Reply

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29 September 1989 and 30 July 1990

The primary result reported in my article "On the Role of Off-equatorial Oceanic Rossby Waves during ENSO" (hereafter B89) is summarized as follows. Coupled atmosphere–ocean modeling studies and theoretical arguments indicate that the essential oceanic wave signal for ENSO is confined within a narrow equatorial band. Off-equatorial wave signals (poleward of 8° latitude) *are* associated with the ENSO events in the coupled model. However, the Rossby wave signals outside of 8° latitude provide virtually no contribution to the returning Kelvin wave that is essential in the model ENSO event (a result recently confirmed in the observations reported by Kessler 1990). Hence, the off-equatorial Rossby waves should be thought of as the product and not the triggering mechanism for an ENSO event.

Graham and White (1991, hereafter GWII) take issue with my conclusion that the off-equatorial wave signals play a passive role during ENSO (with respect to the essential equatorial waveguide processes).[1] Specifically, White et al. (1987, hereafter WPI), Graham and White (1988) and GWII hypothesized that the off-equatorial signals (centered at 12°N) are responsible for triggering ENSO events. They argue that the wave signals in the coupled model differ in character than those observed in the data and hence may indeed be important for the evolution of an ENSO event. To a certain extent it is true that extra-equatorial signals in the coupled model are quantitatively different than those observed (see below). However, one of the explicit results of my paper (see also Battisti 1988) is that *regardless of the character or generation mechanism, the off-equatorial signals should not play a major role in the evolution of an ENSO event.* This conclusion is expected from theoretical arguments, and is supported by analyses of the observational data by Kessler (1989; 1990, 1991), the general circulation ocean modeling study of Harrison et al. (1989), the coupled model study of Zebiak (1989) and the uncoupled model study of Wakata and Sarachik (1991). Nevertheless, in the discussion to follow I will show that the model calculations in GWII, coupled with the observations (not model derived data) found in WPI and Graham and White (1988), are *not* inconsistent with this conclusion and, in fact, support my result.

2. Passivity of the off-equatorial Rossby signal

a. The observed and coupled model off-equatorial Rossby signals during ENSO

In their comment, Graham and White appropriately start with the definition of "off-equatorial." This definition is not arbitrary in B89. Rather it is defined by the extent to which the off-equatorial Rossby signal provides a significant signal back into the Kelvin wave mode in the seasonless periodic integration displayed in Fig. 5 of B89. Graham and White, in referring to the off-equatorial signals, focus on the latitude 12°N (see WPI, GWI, GWII) as this is the latitude of maximum interannual variability of sea level, model derived upper layer thickness (0–400 db) and the 20°C isotherm (Kessler 1989, 1990). This is significantly poleward of that in the coupled model seasonless integration (see Fig. 4 of B89). However, as GWII note, in the Northern Hemisphere the significant integrated zonal transports inferred from the upper layer thickness anomalies (ULT) lie well equatorward of 12°N, between 5° and 9°N. Thus, the location of the mass transport associated with the coupled model off-equatorial signal *is* similar to that inferred from the geostrophic transport associated with the (uncoupled) model derived data. And it is this latter quantity that

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[1] GWII, in their first point of contention, state that I have concluded there is little evidence for systematic ENSO-related variability in Ekman Pumping *from experiments in B88.* This is not correct. This fact, based on observations, was stated in McCreary (1976, 1983) and is also true of the Rasmusson and Carpenter (1982) canonical ENSO event, and applies to the subtropics.

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is crucial for the generation of Kelvin signals through reflection of Rossby signals at the western boundary.

It is worth noting that in a seasonal cycle mode, the coupled model displays more activity in pycnocline *variance* farther from the equator (and hence more consistent with the observed data). However, the model off-equatorial activity provides an insignificant contribution to the reflected Kelvin wave, as expected from linear theory (e.g., see Clarke 1983; appendix B of Battisti 1988; and section 2c).

