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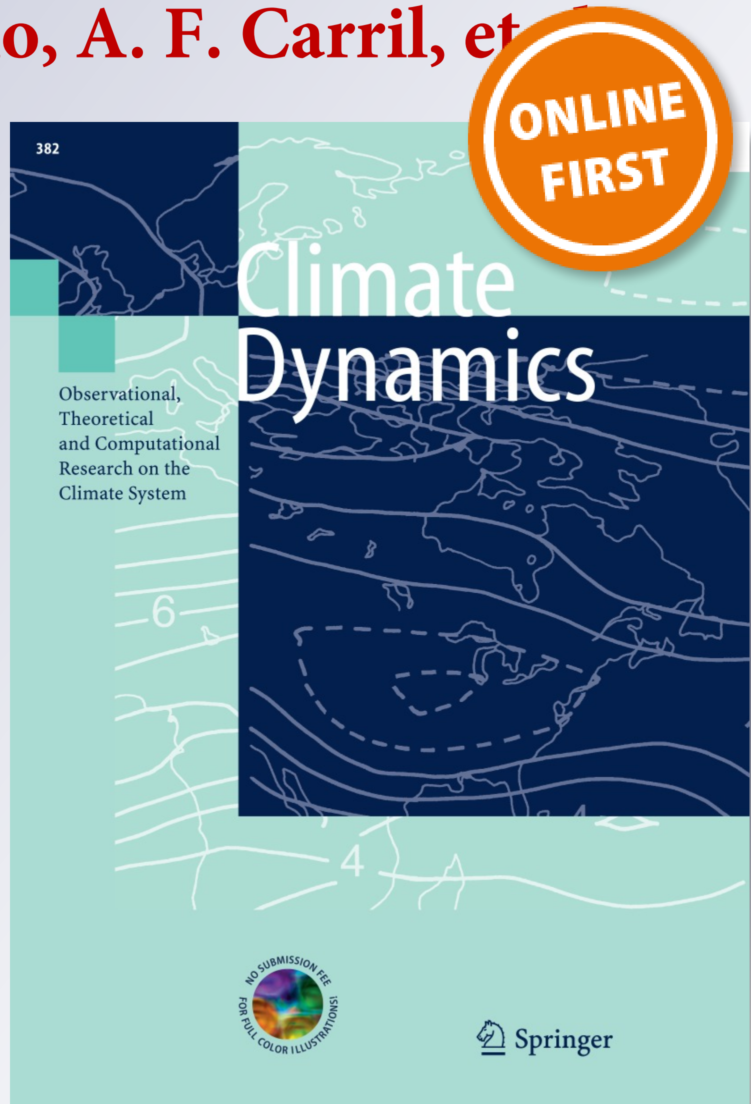

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Abstract The meteorological characteristics of the drought of 2005 in Amazonia, one of the most severe in the last 100 years were assessed using a suite of seven regional models obtained from the CLARIS LPB project. The models were forced with the ERA-Interim reanalyses as boundary conditions. We used a combination of rainfall and temperature observations and the low-level circulation and evaporation fields from the reanalyses to determine the climatic and meteorological characteristics of this particular drought. The models reproduce in some degree the observed annual cycle of precipitation and the geographical distribution of negative rainfall anomalies during the summer months of 2005. With respect to the evolution of rainfall during 2004–2006, some of the models were able to simulate the negative rainfall departures during early summer of 2005 (December 2004 to February 2005). The interannual variability of rainfall anomalies for both austral summer and fall over northern and southern Amazonia show a large spread among models, with some of them capable of reproducing the 2005 observed negative rainfall departures (four out of seven models in southern Amazonia during DJF). In comparison, all models simulated the observed southern Amazonia negative rainfall and positive air temperature anomalies during the El Nino-related drought in 1998. The spatial structure of the simulated

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rainfall and temperature anomalies in DJF and MAM 2005 shows biases that are different among models. While some models simulated the observed negative rainfall anomalies over parts of western and southern Amazonia during DJF, others simulated positive rainfall departures over central Amazonia. The simulated circulation patterns indicate a weaker northeasterly flow from the tropical North Atlantic into Amazonia, and reduced flows from southern Amazonia into the La Plata basin in DJF, which is consistent with observations. In general, we can say that in some degree the regional models are able to capture the response to the forcing from the tropical Atlantic during the drought of 2005 in Amazonia. Moreover, extreme climatic conditions in response to anomalous low-level circulation features are also well captured, since the boundary conditions come from reanalysis and the models are largely constrained by the information provided at the boundaries. The analysis of the 2005 drought suggests that when the forcing leading to extreme anomalous conditions is associated with both local and non-local mechanisms (soil moisture feedbacks and remote SST anomalies, respectively) the models are not fully capable of representing these feedbacks and hence, the associated anomalies. The reason may be a deficient reproduction of the land–atmosphere interactions.

Keywords Amazonia · Drought · Downscaling · Regional models

1 Introduction

Climatic and hydrological records in Amazonia show that occurring in a span of just 7 years, the severe droughts (2005 and 2010) and floods (2009 and 2012) that affected the region are considered to be the most intense extreme events in terms of rainfall and river level anomalies on record. Drought is a recurrent phenomenon in the Amazon region, and its impacts have been detected in various sectors, from the ecology and biodiversity to human activities and health. Historical regional and observational studies have identified drought episodes as early as in 1912, 1925, 1963, and later on in 1983, 1998, 2005 and most recently, 2010. A drought can be consider as an impact, generated either by a combination of deficient rainfall and high evaporation and temperature rates during the peak summer season (meteorological drought), or to anonaloulsy low river level anomalies during the fall-winter season consequence of the deficient previous summertime rainy seasons (hydrological drought). A hydrological drought may be a delayed response to meteorological drought. This study focuses on the meteorological drought in Amazonia in summertme of 2005.

Meteorological drought episodes are related to deficient rainy seasons, and there are some differences among drought episodes, since some of them were related to El Nino (1925, 1983, 1998) or to warming in the tropical North Atlantic (1963, 2005 and 2010). They also can be characterized by late onsets of the rainy season (or longer dry seasons) as in 2005 and 2010 (Marengo et al. 2011a, b). The El Nino related droughts are characterized mainly by rainfall reductions in central and eastern Amazonia, while those related to Tropical Atlantic sea surface temperatures (SST) anomalies favor rainfall reductions mostly over the southwestern Amazon region (see reviews in Ronchail et al. 2002; Marengo et al. 2008a, b, 2011a; Cox et al. 2008; Zeng et al. 2008; Espinoza Villar et al. 2009; Tomasella et al. 2011, 2013; Yoon and Zeng 2010; Samanta et al. 2010; Lewis et al. 2011).

In 2005, rainfall was below normal during the austral summer in southwestern Amazonia and the rainy season started later than normal, as shown by Marengo et al. (2008a). However, rainfall was not exceptionally low, at least as compared to the extremely low rainfall in the same season in 1998. The subsequent drought situation reported in the press in 2005 was better characterized by anomalously low river level anomalies during austral fall and winter (hydrological drought). In fact, in terms of impacts to the population, the perception of drought in Amazonia is based on anomalously low river levels/discharges and not much on rainfall anomalies. Of course, ecological impacts may be related to both; negative rainfall anomalies can increase the risk of fires, while anomalously low river levels can affect humid ecosystems.

The hydrologic response to deficient rainfall seems to be different in each case. The extremely low river levels during the drought of 2005, considered to be one of the most intense drought episodes in the last 100 years (Marengo et al. 2008a, b), affected large sections of southwestern Amazonia. This situation severely affected the population downstream along the Amazon River's main channel and its western and southwestern tributaries—the Solimões and Madeira Rivers. Navigation along these rivers had to be suspended because the water levels fell to historic lows (Tomasella et al. 2011, 2013).

In the present study we analyze simulations from seven regional climate models forced with the common boundary conditions provided by the ERA-Interim reanalyses for 1990–2008 used in the Europe South America Network for Climate Change Assessment and Impact Studies-CLARIS-LPB Project (Boulanger et al. 2011), to assess the meteorological aspects of the drought of 2005. This project aims at predicting the regional climate change impacts in South America, and at designing adaptation strategies for land use, agriculture, rural development, hydropower production, river transportation, water resources and ecological systems in wetlands.

Seven regional models were forced with the common boundary conditions provided by the ERA-Interim reanalyses for 1990–2008. We investigate rainfall, temperature, circulation and evaporation anomalies during the austral summer and fall of 2005, in both northern and southern Amazonia (Fig. 1) during the simulated period. Our emphasis is on the simulation of rainfall anomalies leading to the drought of 2005. Changes in circulation, temperature and energy balance from observations and reanalyses, and from the simulations of the regional models were used to assess rainfall anomalies during that drought episode. In some instances, we compare the meteorological components of the drought of 2005 with the previous intensely dry conditions and drought in Amazonia in 1998. We also discuss the capability and limitations of the regional climate models in simulating the observed interannual rainfall and temperature variability during the simulation period, and in addition the differences among models, focusing on differences among model structure and parameterization schemes (land surface and other physical processes).

While it would be a very interesting exercise to address the origin of the individual model differences, this would go beyond the scope of the present paper. We want to portray the broad inter-model differences mostly for the purpose assessing the quality of the simulated rainfall anomaly features in 2005 that led to low river levels and the hydrological drought in Amazonia observed in that year.

2 Data and methodology

Temperature fields from the Climate Research Unit CRU-University of East Anglia (New et al. 2000), and rainfall fields from the Global Precipitation Climatology project GPCC (orias.dwd.de/GPCC/; Rudolf et al. 2005) were used. There are some small differences in the depiction of rainfall anomalies during 2005 from various rainfall data sets, perhaps due to the interpolation techniques used, and since there is consistency among data sets and we have the experience of working with GPCC on different studies on climate variability (Marengo et al. 2008a, b, 2011a, b) in Amazonia, we decided to use GPCC rainfall for the region for the simulation period.

Time series for rainfall were built for the austral summer (DJF) and fall (MAM) seasons for northern-central Amazonia (75°W–50°W, 5°N–7.5°S) and southern Amazonia (75°W–50°W, 15°S–5°S) previously used in Marengo et al. (2008), and shown in Fig. 1. The wet season in southern Amazonia occurs in December to February (DJF) while for northern-central Amazonia the wet season is from March to May (MAM), according to the region's observed rainfall seasonal cycle (Figueroa and Nobre 1990). The peak of the river level/streamflow in Amazonian rivers in the southern part of the basin, affected by the drought in 2005 occurs in average during April-June.

