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Soil-precipitation feedbacks during the South American Monsoon as simulated by a regional climate model

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Abstract We summarize the recent progress in regional climate modeling in South America with the Rossby Centre regional atmospheric climate model (RCA3-E), with emphasis on soil moisture processes. A series of climatological integrations using a continental scale domain nested in reanalysis data were carried out for the initial and mature stages of the South American Monsoon System (SAMS) of 1993–92 and were analyzed on seasonal and monthly timescales. The role of including a spatially varying soil depth, which extends to 8 m in tropical forest, was evaluated against the standard constant soil depth of the model of about 2 m, through two five member ensemble simulations. The influence of the soil depth was relatively weak, with both beneficial and detrimental effects on the simulation of the seasonal mean rainfall. Secondly, two ensembles that differ in their initial state of soil moisture were prepared to study the influence of anomalously dry and wet soil moisture initial conditions on the intraseasonal development of the SAMS. In these simulations the austral winter soil moisture initial condition has a strong influence on wet season rainfall over feed back upon the monsoon, not only over the Amazon region but

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in subtropical South America as well. Finally, we calculated the soil moisture–precipitation coupling strength through comparing a ten member ensemble forced by the same space–time series of soil moisture fields with an ensemble with interactive soil moisture. Coupling strength is defined as the degree to which the prescribed boundary conditions affect some atmospheric quantity in a climate model, in this context a quantification of the fraction of atmospheric variability that can be ascribed to soil moisture anomalies. La Plata Basin appears as a region where the precipitation is partly controlled by soil moisture, especially in November and January. The continental convective monsoon regions and subtropical South America appears as a region with relatively high coupling strength during the mature phase of monsoon development.

1 Introduction

South America extends across the equator from about 10° N to 55° S and has unique surface features from the world's largest rain forest in Amazonia to the driest desert in northern Chile and a high desert in the Altiplano. The South American Monsoon System (SAMS, Nogués-Paegle et al. 2002) dominates the mean seasonal cycle of precipitation in tropical and subtropical latitudes. The timing of its onset and duration and the frequency and intensity of daily rainfall have important implications for agriculture, hydroelectric power generation, and local ecosystems throughout large regions of tropical and subtropical South America. The large-scale land cover changes and the shift in population to high density urban areas have put supplementary stress on water resources. The correct simulation of the SAMS is essential for seasonal climate forecasting and for studying the interannual variability and the long-term changes of the regional precipitation and for determining the climatic impact of land use.

The first studies that addressed the land surface influence in the Amazon region were deforestation experiments performed with general circulation model GCMs (Dickinson and Henderson-Sellers 1988; Lean and Warrilow 1989; Shukla et al. 1990; Nobre et al. 1991). All these authors found that precipitation decreased as a result of decreased evapotranspiration and moisture convergence. More recent GCM and RCM studies (e.g. Fennessy and Shukla 1999; Costa and Foley 2000; Misra et al. 2002; Baidya Roy and Avissar 2002; Avissar and Werth 2005) have identified the sensitivity of rainfall to changes in vegetation and soil moisture conditions in the region. According to the majority of modelling studies on the effects of large-scale deforestation in Amazonia, desertification results in hydrological cycle weakening. However, assessments also indicate that this effect may be modified by changes in atmospheric moisture convergence, that there are significantly different responses to similar land use changes in different tropical regions and that responses are typically linked to dry season conditions (e.g. Voldoire and Royer 2004; Feddema et al. 2005).

The soil moisture memory potentially contributes to atmospheric variability and seasonal predictability and could influence the development of the SAMS. In a study using ERA15 data, Fu and Li (2004) and Li and Fu (2004) found that the continental surface conditions seem to control the onset date of the monsoon, and in particular that an anomalously dry land surface during the dry season could delay the onset of SAMS with as much as 2 months. Collini et al. (2008) showed similarly that October precipitation was more responsive to reductions than to increases in initial soil moisture using a regional mesoscale model. They found that reductions in initial soil moisture produced almost linear reductions in precipitation over the monsoon region, principally because of the more stable boundary layer that results from the increase of the Bowen ratio. Xue et al. (2006) analyzed the role of vegetation biophysical processes in the structure and evolution of SAMS through (GCM) experiments with different land surface parameterizations. The inclusion of an explicit representation of vegetation processes modified the Bowen Ratio and led to a more realistic simulation of precipitation amount, but also of the spatial and temporal evolution of the monsoon since the division of the surface fluxes influence the continental scale circulation.

However, the simulation of soil moisture–precipitation feedback processes is an issue that has not been fully addressed in South America yet. The purpose of this paper is to summarize the recent progress in regional climate modeling in South America using the Rossby Centre regional atmospheric climate model (RCA3, Kjellström et al. 2005; Samuelsson et al. 2006), with emphasis on the interaction between the simulated rainfall and soil moisture processes. Our objectives are threefold: (a) To isolate the role of including a spatially varying soil depth; (b) to examine the influence of soil moisture initial conditions on SAMS development; and (c) to explore the soil moisture–precipitation coupling strength (defined as the degree to which prescribed soil moisture conditions affect precipitation).

2 Model description and assessment of simulated rainfall

2.1 Model description

The Rossby Centre Regional Atmosphere Model, RCA, is a hydrostatic, primitive equation grid-point limited area model. In this work we use the most recent version of RCA, called RCA3 (Kjellström et al. 2005), modified by including the surface database Ecoclimap (Champeaux et al. 2005) and by adjustments in the atmospheric physics to improve the performance for tropical and subtropical climates (RCA3-E). Details on the physical parameterizations, including changes in the radiation, turbulence and cloud parameterizations in RCA3 compared to earlier versions, as well as recent updates regarding technical aspects, are described in http://www.smhi.se/sgn0106/if/rc/rca.htm. The model domain is based on a rotated grid system with a horizontal resolution of 0.5° and 24 unevenly spaced sigma levels in the vertical.

