TECHNICAL NOTE FOR DCPP-Component C
II. Recommendations for ocean restoring and ensemble generation

This short note provides background and guidelines for restoring strategy to be applied in DCPP-Component C idealized and pacemaker experiments. It also specifies the methodology recommended to generate the ensemble simulations.

Several strategies exist to restore a coupled climate model to specific sea surface temperature (SST) conditions over a given region. None of them are perfect because all introduce energy and seawater density imbalances that could lead to spurious effects. Related perturbations can eventually propagate to the entire coupled system through artificial changes in air-sea interaction and/or oceanic circulation, which could highly alter the coupled model equilibrium and pollute the results. Regional restoring should thus be applied with great caution as in the proposed recommendations.

1-Methodology

The simplest method to constrain the sea surface temperature (SST) consists in adding a Newtonian damping term to the temperature equation at the first level of the ocean model SST_{MODEL} such as:

$$\frac{\partial SST_{MODEL}}{\partial t} = \ldots + \lambda(SST_{MODEL} - SST_{TARGET}) \qquad \textbf{(1)}$$

where λ is the inverse of the relaxation time scale τ_R, which controls the strength of the restoring for the modeled temperature towards SST_{TARGET}, which is the AMV pattern or the IPV pattern in the proposed DCPP-Component C idealized experiments. Such a method is very efficient to constraint the temperature but it introduces artificial sources of heat within the ocean that can lead to spurious perturbations in the coupled system as discussed in Section 3.

A more physical approach is based on a flux formulation that consists in restoring the SST through the introduction of a feedback term added to the non-solar total heat flux following Haney (1971) such as:

$$Q_{ns} = Q_{ns}^{o} + \frac{dQ}{dT}(SST_{MODEL} - SST_{TARGET}) \qquad \textbf{(2)}$$

$\frac{dQ}{dT} = \gamma_T$ is therefore a fixed negative feedback coefficient in $W.m^{-2}.K^{-1}$ that can be estimated from observed temperature/heat flux relationships. The restoring term expressed as a flux is thus acting over the entire mixed layer depth as opposed to (1) where only the first layer of the model ocean is constrained.

2-Recommendation

The ***flux formulation*** is recommended for both idealized and pacemaker DCPP-Component C experiments using a feedback coefficient γ_T ***equal to -40*** $\boldsymbol{W.m^{-2}.K^{-1}}$**.** This value is chosen based on the literature on observational studies (e.g. Frankignoul and Kestenare 2002) and modeling studies (e.g. Rahmstorf 1995, Servonnat et al 2015). This value of γ_T is equivalent to a restoring timescale τ_R of about 60 (12) days for a 50 (10)-meter mixed layer depth.

In the AMV and IPV idealized experiments, SST_{TARGET} correspond to the 12-month climatology estimated from the DECK piControl experiment over at least 500 years of integration, on top of which the anomalous AMV or IPV SST patterns are superimposed. Details on the definition of the AMV and IPV patterns can be found in the technical note I. The mask fields provided in the netcdf files give the spatial domain, over which the restoring should be applied.

Recommendation for the generation of the ensemble members is to use ***macro-perturbations*** following Hawkins et al (2015) nomenclature, which consists in taking different ocean states from the DECK piControl as initial conditions, instead of just only different atmospheric states (micro-perturbation) as done in many "traditional" studies. This ensures a better spread among the ensemble members. For the IPV (AMV) experiments, ocean states should span the distribution of the model AMV (IPV) as much as possible in order to estimate at best the relationship between the two oceans (e.g. the modelers should check that the different ocean states picked from piControl do not by chance all correspond to a maximum of the AMOC, but span at best the AMOC probability density function). In the idealized AMV and IPV experiments, all the external forcings are set to their piControl values.

In the pacemaker experiments, SST_{TARGET} correspond to the 12-month climatology estimated from the historical simulation, on top of which observed monthly SST anomalies from ERSSTv4 (Huang et al 2016) are superimposed. The reference period for computing the observed anomalies is 1950-2014. The use of macro-perturbation is also recommended to define the ensembles. External forcings in the pacemaker experiments evolve like in the historical simulations.

