Moist Teleconnection Mechanisms for the Tropical South American and Atlantic Sector*

J. DAVID NEELIN AND HUI SU

Department of Atmospheric Sciences, and Institute of Geophysics and Planetary Physics, University of California, Los Angeles, Los Angeles, California

(Manuscript received 13 September 2004, in final form 14 February 2005)

ABSTRACT

Teleconnections have traditionally been studied for the case of dry dynamical response to a given diabatic heat source. Important anomalies often occur within convective zones, for instance, in the observed remote response to El Niño. The reduction of rainfall and teleconnection propagation in deep convective regions poses theoretical challenges because feedbacks involving convective heating and cloud radiative effects come into play. Land surface feedbacks, including variations of land surface temperature, and ocean surface layer temperature response must be taken into account. During El Niño, descent and negative precipitation anomalies often extend across equatorial South America and the Atlantic intertropical convergence zone. Analysis of simulated mechanisms in a case study of the 1997/98 El Niño is used to illustrate the general principals of teleconnections occurring in deep convective zones, contrasting land and ocean regions. Comparison to other simulated events shows similar behavior. Tropospheric temperature and wind anomalies are spread eastward by wave dynamics modified by interaction with the moist convection zones. The traditional picture would have gradual descent balanced by radiative damping, but this scenario misses the most important balances in the moist static energy (MSE) budget. A small "zoo" of mechanisms is active in producing strong regional descent anomalies and associated drought. Factors common to several mechanisms include the role of convective quasi equilibrium (QE) in linking low-level moisture anomalies to free tropospheric temperature anomalies in a two-way interaction referred to as QE mediation. Convective heating feedbacks change the net static stability to a gross moist stability (GMS) M. The large cloud radiative feedback terms may be manipulated to appear as a modified static stability $M_{\rm eff}$, under approximations that are quantified for the quasi-equilibrium tropical circulation model used here. The relevant measure of $M_{\rm eff}$ differs between land, where surface energy flux balance applies, and short time scales over ocean. For the time scale of an onsetting El Niño, a mixed layer ocean response is similar to a fixed sea surface temperature (SST) case, with surface fluxes lost into the ocean and $M_{\rm eff}$ substantially reduced over ocean-enhancing descent anomalies. Use of $M_{\rm eff}$ aids analysis of terms that act as the initiators of descent anomalies. Apparently modest terms in the MSE budget can be acted on by the GMS multiplier effect, which yields substantial precipitation anomalies due to the large ratio of the moisture convergence to the MSE divergence. Advection terms enter in several mechanisms, with the leading effects here due to advection by mean winds in both MSE and momentum balances. A Kelvinoid solution is presented as a prototype for how easterly flow enhances moist wave decay mechanisms, permitting relatively small damping terms by surface drag and radiative damping to produce the substantial eastward temperature gradients seen in observations and simulations and contributing to precipitation anomalies. The leading mechanism for drought in eastern equatorial South America is the upped-ante mechanism in which QE mediation of teleconnected tropospheric temperature anomalies tends to produce moisture gradients between the convection zone, where low-level moisture increases toward QE, and the neighboring nonconvective region. Over the Atlantic ITCZ, the upped-ante mechanism is a substantial contributor, but on short time scales several mechanisms referred to jointly as troposphere/SST disequilibrium mechanisms are important. While SST is adjusting during passive SST (coupled ocean mixed layer) experiments, or for fixed SST, heat flux to the ocean is lost to the atmosphere, and these mechanisms can induce descent and precipitation anomalies, although they disappear when SST equilibrates. In simulations here, cloud radiative feedbacks, surface heat fluxes induced by teleconnected wind anomalies, and surface fluxes induced by QE-mediated temperature anomalies are significant disequilibrium contributors. At time scales of several months or longer, remaining Atlantic ITCZ rainfall reductions are maintained by the upped-ante mechanism.

^{*} Institute of Geophysics and Planetary Physics, University of California, Los Angeles, Contribution Number 6208.

Corresponding author address: Dr. J. David Neelin, Department of Atmospheric Sciences, University of California, Los Angeles, 405 Hilgard Ave., 7127 Math Science Bldg., Los Angeles, CA 90095-1565. E-mail: neelin@atmos.ucla.edu

1. Introduction

Teleconnections in the Tropics involve the interaction of atmospheric wave dynamics with moist convective processes in tropical convection zones. While there has been much theoretical work on tropical wave dynamics emanating from a specified heat source that mimics convection (Matsuno 1966; Webster 1972; Gill 1980; Garcia and Salby 1987; Webster and Chang 1988), development of theory for the interaction with convection zones has lagged. Such interactions include feedbacks from convective heating and cloud radiative interaction. They potentially alter the propagation of the teleconnection signal, and they create the precipitation impacts that are important to land ecosystems and human activities. We propose the term moist teleconnections as an appropriate shorthand for this class of dynamics, in which the interaction with moist processes is essential. This paper is part of an ongoing project aimed at assembling a theory of moist teleconnections (Neelin and Zeng 2000, hereafter NZ00; Su et al. 2001, hereafter SNC01; Su and Neelin 2002, hereafter SN02, 2003; Su et al. 2003; Neelin et al. 2003, hereafter NCS03) and related effects in tropical global change anomalies (NCS03; Chou and Neelin 2004, hereafter CN04). Here, eastward teleconnections from El Niño-Southern Oscillation (ENSO) to equatorial South America and the Atlantic sector are treated. For this case, land surface interactions are added to the set of feedbacks. The contrast of land region precipitation response to that of the neighboring ocean sector is of interest.

The angle from which we approach this problem is motivated by two tools: (i) the moist static energy (MSE) budget of the primitive equations and (ii) convective quasi-equilibrium (QE) closures. In the MSE budget of the dry thermodynamic equation and the moisture equation, convective heating and the convective moisture sink cancel when vertically integrated, allowing insight into the more subtle balances that control the moist thermodynamics. The cloud radiative forcing (CRF) terms can also be manipulated to clarify the feedbacks involved. In QE closures, convection tends to remove parcel buoyancy and thus constrains the vertical temperature profile in deep convective regions. As a consequence, temperature and moisture budgets become linked, and the MSE budget becomes the leading thermodynamic constraint in deep convective regions (Emanuel et al. 1997; Neelin 1997).

A primary focus here is on the precipitation impacts and the associated anomalies in vertical motion and the divergent component of the flow in and near deep convective zones. However, as will be seen below, these are closely tied with wave propagation and the horizontal structure of temperature and moisture fields, so an understanding of the moist teleconnection as a whole is essential. There have been a number of studies establishing the link between ENSO anomalies and precipitation anomalies over equatorial South America and Atlantic ITCZ (Hastenrath and Heller 1977; Moura and Shukla 1981; Hastenrath 1990a,b; Giannini et al. 2001; Chiang and Sobel 2002, hereafter CS02; Chiang et al. 2002; Paegle and Mo 2002; Rao et al. 2002; Grimm 2003). The aim here is to contribute an understanding of the interplay of moist dynamical mechanisms that occurs in these teleconnections. An in-depth case study of the onsetting 1997/98 El Niño is used, first evaluating the simulation, then taking apart the mechanisms involved using an intermediate complexity primitive equation model, the quasi-equilibrium tropical circulation model (QTCM) designed specifically for such studies. These mechanisms are believed to be general enough that this provides a prototype for other cases, although the relative contributions may change. Mechanisms are phrased in terms of warm free troposphere temperature anomalies and compensating descent anomalies, although the converse would hold for cold events. Results are then compared to statistics for other events.

The structure of the paper is as follows. After summarizing the model (section 2), the simulation of the July-November 1997 response to specified Pacific SST anomalies is compared to observations as our primary case study (section 3). The MSE budget is then summarized in general and QTCM form (section 4)locations of main symbol definitions are recapped in the appendix. MSE budget anomalies for the 1997 case are examined to identify important factors (section 5), and manipulations of the MSE budget that aid understanding of mechanisms are introduced (section 6). These are used to generate hypotheses regarding mechanisms (section 7) including a prototype for how these interact in a moist Kelvin wave-like solution. Tests of a number of mechanisms by intervening in the model dynamics for the 1997 case study are then presented in section 8a, a mixed layer ocean case is examined in section 8b, and section 8c provides a comparison to other events.

While sections 3–8 provide a detailed account of the evidence for multiple mechanisms in this model's simulated teleconnections, section 9 extracts from this a more general summary of moist teleconnection mechanisms and the analysis approach. It is relatively self-contained for readers preferring the overview.

2. Model and cases

The quasi-equilibrium tropical circulation model (QTCM) version 2.2 with a modified cloud radiation scheme is used. This is the same version as used in SN02 for consistency. The model physics includes a Betts–Miller convective adjustment scheme (Betts and Miller 1986), a weakly nonlinear cloud radiative scheme, and a simple interactive land model, "SLand," with a surface water budget for a single soil layer (NZ00; Zeng et al. 2000, hereafter ZNC00).

Similar to SN02, simulations for the 1997 case are performed with observed SSTs in the tropical Pacific (30°S to 30°N) in regions where SST anomalies are positive and climatological SSTs are specified over other oceanic regions ("POSPAC" runs). Such POSPAC runs provide a simple case for examining teleconnections; the impacts of SST in other regions are discussed in SNC01. An ensemble of 10 members with slightly different initial conditions is conducted. We compare the ensemble means of the POSPAC runs with those of the "CLIM" runs, where climatological SSTs over the whole model domain (60°S-60°N) are used. Runs with a mixed layer ocean over the tropical Atlantic are also conducted. A fixed climatology of implied ocean heat transport divergence is estimated from a previous simulation with climatological SST. Simulations in which particular terms in the model equations are altered for hypothesis testing purposes are described in section 8.

We choose the period July-November 1997 as a primary case because the relative drought over equatorial South America and the Atlantic ITCZ region was large in observations during this time. The 5-month average partially helps in filtering transient variability when comparing to observations. The precipitation anomalies in the time period also have a large spatial extent within equatorial South America, so the problem is not localized to a single region. Furthermore, comparisons between land and the Atlantic ITCZ are facilitated because broad anomalies occurred over both in this case. This choice of season implies that we are not explicitly addressing the well-known Nordeste Brazil drought season of March through June, when the convection zone is farther south. The Nordeste problem is of a more regional scale and has strong dependencies on the Atlantic (Hastenrath 1990b). We believe that some of the mechanisms found here are relevant for the localized Nordeste problem, but we address first this larger spatial-scale case. Equatorial South American and Atlantic ITCZ rainfall are well correlated with ENSO indices during September-November (Hastenrath 1990a; Grimm 2003), so the 1997 event is unusual mainly in

that anomalies were large and started earlier than is typical.

Comparisons to other events are based on a run forced with observed SST (Reynolds and Smith 1994) from 1982 to 1998. SST is specified globally, but the analysis uses cases in which Atlantic SST anomalies are small.

3. Simulation of the South American drought during El Niño

Figures 1 and 2 show the anomalies of precipitation, tropospheric averaged temperature T' (850–200 hPa), 850-hPa wind, 500-hPa vertical pressure velocity (ω'), and outgoing longwave radiation (OLR') for observations and the POSPAC run, respectively, where prime denotes anomalies. All observed anomalies are relative to the climatological means from 1982 to 1998. Temperature, horizontal wind, and vertical pressure velocity data are from the National Centers for Environmental Prediction-National Center for Atmospheric Research (NCEP-NCAR) reanalysis (Kalnay et al. 1996). OLR is obtained from the National Oceanic and Atmospheric Administration Climate Analysis Center (CAC). All modeled anomalies are differences between the ensemble means for the POSPAC run and those for the CLIM run. The regions of interest here, that is, equatorial South America and the Atlantic Ocean, exhibit considerable reductions in precipitation. The region of weak $(-0.5 \text{ mm day}^{-1})$ precipitation reduction in observations extends farther southward and westward over the continent than is simulated. However, the model captures the general features of the drought over equatorial South America, shifted eastward from the region of strongest climatological precipitation and extending across the Atlantic ITCZ.