The ERA-Interim is the latest global atmospheric reanalysis produced by the European Centre for Medium-Range Weather Forecasts (ECMWF). The ERA-Interim project was conducted in part to prepare for a new atmospheric reanalysis to replace ERA–40, which extended back to the early part of the twentieth century (Betts et al. 2009; Uppala 2009; Dee et al. 2011). ERA-Interim covers the period from 1 January 1979 onward, and continues to be extended forward in near-real time. Berrisford et al. (2009) and Dee et al. (2011) provide a detailed description of the ERA-Interim product archive. Information about the current status of ERA-Interim production, availability of data online, and near-real-time updates of various climate indicators derived from ERA-Interim data, can be found at http://www.ecmwf.int/research/era.

There is a growing motivation for downscaling of the simulations and projections provided by global models using regional climate model (Menéndez et al. 2010a, b; Marengo et al. 2011b; Carril et al. 2012; Solman et al. 2013). Previous experiences of downscaling in Central and South America have been performed using various regional models (Eta, MM5, RegCM3, HadRM3, RCA). These regional models were forced using the HadAM3P, HadCM3P or ECHAM5 global models as boundary conditions, for high and low emission scenarios to generate climate projections out to the year 2100, for studies on change in climate and extremes (Marengo et al. 2009a, b, 2011b; Urrutia and Vuille 2009; Vicuña et al. 2011; Soares and Marengo 2009; Garreaud and Falvey 2009; Cabré et al. 2010; Campbell et al. 2011; Karmalkar et al. 2011; Sorensson et al. 2010; Chou et al. 2011) in South America.

A coordinated experiment using four regional models forced with the ERA–40 reanalysis for 1991–2000 (Carril et al. 2012) was done as part of the CLARIS Project, the predecessor of CLARIS-LPB. In other regions of the world, dynamical downscaling has been available for Europe from the PRUDENCE (prudence.dmi.dk) and ENSEMBLES (ensembles-eu.metoffice.com) projects (Jacob et al. 2007; Boberg et al. 2010); over North America, from the NARCCAP Program www.narccap.ucar.edu (Mearns et al. 2009; Wehner 2012); and for Africa, from the CORDEX-Coordinated Regional Climate Downscaling Experiment (hwcrp-cordex.ipsl.jussieu.fr/) experiment (Kim et al. 2012).

For CLARIS-LPB, in order to reduce the spread in the multi-model ensemble, a coordinated approach in terms of model domain, resolution, and boundary conditions for all

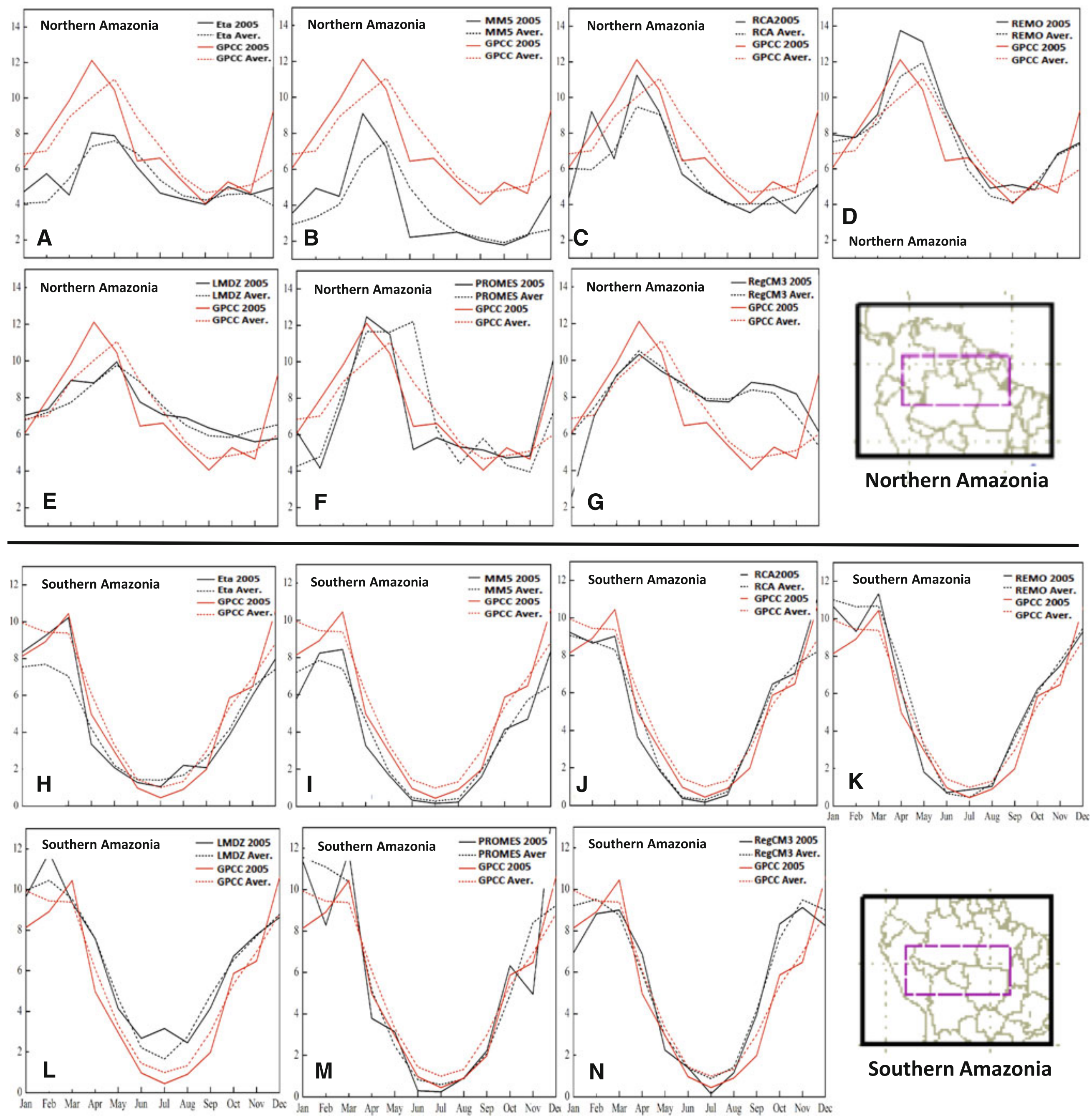

Fig. 1 Mean annual cycle of observed (GPCC) and simulated rainfall from each regional model of rainfall for northern Amazonia (**a**–**i**) and southern Amazonia (**k**–**n**). *Broken lines* represent observed climatology 1990–1998 in *red*, and simulated climatology 1990–2008 in *black*). *Full lines* represent rainfall during 2005(observed GPCC, *red*; and simulated by regional models in *black*)

the simulations was established (Solman et al. 2013). All models were run with horizontal resolution of about 50 km over South America but with slightly varying vertical resolution. Solman et al. (2013) and Carril et al. (2012) provide details of the regional models used in the experiment, and Table 1 shows details of the land surface parameterizations from each model. This information is relevant to the discussion of variations of the energy balance related to rainfall and temperature anomalies during the meteorological drought in the Amazon region. The regional models used the ERA-Interim reanalysis as boundary conditions.

In this present study, simulations from the seven regional models for 1990–2008 were evaluated against GPCC rainfall, CRU temperature observations, and 1,000 hPa low-level circulation, and evaporation,

Table 1 Land surface scheme information from the regional models used in CLARIS LPB

Model	Institution	Land surface scheme	General references
REMO	Max Planck Instute for Meteorology, Germany	LSS: based on the physical parameterizations of the ECHAM4 model. Improved surface runoff scheme, inland glaciers and vegetation phenology. The interface between land surface and atmosphere is a layer of 'infinitesimal' thickness, which is in contact with the atmosphere. The coupling between land and atmosphere is semi-implicit. For vertical surface fluxes, a subgrid scale tile approach for land, water and sea ice surfaces was implemented. Over the land fraction, the big-leaf approach is still applied"	Jacob et al. (2007)
LMDZ	IPSL, Institute Pierre and Simon Laplace, France	ORCHIDEE: It is a complex surface scheme, but only the surface-vegetation-atmosphere transfer module is used in LMDZ and PROMES. The surface hydrology is simulated with two layers covering the first two metres. There are 12 plant function types in addition to bare soil. Leaf area index is prescribed with present-day climatology in all simulations	Krinner et al. (2005)
PROMES	Facultad de Ciencias del Medio Ambiente-Universidad de Castilla-La Mancha, Spain	ORCHIDEE: A dynamic model of the terrestrial biosphere composed by two existing modules, the surface-vegetation atmosphere transfer (SVAT) model SECHIBA and the dynamical vegetation model LPJ, together with one additional module developed recently, the carbon cycle model Saclay Toulouse Orsay model for the analysis of terrestrial ecosystems, which describes photosynthesis, carbon cycle and phenology	Sanchez et al. (2007), Domínguez et al. (2010), Sitch et al. (2003), Krinner et al. (2005)
RegCM3	ICTP and USP, University of São Paulo	BATS (biosphere–atmosphere transfer scheme): It considers one vegetation layer, with 20 vegetation types, and three soil layers. The rooting ratios and upper and total soil depths are functions of land cover type and each vegetation type has its corresponding soil properties. Rooting zones have depths ranging from 1.0 to 2.0 m depending on the vegetation type, while the total soil depth is always 3 m. Modified RegCM3-BATS scheme only for the tropical broadleaf forest. The depth of root zone layer changed from 1.5 to 3.0 m; total soil depth was modified from 3.0 m to 4.5 m; saturated hydraulic conductivity at the bottom of the subsoil layer was defined as 40 % of its default value, reducing therefore the soil water drainage	Pal et al. (2007), da Rocha et al. (2009, 202), Dickinson et al. (1993).
RCA3	Rossby Centre, SMHI, Sweden	LSS: The land-surface scheme belongs to the second generation of LSSs which means that it has fairly advanced treatments of many physical land-surface processes but it does not account for carbon dioxide (CO2) effects on canopy conductance in evapotranspiration calculations. The soil is divided into five layers with respect to temperature with a no-flux boundary condition at 3.0 m depth. The thicknesses of the layers increase from 1.0 cm for the top-most layer to 1.89 m for the deepest layer	
MM5	Centro de Investigaciones del Mar y la Atmósfera (CIMA), Argentina	NOAH: The land-surface model is capable of predicting soil moisture and temperature in four layers with thicknesses of 10, 30, 60 and 100 cm, as well as canopy moisture and water-equivalent snow depth. The land surface model makes use of vegetation and soil type in handling evapotranspiration, and takes into account variations in soil conductivity and the gravitational flux of moisture	Chen and Dudhia (2001), Solman and Pessacg (2012)
Eta	Instituto Nacional de Pesquisas Espaciais INPE, Brazil	NOAH: 4 soil layers for temperature and humidity with 10, 30, 60, and 100 cm depth; 12 vegetation types using map created by Sestini et al. (2002)	Ek et al. (2003), Chou et al. (2011)

precipitation and air temperature from the ERA-Interim reanalysis data set. Precipitation anomalies from GPCC and from each regional model are discussed relative to the observed low water anomalies detected in some rivers of Amazonia, already presented by Marengo et al. (2008a, 2011a) for the hydrological side of the drought of 2005.