The land surface scheme of RCA3 (Samuelsson et al. 2006) employs the tile approach (van den Hurk et al. 2000) for calculation of surface fluxes. The surface of each grid box is decomposed in tiles according to the sub grid vegetation cover and the surface fluxes are calculated separately for each tile. The main tiles are open land and forest, the open land tile being divided in a vegetated and a bare soil sub tile while the forest tile is divided in forest canopy and forest floor. Snow is treated separately in both open land and forest. According to the fractional area of each tile, the individual fluxes from the tiles are weighted to grid-averaged values at the lowest atmospheric layer. The land surface scheme includes processes such as interception of rain and canopy transpiration controlled by photosynthesis to describe the surface water balance. In the current version, the Ecoclimap database has been implemented in the RCA3-E model in order to initialize and drive its soil–vegetation–atmosphere transfer scheme. Ecoclimap is a complete and coherent surface dataset based on a very high-resolution classification of a large number of homogeneous ecosystems and contains all the necessary surface parameters (e.g., roughness length, vegetation fraction, leaf area index, albedo, rooting depth). The soil moisture in RCA3 has two prognostic soil moisture storages, the top layer which has a depth of 7 cm, and the deep layer which depth is given by the database Ecoclimap.

2.2 Evaluation of simulated rainfall

We performed a 20-year simulation of present climate (1980–1999 with 1 year of spin-up) with initial and boundary conditions from the European Centre for Medium-Range Weather Forecasts (ECMWF) 40-year Reanalysis (ERA-40, Uppala et al. 2005). The simulated precipitation was evaluated against high-resolution ($0.5° \times 0.5°$) precipitation data compiled by the Climatic Research Unit (CRU) of the University of East Anglia (New et al. 1999, 2000) focusing on September through March, which is the period of the SAMS. Some caveats should be pointed out concerning this dataset. New et al. (2000) show examples of station densities used in developing the CRU dataset. Over large regions in South America the station density is relatively low. Consequently the interpolation procedure used in developing the CRU dataset might affect the fine scale structure of the actual field. However, our choice of analyzing broad structures/regions compensates for this weakness.

In comparison to CRU data, RCA3-E captures many aspects of the observed annual mean precipitation, but underestimates the rainfall over parts of northern Amazonia and central Brazil, and over some areas of south-eastern South America: southern Brazil, Uruguay and north-eastern Argentina (Fig. 1). The precipitation is overestimated in parts of northern Brazil (around 5° S), western Amazonia and along the Andes. Most of these biases in the simulations of the regional climate

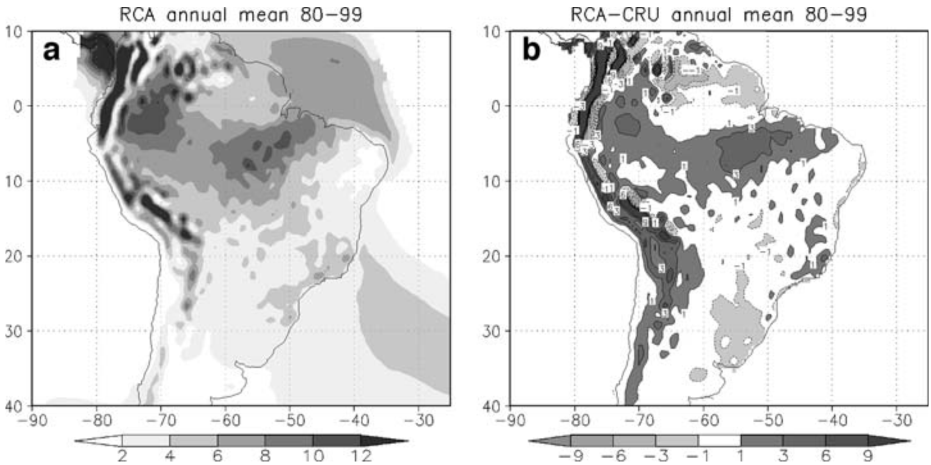

Fig. 1 Annual-mean precipitation (mm/day) 1980-01-01–1999-12-31. **a** RCA3-E, **b** RCA3-E minus CRU

are systematic across many GCMs and regional climate models (RCM) (Christensen et al. 2007). In particular, the underestimation of rainfall over southeastern South America is of special concern for CLARIS. These biases over regions with relatively flat terrain, common to many models, remain for the most part unexplained and the search for their responsible physical mechanisms will be challenging (Menéndez et al. 2009, this issue).

To outline more clearly the premonsoon and the monsoon evolution we show the monthly mean precipitation for the 20 years of simulation, zonally averaged (over land only) between 60° W and 40° W from August through March for CRU, RCA3-E and RCA3-E minus CRU (Fig. 2). The regional model simulates an early

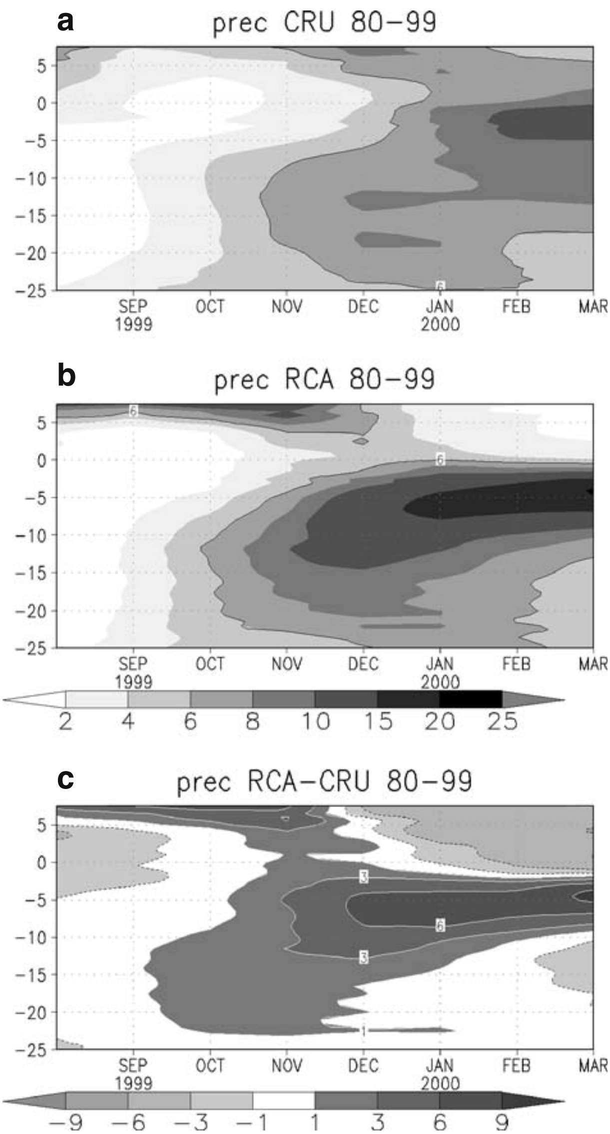

Fig. 2 Temporal evolution of the monthly mean precipitation (mm/day) averaged over 40°–60° W (land only) from August through March: **a** CRU, **b** RCA3-E, and **c** RCA3-E minus CRU. *Vertical axis* shows latitudes

