Chapter 7
Teleconnections into South America from the Tropics and Extratropics on Interannual and Intraseasonal Timescales

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Abstract This chapter presents the mechanisms and the most important effects of tropics-extratropics and tropics-tropics atmospheric teleconnections with South America on intra-seasonal to interannual time scales. The basis for theoretical understanding of teleconnections is reviewed, as well as the observed Southern Hemisphere interannual and intraseasonal variability. The observed teleconnections with South America are described, indicating linkages between the tropics and extratropics and between different regions in the tropics. Possible mechanisms of these linkages are presented and some teleconnections are analyzed with Influence Functions. Special emphasis is placed on the teleconnections associated with ENSO on interannual time scales and the Madden Julian Oscillation on intraseasonal time scales.

Keywords Teleconnections · South America · ENSO · Madden-Julian Oscillation

7.1 Introduction

The term "teleconnection", which denotes climate anomalies related over distances typically of order 10^3 km, was introduced in a climate context by Angstroem (1935) and has been used to describe remote influences on the variability of large-scale features of atmospheric circulation as well as precipitation and temperature. Teleconnections are often represented as geographically dependent wave-like structures. Although the concept of teleconnections only gained widespread usage with the work of Bjerknes in the late 1960s on global-scale impacts of El Niño-Southern Oscillation (ENSO) (see Bjerknes 1969), aspects of remote climate variability now recognized as teleconnections were appreciated much earlier (e.g., Walker 1923, 1924; Walker and Bliss 1932). While teleconnections can be defined on

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synoptic time scales, they are typically associated with lower-frequency variability and are often operationally divided into two frequency regimes: intraseasonal and interannual.

Changes in the tropical atmosphere associated with oceanic changes may trigger responses in the global atmosphere via tropical-extratropical and tropical-tropical teleconnections. In this view, the atmosphere acts as a bridge between the oceanic forcing regions and remote oceanic regions (Lau and Nath 1996; Alexander et al. 2002). Teleconnection studies have provided a good example of link between observations, theory and models, as discussed in Hoskins and Pearce (1983) and Trenberth et al. (1998).

In the early 1980s several studies documented teleconnection patterns (e.g., Wallace and Gutzler 1981), associated some of them with tropical sea surface temperature (SST) variability (e.g., Horel and Wallace 1981), and discussed this association in terms of Rossby wave dispersion theory (e.g., Hoskins and Karoly 1981; Webster 1981; Simmons 1982). These seminal works spurred much interest in the relationship between tropical and extratropical low-frequency variability. Several mechanisms were proposed to explain the origin of the extratropical low-frequency variability. Most of the discussion centered on two hypotheses; namely, that the circulation anomalies are either the product of the Rossby wave dispersion from anomalous tropical heat sources or the result of internal atmospheric fluctuations caused by instabilities of the basic state. Observational and modeling studies have pointed to the contribution of both mechanisms acting in concert, as instability mechanisms can modify and amplify Rossby wave propagation. Instability mechanisms may be taken into account in the Rossby wave propagation by specifying a realistic basic state in the vorticity equation model (e.g., Simmons 1982; Branstator 1983). However, Frederiksen and Webster (1988) discussed the need for a unified theory of Rossby wave propagation and 3-D baroclinic instability to explain the observed teleconnections. According to Schubert (1986), the baroclinic conversion of energy can contribute, but barotropic processes are responsible for most of the total energy that the low-frequency anomalies receive from the basic state.

While the traditional view of teleconnections has emphasized extratropical forcing by the tropics, there is an emergent sense that forcing of the tropics by midlatitude variability is also important. Compared to what is known of tropical impacts on the extratropics, the forcing effects of the midlatitudes on the tropics are less well understood. These processes may involve the possible triggering of equatorially trapped waves by stochastic forcing from higher latitudes (Mak 1969) or the forcing of tropical convection by Rossby waves propagating from midlatitudes (e.g., Liebmann and Hartmann 1984) that can even cross the equator through the westerly duct in the equatorial east Pacific or the Atlantic (e.g., Webster and Holton 1982). The results of different studies on tropical-extratropical interaction suggest a "two-way" relationship in which the tropics influence the extratropics and vice-versa (Matthews and Kiladis 1999; Rasmusson and Mo 1993). A more complete understanding of the processes involved in these interactions and the generation of the atmospheric teleconnection patterns may have a profound impact on the improvement of medium and long-range forecasts (Ferranti et al. 1990).

Interest in the effect of teleconnections on South America (SA) climate increased after the strong 1982–1983 El Niño episode, when extreme floods and droughts affected the southern and northern parts of the continent, respectively. This chapter presents the mechanisms and the most important effects of tropics-extratropics and tropics-tropics teleconnections with SA on intra-seasonal to interannual time scales with more emphasis on the dynamical component of the teleconnections. The focus of Section 7.2 is the theoretical understanding and modeling of the tropics-tropics and tropics-extratropics interactions. Sections 7.3 and 7.4 describe observed teleconnections on interannual and intraseasonal time scales, indicating linkages between the tropics and extratropics and different regions in the tropics. Special emphasis is placed on the teleconnections associated with ENSO on interannual time scales and the Madden Julian Oscillation on intraseasonal time scales.

7.2 Mechanisms of Tropics-Tropics and Tropics-High Latitudes Teleconnections

In the tropics, anomalous positive SSTs increase the heat and moisture flux to the atmosphere, enhance low-level convergence, and therefore increase convection. Consequently, the upper-level tropical divergent circulation is enhanced in its zonal (Walker circulation) and meridional (Hadley circulation) components, resulting in anomalous ascent or descent in the branches of these circulations. Stronger upper-level divergence in the tropics and convergence in the subtropics create an anomalous Rossby wave source. The direct effects of the perturbations in the Walker and Hadley cells, as well as the dispersion and propagation of Rossby waves to the extratropics, induce anomalous circulation in the atmosphere, resulting in changes in precipitation, temperature, and even extratropical SSTs. In this section, the climatological Hadley and Walker circulations are described and the theoretical basis of the Rossby wave propagation is presented.

7.2.1 Walker and Hadley Circulation Cells

The Hadley cell was postulated as early as 1686 by Halley to explain the trade winds; later, Hadley (1735) suggested that thermal convection might drive such a circulation. However, it was only at the end of the 20th century that satisfactory quantitative models of this process were developed (e.g., Held and Hou 1980; see also the thorough review by Diaz and Bradley 2004).

The tropical Hadley cell is generally defined as the zonal mean meridional mass circulation in the atmosphere bounded roughly by 30°S and 30°N. It is characterized by equatorward mass transport by the prevailing trade wind flow in the lower troposphere, and poleward mass transport in the upper troposphere, with convergence and ascending motion in the equatorial zone and subsidence in the subtropics. This cell plays an important role in the large-scale circulation, linking the tropics with the

subtropics. The fundamental forces that determine the Hadley circulation are: pressure gradient forces arising from differential radiative heating between the equator and poles and the Coriolis force associated with the Earth's rotation. The primary energy source of the Hadley cell is the latent heat of water vapor released in the Inter-Tropical Convergence Zone (ITCZ). The adiabatic heating due to downward motion in the subtropics maintains the meridional temperature gradient and large-scale eddy motions in midlatitudes (Sasamori 1981). Also, as suggested by Bergman and Hendon (2000), the impacts of atmospheric cloud radiative forcing on the local stability, which in turn influences convection and latent heating of the atmosphere, are non-negligible and should be taken into account. A more recent discussion of the elementary Hadley circulation can be found in Webster (2004).

The Walker circulation comprises longitudinal direct circulation cells in the equatorial atmosphere. The principal energy source of the Walker circulation is also attributed to the release of latent heat in major convection centers associated with the ocean-continent distribution and the sea-surface temperature distribution maintained by the coupled ocean-atmosphere interactions (Sasamori 1981). In the ascending branch of these longitudinal cells convection and the associated release of latent heat lift isobaric surfaces upward in the upper troposphere and create a high pressure region there. The descending branch is located in a tropical region where less or no convection results in lower pressure in the upper troposphere. A longitudinal pressure gradient is established which, by virtue of its equatorial latitude, cannot be balanced by the Coriolis force. Thus a direct zonal circulation is driven in the equatorial plane with countervailing winds at the surface and in the upper troposphere. The adiabatic heating by downward motion in the Walker circulation is probably mostly balanced by radiative cooling (Bjerknes 1969). The strongest Walker cell is the Pacific cell associated with the pressure difference between eastern and western tropical Pacific, with an ascending branch in the climatological strong convection over the western Pacific warm pool and a descending branch over the cold SST in eastern Pacific. Two other smaller cells are also associated with SA, with ascending branch over the highly convective Amazon region, and descending branches over eastern Pacific and Atlantic oceans.

Diagrams of the climatological Hadley and Walker circulations for the Southern Hemisphere summer and autumn seasons (December-January-February, DJF, and March-April-May, MAM, respectively) are depicted in Fig. 7.1. These seasons were chosen because most of the annual total rainfall observed over SA occurs during these periods.

