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Ecole doctorale en géosciences



Soil-vegetation-atmosphere interactions in the West African monsoon

Dissertation doctorale présentée par

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en vue de l'obtention du grade de docteur en sciences

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4 septembre 2007

A

André Genette et Hélène Hustinx, mes grand-parents maternels,
Jacques van Vyve et Geneviève Huens, mes grand-parents paternels,
in memoriam.

Remerciements

Ces quelques pages sont celles par lesquelles j'aurais souhaité commencer mon travail de rédaction et sont finalement celles par lesquelles je le termine !

Je souhaite débiter en remerciant Jean-Pascal van Ypersele, déjà mon promoteur de DEA et ensuite de ma thèse. Il est rare de rencontrer une personne d'une telle justesse, autant dans la supervision de ses étudiants que dans la pertinence de ses conseils. Malgré un emploi du temps célèbrement surchargé, il a toujours eu du temps à me consacrer et je le remercie infiniment pour cela. De ces années passées sous sa direction, j'en retire plus de *savoir-faire* que de *savoir*, ce dont je lui suis vraiment reconnaissante—merci également pour sa tolérance zéro pour les fautes d'orthographe et erreurs de grammaire, dans quelque langue que ce soit ! Merci enfin d'avoir laissé un certain libre-cours à mon évolution, et de m'avoir poussée à participer à toutes ces conférences et cours internationaux.

Je voudrais ensuite remercier Hubert Gallée, mon co-promoteur, qui m'a également suivie depuis mon DEA. Personne chaleureuse, positive et enthousiaste, il a toujours réussi à me communiquer sa passion pour ce domaine scientifique. La clarté et la simplicité de ses explications, aussi complexe que puisse être le sujet, furent précieuses.

Je voudrais également remercier les membres du jury pour leur lecture attentive de mon manuscrit, en particulier Guy Schayes, Nick Hall et Alain Cornet. J'adresse une attention particulière à Guy Schayes, dont j'avais suivi les cours en DEA et qui m'a vraiment donné le goût de la météorologie. J'ai toujours beaucoup apprécié la limpidité de ses syllabi et explications, ainsi que les quelques discussions durant mon doctorat.

Puisque tous ces paragraphes débutent par la même phrase, cassons la monotonie en changeant de langue. I would like to thank Nick Hall, with who I started a collaboration based on an idea born in 2005 in Dakar,

Remerciements

Senegal. A collaboration, punctuated by typical English humour, which with his optimism and scientific honesty, led to a publication and a chapter of this thesis, but above all a collaboration I have enjoyed a lot.

Merci aussi à Nathalie Philippon, chercheuse au Centre de Recherches de Climatologie à Dijon, France, sans qui la majorité de ce travail sur l'Afrique de l'Ouest n'aurait jamais ni commencé ni continué. Bien que nous n'ayons pas encore atteint le but du projet, je tiens déjà à la remercier pour son soutien, ses idées et son accueil lors de mes séjours à Dijon.

Mes remerciements vont aussi à l'équipe MAR en général: merci à Christophe pour ses encouragements, son aide et nos nombreux échanges par courriels, merci à Wilfran que j'ai rencontré à Dakar, merci à Romain pour son travail, sans lequel cette thèse n'existerait probablement pas non plus, merci à Xavier pour m'avoir enseigné le MAR et avoir toujours été prêt à m'aider, merci à Fidèle, Benjamin et Moussa que je laisse malheureusement ici sur leur parcours de thèse. On a presque appris le MAR ensemble et je garde d'excellents souvenirs de leurs passages en Belgique et de nos échanges sur la vie et le monde, merci enfin à Olivier et Philippe pour leurs conseils judicieux.

Merci aussi au personnel d'ASTR, en particulier à Nicole, Liliane et Jan. Jan, sans qui mes premières années au 4^e étage n'auraient pas été les mêmes. Nicole et Liliane, toujours prêtes à m'aider avec le sourire pour toutes les situations administratives possibles et imaginables. Un merci tout particulier à Nicole, pour m'avoir si bien aidée à régler les démarches pour l'inscription à l'examen du doctorat, alors que j'avais déjà quitté le pays.

Merci infiniment aussi à Valérie, Xavier, Martin, Wouter, Andrew, Manu, Sébastien et Rafiq, compagnons de thèse, pour les dîners sur le toit, dans l'herbe ou à la cafétéria, les pauses de 4 heures improvisées, les dégustations de desserts faits maison, les ras-le-bol généraux, les sorties, etc. ! Un petit clin d'oeil à Xavier qui a toujours répondu avec enthousiasme à mes innombrables questions informatiques sur les scripts, Bash, Ferret, ...

Je voudrais aussi remercier Bernard Van Renterghem et Jean-Paul Minet, que j'ai aussi assommés de questions lorsque je faisais tourner le modèle sur les machines du CISM et qui m'ont toujours répondu clairement et

Remerciements

rapidement.

Dernières personnes mais non les moindres, je voudrais remercier mes parents, frères et soeurs, qui m'ont supportée sans sourciller pendant ces années de thèse et n'ont jamais cessé de m'encourager. Merci pour votre écoute, votre patience et votre foi en tout ce que je fais ! Merci aussi à Geoff, le père de Justin, qui a relu avec une patience et une méticulosité que j'admire, l'entièreté de ma thèse, et qui m'a si bien aidée à faire de ce texte ce qu'il est aujourd'hui.

Et finalement, merci à Justin ! Je ne sais pas où j'en serais sans toi, ni même si ce texte aurait fini par exister. Merci pour ta patience, tes idées, tes conseils ô combien précieux et éclairés, et la bonne dose de flegme anglais dont tu as fièrement fait preuve et qui était bien nécessaire pour me supporter quand je rédigeais...

Contents

Remerciements	i
Contents	v
Introduction	1
Chapter 1 MAR, a regional climate model	15
1.1 Climate models	16
1.2 The Modèle Atmosphérique Régional (MAR).....	18
1.3 Model features	20
1.3.1 Dynamics	20
1.3.2 Physics	23
1.3.3 Land surface scheme	27
1.4 Setup of all simulations	32
1.4.1 Specific regions of interest	33
Chapter 2 Adaptations to the MAR	37
2.1 Vegetation datasets	38
2.1.1 The Normalised Difference Vegetation Index (NDVI)	38
2.1.2 Translation into vegetation properties for the model	40
2.2 The root fraction	43
2.2.1 Modelling of the roots in a land-surface model	44
2.2.2 Results	49
2.3 Conclusion.....	54
Chapter 3 Internal variability	57
3.1 Introduction.....	58
3.2 Experimental setup	60
3.3 Internal variability and temporal scales	62
3.4 Internal variability and spatial scales	67

Contents

3.5 Internal variability and both temporal and spatial scales	68
3.6 Conclusion.....	71
Chapter 4 Soil moisture and spin-up time	75
4.1 Introduction.....	76
4.2 Experimental set-up.....	77
4.3 Soil moisture anomalies	80
4.3.1 At the initialisation	81
4.3.2 Temporal evolution	85
4.4 Spin-up time	89
4.4.1 Atmospheric spin-up time	90
4.4.2 Global spin-up time	91
4.5 Soil moisture influence	107
4.6 Conclusion.....	110
Chapter 5 Validation of the new MAR	113
5.1 Introduction.....	114
5.2 Validation of the MAR2002	115
5.3 Validation of the MAR2006 (1987 and 1988)	119
5.3.1 Experimental setup and observational datasets	119
5.3.2 Rainfall	122
5.3.3 Dynamics	144
5.4 Conclusion: MAR2006 versus MAR2002.....	155
Conclusion and perspectives	159
Bibliography	167

Introduction

Representing around 20 % of the total African landmass, West Africa is the westernmost region of the African continent. It is bounded by the Atlantic Ocean to the west and south, by the Sahara desert to the north, and by Chad and Cameroon to the east. In terms of relief, a number of topographic features are worthy of mention—the Fouta Djallon hills in the south-west that trigger orographic rainfall, the highlands of Cameroon to the east, and the relatively flat region in between. Further into the desert are found the Aïr and Ahaggar mountains, and in Morocco the Atlas mountains. Figure 1 presents the geography of West Africa and surrounding regions, together with the land cover as observed in 2000 (the most recently compiled database). The West African subcontinent is characterised by the latitudinal (zonal) structure of many of its components. Not only is the coast essentially orientated west-east, but the vegetation changes band by band from dense and wooded vegetation along the coast to sparser and more grassy vegetation northwards, with savannas and shrubs dominating until croplands re-appear, particularly of the Sahel. It should be noted that tropical forests are now restricted to equatorial regions—croplands have replaced most of the tropical forests to the south-west of West Africa. The albedo and soil moisture both follow this distribution. Meteorological quantities such as temperature, specific humidity and rainfall are approximately zonally uniform.

The population of West Africa was estimated to be 271 million by mid-2006 (Population Reference Bureau, 2006¹), that is to say a third of the population of Africa. With a rate of natural increase among the highest in the world (2.6 % against 1.2 %), and a third of the population young people, the pressure on food production and water resources is huge. A local economy based mainly on agriculture has caused significant damage to the forests as cultivable areas are predominantly created by deforestation. Additionally,

¹ Most recent evaluation.

Introduction

the conditions are not favoured by a climate in which rainfall essentially occurs in the form of convective storms and during only a few months of the year, which emphasises the need for a detailed understanding of the climatic system in the region, its variability, and its sensitivity to changes in local factors (such as vegetation cover) and global factors (such as greenhouse gas concentrations).

West African climate

West Africa may be considered as a climatic entity dominated by the *West African monsoon*. This system brings moisture from the Gulf of Guinea to the continent, necessary for convection and precipitation. From the late 60's to the mid-90's, an abnormally long drought occurred (see for example Le Barbé *et al.*, 2002), leading to severe socio-economic impacts. Sahelian soils have seen their productivity decrease if not end; severe famines occurred in the 70's and 80's; cattle populations have diminished; and political arguments have also arisen from the transboundary migrations of cattle and of people (De Félice, 1999, chap. 8). Although rainfall amounts appear to have returned to pre-drought levels in the southernmost Sahel² since the end of the 20th century (Nicholson, 2005), the abnormal succession of less-than-normal rainfalls quickly brought the scientists to investigate possible causes and mechanisms (the AMMA³ project is one example).

The West African monsoon finds its origin in the global circulation. The meridional annual net balance of radiative energy is in excess in the tropics and in deficit in the mid and high latitudes. This imbalance is the primary engine of atmospheric motion and leads, in the tropics, to the Hadley cells of the northern and southern hemispheres. These cells carry energy from the low atmosphere to the upper atmosphere and then polewards. This circulation is marked by seasonal migrations with a summer-poleward shift and an intensification of the cell of the winter hemisphere. The zonal distribution of radiative energy on the Earth is not uniform either. Oceans store more energy than the land (heat contributing to the oceanic

² The drought has intensified in the northern Sahel (further north than 16°N).

³ African Monsoon Multidisciplinary Analysis.

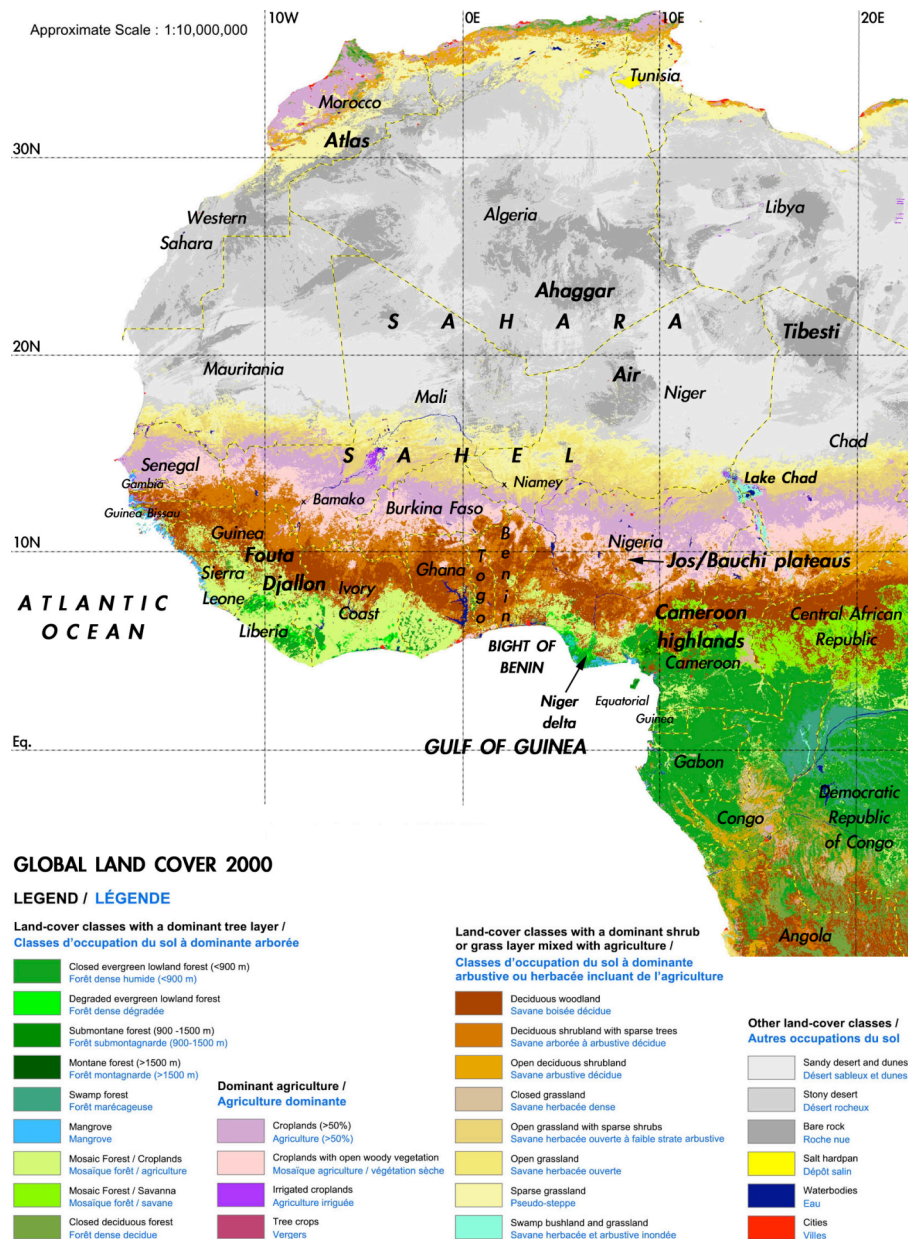


Figure 1 – Map of Africa adapted from Global Land Cover 2000 (GLC, 2000), database combining satellite data (SPOT4, ERS, JERS).

Introduction

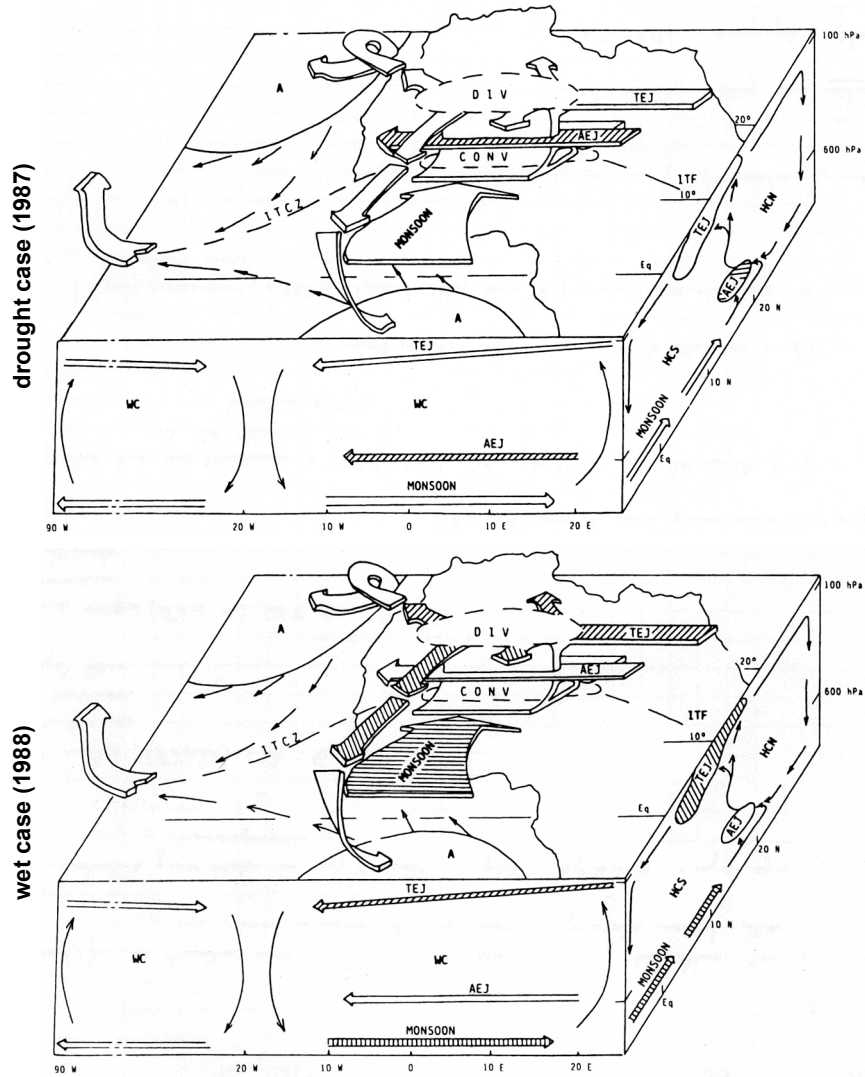


Figure 2 – Schematic diagrams of the West African climate system during abnormally dry (top) or wet (bottom) rainfall seasons. HCN/HCS: Hadley cell of the northern/southern hemisphere, WC: Walker cell, TEJ: tropical easterly jet, AEJ: African easterly jet, DIV/CONV: upper divergence/lower convergence zone, ITF: intertropical front (surface), and ITCZ: intertropical convergence zone. Intensified flows in relation to the normal state are hatched. [Source: Fontaine and Janicot, 1992]

circulation), and land releases heat faster. In the tropics, this leads to the Walker cells that usually have an ascending motion over land and a subsiding motion over oceans.

As illustrated in figure 2, the West African climate system is tightly linked to both the Walker and Hadley cells. Once the trade winds of the Hadley cell of the southern hemisphere (blowing northwards) cross the equator, they are deviated to their right by the Coriolis force, bringing onto the continent a south-westerly moist flow (*monsoon*) which, when linked to convergence and convective motion, brings rainfall onto the continent as far as the Sahel. The region in which the ascending and convective branches of the Hadley cells meet is termed the *Intertropical Convergence Zone* (ITCZ). It corresponds to the area receiving maximum radiative energy and is also an area of concentrated atmospheric moisture—advected by trade winds—and is where the latent heat is released by ascendance and convection. The ITCZ is influenced by the Earth's surface and follows the annual maximum surface temperature, with a one-month delay with respect to the apparent motion of the sun.

The West African monsoon is characterised by strong variability over large space and time scales. At an interseasonal time scale, variability mainly depends on the meridional displacement of a rain band associated to and localised south of the ITCZ. At shorter time scales, spatial variability is due to the convective nature of the rain while temporal variability may be seen as a modulation of the seasonal cycle and the activity of synoptic disturbances.

Three main phases have been identified in the intra-annual behaviour of the monsoon (Fontaine and Janicot, 1992; Sultan and Janicot, 2003) :

- From March to June the *installation phase* takes place, which occurs in steps (Louvet *et al.*, 2003). The ITCZ is located around 5°N and the rain band (previously located over the Atlantic Ocean) extends from the coast northwards. This is the *first rainy season* over the Guinean band (from the coast to around 10°N). A landmark date is the 14th May (± 10 days).
- In late June, the core of the rain band *abruptly shifts northwards* from 5°N to 10°N. This is the end of the first rainy season over the Guinean band and the beginning of the *high rain period* over the Sahel which

Introduction

will last from July to September. The mean date for this shift is the 24th of June (± 8 days).

- From September, the rain band *retreats southwards*, bringing back rainfall over the Guinean band—*second* (and small) *rainy season*.

Four wind regimes have been associated to the West African monsoon (Hastenrath, 1991; Fontaine and Janicot, 1992; see also figure 2). These are:

- The *south-westerly monsoon flow* in the lower atmosphere as previously explained.
- The *African easterly jet* (AEJ) in the middle troposphere (around 600 hPa and 15°N), a jet existing throughout the year but reaching the status of well-defined from April to November with wind speed greater than 10 m/s. It is related to the thermal wind pattern: the increase of easterlies from the surface upwards reflects the juxtaposition of warmer air over the Sahara desert to the north and cooler maritime air to the south.
- The *tropical easterly jet* (TEJ) in the upper troposphere (around 200 hPa and 5-10°N), associated with the Hadley cell of the southern hemisphere and limited to the height of the summer circulation. It lasts from late June to early September with core speeds around 40 m/s. The development of the TEJ is related to the thermal wind pattern during the northern summer. In particular, it owes its origin to the fact that the maximum heating and tropospheric thickness pattern in summer is located in the subtropics rather than near the Equator. It is associated with the upper-level outflow from the Asian monsoon.
- The *subtropical westerly jet* (SWJ) (around 20-35°N). The SWJ is associated with the Hadley cell of the northern hemisphere and located above its descending portion. This jet is best developed in the season of the most intense mean meridional circulation (i.e. northern hemisphere winter time) and peaks at around 30 m/s. It owes its existence to the convergence of the poleward transport of westerly absolute angular momentum in the upper poleward directed branch of that cell.

The monsoon progression is closely linked to the *heat low*, a thermal depression that develops in the region of maximum absorbed solar radiation (Ramel *et al.*, 2006). Figure 3 shows the surface seasonal temperatures and

wind in the boundary layer from January to December, produced by the Modèle Atmosphérique Régional MAR (see chapter 5). As Ramel explains, from March to May, high surface temperatures are found over the Sahel and Sahara, but the maxima are located in the Sahel between 10°N and 15°N. The associated heat low limits the northward extension of the rain band (first rainy season). From early June, although more solar energy is now absorbed by the Sahara soils, the surface temperature increase is slowed down by northerly oceanic winds linked to the intensification of the heat low. In mid-June, the heat low has reached its maximum intensity, the monsoon layer deepens and the moisture convergence around the ITCZ decreases, leading to less rainfall (end of the first rainy season). At the same time, an advection of warm air over the western Sahara is caused by easterly winds associated with high pressures over the Atlas and Ahaggar mountains (subsidence due to the descending branch of the Hadley cell of the northern

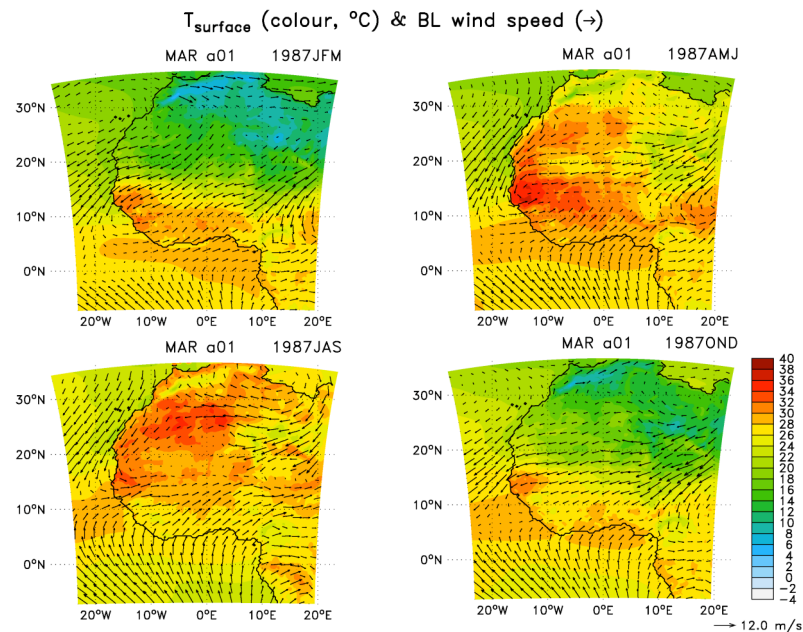


Figure 3 Seasonal evolution of the wind in the boundary layer (BL) and the surface temperature (shaded, °C) modelled by the MAR for JFM (Jan-Feb-Mar), AMJ (Apr-May-Jun), JAS (Jul-Aug-Sep) and OND (Oct-Nov-Dec) 1987.

Introduction

hemisphere). The surface temperatures eventually increase as the upward sensible heat flux decreases, and the maximum quickly shifts from the Sahel to the Sahara. The associated rapid northward displacement of the heat low leads to the observed abrupt northward shift of the rain band in late June (high rainy season over the Sahel). As time passes, the solar energy decreases, the heat low disappears and the ITCZ retreats southwards (second rainy season).

Interannual and intra-annual variability in the monsoon

The monsoon dynamics, and hence the rainfall amount, show a high variability in time and space. The interannual rainfall variability shows irregular succession of *wet* and *dry years*. Wet (dry) years are defined by an annual cumulative rainfall amount greater (lower) than the long-term average. In the Sahel, they show a more northward (southward) position of the ITCZ, stronger (weaker) monsoon flow and TEJ; dry years also exhibit a stronger AEJ (Fontaine and Janicot, 1992; Grist and Nicholson, 2001). This variability is linked to the state of the oceanic and continental surfaces which play an important role in determining the monsoon characteristics and the rainfall variability. The 30-year period of droughts observed in the 20th century has been linked to modification of sea-surface temperature⁴ (Fontaine and Janicot, 1996; Ward, 1998), and to a smaller extent to the continental surface conditions (Eltahir and Gong, 1996; Zheng and Eltahir, 1998). Although the monsoon dynamics are modulated by land-ocean-atmosphere interactions, ocean-atmosphere interactions and land-atmosphere interactions are usually distinguished as they impact on different scales. The former influences the dynamics of wet and dry climate conditions, while the latter may emphasize these conditions.

The *ocean-atmosphere interactions* act through the difference between the sea-surface temperature (SST) and the land-surface temperature. Many studies, based on observations and models, have shown that cold (warm) SST anomalies off the southern coast of West Africa are associated with wet (dry) years in the Sahel region, through an enhancement (weakening) of the

⁴ Increase in equatorial and south subtropical Atlantic Ocean, decrease in north Atlantic Ocean.

meridional gradient of moist-static energy⁵ (MSE) in the boundary layer between the Gulf of Guinea and the continent, which enhances the MSE convergence over the continent and increases convection and rainfall (above references; Lamb, 1978; Folland *et al.*, 1986; Moron *et al.*, 1995; Rowell *et al.*, 1995; Janicot *et al.*, 1998; Vizi and Cook, 2002; Messenger *et al.*, 2004).

