OS52D-08

The Pacific Cold Tongue and ENSO: Sensitivity to the Meridional Wind Stress Climatology

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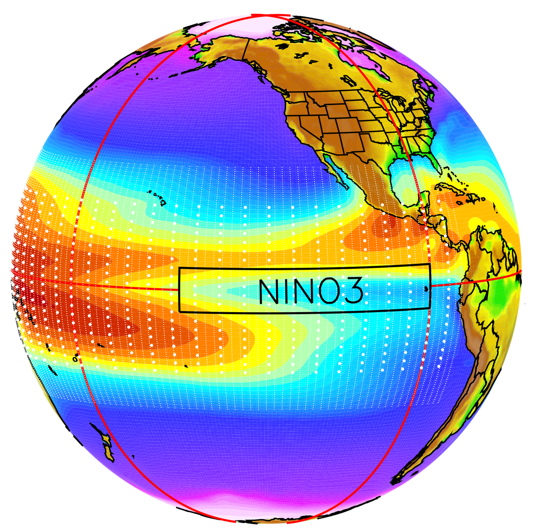
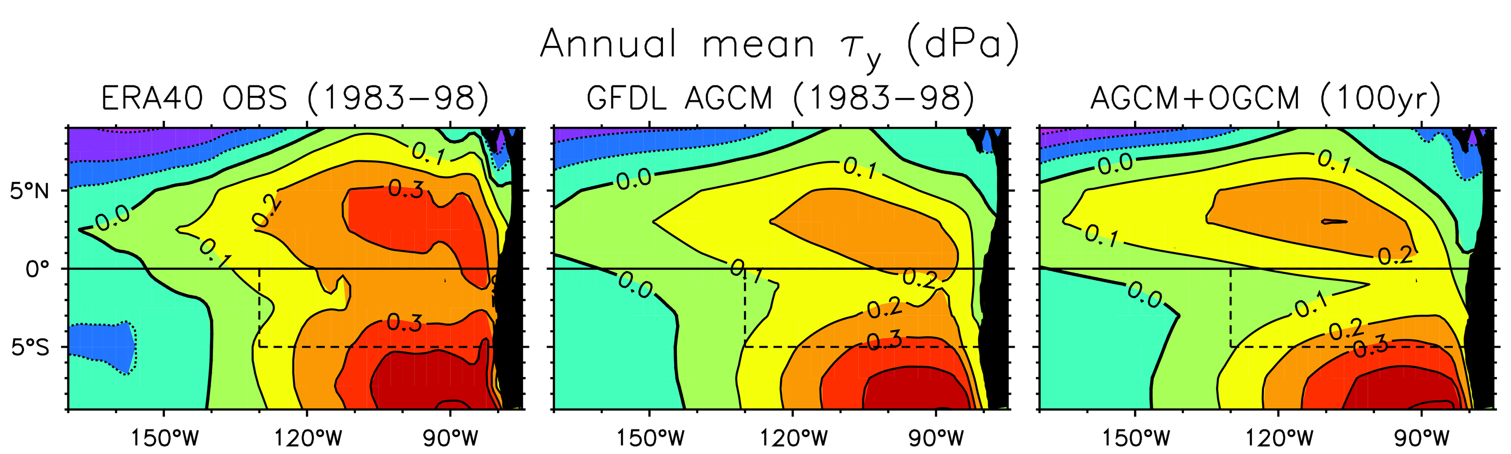
1. Introduction