Graham and White, in their comment, discuss the balance of mechanisms that generate the off-equatorial Rossby signals during ENSO, and point to potential differences between the coupled model and the uncoupled wind forced ocean model results for this balance. The balance between the generation of the off-equatorial signals due to eastern boundary reflection of the Kelvin mode and wind stress curl contributions to the off-equatorial signals is highly sensitive to the observing latitude. This issue is not adequately addressed in either WPI or B89 (since it is not crucial to the findings in the latter). However, GWII produced two hindcasts of sea level using the reduced gravity ocean model forced by FSU surface wind stresses for the period 1961–1987. They note that the *off-equatorial* sea level data for the ocean model driven by observed FSU surface wind stresses agreed better with the observed values if *eastern* boundary reflection (generation) is artificially suppressed (correlation coefficients between observed sea level data and model results without reflection allowed are 0.93, compared to 0.60 to 0.67 with reflection). If this is correct, to explain this result one must appeal to either neglected processes (e.g., nonlinear effects) or question the accuracy of the wind data used to force the model. Certainly this result is not a viable argument in favor of GWII's contention for the dominance of wind stress curl as the generation mechanism for these off-equatorial signals. This result is also inconsistent with the recent calculations presented in Kessler (1989, 1990), who performs similar integrations using the same wind stress data. Comparing the results of these model calculations with the observations Kessler found that, within a band spanning 8° to 15°N, the Rossby wave signal was not significantly degraded by the eastern boundary reflected signal. He further concludes that, within this band the Rossby signal is largely generated by wind stress curl, while poleward of 15°N the signals are severely affected by the reflections from the eastern boundary; the equatorial band spanning 8°N to the equator was not examined. The reader is referred to Kessler (1989, 1990) for a more quantitative discussion of the generation mechanisms for the off-equatorial signals during ENSO.

b. Model and observational results

Cane and Sarachik (1977) showed that the amplitude of a Kelvin signal q_0, generated by reflection of long Rossby waves incident on a meridional boundary, is set by the meridional integral of the Rossby wave zonal transport U_r. Nondimensionally, this is stated:

$$q_0 = -2 \int_{y_S}^{y_N} U_r\,(x = \text{western boundary}, y, t)dy.$$

For an infinite domain, y_S and y_N are replaced by $-\infty$ and ∞, respectively. To illustrate the latitudes which significantly contribute to q_0, B89 presented a Hovmöller plot of integrated zonal mass transport $K(y, y_S)$:

$$K(y, y_S) = -H_0 \int_{y_S}^{y} U_r\,(x = \text{western boundary}, y, t)dy.$$

Significant contributions to q_0, were confined to within 8° of the equator (Fig. 4 of B89). GW presents the complimentary calculation, the *variance* of $K(y_N, y)$ relative to the total (Northern Hemisphere) variance $K(y_N, y_S)$.

Now consider Fig. 4 of GWII, in which they plot the variance of the integrated zonal transport across latitudinal bands as a percentage of the total Northern Hemisphere zonal integrated transport variance. Inasmuch, the curves represent the *potential* for each latitudinal band to effect the amplitude of the reflected Kelvin signal returning to the eastern Pacific (but see below). Both of these curves are based on model results using an identical simple reduced gravity ocean model (and the same dynamical ocean model used in Battisti 1988 and B89). One curve represents the ocean model when forced by the FSU wind data product, the other when the dynamical ocean is coupled to a simple atmospheric model (the Cane and Zebiak (1985) model, denoted LDGO model in GWII).

The two curves in Fig. 4 of GWII indicate a qualitatively consistent estimate for the latitudinal distribution of significant off-equatorial long wave Rossby variance in forming the reflected Kelvin signal. As is evident from Fig. 4 of B89, the curve for the coupled seasonless integration would tail off quicker with increasing latitude, with virtually no variance beyond 8° latitude. However, as noted above, including the seasonal cycle yields a distribution very much like that for the LDGO coupled model (to be expected, since the coupled model used in B89 is a slightly modified version of the Cane and Zebiak (1985) model).[2] Hence, it is clear that, though the variance in ULT and sea level may peak at 12°N, the essential signals contributing to the returning Kelvin wave are strongly trapped to within the equatorial waveguide. By 12°N, fully 81 (88) percent of the variance in the transport contributing to the returning Kelvin mode has been accounted

[2] To be precise, the cutoff cited in B89 for *significant transport* is at 8° latitude which, from Fig. 4 of GWII corresponds to 69% of the variance in signal from within the equatorial band. The discussion in section 4 of GWII concerning 50% ULT *variance* at 5° latitude is misleading.

for in the uncoupled wind forced ocean model (coupled model) integrations.