The *land-atmosphere interactions* act through soil moisture and vegetation, impacting the surface net energy balance (notably via the albedo), as well as the atmospheric energy and water balances, as illustrated in figure 4. These interactions are forming the subjects of an ever increasing number of investigations. In the early to middle Holocene period (12000 to 5000 years ago), the Sahel and Sahara were "green", as proven by fossil pollens, ancient lake sediments and archaeological evidence. Kutzbach *et al.* (1996) showed that if today's orbital forcing were to be replaced by that of the mid-Holocene period, and/or if the desert were to be replaced by grassland and the desert soil by more loamy soil, current summer rainfall in these regions would be increased by a total of 28 % and vegetation would appear north of 20°N (see also Irizarry-Ortiz *et al.*, 2003). The need to include soil-vegetation-atmosphere feedback in current climate models in order to obtain more accurate simulations of the climate in this region had already been underlined by Charney (1975). He proposed a positive biogeophysical feedback mechanism to explain the occurrence of droughts in regions bordering deserts, such as the Sahel. A decrease of vegetation cover in such a region could lead to self-sustaining dry conditions, by increasing the albedo, causing additional radiative cooling that could only be balanced by adiabatic warming of subsiding air, hence inhibiting convective motions and rainfall. Vegetation over the Guinean band is also thought to have a more significant impact, leading to a monsoon collapse in the case of deforestation (Zheng and Eltahir, 1998), as well as playing an important role in enhancing the interdecadal climate variability in the Sahel (Xue *et al.*, 1990; Zeng *et al.*, 1999). Potential multiple equilibriums have been found in

⁵ The moist static energy *MSE* links the geopotential *Z*, the air temperature *T* and the specific humidity *q* as an energy content per mass unit (kJ/kg): $MSE = gZ + c_p T + L_v q$, i.e the sum of geopotential energy, sensible and latent energy, with *g* the gravity, *c_p* the specific heat of dry air at constant pressure and *L_v* the latent heat of vaporisation.

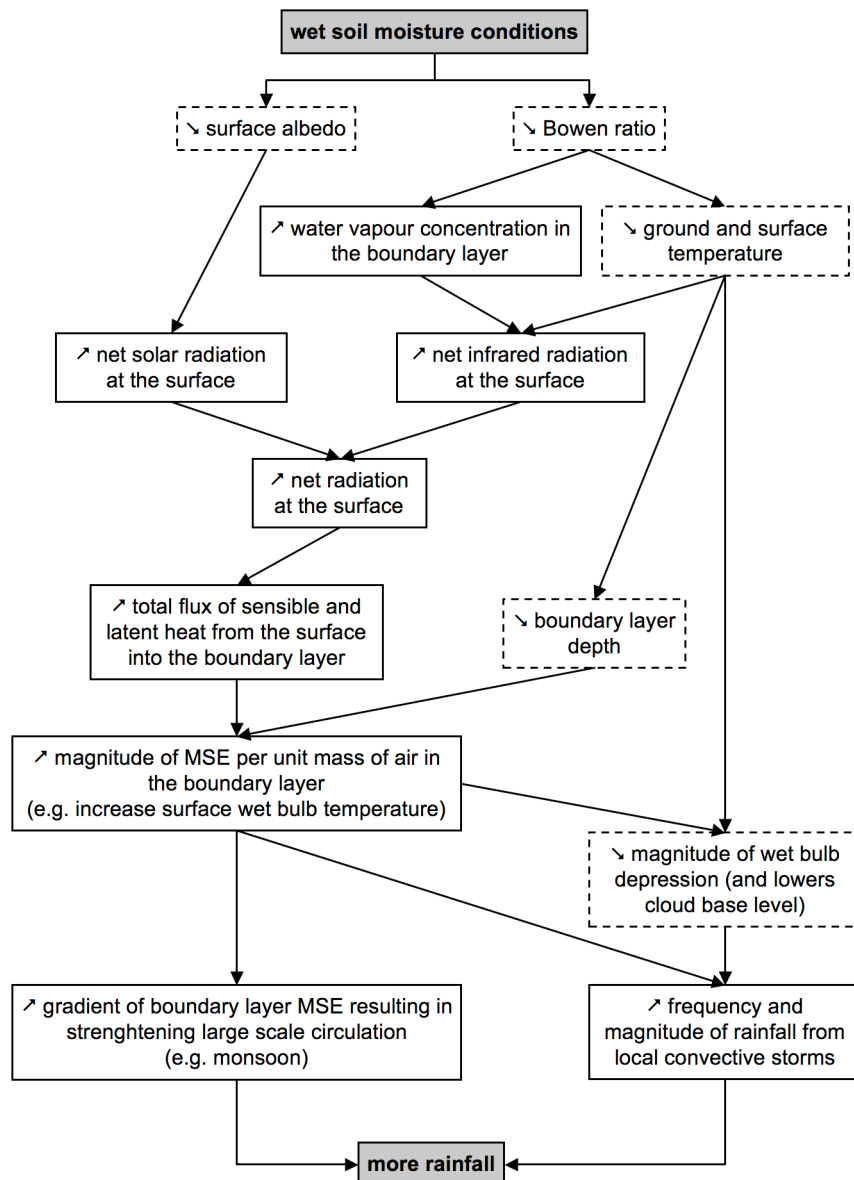


Figure 4 Example of soil moisture and atmosphere interactions, as proposed by Eltahir (1998). Soil contoured frames denote an increase of the quantity, dotted frames a decrease.

the West African climate system (wet or dry and of different intensities). The vegetation would act as a factor for persistence of one equilibrium, or for transition between equilibriums (Kiang and Eltahir, 1999; Wang and Eltahir, 2000a,b). For example, starting from a reduced vegetation, if the consequent lower rainfall is in a sufficient amount, the vegetation is able to fully or partially recover and the system comes back to its previous equilibrium (over several years). If the lower rainfall levels are not enough to allow the vegetation to grow, the climatic system ends up in a different equilibrium (drought).

The influence of soil moisture has not been investigated until recently. There exist both a crucial lack of observations and many questions with respect to its accurate use in climate models and hence its linkage to rainfall and climate in general. The GLACE project⁶ (Koster *et al.*, 2004) investigated the worldwide regions of strong coupling between soil moisture and rainfall with twelve climate models, and the Sahel came out as such a region. Wetter soils lead to smaller albedos and smaller Bowen ratios⁷, increasing the surface net radiation and therefore the total heat flux from the surface. The latter induces in turn a less diluted boundary-layer moist static energy which will increase rainfall by increasing both local evaporation and large-scale moisture convergence (Eltahir, 1998). A more immediate link is the increase of soil moisture due to rainfall, increasing in turn the evaporation and potential for further rainfall, a link that is believed to play a role in the Sudanian band and Sahel (Douville *et al.*, 2001; Douville, 2002).

While previous studies have considered both vegetation cover and soil moisture separately, the Atlantic-Africa MSE gradient (and so the monsoon) is actually modulated by both of them together (Eltahir and Gong, 1996). This result was strengthened by Fontaine and Philippon (2000) and Fontaine *et al.* (2002) who observed that the wettest rain periods in the Sahel are preceded by stronger-than-usual spring meridional MSE gradients. Philippon and Fontaine (2002) also noticed the existence of a significant statistical

⁶ Global Land-Atmosphere Coupling Experiment: joint project involving among others Global Energy & Water Cycle Experiment (GEWEX) and Climate Variability Experiment (CLIVAR); headed up by the Word Climate Research Programme.

⁷ Ratio between the sensible heat and the latent heat.

Introduction

relationship between the boreal summer rainfall in the Sahel (year Y) and the boreal autumn rainfall in the Guinean area (year Y-1). They therefore raised the question of a possible inter-seasonal memory effect carried by the deep soil water content and the vegetation over this latter region in winter. This memory effect would then be transmitted to the atmosphere through a modification of the MSE gradient between the Atlantic and the Guinean region in spring (Philippon *et al.*, 2005).

Modelling the West African monsoon

Different models have been used to investigate the West African climate system and have facilitated an understanding of the land-ocean-atmosphere interactions. In the above references, mainly 2-dimensional models (zonally symmetric) were used, taking advantage of the latitudinal properties of West Africa. Some did and some did not include a detailed surface model, including a vegetation phenology, in order to describe the surface behaviour. The two other categories of models commonly used are the *General Circulation Models* (GCMs) and the *Regional Climate Models* (RCMs), all 3-dimensional models and more realistic than the former.

The first category of models (GCMs) considers the climate over the whole Earth with a typical horizontal resolution of a few hundreds kilometres. They can perform simulations up to several hundreds of years or even more. Over West Africa, such models have been used most commonly to study the interannual variability of the climate. They have helped to relate SSTs patterns to dry and wet years, and have emphasized the need for a proper soil and vegetation description in modelling studies of that region (Vizy and Cook, 2001; Maynard and Polcher, 2003; Osborne *et al.*, 2004; Wang *et al.*, 2004). Recently, more work has been put into climate change in the tropics and has emphasized the uncertainty due to the surface models coupled to the GCMs, SSTs evolution, modelling of ENSO⁸ teleconnections, and the rainfall response (Moron *et al.*, 2003; Douville *et al.*, 2006; d'Orgeval *et al.*, 2006; Paeth and Feichter, 2006). Although the existence of an ENSO teleconnection for eastern Africa is widely accepted, it is not the

⁸ El Niño/Southern Oscillation.

case for West Africa. Some results suggest a correlation between Sahel rainfall and ENSO indexes since the 70's, through connections between the lower and upper troposphere over the world (eastern Pacific, Indian ocean, eastern Atlantic) that increase subsiding air motions over the Gulf of Guinea and West Africa (Janicot *et al.*, 1996; Nicholson *et al.*, 2000). While the temperatures in West Africa are projected to be *very likely*⁹ warmer than the world annual mean warming throughout the 21st century, the signal is much less clear for rainfall, which is not the case for other African regions (IPCC, 2007). Most of the GCMs have struggled to capture the following aspects of the West African climate: climate variability for the past and current climate (some of them do not capture the monsoon), location of the ITCZ (too far south), SST representation (especially in the Gulf of Guinea). The West African climate is particularly hard to model because of the highly convective nature of rainfall and the ocean-land-atmosphere interactions, which play a crucial role in determining the monsoonal circulation. Models with higher spatial resolutions (important for the convective schemes), and parameterisations better adapted to the physics of the region of interest hold considerable advantages.

The second category of models (RCMs) focuses on one region of the Earth. They use a finer resolution (a few tens of kilometres) and perform simulations lasting from a few days to tens of years, using either GCMs results or assimilated observations as forcing at the domain boundaries. They are especially useful to study intra-annual variability of the climate system and local effects, but are also used for interannual variability studies. RCMs currently employed over West Africa are MAR, the model employed in this thesis, MCM, MM5, RegCM, REMO, and RM. Studies have focussed on the feasibility of modelling the West African climate with such models and proven their efficiency. They have investigated the dynamics of the monsoon from different perspectives: intra-annual evolution, convection schemes and rainfall, rainfall accuracy at different time scales, and interannual variability and SSTs (Jenkins, 1997; Druyan *et al.*, 2001; Vizy and Cook, 2002; Gallée

⁹ In the Intergovernmental Panel on Climate Change's terminology, *very likely* refers to more than 90 % probability of occurrence.

et al., 2004; Messenger *et al.*, 2004; Paeth *et al.*, 2005; Ramel *et al.*, 2006; Sijikumar *et al.*, 2006). More recently, the emphasis was brought on vegetation impacts on rainfall. A degradation of the vegetation is thought to lead to a statistically-significant decrease of summer rainfall, threatening food security, and human welfare and health in West Africa (Paeth, 2004).

The present thesis work has been done within this context. It aims to evaluate and improve the results produced by the regional climate model MAR (abbreviation standing for *Modèle Atmosphérique Régional*) over West Africa, paying particular attention to land-atmosphere interactions and how the model represents them.

Chapter 1 describes the features of the MAR and the configuration used for the climate simulations presented in this thesis. Chapter 2 presents a more detailed description of a part of the land model used in the MAR to represent the surface. In particular, an improvement to the modelling of vegetation is presented and validated, as well as the use of new data used to describe the temporal and spatial distribution of the vegetation. Chapter 3 presents a more theoretical study concerning a key element of the model, the internal variability. The internal variability is the inherent uncertainty in the model due to the non-linearity of the atmospheric processes modelled. The aim is to present a quantitative measure of this variability for key variables such as rainfall, surface temperature and wind in the boundary layer, in order to help further analysis of sensitivity experiments. Chapter 4 turns attention to soil moisture, an investigation of how this variable is initialised in the model and how it affects results (spin-up time). It is shown that the method used significantly influences the model results for the first months of simulation, and that caution is required when designing a simulation according to the period of interest. Each of these chapters uses a widely validated version of the MAR. Chapter 5 focuses on a new version of the model, released in 2006. It presents a simulation for two years, respectively categorised as dry and wet, and validates the model climatology. The last section of the thesis concludes and summarises the work and findings and presents a number of perspectives for further work.

Chapter 1

MAR, a regional climate model

Purpose

The main tool of this thesis is the Modèle Atmosphérique Régional, commonly refer to as *MAR*, hence its users are often referred to as *Martians*. Categorised as a *regional climate model* or *limited area model*, it allows us to represent the (past) climate over a specified region at a high spatial and temporal resolution—typically an area of 4000 x 4000 km² with a horizontal precision of 20 to 50 km and a time period cover ranging from a few months to a few or many years. We here briefly talk about this category of climate model and detail the main characteristics of the MAR, emphasizing its tropical adaptations and its configuration for the West Africa region.

Contents

1.1	Climate models	16
1.2	The Modèle Atmosphérique Régional (MAR).....	18
1.3	Model features	20
1.3.1	Dynamics	20
1.3.2	Physics	23
1.3.3	Land surface scheme	27
1.4	Setup of all simulations	32
1.4.1	Specific regions of interest	33

1.1 Climate models

In 1922, the British scientist L. F. Richardson described a numerical method to predict the weather (Richardson, 1922). He expressed the differential equations governing the atmospheric motions as a number of algebraic difference equations for values of the tendencies of various field variables at a finite number of points in space. He showed it was possible to compute numerically the tendencies of each variable, given their observed values at these grid points. These tendencies could then be extrapolated further in time to provide a new set of values for the field variables. The latter data could then be re-used to compute the tendencies, and so on and so forth. At that time, Richardson estimated that a work force of 64000 people would be required for these computations just to keep up with the weather on a global basis...

Nowadays, with computers ever more powerful and fast, Richardson's method has become widely feasible. Most numerical climate models are based on his formulation of the equations of the atmosphere dynamics, equations usually referred to as *primitive equations*. They permit study of the Earth's climate at various spatial and temporal scales, whether for past, current or future climates. Two categories of models are distinguished when performing climatic simulations¹⁰: the *General Circulation Models* (GCMs) and the *Regional Climate Models* (RCMs)—also known as *Limited Area Models* (LAMs). The first category models consider the climate over the whole Earth with a typical horizontal resolution of a few hundreds kilometres. They can perform simulation up to several hundreds of years or even more. The second category of models focusses on one region of the Earth. They use a finer resolution (a few tens of kilometres) and perform simulations lasting from a few days to tens of years.

Our model is an active member of the second category. The modelled region could be seen as a right-angled parallelepiped sitting on the region of interest. In the case of West Africa, this volume is horizontally divided into

¹⁰ As opposed to forecasts that aim to foresee the weather of the next days.



Figure 1.1 – Machine NEC SX-8 of the Institut du Développement et des Ressources en Informatique Scientifique (IDRIS), Paris, France, on which were performed most of the simulations presented in this thesis. [Picture source: www.idris.fr]

134 grid meshes in both x- and y-directions (approximately west-east and south-north), each 40-km long. Vertically, there are 40 divisions and our volume extends to around 40 km in height from ground level. For each of these 718240 points ($134 \times 134 \times 40$) a calculation is performed every 150 s of the model time for a range of variables such as pressure, temperature, air humidity, wind speed, rainfall, cloud cover, incoming solar radiation, surface outgoing radiation, surface temperature, vegetation properties, etc. Results are most commonly stored for every 6 hours of the model time. For a year-long simulation, this leads to approximately a billion data points per variable.

Although technical progress is made year upon year (cf. figure 1.1) and simulations are performed faster and faster, the climate models are still very demanding in time and disk space for generating and storing the results. A one-year long simulation typically requires 150 hours to be executed¹¹. It will be divided into slices of one month or ten days (usually) simulated one after the other as it is simply technically impossible to do the whole in one go. Multiple users of the same computing resources imply sharing and

¹¹ Estimation based on the performances obtained the NEC SX-8 of the IDRIS institute. This machine has 10 nodes with 9 processors each, with 16 GFlops crest per processor, a clock frequency of 2 GHz and a 64 Go memory per node. The operating system currently is SUPER-UK 16.2.

waiting time between the realisations of each slice. The waiting time can easily double the real time needed to perform the simulation.

In total, for the results presented in this thesis, we have performed 244 months of simulation which is equivalent to 20 years. It required 3050 hours of CPU time, that is around 127 entire days and a total amount of data of 732 Gb. That is of course a lower boundary, since in this estimate we have not included waiting time, or the resources needed to prepare the model runs, those needed to postprocess the results or all the additional simulations performed for testing.

1.2 The Modèle Atmosphérique Régional (MAR)

The regional climate model we have used is called MAR, the acronym standing for *Modèle Atmosphérique Régional* (or more precisely *Modèle tridimensionnel Atmosphérique à l'échelle Régionale*). It was initially developed by Gallée and Schayes (1994) at the Université catholique de Louvain in Louvain-la-Neuve, Belgium. It is aimed to study mesoscale processes in the atmosphere over a selected area of the Earth.

So far almost each continent has been approached. Initially developed to study the polar climate of Antarctica¹², it has been progressively adapted to other regions in the world such as Greenland¹³, Europe¹⁴, the Middle East¹⁵, and, last but not least in the current context, Africa.

Over Africa, the very first study was carried out by Ntezimana (1993)

¹² The katabatic winds with the pioneer article on the MAR (Gallée and Schayes, 1994), atmospheric circulation in different parts of the peninsula and more recently the ice-sheet surface mass balance (Gallée *et al.*, 2001).

¹³ Among others, 2-D or 3-D modelling of the atmospheric circulation (Lefebvre *et al.*, 2005), study of the ice sheet mass balance (most recent: Fettweis *et al.*, 2006).

¹⁴ The first studies emphasized the meteorology: wind gusts, extreme events over Belgium, hydrology of the Alpine region (e.g.: Brasseur *et al.*, 2002a,b). Some significant technical work was also achieved for the lateral boundary treatment by Marbaix *et al.*, 2003.

¹⁵ Studies focussed on the land-surface model coupled to the MAR: impact of land-use changes on local climate in Israel (De Ridder, 1997) or more generally land-surface processes and climate change in semiarid region (De Ridder and Gallée, 1998).

who focused on the convection over Lake Victoria. Since then, studies have concentrated on modelling the West African monsoon. Probably the pioneer for detailed studies over West Africa, Moufouma-Okia (2003) dedicated his PhD thesis to investigating the years 1987, 1988 and 1992. He looked in detail at the impact of the boundary forcings on the model results and concluded that the MAR rainfall is very sensitive to the forcing large-scale data. He also analysed the improvements of the convection scheme of Bechtold et al. (2001) with respect to that of Brasseur (1999), in terms of representing the seasonal variability of the precipitation.

Messenger *et al.* (2004) validated the model over 1983 and 1984 and then used the MAR to investigate the link between the sea-surface temperatures (SSTs) and the rainy season for both these years. The model could reproduce what is found in observations and this study helped to identify the physical mechanism at work with the SSTs: for example, warmer SSTs increase the evaporation flux and the meridional moisture transport in the lower atmosphere, leading to more abundant rainfall along the coast. Gallée *et al.* (2004) evaluated the model performance over the year 1992 and could propose, on basis of the model results, a mechanism to explain the oscillations of the rainy regimes (transition between two regimes of the Hadley cell). Finally, Ramel (2005) went further into the validation process. During his PhD, he ran the model over 10 years (1983-1992) and validated very meticulously the rainy regime of the model. He also studied surface-atmosphere interactions over the year 1992, improving the representation of the surface properties in the model (albedo and vegetation). The work presented in the current thesis would not have been achievable without this important step. He finally looked into detail at the physics responsible for the abrupt northward shift of the monsoon late June and proposed a new mechanism for it (see Ramel *et al.*, 2006). The model's results have also been employed at smaller scales: for example, the precipitation was used to force a hydrological model over a basin situated in the Sahelian region (Messenger *et al.*, 2006). The last published study investigates the internal variability inherent in the model in order to support the evaluation of further sensitivity studies (Vanvyve *et al.*, 2007).

1.3 Model features

The MAR is a grid-point hydrostatic¹⁶ primitive equation model in which the atmosphere is considered as a compressible fluid. It is either spatially 1-, or 2- or 3-dimensional depending on the aim of the study.

The model could be described as a dynamic core using (i) large-scale data and (ii) physical parameterisations in order to close and solve a system of differential equations. These equations and parameterisations model all relevant atmospheric processes (see figure 1.2), such as the radiative energy transfer through the atmosphere, the mesoscale transfer of heat/energy/water, the hydrological cycle, the microphysical processes undergoing in clouds and precipitation, and the transfer of heat/energy/water in the boundary layer. The surface behaviour is modelled separately by a 1-dimensional model—i.e. called for each surface grid mesh—known as SISVAT.

The following three sections briefly detail in turn the dynamics of the model (section 1.3.1), the physical parameterisations (section 1.3.2) and the surface model (section 1.3.3).

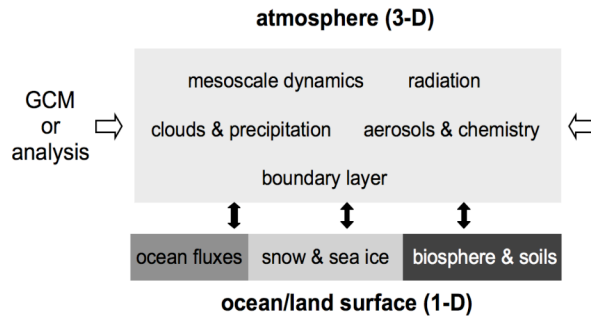


Figure 1.2 – Sketch of the MAR.

1.3.1 Dynamics

a. System of coordinates

The MAR uses a horizontal Cartesian system with a vertical σ coordinate

¹⁶ A non-hydrostatic version is currently being developed.

(also called *terrain following coordinate*) illustrated in figure 1.3. In such a system the ground becomes a coordinate surface ($\sigma = 1$) even in the presence of spatial and temporal surface pressure variations. It is especially useful in regions of strong topographic variations. (Pielke, 2002)

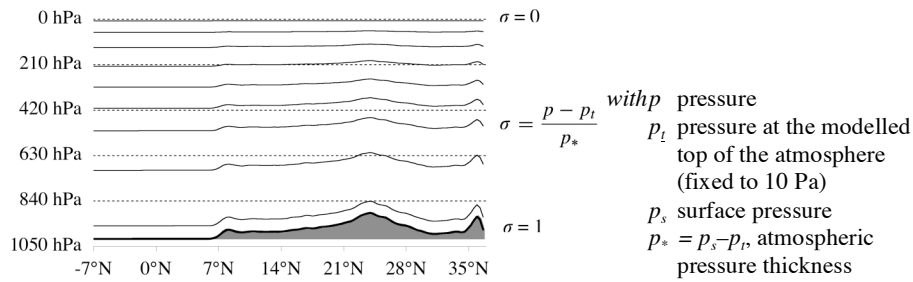


Figure 1.3 – Illustration of the σ -coordinate system. Solid line: σ -levels. Dashed line: pressure levels more or less mimicking the altitude. Shaded area: approximated surface topography.

b. Pronostic equations

The motions of the atmosphere are governed by three fundamental conservation principles for momentum, mass and energy, and these are expressed as pronostic¹⁷ equations. (Holton, 2004)

The equations of **conservation of momentum** are derived from Newton's second law of motion: the rate of change of momentum is equal to the sum of all forces acting—pressure gradient force, gravitational force, and viscous force. Once transposed in a noninertial reference frame (rotating earth), i.e. taking into account the acceleration of the coordinates, two apparent additional forces arise: the centrifugal force and the Coriolis force. The vertical momentum equation is approximated by the hydrostatic equilibrium. The equation of **conservation of mass (continuity equation)** states that there is no loss of mass in the system: the net rate of mass inflow through the sides of a volume element with fixed Cartesian coordinates must be equal to the rate of accumulation of mass within this volume. In the model MAR, the full continuity equation is used, i.e. the atmosphere is considered as a

¹⁷ Time dependent, by opposition to diagnostic equations.

compressible fluid¹⁸. The equation of **conservation of energy** is obtained from thermodynamics: the total thermodynamical energy of a fluid parcel is the sum of the internal energy due to the kinetic energy of the individual molecules and the kinetic energy due to the macroscopic motion of the fluid. The rate of change of this total thermodynamic energy is equal to the rate of diabatic heating plus the rate at which work is done on the fluid parcel by external forces such as pressure, viscosity or gravity.

Those equations of conservation (see Gallée and Schayes, 1994, for their mathematical expression) are complemented by the equations of **conservation of water** which take into account phase changes between solid, liquid and gaseous states of water that mainly occur in clouds. The conservation of water vapour is complemented with equations of conservation of cloud droplets, rain drops, cloud ice crystals, and snowflake (see appendix of Gallée, 1995, for a full description).

This provides us with a set of 12 equations with 12 unknowns: zonal and meridional wind velocities¹⁹, vertical velocity, geopotential, atmospheric pressure thickness, density, potential temperature, concentrations of water vapour and four additional hydrometeors. Pressure and temperature are obtained from the definition of the σ coordinate and the potential temperature respectively.

c. Initial and boundary conditions

By definition of a regional climate model, the domain of simulation is enclosed. In order to solve the pronostic equations—i.e. compute a unique solution per grid point and time step of integration—initial and boundary conditions need to be specified. The upper and lateral boundaries have no physical meaning while the lower boundary corresponds to the surface of the Earth. Logically, the surface model provides numerical values for the latter.

The upper boundary is conditioned by zero gradient of some pronostic variables. Some care must be taken to avoid backward reflection of wave

¹⁸ No approximation is made for the mass conservation.

¹⁹ The zonal wind velocity refers to the east-west component of the wind vector, the meridional wind velocity refers to the south-north component of the wind vector (i.e. along a meridian).

energy on the upper surface that could perturb the physical solutions. This is achieved by introducing a viscous damp layer beneath the upper surface that strongly attenuates this reflection (Klemp and Lilly, 1978).