As a matter of comparison, negative river level anomalies of the Rio Negro in Manaus and Amazonas at Óbidos

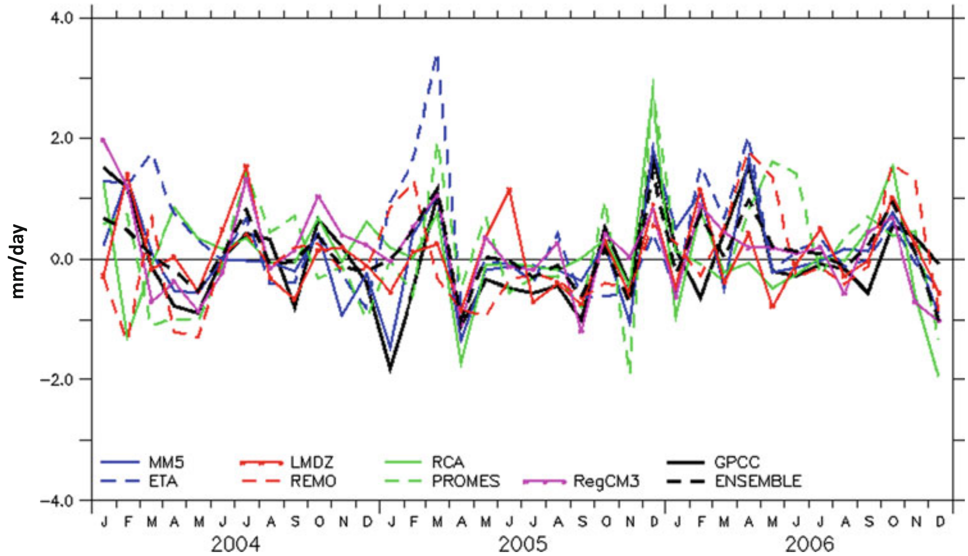

Fig. 2 Rainfall anomaly evolution from January 2004 to December 2006 in southern Amazonia. Thick *black line* represents observed mean and *broken black line* represents the model ensemble. Each individual model is represented by *colored lines*. Units are mm/day. Region is shown in Fig. 1

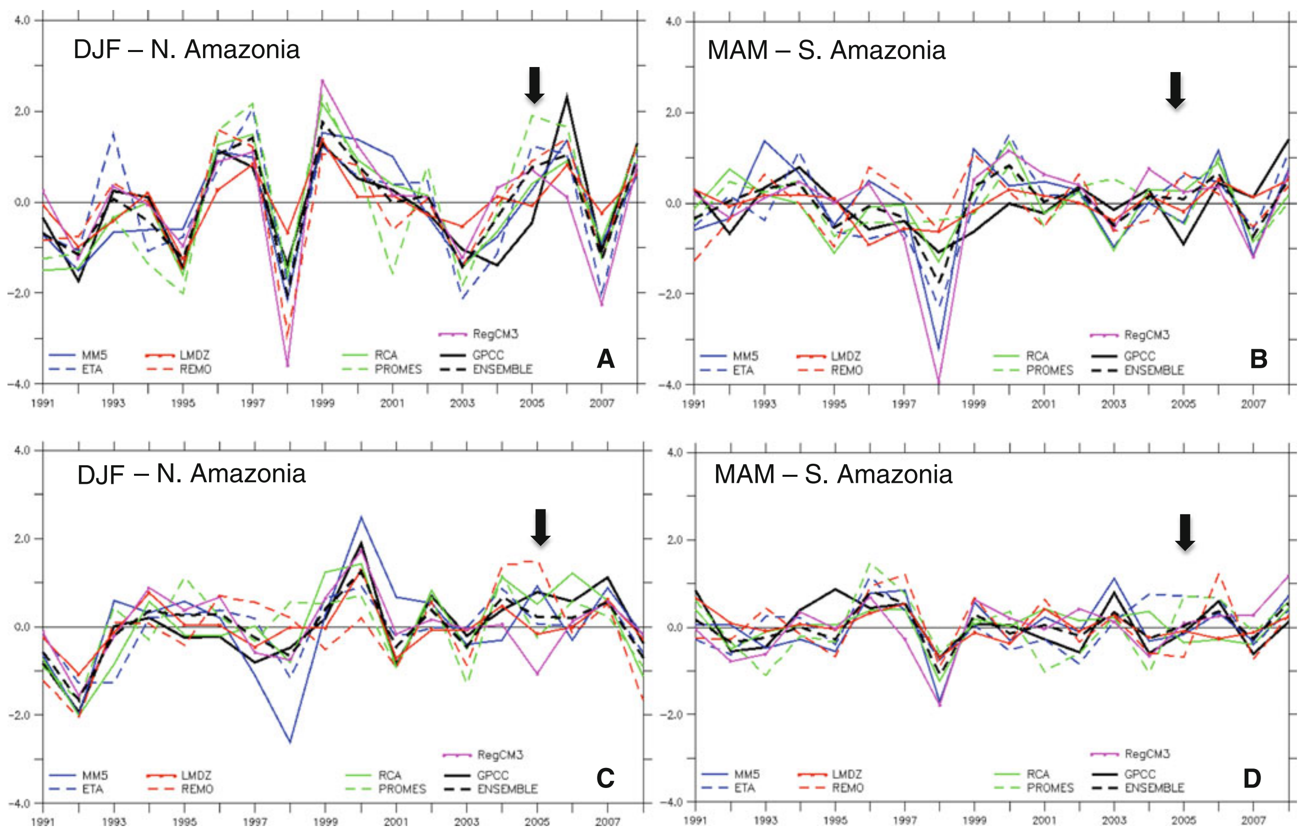

Fig. 3 Interannual variability of observed (GPCC) and simulated rainfall anomalies in northern (**a**, **c**) and southern Amazonia (**b**, **d**), during austral summer DJF and fall MAM during 1991–2008. *Thick black line* represents observed mean and *broken black line* represents the model ensemble. Each individual model is represented by *colored lines*. Units are mm/day. Region is shown in Fig. 1

in 1998 were comparable in magnitude to levels in 2005, but during the 1998 El Nino year, rainfall was extremely low, while it was not exceptionally low in 2005 (Marengo et al. 2011a). We have to remember that changes in river levels are not proportional to the magnitude of the rainfall anomalies, and in one or more sections of the Amazon rivers, short or long-term changes in flow cannot be explained in terms of rainfall variability alone (Sternberg

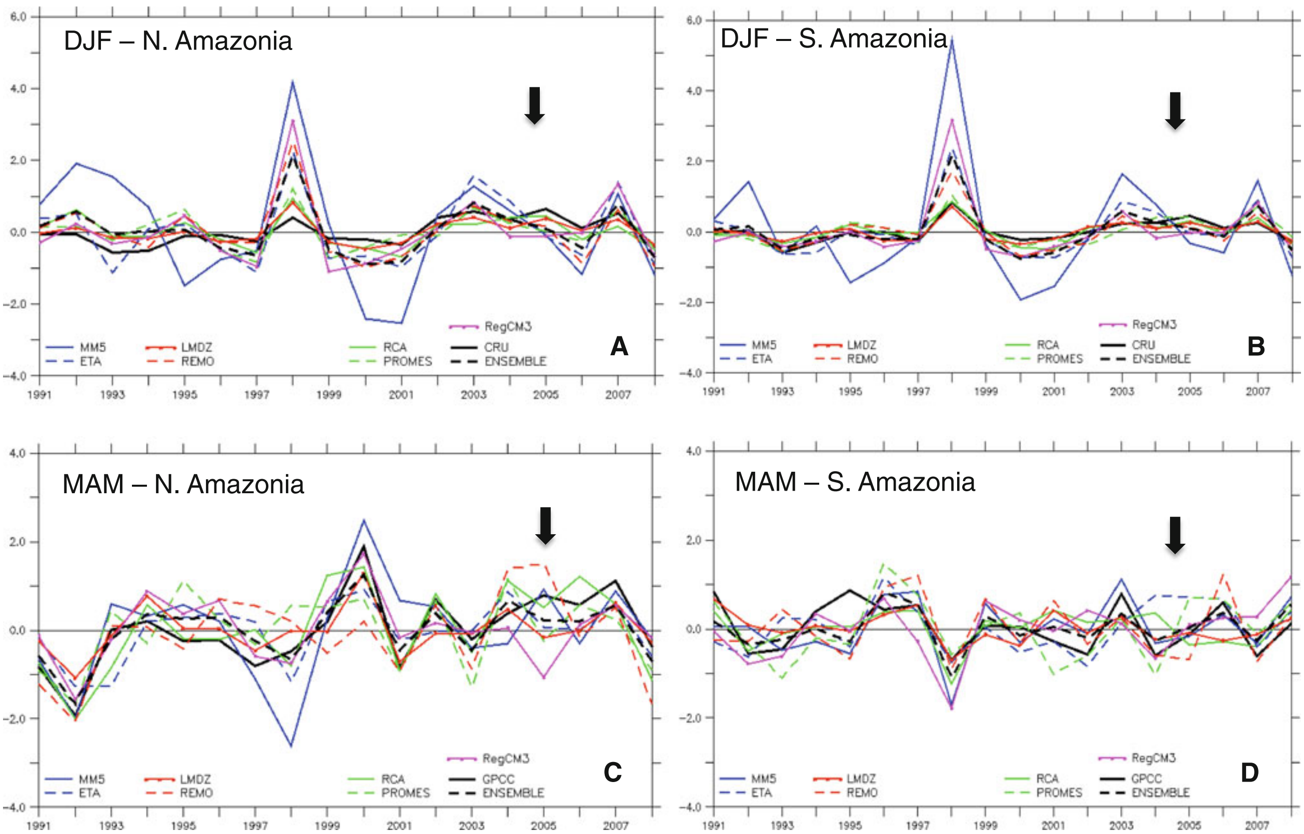

Fig. 4 Interannual variability of observed (CRU) and simulated air temperature anomalies in northern (**a**, **c**) and southern Amazonia (**b**, **d**), during austral summer DJF and fall MAM during 1991–2008. *Thick black line* represents observed mean and *broken black line* represents the model ensemble. Each individual model is represented by *colored lines*. Units are °C. Regions are shown in Fig. 1

1987; Marengo et al. 2011a; Tomasella et al. 2011, 2013). This was the situation for 2005, while for 1998, drought was more related to very strong negative rainfall anomalies in the Amazon region.