During both DJF and MAM, the ascending branch of the Pacific Walker cell is over the western Pacific Ocean, around 160°E–180°, and the sinking branch is located in the eastern Pacific, close to the west coast of SA (between 120°W to 90°W). Over the equatorial portion of SA (including Ecuador, northern Peru, south-central Colombia, Venezuela and much of the Brazilian Amazon) there is strong ascending motion, with compensatory subsidence reaching the east-central tropical Atlantic Ocean. While the Walker circulation changes little near SA from summer to autumn, the regional Hadley circulation shows important differences between these seasons. During DJF ascending motion is observed from 5°S to 30°S with a
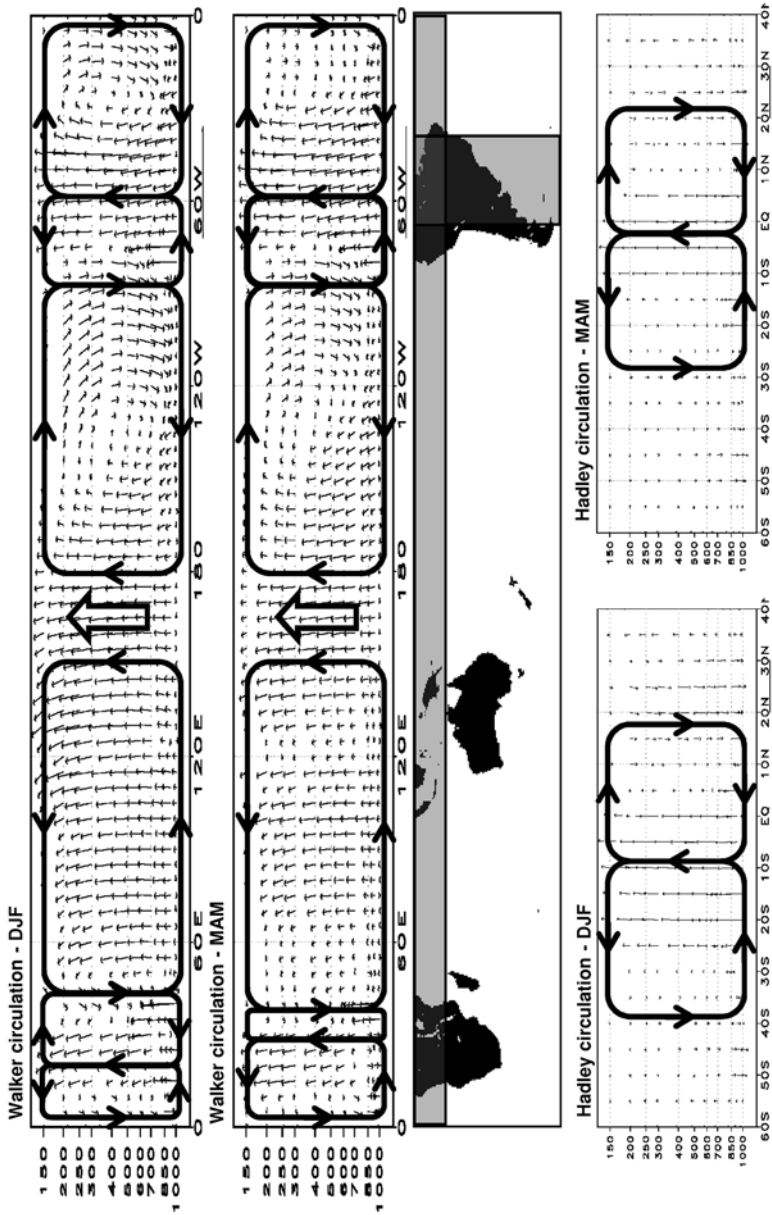
Fig. 7.1 Schematic diagrams of the climatological Walker and Hadley circulations for DJF and MAM, based on wind averages over the latitudinal and longitudinal bands indicated on the map

maximum around 7°S (and a secondary maximum around 20°S). The compensatory descending branches occur over both hemispheres around 20°N and 40°S. This circulation pattern results from the intense convection that occurs over the continent during the austral summer, which reflects the strong monsoon regime that extends into the subtropics of SA. During MAM the ascending motion is nearly centered on the equator. While the descending branch in NH is approximately in the same position as in austral summer, in the SH it is shifted to the north.

It is clear that the seasonal variability of the ascending and descending branches of the regional Walker and Hadley circulations over SA can modulate the rainfall distributions in some regions; further, the intraseasonal and interannual variability of these circulations can significantly affect SA precipitation. The interannual changes associated with ENSO have been intensively studied by many researchers (e.g. Bjerknes 1966, 1969; Arkin 1982; Webster et al. 1998; Philander 1990; Oort and Yienger 1996; Wang 2002; Minobe 2004).

During El Niño (EN) episodes the tropical convection is shifted from western Pacific towards central and east Pacific. Consequently, the Pacific Walker cell is weakened, because the induced anomalous circulation along the equator opposes the climatological circulation. As the anomalous subsidence in the Walker cell associated with anomalous convection over the eastern Pacific occurs over northern SA and Atlantic Ocean, the two smaller cells connected with SA are strongly affected, especially by the weakening of the ascending branch over the Amazon region. The reduction of the convection over this region also reduces the regional Hadley circulation. On the other hand, the Hadley circulation is strengthened over the central/eastern Pacific. During La Niña (LN) episodes the changes are nearly opposite. The perturbations over SA will be analyzed in Section 7.3.2.

The Walker cell variability and associated remote impacts during EN events described above only take in account the dynamical component of the atmospheric circulation. However, as suggested by Chiang and Sobel (2002), impacts from and to the thermodynamics are also important to understanding the teleconnected response to El Niño. Using a single-column model coupled to a slab ocean mixed layer, Chiang and Sobel (2002) investigated the interannual tropical tropospheric temperature (TT) variations during ENSO events. They show evidence that the TT mechanism is applicable to observed SST and precipitation variability in the Tropics outside the ENSO region. In fact, as will be discussed in Section 7.3.2.1, Chiang and Sobel showed that for a given observed TT forcing, the model January-February-March (JFM) averaged precipitation response is correlated to JFM precipitation anomalies in the tropical Atlantic, consistent with previous studies that argue for a dynamical linkage between ENSO and precipitation there via the anomalous Walker circulation (e.g., Ambrizzi et al. 2004 and references therein).

Although the Hadley circulation is referred to in some studies as a global zonally symmetric meridional circulation (e.g. Oort and Yienger 1996; Trenberth et al. 2000; and references therein), and the Walker circulation is considered the zonally asymmetric zonal component of the tropical divergent circulation (Webster 1983), it is possible to analyze their impact and variability on a regional level, though it should be kept in mind that the atmospheric circulations may not be closed cells.

7.2.2 Rossby Wave Propagation

Originally, Rossby (1945) described on a β-plane how a vorticity source in a barotropic atmosphere with a homogeneous westerly basic flow generates a series of ridges and troughs along which the energy is dispersed with the group velocity. The Rossby wave dispersion theory, which still provides the basis for theories of how the tropics influences the extratropics, was later expanded by several authors to more realistic situations. For example, Hoskins et al. (1977) investigated the energy dispersion over a sphere through zonally symmetric basic state, and Simmons (1982) used a realistic zonally asymmetric basic state.

The basic barotropic model to study Rossby wave propagation, with forcing F and damping A, is:

$$\frac{\partial \zeta}{\partial t} + \vec{V}_\psi \cdot \nabla \zeta = F + A \tag{7.1}$$

where ζ is the absolute vorticity and $\mathbf{V}_\psi$ is the rotational component of the wind. The first two terms govern the dispersion of pure Rossby waves. The forcing term is related with divergence. Many of the studies to be discussed are based on the linear wave theory and usually a stationary Rossby wave is assumed.

In the case of tropical-extratropical interactions, the forcing results from anomalous convection over anomalous tropical SST. Convection anomalies result in anomalous atmospheric heating, mainly due to latent heat release in anomalous precipitation, and this leads to upper-level divergence through the stretching of the air column. Equation (7.1) is solved at a pressure level which is both near to the maximum divergence associated with tropical convection and near an equivalent barotropic level in the extratropics (300–200 hPa).

In the earlier studies F = −fD, where f is the Coriolis parameter and D is divergence, and (7.1) is linearized about a zonally symmetric basic state (e.g., Hoskins et al. 1977; Hoskins and Karoly 1981). The resulting wave patterns agreed with certain aspects of the observed responses to tropical ENSO-related convection and those simulated with more complex atmospheric general circulation (AGCM) models: the basic structure and direction of propagation from the tropical forcing, and the stronger teleconnections in winter and weaker in summer (as stronger propagation occurs in stronger westerly flow, but not through easterlies). However, there were also several discrepancies. The principal were: (i) while the tropical forcing is observed in the mean easterlies, the Rossby wave forcing needs to be in mean westerly flow to produce propagation; (ii) this model does not reproduce differences in propagation arising from different longitudes of the forcing, contrary to observed and AGCM-modeled response; (iii) this model response has much weaker amplitude away from the source than the observed response.

Some of these drawbacks can be overcome by relaxing some of the assumptions in this simple model. The use of a realistic basic state, i.e., one that is a function of longitude as well as latitude, can account for the effect of non-uniformities in the basic state on the wave propagation and local instabilities, which allow internal

energy sources to enhance the propagating tropical-stimulated perturbations in the extratropics or to excite internal natural variability modes (e.g., Simmons 1982; Branstator 1983, 1985). Model (7.1) with a climatological basic state was also used by Hoskins and Ambrizzi (1993) and Ambrizzi et al. (1995) to analyze characteristics of the Rossby wave dispersion in terms of preferred propagation paths. They derived the ray path radius of curvature given by the expression

$$r = \frac{K_S^2}{\left(k\frac{dK_S}{dy}\right)} \tag{7.2}$$

where k is the zonal wavenumber, $K_S = \left(\beta^*/\bar{U}\right)^{1/2}$ is the stationary wavenumber, β^* the meridional gradient of absolute vorticity and $\bar{U}$ the mean zonal wind. From a global distribution of K_S, it is possible to indicate preferred propagation paths of Rossby waves.