onset of the monsoon (see e.g. the isohyet of 6 mm/day as a reference) at almost all latitudes south of the equator. Between 0° and 10° S the precipitation increases too rapidly and obtains too high values from November through March. North of the equator a precipitation maximum occurs in December in CRU and shows a weak intraseasonal variability on the monthly timescale. In RCA the maximum occurs in October–November and the model then dries out from December to March. South of 10° S RCA captures better the precipitation in comparison to CRU for the whole period.

Overall the main large-scale spatial patterns of annual and monthly precipitation for South America exhibit a reasonably good agreement with observations and allow us to pursue this study with some confidence in the realism of the results. However, as South America and surrounding oceans is a data-poor region, the actual model skill is masked by existing uncertainties in the lateral boundary conditions used to drive the model and in the observational-based datasets used in its evaluation.

3 Soil depth sensitivity

The amount of water that is available in the soil for evaporation back into the atmosphere will depend, among other factors, on the soil and rooting depth. Land surface parameterizations in both global climate models and RCMs generally use values of about 2 m for rooting depth (e.g. the current version of RCA3 used over Europe employs a constant soil and rooting depth of 2.2 m for all regions but mountain regions where it is set to 1.0 m). This is in contrast to the observational-based data, for example in the Amazon basin deep roots of several meters was found by Nepstad et al. (1994).

The interest in focusing on the soil depth is motivated by two factors: (1), the soil depth of tropical forest that cover large areas of northern South America are increased to 8 m with the incorporation of Ecoclimap in the model, and (2) previous works suggest the importance of soil depth and deep rooted vegetation on the climate system. Kleidon and Heimann (2000) investigated this aspect in the context of the climatic effects of large-scale deforestation in Amazonia. They found that most of the regional and remote changes can be attributed to the removal of deep roots. van den Hurk et al. (2005) analysed the soil hydrological memory in the Rhine basin using large scale analyses of atmospheric water convergence and river discharge. They concluded that the depth of the hydrological soil reservoir in RCMs is indicative for the strength of the hydrological response of the whole river basin to a global temperature increase, and that a proper specification of this depth is an important factor. Therefore, regional simulations with deficient representation in parameters of the underlying physical environment such as soil depth possibly include associated errors not only in the computation of the evapotranspiration and heat fluxes but also in the climatic sensitivity.

In order to estimate the impact of introducing a spatially varying soil depth in the model on the development of the SAMS, we performed two ensembles of five members with different initialization date, each one of the members including the period September 1st 1992 through March 31 1993. An analysis of the time evolution of the soil moisture of a multi-year integration with RCA3-E initialized and forced by ERA-40 showed that the soil moisture spin-up time can be up to 2 years for regions