The seven regional climate models used in this coordinated dynamical downscaling experiment (shown in Table 1) are the following: MM5 (CIMA-Argentina), RCA (Rossby Center/SMHI-Sweden), REMO (MPI-Germany), PROMES (UCLM-Spain), LMDZ (LMD-France), RegCM3 (USP-Brazil) and Eta (INPE-Brazil). A description of each regional model and the diverse aspects of their simulated climatology are given in Menéndez et al. (2010a), Chou et al. (2011), Marengo et al. (2011b), Carril et al. (2012), da Rocha et al. (2012) and Solman et al. (2013).

Uncertainty in rainfall simulations is assessed by comparing simulations against GPCC rainfall observations and in terms of the scatter among the seven regional models and the ERA-Interim reanalyses, considering the biases and degree of agreement or disagreement, as a measure of the confidence in the rainfall simulation during the austral summer and fall of 2005 in Amazonia.

3 Observed and simulated patterns of the drought of 2005 in Amazonia

3.1 Annual cycle of rainfall: 2005 versus climatology

Figure 1 shows the annual cycle of rainfall in northern and southern Amazonia, comparing observations (GPCC) and the ensemble of regional model simulations within two periods, the long-term climatology (1990–2008) and the drought year (2005). The observed dry season in June–August 2005 was more intense than normal in southwestern Amazonia, with rainfall that sometimes decreased to 25 % of the normal value (Marengo et al. 2008).

For northern Amazonia (Fig. 1a–g), comparing the observed and simulated climatology, the Eta model shows a dry bias from December to May, the peak rainfall season, varying between 2 and 4 mm/day while the MM5 and RCA models show this bias all year long (between 2 and 4 mm/day and 1–2 mm/day, respectively). While the REMO model shows a wet bias during March–May (2 mm/day), but closer to observations during the year, the LMDZ model shows a dry bias (2 mm/day) during the summer and

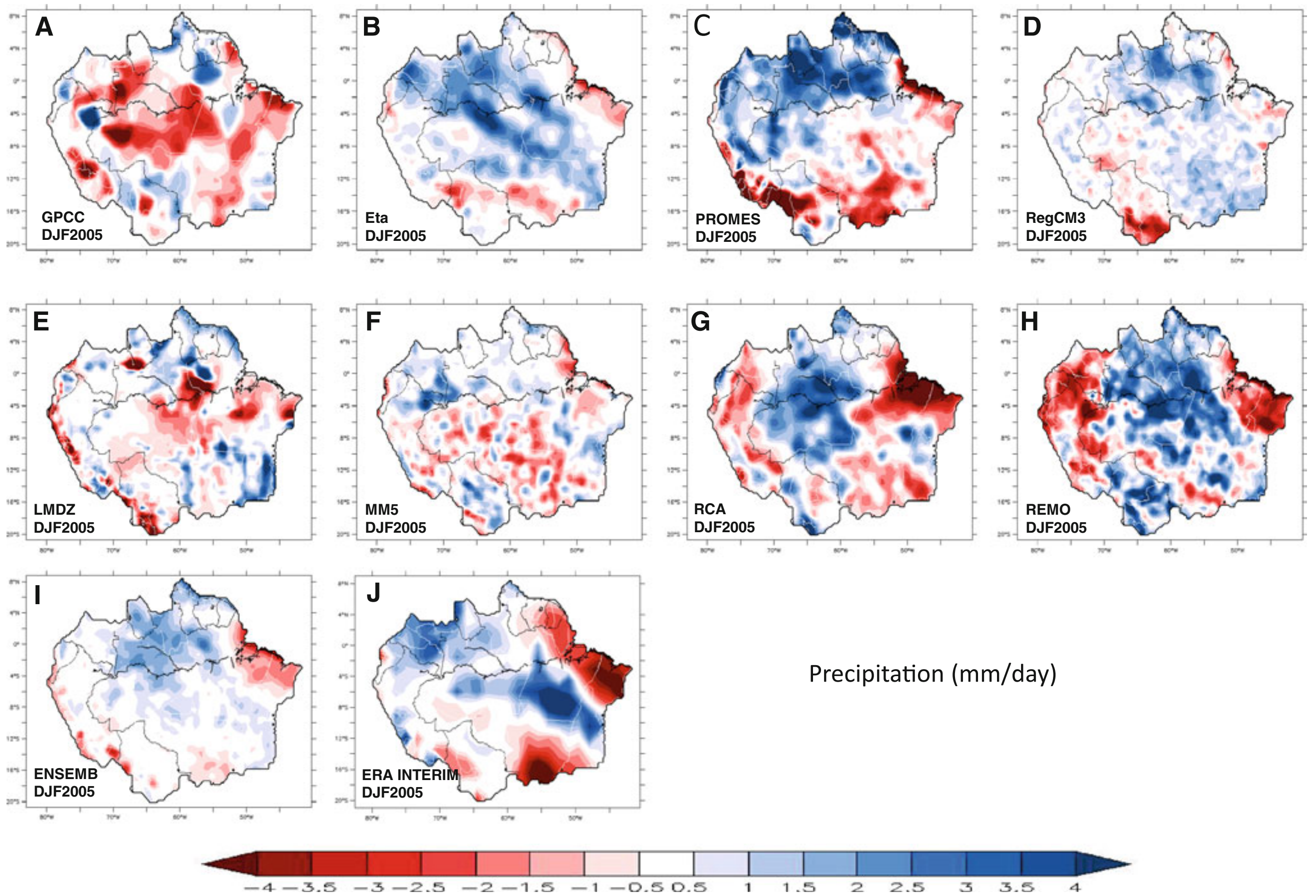

Fig. 5 Geographical distribution of observed and simulated rainfall anomalies for Amazonia for DJF 2004–2005. **a** From GPCC. Simulations for each individual model are organized as follows: Eta (**b**), PROMES (**c**), RegCM3 (**d**), LMDZ (**e**), MM5 (**f**), RCA (**g**), REMO (**h**), ensemble (**i**) and Era reanalyses (**j**). *Color scale* is shown in the lower part of the figure. Units are mm/day

autumn, and wet bias (2.0 mm/day) during winter and spring. The PROMES model shows a dry bias (1–2 mm/day) during January and February and a wet bias from April to June, while the RegCM3 shows two rainfall peaks, one in March–April and another in September–October; observations from GPCC exhibit only one peak during the austral fall. The latter model underestimated observed rainfall by about 2 mm/day during the peak season and overestimated rainfall by about 3.5 mm/day during austral spring.

In southern Amazonia (Fig. 1h–n), the observed annual cycle of precipitation is well simulated by every model. Shortcomings shared by some models are an underestimation of between 1 and 2 mm/day during the peak rainfall season during January–February for the Eta, MM5, RegCM3 and RCA, and overestimation of the same order of magnitude in the other models. In the austral winter, the Eta and LMDZ models show slight overestimation of rainfall (0.5–1 mm/day) while the rest of models show underestimation of the same order of magnitude.

In the austral spring, during which the onset of the rainy season occurs, the Eta and MM5 exhibit rainfall underestimation while the rest of models show overestimation. As pointed out by Solman et al. (2013), rainfall is triggered by convection, and the moisture transport from Amazonia into this region by the South America low level jet south of the Andes (SALLJ) helps in maintaining moisture convergence and rainfall (Marengo et al. 2004). The biases of different signs among regional models suggest that problems in the simulation of the SALLJ by the different models may result in inadequate activation of convective processes affecting the simulation of the onset of the rainy season by the individual models.

In regard to the evolution of rainfall anomalies in southern Amazonia during 2004–2005, Fig. 2 shows that the observed negative rainfall anomalies in southern Amazonia started in December 2004 and then become more intense in January and February 2005. Although rainfall was above normal in March it returned to below-normal values in April and remained there until September. The figure shows that for November 2004 through January 2005, four out of seven models showed negative rainfall anomalies, consistent with observations from GPCC, and while in January 2005 the MM5 rainfall anomalies reached

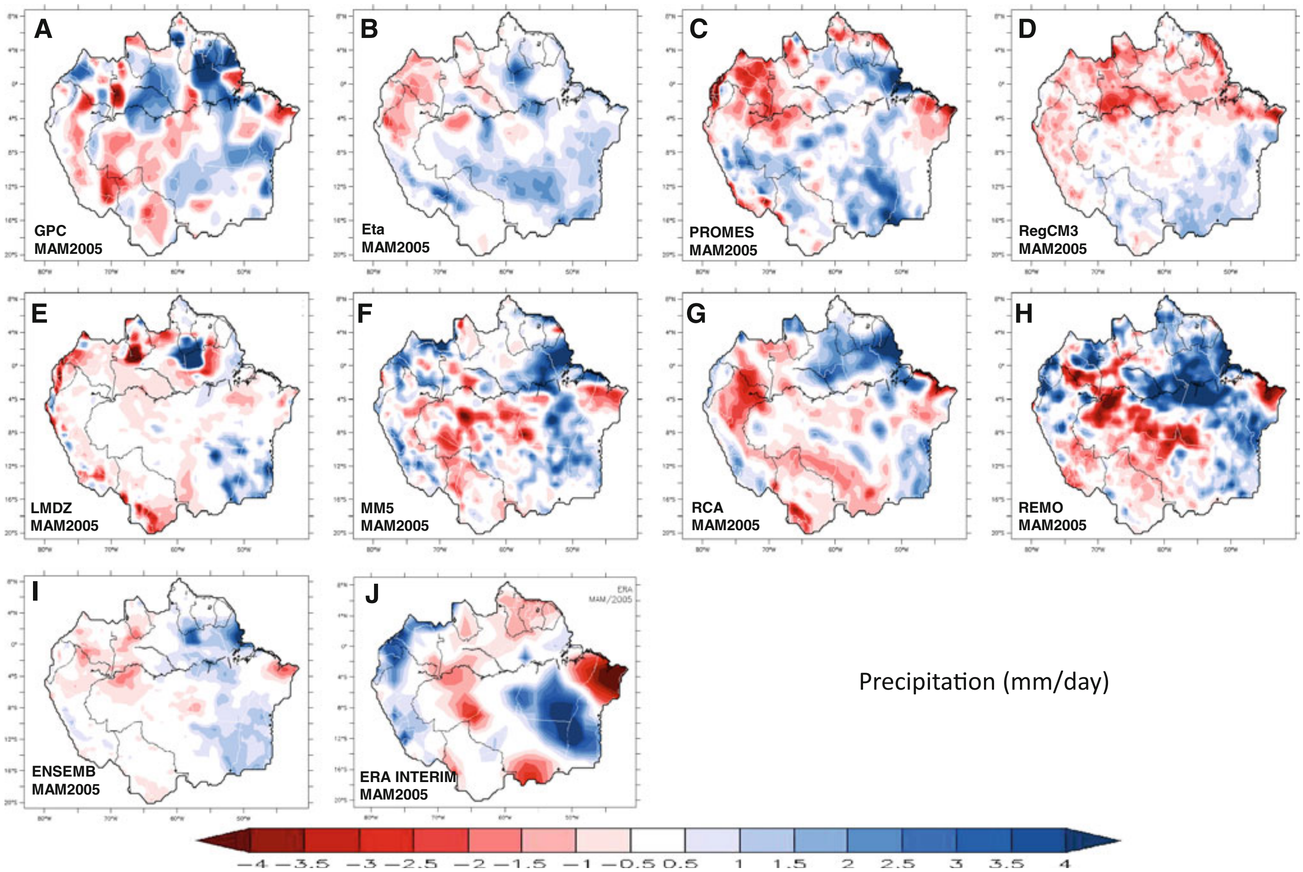

Fig. 6 Same as in Fig. 5, but for MAM 2005

about −1.8 mm/day (observations show −2 mm/day), the Eta and REMO showed about +0.8 mm/day. In February 2005 all but two models show positive rainfall anomalies varying from +0.1 to +2.1 mm/day, while observations show almost −1 mm/day. In March and April 2005, the models capture well the shift from wet conditions in southern Amazonia in March to dry conditions in April, with all models showing this behavior, and from the rest of austral fall to the beginning of spring the average of the models shows negative rainfall anomalies, consistent with the observations.