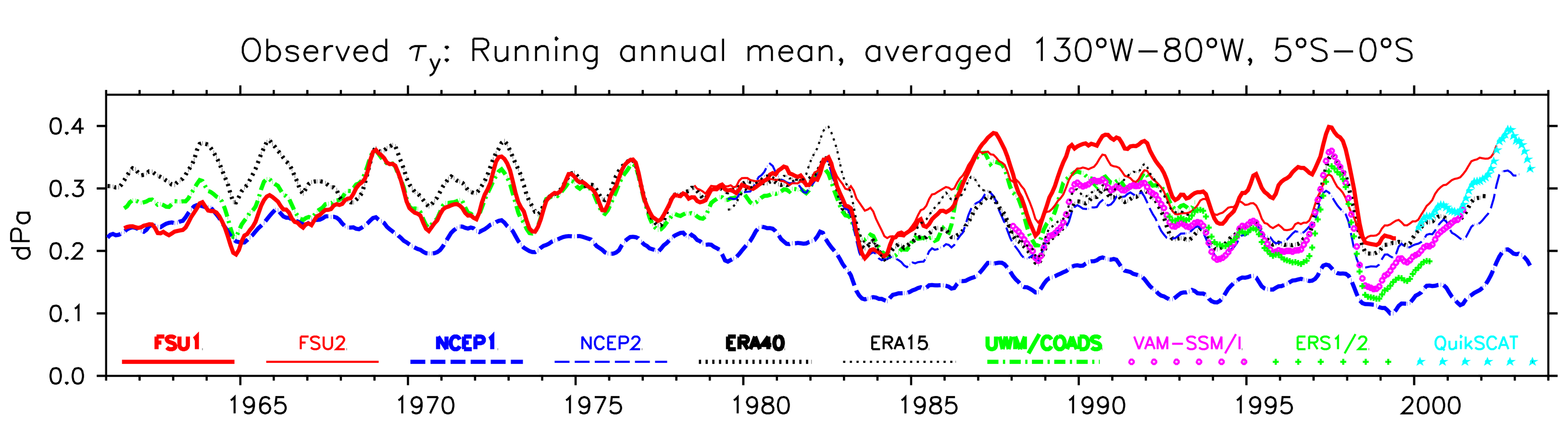


The eastern tropical Pacific displays a distinct meridional asymmetry, with cold sea surface temperatures (SSTs) south of the equator, warmer waters to the north, and an Intertropical Convergence Zone that is mostly north of the equator (Philander et al. 1996; Wang and Wang 1999). The asymmetry is linked to southerly surface winds over a broad span of the equatorial eastern Pacific. How do these cross-equatorial southerlies affect the climatological cold tongue and El Niño/Southern Oscillation (ENSO)?

2. Meridional Wind Stresses in Models and Observations



The asymmetry of the east Pacific depends on processes that are notoriously difficult to capture in general circulation models (GCMs). These include stratus low clouds south of the equator, atmospheric deep convection north of the equator, and oceanic upwelling & mixing in the equatorial zone. The depth of the equatorial thermocline, which depends on surface forcing throughout the Pacific, also controls the asymmetry. Most atmospheric GCMs produce a tropical climate that is too meridionally symmetric, with southerly winds that are too weak—and these biases tend to amplify upon coupling to ocean GCMs (above). In observational analyses (below), the meridional winds over the southeast equatorial Pacific vary from year to year and from decade to decade. Measurements in this region were sparse until very recently, so the actual history of these wind stresses remains uncertain (Wittenberg 2004).