   A step beyond in dealing with the discrepancies between the simple Rossby-wave model simulations and observed tropical-extratropical propagation and in establishing a more realistic relationship between the anomalous tropical heating and the Rossby wave source is using a more complete formulation of the model. From an analysis of the vorticity equation in the upper troposphere, it is possible to neglect the term involving the vertical advection and, after partitioning the horizontal wind into its rotational and divergent components,

$$\vec{V} = \vec{V}_\psi + \vec{V}_\chi$$

and expressing each variable as a sum of a basic state component and a perturbation, such as

$$\zeta = \bar{\zeta} + \zeta'$$

obtain a linearized vorticity equation model by assuming that the perturbations are small compared to the basic state quantities (thereby neglecting products of perturbations):

$$\frac{d\zeta'}{dt} + \bar{\vec{V}}_\psi \cdot \nabla\zeta' + \vec{V}'_\psi \cdot \nabla\bar{\zeta} = F' + A' \tag{7.3a}$$

where

$$F' = -\bar{\zeta}D' - \vec{V}'_\chi \cdot \nabla\bar{\zeta} - \zeta'\bar{D} - \bar{\vec{V}}_\chi \cdot \nabla\zeta' \tag{7.3b}$$

   The left hand side of (7.3a) retains the terms necessary to support Rossby wave propagation, but now the Rossby wave source F includes an additional term besides ζD, namely the advection of vorticity by the divergent flow. Sardeshmukh and Hoskins (1988) advocated the importance of this term and Grimm and Silva Dias

(1995a) showed examples of its influence to the Rossby wave propagation. While the Rossby wave source given by $F' = -fD'$ straddles the equator for anomalous equatorial divergence (or convergence), the Rossby wave source given by (7.3b) extends into the subtropical westerly mean flow (from where the waves can propagate efficiently) due to the vorticity advection by the anomalous divergent flow in regions of strong mean vorticity gradients, such as subtropical jets, or to the advection of vorticity perturbations to the subtropics by the climatological divergent circulation in the tropics (see Grimm and Silva Dias 1995a for the relative contributions of each term in (7.3b) for an EN case). Therefore, in this definition of the source, the upper level anomalous divergence does not need to be placed in the subtropics or in particular regions of the tropics in order to produce a nearly realistic extratropical response. Figure 7.2 shows the difference between the responses to an anomalous divergence over the central equatorial Pacific and convergence over western Pacific using $F' = -fD'$ or the expression (7.3b).

Therefore, the local Hadley circulation, through the associated tropical divergent outflow and subtropical convergence, is important in extending part of the Rossby waves source away from the anomalous tropical heating into the subtropics and the westerly mean flow, where the waves can propagate to higher latitudes.

However, even with the more realistic formulation of the model, there is a major problem when studying tropical-extratropical interactions with the aim of determining the regions where tropical heating anomalies contribute to produce certain low-frequency circulation anomalies: the model demands the correct specification of the upper-level divergence associated with the tropical heat sources. Although
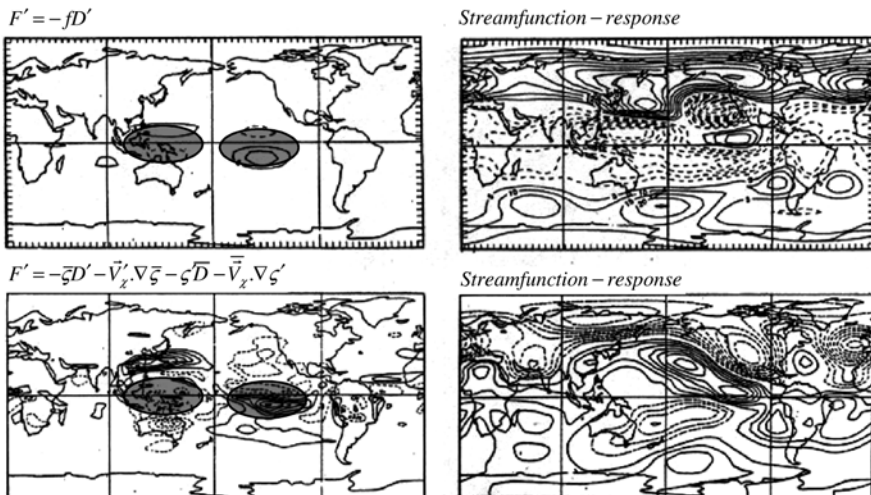
Fig. 7.2 Streamfunction responses (*Right panels*) to the same anomalous divergence in eastern Pacific and anomalous convergence in western Pacific (*indicated by the shaded ellipses in the left panels*), but with different formulations of the Rossby wave source (from Grimm and Silva Dias 1995a, reproduced with permission of the Meteorological Society of Japan)

there is a direct correspondence between upper-level divergence and the tropical heat source, there is not a simple relation with the anomalous upper level convergence, which is as important as the divergence for the rotational response (perturbed convergence in the subtropics is, for example, mainly responsible for the Pacific-North American pattern, PNA, Grimm and Silva Dias 1995a, b). Thus, the correct specification of the anomalous divergence field to achieve certain streamfunction anomalies is very uncertain and small errors or omissions can produce large differences in the response. Apparently insignificant features may have a large impact on the response in remote regions (Grimm and Silva Dias 1995a). Therefore, Grimm and Silva Dias (1995b) proposed the analysis of tropical-extratropical interactions with Influence Functions (IFs) of divergence forcing, which identify the regions where the anomalous upper-level divergence (which is more easily associated with anomalous convection and tropical heating than the vorticity source) has the largest impact on the circulation anomaly around a given point. Therefore, experiments with the barotropic models, which are hardly conclusive when using idealized anomalous divergence fields, can be guided by the indications of the IFs. Model (7.3) was written as

$$\frac{d\zeta'}{dt} + \bar{\vec{V}}_\psi \cdot \nabla \zeta' + \vec{V}'_\psi \cdot \nabla \bar{\zeta} + \bar{\vec{V}}_\chi \cdot \nabla \zeta' + \zeta' \bar{D} - A' = F' \qquad (7.4a)$$

with

$$F' = -\bar{\zeta} D' - \vec{V}'_\chi \cdot \nabla \bar{\zeta} \qquad (7.4b)$$

where F' depends only on the anomalous divergence. The stationary version of model (7.4) may be written as:

$$M\psi' = D',$$

with M a linear operator and ψ' the anomalous streamfunction. Then the IF based on divergence forcing is defined by

$$G_D(\lambda,\phi,\lambda',\phi') = M^{-1} \left[\delta \left(\lambda,\phi,\lambda',\phi' \right) \right], \qquad (7.5)$$

where $\delta \left(\lambda,\phi,\lambda',\phi' \right)$ is the delta function. Thus the IF $G_D(\lambda,\phi,\lambda',\phi')$ for the target point with longitude and latitude (λ, ϕ) is, at each point (λ', ϕ'), equal to the model response at (λ, ϕ) to an upper-level divergence located at (λ', ϕ'). Maps with contours of IF for a given target point indicate the tropical/subtropical regions in which the anomalous upper-level divergence is most efficient in producing streamfunction anomalies at the target point. Using this tool, Grimm and Silva Dias (1995b) indicated the source regions for several teleconnection patterns. Some of them will be analyzed in following sections.

Manifestations of Rossby wave propagation (and thus indirectly of Hadley and Walker circulation variations) can be seen in low-frequency teleconnection patterns. Large scale tropical convection and global circulation undergo variations on several

time scales and characteristic spatial patterns. The low-frequency variability will be analyzed separately on interannual and intraseasonal (10–120 days) time scales, although some of the variability modes oscillate on both time scales (e.g. Lau et al. 1994; Frederiksen and Zheng 2007). The focus is on the Southern Hemisphere variability modes, although they may be part of global modes. Interannual fluctuations are more geographically dependent, while intraseasonal time scales are less anchored and more zonally oriented wave trains generally localized in the strong westerlies of the subtropical and polar jet.

7.3 Teleconnections with South America on Interannual Time Scales

7.3.1 Southern Hemisphere Observed Interannual Variability

Several observational studies on Southern Hemisphere teleconnection patterns have been carried out with different methods, variables, domains, data sets and periods analyzed (e.g., Mo and White 1985; Mo and Ghil 1987; Kidson 1988, 1999; Ghil and Mo 1991; Kiladis and Mo 1998; Mo 2000; Revell et al. 2001; Frederiksen and Zheng 2007). Although there is some diversity in patterns and explained variances, the interannual low-frequency variability in the SH is dominated basically by three or four features. These are a Southern Hemisphere high latitude mode (HLM, also called Antarctic Oscillation (AAO), or Southern Annular Mode (SAM)), a Rossby wavetrain-like mode, referred to as the Pacific South America mode (PSA), and a Southern Hemisphere meridional wavetrain (MW) pattern (sometimes named South Pacific Wave, SPW), linking western Pacific to the western Atlantic Ocean.

Although there are inter-seasonal differences, several studies have verified that there is some similarity between these patterns in different seasons. For example, Fig. 7.3 shows the variability patterns obtained by Kidson (1999) using data from all months and low-pass filtered with a cut-off of 50-days.