Of course, variance in ULT or zonally integrated mass transport does not necessarily indicate a signal significant for ENSO: phase and frequency distribution are not distinguished in this statistic. Wakata and Sarachik (1991) have presented a calculation identical to that in GWII. They forced a model ocean with the observed FSU winds from 1961–89 (GWII used the same winds from 1961–87). The ocean model is identical to the LDGO model. Hence, the model response should be identical to that summarized in Figs. 1 and 4 of GWII. Wakata and Sarachik perform a model decomposition on the model output and clearly demonstrate, in contrast to the variance, roughly 90% of reflected contribution to the Kelvin mode is due to the gravest symmetric Rossby mode—i.e., confined to within 8° of the equator (see their Fig. 4). Thus, it would appear that the uncoupled curve in GWII's variance plot (Fig. 4) overestimates the low frequency transport signals in the higher latitudes.

Kessler (1991) has done a similar calculation using the geostrophic currents calculated from *observed* hydrographic data for the period 1970 to 1987. He concludes that there is no observational support for significant integrated mass transport outside the equatorial waveguide that would lead to large waveguide (Kelvin) signals. Specifically, Kessler found the Kelvin wave signals produced by the observed geostrophic zonal integrated transports poleward of 6°N were about one meter in amplitude—at least an order of magnitude less than those occurring in either the coupled model integration or observed data (15–40 m). Thus, the *observed* integrated zonal transport is more equatorially confined (largely equatorward of 6°N) than in the uncoupled (or coupled) model integrations, and is most consistent with the seasonless cycle results presented in Fig. 4 of B89. Heading all the caveats of using linear-reduced gravity wave models to represent a continually stratified nonlinear media, the conclusion of B89 holds. The important transport signals for the reflected Kelvin mode are indeed trapped within the equatorial waveguide, certainly well equatorward of 10° latitude. Off-equatorial signals (poleward of 8° latitude) do not significantly contribute to the amplitude of the reflected Kelvin mode.

The passivity of the off-equatorial signals for the 1982–83 ENSO event is also noted by Harrison et al. (1989) and Wakata and Sarachik (1991). Harrison et al. use a general circulation model of the tropical Pacific to "successfully reproduce waveguide dynamic heights, while explicitly excluding the possibility of propagation of energy from outside the waveguide in 1981 into the waveguide in 1982" (Harrison 1989). The study by Wakata and Sarachik (1991) has previously been discussed.

The off-equatorial winds associated with ENSO in the LDGO coupled atmosphere–ocean model and the model used in the studies of Battisti (1988) and B89 are not in good agreement with those observed. This is likely to be part of the reason for some of the discrepancy between the observed and coupled model off-equatorial Rossby signal (the wind stress curl associated with the coupled model ENSO is somewhat stronger and shifted to the east of the curl calculated from the data of Rasmusson and Carpenter 1982). However, as expected from theoretical calculations, the off-equatorial Rossby signal in the coupled model is not important to the evolution of a model ENSO event. This was implied by linearly filtering the coupled model off-equatorial winds from the calculated values at 5° latitude to zero poleward of 8° latitude, introducing severe changes in the model wind stress curl but no changes in model ENSO event evolution (Battisti 1989): the essential physics is retained.

GWII, in an attempt to directly address the role of the off-equatorial signals, perform a coupled model integration, filtering the western boundary signal to allow only the transport in the equatorial band spanning 7° about the equator to contribute to the reflected Kelvin signal. Results from this integration are displayed in Fig. 5 of GWII; the western boundary off-equatorial filter is applied at model year 25. Contrary to GWII's conclusion, this calculation does not indicate that off-equatorial signals are important in the evolution or triggering of any individual ENSO events. First, the character of the oscillations is not significantly altered until at least six years (beyond model year 31) after the western boundary condition is modified. Second, the filter is applied at roughly the zero point in the SST field (model year 25, Fig. 5 of GWII)—the trigger point for the ENSO event which peaks about one year later. Yet there is no distinguishable difference between the two runs until *after* the peak of an ENSO event (year 26). However, these two calculations do demonstrate that the *climatological* interannual variability is sensitive to this model change (e.g., compare model years 0–25 and 35–45 in Fig. 5 of GWII). This result is to be expected in the long run as mass is probably not conserved in the filtered integration and, certainly, the climatological basic state has been severely modified (arbitrarily adding 8 meters to the thermocline displacement does not remedy this problem). The sensitivity of the model ENSO events to changing climatology is consistent with the findings reported in Zebiak and Cane (1987) and Battisti and Hirst (1989). However, this experiment, summarized in Fig. 5 of GWII, does not indicate that off-equatorial signals are crucial to either the evolution or triggering of an individual ENSO event.