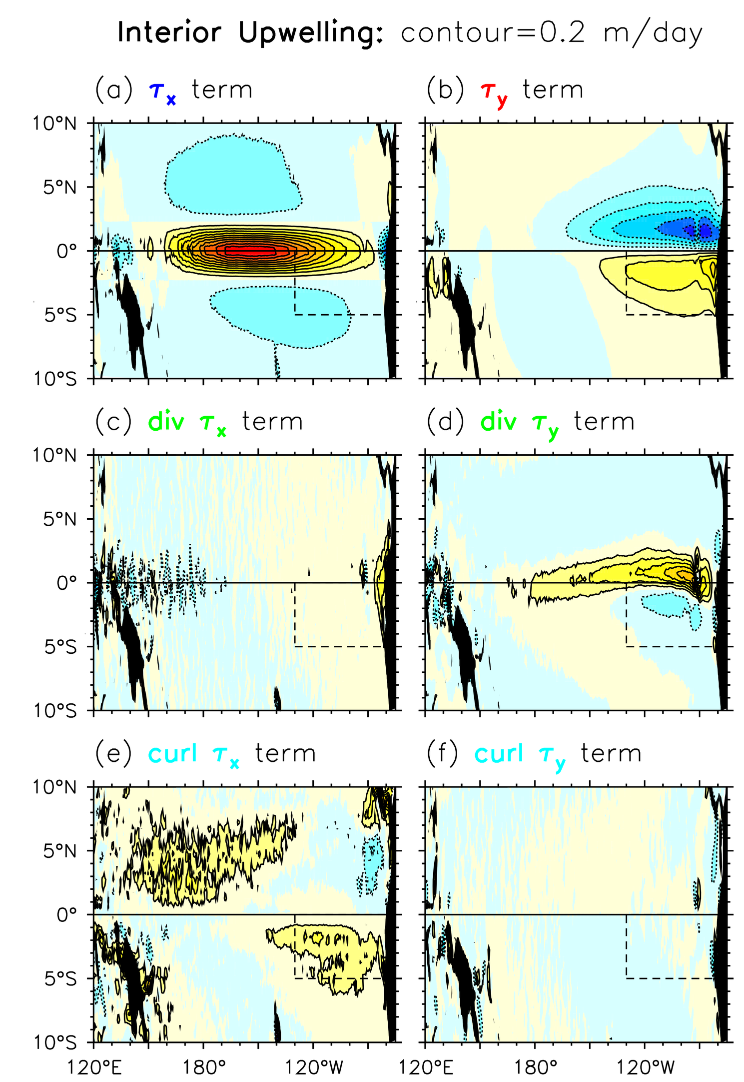


3. Impact on the Time-Mean Ocean State

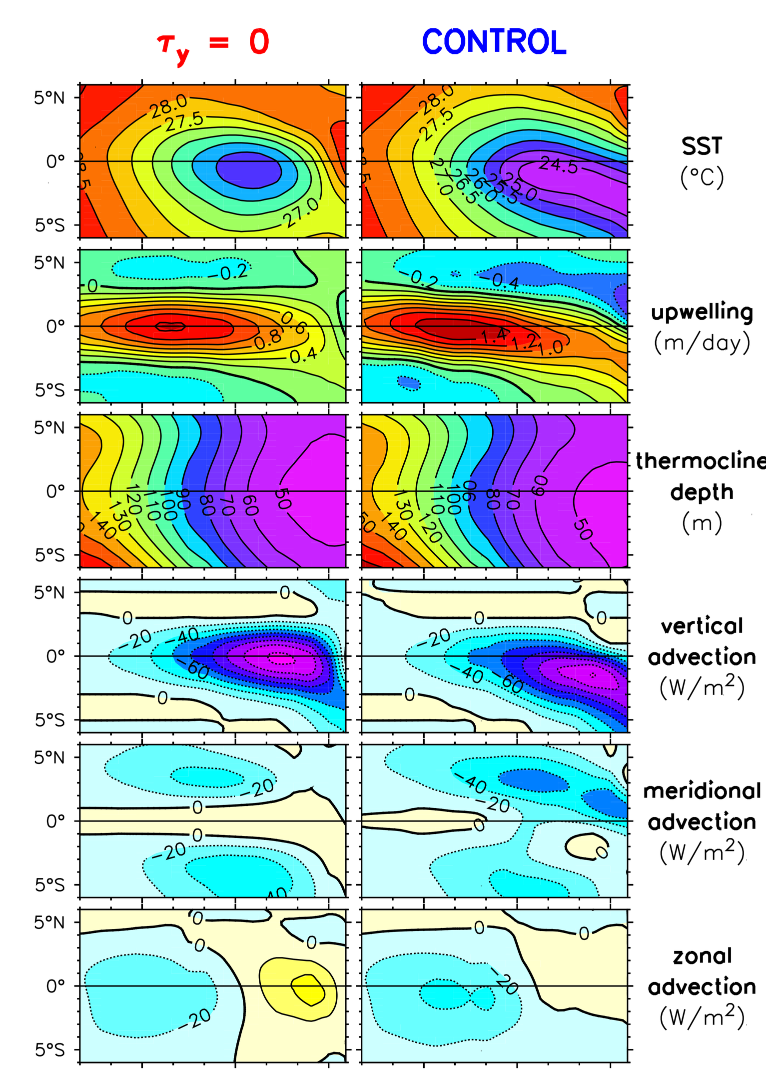
Consider an ocean surface mixed layer of depth H_m, embedded in an active layer of depth H on an equatorial β-plane. Away from coasts, the Ekman upwelling velocity at the base of the mixed layer is approximately