The lateral boundary is a double problem since information travels in both directions: towards the inside (driving fields) or towards the outside (evacuation of the physical and numerical waves generated by the model²⁰). The driving fields are treated with a relaxation boundary scheme (Davies, 1976). Within the *buffer zone*, a zone that borders all lateral boundaries, this scheme constrains the model solution towards the large-scale fields values. See Marbaix (2000) and Marbaix *et al.* (2003) for more detail on the treatment of boundaries.

ERA40

The most commonly used large-scale driving fields for the model MAR are the **re-analyses ERA40** produced by the European Centre for Medium-range Weather Forecast²¹ (Uppala *et al.*, 2005). These data are produced by assimilation of both satellite (SSM/I, ERS, Meteosat GOES and GMS, etc.) and *in-situ* measurements in the spectral global circulation model T159. They cover the period September 1957 to August 2002 with a horizontal resolution of approximately 115 km (over West Africa) and a vertical discretisation of 60 levels.

1.3.2 Physics

The spatial resolution of models (both global and regional) has significantly improved thanks to the fast development of computing resources. However, it is still too coarse to represent explicitly all the dynamics and diabatic processes governing the climate—radiation and its interaction with the atmosphere compounds, vertical mixing through dry and moist convection, the formation of clouds and precipitation, turbulent

²⁰ In the MAR, there is no feedback to the large scale, i.e. what the model produces is not exported towards the driving fields, does not interact with them so that some modified large-scale data could be fed back into the domain. Such a method is called *one-way nesting*.

²¹ ECMWF, www.ecmwf.int.

transfers, etc. These subgrid-scale processes need to be represented in terms of what the model can do at its own scale. Such an approximation of unresolved processes in terms of resolved variables is called *parameterisation*. It is far from an easy task and is a crucial element in climate modelling.

The physics parameterisation schemes in the MAR are numerous. We shall only present the main ones and emphasize the link with the West African climate.

a. Radiation

The radiative transfer scheme handles the solar and infrared radiations all along their travel inside the modelled atmosphere.

The *solar radiation* is computed by the scheme of Fouquart and Bonnel (1980). It includes the transmission/absorption in the wavelength bandwidth 0.25–4 μm and takes the hydrometeors into account in the computation of the optical properties of the clouds following a more recent formulation (see Gallée, 1995 for exact references).

The *infrared radiation* follows the *wide band* formulation proposed by Morcrette (1984) for global circulation models. The heating/cooling rate of each vertical column of atmosphere is determined from the temperature and humidity vertical profiles and from the information on the cloud cover.

b. Convection

Deep convection is the main mechanism for vertical transport of momentum, humidity and energy in the tropical climates. Over West Africa, the individual convective cells tend to organise themselves into mesoscale convective systems²² extending over several hundreds of kilometres (horizontally) with lifetimes ranging from several hours to several days. Over the Sahel, such systems generate up to 90 % of the annual rainfall.

The convection acts upon spatial scales unresolved by the current climate models. Hence, its effects are represented by *deep convection schemes*

²² A mesoscale convective system is a group of cumulonimbus clouds linked to each other by a stratiform anvil.

usually developed according to specific spatial resolutions and needs. Their task is (i) to diagnose the presence of large-scale atmospheric conditions susceptible to the generation of convective activity (*trigger function*) and (ii) to determine the variations of temperature, humidity and momentum that correspond to this activity (*closure hypothesis*). They modify the vertical distribution of heat and humidity, the spatial distribution of clouds, the precipitation, and the regional circulations.

In the model MAR, the best convection scheme for the West African climate is that of Bechtold *et al.* (2001), based on the scheme of Kain and Fritsch (1990). This *mass flux*²³ scheme includes both shallow and deep convective clouds. Its best performance is for horizontal resolutions of 20-25 km. The most important parameter is the Convective Available Potential Energy (CAPE). In order to determine the intensity of the convection, it assumes an equilibrium between the CAPE produced by the motions resolved by the model and its consumption by the deep convection scheme.

The main characteristic of the scheme is the representation of clouds as a pair of stable circulations, composed of down- and updrafts of which the mean properties are determined through a 1-column model of cloud. For each vertical column of the model (above 20 m high), the *trigger function* determines a portion of instable air (in terms of moist convection²⁴) and lifts it above the Lifting Condensation Level²⁵. The *closure hypothesis* postulates that all CAPE in a grid element is removed within an adjustment period, i.e. is consumed by the convection so that the atmosphere is re-equilibrated. For deep convection, the adjustment period depends on the grid mesh surface

²³ Convection schemes allow the representation of the effect of a convective cloud population on its environment. As opposed to *convective adjustment* schemes that adjust the temperature and humidity lapse rates to a reference profile, *mass flux* schemes attempt to model convective processes. They include a cloud model to estimate the convection properties and assume that the convection acts to restore the environment to a stratification based on moist parcel stability.

²⁴ The *moist convection* is convection in which the phase changes of water play an appreciable role. The *dry convection* is convection which occurs without formation of clouds.

²⁵ The level at which a moist air parcel lifted dry-adiabatically would get saturated.

and the mean horizontal wind between the LCL and 500 hPa. See Bechtold *et al.* (2001) for a full description.

c. Microphysics

Cloud systems that develop over West Africa are generated by the convection and/or by the convergence due to the intertropical convergence zone. They are complex—mesoscale convective systems, squall lines, etc.—and impact upon their surroundings at multiple time and spatial scales—modulation of the tropical easterly jet, interaction with easterly waves, modulation of the rainfall evaporation at the surface through radiative effects. They are still not fully understood nor well modelled. As an attempt to improve the results in tropical regions, the model MAR has an explicit representation of the hydrological cycle (as mentioned in 1.3.1b). It was first introduced by Nteziimana (1993) and is closely related to the work of Kessler (1969) (see full description in appendix of Gallée, 1995).

The following microphysical processes are taken into account in the model: condensation of water vapour to form cloud water, evaporation of cloud water, homogenous freezing of cloud water to form cloud ice, homogenous deposition of water vapour to form ice, heterogeneous ice nucleation, sublimation of ice, depositional growth of cloud ice at the expense of cloud water, melting of cloud ice to form cloud water, autoconversion (aggregation) of cloud ice to form snow, depositional growth of snow, sublimation of snow, accretion of cloud ice by snow, accretion of cloud water by snow, accretion of rain by snow, accretion of cloud ice by rain to form snow, accretion of rain by cloud ice to form snow, probabilistic freezing of rain to form snow, melting of snow to form rain, snow fallout, autoconversion of cloud water to form rain, evaporation of rain, accretion of snow by rain, accretion of cloud water by rain, and rain fallout.

d. Turbulence

The boundary layer is the part of the atmosphere directly influenced by the surface of the Earth. It is, among other things, characterised by turbulence generated by (i) convection (due the surface heating) and (ii)

mechanical friction (due to the surface roughness). The turbulence is important for the transport of heat and humidity from the surface upwards and for the transport of momentum towards the surface. It is taken into account in the pronostic equations by replacing variables such as the zonal and meridional winds, the temperature, etc., by their mean (slowly varying mean fields) plus a fluctuation around the mean (rapidly varying turbulent components)²⁶. This leads to additional terms of the kind $\overline{\sigma'\theta'}$ with σ the vertical speed, θ the potential temperature, the overbar the mean, and the prime the turbulent fluctuation. However, this introduces additional unknowns into the set of equations and a closure assumption is needed to express these turbulent fluxes (vertical turbulent heat flux in the above example) in terms of known variables.

Over Africa, the model MAR now²⁷ uses the E - ε turbulent closure model as applied by Duynkerke (1988) and modified according to Bintanja (1998) and Bintanja (2000). E is the turbulent kinetic energy²⁸ and ε is the viscous dissipation (of energy by molecular viscosity), both associated to a pronostic equation. It uses the K theory, according to which the turbulent fluxes are expressed as the local gradient of the mean term scaled by a diffusivity or viscosity term K (eddy-exchange coefficient) and K is evaluated from E and ε ($K \sim E^2/\varepsilon$). The closure order is 1.5. The model MAR previously²⁹ used the closure model of Therry and Lacarrère (1982) (order 1.5).

1.3.3 Land surface scheme

The atmospheric part of the model MAR is coupled to a one-dimensional (in the vertical) Soil-Ice-Snow-Vegetation↔Atmosphere Transfer scheme referred to as SISVAT. It is described in De Ridder and Schayes (1997) and De Ridder (1997). The scheme has been validated over France (Landes), Belgium (Louvain-la-Neuve) and Brazil (Amazonian rainforest) (De Ridder

²⁶ Reynolds averaging.

²⁷ The most recent version of the code.

²⁸ $E = (\overline{u'^2} + \overline{v'^2} + \overline{w'^2})/2$ (per unit mass) with u' , v' and w' the turbulent components of the zonal, meridional and vertical wind speed respectively.

²⁹ The version dating from March 2002 (see subsection 1.4 for details).

and Schayes, 1997), over Israel (De Ridder and Gallée, 1998), and over the Sahel (Derive, 2003).

The SISVAT links the surface and the atmosphere by calculating heat and moisture exchanges between the bare soil, the vegetation and the lower atmosphere—computations are done separately for the bare soil and the vegetation. As illustrated in figure 1.4, it currently has one vegetation layer, one soil skin layer (2 mm thick) and six soil layers reaching 2.888 m in depth. The surface type can be water, land, ice or snow. Note that each mesh of the model MAR at the surface is divided into three submeshes of which the relative importance is described through the *fraction of vegetation (or bare soil)*.

We describe hereafter the main properties of the vegetation and soil and the physics components of the scheme.

a. Vegetation

The vegetation is described through the first two submeshes. Each of them contains a different type of vegetation (the dominant and secondary ones in the mesh). There are 12 types available: three types of crops, grasses, broadleaf trees and needleleaf trees. These types are deduced from the International Geosphere-Biosphere Programme (IGBP) dataset. Figure 1.5 shows the vegetation types for West Africa as seen by the model—note that no needleleaf trees appear and that the vegetation appearing over the Sahara for $k = 2$ (second submesh) will be removed thanks to an associated null fraction of vegetation.

With each submesh is associated a leaf area index (LAI)—the area of soil covered by the vegetation leaves—and a green leaf fraction—the quantity of green leaves, i.e. those alive. These parameters and the fraction of vegetation are obtained by analysing the Normalised Difference Vegetation Index, NDVI (refer to chapter 2 for details). Finally, each type of vegetation impacts on the ground moisture via the root fraction parameter—the amount of vegetation root per layer of soil—as explained in chapter 2.

The no-vegetation case is described through the third submesh that also

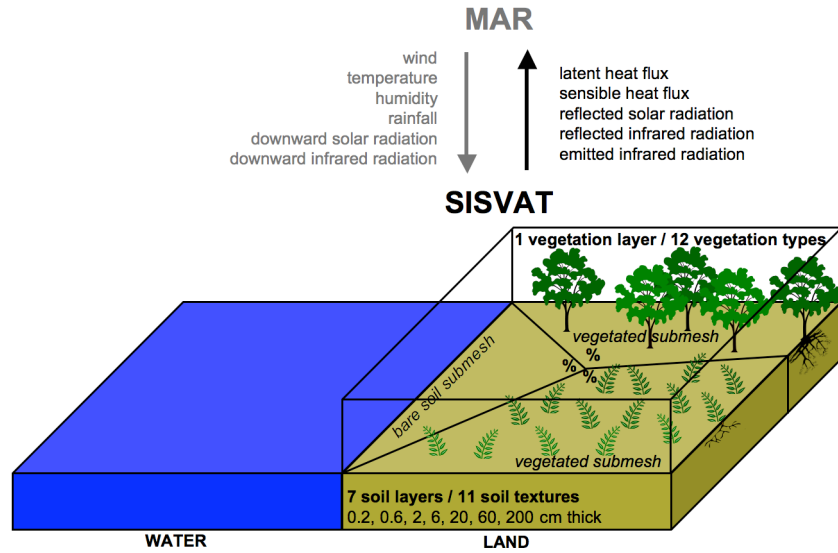


Figure 1.4 – Sketch of the land surface model SISVAT for two meshes (water and land surface types). Arbitrary vegetation types.

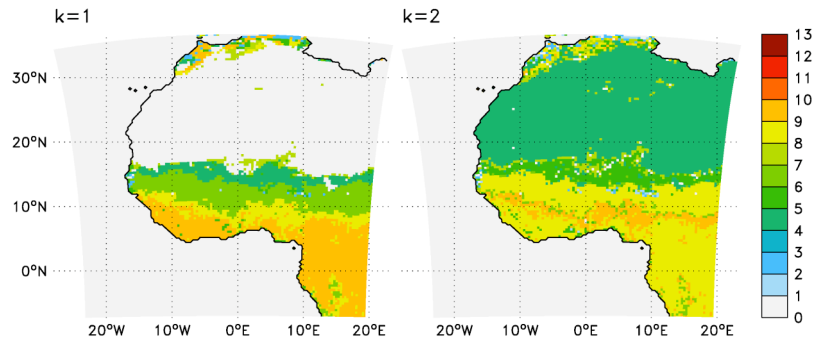


Figure 1.5 – Types of vegetation in West Africa as seen by the model for the first two submeshes ($k = 1, 2$). Types 1 to 3: crops, respectively small, medium and high in size (essentially properties of the millet). Types 4 to 6: small, medium and high grasses, i.e. sparse and dense tiger bus. Types 7 to 9: broadleaf trees, types used to describe fallows. Types 10 to 12: needleleaf trees, types nonexistent in this case.

has an associated "fraction of vegetation" in order to describe its relative abundance with respect to the vegetated submeshes.

b. Soil

There are 11 soil textures available, defined according to the Food and Agriculture Organisation, FAO. The purpose of this is to take into account the soil properties such as albedo, water transfer and thermal properties. Figure 1.6 shows the soil texture for West Africa as computed by the model, alongside the texture triangle.

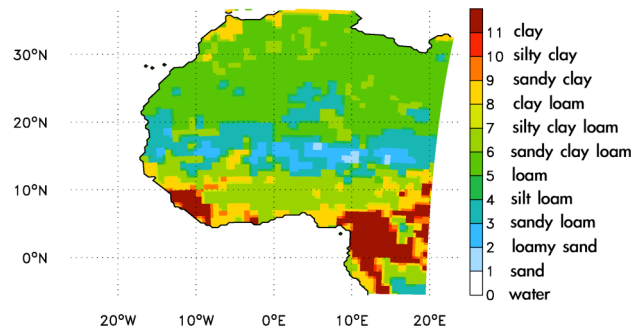


Figure 1.6 – Map of the soil texture in the model MAR.

c. Modelled processes

From the wind speed, temperature and moisture of the first atmospheric layer, the downward shortwave and longwave radiation fluxes, and the rainfall amount, the scheme computes the sensible and latent heat turbulent fluxes, the reflected solar radiation, and the emitted/reflected thermal longwave radiation fluxes at the surface. These depend on the energy and water conservation principles for both the vegetation and the bare soil. We describe hereafter each of the model subcomponents.

The solar energy absorbed by the surface and transmitted to the atmosphere as infrared energy is of main importance for determining the climate. Therefore, the scheme computes what quantity of solar energy is absorbed/reflected by the surface (vegetation and bare soil separately) and what quantity of infrared energy is absorbed/emitted. It takes into account

the soil properties such as the surface moisture (modifying the albedo) and the soil temperatures (modifying the emitted infrared radiation), and the vegetation properties such as the leaf area index, the green leaf index and the leaf optical properties (emissivity, leaf angle, etc.). Green leaves are those interacting with the radiative energy.

The turbulent transfer of heat and moisture in the soil and between the vegetation/bare soil and the atmosphere is calculated through resistances (similarly to electrical resistances). Both sensible and latent heat fluxes appear as the sum of the vegetation and bare ground contributions. A distinction is made between direct evaporation of moist leaves (due to dew or intercepted rainfall) and leaf transpiration. The vegetation transpiration is the expulsion of moisture by the leaves via their stomata³⁰. That water comes from the soil and was absorbed by the roots. In the model, the only flux of water through the plant is from the ground to the atmosphere (through the roots and leaves). There is not excretion of water by the roots nor uptake of air moisture by the leaves. Again with similarity to electrical resistances and potentials, the transpiration is described through leaf and soil water potential (expressed as a negative height) and some resistances (internal plant, soil-root, stomata). These parameters help determine if the roots pump ground water and if the vegetation transpires (cf. chapter 2 for details).

Pronostic equations are used for the water balance in the vegetation layer, the transfers of heat and liquid water in the soil. The heat and water transfer uses hydraulic and thermal conductivities in the seven soil layers, soil water potential, root water uptake, etc. There is no lateral exchange of heat or moisture in the soil, i.e. between the submeshes.

The SISVAT also takes into account snow and ice on the land, but both are irrelevant processes for West Africa.

³⁰ The stomata are microscopic holes in the leaves that serves for gaseous exchanges between the plant and the air. They are controlled by the plant: closed when risks of drought, opened to cool down the plant.

In the SISVAT, the stomata opening is ruled by the effective LAI calculated from vegetation properties. It determines the leaves that receive enough radiative energy ($> 20 \text{ W/m}^2$) in the 400-700 nm wavelength band in order for their stomata to open.

1.4 Setup of all simulations

Two versions of the model MAR have been used to produce the results shown in this thesis. The first one is dated March 2002 and is further referred to as the MAR2002. The second one is dated June 2006 and is further referred to as the MAR2006. Apart chapter 5, which presents the validation of the new version, the MAR2002 has been used throughout. The reader can refer to the chapter 5 where both versions are presented in more detail.

All simulations share a common set of parameters shown in table 1.1. For the sake of simplicity, the table also includes main differences in the physics parameterisation. These will not be repeated subsequently.

Following Ramel's recommendations, the domain has been extended northwards and eastwards (see figure 1.7) in order to improve the representation of the abrupt northward shift of the rain band (linked the heat low over the desert) and rainfall over the coastal band (linked to easterly waves). Indeed, the evaporation over Central Africa is thought to contribute for almost 1/5 of rainfall over West Africa, i.e. almost as much as that over the Atlantic Ocean (Gong and Eltahir, 1996).

	MAR2002	MAR2006	
Dynamics	Slow dynamics time step	150 s	idem
	Spatial discretisation	$\Delta x = \Delta y = 40$ km, 40 σ -levels	idem
	Modelled domain extent	134x134 meshes (i.e. 5360x5360 km ² / $\pm 25^{\circ}\text{W}$ - 22°E , 6°S - 35°N)	idem
	Initial conditions	ERA40 (see subsection 1.3.1c)	idem
	Boundary conditions	ERA40 every 6 hours	idem
	(buffer area)	(6 meshes)	(7 meshes)
Physics	Physics time step	150 s	idem
	Radiative scheme	LMDz	ECMWF
	Convection	Bechtold <i>et al.</i> (2001)	idem
	Turbulence closure	Therry and Lacarrère (1982)	Duynerke (1988)
	Surface model	SISVAT	idem

Table 1.1 – Configuration of the model MAR (both versions).

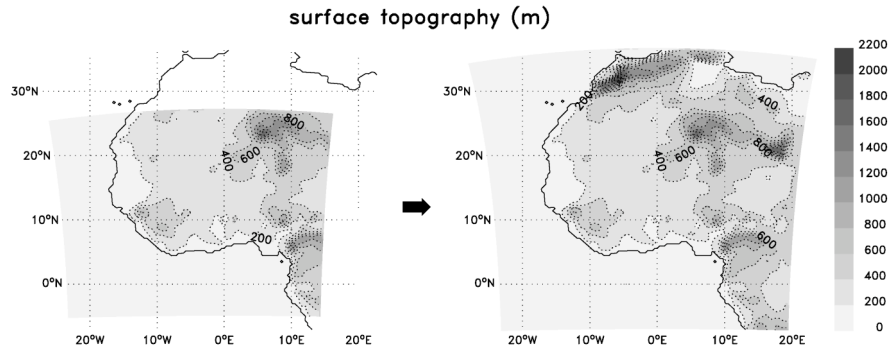


Figure 1.7 – Previous (left) and current (right) domain extent.

1.4.1 Specific regions of interest

Some specific regions are of interest for the analysis of results and presented hereafter.

a. Buffer area and inner domain

Figure 1.8 shows the *buffer area* and the *inner domain*. The buffer area is a band 6- or 7-mesh wide used for relaxing the model values towards the driving fields (cf. previous section). The inner domain corresponds to an inner region of the atmospheric domain, far enough from the boundaries not to be directly affected by the driving fields. *Unless specifically noted to the contrary, all graphics and computations in the thesis refer to the data available in the domain of simulation minus the buffer area.*

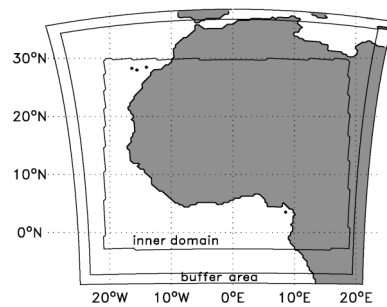


Figure 1.8 – Domain of simulation (external border), buffer area (between the external and internal borders) and inner domain (central square region).

b. AMMA regions and common appellations

AMMA (www.amma-international.org) stands for African Monsoon Multidisciplinary Analyses. The creation of the AMMA project was motivated by the decrease of rainfall observed during the second half of the 20th century in West Africa. Why did these droughts occur? What can be expected during the 21st century? The three main goals of AMMA are (i) *"to improve our understanding of the West African monsoon and its influence on the physical, chemical and biological environment, regionally and globally"*, (ii) *"to provide the underpinning science that relates variability of the West African monsoon to issues of health, water resources, food security and demography for West African nations and defining and implementing relevant monitoring and prediction strategies"*, and (iii) *"to ensure that the multidisciplinary research carried out in AMMA is effectively integrated with prediction and decision making activity"*.

We have used the same regions as those proposed by the AMMA project. Their coordinates are given in table 1.2 and they are illustrated in figure 1.9. The regions are: West Africa Monsoon region (WAM), Western Sahel (WSA), Central Sahel (CSA), Eastern Sahel (ESA), Central Sudan (CSU), Central Guinea (CGU), Ouémé catchment (OUE), Niamey area (NIA), and Gourma (GOU). Note that the WAM region was extended to the east and west to be homogenous with respect to other subregions and hence differs from its AMMA definition.

Figure 1.10 indicates some other common appellations of regions over West Africa. The *coastal region* refers to the continental part of West Africa close to the coast. The *Guinean band*—so called as to include Guinea—is a latitudinal band around 10°N. The *Sahel* is the portion of Africa just before the *Sahara* desert. We shall use the terminology *northern Africa* to refer to the northern part of the modelled domain and *equatorial or tropical or central Africa* to refer to the part of the modelled domain located around 0°N and further than 10°E (approximately).

Acronym	Region	Limits	Scale
WAM	WAM domain	5-20°N, 18°W-18°E	Regional
WSA	Western Sahel	13-18°N, 18-10°W	Subregional
CSA	Central Sahel	13-18°N, 10°W-10°E	Subregional
ESA	Eastern Sahel	13-18°N, 10-18°E	Subregional
CSU	Central Sudan	9-13°N, 10°W-10°E	Subregional
CGU	Central Guinea	5-9°N, 10°W-10°E	Subregional
NIA	Niamey area	13-14°N, 1.6-3°E	Mesoscale
OUE	Ouémé catchment	9-10°N, 1.5-2°E	Mesoscale
GOU	Gourma	14.5-17.5°N, 2-1°W	Mesoscale

Table 1.2 – AMMA regions (except the WAM region).

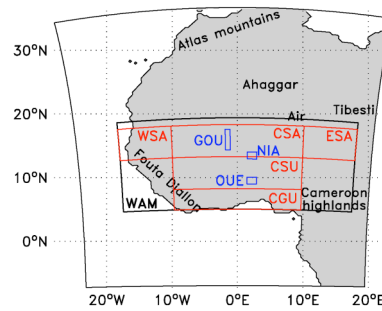


Figure 1.9 – AMMA regions. See text for details.

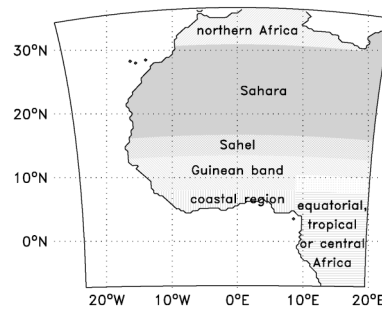


Figure 1.10 – Usual appellations of geographical zones.

Chapter 2

Adaptations to the MAR

Purpose

The model MAR was originally designed to study polar and tropical climates. It has subsequently also been applied to other regions in the world, such as Europe, the Middle East and South America. Each of these regions has its own particularities, which need to be accounted for by applying modifications to the MAR. For example, West Africa is characterised by its complex vegetation which must be represented within the model with as high a degree of accuracy as possible. In the context of this study, use of the model over West Africa has required two adaptations: to enable the use of a new observational dataset to represent both the vegetation extent and temporal cycle, and modification of the calculation of the root distribution, which impacts upon the vegetation transpiration.

Contents

2.1	Vegetation datasets	38
2.1.1	The Normalised Difference Vegetation Index (NDVI)	38
2.1.2	Translation into vegetation properties for the model	40
2.2	The root fraction	43
2.2.1	Modelling of the roots in a land-surface model	44
2.2.2	Results	49
2.3	Conclusion	54

2.1 Vegetation datasets

The description of the vegetation distribution in the model MAR relies upon the Normalised Difference Vegetation Index (NDVI). While vegetation types have been determined by the International Geosphere-Biosphere Program (IGBP), the NDVI index helps to determine how abundant or sparse the vegetation is and how it evolves with time. A new ensemble of NDVI data, which are derived from observational studies, became available in 2005. We chose to use this updated dataset and below provide a comparison of it with respect to the previous version.

2.1.1 The Normalised Difference Vegetation Index (NDVI)

Vegetation indexes are most commonly based upon the radiative properties of vegetation: specifically the red and near-infrared radiances or reflectances of the leaves. Measured by satellites³¹, these quantities are strongly correlated with the photosynthetically active biomass, chlorophyll abundance, and energy absorption. They therefore provide information on the biomass, the leaf area index and the percentage vegetation cover.

The Normalised Difference Vegetation Index (NDVI) is computed from the red and near-infrared channels³². It allows us to provide the MAR with a realistic representation of the vegetation phenology.

The new dataset of the NDVI is available with a 10-day time step for Africa³³ (and bimonthly for other continents) with a 8-km resolution (Tucker *et al.*, 2005). The main differences with the previous NDVI dataset are:

- a different radiometric calibration is employed,
- a pixel by pixel treatment is included to minimize solar zenith angle

³¹ NOAA satellites: 1981 to present, 4-km daily daytime Advanced Very High Resolution Radiometer loaded on board, 5 channels; SPOT-4: 1998 to present, 1-km resolution, 4 channels; MODIS: 2000 to present, 0.25-1 km resolution, 32 channels.