In general, we can say that the observed negative rainfall anomalies beginning in early summer of 2005 (November 2004 to February 2005) are simulated by some models, as are the observed rainfall increases in March 2005 and then the negative rainfall anomalies continuing until September 2005. Figures 1 and 2 show that there are similarities in the observed and simulated annual cycle and monthly rainfall evolution, and that the differences among models may be related to their model structure and different physical parameterizations. In agreement with Carril et al. (2012), the skill of the models in reproducing mean climate conditions over northern and southern Amazonia is weak and the uncertainty is high. Figure 2 suggests that this is not necessarily true for the interannual variability. Nevertheless, this fact highlights that a good agreement with the observations over a given region in particular years does not necessarily imply a correct representation of all the involved physical processes.

3.2 Interannual variability in Amazonia

The interannual variability for rainfall and temperature is shown in Figs. 3 and 4 for DJF (a, b), and MAM (c, d), which are the peak rainy seasons in southern and northern Amazonia, respectively. Figure 3a–d shows rainfall anomalies with respect to the 1990–2008 climatology, for GPCC observations, and for each of the seven regional models and for the multi-model ensemble mean. Observations show negative rainfall departures during 1992, 1995, 1998, 2004 and 2007 in Northern Amazonia, which are years of moderate or strong El Niño events (www.cptec.inpe.br). In this region the signal from strong El Niño events is more intense in austral summer and fall (See Marengo et al. 2011a and references quoted therein).

Most of the simulations tend to agree with the observed negative departures in 1992, 1995, 1998 and 2007 mainly in austral summer (Fig. 3a), while in fall the best consistency

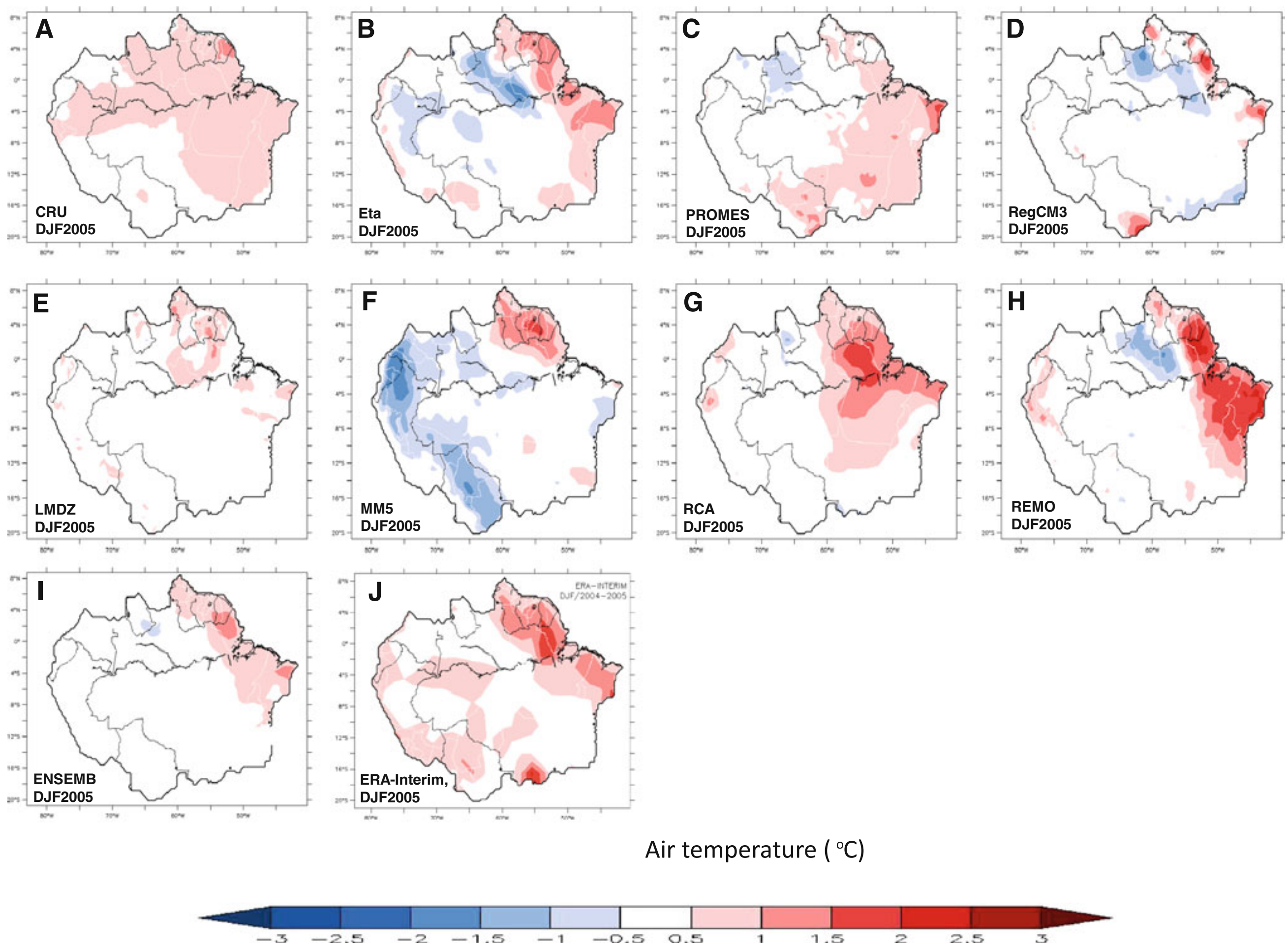

Fig. 7 Geographical distribution of observed and simulated air temperature anomalies for Amazonia for DJF 2004–2005. **a** From CRU. Simulations for each individual model are organized as follows: Eta (**b**), PROMES (**c**), RegCM3 (**d**), LMDZ (**e**), MM5 (**f**), RCA (**g**), REMO (**h**), ensemble (**i**) and Era-Interim reanalyses (**j**). *Color scale* is show in the lower part of the figure. Units are in °C

between observed and simulated negative rainfall anomalies occurs in 1992 and 1998 (Fig. 3c). In both seasons, the scatter among models reaches about ±2 mm/day in DJF, with the largest bias for the RegCM3 and the lowest for the LMDZ. Eta and PROMES show large positive anomalies that are reflected in the positive anomaly for the ensemble mean (Fig. 3a). Most of the regional models produced wetter than normal rainy season in 1996, 1999, 2000 and 2006, which are consistent with the GPCC. For 2005, observed rainfall in northern Amazonia was above normal (approximately 1 mm/day) during MAM and most of the models show positive rainfall anomalies in MAM and DJF (Figs. 3a–c). In MAM the LMDZ, PROMES and RegCM3 simulate slightly negative departures in comparison to the other four models, which show positive anomalies varying from +0.2 (Eta) to +2.0 (REMO) mm/day. This induces an ensemble mean with positive value similar to the GPCP (Fig. 3c).

In southern Amazonia (Fig. 3b, d), observations and simulations from all regional models show large negative departures in 1998 (an El Niño year, also dry in northern Amazonia) during the DJF peak rainfall season. Observed wetter rainy seasons in this region in 1994, and 2006 are depicted by at least four models, while in 2000 all models exhibit large positive departures in DJF that are not shown in the GPCC data (Fig. 3b). For the observed negative anomaly in DJF 2005, three of the models (PROMES, LMDZ and MM5) show negative rainfall departures, while four (REMO, RegCM3, Eta and RCA) other models shows slightly positive rainfall anomalies. For this period, the observed anomaly is −0.4 mm/day and in the ensemble mean it is almost zero (Fig. 3c).

Figure 4a–d shows the observed and simulated interannual variability of air temperature from CRU in northern and southern Amazonia during DJF and MAM. In comparison to the rainfall variability, scatter among models is low and there is close agreement between the values observed and the ensemble means. One model (MM5) tends to overestimate the interannual variability of temperature and all of the models simulated large air