$$w|_{z=H_m} \approx \frac{H-H_m}{\rho H(\tilde{y}^2+1)}\left[\frac{\beta}{r_s^2}\left(\frac{\tilde{y}^2-1}{\tilde{y}^2+1}\tau_x - \frac{2\tilde{y}}{\tilde{y}^2+1}\tau_y\right) + \frac{\mathrm{div}(\boldsymbol{\tau})}{r_s} + \frac{\tilde{y}\,\mathrm{curl}(\boldsymbol{\tau})}{r_s}\right] \qquad (1)$$

where $\tilde{y} = \beta y/r_s$ is a nondimensional latitude scale, r_s is the damping rate for vertical shear across $z = H_m$, ρ is the seawater density, and $\boldsymbol{\tau} = (\tau_x, \tau_y)$ is the vector wind stress. Assuming $H_m/H = 0.4$, $r_s = (2\ \text{day})^{-1}$, and $\rho = 1023\ \text{kg/m}^3$, we use the observed mean $\boldsymbol{\tau}$ from QuikSCAT (Aug 1999–Jul 2003) to compute the upwelling due to each term in (1).



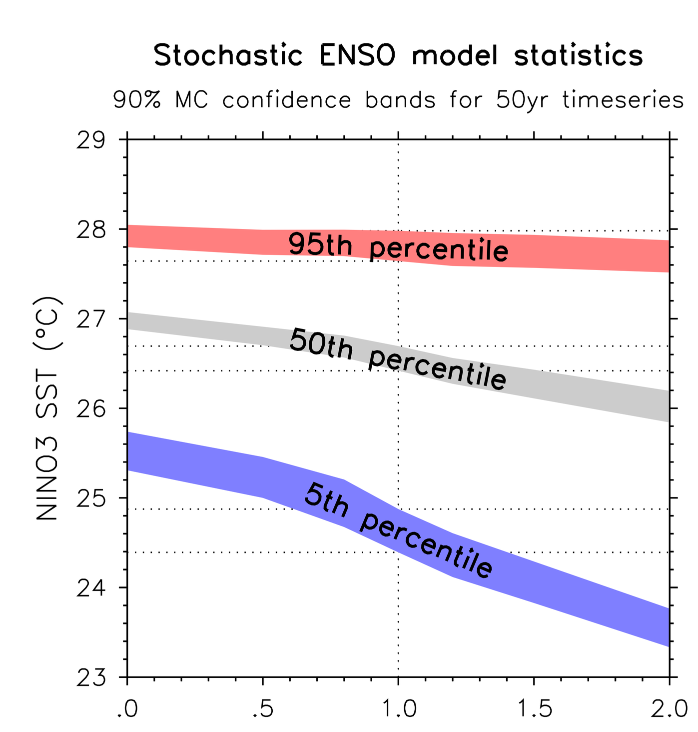
The τ_x term produces strong upwelling in the equatorial central Pacific—but contributes much less in the east, where the thermocline is shallow and particularly susceptible to air-sea interactions. The τ_y term, in contrast, gives east Pacific upwelling just a few degrees south of the equator, where the Ekman drift turns northward to become more parallel with the wind. The divergence of τ_y, which arises from changes in atmospheric boundary layer stability across the cold tongue (Wallace et al. 1989; Liu and Xie 2002), generates additional upwelling in a narrow zone just north of the equator.



The figure at left shows the impact of the observed climatological τ_y in the intermediate ocean model of Wittenberg (2002). Without τ_y, the cold tongue is weak and nearly symmetric about the equator. Adding τ_y enhances both the zonal and meridional asymmetries of the cold tongue: it cools the southeast equatorial Pacific, connects the cold tongue to the South American coast, and generates northward currents that advect cold water across the equator and tilt the thermocline downward toward the north. The extra upwelling also reduces the vertical temperature gradient in the east.