We also note that while the climatological convection zone affects the position of the precipitation anomalies, the anomalies are far from being simply proportional to the climatological precipitation. In evaluating budgets, we define equatorial South America (ESA) and Atlantic ITCZ index regions to be regions with greater than 0.5 mm day^{-1} precipitation anomaly, south of 15° N, for land and the Atlantic Ocean in Fig. 2a. In both observations and the model, this includes much of the Amazon but not the west coast of South America. In model budget analysis, the ESA region includes the Amazon and northern South America from Venezuela and eastern Columbia to northern northeastern Brazil.

The corresponding descent anomalies and increased OLR seen in Figs. 1c,d are consistent with the precipitation anomalies; descent anomalies are roughly proportional to the negative precipitation anomalies. The QTCM anomalies (Figs. 2c,d) reproduce this relation-



FIG. 1. Observed anomalies during Jul–Nov 1997. (a) Xie and Arkin (1997) precipitation anomalies; contoured at -6, -4, -2, -1, -0.5, 0, 1, 2, 4, 6, 8, and 10 mm day^{-1} with light shading below -0.5 and dark shading above 1 mm day⁻¹. (b) Tropospheric temperature (averaged 850 to 200 hPa) and 850-hPa winds; contour interval is 0.25° C with light shading above 0.25 and dark shading above 0.75 C; wind vectors are in m s⁻¹. (c) 500-hPa vertical pressure velocity ω from the NCEP–NCAR reanalysis; contour interval is 10 in units of 10^{-3} Pa s⁻¹ with dark shading below -10 and light shading above 10. (d) OLR from CAC; contoured at -40, -30, -20, -10, and -4 and 4, 8, 12, 20, 30, 40, and 50 W m⁻² with dark shading below -4 W m⁻².

ship. The widespread tropospheric warming extending eastward is captured in the model, although the QTCM T' has a latitudinally narrower pattern over the Atlantic than the reanalysis, in which regions of midlatitude warming may be noted. The modeled 850-hPa wind anomalies are weaker than those of the reanalysis in the equatorial Central Pacific. Over South America and the Atlantic, the modeled response has a coherent pattern of easterly anomalies while the reanalysis has a more complex pattern that includes both easterly and westerly anomalies.

This model version has weaker internal variability (Lin and Neelin 2000) than observed [and than the Lin



FIG. 2. Simulated anomalies from the POSPAC run during Jul-Nov 1997. (a) Precipitation anomalies; contoured at -4, -3, -2, -1, -0.5, 0, 1, 2, 4, 6, 8, and 10 mm day^{-1} with light shading below -0.5 and heavy shading above 1. QTCM climatological precipitation 2 mm day⁻¹ contour overlaid as dark solid line. (b) Tropospheric temperature (averaged 850 to 200 hPa) and 850-hPa winds; contour interval is 0.25° C with light shading above 0.25 and dark shading above 0.75; wind vectors are in m s⁻¹; (c) 500-hPa vertical pressure velocity ω , contoured at -60, -40, -30, -20, -10, -4, 0, 4, 8, 12, and 16 in units of 10^{-3} Pa s⁻¹ with dark shading below -4 and light shading above 4. (d) OLR; contoured at -40, -30, -20, -10, -4, 0, 4, 8, 12, and 16 W m⁻².

et al. (2000) version]; some features of the observed in this 5-month average may not be statistically significant. The general features of the patterns are reproducible for other ENSO events in observations (Wallace et al. 1998) and the QTCM (ZNC00, their Figs. 17 and 18).

4. Moist static energy analysis and implications for land and ocean regions

a. Moist static energy budget

When the thermodynamic Eq. (A1) and moisture Eq. (A2) from the primitive equations are added and ver-

tically integrated, the convective heating and moisture sink terms cancel, resulting in an MSE equation:

$$\partial_t (\langle T \rangle + \langle q \rangle) + \langle \mathbf{v} \cdot \nabla (T+q) \rangle + \langle \omega \partial_p h \rangle = F^{\text{net}},$$
(1)

where the MSE is h = s + q, the dry static energy is $s = T + \phi$, with ϕ as the geopotential, and where temperature *T* and specific humidity *q* are in J kg⁻¹ (i.e., *T* and *q* absorb the heat capacity at constant pressure and latent heat of condensation, respectively). Vertical integrals $\langle \cdot \rangle$ through the depth of the troposphere p_T are here mass integrals $\int \cdot dp/g$ so all terms are in units of W m⁻².

The net flux into the atmospheric column, signed positive when heating the atmosphere, is

$$F^{\text{net}} = F_t^{\text{net}} - F_s^{\text{net}},\tag{2}$$

where the net flux at the top of the atmosphere (TOA) F_t^{net} and the net surface flux F_s^{net} are

$$F_t^{\text{net}} = S_t^{\downarrow} - S_t^{\uparrow} - R_t^{\uparrow}$$
(3)

$$F_s^{\text{net}} = S_s^{\downarrow} - S_s^{\uparrow} + R_s^{\downarrow} - R_s^{\uparrow} - E - H, \qquad (4)$$

with both signed positive downward. Signs on individual flux terms are chosen such that they are typically positive in the climatology. On the solar radiation terms *S* and longwave radiative terms *R*, arrows denote direction of the flux in a two-stream radiation treatment, subscripts *s* and *t* denote surface and model top, and net heating of the atmospheric column terms is $S^{net} = S_t^{\perp} - S_t^{\uparrow} - S_s^{\downarrow} + S_s^{\uparrow}$ and $R^{net} = -R_t^{\uparrow} + R_s^{\uparrow} - R_s^{\downarrow}$, with $R_t^{\downarrow} \approx 0$. Surface evaporation and sensible heat are denoted *E* and *H*, respectively. In GCMs and in the QTCM, there are also horizontal diffusion terms in Eq. (1), which are small but not negligible in large-scale budgets. We omit them for presentation purposes except where necessary.

The MSE budget for the QTCM is a close counterpart to that in the primitive equations. The main difference is that the QE constraints on temperature have lead to a simpler form for the divergence of MSE term. In Eq. (1), $\langle \omega \partial_p h \rangle$ is replaced by $M \nabla \cdot \mathbf{v}_1$, where *M* is the gross moist stability (details in the appendix), here used in units of J m⁻², that is, absorbing (p_T/g) relative to Yu et al. (1998, hereafter YCN98). This is an effective static stability for moist motions that includes the partial cancellation of adiabatic cooling by latent heating (Neelin and Held 1987; Neelin and Yu 1994, hereafter NY94). The baroclinic wind \mathbf{v}_1 is the wind projection on a deep baroclinic structure V_1 (derived from baroclinic pressure gradients that result from deep convection under certain approximations; see NZ00). The divergence $\nabla \cdot \mathbf{v}_1$ gives the horizontal variations of the vertical velocity, the deep vertical structure of which is derived from V_1 . The sign of $\nabla \cdot \mathbf{v}_1$ is positive for upper tropospheric divergence and upward vertical velocity.

For time or ensemble averages such that the time derivative terms are negligible, with anomalies relative to climatology denoted with primes, the anomaly MSE equation in QTCM form is

$$(M\boldsymbol{\nabla}\cdot\boldsymbol{\mathbf{v}}_{1})' + \langle \boldsymbol{\mathbf{v}}\cdot\boldsymbol{\nabla}T\rangle' + \langle \boldsymbol{\mathbf{v}}\cdot\boldsymbol{\nabla}q\rangle' = F_{t}^{\mathrm{net}'} - F_{s}^{\mathrm{net}'},$$
(5)

where the vertical integrals are evaluated on the retained vertical structures. Anomalies in nonlinear terms such as moisture advection $\langle \mathbf{v} \cdot \nabla q \rangle'$ contain changes in the contributions of transients, as discussed with formal notation in SN02. In the runs presented here, changes in the nonlinear products of transient motions appear to be secondary and are omitted for brevity in most of the discussion, though they could be included if needed.

b. Implications for anomalies in land convective zones

Because of the low heat capacity over land, the net surface energy flux on time scales longer than a day obeys

$$F_s^{\text{net}} = 0, \tag{6}$$

so $F_s^{\text{net'}}$ drops out of Eq. (5) over land. Within convection zones, this provides considerable simplification for treatment of land surface feedbacks (Zeng and Neelin 1999, hereafter ZN99; NZ00). Only TOA radiative forcing F_t^{net} and transport processes, that is, advection of temperature and moisture, come into play in balancing the MSE divergence.

To a first approximation, that is, to the extent that convective QE holds and the MSE budget alone controls the dynamics in the convective zones, the atmospheric dynamics is independent of T'_s . The partition between E' and other fluxes does have some effect, especially when departures from QE must be considered.

c. Implications over a passive ocean surface layer

Consider an ocean surface layer governed by

$$\partial_t (c_s \hat{T}_s)' + D'_s = F_s^{\text{net}'},\tag{7}$$

where c_s is the heat capacity of the layer, \hat{T}_s is the temperature averaged through the surface layer, and D_s is the divergence of the ocean heat transport in the layer, including turbulent and advective fluxes from below. In experiments here we use a 50-m fixed-depth

mixed layer, with \hat{T}_s as SST and c_s constant. For a passive ocean surface layer, simply reacting to atmospheric heat flux anomalies with $D'_s = 0$, adding the surface layer energy budget [Eq. (7)] to the perturbation MSE budget [Eq. (5)] and noting that the surface flux term communicating between the two vanishes, yields

$$(M\boldsymbol{\nabla}\cdot\boldsymbol{\mathbf{v}}_{1})' + \langle \boldsymbol{\mathbf{v}}\cdot\boldsymbol{\nabla}T\rangle' + \langle \boldsymbol{\mathbf{v}}\cdot\boldsymbol{\nabla}q\rangle' = F_{t}^{\mathrm{net}'} - \partial_{t}(c_{s}\hat{T}_{s})'.$$
(8)

This implies that while the ocean is in disequilibrium with the teleconnected forcing, to the extent that changes in ocean heat storage are being created by surface fluxes, the SST anomaly can contribute to driving MSE divergence. Once the SST anomaly reaches approximate equilibrium with the atmosphere, that is, when $\partial_t (c_s \hat{T}_s)'$ in W m⁻² is small compared with other terms in the anomaly MSE budget, then the SST is approximately a by-product in convective regions, much like the land surface temperature. Fixed SST experiments can be roughly viewed as a large heat capacity case such that disequilibrium persists.

d. Implied precipitation and the GMS multiplier effect

The moisture budget [Eq. (A2)] gives the precipitation anomalies (vertically integrated convective heating) as

$$P' = (M_a \nabla \cdot \mathbf{v}_1)' - \langle \mathbf{v} \cdot \nabla q \rangle' + E'.$$
(9)

Using the MSE Eq. (5) for the divergence yields

$$P' = \frac{\overline{M}_q}{\overline{M}} \left[-\langle \mathbf{v} \cdot \nabla T \rangle' + F_t^{\text{net}'} - F_s^{\text{net}'} \right] + E' - \left(\frac{\overline{M}_q}{\overline{M}} + 1 \right) \langle \mathbf{v} \cdot \nabla q \rangle' + \left(-\frac{\overline{M}_q}{\overline{M}} M' + M'_q \right) \nabla \cdot \overline{\mathbf{v}}_1.$$
(10)

The factor $(\overline{M}_q/\overline{M})$ is a measure of the moisture convergence relative to the MSE divergence. This factor tends to be several times larger than unity, as noted under slightly different notation in ZN99. We thus refer to the effect of (M_q/M) as the gross moist stability (GMS) multiplier effect. It boosts a small signal in the MSE equation to a convective heating and precipitation impact several times larger. In the observation-based estimates of YCN98, this would be roughly 4 or 5 in the deep convective zones. In results here it is somewhat smaller (although it is amplified by cloud radiative feedbacks as discussed in section 6).