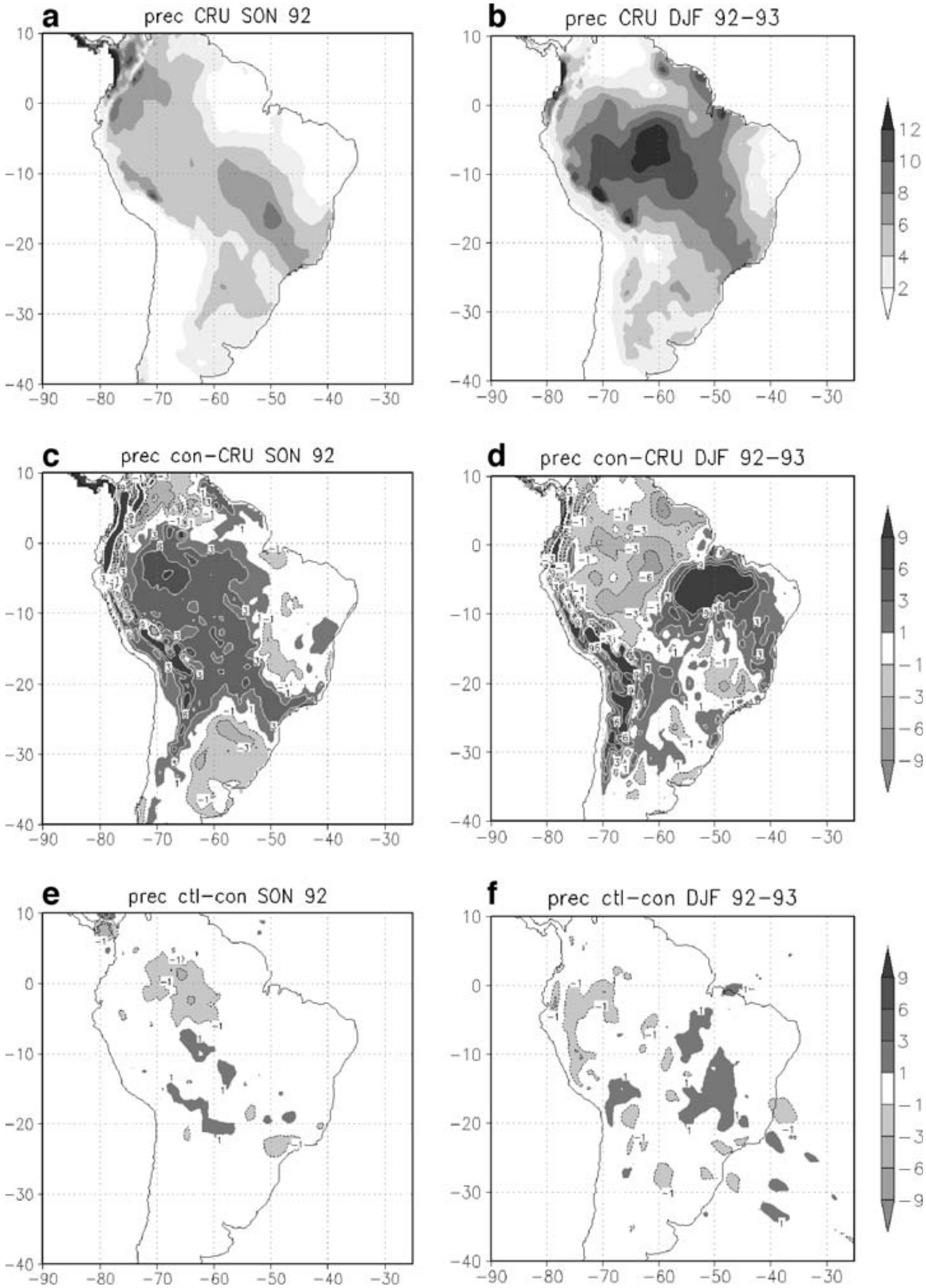

Fig. 3 Mean rainfall (mm/day) for left column the spring (*SON*) 1992 and right column summer (*DJF*) 1992–93: **a** and **b** CRU climatology, **c** and **d** bias with respect to the ensemble with constant soil depth (*CON*), and **e** and **f** difference between the two ensembles (*CTL–CON*: variable soil depth minus constant soil depth)

with deep rooting depth in Amazonia (not shown). To initialize the model with the atmosphere–soil moisture in equilibrium without a long spin-up time, the soil moisture initial conditions are set to the soil moisture fields of corresponding initial date from a RCA3-E/ERA-40 integration initialized 1st September 1990. Ensemble CTL was run with soil depth from the new Ecoclimap database while ensemble CON with the usual constant soil depth (2.2 m). Figure 3 compares the CRU precipitation climatology for the spring and summer 1992–93 with the simulated ensemble means for the simulations performed with constant soil depth (CON) and with variable soil depth (CTL). The inclusion of a spatially varying soil depth tends to reduce the bias in spring and to enhance it in summer over Amazonia (Fig. 3a–c and d–f respectively). Further south, over tropical regions, the positive precipitation bias in spring was increased in CTL, likely due to an enhanced southward transport of atmospheric moisture associated with the South American Low Level Jet (SALLJ). The characteristics of the SALLJ when supplying moisture to subtropical latitudes have been discussed in many articles (e.g., Berbery and Collini 2000; Marengo et al. 2004). During summer the difference between both ensembles is largest over Brazil and tends to increase the precipitation over the area affected by the South Atlantic Convergence Zone (SACZ, a south eastward extension of cloudiness and precipitation from the southern Amazon towards southeast Brazil and the neighbouring Atlantic Ocean).

4 Sensitivity to soil moisture initial conditions

In this section we explore the influence of anomalous soil moisture initial conditions in late austral winter on the intraseasonal development of the SAMS through two ensembles of simulations initialized with highly idealized and extreme anomalous surface conditions of soil moisture. Our study covers the monsoon of 1992–93; however, some authors have suggested that the surface and dynamical processes of the SAMS act independently of the large-scale conditions. Fu et al. (1999) analyzed the onset of the monsoon using satellite radiances, radiosondes, and assimilation data and found that the forcings that control the onset of the monsoon are the same for El

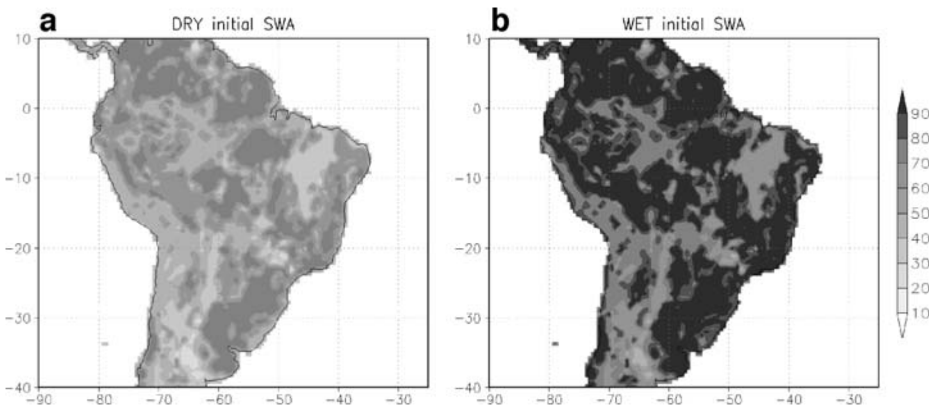

Fig. 4 Initial soil water availability (*SWA*) of **a** ensemble DRY and **b** ensemble WET

Niño and La Niña event. Collini et al. (2008) draw similar conclusions in a regional climate model study of several October months.

We performed two ensembles with anomalously dry and wet land surface initial conditions over the whole domain. As in the rooting-depth study above, the ensembles have five members initialized on different dates, all members including the period 1 September 1992–31 March 1993. These ensembles will in the following be called "DRY" and "WET" respectively. The initial soil water availability (SWA) for the two simulations was modified from the SWA of the driving reanalysis of the corresponding initialization dates (SWA_{ERA40}). The SWA_{ERA40} was multiplied by a factor 0.2 to generate dry conditions, and to generate wet conditions without allowing supersaturation we used the formula $SWA_{WET} = SWA_{ERA40} + (1 - SWA_{ERA40}) * 0.8$. The initial SWA fields of the two ensembles are shown in Fig. 4.