³² $NDVI = (\text{visible channel} - \text{near-infrared channel}) / (\text{visible channel} + \text{near-infrared channel})$.

³³ Available on the Africa Data Dissemination Service website of US AID's Famine Early Warning System (igskmncnwb015.cr.usgs.gov/adds). See Tucker *et al.* (2005) for more information.

effects,

- an atmospheric correction is applied during volcanic stratospheric aerosol periods,
- satellite tracks have been removed,
- data covers a longer period: 1983–present instead of 1983–mid-1994, which therefore allows more recent studies of the West African climate.

The old and new NDVI values are presented on figure 2.1 for August 1987, as is the difference between them. With respect to the old dataset, the new set shows enhanced spatial variability. The index is much lower over bare soils (~25) in the Sahara desert and central Africa (Gabon, Equatorial Guinea) (previously set to a constant of 135). The central part of the Guinean region also shows a slightly less intense index, indicating modification of the vegetation characteristics in that region.

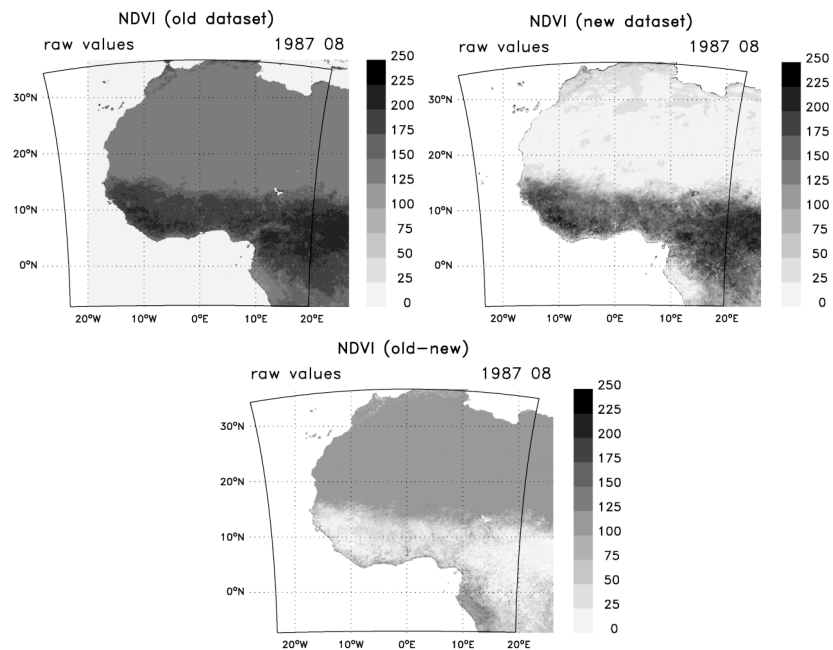


Figure 2.1 – Raw NDVI provided by the old dataset (upper left), the new dataset (upper right), and the difference between both (lower).

2.1.2 Translation into vegetation properties for the model

The NDVI are converted into vegetation properties for the MAR: the fraction of vegetation (or bare soil), the leaf area index, and the green leaf fraction per grid and/or subgrid mesh.

These three variables are of principal importance for the biosphere-atmosphere interactions. The fraction of vegetation determines the partitioning of heat and radiation between the vegetation and the bare soil and the heat and radiation fluxes that are subsequently returned to the atmosphere. The leaf area index provides an indication of the vegetation phenology. The green leaf fraction represents the leaves active in photosynthesis (i.e. those that influence the radiative balance) and evaporation (i.e. dead leaves are empty of chlorophyll and do not evaporate).

Ramel (2005) improved the representation of the plant properties in the MAR by modifying the computation of these three variables. This had to be adapted for the new dataset of NDVI.

The fraction of vegetation is computed using

$$veg. fraction(x, y) = v(x, y) \left(2 - v(x, y) - e^{-2.5 v(x, y)} \right) \quad (2.1)$$

$$where v(x, y) = \frac{NDVI_{max}(x, y) - NDVI_{min}}{NDVI_{max} - NDVI_{min}} \quad (2.2)$$

$NDVI_{max}(x, y)$ is the maximum temporal NDVI observed in the grid mesh. $NDVI_{min}$ and $NDVI_{max}$ are respectively the minimum and maximum global temporal NDVI. Hence the fraction of vegetation is time independent.

The non-linear relationship between the NDVI and the fraction of vegetation allows one to represent vegetation types such as dense herbs and forest which have similar vegetation fractions, but very different NDVI indexes. The coefficient 2.5, which was previously adjusted to get a vegetation fraction in the Sahel close to observations, was kept for the new NDVI dataset as the new vegetation fraction range was similar in magnitude to that used previously.

The green leaf fraction is set to 1 independently of the NDVI. It prevents the green leaf fraction from becoming less than 1 during the growing phase of the vegetation, even if the corresponding NDVI are lower than when plant

maturity is reached. It also assumes the immediate disappearance of dead leaves.

The leaf area index (LAI) is computed on the basis of the NDVI through an empirically determined exponential relationship (Baret and Guyot, 1991) :

$$LAI = -\frac{1}{k} \ln NDVI \quad (2.3)$$

Because the green leaf fraction is set to 1 and the vegetation fraction is determined on an annual basis, all temporal variability in vegetation is therefore brought about by changes in the leaf area index. In the above equation, the coefficient k was set to 0.15 for the old NDVI dataset in order to get a leaf area index of 7-8 for tropical forests, as seen in observations. This coefficient was adjusted for the new NDVI datasets to a new value of 0.20.

Figure 2.2 shows the modelled vegetation properties deduced from the old and new NDVI datasets. We recognize the difference in magnitude of the NDVI for bare grounds and vegetated grounds for both datasets as well as a more precise spatial distribution for the new one.

The fraction of vegetation is similar for both datasets and reaches a maximum over tropical forests (central Africa: Cameroon, western Africa: Liberia). These plots show data for August, a month during which the vegetation is (almost) at its maximum, as visible over the coastal band. The main differences between the datasets also occur in this band and predominantly concern the spatial variability of the coverage. It is reassuring to see that the expected correlation between the NDVI and vegetation fraction is clearly visible, for example in central Africa. Globally, the previous NDVI dataset seems to have slightly overestimated the vegetation fraction.

Both old and new leaf area index values, although showing slightly different spatial distributions, present similar intensities. The LAI reaches 1-2 over the Sahel and progressively increases as one moves southwards. The spatial distribution differences are mainly due to the spatial distribution of the NDVI. The Cameroon mountainous arc is less pronounced and lake

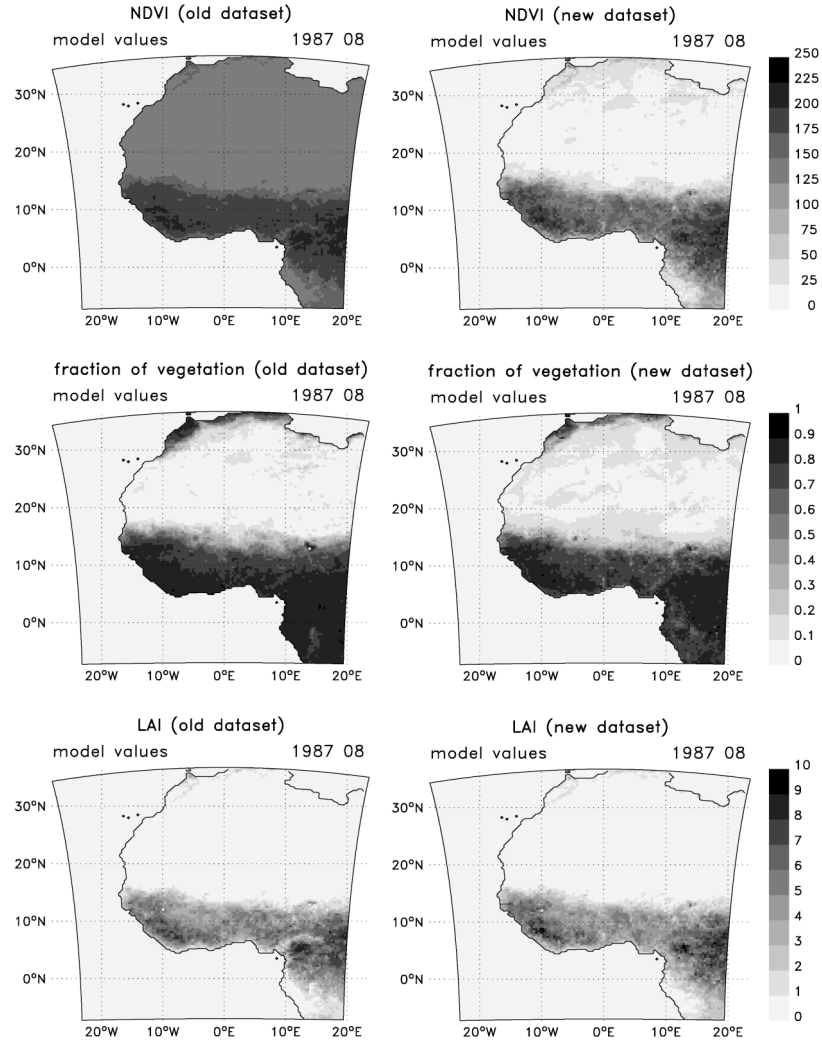


Figure 2.2 – NDVI read by the model (top), fraction of vegetation (middle) and leaf area index (LAI, bottom) computed from the NDVI for August 1987 and the old/new (left/right) datasets.

Volta in Ghana is visible (8°N-0°E). The maximum northward extension of vegetation is the same for both datasets (13-14°N) and the central African forest appears more rounded.

The new dataset of NDVI hence appears to have a number of interesting improvements. Its incorporation into the model is expected to help the representation of vegetation over West Africa both in spatial and temporal dimensions.

2.2 The root fraction

From a biological point of view, the roots of plants ensure the anchorage of the plants in the soil, the absorption of water and mineral substances— from shallow and deep soil layers—, and the accumulation of any reserve materials (Germain and Renard, 1976). From a climatic point of view, the roots constitute the only link between the atmosphere and the ground water. They impact on the surface energy balance (i.e. mainly latent and sensible heat fluxes) through modifications of the vegetation transpiration and on the soil hydrological balance through modifications of the soil wetness.

Each plant type has a particular root profile in the ground. Misrepresenting this is similar to misinterpreting the heterogeneity of the vegetation cover, which also plays an important role. In the absence of precipitation in a vegetated area, variation in soil wetness is caused only by the root water uptake since horizontal water fluxes in the ground and water input from deeper soil layers are usually negligible, and are not represented in land models. For example, Williams and Albertson (2004) studied the effects of a rainy event followed by a drought at both a grass site and a grass+tree site in Botswana. They showed that there was a higher contribution from root water uptake than evapotranspiration at the grass+tree site, and that the soil was significantly drier at this site because the trees, having deeper roots, were able to tap water at greater depths. Zeng *et al.* (1998) showed that the use of observational datasets for the root distribution in the Biosphere–Atmosphere Transfer Scheme (BATS³⁴) improved its

³⁴ The BATS is one of the most widely-used land models.

simulation of the surface water and energy balances when the model was both run in a stand-alone mode and when coupled to a climate model. This was in spite of the fact that their model coarsely represented the ground through three soil layers of which only two were in the root zone.

From the discussion above, it is clear that any parameter influencing the surface energy and water budget must be carefully considered in a climate model. In the model MAR, the roots of the plants are represented in terms of root fractions per layer of soil. Initially conceived for a 4-layer soil system, the initial formulation of the root distribution has presented problems when applied to a 7-layer soil system, which is now the standard configuration of the model. The correction of this formulation is the topic of this section.

2.2.1 Modelling of the roots in a land-surface model

Producing a parameterised root scheme for a model is a non-trivial task. At best, existing land models describe vegetation using only a limited number of classes—for example in the MAR, 3 subclasses for each of grasses, crops, trees—, but observations on roots are typically made on one particular plant species in one particular biotope.

In the SISVAT, the root fraction per layer was initially computed from the fraction of roots in the upper 0.1 m of soil. The roots were assumed to be distributed with depth according to uniform profiles from the surface to 0.1-m deep and from 0.1-m to 1-m deep, one metre in depth being the maximum depth represented (De Ridder and Gallée, 1998). Observational data were adapted from Wilson *et al.* (1987) who provided values for the root fraction in the upper 0.1 m of soil of 0.3 for crops, 0.5 for shrubs and trees, and 0.7 for herbaceous plants. The SISVAT is now used with seven soil layers reaching 2.888 m in depth and subsequently the initial formulation leads to incoherence in the root fraction computation as soon as the soil depth exceeds 1 metre (see table 2.1). More precisely, the calculated root fraction per soil layer exceeds 1 beyond a depth of 1 m indicating that most of the roots are located in the deep ground, contradicting both mathematics and observations. Indeed, most plants have maximum root quantity in the first metre of soil (Jackson *et al.*, 1996; see figure 2.3).

7-layer soil				4-layer soil			
depth (m)	crops	grasses	trees	depth (m)	crops	grasses	trees
0.002	0.002	0.001	0.001				
0.008	0.005	0.002	0.003				
0.028	0.02	0.01	0.01	0.020	0.02	0.01	0.01
0.088	0.05	0.02	0.03	0.080	0.05	0.02	0.03
0.288	0.16	0.07	0.11	0.280	0.16	0.07	0.11
0.888	0.47	0.20	0.33	1.000	0.78	0.91	0.84
2.888	1.78	1.33	1.56				
Σ :	2.47	1.63	2.05	Σ :	1.00	1.00	1.00

Table 2.1 – Root fraction as a function of depth, computed with the initial formulation, for each different SVAT vegetation type (1-3 = crops, 4-6 = grasses, 7-12 = trees). Sum errors due to rounded numbers.

In order to correctly model the root distribution, we use the non-linear function proposed by Gale and Grigal (1987). The cumulative root fraction Y from the surface to a depth d (in cm) is given by

$$Y = 1 - \beta^d \xrightarrow{\text{discretisation}} \Delta Y = \beta^{d_1} - \beta^{d_2} \quad (2.4)$$

where the regression coefficient β is considered as a measure of the vertical root distribution. High values of β (~ 0.97) are associated with a larger quantity of deeper roots than low values of β (~ 0.92). ΔY represents the root fraction between depths d_1 and d_2 . Figure 2.3 shows Y for different plant types.

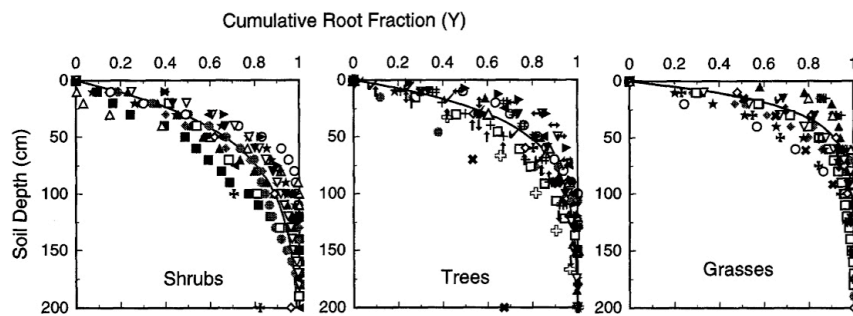


Figure 2.3 – Cumulative distribution of grass, tree, and shrub roots as a function of soil depth. Symbols refer to different plant species. [Source: Jackson *et al.*, 1996]

The original study of Gale and Grigal (1987) was illustrated using a northern tree species only and as such is not directly applicable to modelling the diverse range of vegetation found over West Africa. Therefore, the coefficient β is computed on the basis of Jackson *et al.* (1996). These authors compiled 250 studies of root distributions, subdivided results into 11 biomes and finally fitted the coefficient β to the data for each biome. This allows us to determine the coefficient β for each of the SVAT vegetation types over West Africa³⁵ (see table 2.2).

Figure 2.4 shows the root fractions per layer and per SVAT vegetation type computed in this new way. All vegetation types show maximum root fraction between depths of 0.1 and 1 m, which corresponds well to observations. At greater depths, the root fraction is far lower. Certainly with respect to the initial formulation, the root representation is vastly improved. We can look a little closer at the differences between specific vegetation types. Low grasses have shallow roots and as such only benefit from water in the shallow layers (for example water brought by short periods of precipitation). High grasses on the contrary present the greatest deep root fraction. They are present in the Sahelian region where vegetation needs to tap water deeper during the dry season when the shallow soil layers have dried out. Soils are usually shallow in tropical regions, therefore plants such as trees, which are normally able to send roots deep into the ground are limited. A special case is that of broadleaf trees, which show a root profile whereby most roots are within the top 1 m of the soil.

An unavoidable repercussion of adopting this approach is that the definition of the soil layers has an impact on the root fraction formulation. The Jackson's database and Gale and Grigal formulation stipulate that roots will not be present much further than 2 m in depth. If a soil layer originating from a depth above 2 m extends to a depth well below 2 metres, then this will result in an overestimation of the root fraction at that lower depth. Therefore, the user must apply care when defining the soil layers.

³⁵ We advise the following standard configuration for any other region: $\beta = 0.961$ for crops (1–2–3) on the basis of biome 2, $\beta = 0.943$ for grasses (4–5–6) on the basis of biome 7, $\beta = 0.966$ for broadleaf trees (7–8–9) on the basis of biome 6 and $\beta = 0.976$ for needleleaf trees (10–11–12) on the basis of biome 5.

Jackson <i>et al.</i> 's biomes		β	root biomass in 0-30 cm (%)
1	boreal forest	0.943	83
2	crops	0.961	70
3	desert	0.975	53
4	sclerophyllous shrubs	0.964	67
5	temperate coniferous forest	0.976	52
6	temperate deciduous forest	0.966	65
7	temperate grassland	0.943	83
8	tropical deciduous forest	0.961	70
9	tropical evergreen forest	0.962	69
10	tropical grassland savanna	0.972	57
11	tundra	0.914	93

SVAT vegetation type		β	corresponding biome(s)
1	crops low (millet)	0.961	2
2	crops medium (millet)	0.961	2
3	crops high (millet)	0.961	2
4	grass low (sparse tiger bush)	0.943	7
5	grass medium (dense tiger bush)	0.964	4
6	grass high (dense tiger bush)	0.972	10
7	broadleaf trees low (fallow)	0.968	4+10
8	broadleaf trees medium	0.962	8+9
9	broadleaf trees high	0.962	8+9
10	needleleaf trees low*	0.971	5+6
11	needleleaf trees medium*	0.976	5
12	needleleaf trees high*	0.976	5

Table 2.2 – Upper: coefficients β and root biomass in the upper 30 cm of soil (Jackson *et al.*, 1996); lower: coefficients β for the SVAT vegetation types and corresponding biome number used for the computation. An asterisk denotes a nonexistent vegetation type over West Africa.

Chapter 2: Adaptations to the MAR

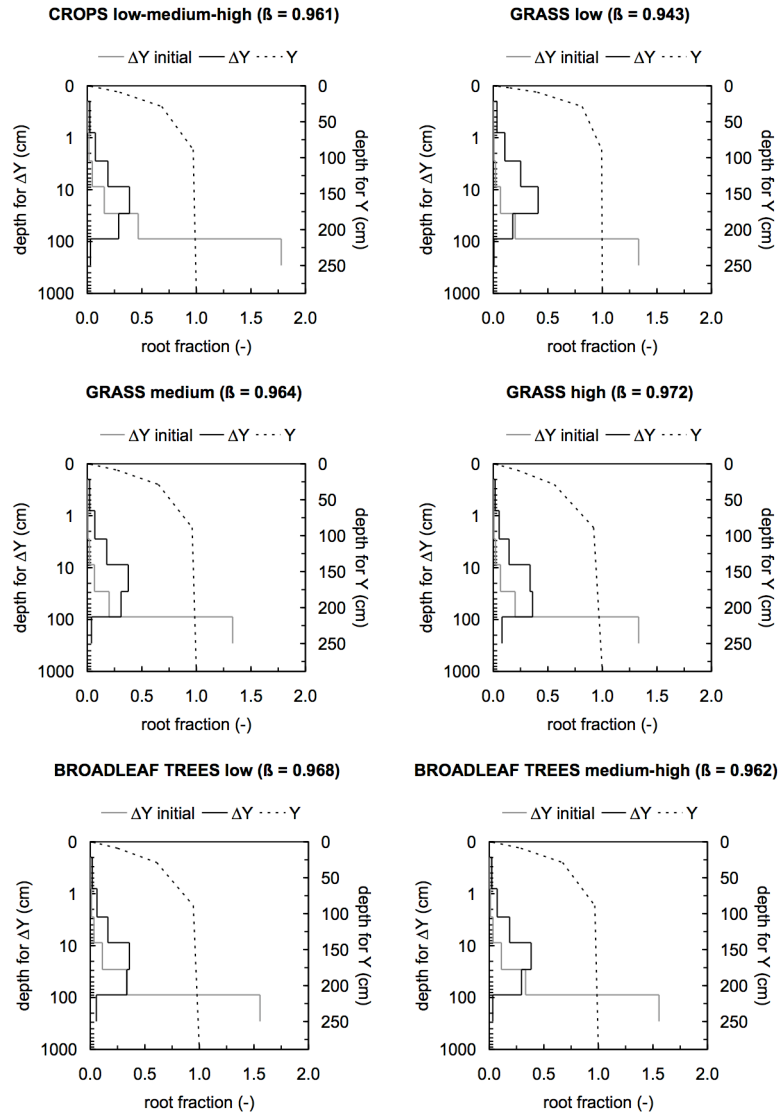


Figure 2.4 – Root distribution per soil layer and SVAT vegetation type as computed with the Gale and Grigal's formulation (solid curves). Dashed curves show the cumulative root fraction. Gray curves indicate the root distribution computed with the initial formulation.

It could be argued that an additional factor has to be taken into account to compute the root fraction: the maximum rooting depth. Indeed, plants from low rainfall environments show the deepest rooting habits (Canadell *et al.*, 1996). Some trees in deserts can tap water as deep as 50 m—roots of a tree in the dry savanna of the central Kalahari were measured to reach 68 m in depth but thought to actually reach 140 m (water table depth). We decided not to explicitly use this additional parameter for two reasons. Firstly, the formulation of Gale and Grigal (1987) already represents the deep root fraction, as the coefficient β is determined from observations and thus takes into account the maximum rooting depth of the plants. In the model, roots can not go deeper than 2.888 m and deep roots are represented by the value of the root fraction in the deepest soil layer (0.888–2.888 m deep) and no further. Secondly, observations have shown that the shallow roots seem to absorb most of the water that the plants need, but the exact proportion of water absorbed by each type of roots—fixative (short, anchorage, storage of shallow water), fibrous (long, take advantage of the weakest water supply), vertical (long, deep water contact)—is still unknown (Prof. X. Draye, UCL, personal communication). Another noteworthy point is that the water extraction by roots is limited by the water availability. As the soil water content decreases, water becomes more strongly bound to the soil matrix (which also depends on the soil texture) and becomes more difficult to extract. As such, a large proportion of deep roots does not automatically indicate increased water pumping, especially during the rainy season where the plants preferably pump shallow water. We have observed this in the model by comparing the soil moisture of the deepest layer computed from our results that overestimated (new formulation doubled) and underestimated (old formulation) the root fraction in the lowest layer respectively. We found that there was no significant difference in the calculated soil moisture of the deepest layer between the two runs.

2.2.2 Results

The root fraction is used in the MAR to compute the water flux through the plants i.e. the sum of the transpiration fluxes from each soil layer. Each

transpiration flux is proportional to the water potential difference between the leaves and the soil layer and depends on the root–soil resistance and the internal plant resistance. The higher the resistances are, the smaller the fluxes are. It is to be expected that the modifications to the formulation of the root fraction will impact upon calculated surface fluxes, such as the evapotranspiration and the latent heat.

We present here the results of two simulations for the period July-August-September 1987³⁶. The simulation Fr0 was performed with the initial formulation of root fraction. The simulations Fro and Frp were performed with the new formulation, as presented above, with the simulation Frp having a doubled root fraction in every soil layer.

Figure 2.5 shows the differences between the new formulation and the initial one for the evapotranspiration, latent heat flux, and soil moisture in the 0-28.8 cm soil layer for August. Other months present similar patterns (not shown). At a monthly scale, the new formulation leads to an increase in evapotranspiration of around +0.2 mm/day, in latent heat of around +8 W/m² and a drying of the shallow soil layers of around -0.02 m³/m³. Differences are rather weak in amplitude as the averaging is performed on a monthly scale, where as these variables show variability on time scales of typically a few days, if not hours. As one might expect, the results for the doubled root fraction experiment corroborates the results described above: the same changes are observed but more stronger (+0.4 mm/day, +16 W/m² and -0.03 m³/m³ respectively).

Figure 2.6 presents the same results, but from a different perspective. A spatial average is performed over two AMMA regions: the Ouémé basin (black curves) located in the north of Benin (9-10°N, 1.6-3°E) and the Niamey area (grey curves) located in Niger (13-14°N, 1.6-3°E) (see chapter 1, section 1.4). We can make statements similar to those above: for example in the Ouémé basin, evapotranspiration and latent heat flux are increased (+0.2 mm/day and +7 W/m² resp. in average for the three months) with the new formulation while the soil moisture until 30 cm deep is decreased (-0.02 m³/m³). The change in latent heat leads to a change in sensible heat

³⁶ Both started on the 1st January 1986.