For both idealized and pacemaker experiments, ***the heat flux restoring term (hfcorr) defined as:***

$$hfcorr = \frac{dQ}{dT}(SST_{MODEL} - SST_{TARGET}) \qquad \textbf{(3)}$$

must be diagnosed at monthly timescale and provided to the community as a standard CMORized output. This is critical to evaluate the energy imbalance introduced into the coupled system by the protocol in order to correctly estimate the model response to the restored SST.

3-Justification about the recommended parameters

3.1-Strength of the restoring term

There is always a trade-off between keeping the magnitude of the SST anomalies as close as possible to the values it is restored to, and minimizing the perturbation that is introduced in the coupled model because of the restoring. One must find a set of restoring parameters that perturbs the least possible the equilibrium and the intrinsic physics of the coupled climate model. To illustrate the issues that can be encountered, an example is given below based on the CNRM-CM5 model (Voldoire et al 2012) for the Atlantic Pacemaker experiment over the 1986-2005 period.

In order to choose objectively the best value for the restoring strength, we tested several values of γ_T namely -960, -240, -120 and -40 W.m^{-2}.K^{-1} corresponding to a restoring timescale τ_R of 2, 10, 20 and 60 days for a 50-m mixed layer depth, respectively. We found that a too strong restoring can lead to spurious oceanic circulation and altered air-sea interactions, leading to a spin up of the model and a drift, that are detrimental. As an example, the Atlantic Meridional Overturning Circulation (AMOC) strength in CNRM-CM5 undergoes a significant increase in the 2-day relaxation case (which is close to AMIP condition over the restored domain) and it stabilizes at a mean value (+5 Sv on average) that is outside the range of the model equilibrium as assessed from historical ensemble simulations (Fig. 1a). The AMOC strengthening and the associated enhanced northward heat transport leads to a progressive depletion of the ocean heat content (Fig.1b) in the southern Tropical Atlantic, creating an artificial latitudinal dipole that progressively affects the entire ocean circulation and the atmosphere through coupling (ITCZ changes etc.).