Figures 5 and 6 shows the monthly mean precipitation for the DRY and WET experiments of the period October through January (regions of more than

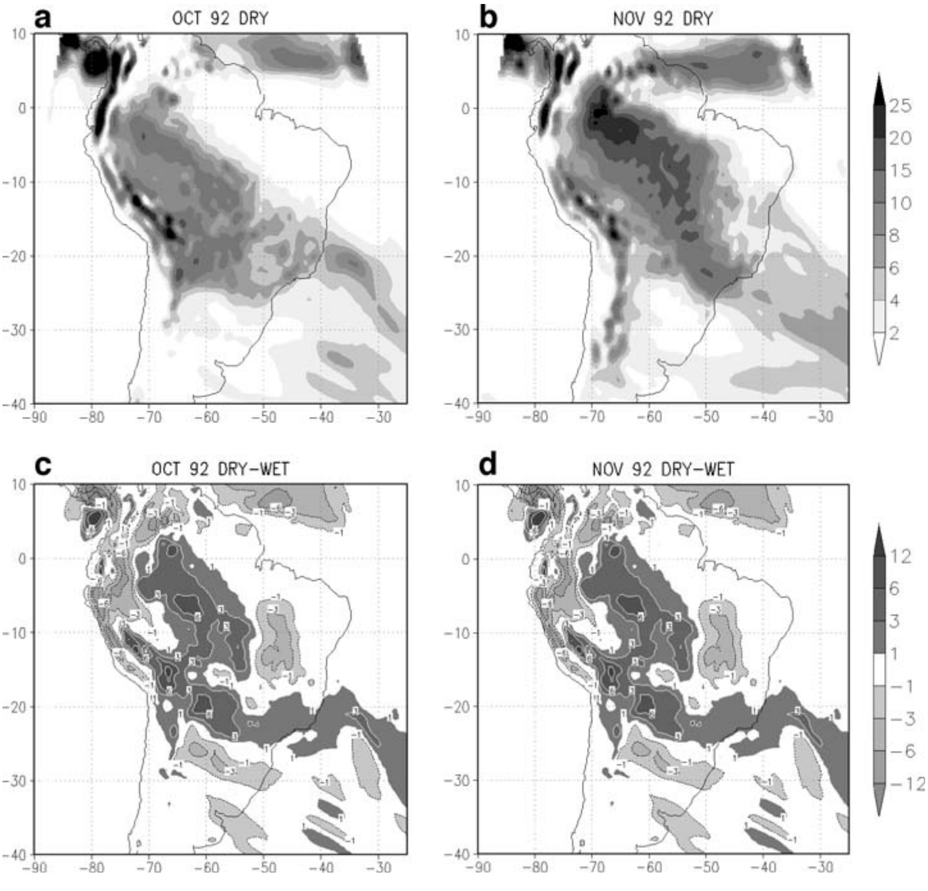

Fig. 5 Monthly precipitation for October and November (mm/day) for upper panel DRY and lower panel DRY–WET. *Grey shading* levels are 1, 3, 6, 10, 15, 20 and 25 mm/day (*upper panel*) and ±1, ±3, ±6, ±12 mm/day (*lower panel*). The line ±6 mm/day is *highlighted*

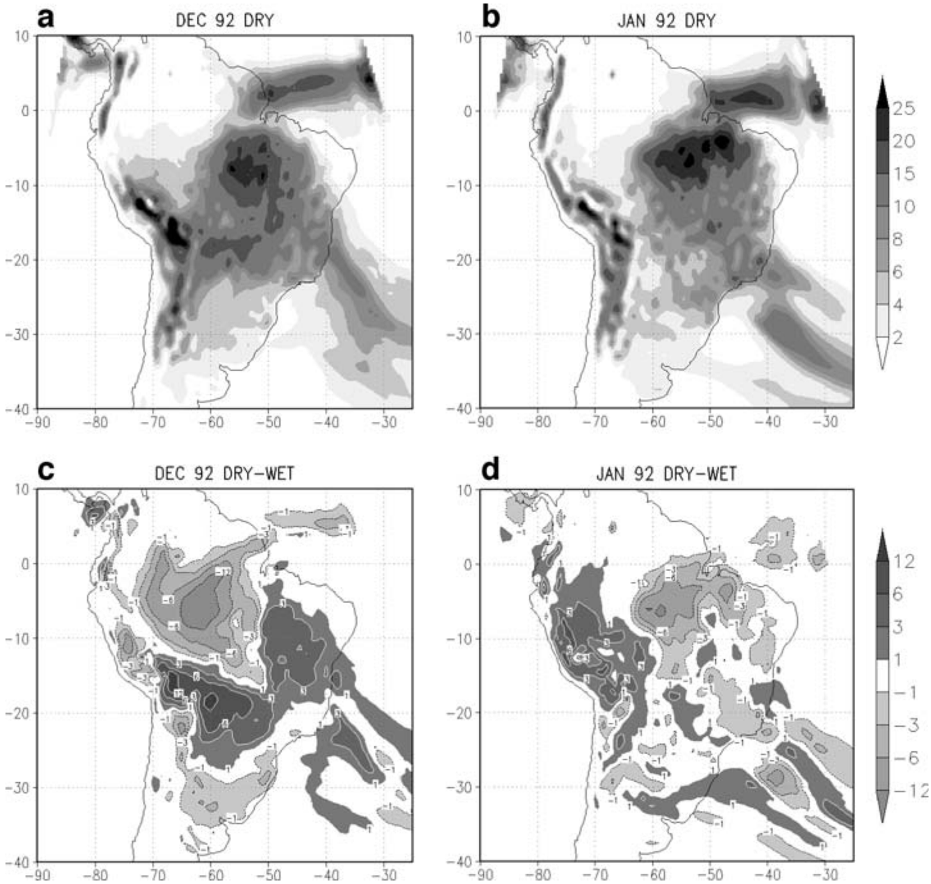

Fig. 6 As Fig. 5 but for December and January

±6 mm day^{-1} are highlighted for visual reference). This is a complex system with land surface–atmosphere interactions depending on numerous factors. The figures suggests that the soil moisture initial condition has a strong influence on wet season rainfall over the continental convective monsoon regions. However, a wetter (dryer) land surface does not always coincide with more (less) precipitation. In the comparison DRY – WET, some areas are drier where the monsoon either was delayed or could not reach any further development, and others are wetter due to redistribution of the circulation or changes in position of the maximum precipitation band. The changes of land surface conditions also affected the precipitation over ocean due to the impact of land–atmosphere interaction on circulation, similar to Sato et al. (1989) and Xue et al. (2006). This is consistent with recent studies on tropical deforestation in the Amazon Basin suggesting that land surface conditions affect the sea surface temperature in the nearby ocean, further amplifying teleconnections (Avissar and Werth 2005; Feddema et al. 2005; Voldoire and Royer 2005). Compared to experiment DRY, experiment WET increases precipitation along the Intertropical

Convergence Zone (ITCZ) in the development phase of SAMS (October–November). Compensating subsidence produce large areas of decreased precipitation further south in tropical South America. During the mature phase of monsoon development, experiment WET enhances rainfall in western Amazon Basin (December) and in central Amazon Basin and in the SACZ region (January). Experiment WET tends to produce a weakened SACZ shifted southward and increased rainfall over large areas of subtropical South America.