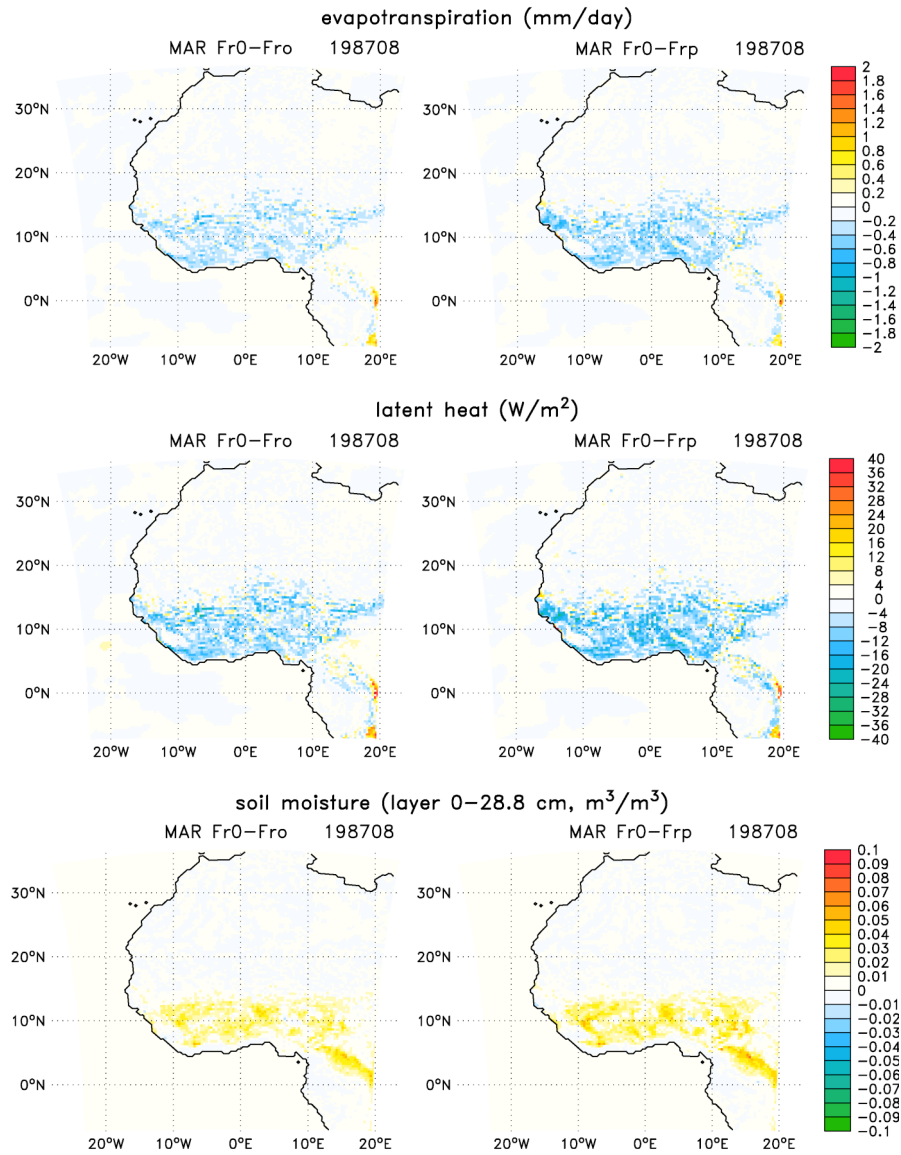


Figure 2.5 – Differences in evapotranspiration, latent heat and soil moisture between the new formulation (left: normal, right: doubled) and the initial formulation. Monthly averages presented as daily values.

Chapter 2: Adaptations to the MAR

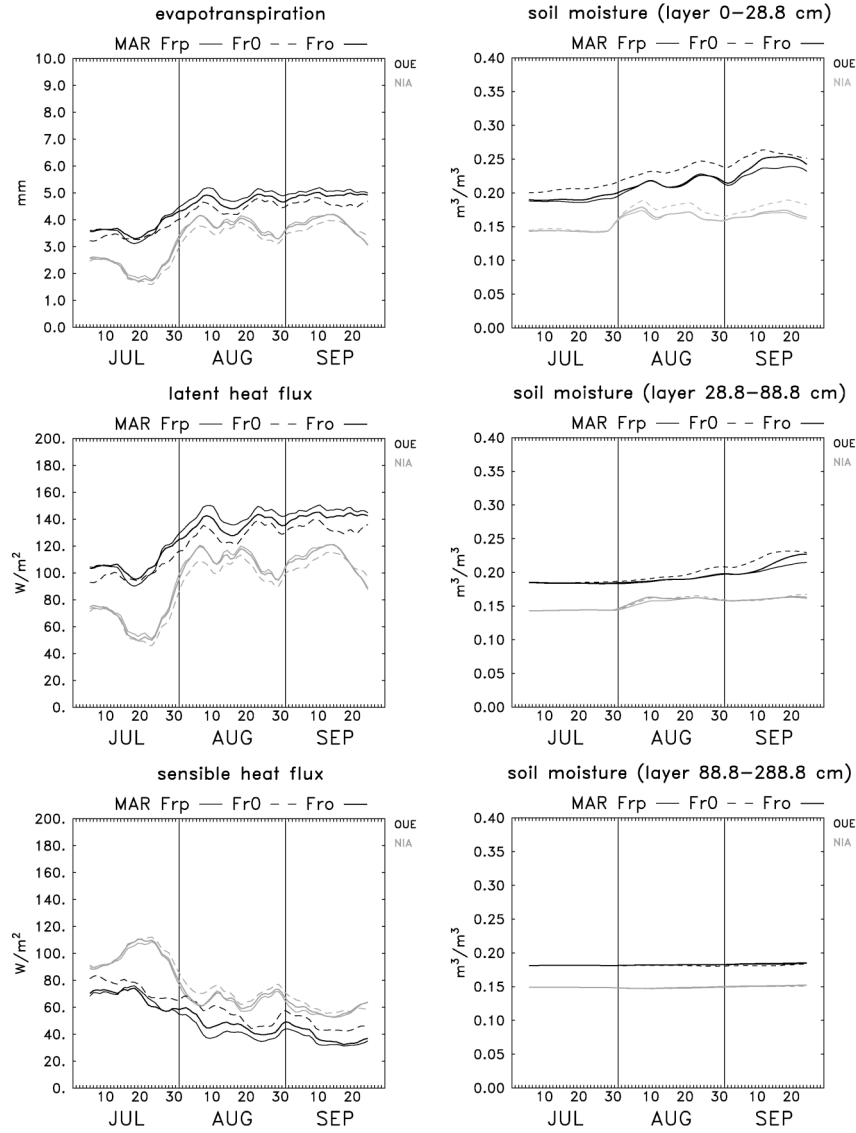


Figure 2.6 – Evapotranspiration, heat fluxes and soil moisture over the Ouémé basin (black) and the Niamey area (grey) with the new formulation (thick solid curves), the new formulation and doubled root fraction (thin solid curves) and the initial formulation (dashed curves). Running mean with a 10-day period.

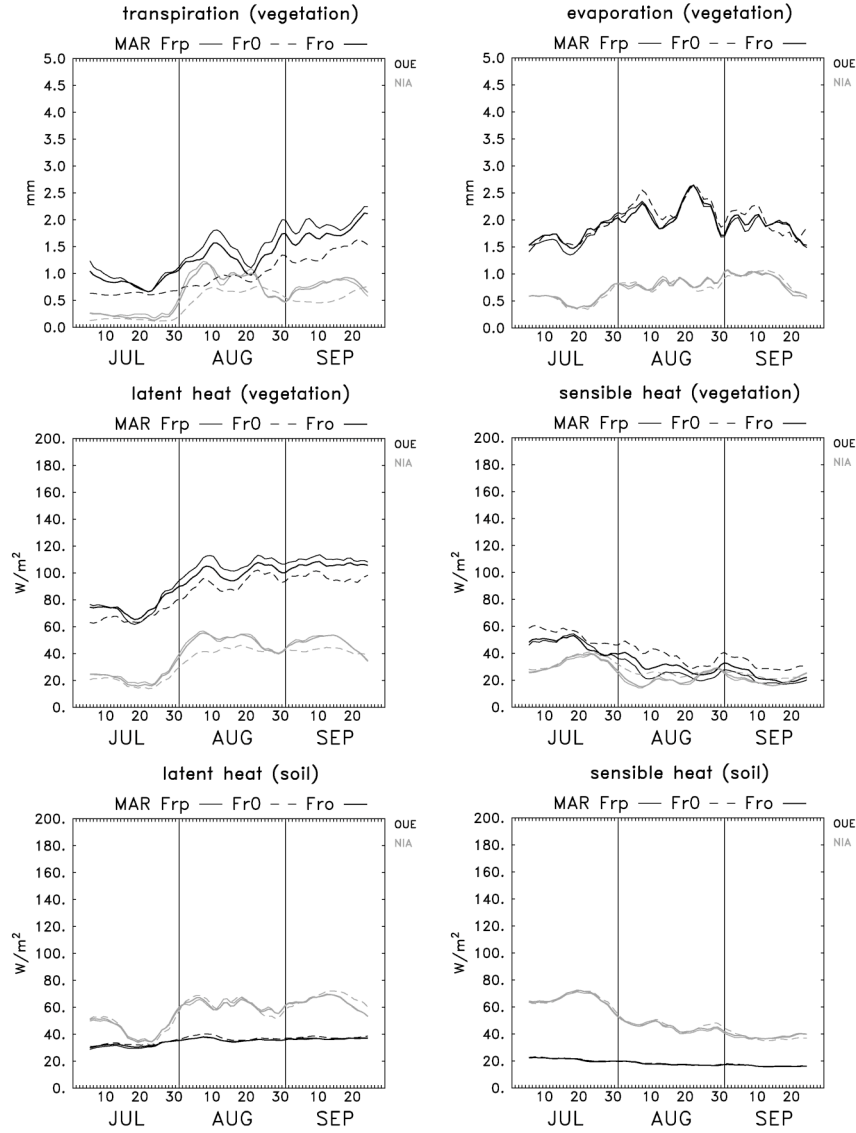


Figure 2.7 – Surface water and heat fluxes for the Ouémé basin (black) and the Niamey region (grey) with the new formulation (thick solid curves), the new formulation and doubled root fraction (thin solid curves) and the initial formulation (dashed curves). Running mean with a 10-day period.

(-8 W/m^2). It is again clear that deeper soil layers are less affected by the root intake of water, as the root fraction is small at depth and the water is more strongly associated with the soil matrix. The Niamey region presents similar changes (although with a slightly smaller amplitude). This can be reconciled by considering that this region has less vegetation, it is located in the Sahel, to the north of the Ouémé basin.

A separation of the vegetation and soil contributions to the evapotranspiration and heat fluxes for the three different runs is presented in figure 2.7. The vegetation evaporation, i.e. evaporation of rainfall intercepted by the vegetation, is very similar for each of the runs (-0.1 mm/day in average over the three months for the Ouémé basin). The variation that is observed is due to rainfall variability. Comparing run Fro to Fr0 we see that the vegetation transpiration is increased by $\sim 0.4 \text{ mm/day}$. The latent and sensible heat fluxes from the soil vary by -1 W/m^2 and 0 W/m^2 while the latent and sensible heat fluxes from the vegetation vary by $+8 \text{ W/m}^2$ and -8 W/m^2 respectively. The evapotranspiration is the sum of the vegetation evaporation, the vegetation transpiration and the soil evaporation (here unavailable), giving $+0.2 \approx -0.1 + 0.4 \text{ mm/day}$. The latent (sensible) heat flux is the sum of the soil and vegetation contributions; which gives $+7 = -1 + 8 \text{ W/m}^2$ ($-8 = 0 + 8 \text{ W/m}^2$). Similar relationships can be computed for the doubled root fraction experiment. The net radiation balance is met to within $\pm 1 \text{ W/m}^2$ and the equation for the evapotranspiration to $\pm 1 \text{ mm/day}$.

2.3 Conclusion

In this chapter we have presented two modifications made to the model MAR in order to improve its representation of the biosphere over the West Africa region.

We have described a new dataset of the Normalised Difference Vegetation Index (NDVI) that has been used to represent some vegetation characteristics in the model. The new dataset is superior in quality to the previous version and in addition now extends to cover a period up to the present day (instead of mid-1994). As a consequence of adopting the new

NDVI dataset, the spatial and temporal properties of the representation of vegetation by the model have been improved. The fraction of vegetation as well as the leaf area index are represented with more detail and the northward extension of the vegetation is maintained with a leaf area index matching those from observations—this is encouraging, as the empirical relationship of Baret and Guyot (1991) was calibrated to get realistic values over tropical forests only.

We have also corrected and improved the computation of the root fraction within each soil layer. The newly-implemented formulation of Gale and Grigal (1987) has enabled the computation of the root distribution at any depth in the ground. Its application to each of the model classes of vegetation has been made according to Jackson *et al.* (1996). Results have shown coherent and realistic plant root structure that has impacted upon the heat fluxes computed by the model. Indeed, this was to be expected as the plant roots constitute the only link between the soil water and the atmosphere water in land-surface models. Fluxes such as latent heat and evapotranspiration have increased—but no significant impact was found for rainfall—, which is expected to help to improve the representation of rainfall over West Africa.

Chapter 3

Internal variability

Purpose

Running regional climate models requires powerful computing resources both in terms of data generation and storage. Experiments have to be carefully planned in order to be achievable within reasonable time scales. Hence, when performing sensitivity studies, one often runs only a limited number of simulations (typically two to four), analyses the difference between them, and relies upon experience and common sense to assess their significance. However, there is an internal variability inherent in such models that has to be taken into account in such simulations. We here investigate the internal variability displayed by the model MAR. The intention of this work is to provide a structured framework with which to assess the variability, thereby enabling the results from sensitivity experiments to be interpreted with added insight and confidence.

Contents

3.1	Introduction.....	58
3.2	Experimental setup	60
3.3	Internal variability and temporal scales	62
3.4	Internal variability and spatial scales	67
3.5	Internal variability and both temporal and spatial scales	68
3.6	Conclusion.....	71

3.1 Introduction

Regional climate models can be used in experiments designed to isolate the effects of different boundary influences in a setting that is more realistic than typical global circulation model sensitivity experiments because of the prescription of observed conditions at the boundaries. Since regional climate models are relatively expensive to run, sensitivity experiments are usually limited to two runs, with differing boundary conditions, internal processes or both. Conclusions are drawn from the differences between the two runs and if further runs are to be carried out, resources are usually put into another study: for example, a different year or use of different parameterisations. The statistical significance of the result is taken for granted, and general experience is used to decide which diagnostics are reliable.

Regional models can be used as scale filters between large-scale climate variations and the impact scale. In this application they become a more interdisciplinary tool, shared by different communities, and it becomes more important to establish some guidance to the interpretation of their results. The purpose of this chapter is to give an objective guide as to which diagnostics are reliable, and how large a signal needs to be before conclusions can be drawn from it with confidence. This type of information is equally applicable to assessing confidence in sensitivity experiments and to providing a reasonable range of inputs to impact models.

We present a pair of experiments over West Africa which differ only in their internal variability. All boundary conditions, surface conditions and parameterisations are exactly the same for the two runs. Only the initial condition is changed. This gives rise to differences between the two experiments that are independent of all the factors usually considered in sensitivity experiments. After an initial spin-up time (three months, which are not considered in the following analysis) this arbitrary change in initial conditions is forgotten and the differences between the two experiments are small. They represent the noise level, an amplitude for random unpredictable perturbations which is referred to as *internal variability*. This noise level

depends on the variable and the time period under consideration, and a range of diagnostics will be presented here that are of interest to impact modellers. For sensitivity studies the results give a measure of significance that is necessary if inferences are to be drawn about mechanisms. For smaller scale impact models that rely on these regional simulations, the results provide a lower limit for meaningful changes in input, or alternatively a measure of the range of inputs that should be provided for an ensemble study.

Not many studies about internal variability in regional climate models have been conducted so far. Giorgi and Bi (2000) performed 3-month long simulations for a year over China with the RCM RegCM. Random time- and space-varying perturbations were applied to the initial and/or lateral boundary conditions (atmospheric variables only) over selected parts of the modelled domain such as the buffer area (boundary zone where the large-scale driving fields are progressively transmitted to the RCM domain), the inside domain only or some specific regions. They concluded that the model response was barely sensitive to the location of the perturbation, its origin and its magnitude. This result was further confirmed by Caya and Biner (2004) using the Canadian RCM CRCM over the eastern half of North America, and especially by Rinke *et al.* (2004) using HIRHAM over the pan-Arctic region. The latter authors tested different driving fields (NCEP or ERA), perturbed lateral boundary conditions similarly to Giorgi and Bi (2000) and initial conditions—simulation started one day later as Caya and Biner (2004) also examined.

Christensen *et al.* (2001) used a different approach. They simulated 6 times the same year with identical lateral boundary conditions and with the soil variables taken from the model results year after year. Hence, they tested the internal variability generated by a difference in surface conditions in two RCMs, HIRHAM and PROMES, over a European domain centred in the Mediterranean Sea. They obtained similar internal variability magnitude in both models but with different geographical distribution. Hydrological variables show variability similar to that found in GCMs. They highlighted the importance of knowing more about the properties of the internal variability scales, especially with resolution, but also temporally and

vertically. Weisse *et al.* (2000) gives one more proof that the internal variability of a model must be known before any sensitivity experiment is carried out. Investigating two computations of the sea-surface roughness in HIRHAM coupled to a wave model and an ocean model over the Northern Atlantic Ocean, they found out that most of the differences obtained by their results were inside the internal variability of their model. The differences in results were significant only for a few periods and this was linked with the atmospheric circulations. This was also found by all previous authors. The internal variability shows the strongest magnitude in the most "meteorologically active" period of the year: winter for the Arctic, summer for China, Europe or North America.

The experimental design is described in section 3.2. Results are shown and analysed in sections 3.3 to 3.5. Conclusions are given in section 3.6. The content of this work is *in press* in *Climate Dynamics* (see Vanvyve *et al.*, 2007).

3.2 Experimental setup

Two simulations were performed during the year 1986 (1st January to 31st December). Within the context of the dry conditions experienced during the 70-80's, this year is considered as neither wet nor dry—see Nicholson (1993), in which the wet year 1987 and the dry year 1988 are compared.

	EXP1	EXP2
period of simulation	01/01 → 31/12/1986	01/01 → 31/12/1986
boundary forcings	every 6 hours, ERA40	every 6 hours, ERA40
dynamic field initialisation time	01/01/1986 (18h UT)	01/01/1987 (18h UT)
soil moisture initialisation field	ERA40	ERA40
spin-up	first 3 months	first 3 months
NDVI for LAI	as observed	as observed

Table 3.1 – Additional parameters used for the simulations.

Both simulations use the standard characteristics described in chapter 1, section 1.4. In addition, some specific parameters are given in table 3.1. The

first run (EXP1) is a standard run. The second run (EXP2) is identical to the first in every respect except the initial conditions, i.e. the model state (including the boundaries) at the first time step. In order to introduce an initial perturbation, the zonal wind, meridional wind, potential temperature, pressure, specific humidity and surface temperature were initialised with the values of the 1st of January 1987 (18:00 UT). This effective initial perturbation is shown in figure 3.1. It represents a large perturbation and for our purposes is completely arbitrary. As indicated in table 3.1, the spin-up period is taken as three months. This period refers to the time during which the initial perturbation influences the results to an extent that prevents quantification of internal variability.

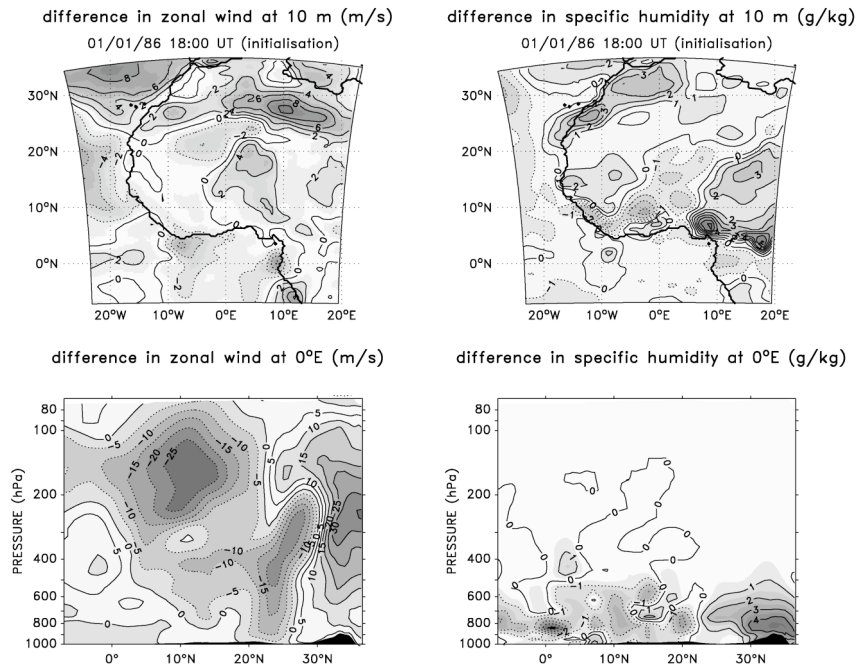


Figure 3.1 – Differences between the standard run EXP1 and the initially-perturbed run EXP2, at the first time step, for the zonal wind (left) and specific humidity (right), at 10 m in height (top) and at 0°E (bottom). Dotted curves denote negative values.

Chapter 3: Internal variability

Figure 3.2 shows the development of the area averaged kinetic energy of the difference field EXP1-EXP2 in the mid-troposphere (500 hPa) as a consequence of the initial perturbation. After the initial shock, the value diminishes over a period of a few months (January until March) and reaches an equilibrium level from April. It is the size of this "equilibrium" signal that interests us here for variables relevant to process and impact studies. This is investigated in the next sections.

To avoid the need to refer back to a previous chapter, figure 3.3 shows the modelled area, some geographical features and some regions of interest. These are the West Africa monsoon region (WAM), the Western/Central/Eastern Sahel (WSA/CSA/ESA), the Central Sudan/Guinea (CSU/CGU), the Ouémé basin (OUE), the Niamey area (NIA) and Gourma (GOU). More details regarding the modelled regions may be found in chapter 1, section 1.4.

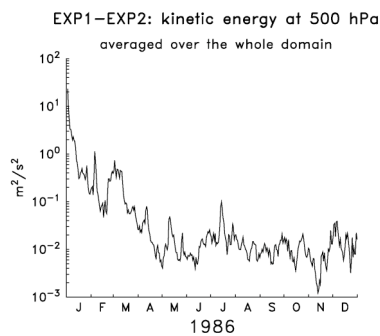


Figure 3.2 – Kinetic energy of the difference field EXP1-EXP2 at 500 hPa, averaged over the whole domain (daily timescale).

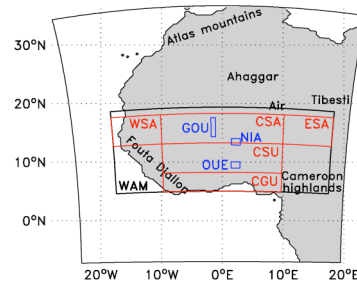


Figure 3.3 – Modelled domain, geographical features and regions of interest for the further analysis.

3.3 Internal variability and temporal scales

The magnitude of the internal variability of the model is first established by looking at the difference EXP1-EXP2 for a selection of variables.

Time averages for this difference are calculated for various time intervals, for the rainfall, the air temperature (T) at 10 metres and the wind

speed in the boundary layer (BL). The time intervals considered are daily, 5-day, 10-day, monthly, "seasonal" (3-month averages) and "annual" (April to September average). Table 3.2 summarises the maximum absolute differences observed across the domain. This absolute difference decreases as the averaging period gets longer. At a daily time interval, the rainfall can show variability up to 27 mm/day, but usually typical values of the daily difference are around 10 mm/day.

It should be remembered that table 3.2 shows maximum values. These values represent a worst case for uncertainty for a given time interval if one is interested in the smallest scale present in the model (remember the resolution is 40 km). For example, if a sensitivity experiment yields a 5-day mean change in rainfall of 8 mm/day at a given location, this result may have no statistical significance. If the same value appears in the monthly mean signal it is probably safe to attach a physical interpretation to it and examine its effect in an impact study. The same general principle applies to the values for temperature and wind also shown in table 3.2.

	daily	5-day	10-day	monthly	seasonal	annual
rain (mm/day)	26.5	8.3	4.3	1.6	0.9	0.5
10-m air temperature (°C)	2.9	0.8	0.7	0.5	0.2	0.1
wind in the BL (m/s)	4.4	1.5	0.8	0.5	0.4	0.3

Table 3.2 – Absolute maximum EXP1-EXP2 differences for each considered variable and time interval, across the whole domain, from April to September.

In order to illustrate the spatial characteristics of these anomalies, the 14th of July has been chosen as a typical day, and the difference EXP1-EXP2 is shown in figure 3.4 along with results averaged over longer periods that span this day. The boundary-layer wind, temperature and precipitation are shown and the structures seen are representative of the rainy season in general. Although the origin of the difference signal can be considered random, the signal itself shows clear synoptic-scale organisation associated with dynamical balances for wind and temperature (if both runs are dynamically balanced at the large scale then so must the difference between

them). The precipitation anomalies are at a smaller scale, consistent with their convective generation, and are co-located with areas of maximum rainfall in the individual runs. The typical annual mean ratio between convective and stratiform rainfall in the model is 87 % versus 13 % over West Africa, and separate diagnostics confirm that most of the differences in rainfall seen here are of convective origin. Longer averaging leads to a universal reduction in the rainfall signal (note the changing colour scales) as small displacements of the grainy signal lead to cancellations. Owing to their greater coherence at larger scales and slower temporal evolution, the signal for the dynamical variables diminishes less rapidly as the averaging period increases.

To quantify these effects and their geographical distribution over the whole of the rainy season, the standard deviation (SD) was calculated according to:

$$SD = \sqrt{\left\{ \left[(\text{EXP1} - \text{EXP2})_{\Delta t} - \overline{(\text{EXP1} - \text{EXP2})}_{\text{Apr-Sep}} \right]^2 \right\}_{\text{Apr-Sep}}}$$

where EXP1-EXP2 is the anomaly signal, Δt is the averaging period considered (Δt = daily, 5-day, 10-day, monthly, seasonal) and the overbar denotes a temporal average over the period indicated by the subscript. In each case the calculation is made relative to the April to September mean.