In identifying mechanisms, it is usually important to expand the nonlinear terms of the form $\langle \mathbf{v} \cdot \nabla q \rangle'$ into \mathbf{v}'

and q' contributions (and an additional term for changes in transients). Each term in Eq. (10) is typically associated with some particular pathway for precipitation anomaly impacts. The most important to the results here will be elaborated in section 7.

5. Moist static energy budget of South American/Atlantic anomalies

Figures 3–8 show the spatial patterns of terms relevant to the MSE budget analysis for the POSPAC run, with Table 1 and Table 2 summarizing area-averaged MSE budget and flux terms for the ESA and Atlantic ITCZ land and ocean index regions indicated as dashed outlines in Figs. 6–8.

Figure 3 shows the anomalies of surface temperature, net surface and TOA energy fluxes, and the net heating rate of the atmospheric column by these fluxes. The observed warm SST anomalies within the tropical Pacific are prescribed. Over northern South America, land surface temperature (Fig. 3a) increases slightly, but the main warming is confined to the region with strong negative rainfall anomalies. This warming is primarily caused by increased surface solar radiation due to reduced cloudiness (Fig. 3b). The net surface energy flux anomalies $F_s^{\text{net'}}$ (Fig. 3b) are approximately zero over land, as anticipated in Eq. (6). In contrast, positive net surface energy fluxes are seen in regions of descent anomalies over the northern Tropical Atlantic, especially the convective zones, and similarly in the western Pacific. These would tend to warm SST and represent a loss of heat from the atmosphere compared to neighboring land regions. The large negative $F_s^{\text{net'}}$ anomalies within the main El Niño warm region provide the forcing for the atmospheric teleconnection.

The TOA net flux (Fig. 3c) is generally much weaker, especially over the teleconnection regions of interest. Some of the individual components of $F_t^{\text{net}'}$, however, are large, as will be discussed below for Fig. 4 and Fig. 5, where these are broken out.

A contrast in the energetics of descent anomalies over land and ocean is seen in the net flux into the

TABLE 1. Area-averaged moist static budget anomalies for the eastern ESA region and the Atlantic ITCZ region for the POSPAC run. Values are in W m⁻². Net flux $F^{\text{net}} = F_t^{\text{net}} - F_s^{\text{net}}$ is signed positive when heating the atmospheric column.

	ESA region	Atlantic ITCZ
$(M\nabla \cdot \mathbf{v}_1)'$	-20.5 (100%)	-22.1(100%)
$-\langle \mathbf{v} \cdot \nabla T \rangle'$	-8.0 (39%)	-2.3(10%)
$-\langle \mathbf{v}\cdot \nabla q \rangle'$	-8.7 (42%)	-5.9(27%)
F ^{net'}	-1.3 (6%)	-11.3(51%)
$F_{s}^{net'}$	0.0	12.4 (56%)
$F_{t}^{net'}$	-1.3 (6%)	1.1 (-5%)

TABLE 2. Area-averaged flux anomalies at the TOA and at the surface for the eastern ESA region and the Atlantic ITCZ region for the POSPAC run. Values are in W m⁻². Some flux terms appear with a negative sign so that all values appearing in the table are signed positive when tending to heat the atmospheric column. Additional subscripts on R_t^{\uparrow} and $R^{\text{net'}}$ denote the linearized contributions from cloud, *T*, *q*, and surface temperature *T_s* anomalies, respectively. The boldface on these is a reminder that TOA terms are most relevant over land (or equilibrated ocean), while $R^{\text{net'}}$ is the most relevant over fixed-SST ocean.

	ESA region	Atlantic ITCZ
1) Surface		
$F_s^{\rm net'}$	0.0	-12.4
$-S_{s}^{\downarrow \prime}$	-10.0	-11.4
$S_{s}^{\uparrow \prime}$	1.5	0.7
$-R_{s}^{\downarrow \prime}$	-2.9	-0.8
$R_s^{\uparrow'}$	6.4	0.0
E'	-4.8	1.4
H'	-9.8	-2.3
2) TOA		
$F_t^{\rm net'}$	-1.3	1.1
$-S_t^{\uparrow'}$	8.6	10.8
$- \dot{R_t^{\uparrow'}}$	-9.9	-9.7
3) Longwave contributions		
$-R_{\rm cld'}^{\uparrow\prime}R_{\rm cld}^{\rm net'}$	- 6.5, -5.3	-7.2, - 5.9
$-R_{Tt}^{\uparrow \prime}R_T^{\text{net}\prime}$	- 2.9, -6.1	-2.1, - 4.4
$-R_{at}^{\uparrow\prime}, R_{a}^{\text{net}\prime}$	0.8, -1.6	0.4, -0.8
$-R_{T_s}^{\uparrow}t'$	-0.3	0.0
4) Cloud (shortwave and		
longwave) radiative feedback		
$CRF'_t, CRF^{net'}$	2.1, -5.2	3.6, - 5.8

atmospheric column $F^{net'}$ in Fig. 3d. Warming of the atmospheric column over the warm SST in the Pacific is seen as expected, due to increased surface heat fluxes and reduced OLR. The net flux for anomalous descent regions over ocean and over land are quite different. The atmosphere over certain ocean regions (to the north and south of the main El Niño warm area in the tropical Pacific and in the tropical Atlantic, especially north of the equator) experiences substantial cooling due to the heat loss into the ocean at the surface. Reference to Fig. 4b shows that the shape of these regions is strongly controlled by the region of reduced cloudiness, where less solar radiation is reflected. The pattern of $F^{net'}$ in Fig. 3d is generally controlled by that of the net surface flux (Fig. 3b), with relatively small contributions from the top of atmosphere. The exception is over land, where the top of the atmosphere balance entirely controls $F^{\text{net'}}$, as noted in section 4b. A very slight cooling of the column may be noted over South America, but this is sensitive to canceling cloud radiative effects.



FIG. 3. Anomalies during Jul–Nov 1997 for the POSPAC run. (a) Surface temperature; contour is 0.5°C, shaded above 1°C. (b) Net surface flux F_s^{net} ; contour internal is 10 W m⁻², plus 4 W m⁻² contour, with light (dark) shading below (above) ± 10 W m⁻² with stippling for net surface fluxes of magnitude smaller than 0.2 W m⁻². (c) Net TOA flux F_t^{net} ; contoured at ± 2 and ± 4 ; otherwise interval is 10 W m⁻², with light (dark) shading below (above) ± 2 W m⁻². (d) Net flux into the atmosphere $F^{\text{net}} = F_t^{\text{net}} - F_s^{\text{net}}$ (contoured at -2, -4, and -10 W m⁻²; otherwise interval is 10 W m⁻². Positive sign indicates downward flux anomalies in (b) and (c) and atmospheric heating by fluxes in (d).

The TOA cloud radiative forcing contributions may be seen in Fig. 4. There is strong cancellation between reduced solar reflection (Fig. 4b) and increased OLR (Fig. 4a) associated with reduced cloudiness over anomalous descending regions. The net TOA CRF (Fig. 4c) actually contributes a heating of a few watts per meter squared to the land–atmosphere column over regions of negative precipitation anomalies [roughly consistent with the central Pacific results of Kiehl (1994, his Fig. 6a)]. While it might seem counterintuitive to have a heating contribution in regions that are anomalously warm, this simply opposes the anomalous de-



FIG. 4. (a) OLR anomalies and (b) upward solar radiation anomalies at TOA due to cloud fraction changes. Contour interval is 5 W m⁻², with light (dark) shading for values below (above) \pm 5 W m⁻². (c) Total cloud radiative forcing at TOA, signed positive when downward. Contours and shading are similar but at 2 W m⁻².

scent as shown in section 6. The magnitude and sign of this effect depend on the details of the shortwave versus longwave cancellation.

Over the Atlantic ITCZ, TOA CRF in Fig. 4c is similar to that over land but the effect of cloud forcing is very different because of the role of surface fluxes. The



FIG. 5. The OLR anomalies due to (a) atmospheric temperature anomalies and (b) moisture anomalies.

largest contribution to the net surface flux (Fig. 3b) in this region is downward shortwave, as may be seen in Table 2. Thus the bulk of the $F^{net'}$ column cooling (Fig. 3d) in the Atlantic ITCZ is due to net cloud radiative cooling as surface solar is (temporarily) lost into the ocean. Rather than TOA balances, the atmosphere column balances of radiative terms are the more important. The increase in OLR associated with reduced cloud fraction is the dominant term in the atmospheric column balances, creating a substantial cooling tendency (Table 2) that reinforces the descent anomaly.

The contributions of temperature and moisture anomalies to OLR are shown in Fig. 5. The tropospheric warming increases OLR by up to 3 W m⁻², while the moistening would reduce OLR by about 1 W m⁻² over equatorial South America. The net TOA cooling due to temperature and moisture anomalies is only about 2 W m⁻². This proves small compared to other terms in the MSE budget seen in Figs. 6–8, suggesting a modest role for the traditional balance of radiative cooling—creating descent.

The magnitude of moisture and MSE divergence terms due to descent/divergence anomalies may be seen in Fig. 6. The column-integrated moisture convergence anomalies (Fig. 6a) correspond well to the precipitation anomalies in Fig. 2a and have a magnitude of about -40 W m^{-2} over ESA and a slightly larger magnitude over the Atlantic ITCZ. To compare Fig. 2a and Fig. 6a in the same units, multiply precipitation in millimeters per day by 28 to obtain vertically integrated convective heating in watts per meter squared. Thus the descent



FIG. 6. Anomalies of (a) moisture convergence $(M_{q1} \nabla \cdot \mathbf{v}_1)'$ and (b) MSE divergence $(M\nabla \cdot \mathbf{v}_1)'$. The heavy dashed line indicates the areas where negative precipitation anomalies exceed 0.5 mm day⁻¹, south of 15°N, the land and Atlantic Ocean portions of which define the ESA and Atlantic ITCZ index regions, respectively. Contour interval is 10 W m⁻² below 20 W m⁻² and 50 W m⁻² above. Dark (light) shading denotes above (below) 10 W m⁻².

anomaly effect on moisture convergence anomaly dominates the precipitation anomaly, and we turn to the MSE budget for an account of how the descent anomaly is balanced. Because of the cancellation of dry static energy divergence with moisture convergence for convective regions, the anomalies of net divergence of MSE (Fig. 6b) are on the order of 20 W m⁻² over ESA and the Atlantic ITCZ (dashed outline region). From Fig. 3d, the cooling due to net flux anomalies $F^{\text{net}'}$ is only 2 W m⁻² over ESA and 10 W m⁻² over the Atlantic ITCZ. The adiabatic warming associated with strong descent anomalies is thus not balanced by radiation and surface heat exchange, especially over ESA, and advection terms must play a lead role.

Figure 7 displays the anomalous advection of temperature and moisture from the POSPAC run. The temperature advection anomalies over ESA are substantial (Fig. 7a). The moisture advection anomalies are even stronger, reaching up to 25 W m⁻², and are concentrated near the southeast margin of the convection zone. Both provide a negative, "cooling" tendency in the MSE budget, that is, they tend to create descent anomalies. Analysis of $-\langle \bar{\mathbf{v}} \cdot \nabla(T, q') \rangle$ versus $-\langle \mathbf{v}' \cdot \nabla(\overline{T}, \overline{q}) \rangle$ shows that the former (Fig. 8) dominates in the region of interest. The anomalous cold and dry advection terms of Fig. 7 are well explained by the climatological winds acting on anomalous temperature and moisture gradients in Fig. 8.

This is of additional interest because it runs contrary to the general assumption of weak-temperature-gradient theory for the tropical atmosphere (Sobel and Bretherton 2000; Sobel et al. 2001; Neelin and Held 1987), in which spatial variations of tropospheric temperature are neglected when considering the thermo-



FIG. 7. Anomalies of horizontal advection of (a) temperature $-\langle \mathbf{v} \cdot \nabla T \rangle'$ and (b) moisture $-\langle \mathbf{v} \cdot \nabla q \rangle'$. Contour interval is 5 W m⁻², with dark (light) shading above (below) 5 W m⁻².