The SAM, the dominant mode on interannual time scales (IA1 in Fig. 7.3), has a strong zonally symmetric component, with out of phase relationship between heights in a zonal band between 55°S and 45°S and heights south of 60°S and in the subtropics, leading to a maximum negative correlation between the anomalous zonal westerlies at about 40°S and 60°S. In its positive phase, it is associated with an equatorward shift of the polar jet. This mode shows a strong connection to ENSO events (Kiladis and Mo 1998; Kidson 1999; Carvalho et al. 2005, and references therein). Enhanced Hadley circulation, driven by strong convection over the central Pacific, strengthens the subtropical jet stream by the large scale meridional flux of westerly momentum. The large anomalies in the extratropical tropospheric zonal wind are brought about by the anomalous rotational circulation forced by perturbed divergence in the subtropics, as the Rossby wave source gets stronger there. The polar jet is weakened and the tendency is for a single
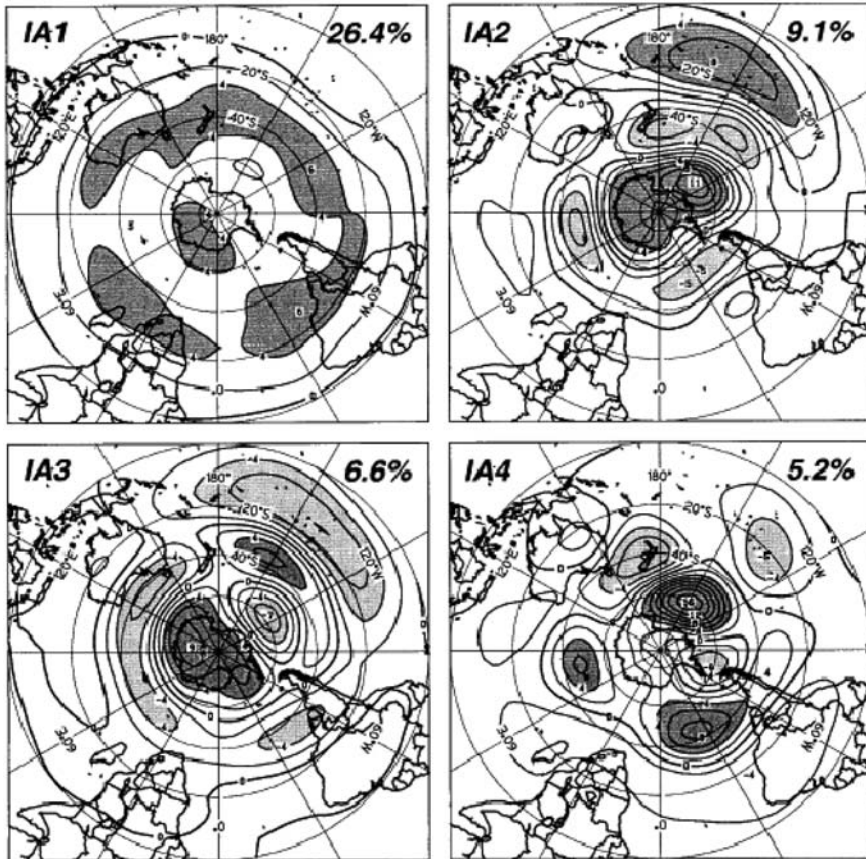
Fig. 7.3 The four leading Empirical Orthogonal Functions of 50-day low-pass filtered 300-hPa streamfunction (from Kidson 1999, reproduced with permission of the American Meteorological Society)

jet, resulting in an equatorward shift of the storm track, which affects southern SA. During cold events approximately opposite anomalies occur and the double jet is more pronounced, as the polar jet gets stronger in subantarctic latitudes (e.g., Kiladis and Mo 1998). Revell et al. (2001) found the strongest upper-level divergence/convergence associated with this mode in the central tropical and subtropical Pacific. Besides the impact of the perturbed Hadley cell and Rossby wave propagation due to anomalous tropical heat sources, it is also necessary to take into account the feedback by the transients in strengthening the westerlies (e.g. Trenberth 1987).

The second and third leading patterns are also connected to ENSO, corresponding to the wave train response that propagates southeastward from the central/western tropical Pacific into the extratropics and then equatorward into SA. (e.g., Kidson 1999; Frederiksen and Zheng 2007). It is referred to as PSA patterns (Mo and Ghil 1987; Mo 2000, and references therein) in analogy to the

Northern Hemisphere PNA pattern (Wallace and Gutzler 1981). These modes seem to originate from the tropics or subtropics, are stronger over the Pacific sector, have wavenumber 3 structure at midlatitudes and are in quadrature, although there is no evidence that this represents a travelling oscillation (Kidson 1988). While one of them seems to originate in tropical central Pacific, the other seems to extend from subtropical western Pacific, which is coherent with the results by Revell et al. (2001). These modes are instrumental in the ENSO impact on southeastern SA. The interannual PSA modes in summer exhibit a much more zonally symmetric structure and weaker magnitude than in other seasons, with less indication of wave train turning northeastward from southern Pacific into subtropical SA and Atlantic Ocean (Kidson 1999; Frederiksen and Zheng 2007). This indicates that the extratropical ENSO teleconnection with southeastern SA in austral summer is weaker, as already suggested in Grimm (2003, 2004).

The fourth leading mode (MW) is a nearly meridional South Pacific wavetrain, with wavenumber 3 (winter) or 4 (summer), extending from Australia into the Atlantic Ocean. It is not related with ENSO, but seems to be connected with anomalous SST east of Australia. It also seems to be a product of Rossby wave dispersion (Karoly et al. 1989). Influence functions of the action centers of this pattern confirm the region between Australia/New Guinea and the Date Line as the region in which upper-level divergence is most efficient in producing the streamfuntion anomalies associated with this mode (not shown).

As seen, a large portion of the interannual variability of the atmosphere in Southern Hemisphere is linked to variations in the tropical oceans, and the most important influence is ENSO. Therefore, the impacts of ENSO will be described in next section.

7.3.2 El Niño/Southern Oscillation Teleconnections with South America

The most important mode of interannual variability of the coupled ocean-atmosphere system is ENSO. Literature on ENSO is extensive, and the reader can find comprehensive reviews in Philander (1990), Diaz and Markgraf (1992), Glantz et al. (1991), Allan et al. (1996) and Trenberth and Caron (2000). Here we will focus on its influence on circulation and precipitation.

The first interannual variability mode of precipitation in SA, which is connected with ENSO, shows the general tendency to opposite ENSO-related anomalies in northern/northeastern and southeastern SA (Fig. 7.4). This leading mode explains 19.0 % of the variance. To calculate the variability modes, precipitation data from about 10,000 stations over most of SA are averaged on a 2.5° × 2.5° latitude-longitude grid. The analyzed period is 1961–2000. These data are obtained mainly from the Brazilian Agência Nacional de Águas (ANA) but also from some other Brazilian agencies and meteorological services of other countries in SA.

Fig. 7.4 First interannual
variability mode of annual
rainfall over South America
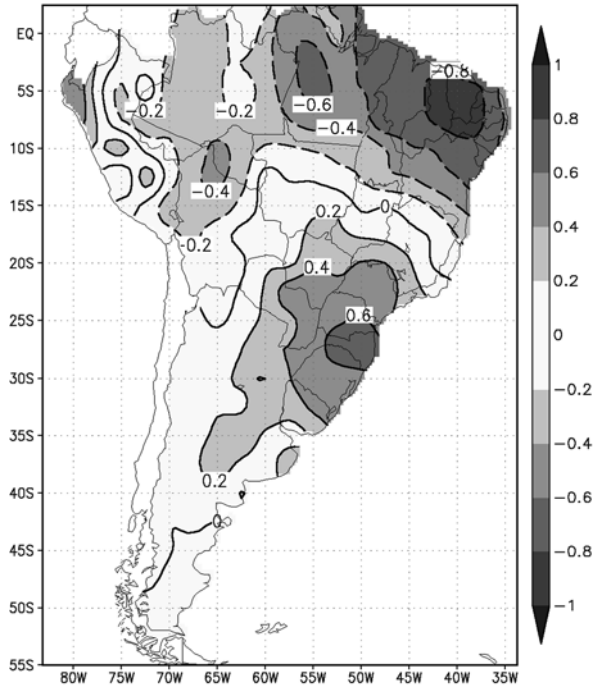
The anomalous tropical heat sources associated with ENSO events perturb the
Walker and Hadley circulation over SA, and generate Rossby wave trains that pro-
duce important effects in the subtropics and extratropics of the continent. This
response is represented in the leading Empirical Orthogonal Functions (EOFs) of
circulation anomalies in Southern Hemisphere (such as the SAM and PSA). It is
worth emphasizing that, in spite of the general similarity in the EOF patterns for
different seasons, there are significant seasonal differences in these patterns. There-
fore, there is some disagreement between the patterns calculated with all months
(as in Fig. 7.3) and patterns for one season or month. The ENSO teleconnections
described here will focus on the austral spring season, when they are strongest and
the largest relative impacts occur.

7.3.2.1 Northern South America and the Tropics-Tropics Teleconnection

Areas located in northern/northeastern SA (including Venezuela, French Guiana,
Surinam, Guyana and Brazilian Amazon and Northeast regions) have one of the
most consistent ENSO-precipitation relationships (Ropelewski and Halpert 1987,
1989). Several authors (e.g., Kousky et al. 1984; Aceituno 1988; Rao and Hada
1990; Uvo et al. 1998; Souza and Ambrizzi 2002; Grimm 2003, 2004; Poveda
et al. 2006; Grimm and Tedeschi 2009) reported observational evidence that EN
(LN) episodes are related to deficient (abundant) rainy season in these continental

areas. Ambrizzi et al. (2004) and Magaña and Ambrizzi (2005) have associated ENSO impacts on the seasonal rainfall with the tropical Hadley and Walker regional circulation.

The tropics-tropics teleconnection is instrumental in the impact of ENSO over northern SA. In warm ENSO episodes, warm SST anomalies occur over the tropical central and eastern Pacific (Fig. 7.5). These warm anomalies induce more evaporation, lower sea level pressure, more low-level convergence and more convection in a tropical region with climatological subsidence. The enhanced convection translates into more precipitation, which is represented as negative OLR anomalies in Fig. 7.5. Consequently, there is upper level anomalous divergence over the eastern Pacific and convergence over northern SA (Fig. 7.5), which is associated with the descending branch of a stationary Kelvin wave that inhibits convective activity. The upper-level streamfunction anomalies show the expected pair of antyciclones straddling the tropical anomalous convection (Gill 1980), with a weaker pair of cyclonic anomalies to the east (Fig. 7.5). This anomalous convection over the central/eastern Pacific and its associated tropical circulation perturbs the Walker circulation from its climatological situation (compare Fig. 7.5 with Fig. 7.1) and the Pacific descending branch is shifted from eastern Pacific to northern SA and the increased subsidence over this region is associated with the observed rainfall deficit. Therefore, the Hadley cell is also perturbed (Fig. 7.5). However this canonical El Niño result for a typical episode may be altered by other factors that can modify this pattern, such as anomalous SST in the tropical Atlantic or other inter-episodes differences (Ambrizzi et al. 2004; Magaña and Ambrizzi 2005; Nobre and Shukla 1996).