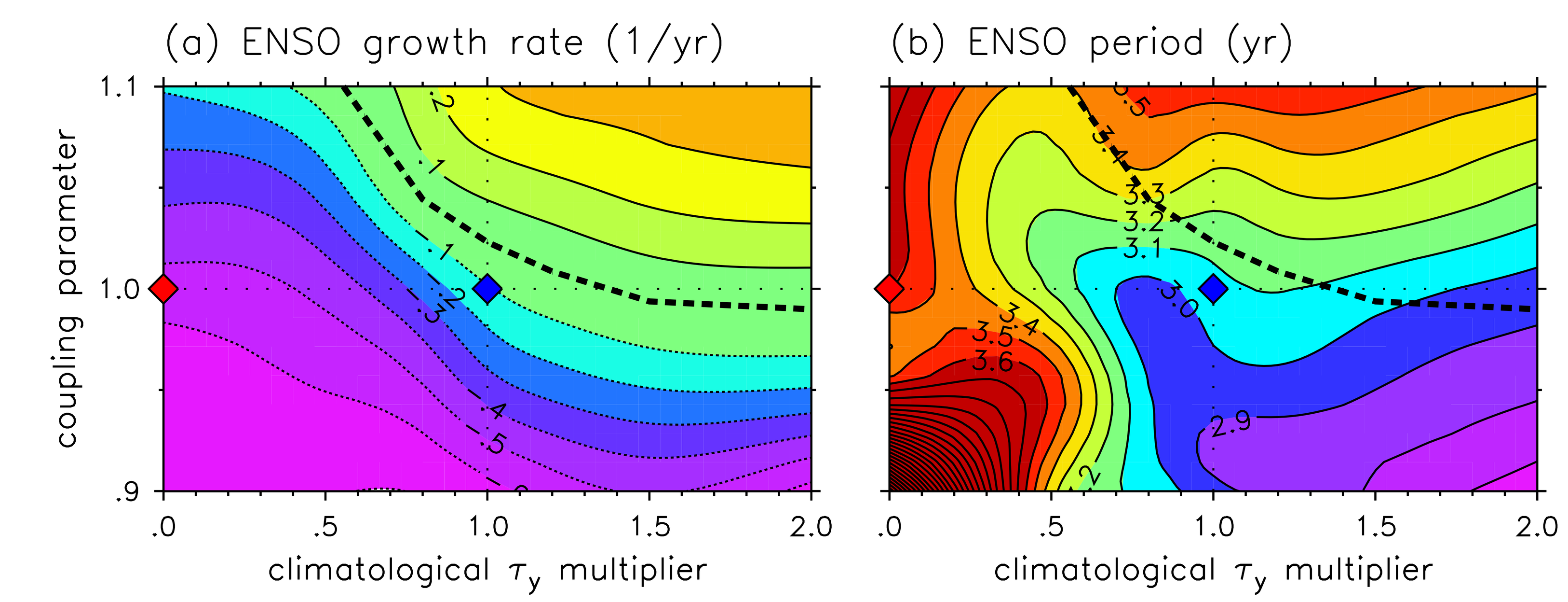
4. Impact on a Stochastically-Driven ENSO

Intensifying the background τ_y in a stochastically-driven hybrid coupled intermediate model of ENSO (Wittenberg 2002) increases the SST variability and shifts it eastward. A 40% change in τ_y alters the cold extremes enough to be detected in timeseries as short as 50 years. The results suggest that improving τ_y in coupled GCMs could also improve their ENSOs, which typically are too weak and exhibit SST anomalies too far west.