In Fig. 7 we illustrate the monthly mean precipitation, zonally averaged between 60° W and 40° W, from August through March. The development phase of SAMS

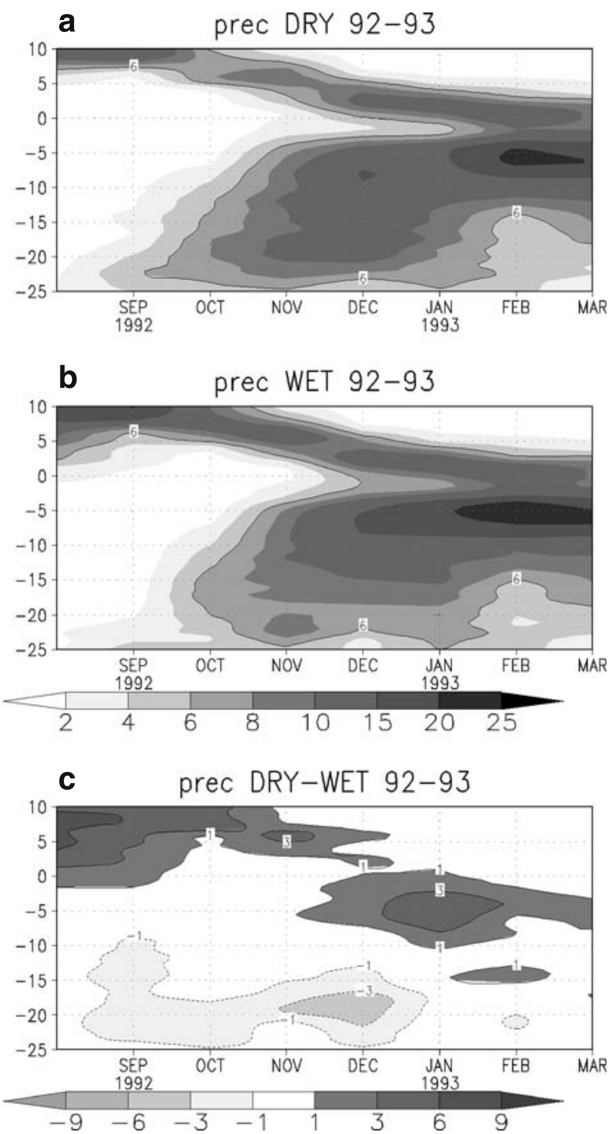

Fig. 7 Temporal evolution of the monthly mean precipitation (mm/day) averaged over 40°–60° W from August through March: **a** DRY, **b** WET, and **c** DRY minus WET. *Vertical axis* shows latitudes

during austral spring of 1992 was characterized by the presence of the ITCZ in the northern part of the domain; with a tendency toward moving southward especially in experiment DRY. A strong precipitation band between 15° S and 25° S appears from October in experiment DRY and somewhat later and weaker in experiment WET. Meanwhile, the rapid southward shift of the region of intense convection from the equator toward the southern Amazon Basin is manifested earlier in experiment WET (in December). During the mature phase, rainfall intensity is heavier in case WET over the SAMS core region, but case DRY simulates stronger precipitation further south over eastern Brazil and the nearby Atlantic.

In order to provide a more detailed picture of the model's sensitivity than the monthly mean values, Fig. 8 displays the histograms of daily rainfall on different intensity classes over two continental areas where January sensitivity is particularly strong: Amazonia and the upper la Plata Basin (monthly mean rainfall decreases in Amazonia and increases in upper la Plata Basin in experiment DRY). The methodology is to count for each grid point, the total number of days within each interval representing dry days (0–0.5 mm/day) and light (0.5–6 mm/day), moderate (6–15 mm/day), strong (15–30 mm/day) and heavy (>30 mm/day) precipitation days. The effect of soil moisture late winter initial conditions on the frequency distribution of the daily rainfall rates in January shows a considerable spread among the different regions. Over Amazonia, dry surface initial conditions tend to decrease the number of intense convective rainstorms (i.e. heavy rainfall days), consistently with a monthly

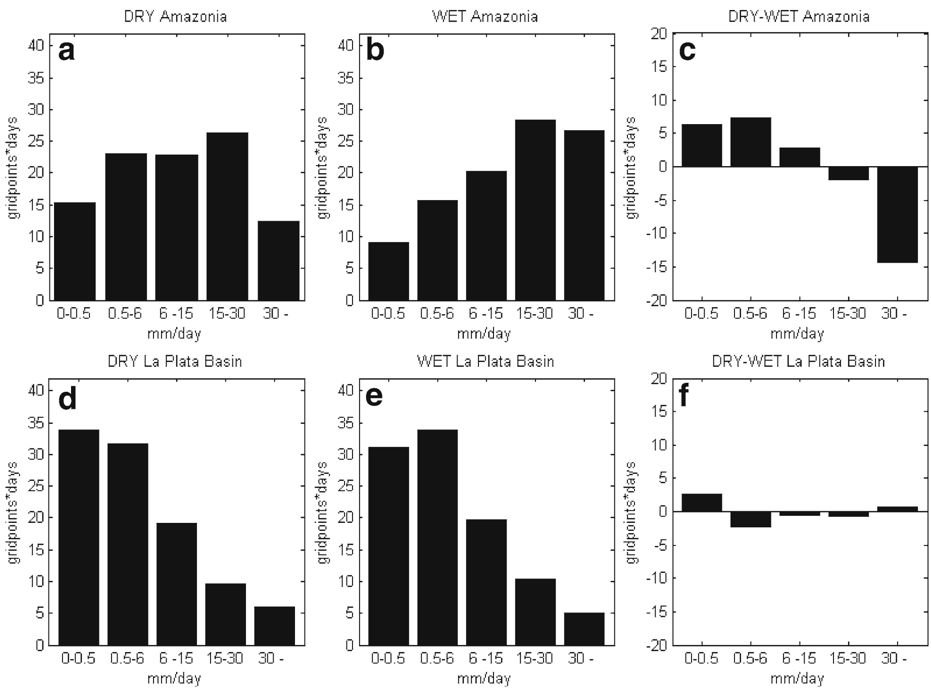

Fig. 8 Histograms of daily January precipitation rates (mm/day) over **a**–**c** Amazonia (3° S–8° S, 60° W–50° W) and **d**–**f** la Plata Basin (25° S–20° S, 55° W–48° W), *left column* DRY, *middle column* WET and *right column* DRY–WET

mean decrease in the rainfall amount. In upper la Plata Basin, the number of dry and light precipitation days (0.5–6.0 mm/day) increases in experiment. We in effect notice an increased number of strong and heavy rainfall days as a response to decreased initial soil moisture (probably due to the increased convergence of water vapour content in the region).