FIG. 8. The anomalies of horizontal advection due to climatological winds acting on anomalous gradients of (a) temperature – $\langle \mathbf{v} \cdot \nabla T \rangle'$ and (b) moisture – $\langle \mathbf{v} \cdot \nabla q \rangle'$. Contours and shading are the same as in Fig. 7.

dynamics of the tropical circulation. We have found other cases of tropical teleconnection in which temperature gradient anomalies are not crucial for setting descent anomalies (SN02). Here, the importance of temperature gradients reinforces our conclusions regarding the complexity of descent anomalies. We note that while the QTCM temperature gradient across South America is smoother than in the reanalysis, temperature anomalies drop substantially from the Pacific to the Atlantic in observations.

To compare the difference between land and ocean regions, Table 1 may be summarized as follows. Over the ESA region, about 80% of the total MSE convergence anomalies are contributed by anomalous advection of temperature and moisture. The net flux anomalies at TOA, at 1 W m⁻², account for only 6% of the descent anomaly as measured by MSE convergence. The residuals (about 13%) are due to transient variability and diffusion anomalies. Net surface flux is zero as expected for land. Over the Atlantic ITCZ, moisture advection anomalies remain important in contributing to anomalous descent (27% of the MSE divergence term) while temperature advection anomalies drop to a 10% contribution. The small role of TOA anomalies remains similar (1 W m^{-2} , or 5%). The obvious difference is the importance of surface flux anomalies, which jump to the leading role at 56% in this fixed SST case, mainly due to downward shortwave flux.

For the land versus ocean regions, the flux contributions in Table 2 may be viewed as follows. In both regions, the increased OLR due to warm atmospheric temperature anomalies is small and is partly cancelled by the moisture contribution to OLR. The large cloudinduced downward surface shortwave anomaly $S_s^{\downarrow \prime}$ is of similar size over ESA and Atlantic ITCZ, tending to heat the surface. However, over land this term plus downward longwave and any tropospherically induced evaporation anomaly are necessarily cancelled by changes in upward longwave, evaporation, and sensible heat as land surface temperature compensates. The net solar plus longwave CRF contribution at TOA is also similar in size over the ESA and Atlantic ITCZ regions, but while over land the net TOA effect is the most relevant, over the ocean it is more productive to do the accounting differently. The CRF shortwave anomaly at surface and TOA is similar since there is little shortwave absorption. The CRF longwave anomaly contribution to $F^{net'}$ is thus left as a strong cloud radiative feedback over ocean (with prescribed SST). Because the increase in downward solar flux is lost to the surface, the increased OLR tends to cool the atmosphere, enhancing descent anomalies.

In Table 2, contributions to longwave radiation by cloud and tropospheric temperature and moisture, respectively, are provided with two measures side by side: the TOA forcing, which is more relevant to land, and the net forcing on the tropospheric column, which is more relevant over ocean (for fixed SST or short time scales). The net CRF, including the sum of shortwave plus longwave effects, is provided in similar format. When comparing the same term, such as $R_{T_{t}}^{\uparrow'}$ between the two regions, the differences tend to be modest. The more striking contrast, indicated by boldface type, is when comparing, say, $R_{Tt}^{\uparrow'}$ over land to $R_T^{\text{net}'}$ over ocean, associated with the switch to nonzero net surface flux. The contribution of temperature effects on longwave viewed this way roughly doubles and is reinforced, rather than opposed, by the moisture contribution to longwave. Comparing CRF' to CRF^{net'} between the regions offers a similar contrast.

Of the surface flux anomaly contributions in the Atlantic ITCZ, evaporation anomalies are relatively modest because several effects compete, as elaborated in section 7. Shortwave is by far the largest surface contribution and thus a leading source of surface warming.

6. MSE analysis with cloud radiative feedbacks as a modification to the effective static stability

a. Manipulation of cloud radiative effects

As seen in Fig. 4 and Table 2, cloud radiative forcing contributions are individually large terms, and they behave differently over land and ocean. Furthermore, cloud effects depend strongly on descent/precipitation anomalies and thus are not a forcing but a feedback. Fortunately, there is a manipulation of cloud radiative effects that provides insight into the nature of this feedback. ZN99 presented the land case under different notation and SN02 presented the ocean case. Here we contrast the land and ocean cases under uniform notation and quantify the approximations involved. Versions have been used in Bretherton and Sobel (2002, hereafter BS02) and Sobel and Gildor (2003, hereafter SG03).

Both solar and longwave feedbacks involving deep clouds and the associated cirrostratus/cirrocumulus (CsCc) have a strong linkage to convection and thus to precipitation. The CRF anomaly at TOA CRF'_t and the net cloud radiative forcing anomaly on the tropospheric column CRF^{net'} can thus be approximated by

$$\operatorname{CRF}_{t} \approx C_{t}P', \quad (\text{land})$$
 (11)

$$CRF^{net'} \approx C_{net}P'$$
, (ocean). (12)

Note that while Eqs. (11) and (12) can be estimated for either land or ocean, for land regions, only the top-ofatmosphere radiative fluxes count in the MSE budget, while over the ocean, in presence of nonzero net surface flux, the net forcing at TOA minus surface is important. Thus Eqs. (11) and (12) are used in the MSE budget over land and ocean, respectively, as noted in brackets.

For the parameters in this model,

$$C_t \approx -0.05, \quad C_{\text{net}} \equiv (C_s + C_t) \approx 0.12.$$
 (13)

These linearizations depend on basic state. For shortwave, mean cloudiness and surface albedo are important. For longwave, nonlinearity between cloud, T, and q effects is nonnegligible, which is why the cloud, T, q, and T_s contributions to $R_t^{\uparrow \prime}$ in Table 2 sum only approximately to the total. The value of C_t given is appropriate for ESA; $C_t \approx -0.7$ would apply over the Atlantic, where clouds shield a lower albedo surface. Because C_t results from cancellation of large individual longwave and shortwave terms, it is more sensitive than C_{net} . Longwave top and surface contributions individually tend to be more sensitive than is their sum in $R^{\text{net}'}$.

The magnitude of C_{net} is larger than C_t , consistent with the difference between CRF^{net'} and CRF'_t in Table 2. This is because net solar flux at the surface is taken to be absorbed (temporarily) by the ocean surface layer, and cloud longwave heating dominates C_{net} . The magnitude of C_t is much smaller than the longwave or shortwave contributions, due to cancellation between them (Kiehl and Ramanathan 1990). The sign of C_t implies a small net cooling in the TOA cloud radiative forcing when precipitation increases. This occurs because C_t is dominated by deep cloud and CsCc for which shortwave reflection is slightly larger than TOA longwave effects.

We now explore the consequences of Eqs. (11) and (12) for the MSE budget. Using the moisture Eq. (A2) to rewrite P' in Eqs. (11)–(12) as moisture convergence and other terms, and substituting for the CRF in the MSE budget [Eq. (5)] yields an approximate MSE budget:

$$(M_{\rm eff} \nabla \cdot \mathbf{v}_1)' + \langle \mathbf{v} \cdot \nabla T \rangle' + (1+C) \langle \mathbf{v} \cdot \nabla q \rangle' = F_{\rm eff}^{\rm net'}.$$
(14)

An effective moist stability M_{eff} that *includes cloud* radiative feedbacks has been defined as

$$M_{\rm eff} = M - M_{\rm CRF} = M - CM_q,\tag{15}$$

$$C = C_t$$
, land; or $C = C_{net}$, ocean. (16)

For the ESA region in the results here, $(M_q, M) \approx (19.6, 9.2) \times 10^6$ J m⁻², so the land value of $M_{\rm CRF} \approx -0.11M$. The negative value occurs because the net TOA CRF warming when cloud cover is reduced tends to increase adiabatic warming by descent, thus slightly increasing the effective moist static stability. Even though a heating occurs in a region that is warm, this is not a conventional positive feedback but simply a factor opposing descent.

For the Atlantic ITCZ region in this model, (M_a, M) \approx (19.0, 9.7) \times 10⁶ J m⁻², so the ocean value of $M_{\rm CRF}$ $\approx 0.24M$, that is, the effective stability is reduced by almost a quarter relative to the moist value. In the YCN98 observation-based estimates, M is smaller relative to M_a than is simulated here for the Atlantic sector. Thus the role for cloud feedbacks is potentially larger than the already substantial role indicated in these results. The oceanic effective moist stability is reduced because the radiative cooling associated with decreased deep convective cloud opposes adiabatic warming in the descent regions. As indicated in Table 2, the shortwave anomaly largely passes through the atmosphere into the ocean $(S_s^{\downarrow\prime} - S_s^{\uparrow\prime} = 10.7 \text{ W m}^{-2}$ in the Atlantic region), leaving only a small absorption heating (0.1 W $m^{-2)}$ in the atmosphere. Downward surface longwave flux is slightly reduced ($R_{clds}^{net'} = -R_{cldt}^{\uparrow'} - R_{cld}^{net'} = -1.3$ W m^{-2} in the Atlantic ITCZ due to cloud) when cloud fraction is reduced. The increased cooling by OLR (7.2 W m⁻²) dominates, yielding $R_{cld}^{net} = -5.9$ W m⁻² in Table 2. The cloud radiative feedbacks thus are large contributions to both the net surface heating and the tropospheric cooling that enhances the suppression of convection.

b. Effective flux forcing

The effective flux forcing in the modified MSE budget [Eq. (14)] after rearrangement of CRF terms is approximately

$$F_{\text{eff}}^{\text{net}\prime} = \epsilon_{Tt}T_1' + \epsilon_{qt}q_1' + \epsilon_{T_st}T_s', \quad \text{land}$$
$$= \epsilon_T T_1' + \epsilon_q q_1' + \epsilon_{T_s}T_s' + (1+C)E' + H' \quad \text{ocean.}$$
(17)

As in Eq. (16), the land case takes into account the zero net surface flux condition, while the ocean case includes nonzero net surface flux, as for fixed SST or for an ocean surface layer that has not equilibrated. Here $\epsilon_T = \epsilon_{Tt} + \epsilon_{Ts}$ is the longwave cooling rate of the troposphere per unit temperature, given by TOA and surface contributions ϵ_{Tt} and ϵ_{Ts} and similarly for moisture longwave coefficients $\epsilon_q = \epsilon_{qt} + \epsilon_{qs}$; ϵ_{Ts} is the coefficient for tropospheric absorption of longwave radiation due to surface temperature anomalies T'_s ; ϵ_{Ts} is the (small) coefficient of TOA contribution of T'_s to OLR; and T_1 and q_1 are projections of temperature and moisture on their respective typical vertical profiles (ZNC00). Nonlinear cross terms in the radiation (notably between cloud and moisture) are neglected in Eq. (17).

In the ocean case, the addition of downward longwave losses to the surface results in substantially stronger radiative cooling effects than in the land case. As noted in Table 2, the cooling $R_T^{\text{net}'}$ associated with ϵ_T is roughly double its TOA counterpart, $R_{TI}^{\uparrow_I}$, associated with ϵ_{TI} .

Furthermore, evaporation in Eq. (17) can be linearized as $E' = \epsilon_E(q'_1 - \partial_{T_s}q_{sat}T'_s) + (\partial v E)V'$, which has a damping effect toward SST (aside from effects of wind speed anomalies V'). However, all of the surface flux effects that tend to enhance descent in the short term disappear when the ocean surface equilibrates. Some interesting effects of shortwave feedbacks in the presence of such an ocean surface layer lag have been noted in SG03.

c. Modified MSE budgets

Table 3 shows the area-averaged MSE budgets when CRF is incorporated into the effective moist static stability, using the land case of Eqs. (16)–(17) for ESA and the ocean case for the Atlantic. The effective MSE divergence is increased by 9% in the ESA region, compared to $(M\nabla \cdot \mathbf{v}_1)'$ in Table 1. The relative contribution of the various terms is only modestly changed in ESA, since the M_{eff} differs from M by only 11%. With the cloud radiative effect absorbed in the effective moist static stability, the net flux anomalies contribute

TABLE 3. Area averaged MSE budget anomalies using the "effective moist static stability" $M_{\rm eff}$ and corresponding modifications to other terms when cloud radiative effects are rewritten as a feedback. Regions and units are as in Table 1. Note that the ESA region uses the land case and the Atlantic ITCZ region uses the fixed-SST ocean case described in Eqs. (11)–(17).