Many studies have already indicated that the interannual variability of SSTs over the tropical Atlantic has a strong influence on the distribution of rainfall over the tropical Americas (Moura and Shukla 1981; Hasternrath 1984; Nobre and Shukla 1991). In particular, Nobre and Shukla (1996) presented observational evidence that ENSO phenomenon in the Pacific influences atmospheric circulation and SST anomalies over northern tropical Atlantic through atmospheric teleconnection patterns into higher latitudes of the Northern Hemisphere. They suggested that the Pacific-North American pattern (PNA – Horel and Wallace 1981) induces Sea Level Pressure (SLP) of the opposite sign between the Pacific and Atlantic tropical basins. Once SLP anomalies of opposite sign are set over the northern and southern Atlantic, then hemispherically asymmetric trade winds and SST anomalies follow. The observational and modeling studies done by Alexander et al. (2002) have also shown the connection between ocean basins via the "atmosphere bridge".

Besides the dynamics view of the link between the ENSO events in the Pacific and its influence over the Atlantic SSTAs one can also use thermodynamic arguments to explain this association. For example, one suggestion of Chiang and Sobel (2002), and expanded upon in Chiang and Lintner (2005), is that the thermodynamic view of the tropical teleconnection underscores the role of surface-troposphere disequilibrium: in the presence of a warm troposphere (which occurs within weeks after development of El Niño phase Pacific SSTAs because of anomalous temperature propagation via efficient Kelvin wave dynamics), the ocean surface temperature, with its large thermal inertia, initially lags. This produces certain signatures in the
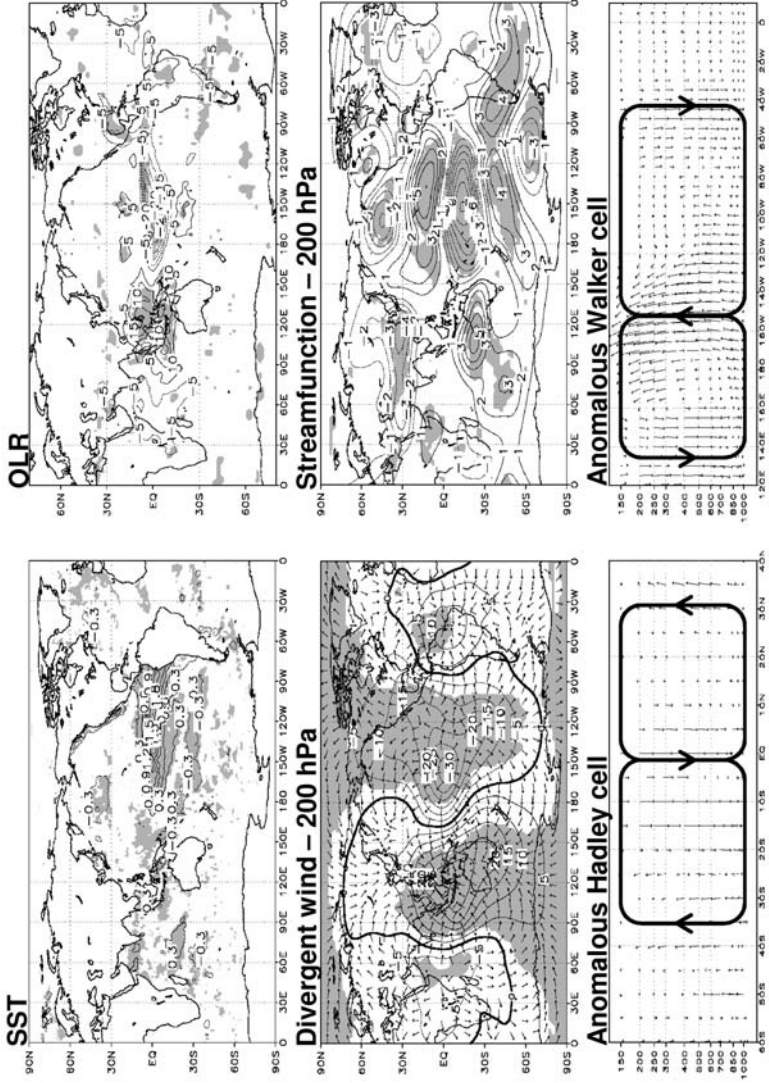
Fig. 7.5 Average anomalous fields associated with El Niño episodes in the austral spring (OND). Regions *shaded in gray* present consistent anomalies with significance level better than 0.10 (After Grimm 2009)

ocean surface flux budget, e.g., latent heating anomalies are initially negative (into the surface) as the surface warms, but the sign of the latent heating anomaly may, at some point, reverse, with positive anomalies acting to limit further surface warming. The way in which latent heating anomalies come about also encompasses both a dynamical effect – i.e., changes associated with low-level windspeed – as well as a thermodynamic effect – the difference in specific humidity at the surface relative to saturation. Chikamoto and Tanimoto (2006) noted that the latter appears to be important to the generation of the Atlantic SST response to El Niño.

The simplified Gill type circulation (Gill 1980) described in this section, i.e., anomalous forced ascent over the Pacific is balanced by compensatory descent elsewhere, is not able to fully explain many of the observed features in the tropical teleconnection where more realistic land-ocean-atmosphere interactions, moist convective dynamics, and inhomogeneous basic states have to be considered (Lintner and Chiang 2007). In fact the "atmospheric bridge" as defined by Lau and Nath (1996) or more recently the "remote tropics" discussed by Lintner and Chiang (2007) indicated the teleconnected dynamical and thermodynamic linkages between warm SST anomalies in the eastern equatorial Pacific and their remote tropical influence. The role of thermodynamics in propagating the ENSO signal to remote Tropics was suggested by the TT mechanism in the work of Chiang and Sobel (2002) and the importance of regionally variable moisture-related processes (e.g., advection and evaporation) in interfering with El Niño-related precipitation changes was studied by Neelin and Su (2005 and references therein).

In summary, as mentioned by Neelin and Su (2005), teleconnection in the tropics, and specially the important impacts on land precipitation are complicated by interactions with convective heating, shortwave and longwave cloud radiative feedbacks (Bergman and Hendon 2000) and land surface feedbacks. The present review has concentrated mostly on the dynamical aspects of the tropics-tropics interaction. However, a general view of the thermodynamic features can be obtained in some of the main studies described above.

7.3.2.2 Southeastern South America and the Tropics-Extratropics Teleconnection

Southeastern SA exhibits a strong and consistent impact of ENSO episodes on precipitation (Ropelewski and Halpert 1987, 1989; Aceituno 1988; Montecinos et al. 2000; Grimm et al. 1998, 2000a; Grimm 2003, 2004). Abundant (deficient) rainfall prevails during EN (LN) episodes, but the signal in precipitation is not uniform throughout an episode, because the atmospheric basic state in which the Rossby waves propagate undergoes seasonal variation. The season of strongest impact is spring, when the Rossby wave propagation is more intense towards subtropical SA, and a PSA-like pattern is more evident in the anomalous fields associated with the Southern Oscillation, as can be seen in Fig. 7.5. (e.g., Trenberth and Caron 2000). The focus of our analysis will be November, the month of strongest impact in spring.

The upper-level circulation anomalies during November of EN episodes show one wave train extending from central Pacific towards the extratropics with a small
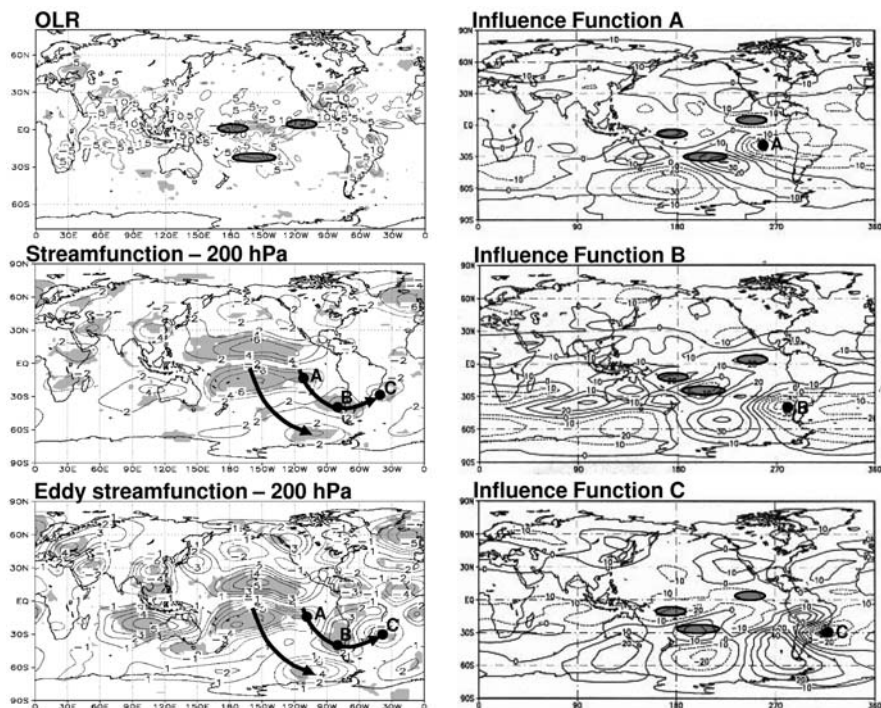
Fig. 7.6 (*Lefts panels*) Average anomalous OLR (with maximum convection over the tropical Pacific and maximum subsidence over the southern Pacific indicated by *shaded ellipses*) and streamfunction (total and without zonal mean) during November of EN episodes. Regions *shaded in gray* present consistent anomalies with significance level better than 0.10. (*Right panels*) Influence functions for the action centers of the wave train extending from eastern Pacific towards SA (*indicated by black circles*), with tropical/subtropical common regions with maximum positive (negative) values indicated by shaded ellipses (After Grimm 2009)

curvature, and another one emerging from the eastern Pacific and then, at the latitude of the subtropical jet, turning northeastward into subtropical SA, coherently with the characteristics of the propagation of a Rossby wave (Fig. 7.6). This wave train produces a cyclonic anomaly over southwest SA, and an anticyclonic anomaly off the eastern coast. This pattern of vertical equivalent barotropic structure favors the strengthening of the subtropical jet over SA, the advection of cyclonic vorticity over southeastern SA, and the transport of northerly warm and moist air into this region, making available the essential ingredients for excessive rainfall: moisture and dynamic lift (Grimm 2003). In LN episodes, the anomalies are nearly opposite, although the patterns are a little shifted; the result is deficient rainfall, with even stronger and more consistent anomalies than in EN (Grimm 2004). Besides the ENSO impact on total monthly rainfall, Grimm and Tedeschi (2009) disclosed an even stronger impact on extreme daily rainfall events, due to the greater sensitivity to ENSO in the heavy rainfall tail of the daily rainfall frequency distribution.