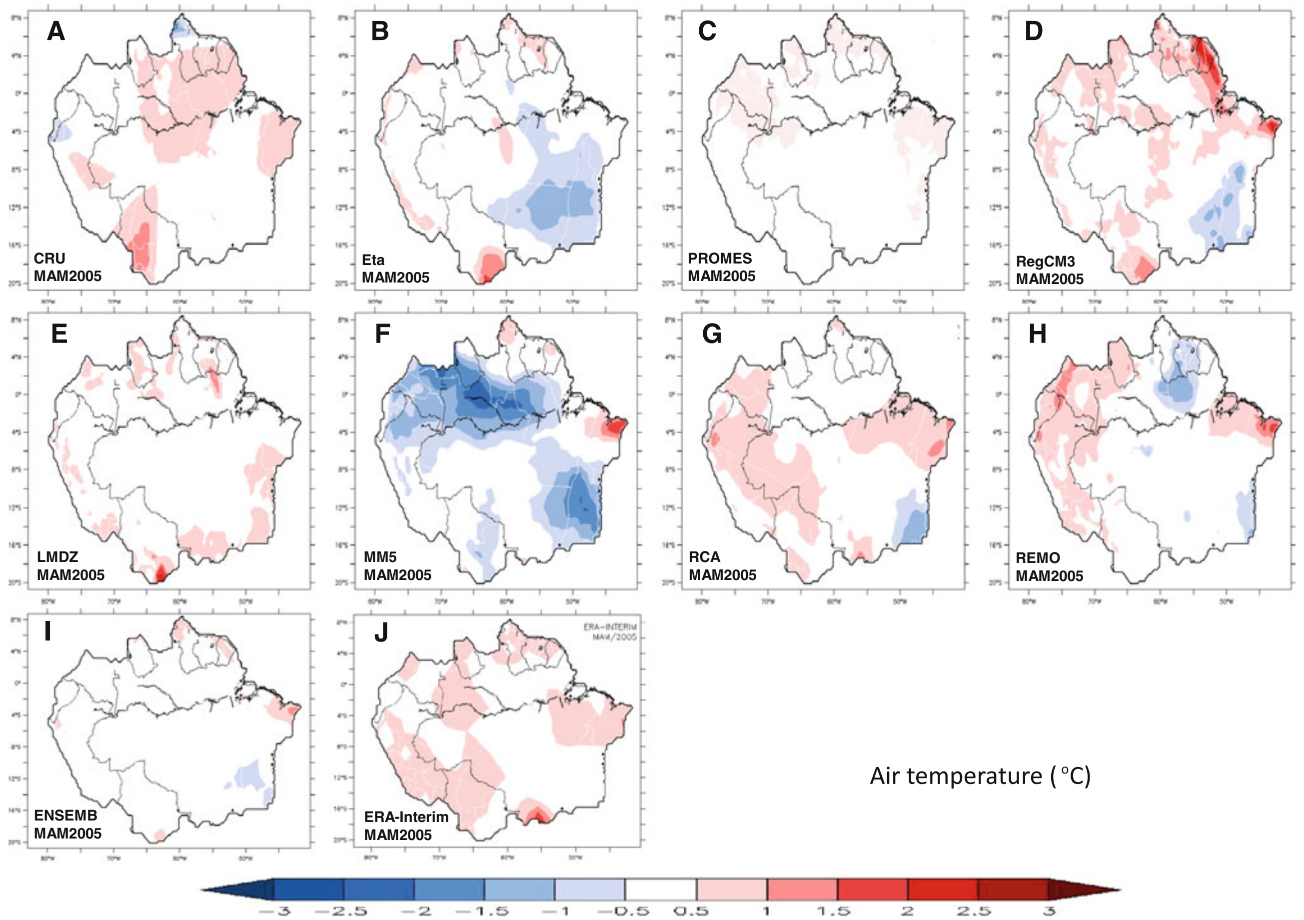

Fig. 8 Same as in Fig. 7, but for MAM 2005

temperature increases during the drought of 1998. However, in 2005, six out of seven models show warming in both regions of Amazonia.

The increase of sensible heating and smaller availability of moisture due both decreases of evaporation and anomalous low level circulation (as discussed below) can help to explain the warming in 2005. Most of the models were able to capture these processes, as is reflected in the positive temperature anomalies during 2005; however 1998 was more interesting in that all models show warming, consistent with observations—a model consensus that is not found in 2005.

In summary, the strong El Niño signal appears at the peak of the rainy season in both northern and southern Amazonia in 1998, where rainfall (temperature) anomalies are well below (above) normal. In 2005 three of the seven models depict rainfall below normal in southern Amazonia during DJF. Some biases are evident and not all models show the same bias. However, considering the two areas and seasons, in general the ensemble mean represents the observed signal of the anomaly better than individual models. The most important condition affecting the climate and it variability in northern Amazonia is the SST in the equatorial Pacific and in the tropical Atlantic. However, in the inland regions, including southern Amazonia, land-surface conditions (such as soil wetness, leaf area index, stomatal resistance, etc.) and anomalies of circulation (discussed in the following sections) may also play an important role in seasonal and year-to-year climate variability. These features may be even more important than the SST forcing, as occurred in 2005. An exception would be when a strong El Niño event occurs, as in 1998, where both sections of Amazonia experienced drier conditions.

3.3 Rainfall and temperature anomaly distribution across the basin

Figures 5a–j and 6a–j show rainfall anomalies observed and simulated from each regional model, and the model ensemble for the austral summer (Fig. 5) and fall (Fig. 6) for 2005, respectively, relative to the long-term mean period 1990–2008. The maps with GPCC observations indicate that the basins in the southwestern and extreme eastern Amazon region were the most affected by the drought during 2005, especially during the peak of the

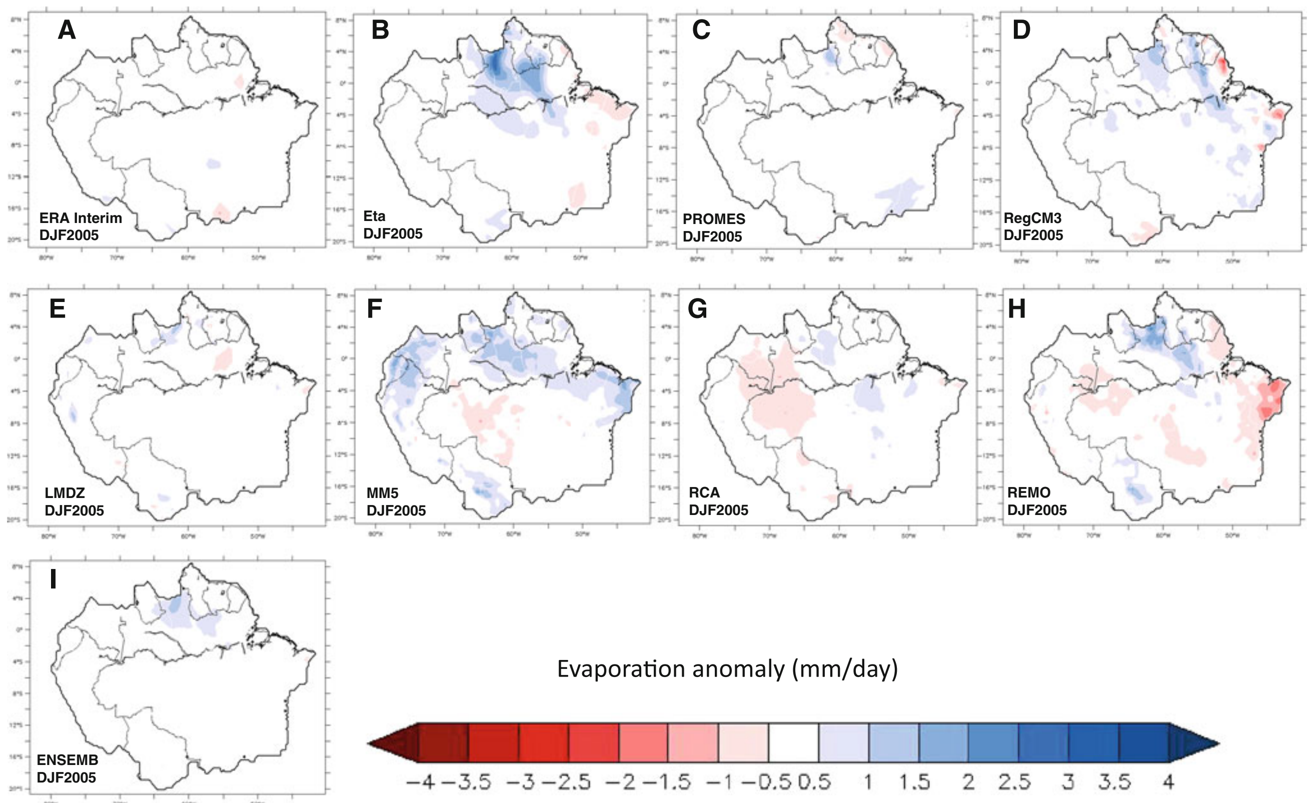

Fig. 9 Geographical distribution of observed and simulated evaporation anomalies for Amazonia for DJF 2004–2005. **a** From Era-Interim Reanalyses. Simulations for each individual model are organized as follows: Eta (**b**), PROMES (**c**), RegCM3 (**d**), LMDZ (**e**), MM5 (**f**), RCA (**g**), REMO (**h**), ensemble (**i**). *Color scale* is show in the lower part of the figure. Units are mm/day

rainy season in early austral summer (between −1 and −3 mm/day), while in fall, rainfall was above normal in northern Amazonia (above +3 mm/day).

During the DJF season the Eta model shows rainfall anomalies above +3 mm/day in central and eastern Amazonia, while the observations show deficits of about −3 mm/day. Similar overestimation of rainfall is noticed in central Amazonia for the RCA and REMO models. On the other hand, the RegCM3 shows negative rainfall anomalies in southwestern Amazonia (between −1 mm/day and −3 mm/day), comparable in magnitude with the observed anomalies in that region during DJF. During MAM almost every model reproduces the positive rainfall anomalies over the northern part of the Amazon basin, but the RegCM3 shows negative rainfall anomalies in western Amazonia, consistent with observations; however the simulated negative rainfall anomalies in central and northern Amazonia (1–2 mm/day) are in contradiction to the observed positive rainfall anomalies (1–3 mm/day) in the same region. Several biases are evident in this figure and it is interesting to note that not all the models share the same biases.

The ensemble mean during DJF shows some agreement with the observed negative rainfall anomalies in extreme eastern Amazonia as well as the positive rainfall anomalies in northern and southeastern Amazonia during MAM. However, the ensemble does not simulate the extension of the observed negative rainfall anomalies in southwest Amazonia during DJF. The Era-Interim rainfall maps show similarities to most of the simulated maps, and to the ensemble, as expected, since these are boundary conditions for the regional models. However, the Era-Interim does not depict the observed negative rainfall anomalies in central and southwestern Amazonia during DJF and MAM, and the two data sets agree in depicting the observed negative rainfall anomalies over eastern and extreme southeastern Amazonia.

The observed temperature anomaly maps (Figs. 7a–j, 8a–j) show warming over all of Amazonia (between 1 and 1.5 °C) during DJF and concentrated mostly in southern and eastern Amazonia (up to 1.5 °C) during MAM, consistent with the negative rainfall anomalies. The PROMES, RCA, RegCM3, Eta and REMO show warming in various parts of Amazonia, particularly in the eastern section, and the Eta model also shows cooling in northern and southern Amazonia during DJF and MAM, respectively (consistent with the positive rainfall anomalies in Figs. 3, 4). The MM5 also shows cooling in most of Amazonia. The ensemble mean depicts the observed warming in eastern

Amazonia during DJF. The interannual variability of temperature indicates that northern and southern Amazonia experienced an observed warming of about 0.5 °C during both seasons, of which five models show similar warming in 2005 of about 0.3 °C and the variability among models is very low. In comparison, during the drought of 1998, the observed warming varied between 0.8 and 1 °C in southern Amazonia, and about 0.5 °C in northern Amazonia. All models exhibit warming, which ranges from 0.6 to 5.5 °C in both regions and seasons. The warmest model in 1998 is the MM5, while the same model shows cooling in both regions and seasons in 2005.