	ESA region	Atlantic ITCZ
$\overline{(M_{\text{off}} \nabla \cdot \mathbf{v}_1)'}$	-21.9(100%)	-17.0(100%)
$-\langle \mathbf{v} \cdot \nabla T \rangle'$	-8.0 (37%)	-2.3 (14%)
$-(1+C)\langle \mathbf{v}\cdot \nabla q\rangle'$	-8.3 (38%)	-6.6(39%)
$F_{\rm eff}^{net'}$	-2.4 (11%)	-6.0 (35%)

slightly more to the MSE divergence anomalies in this land case.

Over the Atlantic ITCZ, the effective MSE divergence $(M_{\text{eff}}\nabla\cdot\mathbf{v}_1)'$ is reduced by 23% compared to $(M\nabla\cdot\mathbf{v}_1)'$. The moisture advection contribution to effective MSE divergence anomalies is increased to 39% of $(M_{\text{eff}}\nabla\cdot\mathbf{v}_1)'$ in Table 3, while that of net flux anomalies is decreased to 35%. This is because the cloud effects that were the largest contribution to net flux anomalies in this ocean case have been taken into account in M_{eff} . This suggests that mechanisms related to $\langle \mathbf{v} \cdot \nabla_q \rangle'$ have comparable importance to noncloud fluxes. From Table 2, noncloud radiative terms $(R_T^{\text{net}'} + R_q^{\text{net}'} = -5.2 \text{W m}^{-2})$ are the largest contributor of effects from wind speed anomalies and q', each of which is individually substantial.

7. Hypothesized mechanisms for descent anomalies

a. QE-mediated effects

Within convection zones, temperature in the free troposphere and low-level moisture interact through convection. Under convective QE, it is hypothesized that convection acts on a relatively fast time scale to remove the buoyancy available to convective elements, of which convective available potential energy (CAPE) is one measure. This tends to establish a relationship among lower-tropospheric moisture and the column temperature profile. When the free troposphere is warmed by wave dynamics, this implies adjustment of low-level moisture. We note that this can hold for parameterizations that do not explicitly use convective QE closures or for conditions that depart from QE, so long as warming aloft tends to increase low-level moisture and temperature through convection. CN04 referred to this as CAPE mediated, but since drying or moistening in the lower troposphere above the ABL

can also impact convection, we use the more general term "QE mediated."

This QE-mediated pathway is common to at least three of the mechanisms below: the upped-ante mechanism (NCS03), QE-mediated surface flux mechanisms (CS02), and the M' mechanism. Quasi equilibrium also affects the longwave contribution by q'. All QE-mediated mechanisms can be interrupted simultaneously (experiment "CLIMCAPET") by suppressing the effect of tropospheric temperature anomalies in the convective parameterization. The climatological tropospheric temperature is passed to the convection scheme, so teleconnected warming cannot directly produce low-level moisture anomalies via convection.

Figure 9a shows POSPAC moisture anomalies in the QTCM. In the main ENSO region, moisture is high due to warm SST anomalies. Moisture also tends to be high within convection zones through other parts of the Tropics, including tropical South America and the Atlantic ITCZ. Figure 9b shows the anomalies in the QE value of moisture; that is, the difference between the values that the convection scheme adjusts toward given the temperature in the control and POSPAC runs, re-



FIG. 9. (a) Moisture anomalies for the POSPAC experiment relative to control; (b) anomalies of the convective equilibrium value of moisture computed from control and POSPAC temperatures; and (c) the difference between these, showing departures from QE. Specific humidity is averaged from 1000 to -850 hPa in units of grams of water per kilograms of air. Contours are -0.5, -0.1, 0, 0.1, 0.2, 0.5, 1, and 2 g kg⁻¹, and shading signifies above 0.2 in (a) and (b) and below -0.2 in (c).

spectively. This looks very much like the temperature anomaly and is considerably smoother than the actual moisture field. The difference between the actual moisture anomaly and this QE anomaly (Fig. 9c) indicates departures from QE. In the eastern equatorial pacific, departures are large because moisture is not constrained by convection under normal conditions. In the region along the southeastern edge of the South American convection zone and the Atlantic ITCZ, there are substantial negative departures from QE. These are induced by moisture advection from the adjacent drier, nonconvecting regions by the upped-ante mechanism, discussed below. Within the Atlantic ITCZ, departures of the moisture anomalies from QE are smaller but not entirely negligible, even for the short 2-h convective time scale used in this model. The QE value computed here is based on the mean temperature, not on the mean of the QE values, so transients potentially contribute to these departures.

b. Radiative cooling (traditional mechanism)

While it has been traditional to assume that descent anomalies are due to radiative cooling due to warmer temperatures, the flux anomaly contributions in Table 2 make it clear that the OLR anomalies associated with temperature change are small, less than 3 W m⁻². We may anticipate that radiative cooling will have some effect but that other mechanisms associated with larger terms in the MSE budget will dominate on regional scales, at least over land. Over ocean, the downward longwave anomalies associated with T' and q' [Eq. (17)] contribute additional cooling to the atmosphere (on time scales less than SST equilibration). These roughly double the radiative damping over ocean during periods when SST is in disequilibrium (see Table 2).

c. The upped-ante mechanism

In convective regions, an increase in tropospheric temperature would tend to decrease parcel stability unless the boundary layer MSE also increases. In NCS03 terminology motivated by a poker analogy, this "ups the ante" for the amount of moisture a region must have to be able to compete for available energy and moisture supply and continue to precipitate. In regions where there is no other impact on the moisture or energy budget, the boundary layer moisture simply increases and precipitation continues. The moisture supply required for the increase is small and can be met by a small, temporary (on the order of the convective time scale) decrease in precipitation that does not persist after the moisture returns to convective QE with the warmer troposphere. However, because moisture is not being increased in neighboring nonconvective regions, moisture gradients are created. In regions where there is low-level inflow of air from a nonconvective region, a $\mathbf{v} \cdot \nabla q'$ moisture advection term creates a drying effect in the margins of the convective zone. We have seen diagnostically (Fig. 7b; Tables 1 and 2) that this term is a large contributor to the anomaly MSE budget over much of the ESA and Atlantic ITCZ region. We test this pathway in two experiments, one suppressing anomalies of moisture advection and the other suppressing the temperature anomalies within the convective scheme.

d. M' mechanism

Quasi-equilibrium increases in moisture during warming can act to reduce the gross moist stability M in convection zones. CN04 found that this tends to be balanced by increased convergence, yielding a contribution to precipitation $P' = \nabla \cdot \overline{\mathbf{v}}_1(-(\overline{M}_a/\overline{M})M' + M'_a)$ in Eq. (10), and that this can be important in greenhouse warming experiments. Contributions are largest where mean convergence is large. Evaluation of the contribution of $M' \nabla \cdot \overline{\mathbf{v}}_1$ to $(M \nabla \cdot \mathbf{v}_1)'$ in results here shows that this contributes most strongly in the Amazon and is weaker over the Atlantic where, in this model, the mean convergence zone is weaker. Averaged over ESA, the contribution of $M' \nabla \cdot \overline{\mathbf{v}}_1$ to $(M\nabla \cdot \mathbf{v}_1)$ in Table 1 is about -4.7 W m⁻², or 23% of the negative anomaly. Over the Atlantic ITCZ, the contribution is less than 4%. The effect of this is to reduce the descent and rainfall anomalies in ESA, as may be seen from Eq. (10). The QE-mediated moisture increase reduces M, which partly offsets the need for descent anomalies in balancing other terms.

e. Moist wave decay mechanisms

The temperature signal in Fig. 2 is suggestive of a Kelvin wave decaying eastward, and the MSE budget suggests a significant role for gradients of temperature and moisture. Some consideration of moist Kelvin-like wave behavior is useful in interpreting these results. A simple case of the momentum equation for baroclinic mode zonal wind anomalies u'_1 has the form

$$\partial_t u_1' + \overline{u}_u \partial_x u_1' + \kappa \partial_x T_1' = F_u \tag{18}$$

for a solution with negligible v', where \overline{u}_u is a suitable projection of the mean wind, and F_u is a projection of the vertical momentum flux that includes a dependence on surface drag and tends to act as a damping. Terms and assumptions are given in the appendix.

In a time-dependent setup, an atmospheric internal Kelvin wave packet must propagate information eastward from the Pacific forcing region across South America and the Atlantic, interacting with moist convection as it goes and spinning up other contributions that in an inviscid problem would project on Rossby waves. Here we are concerned with the near-steady, eastward-decaying solution that is left by this spinup, which is not purely a moist Kelvin wave but is similar enough to refer to as a "moist Kelvinoid solution." In steady state, with small F_u , the dominant balance in Eq. (18) is between momentum advection and the baroclinic pressure gradient term $\kappa \partial_x T'_1$. This creates a relationship between anomalies of wind and temperature that depends on the mean wind.

Combining steady Eq. (18) with a similar form of Eq. (14) and using the QE temperature–moisture relation yields a moist wave equation of the form

$$(c_{\rm eff}^2 - \overline{u}_u \overline{u}_h) \partial_x u_1' = \kappa F_{\rm eff}^{\rm net\,\prime} - \overline{u}_h F_u', \tag{19}$$

where $c_{\rm eff}$ is an effective moist phase speed proportional to $M_{\rm eff}^{1/2}$. The u_h term comes from $\langle \overline{\mathbf{v}} \cdot \nabla T' \rangle$ + $\langle \overline{\mathbf{v}} \cdot \nabla q' \rangle$ in the MSE Eq. (5), and \overline{u}_h is a vertical projection of the mean wind in these terms under QE. Both \overline{u}_h and \overline{u}_u are easterly here, so the effective mean wind for the wave equation is $(\overline{u}_{h}\overline{u}_{u})^{1/2}$. The effective flux $F_{\rm eff}^{\rm net}$ is that given in Eq. (17) and Table 3 when CRF associated with convergent motions is moved into the effective moist stability $M_{\rm eff}$. The terms on the rhs tend to act as a damping and include noncloud radiative cooling and evaporation per Eq. (17) and surface drag (see the appendix). The form (19) is useful for the current discussion; elimination of the T'_1 dependence in $F_{\rm eff}^{\rm net'}$ requires that ∂_x be taken, so the wave equation is second order, with one solution eliminated by y-boundary conditions.

Because of partial cancellation between the moist Kelvin wave phase speed and the mean wind, seemingly small damping terms can create substantial divergence anomalies and zonal temperature gradients. This may be interpreted by considering the spinup of the steady solution-because the wave propagates more slowly against the easterly flow, the damping terms have longer to act in a given distance of propagation. The slow moist phase speed is important, since the closer $c_{\rm eff}$ is to $(\overline{u}_h \overline{u}_u)^{1/2}$, the greater the effect. This is thus a subresonant moist Kelvinoid solution. The $\overline{\mathbf{v}} \cdot \nabla T'$ and $\overline{\mathbf{v}} \cdot \nabla q'$ terms that prove to be substantial in the MSE budget appear as the $\overline{u}_{h}\overline{u}_{u}$ term on the lhs of Eq. (19), creating descent in a manner that interacts with other terms. Radiative damping is also seen as part of a moist wave decay contributing to the gradients. Furthermore, the F_u damping term on the rhs of Eq. (19) suggests that surface drag, even if relatively small, may

have an impact on a subresonant moist Kelvin wave. This latter hypothesis can be tested in the model context by suppressing anomalies of surface drag (experiment "CLIMDRAG").