Figure 3.5 shows the standard deviation of the wind speed in the boundary layer, the temperature at 10 metres and rainfall for the five different timescales. The signal shows large-scale structure and in all cases decreases steadily in magnitude with increasing timescale (for these plots the colour scales have been kept constant so they can be directly compared). There is a great deal of variance in the wind off the coast near Fouta Djallon. There are corresponding variations in temperature, which remain strong out to the 10-day timescale despite the fixed sea-surface temperatures. Variations in temperature are strongest over the southern part of the continental surface, where the strongest rainfall variations are located. Variations in rainfall appear to mimic the actual rainfall distribution and can probably be understood as essentially random variations in convection. The spatial distribution does not change with increasing timescale. It should be remembered that these variations are random in origin even though they

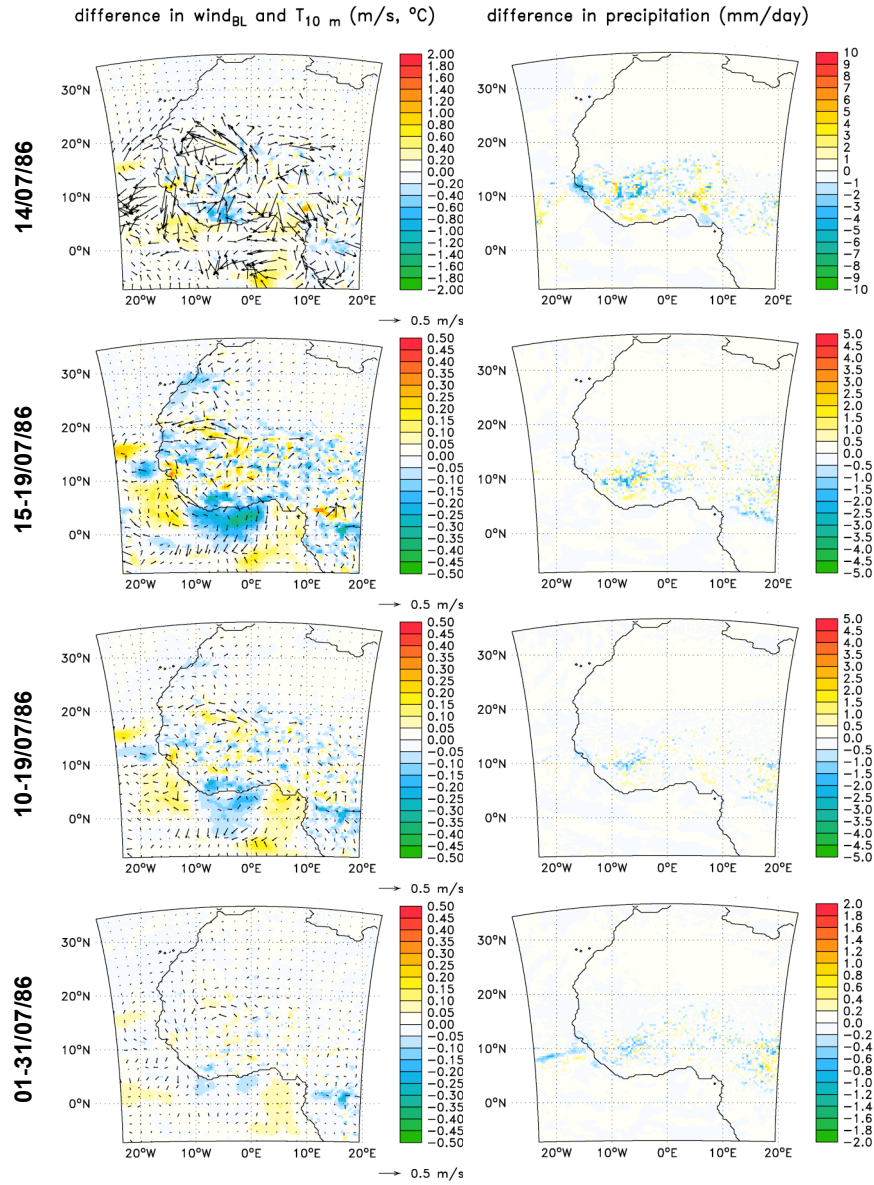


Figure 3.4 – Differences between EXP1 and EXP2 for the wind in the boundary layer (BL) [left, arrows], the temperature (T) at 10 metres [left, colours], and the precipitation [right] at daily, 5-day, 10-day and monthly timescales. Note the changing colour scales.

Chapter 3: Internal variability

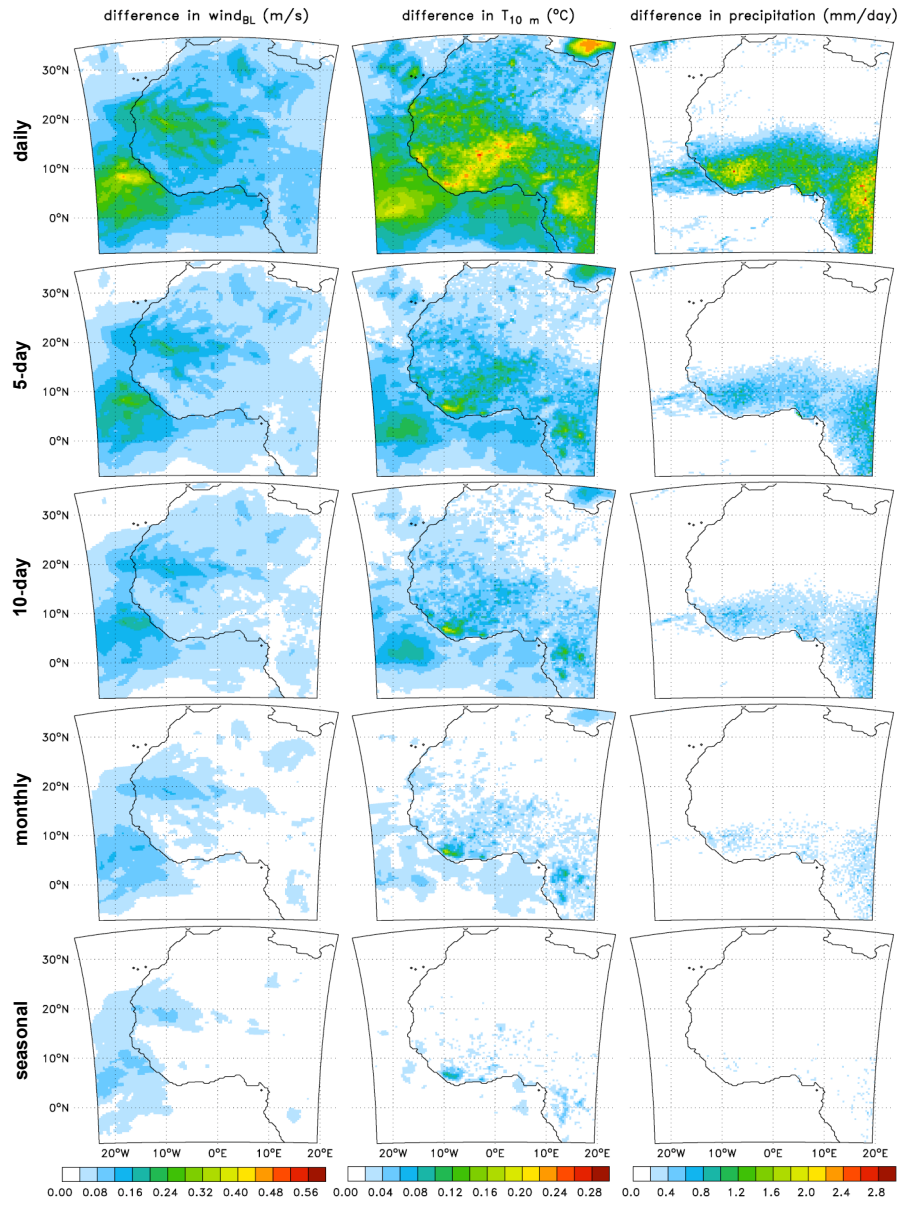


Figure 3.5 – Standard deviations of the anomaly field EXP1-EXP2, for the boundary-layer (BL) wind [left], the 10-m air temperature (T) [middle] and rainfall [right] at daily, 5-day, 10-day, monthly and seasonal timescales. Each colour bar corresponds to its column.

display some spatial structure. They are different from the systematic errors also present in the model and must be considered separately when assigning confidence measures to the model output. Like many other models, the MAR has a tendency to produce too many weak rainfall events and not enough strong ones (see Ramel, 2005). If this systematic error was not present, the random error measured here may indeed be different, but it would still be a source of uncertainty.

3.4 Internal variability and spatial scales

If the internal variability in the model is truly random, we can reasonably expect the signal to decrease when considering larger spatial scales in exactly the same way as when considering longer timescales. The fact that the signal we are interested in has various physical and dynamical origins means that this is not necessarily so, and it is interesting to consider separately the question of spatial scales. Here we examine rainfall and temperature on daily timescales, using the standard deviation as defined above for the rainy season. To measure the variability associated with spatial scales larger than the model grid size, the model resolution was progressively degraded by spatial averaging prior to calculating the standard deviation. The spatial degradation was performed as follows (see figure 3.6). The coordinates of each grid box centre in the new resolution were chosen to match the coordinates of the previous grid box centre (i.e. in the 40-km resolution). For the new enlarged grid box, the mean was computed on basis of a 80 by 80, 160 by 160 or 320 by 320 km area, as the sum of:

1. the n^2 full grid boxes forming a square centred on the 40-km initial grid box (where $n = 1, 3$ or 7 for a 80, 160 or 320-km resolution respectively),
2. half the m adjacent grid boxes to that new square (where $m = 4, 12$ or 28 respectively), and
3. a quarter of the four remaining corners.

The sum was then divided by the number of full grid boxes equivalent to the new area (4, 16 or 64 for a 80, 160 or 320-km resolution respectively).

Results are shown in figure 3.7 for the model resolution of 40 km (as

before) and the degraded resolutions of 80, 160 and 320 km. Unsurprisingly, the precipitation signal decreases uniformly without any obvious change in its spatial distribution. The temperature signal also decreases, but it decreases more over the continent than over the sea. The structures associated with temperature variance on the daily timescale are more spatially coherent on large scales over the sea than over the continent. We may conclude that over the continent the temperature variance is driven by changes in convection, whereas over the sea it results from propagating synoptic scale systems, and the convection signal is less influential.

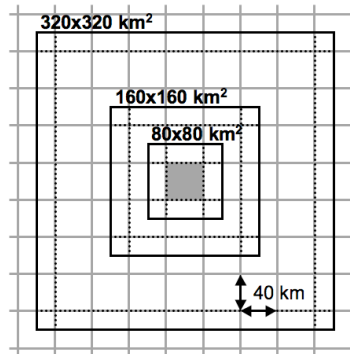


Figure 3.6 – Sketch of the spatial degradation process. The greyed grid shows the original grid, the central initial grid box is highlighted in grey. The dotted lines show limits of full/half/quarter of grid boxes.

3.5 Internal variability and both temporal and spatial scales

We have so far investigated temporal and spatial scales separately. To bring together information about internal variability on different spatial and temporal scales in a practical way, we shall present the results as averaged per region.

Firstly, in table 3.3 are shown the maximum difference values of standard diagnostic and field data averaged over nine regions. In table 3.4 the standard deviation values for the same parameters averaged over the same regions are shown. This is done for the five standard temporal categories,

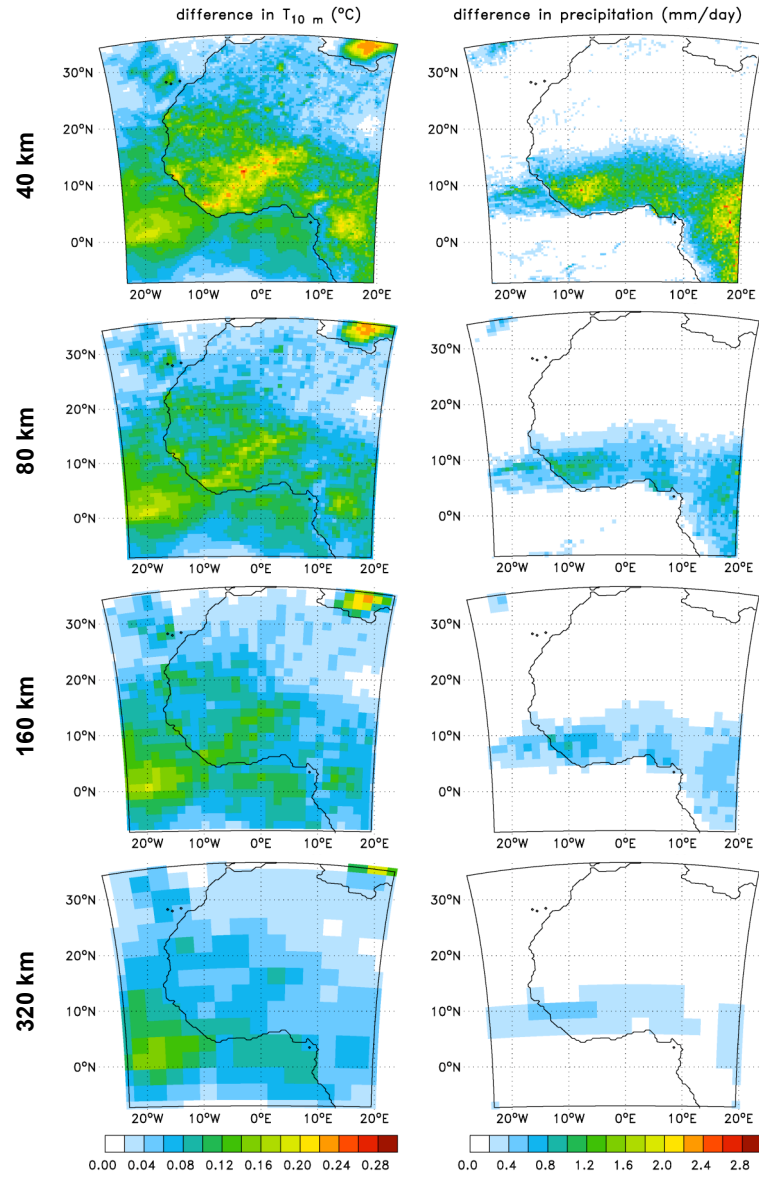


Figure 3.7 – Spatially degraded standard deviation of the anomaly field EXP1-EXP2 for the 10-m air temperature and rainfall at a daily timescale. The spatial resolution was progressively degraded from 40 km to 80, 160 and 320 km before the calculation of the standard deviation.

Chapter 3: Internal variability

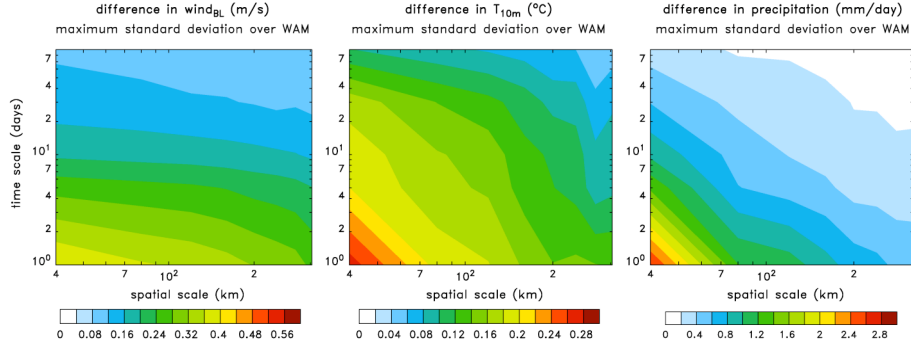


Figure 3.8 – Maximum standard deviations of the anomaly field EXP1-EXP2 over the WAM subregion, shown in the space-time domain for the wind in the boundary layer (BL), the 10-m air temperature (T) and the precipitation.

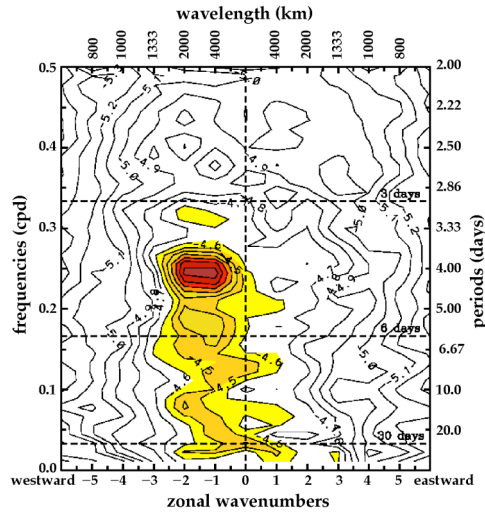


Figure 3.9 – Power spectrum of the anomaly field EXP1-EXP2 for the wind in the boundary layer, averaged over 5°N-20°N in the region WAM. Contours drawn on basis of the base-10 logarithm of the power in arbitrary units (Wheeler and Kiladis, 1999); frequency units in count per day. [Results plotted by Stéphanie Leroux, LTHE, Grenoble]

plus when relevant the "annual" mean—i.e. rainy season: April to September. As expected all values decrease as the averaging period increases. The regions are not all the same size, and geographically they span maxima and minima in our observed variability signals. This is reflected in the wide range of magnitudes in table 3.3 and table 3.4, with the smaller study areas obviously showing much larger values.

Secondly, the same information is presented in figure 3.10 and figure 3.11 this time normalised prior to plotting in order to compare the difference behaviour with time for all regions collectively. The values have been normalised with respect to the area under the curve for each respective region, i.e. the sum of the daily, 5-day, 10-day, monthly and seasonal timescales for each considered region and variable. These figures confirm that the signal we observe here is true internal variability. For each variable, all region-averaged values decreased similarly with the averaging period.

To summarise, figure 3.8 provides a visual summary of the findings in the form of a plot of the standard deviation in the time-space domain for the three variables considered. Additionally, figure 3.9 shows a spatio-temporal spectrum for the anomaly of the low-level meridional wind. The peak for internal variability is around four days with a spatial scale of about 2500 km, consistent with the internal generation of African easterly waves. This further illustrates the dynamical coherence of the internal variability signal for the circulation, but again this result is not reproduced for the precipitation.

3.6 Conclusion

It is necessary to have a measure of the uncertainty associated with any experiment. This is certainly true for experiments with regional climate models where the influence of some change in physical processes or boundary conditions is under investigation. It is also true for simulations of climate change and for impact studies, which can amplify uncertainties in small-scale input data. In this chapter we provided a practical measure of random uncertainty suitable for both applications by performing an initial condition perturbation experiment with a regional climate model. Unlike an

Chapter 3: Internal variability

ensemble forecast, where ensemble members diverge, the regional climate model experiments converge because they are controlled by identical boundary conditions. The initial condition is not of primary interest, it serves only to perturb the system. An equilibrium state is reached in which any further difference signal is internally generated and self-sustaining, and is independent of the initial condition. It is sufficient that the initial perturbation is large enough to allow this to occur.

This signal represents the extent to which two identical integrations can differ and is referred to as *internal variability*. Non-identical pairs of experiments performed to evaluate a sensitivity or a response must differ by more than this amount for the conclusions to be reliable. Ensembles of inputs for impact studies must span at least a range of values equal to the size of this signal.

We find that the daily signal for precipitation at the 40 km scale can reach values of about 25 mm. The rainfall signal has a great deal of small scale structure and we deduce that it is associated with essentially random variations in convective triggering. Temporal or spatial averaging thus always reduces this signal, and the reduction occurs relatively rapidly as we move to longer periods and larger scales. Daily temperature and wind variations at the 40 km scale can reach values of around 3°C and 4 m/s. These values persist in time and space to a greater extent than the precipitation signal, particularly over the ocean where the convective influence is weaker and large-scale dynamically coherent structures appear in the difference field.

Naturally these results are most relevant for future experiments with the same model. Other models may produce different results, especially if there is a large difference in simulation of physical processes (the location of maxima in convection, for example). Nevertheless, to the extent that MAR is realistic, or at least to the extent that it shares systematic errors with other models, our findings may have a more general applicability.

	WAM	WSA	CSA	ESA	CSU	CGU	GOU	NIA	OUE
<i>rainfall (mm/day)</i>									
daily	0.14	0.46	0.18	0.13	0.33	0.61	1.42	1.25	1.56
5-day	0.04	0.12	0.03	0.05	0.11	0.13	0.25	0.37	0.36
10-day	0.03	0.05	0.02	0.04	0.06	0.10	0.12	0.26	0.28
monthly	0.01	0.03	0.01	0.02	0.03	0.04	0.03	0.11	0.07
seasonal	0.01	0.02	0.003	0.01	0.02	0.01	0.01	0.05	0.03
"annual"	0.01	0.01	0.0001	0.003	0.01	0.00	0.01	0.03	0.03
<i>10-m air temperature (°C)</i>									
daily	0.05	0.19	0.08	0.08	0.12	0.18	0.36	0.49	0.39
5-day	0.03	0.10	0.05	0.03	0.05	0.09	0.13	0.18	0.07
10-day	0.02	0.05	0.03	0.02	0.04	0.04	0.09	0.14	0.06
monthly	0.01	0.01	0.01	0.01	0.02	0.01	0.04	0.05	0.01
seasonal	0.003	0.01	0.01	0.003	0.01	0.004	0.01	0.03	0.01
"annual"	0.002	0.005	0.0001	0.0008	0.01	0.003	0.005	0.02	0.004
<i>wind in the BL (m/s)</i>									
daily	0.09	0.49	0.23	0.18	0.16	0.28	0.68	1.10	0.52
5-day	0.06	0.29	0.14	0.12	0.08	0.15	0.31	0.30	0.22
10-day	0.04	0.19	0.11	0.08	0.06	0.11	0.26	0.23	0.17
monthly	0.03	0.14	0.08	0.06	0.05	0.08	0.23	0.20	0.16
seasonal	0.03	0.12	0.06	0.06	0.05	0.06	0.17	0.15	0.12
"annual"	0.02	0.09	0.06	0.05	0.04	0.06	0.16	0.15	0.11

Table 3.3 – Maximum absolute EXP1-EXP2 differences for nine regions of interest and each considered variable and timescale. See text for explanation of regions. "Annual" refers to April to September.

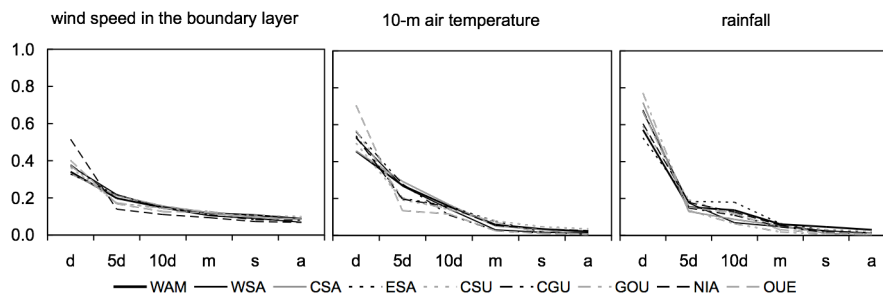


Figure 3.10 – Normalised maximum absolute EXP1-EXP2 differences per region, for each considered variable and timescale: daily (d), 5-day (5d), 10-day (10d), monthly (m), seasonal (s) and "annual" (a).

Chapter 3: Internal variability

	WAM	WSA	CSA	ESA	CSU	CGU	NIA	GOU	OUE
<i>rainfall (mm/day)</i>									
daily	0.04	0.03	0.03	0.02	0.11	0.18	0.24	0.13	0.61
5-day	0.01	0.03	0.01	0.01	0.04	0.06	0.11	0.07	0.13
10-day	0.01	0.02	0.01	0.01	0.03	0.04	0.09	0.05	0.10
monthly	0.01	0.01	0.01	0.01	0.01	0.02	0.06	0.02	0.04
seasonal	0.004	0.01	0.003	0.003	0.004	0.01	0.02	0.01	0.01
<i>10-m air temperature (°C)</i>									
daily	0.02	0.05	0.04	0.04	0.03	0.04	0.12	0.09	0.11
5-day	0.01	0.03	0.01	0.01	0.01	0.02	0.05	0.04	0.03
10-day	0.01	0.02	0.01	0.01	0.01	0.01	0.04	0.04	0.02
monthly	0.002	0.005	0.01	0.01	0.01	0.005	0.02	0.03	0.01
seasonal	0.001	0.001	0.005	0.002	0.003	0.001	0.01	0.003	0.001
<i>wind in the BL (m/s)</i>									
daily	0.02	0.07	0.06	0.05	0.05	0.08	0.13	0.15	0.19
5-day	0.01	0.06	0.02	0.02	0.02	0.03	0.05	0.06	0.04
10-day	0.01	0.05	0.02	0.02	0.01	0.02	0.04	0.05	0.03
monthly	0.01	0.03	0.01	0.01	0.01	0.01	0.03	0.04	0.02
seasonal	0.01	0.03	0.01	0.01	0.004	0.002	0.01	0.01	0.01

Table 3.4 – Standard deviation of the difference between EXP1 and EXP2 for nine regions of interest and each considered variable and timescale. See text for explanation of the regions.

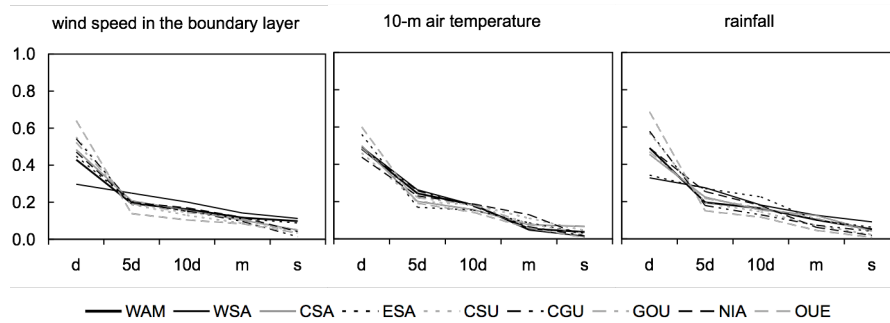


Figure 3.11 – Normalised standard deviation of the difference between EXP1 and EXP2 for the nine regions of interest and each considered variable and timescale: daily (d), 5-day (5d), 10-day (10d), monthly (m), seasonal (s) and "annual" (a).

Chapter 4

Soil moisture and spin-up time

Purpose

Initialising a model is a crucial step in modelling the climate of a region. Although the regional climate models benefit from their driving fields which force them to stick to a certain reality throughout the simulation, the initial state from which computation starts can significantly perturb the model climatology over the first few months of simulation. In this chapter, we investigate one particular variable which stands among the most critical ones for initialisation: the soil moisture.

Contents

4.1	Introduction.....	76
4.2	Experimental set-up.....	77
4.3	Soil moisture anomalies	80
4.3.1	At the initialisation	81
4.3.2	Temporal evolution	85
4.4	Spin-up time	89
4.4.1	Atmospheric spin-up time	90
4.4.2	Global spin-up time	91
4.5	Soil moisture influence	107
4.6	Conclusion.....	110

4.1 Introduction

Any simulation performed with a regional climate model requires information about the climatic situation at the initialisation and throughout the length of the simulation—this is in order to solve the primitive equations of the dynamics in the domain. As time elapses, the model results progressively move away from the initial conditions until a dynamic equilibrium is reached between the large-scale boundary data and the model physics and dynamics. From that moment, the model's ability to represent its own circulation is stable. In contrast to forecast models and global circulation models, this model's departure from the large-scale data does not grow any further (Errico and Baumhefner, 1987). The time needed to reach this equilibrium is known as (*atmospheric*) *spin-up time*. Its length depends on various factors, such as the domain size, the season and the circulation strength. (Anthes *et al.*, 1989; Giorgi and Mearns, 1999; Laprise *et al.*, 2000)

While the atmospheric equilibrium is usually reached within a few days, the inclusion of the surface extends the spin-up time to a much longer period (months, seasons, etc.). Indeed, the continental surface a lower boundary for approximately 30 % of the atmosphere. It directly affects the vertical fluxes of momentum, heat, and water vapour into the lowest layers of the model, and hence constitutes sources and sinks of energy. The land surface model of the MAR is not forced at the boundaries, unlike the atmosphere, and the time it needs to reach a thermal and hydrological equilibrium with the forcing meteorology will highly depend on its initial state. (Yang *et al.*, 1995, Pitman and Henderson-Sellers, 1998³⁷)

Among the soil physical processes, the soil moisture content has been recognised as one of the most critical variables (Yang *et al.*, 1995). Not only do the attached processes have characteristic times of fortnights, months or seasons, but also, moreover, reliable observations that could be used to initialise models are not available.