5 Coupling strength

We explored another way of determining soil moisture influence on the South American climate through calculating the coupling strength between soil moisture and precipitation. Coupling strength is defined as the degree to which all prescribed boundary conditions affect some atmospheric quantity. Within the GLACE project (Koster et al. 2003, 2004, 2006; Guo et al. 2006); the coupling strength between soil moisture and atmosphere for global atmospheric models has been explored over the northern hemisphere for boreal summer, a season where soil moisture–land coupling could be comparable or even stronger than sea surface temperature (SST)–land coupling for midlatitudes (Koster et al. 2000). In our case, we are interested in documenting the degree to which the precipitation responds to soil moisture anomalies during the SAMS. Coupling strength is still largely unknown for South America and is a very uncertain aspect of regional modelling. The methodology essentially follows the GLACE study.

Two ensembles (called W and S) of ten members each were created, starting from different initial dates. Each member includes the 120-days-period November 1st 1992 through March 31 1993. To avoid a long spin-up time, the soil moisture is initialized as was described in Section 3. Other initial and boundary conditions are taken from ERA-40.

Ensemble W: Model with a fully land surface–atmosphere interaction. The soil moisture that is “seen” by the atmosphere is calculated by the model at each time step and the only difference between members is the initialization date. Ensemble S: The ensemble members are forced, at each time step, to maintain the same space–time varying series of top and deep soil moisture. These series are obtained from a previous simulation of this period from which we save top and deep soil moisture every 30 min. Consequently, between the soil moisture and other components of the system, and in particular the water budget, there is only a one way interaction. The soil moisture influence the precipitation, evaporation and surface temperature e.g., but these variables do not feed back upon soil moisture.

Since the initial dates are the same for the two ensembles the only difference between ensemble W and S is that soil moisture is equal among members in ensemble S while it differs among members of ensemble W. The similarity between members of one ensemble for any atmospheric variable x is calculated as follows for all grid points

$$\Omega_x = \frac{m\sigma_{x^\wedge}^2 - \sigma_x^2}{(m-1)\sigma_x^2}$$

where $\sigma_{x^\wedge}^2$ is the variance of the mean time series of all members of one ensemble, σ_x^2 is the ensemble intermember variance which is obtained by calculating the variance among all time steps and ensemble members and m is the number of ensemble members. Ω is interpreted as the fraction of the variance that is explained by boundary and initial conditions (the total variance depend on internal variability of the model and on boundary and initial conditions). The similarity is 0 if there is no correlation among ensemble members and 1 if the time series of x are equal for all ensemble members. From this interpretation and from the fact that ensemble S is driven by a larger set of forcing variables than W, we expect that Ω will be larger for ensemble S in regions were the soil moisture explains some of the variance of the variable x. The coupling strength (CS) between soil moisture and x is defined as the difference between the similarities of the two ensembles:

$$\mathrm{CS} = \Omega_x(S) - \Omega_x(W)$$

A property of this index worth noting is that the coupling strength is a measure of the degree to which the whole forcing field of soil moisture influences on the variable, and is not an estimate of the local evaporation–precipitation recycling.

In Fig. 9 we show the coupling strength for November, January and March, representing different stages of the monsoon development. We can identify Northeastern Brazil as a region with relatively high coupling strength in all months. At other low latitudes it is difficult to identify any consistent spatial or temporal pattern; the coupling strength varies between small positive and small negative values. This is because the model's internal variability is very high in this tropical region during austral summer (not shown), so that the fraction of the variance of precipitation that is explained by the atmospheric noise is much higher than the fraction explained by the boundary condition. Ω is therefore low in both ensembles in this region (not shown). The region La Plata Basin, which is of special interest for the CLARIS project, includes eastern and northern Argentina, Southeastern Bolivia, Paraguay, Uruguay and southern Brazil. In this study, La Plata Basin appears as a region where the precipitation is partly controlled by soil moisture, especially in November and January.

It has been suggested that the coupling strength between precipitation and soil moisture should be highest in transition zones between arid and humid climates (e.g. Koster et al. 2004; Guo et al. 2006). To test this hypothesis with RCA3-E, we organized the grid points of the whole area shown in Fig. 9 according to five intervals of SWA and calculated the average coupling strength for each interval (Fig. 10) for the 3 months. SWA is a measure of the degree to which the transpiration is regulated by soil moisture stress and is expressed as a fraction between 0 (wilting

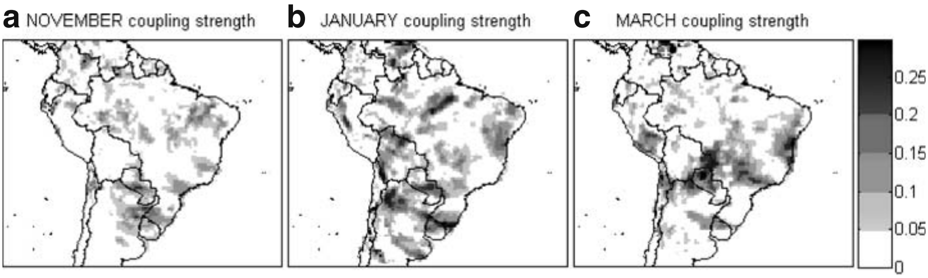

Fig. 9 The coupling strength between precipitation and soil moisture for **a** November, **b** January and **c** March