One caveat on the simple case that yields Eq. (19) is that horizontally uniform QE assumptions are made that greatly reduce the upped-ante mechanism. In regions where the mean flow crosses a boundary from a nonconvecting to a convecting region, the upped-ante mechanism operates quite differently than is captured in Eq. (19).

Finally, this simple prototype provides a means of viewing differences in land versus ocean balances and suggests the importance of cloud radiative effects over oceans. For the ocean case of $M_{\rm eff}$ [Eqs. (15)-(16)], cloud effects reduce $M_{\rm eff}$ by almost 25% compared to what would happen if cloud radiative feedbacks were not active. When the partial cancellation between mean wind and $M_{\rm eff}$ on the 1hs of Eq. (19) is taken into account, the overall descent anomaly response will be amplified relative to either effect acting alone. Thus the cloud radiative feedback interacting with wave effects may be inferred to contribute substantially to the descent anomalies in the Atlantic ITCZ. Furthermore, Eq. (17) implies that $F_{\text{eff}}^{\text{net}}$ on the rhs of Eq. (19) roughly doubles in magnitude over oceans because of surface fluxes (both noncloud radiative damping and evaporation), providing stronger driving of descent.

f. Anomaly wind mechanisms

In the case of tropical Pacific descent anomalies, remote wind anomalies are active players in creating descent anomalies through $\mathbf{v}' \cdot \nabla \overline{q}$, $\mathbf{v}' \cdot \nabla \overline{T}$, and wind anomaly effects on evaporation. In the current study, wind anomalies are not associated with leading terms in the MSE budgets. Nonetheless, it is worth examining the impact of teleconnected wind anomalies on the ESA and Atlantic precipitation anomalies. Experiment "CLIMWIND" tests this by suppressing the effect of wind anomalies in the thermodynamics by using climatological winds in the advection and surface heat fluxes.

g. Troposphere/SST disequilibrium

As seen in Eq. (8), mechanisms associated with surface fluxes operate only while the SST is in disequilibrium with the tropospheric teleconnection. Mixed layer experiments (section 8b) are used to determine, first, to what extent the fixed SST ocean case approximates a mixed layer ocean response for the ENSO onset time scales here and, second, the response in absence of disequilibrium surface flux, by seeking the equilibrium solution for the same teleconnection forcing.

8. Tests of hypothesized mechanisms

a. Tests by intervention in model dynamics

The experiments suppressing terms associated with hypothesized mechanisms are shown in Fig. 10, which may be compared to the precipitation anomaly of the control POSPAC run in Fig. 2a. MSE budgets averaged over the ESA and Atlantic ITCZ regions are summarized in Table 4, which may be compared to the control run anomalies in Table 1. The figures tend to emphasize the impacts on the parts of each region with the strongest anomalies, while the regional averages in



TABLE 4. Area-averaged MSE budget anomalies as in Table 1, but for the mechanism-testing experiments: CLIMRADT (no anomalies of longwave radiation due to temperature perturbations); CLIMGRDT (no anomalies of temperature gradient in advection); CLIMGRDQ (no anomalies of temperature gradient in advection); CLIMDRAG (no anomalies of surface stress); CLIMCAPET (no anomalies of temperature in convective parameterization); and CLIMWIND (no wind anomalies in surface fluxes or advection). Square brackets give the changes (W m⁻²) from the control run values of Table 1 for the MSE divergence term and the one or two largest contributing changes (boldface for MSE divergence changes exceeding 30% of the control value).

	ESA region	Atlantic ITCZ
CLIMRADT		
$(M\nabla \cdot \mathbf{v}_1)' - \langle \mathbf{v} \cdot \nabla T \rangle' - \langle \mathbf{v} \cdot \nabla q \rangle'$	-17.1 [3.4] -7.2 [0.8] -8.5	-20.5 [1.6] -2.1 -5.8
	1.2 [2.5]	-10.1 [1.2]
CLIMGRDQ		
$\begin{array}{l} (M \nabla \cdot \mathbf{v}_1)' \\ -\langle \mathbf{v} \cdot \nabla T \rangle' \end{array}$	-10.1 [10.4] -6.2	-19.8[2.3] -1.8
$\frac{-\langle \mathbf{v} \cdot \boldsymbol{\nabla} q \rangle'}{F^{\mathrm{net}'}}$	0.6 [9.3] -2.5	-0.7 [5.2] -13.9 [-1.3]
CLIMCAPET		
$(M\nabla \cdot \mathbf{v}_{1})' - \langle \mathbf{v} \cdot \nabla T \rangle' - \langle \mathbf{v} \cdot \nabla q \rangle' F^{\text{net}'}$	-10.2 [10.3] -5.5 [2.5] -1.2 [7.5] -2.8	-9.8 [12.3] -1.4 -2.5 [3.4] -4.6 [6.7]
CLIMGRDT		
$(M\nabla \cdot \mathbf{v}_1)' - \langle \mathbf{v} \cdot \nabla T \rangle' - \langle \mathbf{v} \cdot \nabla q \rangle' F^{\text{net}'}$	-12.7 [7.8] -0.6 [7.4] -8.5 -1.9	-20.5 [1.6] 1.1 [3.3] -6.7 -12.6 [-1.3]
CLIMDRAG		
$(M\nabla \cdot \mathbf{v}_{1})' - \langle \mathbf{v} \cdot \nabla T \rangle' - \langle \mathbf{v} \cdot \nabla q \rangle' F^{\text{net}'}$	-13.4 [7.1] -3.7 [4.3] -4.3 [4.4] -1.7	-19.8 [2.3] 2.3 [4.6] -5.1 -14.4 [-3.1]
CLIMWIND		
$(M \nabla \cdot \mathbf{v}_1)' - \langle \mathbf{v} \cdot \nabla T \rangle' - \langle \mathbf{v} \cdot \nabla q \rangle' F^{\text{net}'}$	-18.8 [1.7] -8.8 [-0.8] -5.2 [3.5] -1.2	-23.1 [-1.0] -2.5 -4.7 [1.2] -15.2 [-3.9]
MLEQBM		
$(M\nabla \cdot \mathbf{v}_{1})' - \langle \mathbf{v} \cdot \nabla T \rangle' - \langle \mathbf{v} \cdot \nabla q \rangle' F^{\text{net}'}$	-17.5 [3.0] -5.4 [2.6] -7.1 [1.6] -2.9	-15.3 [6.8] -1.5 -5.5 -2.2 [9.1]

FIG. 10. Precipitation anomalies for experiments (a) CLIMRADT, (b) CLIMGRDQ, (c) CLIMCAPET, (d) CLIMGRADT, and (e) CLIMDRAG. Contouring is the same as in the control in Fig. 1a, except 0.2 mm day⁻¹ contour is added in (c).

Table 4 include impacts of smaller anomalies over the whole region. The region in which target terms are artificially set to climatological values (outlined on the figures) covers tropical South America and Atlantic and extends to $\pm 25^{\circ}$ and far to the east (60°E) to avoid potential edge effects near the regions of interest.

A caveat common to all these experiments is that when one mechanism is suppressed, the others tend to counteract the change. This is particularly noticeable over the ocean region because surface fluxes operate in addition to the mechanisms that operate over land. In terms of the Kelvinoid wave decay prototype [Eq. (19)], the experiments alter an operator on u', T' rather than altering a forcing. If the alterations were small relative to the operator of the control case, effects would be approximately additive, but the alterations are large. The experiments are thus not additive but, suitably interpreted, can give a good sense of the importance of a mechanism. We also note that the diffusion terms are not shown in the budgets; their contributions, while secondary, can be nonnegligible, and so the terms shown do not necessarily cancel exactly.

Experiment "CLIMRADT," similar to one conducted in SN02, suppresses the longwave cooling due to temperature anomalies within the outlined region (cloud and moisture impacts on radiation are still active). Comparing the precipitation (Fig. 10a) to those of the standard POSPAC run (Fig. 2a), changes are relatively small. In regional average MSE budget anomalies (Table 4), the total MSE divergence anomalies are 16% smaller in the ESA region and 7% smaller in the Atlantic ITCZ region. Thus the traditional radiative cooling mechanism plays a rather modest role in creating these regional precipitation anomalies. In the ESA region, changes in temperature gradient enhance the direct effect of $F^{\text{net}'}$.

In the "CLIMGRDQ" experiment, the moisture gradient within the $\mathbf{v} \cdot \nabla q$ advection term is replaced by its climatological value. This tests the effects of the uppedante mechanism, which is mediated by this term. Note that we only suppress the moisture gradient anomalies in advection; changes in $\langle \mathbf{v} \cdot \nabla q \rangle'$ are small but not zero because of the contribution of anomalous winds and transient terms. There is a large impact on descent anomalies over the ESA region. In Fig. 10b, only a small area with negative precipitation anomalies less than -0.5 mm day^{-1} remains. In ESA average MSE budget anomalies (Table 4), the reduction exceeds 50% (the remaining anomalies are largely due to coastal points) and the changes are unambiguously due to the moisture advection changes. Over the Atlantic, the reduction in anomalies is smaller. A sizable reduction in $\langle \mathbf{v} \cdot \nabla q \rangle'$ is partly offset by changes in surface fluxes. The loss of energy into the ocean helps to maintain the descent anomalies in the Atlantic.

Experiment CLIMCAPET tests all QE-mediated effects by specifying climatological tropospheric temperature values in the convective parameterization. This suppresses the upped-ante mechanism, the M' mechanism, and QE-mediated surface flux mechanisms, since warming aloft has no way of producing low-level moisture anomalies through convection. The negative precipitation anomalies are greatly reduced (Fig. 10c). Over the ESA region, the reduction relative to the control is comparable to the CLIMGRDQ run, consistent with suppressing the dominant upped-ante mechanism in each. Over the Atlantic ITCZ region, the reduction is the largest achieved in any of the experiments (Table 4), since two leading mechanisms are suppressed simultaneously. The remaining precipitation anomalies in ESA can be attributed largely to $\langle \mathbf{v} \cdot \nabla T \rangle'$ and tend to be broadly spread (see 0.2 mm day^{-1} contour in Fig. 10c). Remaining precipitation anomalies over the Atlantic are due to a combination of terms, including radiative damping and CRF, which maintain the still substantial negative $F_s^{net'}$ anomaly. The moisture anomaly pattern (not shown) is greatly altered from the control seen in Fig. 9. In particular, it lacks the increased moisture within the convection zones, confirming the importance of OE mediation in this feature, at least in this model. While moisture anomalies tend to be smaller in magnitude within the experimental region than in the control, significant negative moisture anomalies (which depart from convective equilibrium values) occur along the edge of the convection zone, associated with the remaining negative precipitation anomalies.

In experiment "CLIMGRDT," the temperature gradient within the $\mathbf{v} \cdot \nabla T$ advection term is replaced by its climatological value. In Fig. 10d, the negative precipitation over ESA is substantially reduced. The descent anomaly (as measured by the MSE divergence term in Table 4) is reduced by 38% relative to the control. The negative precipitation anomalies over the Atlantic are affected less strongly, with only a 7% reduction in descent anomaly (Table 4). This is consistent with the diagnosed contributions of $\mathbf{v} \cdot \nabla T$ anomalies in the control (Table 1).

Experiment CLIMDRAG, motivated by the role of surface drag in Eq. (19), has surface stress anomalies suppressed in the target region. Noting from Eq. (18) that the surface stress anomalies must act by slightly altering the balance between wind anomalies and temperature gradient anomalies, and that the latter are more active in the MSE budget, one anticipates similarities to run CLIMGRDT. The negative precipitation anomalies in Fig. 10e indeed bear a resemblance to those in Fig. 10d. The MSE budget (Table 4) shows smaller temperature advection anomalies over the ESA region than the control, as anticipated. They are reinforced by changes in moisture advection anomalies, which may be due to the fact that the temperature field changes and has implications for moisture via QE. Because the drag over ocean is considerably smaller than over land, the smaller changes in precipitation anomalies seen over the Atlantic are expected. However, the MSE budget (Table 4) shows that temperature advection anomalies over the Atlantic ITCZ change significantly relative to the control and are compensated for by surface flux changes. Overall, the indication is that drag, at least in this model, can play a secondary but nontrivial role in establishing descent anomalies.