The discussions about the interannual variability and the geographical distribution of temperature anomalies and rainfall anomalies during summer and fall 2005 suggests some interesting aspects which should be analyzed in the context of the energy balance and land surface processes. Table 1 describes the land surface schemes from each of the regional models, and while some of them use modified versions of the same scheme, there are some differences in the vegetation and soil types, as well as in the number of soil layers, surface hydrology, and schemes for coupling between land and atmosphere, that may in the end have an effect on the depiction of sensible and latent heat, and subsequently, on the near surface temperature fields, by means of changes in evaporation. Evaporation shows strong correlations with precipitation and temperature, since they are measure of the surface-atmosphere coupling (Jung et al. 2010). In the next section we will discuss the distribution of evaporation.

3.4 Evaporation

For the summer of 2005 in southern Amazonia, evaporation estimates calculated using the Penman–Monteith equation for a free water surface (Tomasella et al. 2011) reached 4.9 mm/day, leading to drastic lake depletion during the 2005 drought. In the same region, evaporation from the Era-Interim reanalyses (not shown) for the summer season varies between 4 and 5 mm/day in the same region. While PROMES and RCA show the lowest evaporation values—between 2.5 and 3.5 mm/days, respectively—the highest values are from the Eta, RegCM3 and REMO, reaching up to 6 mm/day. The LMDZ and MM5 evaporation show values similar to those from the Era-Interim derived evaporation. However, uncertainties still remain in the representation of evaporation from the ERA-Interim reanalyses. Table 1 shows the different land surface schemes from each regional model, that at the end affect the simulation of evaporation among models.

The evaporation anomaly maps (Figs. 9a–i, 10a–i), show anomalies smaller than those of precipitation, which

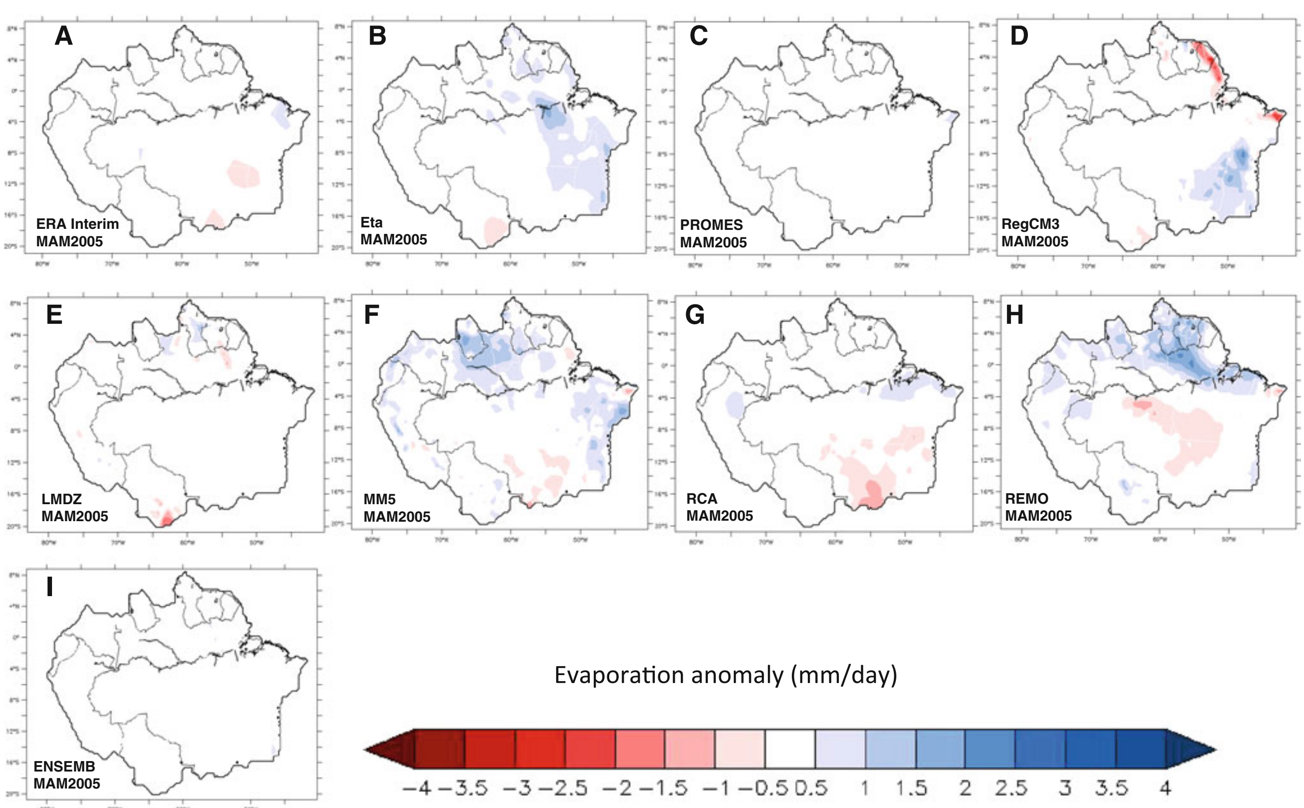

Fig. 10 Same as in Fig. 9, but for MAM 2005

means that changes in evaporation are less intense than changes in precipitation, relative to the mean climatology. The maps show negative anomalies over northern Amazonia in the MM5, Eta and RegCM3 and REMO for the DJF maps, while RCA and REMO show positive anomalies in western and southern Amazonia in DJF (Fig. 9). In MAM, Fig. 10 presents positive (negative) evaporation anomalies in southeastern Amazonia for the Eta, MM5 and RegCM3 (for the RCA). The evaporation anomaly fields are in some cases consistent with changes in precipitation and temperature, where rainfall reductions are consistent with increase in temperature, but not necessarily with evaporation decreases.

For DJF, in the RegCM3, MM5, REMO and Eta, the precipitation increases in northern Amazonia, consistent with temperature and evaporation reductions. However, in the RCA there are precipitation reductions and increases in temperature while evaporation shows barely any change. These different feedbacks could be related to the poor representation of processes linked to tree roots in the RCA, that do not seem to be capable of transporting water from deep soil levels to the plant, affecting evaporation rates. In the RegCM3, da Rocha, (personal communication) discussed whether the coupling between surface-atmosphere is better solved in northern and central Amazonia than in southern Amazonia, suggesting that the soil moisture anomalies could be better represented in northern Amazonia.

For instance, the interannual rainfall variability is more similar to the observations in PROMES than RCA, what can result from PROMES capturing the surface-atmosphere coupling for evaporation and latent heat as discussed by Jung et al. (2010), while RCA does not. However, it is still hard to say which models have the best skill in simulating rainfall, temperature and evaporation anomalies for depicting the anomalies in 2005. These suggest that the coupling of precipitation with surface processes does not seem to be consistent in all models. Part of the inability of several models to properly describe the evaporation in the Amazon forest are due to problems in representing the extraction of ground water by the root systems in the models (Harper et al. 2010).

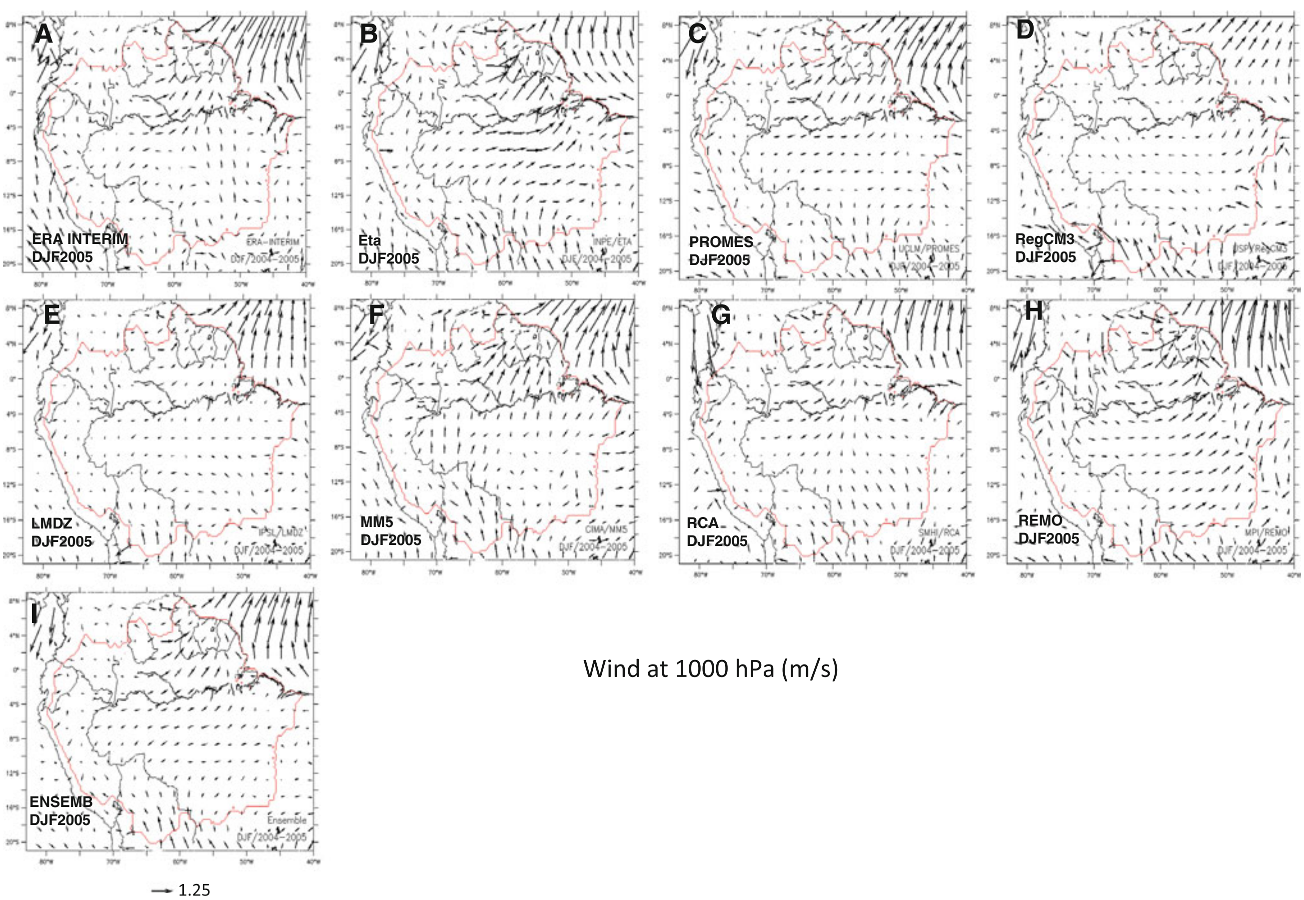

Fig. 11 Geographical distribution of observed and simulated rainfall low-level (10,000 hPa) circulation anomalies for Amazonia for DJF 2004–2005. **a** From Era-Interim reanalyses. Simulations for each individual model are organized as follows: Eta (**b**), PROMES (**c**), RegCM3 (**d**), LMDZ (**e**), MM5 (**f**), RCA (**g**), REMO (**h**), ensemble (**i**). *Arrow scale* is shown on the lower part of panels. Units are m/s