Figure 7.6 shows the influence functions (IFs, Grimm and Silva Dias 1995b) for the action centers of the eastern wave train. They indicate the importance of the anomalous convection (or upper level divergence) over eastern and central tropical Pacific (shown by the near equatorial ellipses) and the anomalous subsidence (or upper level convergence) over subtropical central Pacific (shown by the subtropical ellipse) in generating the streamfunction anomalies of this wave train, as the strongest values in the tropics, common to the three centers, are in these locations. The sign of the IFs is also coherent with the sign of the streamfunction anomaly around each center. For example, in the eastern and central tropical Pacific, the IFs are negative for action centers A and C, coherent with the negative streamfunction anomaly generated around these centers by positive divergence anomalies in those regions; on the other hand it is positive for action center B, coherent with the positive streamfunction anomaly generated around this center. Yet in the subtropics of central Pacific the IFs show opposite sign, indicating that negative divergence anomaly in this region (upper level convergence associated with subsidence) is able to generate streamfunction anomalies of the same sign as those generated by positive divergence anomalies in tropical eastern and central tropical Pacific. Frequently the IFs in the subtropics are stronger than in the tropics, indicating the importance of the compensatory subsidence in the local Hadley cell.

The regions with the strongest common IFs in Fig. 7.6 are also those with the strongest convection anomalies, as indicated by the OLR anomalies, which explains the significant wave train responsible for the consistent precipitation anomalies in southeastern SA. The IFs of the action centers of the western wave train (not shown) indicate the importance of anomalous upper level convergence over Australia as a possible source for this pattern.

A comment is necessary related to the influence function analysis: the large values of the influence functions at middle and higher latitudes do not imply that these regions are important for generating rotational flow by anomalous divergence at upper levels of the troposphere. It simply means that anomalous divergence in these regions is consistent with high values of the streamfunction at the target point. In these latitudes the upper level anomalous divergence is not directly related to anomalous heat sources as in the tropics, but divergence and vorticity are coupled. Thus, more attention is given to the tropics/subtropics, where the anomalous upper level divergence/convergence is directly related to anomalous tropical heat sources and nearly independent of the rotational flow.

From November to January the ENSO signal in rainfall weakens in southern Brazil and Uruguay, and even changes sign in the northern part of this region (see Section 7.3.2.3). However, in austral autumn, especially in April, remote influences are again instrumental in producing above normal rainfall in southeastern SA (Grimm et al. 2000a; Grimm and Tedeschi 2009).

Another example of information given by the IFs is the influence of tropical convection on the circulation in southern SA in winter (July). During winter of EN events there is increased precipitation in southeastern SA, especially when there is increased SST in the eastern Pacific, off the western SA coast. This caused extreme
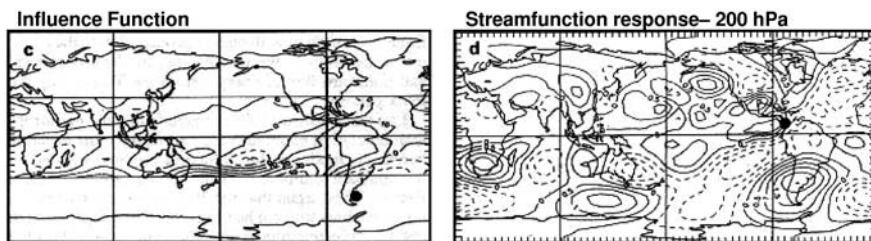
Fig. 7.7 (*Left*) Influence function of a target point in southern SA, indicated by a black circle (from Grimm and Silva Dias 1995b, reproduced with permission of the American Meteorological Society). (*Right*) Streamfunction response to an upper level anomalous divergence centered on the black circle over northern SA

floods in 1983, 1987, 1992, for instance. This enhanced precipitation is associated with a cyclonic anomaly over southern SA (Grimm et al. 2000a) which appears in EN episodes composites for winter (e.g., Karoly et al. 1989; Kiladis and Mo 1998). The IFs for the centers of the winter PSA pattern (not shown) display the right sign for producing the correct positive or negative streamfunction anomalies around each action center and indicate common influence region near the Date Line, and in the subtropics (Grimm and Silva Dias 1995b). It is noteworthy that, besides being influenced by upper level tropical divergence near the Date Line, as the other centers, the circulation around the center in southern SA is connected with the anomalous convection (or upper-level divergence) in eastern tropical Pacific and northern SA (Fig. 7.7, left). Anomalous convection in this region was strong in the EN years with strong winter rainfall anomalies in southern Brazil. Figure 7.7 (right) shows the simulated response to an upper level anomalous divergence in eastern equatorial Pacific/northern SA, featuring the cyclonic anomaly over southern SA. This inter-hemispheric connection, which is not present in austral summer, can enhance winter rainfall in southern Brazil even in non-ENSO years, when there are anomalous heat sources in western equatorial Atlantic and northern SA during the Northern Hemisphere summer.

7.3.2.3 Central-East Brazil and the Influence of Regional Processes During the Summer Monsoon

The Rossby wave tropical-extratropical teleconnection weakens in summer because the atmospheric basic state is not favorable to propagation, as the stronger westerlies are shifted to higher latitudes. There is a reversal of ENSO-related anomalies in central east Brazil (which is part of the SA monsoon core region) and in part of southern Brazil, from November to January, due to surface-atmosphere interactions that are stronger during summer. These processes favor a regional circulation pattern that overcomes the remote influences of spring (Grimm 2003, 2004; Grimm et al. 2007; Grimm and Zilli 2009).

7.4 Teleconnections with South America on Intraseasonal Time Scales

7.4.1 Southern Hemisphere Observed Intraseasonal Variability

The intraseasonal low-frequency variability (10–120 days) in the Southern Hemisphere has also been analyzed in a number of observational studies (e.g., Farrara et al. 1989; Kidson 1991; Ghil and Mo 1991; Mechoso et al. 1991; Lau et al. 1994; Kiladis and Mo 1998; Kidson 1999; Frederiksen and Zheng 2007).

The intraseasonal variability exhibits different behavior for the lower and higher frequency bands within the intraseasonal time-scales. The higher frequency intraseasonal variability (10–30 days), which concentrates much of the intraseasonal variability, is less anchored to the underlying geography. The variability patterns, usually in spatial quadrature, represent zonally oriented eastward (or equatorward) progressing wave trains in the extratropical westerly belt (near 50°S), with nearly equivalent barotropic vertical structure, usually generated by internal barotropic instability. Wavenumbers 3 and 4 are the most frequent (e.g. Kidson 1991, 1999; Berbery et al. 1992; Kiladis and Mo 1998; Frederiksen and Zheng 2007). According to Revell et al. (2001) there is no indication of the forcing of these higher frequency modes coming from the tropics, since the divergence field associated with them is essentially the divergence (convergence) ahead of each trough (ridge). This configuration generates the vorticity tendency needed to hold the upper-level component of a barotropic Rossby wave in check and prevent it being torn apart by the vertical shear in the background zonal wind.

Based on the observational intraseasonal teleconnection patterns for the Northern Hemisphere winter obtained by Hsu and Lin (1992), and applying the Rossby wave theory using simple barotropic models, Hoskins and Ambrizzi (1993) and Ambrizzi et al. (1995) draw diagrams of global teleconnection patterns of intraseasonal variability in the 10–30 day band for austral summer and winter (Fig. 7.8).

Figure 7.8, which is a conceptual diagram of teleconnection pathways obtained from global lag correlation analysis, suggests that the South American continent can be influenced by the passage of the intraseasonal waves in the subtropical and polar jets which may act as wave guides (e.g., Berbery et al. 1992; Hoskins and Ambrizzi 1993; Ambrizzi and Hoskins 1997). It is also indicated that inter-hemispheric propagation and extratropics-tropics interaction can affect the atmospheric circulation over SA, and are dependent of the season. For instance, the wave propagation pattern linking the tropics with the extratropics is clearly observed from the streamfunction 200 hPa map depicted in Fig. 7.6.

In this frequency band the extratropics-tropics interaction is frequent. Troughs and associated frontal zones propagate into the tropical SA from the Southern Hemisphere westerlies, producing cloud band activity over the South Atlantic Convergence Zone (SACZ), connecting extratropics and tropics (e.g., Kiladis and Weickmann 1992a, b, 1997; Liebmann et al. 1999; Todd et al. 2003).