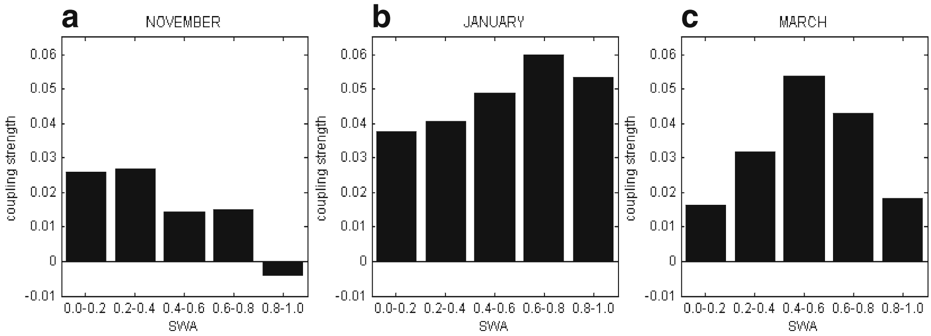

Fig. 10 Average of coupling strength for all grid points in the area shown in Fig. 8 binned according to different soil moisture availability **a** November, **b** January and **c** March

point) and 1 (field capacity). SWA depends on both the soil moisture content and the soil type. In November, the coupling strength was highest in arid regions and lower for regions with intermediate and high SWA, but in January and December the coupling strength is higher for intermediate values of SWA than for arid and wet regions.

The coupling strength between precipitation and soil moisture depends on the degree to which the evapotranspiration responds to soil moisture anomalies, and how these evaporation variations influences on precipitation. In humid zones the soil is saturated or almost saturated with water and the atmosphere moisture content and temperature is the limiting factor on evapotranspiration. Water that is added to the soil will therefore not result in higher evapotranspiration rates. On the contrary, in arid zones, the evapotranspiration rates are highly dependent on soil moisture anomalies since the atmosphere is not saturated with water vapour. However, the amounts of evapotranspiration will likely not be enough to generate precipitation in a region with a dry and stable boundary layer. Only in transition zones the evapotranspiration rate generated by soil moisture anomalies are sufficient to trigger precipitation.

6 Final remarks

We have described ongoing research using RCA3-E in South America with emphasis on soil moisture processes. Various experiments, initialized with the new Ecoclimap database and driven with ERA-40 reanalysis, were carried out using a continental scale domain. A present-day climate simulation was verified against available climatological precipitation. RCA3-E exhibits a reasonably good agreement with observations, although some deficiencies (often also found in other state-of-the-art global and regional models) are evidenced in the simulation of the regional precipitation. This study covers only one monsoon cycle (1992–93) but since previous studies have shown (Fu et al. 1999; Collini et al. 2008) that the surface and dynamical processes of the SAMS act in the monsoon region independently of the large-scale conditions we consider that our results have a certain degree of robustness although we do not dismiss the interest of repeating the experiments with different boundary forcing.

An objective of this work was to isolate the role on SAMS development of including a soil depth that extends as deep as 8 m in some areas of Amazonia. We have compared an ensemble of simulations which includes spatially varying soil depth to another with the standard constant soil depth. The role of the soil depth depiction was relatively less critical than expected, with both beneficial and detrimental effects on the simulation of the seasonal mean rainfall. However, it should be considered that the simulations were initialized in late winter, extending only throughout spring and summer. Kleidon and Heimann (2000) suggest that the incorporation of deep roots into a climate model would be important especially during the dry season (i.e. austral winter in South America), since during the wet season the soil moisture content is near field capacity due to heavy rains and the evapotranspiration is not limited by soil moisture. During the dry season though, the ever-green forest would be capable of transpiring considerable amounts of water throughout the dry season if deep soil depth and deep roots are included in the model. According to Kleidon and Heimann (2000), in that case, evapotranspiration and the associated latent heat flux are considerably increased and the enhanced atmospheric moisture is transported towards the main convection areas in the inner tropical convergence zone where it supplies more energy to convection thus intensifying the tropical circulation patterns. This effect still needs to be verified with RCA3-E and will be the subject of future research.

Another objective was to examine the influence of soil moisture initial conditions on the SAMS development. In this case, we have compared two simulations of the period 08/1992 to 03/1993 with modified initial soil moisture. Of course, studying the impact of soil moisture initial conditions constitutes a limited approach as part of the difficulty for understanding and simulating the hydrologic cycle in this region. In this simple and qualitative assessment of the soil–precipitation feedback, we have analyzed simulations with opposite soil moisture initial conditions in order to represent two highly idealized and extreme anomalous surface conditions during the late austral winter. Our results suggest that the initial springtime soil moisture conditions feed back upon the SAMS during the warm months, not only over Amazonia but in subtropical South America as well. This could be related with different mechanisms, e.g.: (1) anomalies in the Bowen ratio could affect the low-level jet and the associated transport of atmospheric moisture (as in Collini et al. 2008); and (2) changes in convection patterns can affect the Hadley Circulation and thus propagate climate perturbations into the subtropics (as suggested e.g. in Branstator 1983; Sadershmukh and Hoskins 1985; Figueroa et al. 1995). While future research needs to be developed to further assess these mechanisms, the fact that tropical regions have the potential to affect climates beyond their neighbouring area has been discussed in the recent literature (e.g. Voldoire and Royer 2004; Avissar and Werth 2005; Feddema et al. 2005).

In Koster et al. (2003, 2004) the coupling strength simulated by a dozen atmospheric GCMs was evaluated for the Northern Hemisphere summer. The results differ widely from model to model. This model dependency suggests that the physical processes occurring at the continental land surface are not properly represented in current climate models. These processes are indeed very numerous and intricately linked, being a function of the parameterizations controlling e.g. the land surface energy balance, the development of the boundary layer and the precipitation generation. Based on the analysis of two ensembles of simulations of the austral summer, our results indicate that the degree to which the atmosphere responds to soil moisture conditions is relatively large in North eastern Brazil and in the la Plata Basin. The reasons for the geographical variations in the coupling strength are not sufficiently clear and require additional analysis. Koster et al. (2004) suggest that in continental transition zones between wet and dry climates during summer, where boundary layer moisture can trigger moist convection and where evaporation is suitably high but still sensitive to soil moisture, we can expect soil moisture to influence precipitation. The regions of maximum coupling strength simulated by RCA3-E in January and March tend to be located in transition zones in South America and are qualitatively compatible with Koster et al. (2004). In order to address the realism of RCA3-E's coupling strength and sensitivity to soil moisture conditions more diagnostics and simulations are needed and, in particular, it would be useful to determine how it compares with other RCMs in this region.

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