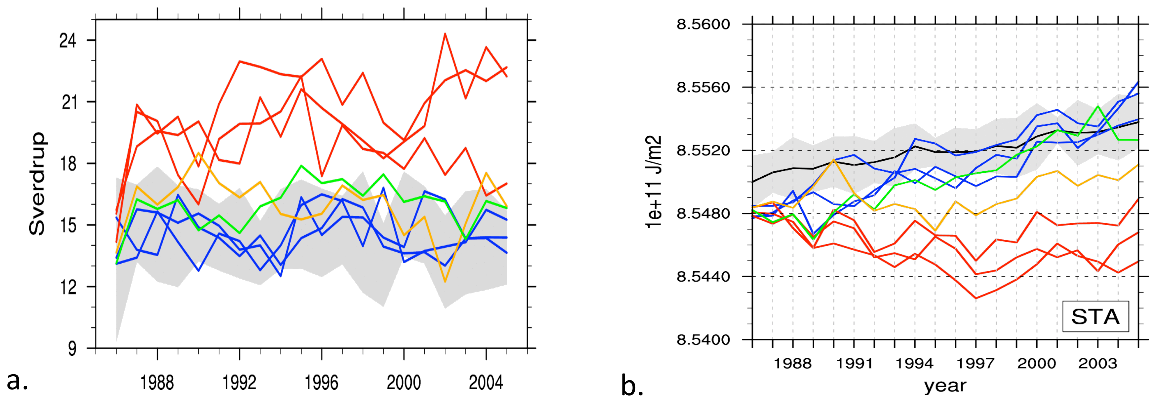

Figure 1: ***a.*** *AMOC time series for Atlantic pacemaker experiments in the CNRM-CM5 model using different restoring strengths: -960 (red), -240 (orange), -120 (green) and -40 W.m^{-2}.K^{-1} (blue). An ensemble of 3 members has been carried out for the -960 and -40 W.m^{-2}.K^{-1} cases, only one member for the others as these were just preliminary experiments to test the protocol. Gray shading stands for the AMOC Min-Max range of the CNRM-CM5 model as estimated from the 10-member CMIP5 historical simulations. The nudging protocol must not disrupt the model equilibrium while controlling the temporal evolution of the surface ocean. Therefore, the AMOC in the pacemaker experiments must be comprised within the gray shading.* ***b.*** *Same but for the ocean heat content over the so-called Tropical South Atlantic (TSA, 30°W-10°E, 20°S-0°) region. The black curve stands for the historical ensemble mean.*

The high frequency air-sea interactions are also significantly affected by a strong restoring as illustrated in Fig.2. At midlatitudes and in the subtropics, it is well known that the atmosphere has a dominant forcing role upon the surface ocean (Deser and Timlin 1998, among others). We recall here that for the idealized and the pacemaker experiments the goal is respectively to constrain the surface ocean mean state and the temporal evolution of its low-frequency change, while preserving at best the intrinsic high frequency ocean/atmosphere relationship. Alteration in the latter can be tracked in surface flux imbalance and because experiments are performed in coupled mode, this inevitably leads to spurious drift and model adjustment, which can *in fine* pollute the interpretation of the results.

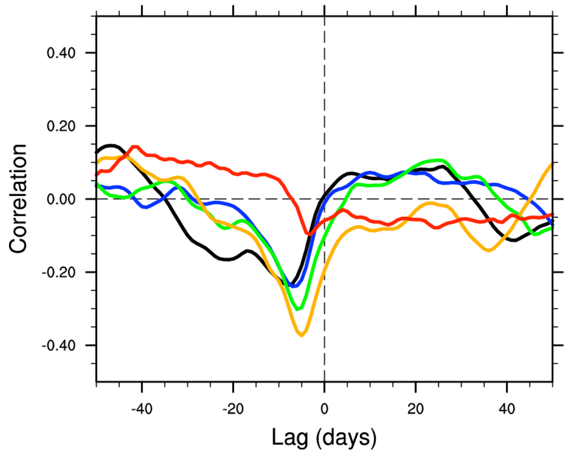

Figure 2: *Lead-lag correlation between daily anomalous mean sea level pressure and SST over the North Tropical Atlantic domain (TNA, 55°W-15°W, 5°N-25°N) for the summer (June-September) season in the Pacemaker Atlantic experiments shown in Fig.1. Positive lags mean that the ocean leads the atmosphere. The color code is the same as in Fig.1. Weak restoring (blue curve) fits best the historical equilibrium (black curve) whereas strong restoring leads to very weak and opposite sign correlation values.*

For those reasons, the recommendation for DCPP-Component C experiments is to use a weak restoring equal to -40 W.m^{-2}.K^{-1}. Note that the efficiency of the latter is dependent on the mixed layer depth: it is very strong in stratified shallow mixed ocean (equivalent to a relaxation timescale τ_R of ~10 days for a 10-m mixed layer depth) but almost inactive when deep ocean convection occurs (equivalent to ~3 years for a 1000-m mixed layer depth). Because of the strong seasonal cycle of the mixed layer depth and its regional signature, it turns out that the efficiency of the restoring feedback term is thus very much seasonal-dependent and region-dependent. One alternative following Ortega et al. (2016) would be to introduce a variable γ_T^{VAR} proportional to the mixed layer depth such as:

$$\gamma_T^{VAR} = \max\left(\gamma_T, \gamma_T \frac{h}{h_o}\right) \qquad (4)$$

where h is the prognostic mixed layer depth given by the model at each time step, h_o is the standard reference mixed layer depth (here 50m) associated with the restoring timescale τ_R and equal to 60 days when γ_T= -40 W.m^{-2}.K^{-1}. There is no strict recommendation from DCPP about this adjusted-restoring technique but modeling groups are encouraged to test it.