The CLIMWIND experiment (with climatological winds in the advection and surface heat flux computations) is anticipated from the budgets not to produce as large impacts as other experiments. This proves true in precipitation (figure not shown) and descent anomalies (Table 4), although there is a slight decrease in descent anomaly in ESA and a slight increase in the Atlantic ITCZ. The positive evaporation anomalies in the Atlantic in the control (Table 2) are mostly caused by wind anomalies, which overcome an opposite tendency by QE effects. Suppressing these does cause a nonneg-ligible impact in the surface fluxes in CLIMWIND, partially compensated for by other terms, and this creates the slight increase in the descent anomaly.

b. Tests of mixed layer ocean effects

Given that surface fluxes play an important role in the Atlantic ITCZ region, tests using a mixed layer ocean provide an additional step in assessing the role of ocean surface layer in the teleconnection response. Figure 11 shows an experiment using a mixed layer ocean, along with an experiment in which surface flux effects are suppressed by running the mixed layer to equilib-



FIG. 11. Precipitation anomalies for experiments with a mixed layer ocean. (a) Control forced by Jul–Nov 1997 POSPAC SST; (b) equilibrium case (20-yr average).

rium. To simplify the comparison, the July-November 1997 average SST anomaly is switched on in the POSPAC region at the beginning of July and then maintained. The anomaly is close to equilibrium by the second year (an equilibrium modulated by the seasonal cycle). The July-November average over 20 equilibrated years is used although interannual variability is relatively small. The response during 1997 is similar to the fixed SST case, indicating that a 50-m mixed layer can slow the ocean response enough to achieve substantial negative precipitation anomalies by surface heat fluxes. A 50-m layer is slightly deeper than the typical tropical Atlantic values in Saravanan and Chang (1999), so the surface flux effects may be slightly overestimated. A case (not shown) with 0.5-m mixed layer depth equilibrates within the first year to very similar values.

The equilibrium case may be compared either to the first year or to the fixed SST control; differences from the latter provide a limiting case and are used in Table 4 (last entry). An interesting feature is that despite near absence of surface heat fluxes, substantial negative precipitation anomalies remain (Fig. 11b). Only a 30% reduction in descent anomalies is achieved in the Atlantic ITCZ relative to the fixed SST case. The MSE budget terms (Table 4) indicate that the upped-ante mechanism is the leading effect in sustaining the negative precipitation anomalies in absence of surface fluxes. The remainder terms not included in the table are largely due to moisture diffusion, which acts similarly to advective inflow in preventing moisture rise in the margins of the convection zone via the upped-ante mechanism.

c. Comparison to other events

The question arises as to what extent mechanisms from the case study for the strong 1997 El Niño hold for ENSO events in general. MSE budget analysis of a run with specified observed SST indicates that overall the 1997 case provides a good prototype for other events in terms of the mechanisms that are active. With several mechanisms present, some quantitative variation in the contribution of each effect of course occurs. To illustrate the relationship to the 1997 results, Fig. 12 shows composite ENSO anomalies for precipitation and selected MSE budget terms for September-November (SON). The composite is for SON warm cases 1982, 1986, 1987, 1991, and 1994 minus cold cases 1983, 1984, 1985, 1988, and 1995. These are chosen based on the criterion that the magnitude of the Niño-3.4 SST index anomalies exceeds 0.5°C for SON. The composite SST anomaly (not shown) exceeds 2°C in the Pacific cold tongue and is less than 0.2°C in the northern tropical



FIG. 12. QTCM ENSO composite anomalies for SON of the mean of the warm ENSO events of 1982, 1986, 1987, 1991, and 1994 minus the mean of the cold ENSO events of 1983, 1984, 1985, 1988, and 1995. (a) Precipitation; contour is 1 mm day⁻¹ except for light shading below $-0.5 \text{ mm} \text{ day}^{-1}$. (b) MSE divergence $(M\nabla \cdot \mathbf{v})'$; contour interval is 4 W m⁻², with light (dark) shading below (above) $\pm 8 \text{ W m}^{-2}$. (c) Moisture advection $\langle \mathbf{v} \cdot \nabla q \rangle'$; contour interval is 2 W m⁻², with light (dark) shading below (above) $\pm 2 \text{ W m}^{-2}$. (d) Net flux into atmosphere $F^{\text{net}'}$; contour interval is 4 W m⁻² except for 10 W m⁻² above 20 W m⁻², with light (dark) shading below (above) $\pm 8 \text{ W m}^{-2}$.

and equatorial Atlantic and the southern tropical Atlantic except for a small region near the African coast. SON is chosen rather than July-November, which was used in the 1997 case, because this season better matches the typical onset of events, and with several events, statistical significance is easier to obtain in comparison to observations. Regression analysis on the Niño-3.4 index yields similar patterns and indicates statistical significance of all the main anomalies in the ESA and Atlantic ITCZ regions of interest. The precipitation anomaly in Fig. 12a compares reasonably well at large scales to a similar composite of observed rainfall (not shown), although the ITCZ anomaly is broader in the model. The negative anomalies in the southern Caribbean are correct. The intrusion of negative precipitation anomalies into South America along the southeastward flank of the convection zone is captured, although in the model it is shifted slightly toward northeastern Brazil relative to the observed.

Figures 12a,b,c,d may be compared to Fig. 2a, Fig. 6b, Fig. 7b, and Fig. 3d of the 1997 case, respectively. The anomalies are generally smaller, as expected since the forcing is not as large, but we can assess the similarity of the relationship among contributing terms. The precipitation and MSE divergence anomalies inland in the Amazon are weaker relative to those nearer the coast than in the 1997 case. The anomalies in eastern equatorial South America and across the Atlantic ITCZ have generally similar spatial structure to those of 1997. The moisture advection is important in a similar region along the southern and eastern margin of the convection zone, indicating a similar role for the upped-ante mechanism, especially over land convective margins. Temperature advection (not shown) has a rather similar pattern to Fig. 7a over equatorial South America although it is slightly weaker in proportion to other effects. Tropospheric net flux anomalies $F^{net'}$ have a similar land–ocean contrast due to the role of surface fluxes. The pattern is generally similar although surface fluxes are slightly larger in relative importance over the Atlantic ITCZ than in the 1997 case. The region of $F^{net'}$ negative anomalies also covers a somewhat broader region in the northern tropical Atlantic.

Generally similar results hold for the December– February season. The location of the anomalies tends to move with the convection zones, but the contributing terms tend to recur in similar relationships, at least in this model context.

9. Overview and discussion

a. Overview

Teleconnections in the Tropics, and especially the important impacts on land precipitation, are complicated by interactions with convective heating, shortwave and longwave cloud radiative feedbacks, and land surface feedbacks. Traditional notions of descent anomalies being balanced by radiative cooling are shown here and in SN02 to not be leading balances in the regions of strongest precipitation and descent anomalies. Fortunately, using diagnostics based on the moist static energy (MSE) budget and convective quasi equilibrium (QE), it proves possible to understand the contributing mechanisms in considerable detail. Because some of the principles and mechanisms are spread over several papers, we summarize more broadly than the specific South American/Atlantic sector results.

In these moist dynamical interactions, three powerful simplifications come from the following:

- 1) *The gross moist stability (GMS) M.* Following NY94, YCN98, and NZ00, *M* permits convective heating feedbacks to be handled simply. It arises under certain approximations in the MSE budget and takes into account the partial cancellation of adiabatic cooling by convective heating as a reduced static stability, that is, a reduction in the energy per unit mass required to balance vertical motion.
- 2) The zero net surface flux condition over land (or for an ocean mixed layer if approximate equilibrium is established). The occurrence of $F_s^{\text{net}'} \approx 0$ has particularly useful implications within deep convection zones. As noted in NZ00 and ZN99, if QE holds sufficiently well, the moisture equation and dry thermodynamic equation are linked by convection so the MSE equation governs the tropospheric thermodynamics. The partition of $F_s^{\text{net}'}$ between evaporation, radiation, and sensible heat is of negligible importance to leading dynamical balances (and thus surface temperature T_s can be viewed as a by-product).
- 3) Cloud radiative feedbacks as a change in M. Although cloud radiative feedbacks constitute large terms in the budgets, CRF associated with changes in deep convection can be handled succinctly using a trick from ZN99, SN02, BS02, and SG03. For deep cloud feedbacks approximately proportional to precipitation, the largest contribution is in turn related to moisture convergence and thus acts mathematically like a modification to the gross moist stability, replacing M by $M_{\rm eff} = M - M_{\rm CRF}$. The required linearization of the radiative effects is found to work adequately in this model. Longwave feedbacks tend to reduce $M_{\rm eff}$. Over land surfaces, using the zero net heat flux condition, shortwave plus land surface feedbacks tend to increase $M_{\rm eff}$. The TOA cancellation between shortwave and longwave CRF results in modest effects on land region descent and precipitation anomalies, although they are important for land surface temperature. Over ocean, for time scales short enough that SST does not equilibrate (or for fixed SST), the cloud radiative feedbacks

provide a shortwave forcing to the ocean and a longwave feedback in the atmosphere. Absorbing CRF in $M_{\rm eff}$ implies that cloud feedbacks do not initiate vertical motion anomalies but act to amplify descent anomalies initiated by other mechanisms (i.e., CRF affects the amount of vertical motion per unit forcing). Over the Atlantic ITCZ, results here suggest that amplification by CRF contributes roughly 25% of the descent anomalies during ENSO onset.

The above simplifying principles serve to remove distractions and to handle some of the feedbacks, while analysis of the MSE budget shows which terms balance anomalous descent. Combining this budget analysis with considerations of wave dynamics and QE mediation permits causal pathways to be hypothesized. These can be tested numerically in experiments that artificially suppress terms involved in the mechanism.

We find a small "zoo" of mechanisms, many of which have overlapping steps in their pathways. Mechanisms are stated here for an El Niño warming case, but the mechanisms also tend to work for a La Niña case with signs reversed, for example, cool troposphere is in QE with reduced low-level moisture.

The main principles common to all mechanisms are as follows:

- Wave dynamics. Wave dynamics acts to reduce baroclinic pressure gradients that would otherwise be created by the local heating by surface fluxes in the source region, spreading warm temperatures. For the present case of El Niño influence on South America and the Atlantic, Kelvin-like wave dynamics modified by moist convection is the most relevant.
- 2) *QE mediation.* The direct effects of the wave dynamics are expressed by tropospheric temperature anomalies T' and by the associated wind anomalies \mathbf{v}' . Convective QE links moisture anomalies q' to T' within convection zones; we use the term QE mediation for mechanisms that depend on this effect. The teleconnected T' and \mathbf{v}' and the QE-mediated q' impact various terms in the MSE budget in particular regions. For instance, $\mathbf{v}' \cdot \nabla \overline{q}$ is important to Pacific descent anomalies (SN02), while QE q' effects act via moisture advection and evaporation. The temperature gradient determined by wave dynamical interactions also has consequences for descent anomalies.
- 3) GMS multiplier effect. Anomalies in the vertical motion determined by interplay with other terms in the MSE equation have disproportionate consequences for precipitation (as in ZN99). This is due to the ratio of moisture convergence to MSE divergence,

measured by the ratio of the gross moisture stratification to the gross moist stability M_a/M , which is considerably larger than one. If the MSE equation has a balance $\overline{M} \nabla \cdot \mathbf{v'}_1$ = term', where term' is any of the anomaly terms involved in creating the descent motion, then the moisture equation has a contribution to precipitation $P' = (\overline{M}_a / \overline{M})$ term'. We term this the GMS multiplier effect since it has the effect of taking a term in the MSE equation whose amplitude may be modest and turning it into a substantial effect on precipitation, on the order of 5 times as large (in W m^{-2}). In brief, a term in the MSE equation tends to be balanced by a relatively large convergent motion because M is small, and this produces precipitation by moisture convergence.

b. Mechanisms

In the South American/Atlantic problem, the following mechanisms have been found to be active. Note that the mechanisms are not additive or fully independent. See section 7 for additional description.