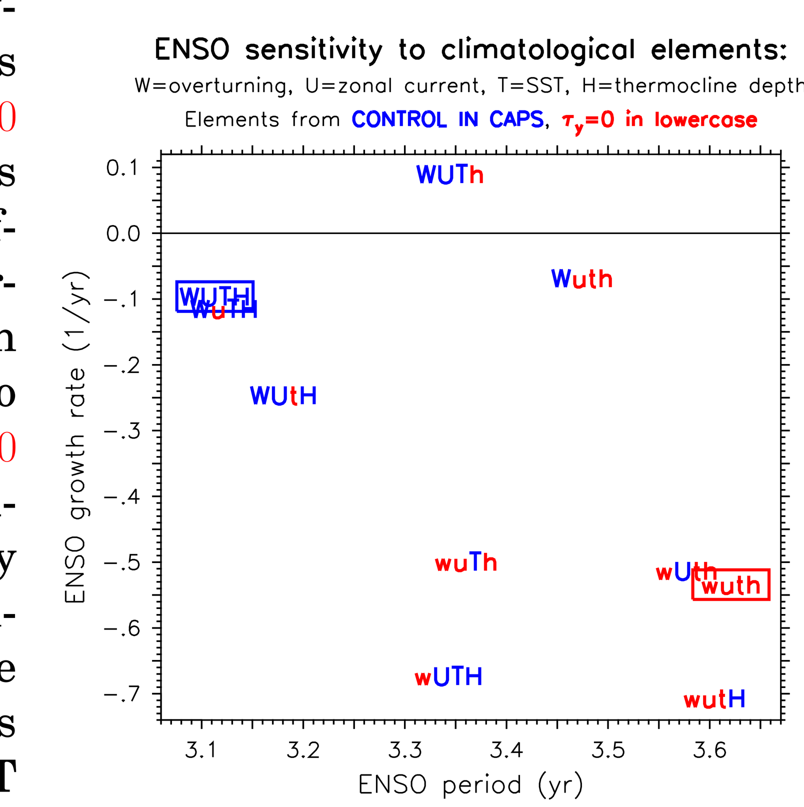
5. Impact on Linear Stability

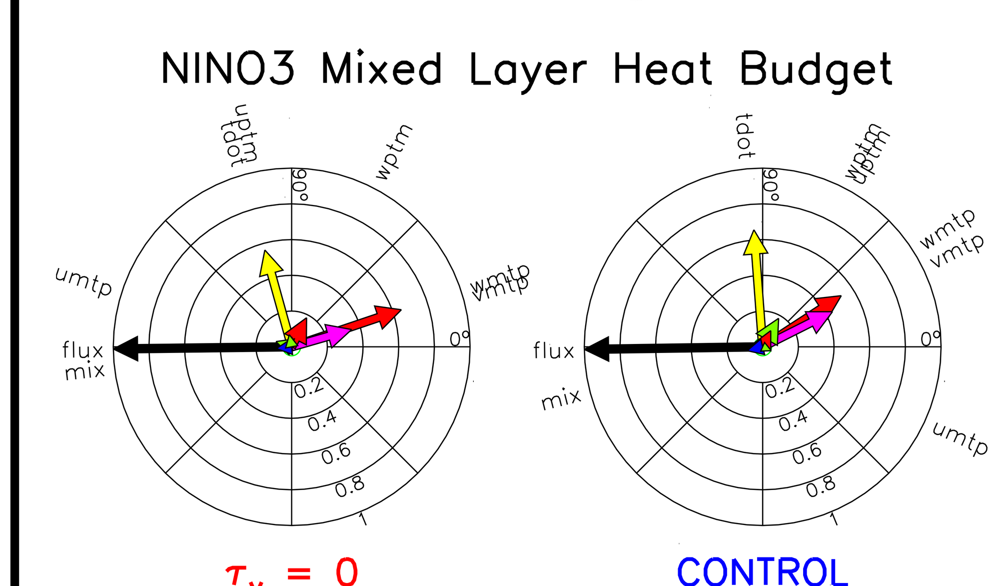
With the stochastic forcing turned off, the evolution of a tiny initial perturbation reflects the linear stability of the model ENSO. Variability is strongly damped in the absence of background τ_y, but as τ_y increases or the wind stress response to SST anomalies strengthens, ENSO grows more unstable. At the critical coupling for instability (dashed), the ENSO period decreases slightly with increasing τ_y.



6. Understanding the ENSO Sensitivity

To understand the model's ENSO sensitivity to background τ_y, we substitute parts of the control climatology into the $\tau_y = 0$ case and vice versa (right). Clearly, it is the meridional overturning that most affects the stability, by determining the air-sea feedback strength in the east Pacific. In the control the ENSO period is sensitive to thermocline depth as well, while for $\tau_y = 0$ it is sensitive to SST. The NINO3 heat budget (below) shows why. Terms are scaled by the surface heat flux which acts as a linear damping on SST anomalies. Zero phase corresponds to the SST peak and indicates a destabilizing term, while 90° leads SST and indicates a transitioning term. Advection by mean overturning (wmtp, vmtp) dominates the growth & transitioning that drive the SST tendency (tdot); hence the stability dependence on this overturning. By enhancing SST gradients and reducing the vertical temperature gradient, τ_y alters other feedbacks that control the period—like those due to zonal current and upwelling anomalies (uptm, wptm).





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