In this chapter, we study in more detail the influence of the soil moisture

³⁷ The project PILPS achieved an intercomparison of 22 land surface parameterisation schemes.

initialisation over the first months of simulation and its impacts on the spin-up time of the model. Section 4.2 details the simulation performed with this aim in mind. Section 4.3 investigates the characteristics and behaviour of the soil moisture anomalies produced by the initialisation, and section 0 links this with the spin-up time of the model. Section 4.5 summarises the influence of the soil moisture on the simulations, with section 4.6 concluding the chapter.

4.2 Experimental set-up

Seven simulations were performed with the model MAR using two different methods of initialising the soil moisture and starting at different dates. We detail here the simulation characteristics, the methods of soil moisture initialisation and some particularities useful for the further analysis.

a. Simulations

Table 4.1 presents the simulations performed for this study. Each simulation started during the year 1986 and ended on the 31st of December 1987. The simulation REF covers the longest period (two years) and it constitutes our reference simulation as it allows the longest spin-up time. Two sets of three runs were then performed, corresponding to each of the

simulation	start date	end date	soil moisture initialisation
REF	01/01/1986	31/12/1987	fixed to 70 %
F05	01/05/1986	31/12/1987	fixed to 70 %
F09	01/09/1986	31/12/1987	fixed to 70 %
F11	01/11/1986	31/12/1987	fixed to 70 %
E05	01/05/1986	31/12/1987	ERA40
E09	01/09/1986	31/12/1987	ERA40
E11	01/11/1986	31/12/1987	ERA40

Table 4.1 – Performed simulations. Those using a fixed soil water content are labelled with "F". Those using the soil water fields provided by the ERA40 are labelled with "E".

two methods of soil moisture initialisation and all beginning with some soil moisture anomalies.

Different start dates were chosen in order to study the influence of the monsoon rainfall on the evolution of these anomalies. Indeed, an excess of soil moisture could reduce quickly, prompted by dry and warm conditions, or persist or even be enhanced due to rainfall, thereby impacting on the ground hydrological balance and the lower atmosphere water budget. In a similar logic, a lack of soil moisture could significantly perturb the lower atmosphere by pumping its water, or could take months to be filled if rainfall in the region does not arrive for a few months. For these reasons, we chose (i) May 1986 (F05 and E05), corresponding to the time the soil moisture content enhances—the rainy season starts—; (ii) September 1986 (F09 and E09), corresponding to the annual maximum soil moisture content—toward the end of the rainy season—; and (iii) November 1986 (F11 and E11), very close to the annual minimum soil moisture content—after the rainy season (see figure 4.1).

The underlying idea is also to determine which latest start date could be used so that the spin-up time ends before the year of interest (1987 in this case), making that year relevant for analysis straight from January. Moreover, it has often been assumed that a few months—January to April

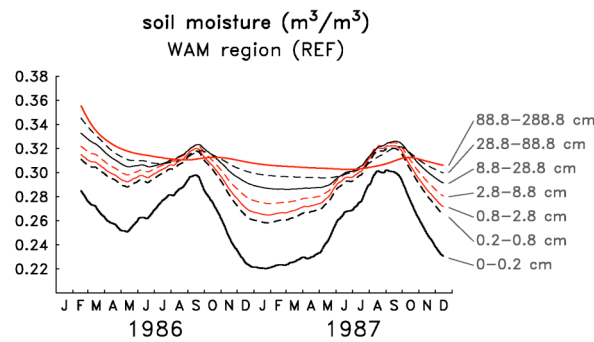


Figure 4.1 – Time series of the soil moisture in each soil layer as in REF, averaged over the WAM region³⁸ (running average of 30 days).

³⁸ The region WAM extends over 18°W-18°E and 5-20°N. See chapter 1, section 1.4 for more detail.

before the monsoon period begins—suffice to get rid of the soil water anomalies introduced at the initialisation, although it is possible that this is not the case.

b. Methods for the initialisation of the soil moisture

Two methods are available in the model to initialise the soil moisture: the soil water content is calculated either with a fixed percentage of saturation (method F) or with the ERA40 soil moisture fields (method E).

Method F consists of initialising the soil moisture to 70 % of the saturation, i.e. the minimum soil water content plus 70 % of the difference between the maximum and minimum soil water contents—which depend on the soil texture. No differentiation is made according to depth. This method was used for the control experiment because it was part of the standard configuration of the model when we started this study. It is possible to change the value of this percentage, but the choice was made to compare both methods E and F rather than one method within itself, the amount of data for analysis being already considerable enough.

Method E uses the fields of the soil water content provided by the ERA40 for the layer 0-7 cm and the layer 7-28 cm (among others). These values, converted to a percentage of humidity, are used as explained for the method F. However, a differentiation is made according to depth. The first four soil layers are calibrated with the first ERA40 layer, the last two layers with the deep one, and the middle one(s) (fifth in our case) with the arithmetic mean of both ERA40 values.

The initial soil moisture fields so obtained are presented in section 4.3.

c. An additional source of information

In the further analysis, we also consider the results of the simulation EXP2 (cf. chapter 3, section 3.2). Remember, it is identical to our reference simulation REF³⁹ in every respect except the initialisation. The initial state of some dynamic variables was modified in order to give rise to differences

³⁹ For the sake of simplicity, the simulation EXP1 of chapter 3 has been renamed here REF.

representative of the internal variability in the model—at least a lower limit. For our current purpose, the simulation was extended until the 31st of December 1987 providing a 21-month simulation (first 3 months not considered) and giving a good estimation of acceptable biases in the model.

4.3 Soil moisture anomalies

Both methods of initialising the soil moisture lead to anomalies of moisture in all soil layers. While method F produces a global wet anomaly over the whole continent—excess of water in simulations F* with respect to REF—, method E produces both dry and wet anomalies.

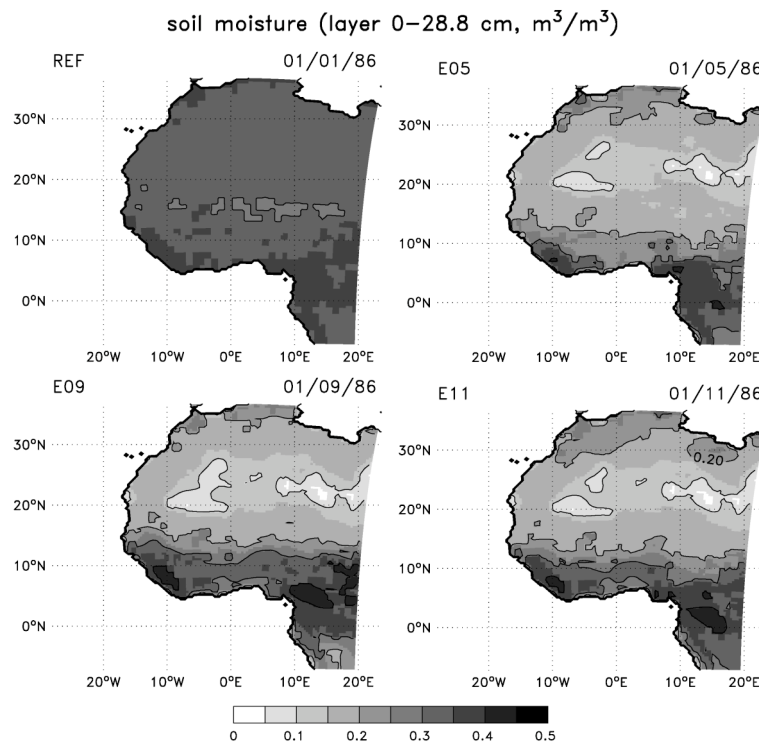


Figure 4.2 – Soil moisture fields at the initialisation in layer 0-28.8 cm, established with method F (top left) or method E (top right and bottom).

4.3.1 At the initialisation

Figure 4.2 shows the soil moisture fields obtained for the initial time step of each simulation. Only the fields for the first five soil layers gathered (surface to 28.8 cm in depth) are illustrated here. Both methods produce the same general features: wet soil over the Fouta Djallon and equatorial Africa, moderately wet soil on the north coast, and drier soils in between.

With method F, equal values are found for all simulations (hence only REF is shown here). The soil texture is responsible for the spatial relative heterogeneity (chapter 1, section 1.4). If not taken into account, it would have led to an equal moisture content all over the domain. The (loamy) sand at the southern boundary of the Sahara corresponds to the smallest amount of soil moisture, whereas the clay (Fouta Djallon, equatorial Africa) and clay loam (north coast and regions around the clay) equate to the greatest soil moisture. As expected, there is no differentiation due to the time of the year.

With method E, spatial heterogeneity is of course emphasized and the time differentiation is clear. May and November are relatively similar to each other and drier than September. November tends to be slightly wetter than May over northern Africa and the Guinean band, which is expected since the month corresponds to the end of the rainy season.

Figure 4.3 presents a comparison of these soil moisture fields at the start date with respect to the reference simulation. The differences between each simulation and REF is calculated for the shallow layer (0-28.8 cm) and the deep layer (88.8-288.8 cm) at the different start dates, assuming that REF is the reality or at least very close to it.

Method F clearly generates a strong wet anomaly over the whole domain and especially in dry regions such as Sahel, Sahara and Northern Africa—in all three simulations and both considered layers. Over the cited regions, the strength and spatial distribution of the anomaly is surprisingly linked to the soil texture: in the shallow layer, the wettest anomaly (around $0.18 \text{ m}^3/\text{m}^3$) corresponds to loamy sand and sandy loam (south border of the Sahara), and the anomaly patches of $\sim 0.12 \text{ m}^3/\text{m}^3$ in the Sahara and further northwards correspond to a sandy clay loam. This structure can only have been produced if the soil moisture in REF is spatially homogenous (as well as temporally):

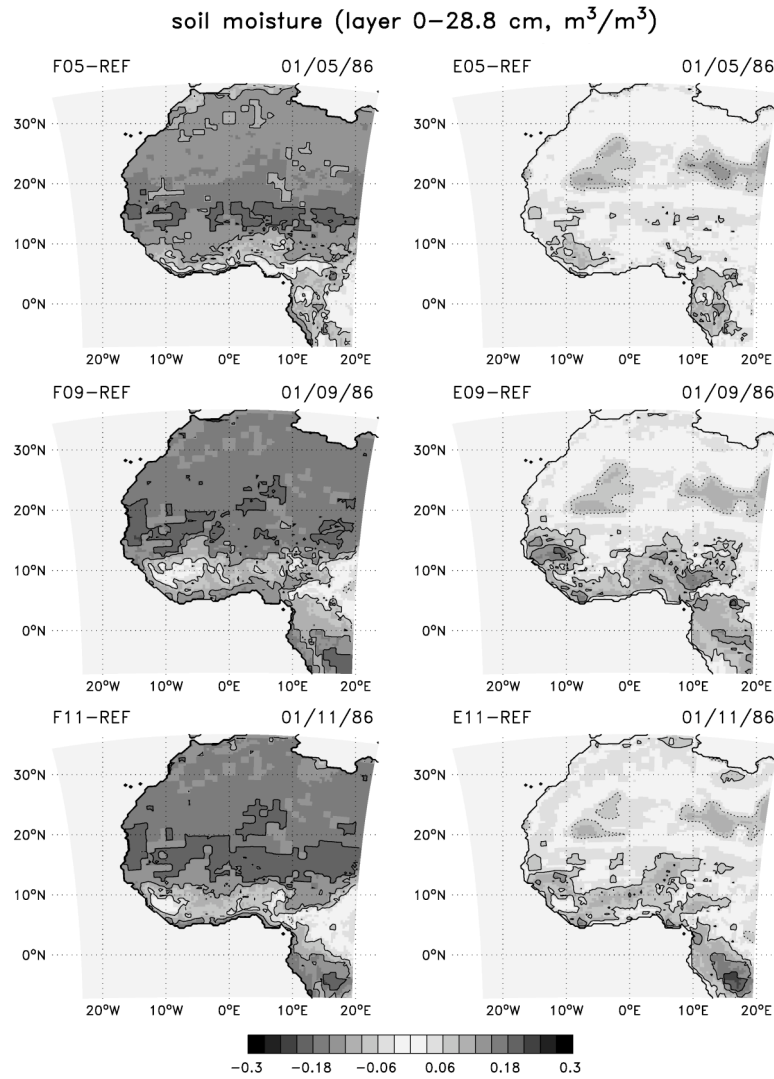


Figure 4.3 – Per-start-date anomalies of soil moisture in the layer 0–28.8 cm, between each simulation and REF. Dotted contours denote negatives values (see Sahara for example).

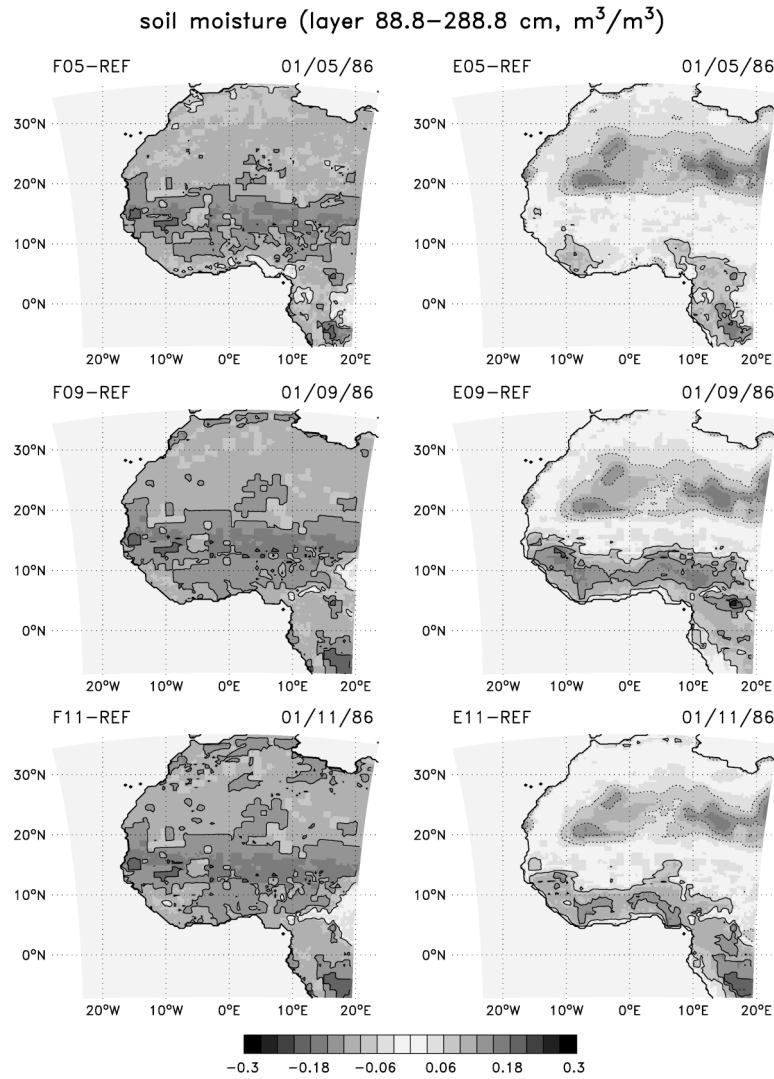


Figure 4.4 – Per-start-date anomalies of soil moisture in the layer 88.8–288.8 cm, between each simulation and REF. Dotted contours denote negatives values (see Sahara for example).

Chapter 4: Soil moisture and spin-up time

the soil is similarly dry all year long. Over the Guinean band, the situation is different. Soils are generally continuously wet and the intra-annual variability of the soil moisture content depends on rainfall distribution. The wetter the soil, the weaker the anomaly; the minimum anomaly follows the location of the rain band. In May, this location is more southern than in September and rather similar to that of November. However, in November, soils have been refilled by the rainy season—they are wet and the anomaly is hence more widely weaker than in May. Starting a simulation with 70 % of moisture saturation is hence close to reality for this region. Two remarks can be added. Firstly, in September and November, the regions with the smallest anomaly match maxima of rainfall in the model, especially the eastern border of the domain which typically experiences high rainfall with this version of the model. Secondly, in between the Fouta Djallon and the Cameroon region, the regions have received a smaller amount of rainfall, the soils are less wet and hence the anomaly is higher. In the deep layer, the anomaly over the whole domain tends to be weaker than in the shallow layer. Indeed, the deep ground is characterised by a global higher moisture content and hence the excess of water introduced at the initialisation is smaller. For the three runs, the wet anomaly reaches $0.21 \text{ m}^3/\text{m}^3$ in the shallow and deep layers.

Method E globally leads to less extended anomalies in both layers than method F. The main characteristics are a moisture dipole with dry anomalies north of 15°N (Sahara) and wet anomalies south of 15°N (Fouta Djallon, Guinean band, equatorial Africa). Differences are the smallest in May, suggesting that REF is actually close to the reality. Indeed, two independent models, SISVAT for the fields shown in REF and ERA40 for the fields given to E05, provide very similar moisture content by independent methods. September presents the strongest wet anomaly. The water excess is much more spread to the south of 15°N than in May and November. At this period of the year, the monsoon is retreating southwards, so it can be assumed that any surplus water is not further amplified. November is an intermediate situation: the lack of soil moisture over the Sahara is similar to both previous cases and the wet soil anomaly is slightly weaker than in

September, although a strong excess of water can be observed in equatorial Africa. The rain band is actually located at this latitude and should have led to a weak anomaly. Since it is the first time step in the model, the excess could be linked to an excess of rainfall in the ERA40 over that region in November (leading to excessive soil moisture in the ERA40 fields) or a lack of rainfall in REF, hence in the MAR. For the three runs, the dry (wet) anomaly reaches $0.15 \text{ m}^3/\text{m}^3$ ($0.24 \text{ m}^3/\text{m}^3$) in the shallow layer and $0.21 \text{ m}^3/\text{m}^3$ ($0.21 \text{ m}^3/\text{m}^3$) in the deep layer. Importantly, both anomalies extend until the deepest layer, which does not favour a quick recovery.

To sum up, the main differences between the methods are (i) the presence of the dry soil anomaly over the Sahara with method E, (ii) the similarity of the intensity of wet soil anomalies with both methods. So far, method E seems to produce less strong and widespread anomalies than method F and is potentially a better method of initialising the soil moisture.

4.3.2 Temporal evolution

Only the first time step has been taken into account so far and we now look in more detail at how the soil moisture anomalies evolve with time. Figure 4.5 and figure 4.6 show the temporal evolution of the monthly soil moisture anomalies in the shallow layer (0-28.8 cm) and deep layer (88.8-288.8 cm). For each simulation, the temporal series starts with the first four months followed by intervals of two months, ending with the last month of simulation. The months filling the gaps until December 1987 are similar to the last month shown.

a. First time step versus first month of simulation

The monthly averages for the first month of simulation indicate a general decrease of intensity of the anomalies (as in figure 4.3; note the change in colour scales). From a wet anomaly reaching $0.21 \text{ m}^3/\text{m}^3$ in both considered layers for all F* runs, values have diminished to a maximum of $0.18 \text{ m}^3/\text{m}^3$ and $0.15 \text{ m}^3/\text{m}^3$ for the shallow and deep layers respectively. For the E* runs, the dry (wet) anomaly has reduced from $0.15 \text{ m}^3/\text{m}^3$ ($0.24 \text{ m}^3/\text{m}^3$) and $0.21 \text{ m}^3/\text{m}^3$ ($0.21 \text{ m}^3/\text{m}^3$) in the shallow and deep layers respectively to

0.15 m³/m³ (0.12 m³/m³) and 0.21 m³/m³ (0.15 m³/m³). In other words, the wet anomalies seem to be decreasing at the same rate in all simulations, while the dry anomalies decrease at a slower rate (see its spread).

Comparing this with the initial anomalies in the shallow layer, the E* runs present similar patterns of anomalies with a reduced intensity. E09 maintains the strongest anomalies and E05 the weakest ones. The F* runs are slightly different. The wet anomalies have reduced more quickly over the hot regions (Sahara) than over the Guinean band. The regions of maximal anomalies actually receive more rainfall in F05 and F09 than in REF—this does not happen in the E* runs. The deep layer shows patterns of anomaly almost identical to those of the initial time step, but with reduced values.

b. Intra- and interannual variability

The first major statement is the progressive and steady decrease of the wet anomalies with time, for both layers. The only ways to modify the soil moisture in the model are the following: motion of water from a lower/upper layer⁴⁰ by gravity, soil suction or root suction; water arrival through rainfall; and water removal by evaporation or by canopy transpiration.

The F* runs reach a similar situation in around 8 months. F09 and F11 started with a wetter anomaly than F05. However, they take advantage of the dry season to eliminate the anomalies over the Guinean band more quickly. F05 sees the wet anomaly maintained over that region until August 1986, from which time it starts to vanish at the same rate as before. The wet anomaly persists there because it already existed before the rainy season and the latter one prevents the soil moisture excess to be removed as quickly as during the dry season. A similar explanation holds for the deep layer.

The dry regions (Sahara, Sahel) see their wet anomaly disappearing at a similar rate for all three runs—note that F09 and F11 started with a greater soil moisture excess. However, it takes longer to disappear than in the Guinean band because the soil water can only be removed through direct evaporation from the barren ground—no vegetation pumping water through its roots.

⁴⁰ No lateral exchange of moisture from a mesh to a neighbour one is allowed.

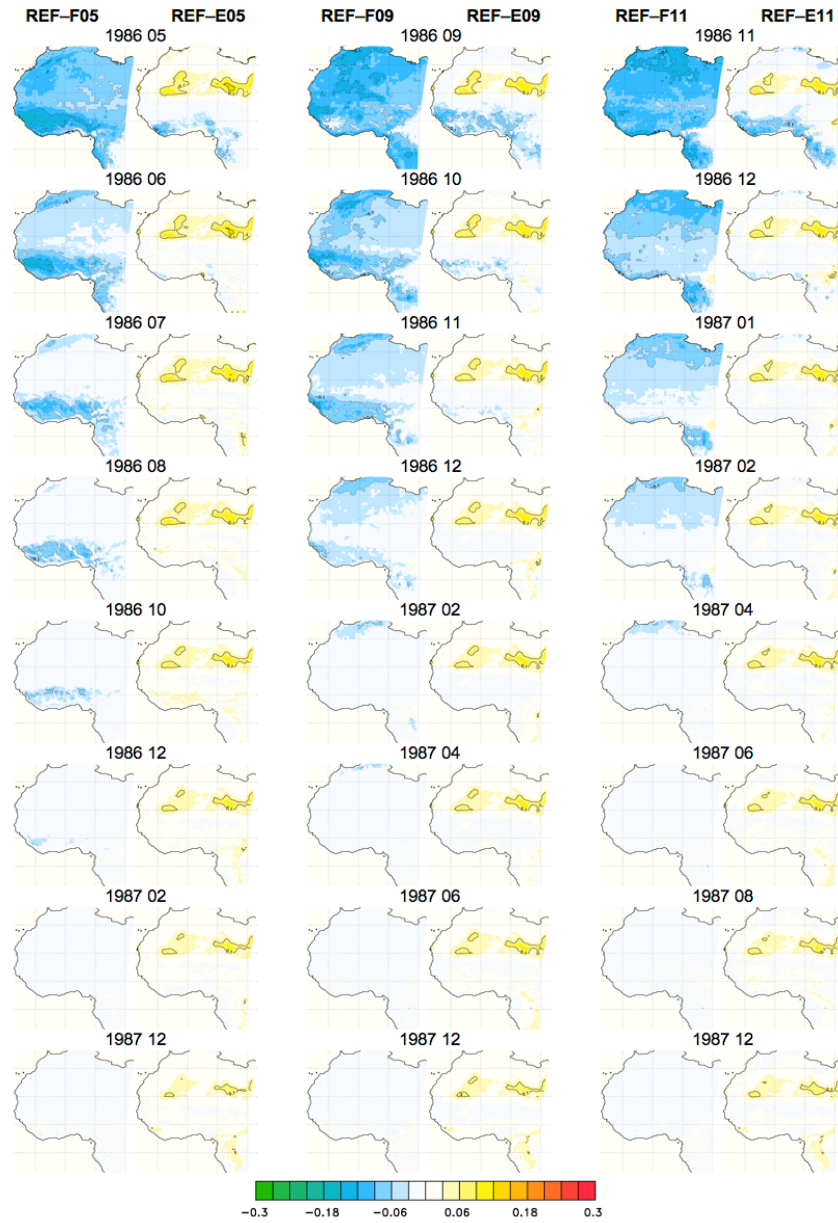


Figure 4.5 – Monthly anomalies of soil wetness in layer 0-28.8 cm (m^3/m^3) between each simulation and REF. Irregular time scale.

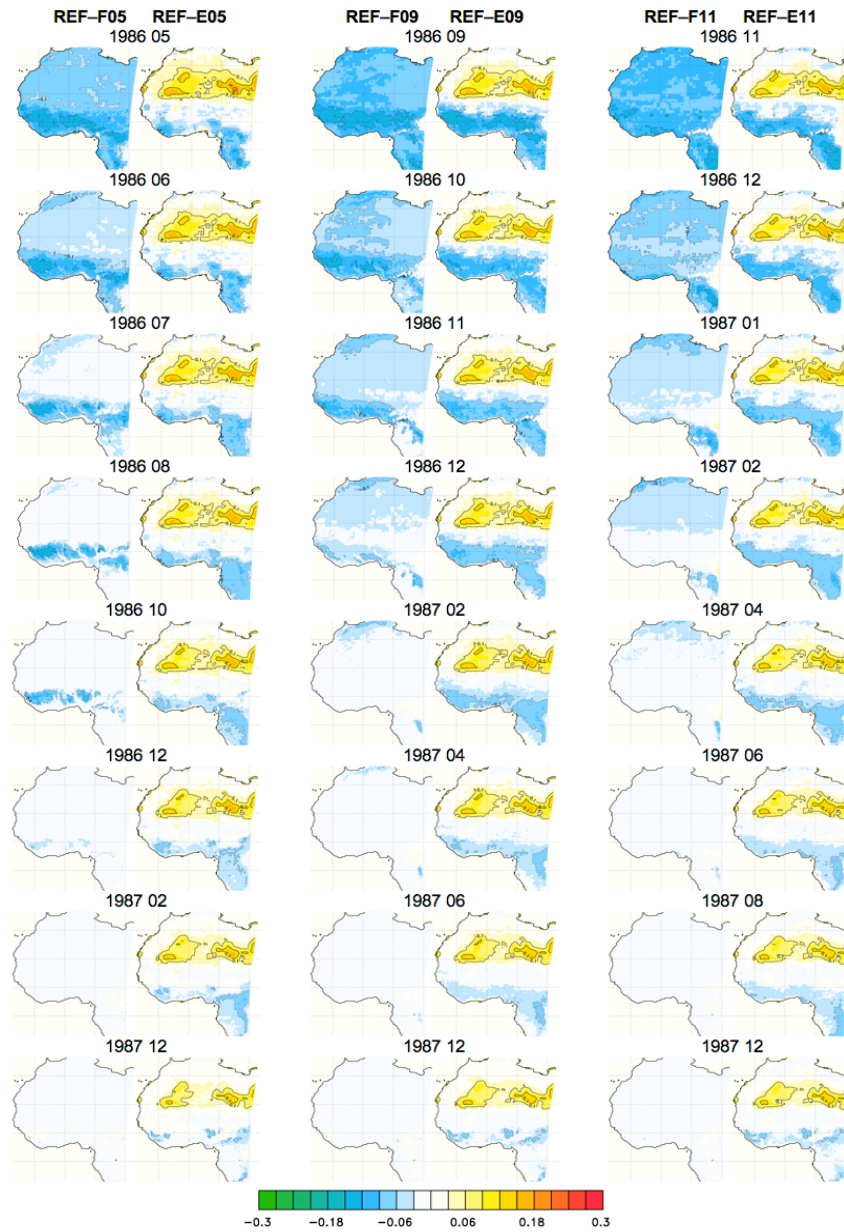


Figure 4.6 – Same as figure 4.5 but in layer 88.8-288.8 cm (m^3/m^3).