- 1) The upped-ante mechanism. The upped-ante mechanism described in NCS03 (and recapped in section 7) involves warmer temperatures in the free troposphere implying an increased "ante" in terms of the QE moisture required to maintain convection. This creates moisture gradients relative to nonconvective regions, and regional drought occurs by the $\overline{\mathbf{v}} \cdot \nabla q'$ term. This impacts vertical motion via the GMS multiplier effect, leading to a substantial precipitation reduction from a modest $\mathbf{v} \cdot \nabla q'$. In the simulations here, the upped-ante mechanism is the leading contributor to the precipitation reduction over the eastern equatorial South American region. It is also a major contributor over the Atlantic ITCZ region. In coupled ocean mixed layer experiments, as the SST equilibrates, it becomes the leading mechanism for the remaining anomalies.
- 2) Moist Kelvinoid wave subresonant in easterly flow. An eastward-decaying, steady, small v solution provides a prototype for understanding a number of interactions. The smaller effective moist stability $M_{\rm eff}$ in convection zones implies greater descent per unit cooling anomaly. Over ocean where $M_{\rm eff}$ is further reduced by cloud feedbacks, this is enhanced. Easterly mean wind tends to further enhance descent. The mean wind plays a more important role than in traditional solutions for two reasons: for steady solutions, momentum advection terms are the main balancing terms for baroclinic pressure gradients in the free troposphere; and the slow moist

wave speed, on the order of 10–15 m s⁻¹, implies that the zonal wind can be comparable to the phase speed. For an eastward Kelvin-like wave, the mean easterly flow can greatly slow the wave, allowing more time for the weak damping processes to act. The cancellation between \bar{u} terms, such as $\bar{u}\partial_x T'$, and the $M_{\rm eff}$ term that gives the phase speed yields greater gradients in the anomalies, that is, faster zonal decay in T' and greater descent and precipitation anomalies.

- 3) Radiative damping. A traditional assumption is that radiative damping due to warmer temperatures would be important to anomalous descent. However, in MSE budgets, radiative cooling associated with temperature is only a few watts per meter squared, considerably smaller than other terms, and at TOA it is partially cancelled by the contribution from moisture. Over ocean, the radiative damping effect is roughly doubled since loss occurs (temporarily) to the surface as well as TOA. In experiments in which anomalies of radiative cooling are suppressed, the overall precipitation anomaly pattern is changed only modestly. We note that this applies at the scale of regional precipitation anomalies; at larger spatial scale, where transport terms become smaller, radiative cooling contributions can be important.
- 4) Surface drag. Observed ENSO tropospheric temperature anomalies decrease considerably across South America (Fig. 1), and a smoother version occurs in the QTCM simulations. Observational and GCM-based estimates also show large ENSO surface stress anomalies across South America (Munnich and Neelin 2004). Although momentum advection terms such as $\overline{u}\partial_x u'$ play a leading role in the vertically integrated momentum budget, small damping by surface stress, communicated upward in the advection terms (as in Bacmeister and Suarez 2002), can potentially be important to the eastward decay over South America. This causes precipitation impacts via the temperature gradient and interacts with other mechanisms. In the QTCM, experiments suppressing anomalies of surface drag show a reduction in the eastward decay of the temperature field and a reduction of negative precipitation anomalies over South America.
- 5) *M' mechanism*. Moisture increases in convective regions in QE balance with tropospheric warming tend to increase M_q , thus reducing *M*. This tends to increase precipitation in climatological convergence zones. While important in global warming experiments (CN04), here it is a secondary but not negligible contributor over land, acting to reduce the

drought in the center of the Amazon convergence zone.

- 6) Anomaly wind mechanisms. Anomalous wind mechanisms occur via teleconnected v' contributions to moisture and temperature advection, v' · ∇(q, T), and to surface fluxes. In the case of surface fluxes, they fall under the category of troposphere–SST disequilibrium mechanisms below. While in the Pacific case these were leading contributors to precipitation anomalies, for ENSO impacts on South America and the Atlantic they appear to be secondary, though not negligible. The main effect in results here is to oppose negative precipitation anomalies in the Atlantic ITCZ.
- 7) Troposphere/SST disequilibrium (surface flux) mechanisms. In the Atlantic ITCZ region, the net surface flux anomaly is a leading MSE budget contributor, and the GMS multiplier effect acts on this to yield a substantial negative precipitation contribution. However, there are several mechanisms active in producing the net surface flux. As deep convection acts to bring the ABL and free troposphere into QE, surface fluxes act to bring the ABL and ocean surface layer into equilibrium (CS02). Cloud radiative feedbacks, especially shortwave, yield a substantial surface-warming tendency, while longwave radiative fluxes associated with temperature and moisture contribute a weaker tendency toward equilibrium with the teleconnected tropospheric warming. Additionally, teleconnections via wind create a contribution to surface flux anomalies. Of these, we infer from results here that the QE-mediated surface flux mechanism and shortwave feedback are of leading importance, along with the upped-ante mechanism.

Elaborating on the last item, mechanisms associated with surface fluxes have often been considered as a means of teleconnected warming of SST (e.g., Enfield and Mayer 1997; Klein et al. 1999; Lau and Nath 2001; CS02). The MSE budget makes clear that the flux warming the ocean is lost from the troposphere, and in deep convective regions this tends to be balanced by descent anomalies that yield negative precipitation anomalies. Once the ocean surface layer equilibrates with the troposphere, if ocean transport divergence anomalies are negligible, surface heat flux anomalies are small and effects of surface fluxes disappear in the MSE equation. In such an equilibrated state, SST can be viewed as a by-product of surface heat flux equilibrium much the same as in the land surface case. The proposed terminology, troposphere/SST disequilibrium, serves as a reminder that surface flux effects are

temporary (unless supported by ocean heat transport) and that the free troposphere is involved as well as the ABL. For the time scale of the onsetting El Niño during July-November, results from a 50-m ocean mixed layer experiment in the Atlantic are similar enough to a fixed-SST case that the latter can be used as a prototype, with the caveat that any surface flux mechanisms depend on ocean mixed layer heat capacity and $\partial_t T_s$. In experiments where the ocean mixed layer is permitted to equilibrate for the same teleconnection forcing, the upped-ante mechanism maintains a substantial negative precipitation anomaly over the Atlantic ITCZ despite absence of surface fluxes. Equilibration is essentially complete after 1 yr for a 50-m mixed layer. A rough estimate of the time scale for disequilibrium mechanisms to disappear is thus several months, unless ocean heat transport anomalies are induced.

To summarize leading mechanisms for precipitation and descent anomalies during El Niño for equatorial South America versus the Atlantic ITCZ, the uppedante mechanism is a leading mechanism over both, especially at the southeastern margin of the convective zone; surface drag and temperature gradients contribute over ESA; troposphere–SST disequilibrium mechanisms including radiative cooling and QE-mediated evaporation are important over the ocean; cloud radiative feedbacks are a significant amplifier over the ocean; easterly flow effects are a significant amplifier over both; and the GMS multiplier effect is a very important amplifier in all cases.

Acknowledgments. This work was supported under National Science Foundation Grants ATM-0082529 and DMS-0139666 and National Oceanographic and Atmospheric Administration Grants NA16-GP2003, NA05OAR4310007, and NA04OAR4310013. The subject of the present study was suggested by earlier work with N. Zeng, and discussions are gratefully acknowledged. Earlier versions of this work have previously been presented at the December 1998 American Geophysical Union, Union Session; the 1999 International Union of Geodesy and Geophysics; and the 2000 Canadian Meteorological and Oceanographic Society Plenary session. Thanks are due to J. Meyerson for graphics.

APPENDIX

Definitions and Derivations

Locations in the text of *definitions of main symbols* are summarized by the equation near which they appear. In order of appearance: h, s, T, q, and $\langle \cdot \rangle$ are

The vertically integrated thermodynamic equation and moisture equation from which Eq. (1) is derived are

$$\partial_t \langle T \rangle + \langle \mathbf{v} \cdot \nabla T \rangle + \langle \omega \partial_p s \rangle = \langle Q_c \rangle + S^{\text{net}} + R^{\text{net}} + H$$
(A1)

$$\partial_t \langle q \rangle + \langle \mathbf{v} \cdot \nabla q \rangle + \langle \omega \partial_p q \rangle = \langle Q_q \rangle + E, \tag{A2}$$

where Q_c and Q_q are convective heating and moistening, with other terms defined following Eq. (1). Precipitation $P = \langle Q_c \rangle = -\langle Q_q \rangle$. The gross moist stability has two contributions resulting from the vertical motion terms in Eqs. (A1) and (A2), respectively. Defining a typical vertical velocity profile Ω such that $\omega \approx \Omega \nabla \cdot \mathbf{v}_1$ yields $M \equiv M_s - M_q$ with $M_s = -\langle \Omega \partial_p s \rangle$ and $M_q = \langle \Omega \partial_p q \rangle$.

Equation (18) is derived from that in NZ00, which projects the primitive equation momentum equation onto velocity basis functions yielding one barotropic and one baroclinic velocity component. The advecting mean wind term $\overline{u}_{\mu} = \overline{u}_0 + (3/2) [\langle V_1^3 \rangle / \langle V_1^2 \rangle] \overline{u}_1$ includes contributions from projection of horizontal advection by barotropic and baroclinic components of the mean wind \overline{u}_0 , \overline{u}_1 [vertical structure constant and $V_1(p)$, respectively] and from vertical advection of the mean wind by $\partial_x u'_1$. The coefficient of baroclinic pressure gradients $\kappa = Rc_p^{-1}$ is the ratio of gas constant to heat capacity for air. The remainder momentum term on the rhs of Eq. (18) has the form $F_{\mu} = -([\langle V_1^3 \rangle / 2 \langle V_1^2 \rangle] \nabla \cdot \overline{\mathbf{v}}_1$ $+ \epsilon_1$) u'_1 , where ϵ_1 is proportional to the surface drag coefficient (see NZ00). The basic state is assumed independent of longitude; the basic state northward velocity \bar{v} is assumed small except in the mean divergence term $\nabla \cdot \overline{\mathbf{v}}_1$. Anomalies of v' are assumed negligible for a Kelvin-wave-like solution. The v equation is assumed dominated by geostrophic balance. Terms caused by interaction with the barotropic mode due to vertical shear and surface stress are neglected for heuristic purposes. In practice these are important. Easterly mean wind is assumed, which yields u'_1 of the same sign as T'_1 , and mendional decay at a scale combining $[(-\overline{u}_{u})/\beta]^{1/2}$ (for small F_u) with moist deformation radius $(c_{\text{eff}}/\beta)^{1/2}$. A full theory including interaction with the barotropic mode and spatially varying basic state and is beyond the scope of this paper.

In the MSE equation, a QE relationship $\hat{b}_1 q_1 = \hat{B}_1 T_1$

is assumed in Eq. (19). The advecting term in the MSE equation under these conditions and with CRF absorbed in M_{eff} and other terms is $u_h = \overline{u}_0[\hat{a}_1 + (1 + C)\hat{B}_1] + \overline{u}_1[\langle a_1V_1 \rangle + (1 + C)\langle b_1V_1 \rangle \langle B_1/b_1 \rangle]$. The effective phase speed $c_{\text{eff}}^2 = \kappa M_{\text{eff}}$ in Eq. (19) is equal to the actual moist Kelvin wave phase speed times $[\hat{a}_1 + (1 + C)\hat{B}_1]$; for the steady problem, it is simpler not to have to carry this extra factor, which acts as the QE moist heat capacity of the column in time derivative terms.

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