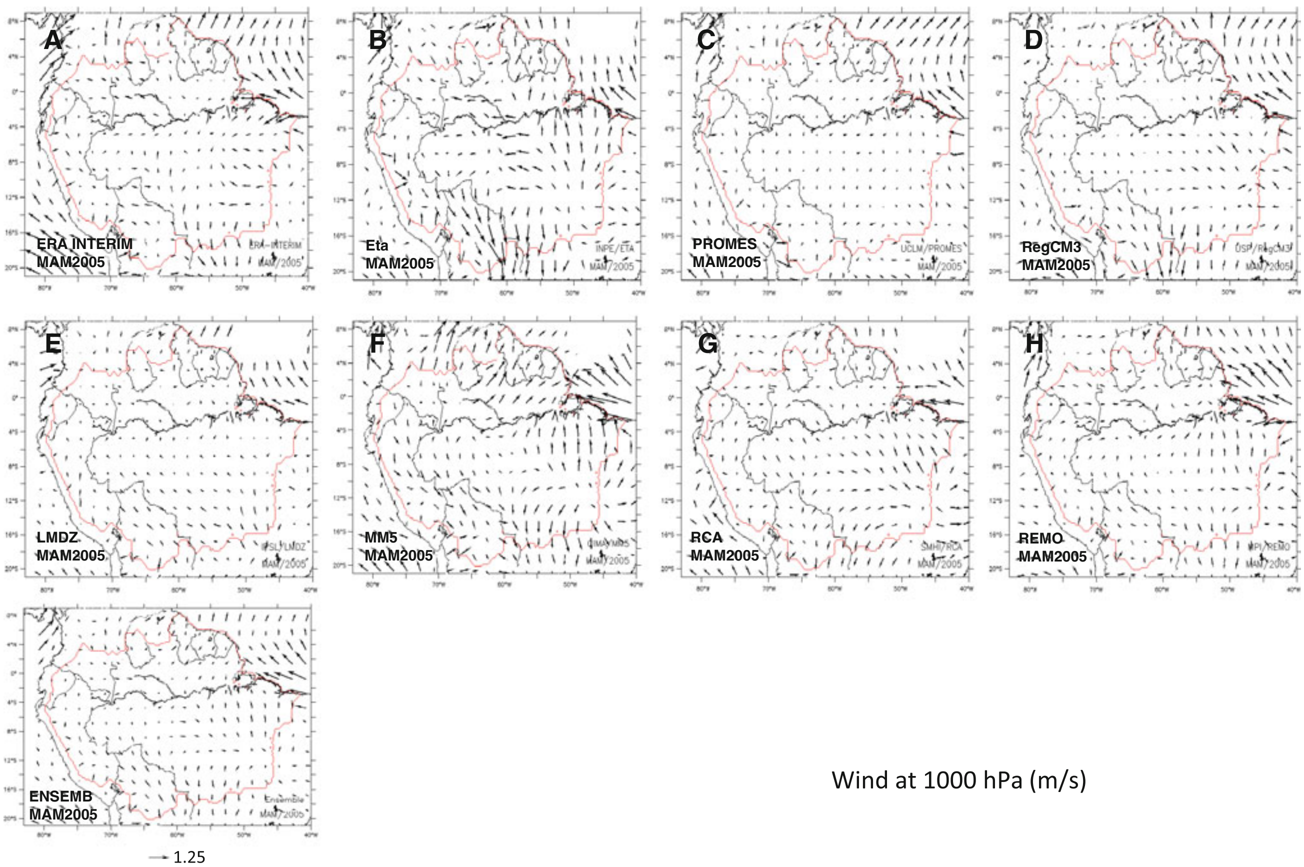

Fig. 12 Same as in Fig. 11, but for MAM 2005

3.5 Low level circulation

Observational studies of the drought of 2005 (Marengo et al. 2008a, b; Zeng et al. 2008) have shown that the northeast trades weakened during austral summer (DFJ), bringing less moisture into the Amazon region because the ITCZ was located anomalously northward, lying over a warmer tropical North Atlantic Ocean. This is observed in the ERA-Interim circulation maps in Fig. 11a–i, where the wind anomalies over tropical South America east of the Andes and over the tropical Atlantic suggest weaker northeast and stronger southeast trades over Amazonia and the tropical North Atlantic during DJF, respectively. The moisture transport during this season was assessed by Marengo et al. (2008a), and shows the reduced moisture transport from the tropical North Atlantic into the Amazon region during the DJF 2005 season, which is consistent with the negative rainfall anomalies observed in central and southern Amazonia.

The LMDZ, PROMES, RCA, RegCM3, MM5 and REMO models reproduce this circulation pattern, with some underestimation in the Amazon region and some overestimation over the tropical North Atlantic. The Eta model shows intensified westerly flow over eastern Amazonia, which converges with southeasterly flow in eastern Amazonia, indicating convergence and rainfall over eastern Amazonia. The anomaly circulations, which are inherited through the boundary conditions from ERA-Interim reanalyses are fairly well represented by almost every model. The individual models and the ensemble show southerly flow anomalies south of Amazonia, especially over Bolivia, suggesting a reduction of the moisture flux from Amazonia to the South, which in fact was true in DJF 2005, since no episodes of SALLJ were detected during that season (Marengo et al. 2008a). In MAM (Fig. 12a–i), the circulation from northern to southern Amazonia and to the La Plata basin is intensified, as shown in both observations and the ensemble.

4 Conclusions

In this study we analyze rainfall anomalies during the drought of Amazonia in 2005, simulated various regional models. The perception of drought in Amazonia was related to anomalously lower river levels in the region in fall and winter, and not so much to rainfall anomalies in previous rainy austral summer season, a hydrological drought. While precipitation anomalies in 2005 were not exceptionally low (as compared to 1998), the combination of low rainfall during austral summer in southwestern Amazonia and the subsequent low river levels in this region lead to extremely low levels of the Solimões and Madeira Rivers, characterizing the hydrological component of the drought of 2005. This study was directed at analyzing the simulation of the meteorological component of the drought of 2005 in southern Amazonia.

We used a combination of GPCC and CRU data sets and the ERA-Interim reanalyses, and the simulations from the seven regional models to assess the simulated meteorological patterns during the drought that affected southwestern Amazonia during the austral summer of 2005. The models reproduced in some degree the observed annual cycle of precipitation and the geographical distribution of negative rainfall anomalies during the summer months in 2005. On the evolution of rainfall during 2004–2006, some of the models simulated the negative rainfall departures during early summer of 2005 (November 2004 to February 2005) in southern Amazonia, and all models simulated the abundant rainfall in March 2005 and the drier conditions in April 2005, as well as the below-average rainfall conditions during May–September 2005.

The interannual variability of rainfall anomalies for both DJF and MAM seasons over northern and southern Amazonia showed a large spread among models, with some of them capable of reproducing the 2005 observed negative rainfall departures (four out of seven models in southern Amazonia during DJF). For interannual temperature anomaly variability almost all models showed a good agreement compared to observations. Moreover, six out of seven models showed warming in both regions of Amazonia in 2005. In comparison, over southern Amazonia all models simulated the observed negative rainfall and positive air temperature anomalies during the El Nino related drought in 1998.

For seasonal climate prediction, model skill in the southern part of the basin during the southern hemisphere summer peak was lower as compared to skill in northern-central Amazonia where precipitation peaks in the March–May (fall) season (Paegle and Mo 2002; Grimm 2010 and references cited therein). Southern Amazonia shows lower seasonal climate predictability because rainfall in that region seems to be more sensitive to regional and local influences (e.g. soil moisture and land surface processes) than to remote influences from the SST anomalies. This also could highlight deficiencies among the regional models, particularly in the land surface interactions. However, another possibility is in the fact that rainfall in southern Amazonia depends on SST anomalies in regions other than the tropical oceans, and therefore predictability is more complex because a small change in the relative magnitude of SST anomalies may change the sign of precipitation anomaly.

This is confirmed by the fact that the drought of 2005 was more connected to SST anomalies in the tropical North Atlantic, while rainfall anomalies during the droughts of 1998 and 2010 were more connected to the tropical Pacific. As explained by Mo and Berbery (2011), local controls may be important, but the fact that there is no single oceanic region that controls precipitation anomalies in the southwestern Amazonia makes predictability lower.

The spatial structure of the simulated rainfall and temperature anomalies in DJF and MAM 2005 showed biases that are different among models. While some models simulated the observed negative rainfall anomalies over parts of western and southern Amazonia during DJF, others simulated positive rainfall departures over central Amazonia. The simulated circulation patterns indicate a weaker northeasterly flow from the tropical North Atlantic into Amazonia, and the reduced flows from southern Amazonia to the La Plata basin in DJF, which are consistent with observations. At this time, the northeasterly flow was re-established in MAM as shown by models and observations, allowing it to bring atmospheric moisture into Amazonia and contributing to the rainfall in northern Amazonia, as observed in the positive rainfall anomaly during MAM.

In general, we can say that the regional models are able to capture in some degree the response to the forcing in the tropical Atlantic during the drought of 2005 in Amazonia. Moreover, extreme climatic conditions in response to anomalous low-level circulation features, are also well captured, since the boundary conditions come from reanalysis and the models are largely constrained by the information provided at the boundaries. So, though the reliability in simulating rainfall in southern Amazonia may be low, as shown in Solman et al. (2013), the reliability of simulating the anomalous condition of an extreme drought in response to an anomalous circulation feature may be high.

The performance of each individual model and the ensemble in depicting the climate anomalies leading to the drought situation in summer of 2005 can be summarized thus: Some models depict well enough the observed rainfall anomalies in summer and fall of 2005, as well as air temperature and evaporation features, while the low-level circulation anomalies, strongly controlled by the boundary conditions, are also well reproduced. The analysis of the 2005 drought suggests that when the forcing leading to extreme anomalous conditions is associated with both local and non-local mechanisms (soil moisture feedbacks and remote SST anomalies, respectively) the models are not fully capable of representing these feedbacks and hence, the associated anomalies. The reason may be due to the deficiency of reproducing the land–atmosphere interactions. Perhaps, the major contribution of this paper is to clearly show that the regional models and the ERA reanalysis seem to have a major problem in describing the coupling of the surface processes, that affect simulation of precipitation and evaporation in the Amazon.

We expect that understanding model uncertainties and biases, and model development directed to a better representation of land surface and other physical processes will lead to a better simulation of seasonal climate extremes—mainly droughts and floods. This would have strong impacts on seasonal climate prediction and in the generation of projections of future climate change scenarios for extremes.

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