c. *Theoretical results*

Many of the studies concerning oceanic motions during ENSO are based on the linearized equations of motion on an equatorial β-plane, including the modeling work of McCreary (1983), Cane and Zebiak (1985), Battisti (1988), and B89, and the observational interpretations from WPI and Graham and White (1988). The latter two observational studies focus on the off-equatorial signal (12°N) in ENSO. These signals take over one year to transit from the central Pacific to the western boundary, where they are postulated to return as a significant Kelvin signal in the equatorial waveguide. Appropriately, the work of McCreary (1983) is cited as the framework for this scenario. In my note (B89, section 2a) I describe the evolution of the simple model studies from McCreary's earlier work to our current conceptual framework for ENSO. In particular, the unrealistically large off-equatorial signals needed to produce interannual variability in McCreary (1983) was reconciled as a result of an overly crude coupling between the ocean and atmosphere in the waveguide.

Presently, the simple coupled model studies indicate there are two important processes for describing the model ENSOs: the gravest mode Rossby-Kelvin transit time and the coupled localized (to the equatorial eastern Pacific ocean) ocean–atmosphere instability. These processes have short time scales (about six months) which, together, act to produce the interannual signal (ENSO) (Schopf and Suarez 1988; Battisti and Hirst 1989). The model ENSO events are governed by dynamical and thermodynamics processes within the equatorial waveguide; off-equatorial signals are not necessary (see also section 4e of Battisti 1988).

On annual and longer time scales, an off-equatorial (12°N) thermocline anomaly of order 250 m is required to obtain a reflected Kelvin signal comparable to those observed in the data (about 20 m in ULT) (e.g., see Cane and Gent 1984; Clarke 1983; or Battisti 1988, appendix B). This is much larger than any observed signals, and would severely compromise the linear assumptions used in the model. An even larger off-equatorial signal is required if the meridionally straight boundary is replaced with a piece-wise continuous western boundary that better approximates the Indonesian Archipelago (e.g., see McCalpin 1987). Interestingly, the reflection of the low frequency, gravest mode Rossby wave which is crucial for the coupled model ENSO events (Battisti 1988) is not significantly degraded with this more realistic western boundary. [More than 83% of the energy from the straight boundary case is realized for motions with periods greater than one year (Clarke 1990; Cane and du-Penhoat 1990).]

3. Discussion and conclusion

Graham and White note that the off-equatorial Rossby signal in the coupled atmosphere–ocean model used in Battisti (1989) is not entirely consistent with those in the observations. This is a fair statement if one focuses on the upper-layer thickness (ULT) response in the seasonless integration in Battisti (1989) (for the seasonal cycle the agreement between uncoupled and coupled model ULT variance is much better). However, the essential signal for the returning model Kelvin mode is the zonal integrated transport (not ULT) which is indeed represented reasonably well in the coupled model (in both seasonal and seasonless mode). The coupled atmosphere–ocean model and observational studies indicate all of the significant integrated zonal transport, responsible for the returning Kelvin signal, is located within a narrow equatorial waveguide, well equatorward of 12° latitude stressed in the studies of White et al. (1987) and Graham and White (1988). Nevertheless, linear theory and modeling results indicate that if there was a significant transport at higher latitudes (there is not), it would not be essential for a model ENSO cycle. The contention of White et al. (1987) and Graham and White (1988) that off-equatorial signals (at about 12°N) may trigger ENSO events is not supported by the observations and calculations in Kessler (1989, 1990), the new observations of Kessler (1991) or the uncoupled wind-forced ocean model study of Wakata and Sarachik (1991). It is also inconsistent with the ocean GCM hindcast study of Harrison et al. (1989) and the coupled model results presented in Battisti (1988, 1989), Battisti and Hirst (1989) and Zebiak (1989). Finally, the hypothesis that the off-equatorial signals play a leading role in an ENSO event is not consistent with the linear theory the authors refer to in their arguments. The real world presents a more complicated setting than described by linear theory or prescribed in our coupled models. However, an argument for a significant role for the off-equatorial signals in the ENSO cycle must incorporate exotic effects which are not well understood (or even postulated) at this point.

Acknowledgments. The author thanks W. S. Kessler and E. S. Sarachik for numerous discussions during the preparation of this note. This work was supported by a grant from NSF Climate Dynamics, ATM-8822980.

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