3.2- Avoiding drifts

Some model can be sensitive to the local alteration of the ocean density due to the additional flux term introduced by SST restoring. Because experiments are run in coupled mode, this could progressively alter the mean ocean circulation and mean thermodynamical balance. In that case, the recommendation is to use a relaxation term for SSS to maintain a zero density anomaly. By analogy to SST, the SSS restoring can be done through a flux formulation as follows:

$$EMP = EMP^o + \gamma_S \frac{(SSS_{MODEL} - SSS_{TARGET})}{SSS_{MODEL}}$$

The SSS restoring term γ_S can be viewed as a flux correction on freshwater fluxes; it has no physical meaning by contrast to γ_T and is just chosen to match the same relaxation timescale as in temperature for a given mixed layer depth and thus to maintain the density. In that case, SSS_{TARGET} stands for salinity values that counterbalance the density anomalies generated by the SST restoring.. Another alternative is to compute a salt correction flux at each time step of the model via the equation of state in order to maintain the density constant at the surface.

The role of the SSS restoring for neutral density is illustrated in the GFDL-CM2.1 model in Fig.3 for simulations where the model North Atlantic SSTs are restored to their climatology. By construction, these experiments allow to isolate the perturbations introduced by the restoring protocol (cf. Supplementary Material in Ruprich-Robert et al 2016). We show that without SSS restoring, the heat flux correction term (*hfcorr*) over the North Atlantic Subpolar Gyre acting to maintain the SST climatology is not equal to 0 but drifts significantly and reaches values as high as 15 $W.m^{-2}$ after 20 years of integration. It turns out that *hfcorr* tends to compensate a developing surface cooling associated with a weakening of the AMOC and a slackened deep-convection because of lower surface density (not shown). When SSS restoring is applied together with a weak SST relaxation, *hfcorr* does not introduce any significant artificial heat source into the model ocean and the surface temperature and large-scale circulation are well preserved (not shown).

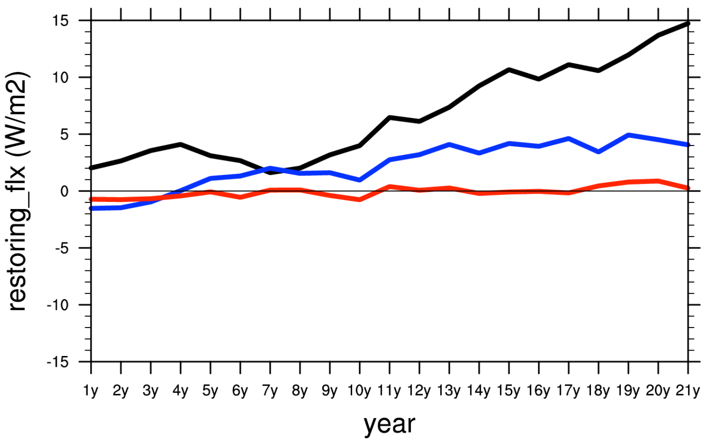

Figure 3: *20-yr time series of the restoring flux term (hfcorr) averaged over the North Atlantic Subpolar gyre in GFDL.CM2.1 simulations for which the model North Atlantic SSTs are restored to their own climatology (this experiment corresponds to the control Atlantic experiment in Table 1 of the GMD paper).* γ_T *is equal to -120* $W.m^{-2}.K^{-1}$ *without (black) and with (blue) neutral density control, and to -40* $W.m^{-2}.K^{-1}$ *with neutral density control in red. hfcorr should be as close possible to 0 to avoid the introduction of artificial heat source due to the restoring protocol.*

The necessity for SSS restoring is very much model dependent. For instance, tests have been done with the CNRM-CM5 and CESM1 models and those do not require such a term for neutral density. Therefore, it is recommended that each modeling group ***evaluates the drift of their coupled system and makes use of ad-hoc neutral density control whenever necessary***. Drifts can be evaluated from simple metrics such as the temporal evolution of the AMOC, of the surface temperature in critical regions such as in the subpolar gyre etc. and from the computation of the flux correction term that should be as close as possible to 0 when averaged over the full period for the pacemaker experiments.

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