind natural variability. Thus, it is important to recognize possible limitations of this approach when previous and future modeling simulations with flux adjustment are considered.

[13] **Acknowledgments.** We would like to thank U. Bhatt for help in data preparation and J. Moss for help with illustrations. Comments of J. Walsh and A. Weaver were most useful. This project was supported by the International Arctic Research Center, University of Alaska Fairbanks (IARC-NSF grant 0327664) and Geophysical Fluid Dynamics Laboratory, NOAA. IP thanks the Frontier Research System for Global Change for financial support.

References

Bryan, K., and L. J. Lewis (1979), A water mass model of the world ocean, *J. Geophys. Res.*, *84*, 2503–2517.

Deser, C., and M. Blackmon (1993), Surface climate variations over the North Atlantic Ocean during winter: 1900–89, *J. Clim.*, *6*, 1743–1753.

Duffy, P. B., J. Bell, C. Covey, L. Sloan, and CMIP investigators (2000), Effect of flux adjustment on temperature variability in climate models, *Geophys. Res. Lett.*, *27*(6), 763–766.

Enfield, D. B., A. M. Mestas-Nunez, and P. J. Trimble (2001), The Atlantic multidecadal oscillation and its relation to rainfall and river flows in the continental U.S., *Geophys. Res. Lett.*, *28*(10), 2077–2080.

Gent, P. R., and J. C. McWilliams (1990), Isopycnal mixing in ocean circulation models, *J. Phys. Oceanogr.*, *20*, 150–155.

Griffies, S. M., M. J. Harrison, R. C. Pacanowski, and A. Rosati (2003), A technical guide to MOM4, technical report, NOAA/Geophys. Fluid Dyn. Lab. Ocean Group, Princeton, N. J. (Available at http://www.gfdl.noaa.gov/fms/)

Large, W. G., J. C. McWilliams, and S. C. Doney (1994), Oceanic vertical mixing: A review and a model with a nonlocal boundary layer parameterization, *Rev. Geophys.*, *32*, 363–403.

Pierce, D. W., T. P. Barnett, and U. Mikolajewicz (1995), Competing roles of heat and freshwater fluxes in forcing thermohaline oscillations, *J. Phys. Oceanogr.*, *25*, 2046–2064.

Redi, M. H. (1982), Oceanic isopycnal mixing by coordinate rotation, *J. Phys. Oceanogr.*, *12*, 1154–1158.

Steele, M., W. Ermold, S. Hakkinen, D. Holland, G. Holloway, M. Karcher, F. Kauker, W. Maslowski, N. Steiner, and J. Zhang (2001), Adrift in the Beaufort Gyre: A model intercomparison, *Geophys. Res. Lett.*, *28*(15), 2935–2938.

Trenberth, K. E., and J. M. Caron (2001), Estimates of meridional atmospheric and ocean heat transports, *J. Clim.*, *31*, 3343–3443.

Zhang, J., W. D. Hibler III, M. Steele, and D. A. Rothrock (1998), Arctic ice-ocean modeling with and without climate restoring, *J. Phys. Oceanogr.*, *28*, 191–217.

H. Simmons and I. V. Polyakov, International Arctic Research Center, University of Alaska Fairbanks, P. O. Box 757220, Fairbanks, AK 99775-7220, USA. (igor@iarc.uaf.edu)