The lower frequency intraseasonal variability (30–120 day) is mainly dominated by the eastward propagating Madden-Julian Oscillation (MJO) on the 30–60 day
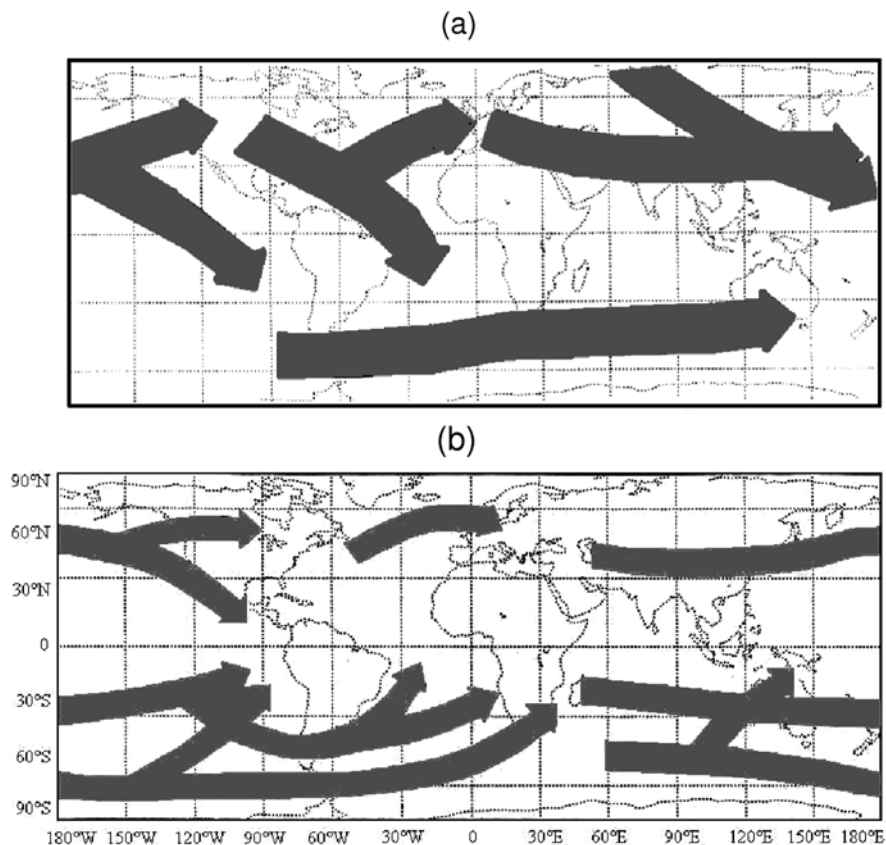
Fig. 7.8 Conceptual diagram of the global teleconnection pathways of intraseasonal influence obtained from lag correlation maps for (**a**) DJF and (**b**) JJA. (After Hoskins and Ambrizzi 1983; and Ambrizzi et al. 1995)

time scale (Madden and Julian 1971). This oscillation dominates the tropical intraseasonal variability, with spatial scale of wavenumber 0 through 2, but also significant extratropical signals are observed during its evolution. The MJO relationships to extratropical intraseasonal variability over the Southern Hemisphere will be analyzed in the following section. The PSA patterns are also present in association with the MJO.

In spite of the differences, there are similarities between the interannual variability patterns described previously and the intraseasonal patterns mentioned above. Both are dominated by the wave 3 and 4 structures and there are fluctuations in similar locations in a broad range of time scales, as in the action centers of the PSA pattern and the MW pattern. The PSA patterns, for example, are present in time-scales ranging from daily (Mo and Ghil 1987), intraseasonal (Ghil and Mo 1991; Mo and Higgins 1998), interannual (Kidson 1988, 1999) to interdecadal

timescales (Garreaud and Battisti 1999). Therefore there can be a periodic amplification/suppression of the stationary waves in those locations (Kiladis and Mo 1998; Frederiksen and Zheng 2007).

7.4.2 Madden-Julian Oscillation (MJO)

There is vast literature on the MJO. Observational and theoretical aspects can be found in Knutson and Weickmann (1987), Lau and Chan (1988), Rui and Wang (1990), Ferranti et al. (1990), Hendon and Salby (1994), Madden and Julian (1994). Here the focus is on the signals related to SA in austral summer, as this is rainy season over most of the continent.

The MJO can be described as a planetary-scale mechanism with disturbances in both tropical deep convection and tropospheric circulation, which propagate eastward in the tropics, taking approximately 30–60 days for a complete cycle around the globe (Madden and Julian 1994), but with related perturbations also in the extratropics.

Observational analysis has disclosed significant intraseasonal variability in the 30–60 day time scales over tropical (Kousky and Kayano 1994; Carvalho et al. 2002; Ferraz 2004; Souza et al. 2005; Souza and Ambrizzi 2006) and subtropical SA, with emphasis on SACZ (Casarin and Kousky 1986; Nogués-Paegle and Mo 1997; Grimm et al. 2000b; Nogués-Paegle et al. 2000; Ferraz 2004; Carvalho et al. 2004; Cunningham and Cavalcanti 2006), which is dynamically linked with MJO events (Kiladis and Weickmann 1992a; Grimm and Silva Dias 1995b). In addition, Jones et al. (2004) reported robust signals of increased frequency of precipitation extremes in the eastern part of SA during active MJO situations.

The intraseasonal variability of summer precipitation (November through March) over SA in the 30–60 day band exhibits two main features: the variability in central-east Brazil and the variability of the SACZ. The first two rotated Empirical Orthogonal Functions (EOFs) represent these features (Fig. 7.9, Ferraz 2004). The first rotated mode (10.6% of the variance) represents oscillations in SACZ, while the second one is concentrated in Central-East Brazil (10.2% of the variance). Both modes also feature anomalies in the subtropical plains of SA that are opposite in sign to those in the main center, but much weaker in magnitude. There is, however, a significant "dipole"-like relationship between precipitation anomalies in Central-East Brazil/SACZ and the subtropics to the south, although with different magnitude in both centers.

The local circulation anomalies associated with the first two modes are consistent with a seesaw pattern in precipitation. In one extreme phase of the first (second) mode, a cyclonic anomaly around 25°S, 45°W (20°S, 50°W) directs the northwesterly moisture flux into SACZ (Central-East Brazil) and decreases the southward transport to southeastern SA. In the opposite phase, an anticyclonic anomaly enhances the moisture flux towards the subtropical plains (In Fig. 7.9, right, the arrows of moisture transport are very similar to the low-level wind). In both modes the positive precipitation anomalies are connected with a cyclonic anomaly
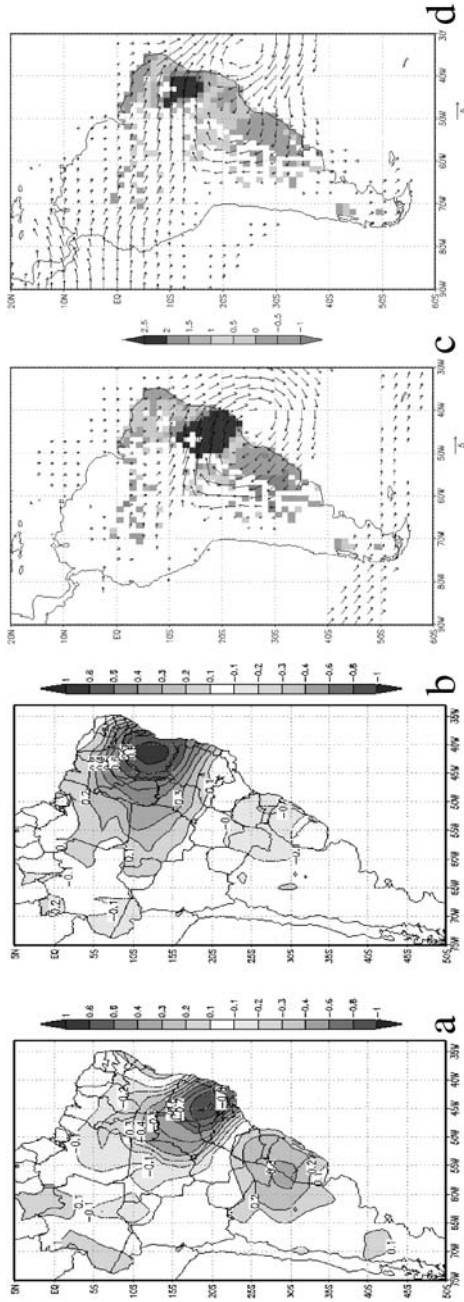
Fig. 7.9 (*Left*) First two rotated modes of intraseasonal variability in the 30–70 day band. (*Right*) Composites of rainfall anomalies and vertically integrated moisture flux for wet phases of these modes in the SACZ and central-east Brazil. Only significant anomalies are represented by arrows (From Ferraz 2004)
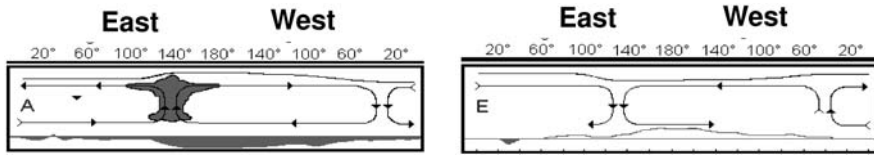
Fig. 7.10 Diagram of the anomalous Walker circulation associated with the phases of the MJO in which there is reduced (*left*) and increased (*right*) rainfall in Central-East Brazil, in the second mode of Fig. 7.9

to the southwest. For the first mode, there is still a connection with an anticyclonic anomaly southwest of this cyclonic anomaly, over southeast Pacific and southern SA, while for the second mode there is a stronger connection with winds in the tropical Pacific.

The first mode seems to be result of a tropical-extratropical interaction, while the second one is probably the product of a tropical-tropical interaction. The low-level circulation connected with these modes in Fig. 7.9 shows that there are no significant winds connected with the first mode in the tropics, but there is a significant connection with the extratropics. The second mode shows opposite behavior, with stronger connection with the tropics, and weaker with the extratropics.

The connection of the second mode with the tropics is stronger when the increased (reduced) convection associated with MJO is over New Guinea and the subsiding (ascending) branch of the anomalous Walker cell is over northern SA, as represented by Madden and Julian (1994), and reproduced in Fig. 7.10. Therefore, enhanced (reduced) convection over central-east Brazil would approximately correspond to reduced (enhanced) convection over New Guinea (corresponds to the first and third panels in the evolution of the MJO convection displayed by Ferranti et al. 1990).

From Fig. 7.10, it is clear that tropics-tropics interaction over South America seems to be important in modulating the precipitation. A further example of the MJO signature is the composite of rainfall anomalies over tropical Brazil keyed to the MJO cycle (Fig. 7.11) presented in Souza and Ambrizzi (2006). A large contribution to the intraseasonal variability over this region can also be due to the interactions with the submonthly variability related to SACZ and ITCZ activity (Souza et al. 2005).

The connection of the first mode in Fig. 7.9 (SACZ mode) to the extratropics is stronger when the convection signal of MJO reaches the SPCZ, enhancing it and shifting it eastward, at the same time that subsidence prevails over Indonesia. IF analysis indicates that the maximum influence in generating the cyclonic anomaly associated with the enhanced SACZ comes from the shifted SPCZ region with subtropical convection over the Pacific Ocean (Fig. 7.12). Anomalous divergence associated with enhanced SPCZ in this phase of MJO is able to generate an extratropical wave train that can stimulate the SACZ by generating the cyclonic anomaly southwest of it (Fig. 7.12, Grimm and Silva Dias 1995b), while opposite effects are associated with the opposite phase of the MJO (these phases correspond
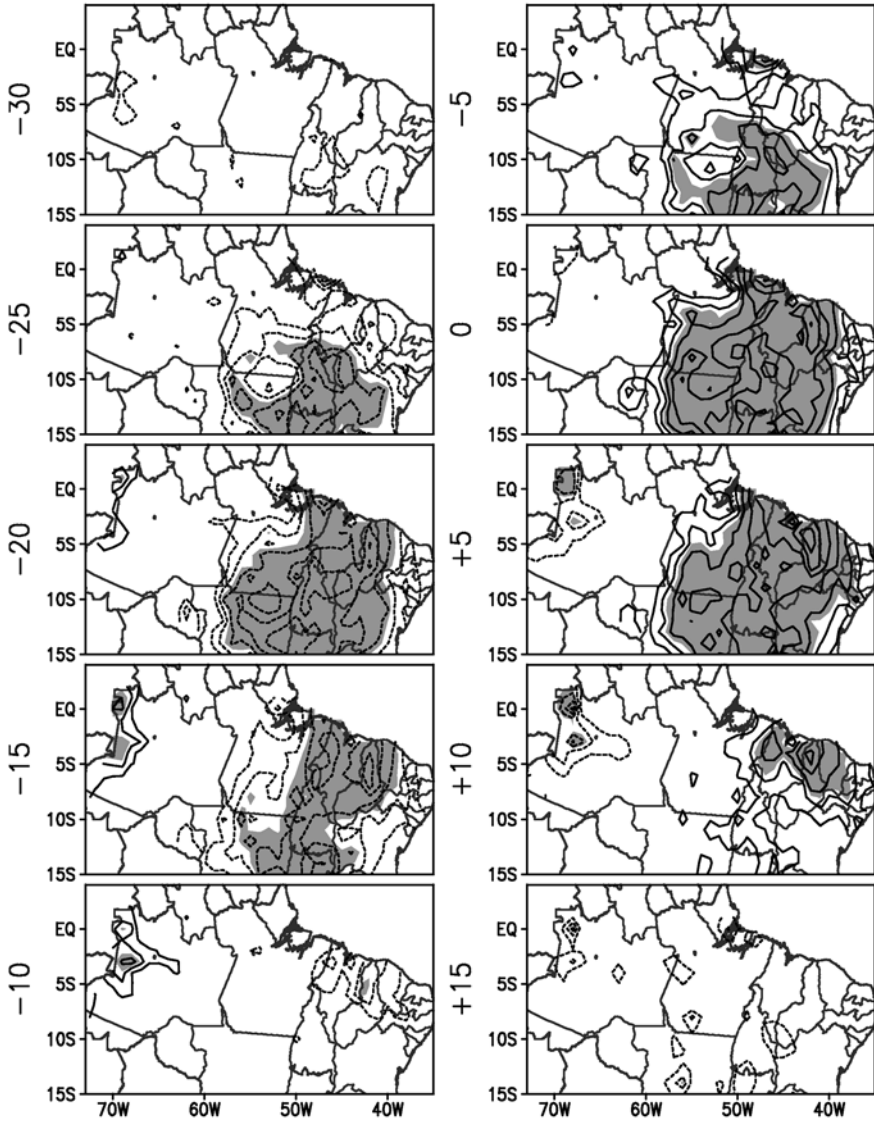
Fig. 7.11 Composites of rainfall anomalies over Tropical Brazil for pentads from day –30 to +15. Nonzero values are contoured every 0.5 mm/day with solid (dashed) contours for positive (negative) anomalies. Shaded areas indicate statistically significant anomalies at 95% local confidence level (After Souza and Ambrizzi 2006)

approximately to the second and fourth panels in the MJO-related OLR evolution displayed by Ferranti et al. 1990). Observational studies confirm this association between enhanced SACZ and an extratropical wave train in the MJO time scale (e.g., Casarin and Kousky 1986; Horel and Jones 1990, Fig. 7.12).
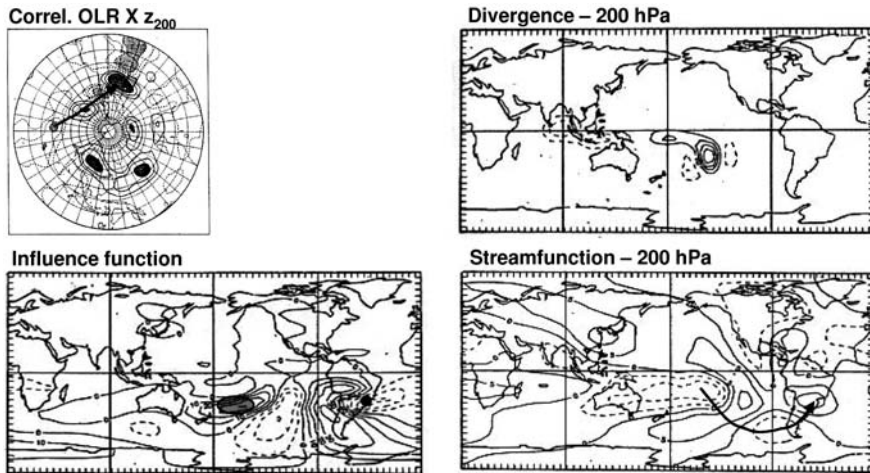
Fig. 7.12 (Upper left) Correlation analysis between OLR at the SACZ and geopotential height at 200 hPa, both filtered to retain only intraseasonal variability (from Jones and Horel 1990). (*Lower left*) Influence function for the target point at the center of the cyclonic anomaly associated with enhanced SACZ, with the region of maximum values indicated by the shaded ellipse. (*Upper right*) Prescribed anomalous divergence at 200 hPa, reproducing the shifted SPCZ at one of the MJO phases, and (*lower right*) resulting streamfunction at 200 hPa (From Grimm and Silva Dias 1995b, reproduced with permission of the American Meteorological Society)

The examples of tropical-extratropical and tropical-tropical teleconnections presented so far described remote influences on the SA circulation and precipitation. However, it is important to mention that SA anomalous tropical convection is also able to excite Rossby wave dispersion, even inter-hemispheric, through the region of climatological upper-level westerlies in the equatorial Atlantic. For example, Grimm and Silva Dias (1995b) show (Fig. 7.13) that the anomalous convection over central-east Brazil (or the SACZ shifted northward) associated with one phase of the MJO generates a wave train that propagates over North Atlantic and Eurasia, which shows reasonable agreement with the composites of streamfunction associated with this phase of the MJO (e.g., Weickmann et al. 1985; Knutson and Weickmann 1987). As this wave train is able to influence convection in the western Pacific/ Indian Ocean
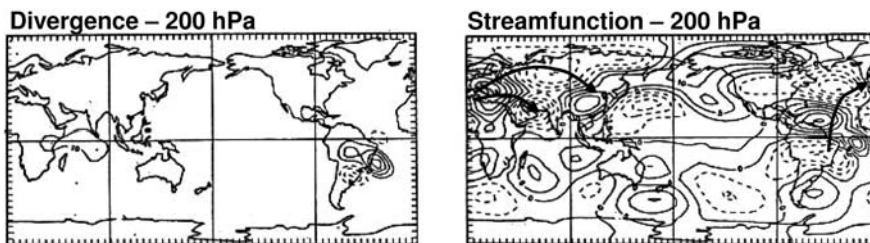
Fig. 7.13 (*Left*) Anomalous divergence at 200 hPa corresponding to enhanced SACZ at one of the MJO phases; (*Right*) resulting streamfunction at 200 hPa (From Grimm and Silva Dias 1995b, reproduced with permission of the American Meteorological Society)

and thus trigger a new pulse of the MJO, there is another dynamical mechanism for explaining the intraseasonal variability associated with the MJO (Grimm and Silva Dias 1995b).

7.5 Final Remarks

Tropical-tropical and tropical-extratropical teleconnections have an important role in generating low-frequency variability of the atmosphere, as demonstrated in many studies. The relationship between tropical anomalous heat sources and atmospheric circulation anomalies provides longer term predictability, which needs to be realized.

Observational and theoretical studies have done much to examine the role of tropical forcing (e.g. review by Trenberth et al. 1998), but several issues remain to be addressed in more detail in order to better understand the interannual and intraseasonal variations derived both from tropical forcing and from boundary forcing or internal variability at higher latitudes. There are a number of factors influencing the teleconnections through the atmosphere, such as the basic atmospheric state, feedback from transients, baroclinic and non-linear effects, feedbacks from induced changes in the underlying surface, such as changes in extratropical SSTs and land surface hydrology. Land-surface processes may be especially important in summer (e.g., Grimm et al. 2007; Grimm and Zilli 2009). Much research is still needed to explore all these issues.

This chapter provides only a brief summary of the main observed teleconnections affecting SA. However, besides ENSO and MJO, there are other possible sources of teleconnections with SA, generating regional patterns of interest, but they may have insufficient variance to be identified by the EOF analyses.

An important issue to be aware of is the dependency of the teleconnections on the basic atmospheric state through which they propagate. Slow variations on interdecadal time scales or longer can affect teleconnections that are effective in present time, shifting the teleconnection patterns and the resulting impacts (e.g., Grimm et al. 2006).

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