The E* runs which had globally weaker absolute anomalies at the initialisation are surprising. The dry anomaly is sustained while the wet anomalies reduce much quicker than in the F* runs. The dry anomaly presents a particularly steady behaviour. It lasts until the end of each simulation, decreasing slowly from month to month. This is explained by its location. Indeed, it relates to a dry anomaly observed over the Sahara, a region known to receive no rainfall. Although the high temperatures favour a high saturating water vapour pressure, the air is dry (~ 4 g/kg). The soil absorbs the air moisture, but this process is slow and the dry anomaly is therefore very slowly absorbed. Since it extends to the last soil layer, it will take even longer to be totally removed. On the other hand, the wet anomaly over the Guinean band disappears more quickly in these runs—three months seem sufficient for the shallow layer. All six simulations show the first month experiencing more rainfall than in REF over that region (not shown). This wet rainfall anomaly is weaker in the E* runs than in the F* runs, and is insignificant after one month—it maintains for 2-3 months in the latter ones. This could explain why the soil moisture excess disappears more quickly in the E* runs. It is not amplified by abnormal rainfall and disappears in a shorter period. Although the wet anomaly vanishes quickly in the shallow layer, it maintains longer in the deep layer in all three E* runs compared to F* runs.

Such long-lasting and wide-spread anomalies of soil moisture—dry or wet—are expected to impact on the model climatology through heat and water exchanges between the surface and the low atmosphere. The next section looks in more detail at a number of atmospheric variables, especially the model's spin-up behaviour.

4.4 Spin-up time

Using any of the two methods of soil moisture initialisation, the simulations start with anomalies in soil water content. These are wet or dry, small or great, and spread all over the continent. It takes several months for each simulation to reach a state close to REF. How the soil moisture anomalies impact on the model's ability to represent the climate over West

Africa is the topic of this section. The spin-up time is a very good indicator for this: firstly, we consider only its atmospheric component; secondly, we consider it globally and characterise the influence of the soil moisture on its length and behaviour.

4.4.1 Atmospheric spin-up time

As mentioned in the introduction, the *atmospheric spin-up time* is the time needed by the (atmospheric) internal physics and dynamics of the model to reach a dynamical equilibrium with respect to the driving large-scale fields. Figure 4.7 shows the time series of the root-mean-square (RMS) errors between the modelled and analysed sea-level pressures and geopotential heights at 500 hPa for the seven simulations with respect to the number of days since their start date. This is calculated over the inner domain (see chapter 1, section 1.4).

At the initialisation, the RMS errors are small, but not strictly zero as expected. This is because of the vertical and horizontal grid interpolations of the large-scale fields onto the MAR grid. The RMS errors then increase for a few days until they reach a stable value. This asymptotic behaviour is typical of other atmospheric variables too and is very closely similar to the one found for the RCMs MM4 by Anthes *et al.* (1989) and RegCM by Giorgi and Mearns (1999) over the U.S.A..

Within the first month of simulation, the mean RMS errors are 2.8 hPa for the sea-level pressure and 13.5 m for the geopotential height at 500 hPa. Not shown, these errors are 0.5 and 2.0 g/kg for the specific humidity at 500 and 850 hPa respectively, and 0.5 and 2.5°C for the air temperature at these same levels respectively. All these values are in very good agreement with those found by Anthes *et al.* (1989) for the same variables: 3.5 hPa (around 2 hPa for RegCM by Giorgi and Mearns (1999)), 22.8 m, 0.5 and 1.8 g/kg, 1.2 and 2.1°C (see hereafter for the sea-level pressures and geopotentials). For another nested grid model, Anthes (1983) found RMS errors of 24.7 m for the same geopotential and of 1.5 and 4.5°C for the temperature at 500 and 850 hPa.

On basis of the curves of figure 4.7, the atmospheric spin-up length can

therefore be estimated as 3–4 days, which is a typical value also obtained by Giorgi and Mearns (1999) for RegCM (not provided for MM4). In the long term, the global stabilisation of the model's departure from the driving fields is confirmed by figure 4.8, where the entire length of the simulations is considered and its climatological ability remains stable.

After the initial atmospheric spin-up time of a few days, all seven simulations show RMS errors of the sea-level pressures oscillating around 2.2 hPa and those of the geopotential at 500 hPa oscillating around 13.3 m, values still in good agreement with those aforesaid.

4.4.2 Global spin-up time

Figure 4.7 and figure 4.8 suggest that the initialisation of the surface plays a role during the first months of simulation. Indeed, all F* simulations have a positive departure from REF no matter their start date, while all E* simulations tend to require much less time to get closer to REF. We define here the *spin-up time* as the total time the model needs to get independent of its initial conditions and stable in skill.

In our current set of experiments, soil moisture is the only factor potentially responsible for the differences observed between the curves in the previous figures. If the season (start date) or the boundary conditions were responsible, then a similar behaviour would be found for E05 and F05, E09 and F09, E11 and F11 —since the atmospheric forcings are the same.

Figure 4.9 shows the difference between the RMS errors with respect to the reference simulation (i.e. $\text{RMS}_{\text{simulation}} - \text{RMS}_{\text{REF}}$). The EXP2 simulation clearly indicates that our set of experiments is under the influence of another parameter other than purely internal variability. The E* runs show a small departure from REF over the first months and quickly get close to REF. The F* runs present a greater departure from REF and take a longer time (potentially 7 months) to get similar to REF. This figure also suggests that the element responsible for these differences is close to the surface. Indeed, the differences of RMS errors for the geopotential at 500 hPa do not show the same strong departure from REF, except F05 and F09.

In order to assess the significance of these differences, a two-sample *t* test

Chapter 4: Soil moisture and spin-up time

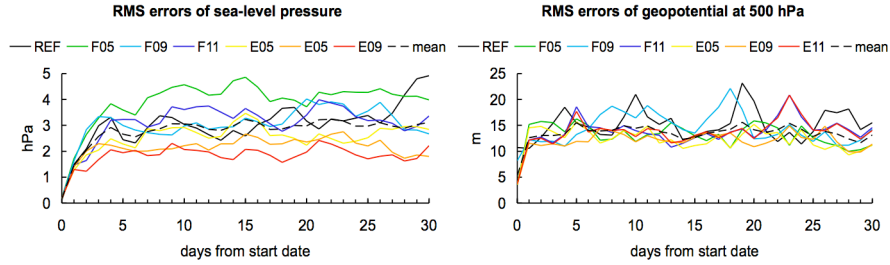


Figure 4.7 – RMS errors between MAR and ERA40, computed over inner domain, first 30 days of each simulation; mean of all curves.

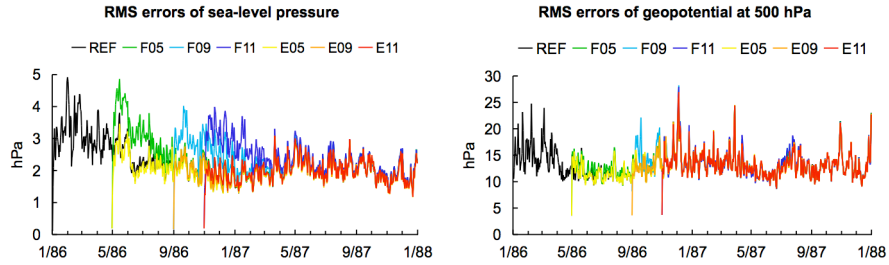


Figure 4.8 – Same as figure 4.7 for the entire simulation length.

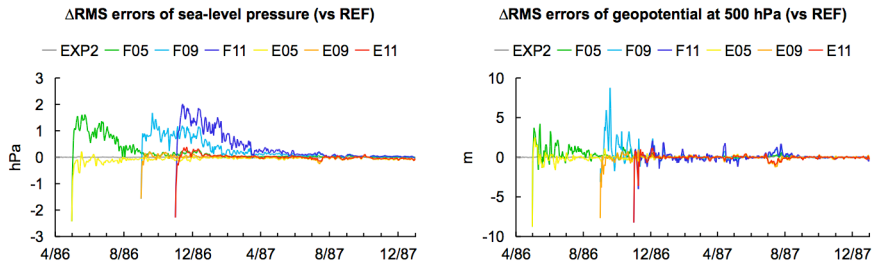


Figure 4.9 – Differences between the RMS errors of REF and the RMS errors of the other simulations.

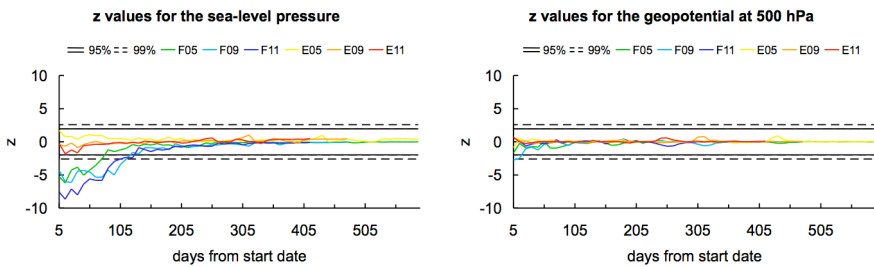


Figure 4.10 – z values of the 2-sample t test for the series of figure 4.9 compared to REF. The black lines show the intervals of confidence.

for correlated data was performed on the data (Wilks, 1995: chapter 5, *Hypothesis testing*), using a similar technique to that used by Anthes *et al.* (1989). The first 90 days of each simulation were compared to the corresponding first 90 days of the simulation REF, thanks to:

$$z = \frac{\overline{X}_{REF} - \overline{X}_{run}}{\sqrt{\frac{s_{REF}^2}{n'_{REF}} + \frac{s_{run}^2}{n'_{run}}}} \quad (4.1)$$

with z the test statistic. The overbar denotes the average of the considered variable over the first 90 days of the simulation, s^2 is the sample variance over the same length and n' is the effective sample size as proposed by Wilks (1995) (equation 5.12). The null hypothesis is that the difference between a run and REF equals zero.

For the sea-level pressure, all F* runs are significantly different from REF at a 99 % level, while all E* runs are not. The EXP2 run is significantly equal to REF at a 99 % level⁴¹. For the geopotential at 500 hPa, none of the differences are significant at a 99 % level (nor 95 %).

Therefore, there is clear evidence that the spin-up time is much longer than the atmospheric spin-up time. In order to determine its length with precision, we shall consider two different approaches. On the one hand, we determine how long the previous errors are significantly different from REF (statistical approach). On the other hand, we look at how long the physical differences may last and combine this with the results of EXP2.

a. Statistical approach

A way to estimate the length of the global spin-up time is to compute the extent to which the differences between the runs and REF are statistically significant. Our two-sample t test was modified so as to be performed per 10-day period with the results shown in figure 4.9. The statistic test z is now expressed as:

⁴¹ $z = -3.54, -6.74, -11.20$ for F05, F09 and F11 respectively. $z = 0.74, -0.46, -1.43$ for E05, E09 and E11 respectively. $z = -0.003$ for EXP2. At a 99 % level, $z = \pm 2.58$; at a 1 % level, $z = \pm 0.01$.

$$z(t) = \frac{\overline{X}_{REF}(t) - \overline{X}_{run}(t)}{\sqrt{\frac{s_{REF}^2}{n_{REF}(t)} + \frac{s_{run}^2}{n_{run}(t)}}} \quad (4.2)$$

where the overbar denotes the average of the considered variable over a 10-day period, s^2 represents the variance of the variable over the whole simulation length (computed on the basis of the daily data) and n represents the number of points used to compute the average (i.e. 10). The null hypothesis is that the difference between a run and REF equals zero. Results are shown on figure 4.10.

At 500 hPa, only the simulation F09 is significantly different from the simulation REF at a 95 % level of confidence for the first two 10-day periods. This is one more indication that the factor increasing the spin-up time is related to the surface. Indeed, close to the surface (sea-level pressure), the F* simulations do present a significant difference from REF at a 99 % level of confidence for a few months. All E* simulations do not, even at a 95 % level. F05, F09 and F11 are significantly different from REF for 75, 115 and 125 days respectively (at both 95 and 99 % level). The spin-up time would therefore end around mid-July 1986 for F05 (after 2.5 months), late December 1986 for F09 (after 4 months), and end of February for F11 (after 4 months).

This constitutes a first estimation of the spin-up time showing that a minimum of 2.5 months is necessary to let the surface model adjust with respect to its initial conditions and the atmospheric forcings. However, this was established on basis of one atmospheric variable. We shall see that surface or close-to-the-surface variables seem to require a longer time.

b. Physical approach

Another approach is to look at the temporal persistence of the differences. In this section, we search for the maximum absolute differences between each simulation and REF, observed across the whole domain with a 10-day period time step (see figure 4.11). Considered variables are the soil moisture (layers 0-28.8 cm, 28.8-88.8 cm and 88.8-288.8 cm), the albedo, the evapotranspiration, the surface temperature, and three dynamic variables—

specific humidity, air temperature, and zonal wind⁴² (all at 10 metres high). The shaded area on each graph of figure 4.11 corresponds to the maximum absolute difference observed between REF and EXP2 at a 10-day period time scale, over the 21 months for which these simulations are considered.

The maximum values obtained are not representative of a single point (random) variability, at least for the first months of simulation. Indeed, as will be shown, differences between the simulations and REF are characterised by large spatial scales and temporal persistence. They are related to the soil moisture anomalies for a long period until they finally represent local random variability. Of course, these maxima might not be located in the same places and this point will be taken into account in the further analysis.

A quick look at figure 4.11 shows the expected global decrease (with time) of the maximum differences between simulations and REF. A distinction can usually be noticed between E* and F* simulations as well as a rather large variability in the time at which the differences enter the range of the internal variability. Locating when the maximum differences enter the range of the internal variability should provide us with an estimation of the length of the spin-up time.

The difference in **soil moisture** presents similar behaviour for all layers. The resorption of the anomalies in F* simulations tends to be faster than in E* simulations. The particularly steady behaviour of E* simulations is explained by the location and type of anomaly (see section 4.3.2). Until the end of each simulation, the dry anomaly steadily and slowly decreases. In terms of spin-up length, nothing can be estimated. The information provided by EXP2 is only a lower limit to the internal variability in the model; it can be higher. This particular behaviour suggests that the spin-up time could require a longer period than the length of the simulations themselves. Looking at the wet soil anomaly (see figure 4.5 and figure 4.6), a period of 3 months seems enough to remove the anomaly in the shallow layer, while the whole length of each simulation does not suffice in the deep layer. All

⁴² The meridional wind behaves very similarly and will not be used.

Chapter 4: Soil moisture and spin-up time

F* simulations show a decrease of the maximum differences, followed by a rather steady period and then a final increase. This corresponds to the time the wet anomaly has been removed, and further maxima found are representative of the variability of one or two meshes. While figure 4.5 suggest a global spin-up time of 8 months (240 days) for all three simulations, figure 4.11 and the inclusion of EXP2 reduce this length to 70-220 days for all layers (the exact values are given in table 4.2).

The difference in **albedo** exhibits similar behaviour to that of the soil moisture anomalies. As expected, the albedo anomaly is much more intense in the F* simulations than in the E* simulations. The albedo anomaly is located over the Sahara and is maintained as long as the wet soil anomaly persists in the region (wetter soils have a lower albedo). It decreases more quickly in F05 since the wet anomaly over the Sahara decreases faster (it started with a smaller value than in F09 and F11). The soil moisture fields of E* simulations are much closer to REF in terms of spatial characteristics; the albedo anomalies with respect to REF start at smaller values and quickly reach the range of the internal variability. Although the dry anomaly persists, it does not seem to lead to a sustained albedo anomaly. The spin-up time ranges between 100 and 260 days for all six simulations (more details in table 4.2). It has to be added that the precision for the E* simulations is very low due to the flatness of the curves, and the range of internal variability is likely to be wider.

The difference in **evapotranspiration** is more variable. Remember, evapotranspiration includes three contributions: evaporation from bare soil, evaporation of water deposited on the canopy, and vegetation transpiration. The previous variable (albedo) showed a progressive and quiet decrease. Note that the area of maximum anomaly was constant in location and directly related to the soil moisture anomaly for at least 200 days. It is different for evapotranspiration: it is of course related to places of strong soil moisture anomalies, but these actually move with time (see figure 4.12 which shows the monthly anomalies in evapotranspiration with respect to REF).

Chapter 4: Soil moisture and spin-up time

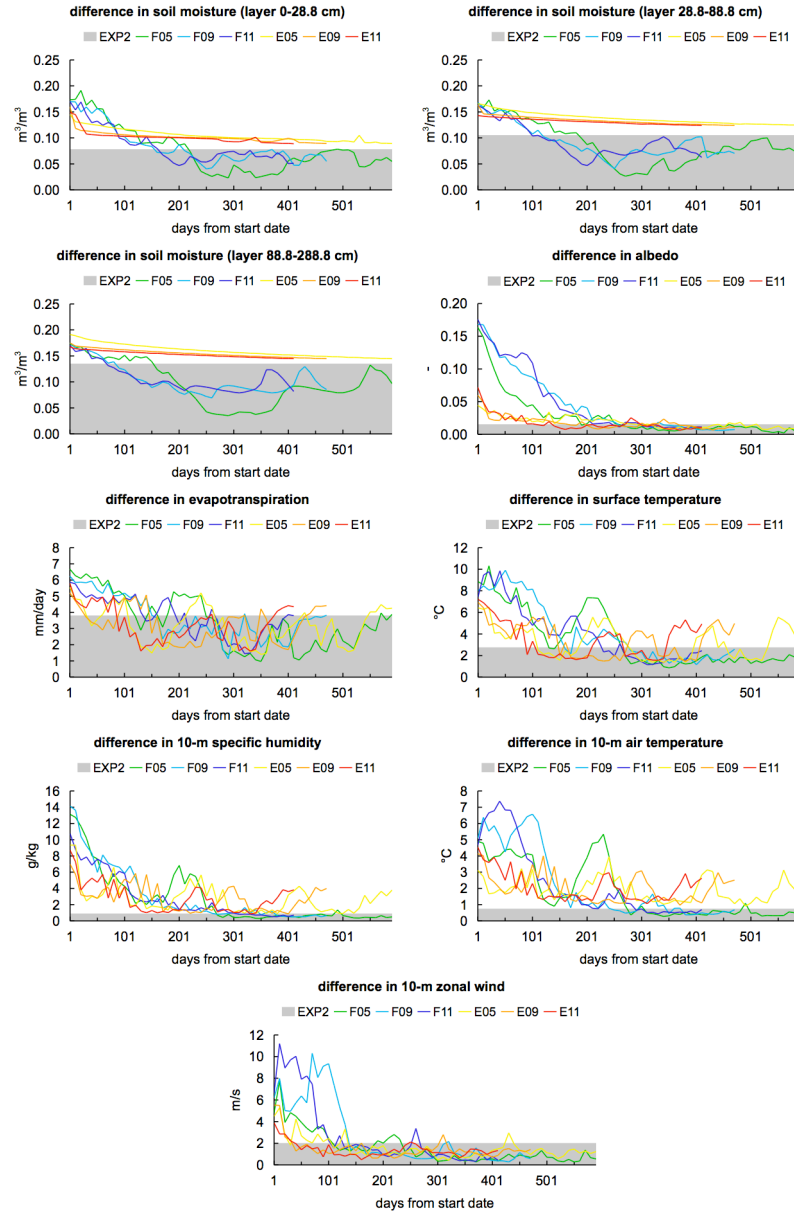


Figure 4.11 – Maximum absolute differences between REF and each simulation, observed across the domain (10-day values). The grey shaded area represents the lower limit of the internal variability.

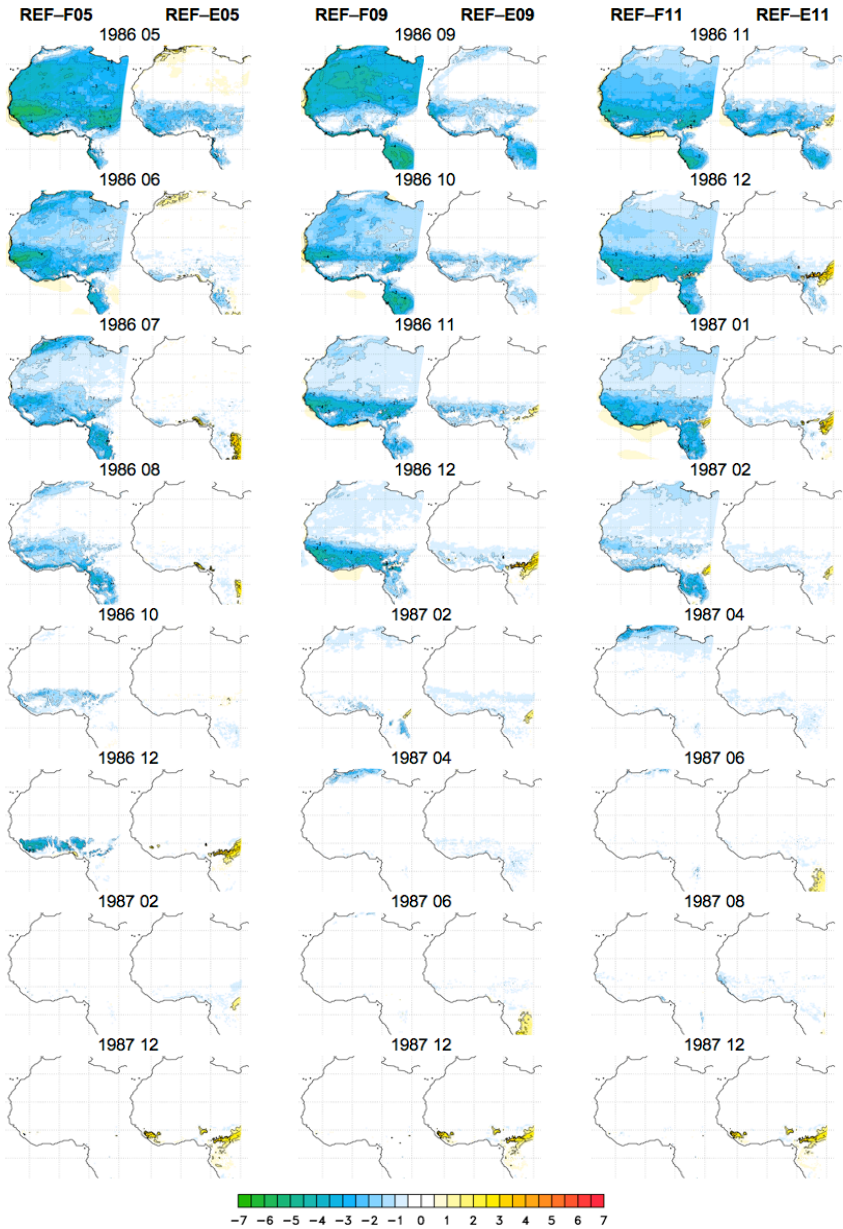


Figure 4.12 – Monthly anomalies of evapotranspiration (mm/day) between each simulation and REF. Irregular time scale.

In F11, the evapotranspiration negative anomaly (i.e. more evapotranspiration than in REF) is initially stronger over the Guinean band albeit the maximum soil moisture anomaly is located in the northern half of the domain. This is explained by two combined factors: (i) the vegetation contribution through soil water extraction by the roots is nonexistent over desert regions, (ii) cold temperatures in the northern part of the domain limit the evapotranspiration. This is especially true over the Atlas mountains where the wet soil anomaly will start to reduce significantly only from March 1987 (day 120). This month also corresponds to the point at which the excess of evapotranspiration in the Guinean region has been totally removed. Hence, in figure 4.11 the maximum values are those of the Guinean band until day 120, and of the Atlas mountains from day 120. At that moment, evapotranspiration can increase over the Atlas and this leads to the observed increase in the maximum difference. It lasts approximately a couple of months until the soil moisture anomaly is gone. From July 1987 (day 180), evapotranspiration anomalies are local and independent of the initial soil moisture anomalies. The global spin-up time is estimated to be 190 days.

The behaviour of the evapotranspiration in F09 also shows some interesting features. In September 1986 (initial month), the evapotranspiration anomaly pattern is similar to that of soil moisture (see figure 4.5 and figure 4.12). The evapotranspiration can not be very efficient over the Guinean band due to the rainy season, where saturation of the canopy evapotranspiration is likely to have been reached. Nevertheless, unlike F11, conditions are more favourable over the northern part of the domain and no null difference is observed this time. As months pass, the conditions in the north tend to reduce the evapotranspiration while those in the south tend to favour it. The maximum anomalies shift southwards over the Guinean band. Again, in March 1987 (day 180), the soil moisture anomalies over the Guinean band have disappeared. Conditions again get better in the north, the still present excess of soil moisture being evacuated through evaporation (secondary maximum in figure 4.11). The increase is within the internal variability of the model as less water remains to be

Chapter 4: Soil moisture and spin-up time

evaporated at this stage than in F11 (due to a longer previous period with little evaporation). From July 1987 (day 300), the maximum anomaly is representative of a very local signal not linked to the initial soil moisture anomalies. The spin-up time is estimated to be 150 days.

The case of F05 is probably the most interesting in terms of the rainy season impact on the soil moisture anomalies (and hence evapotranspiration). For the first few months, the evapotranspiration anomaly matches the soil moisture anomaly in spatial distribution. The excess of soil moisture over the Atlas mountains is evacuated much more quickly than in F09 and F11 due to favourable conditions. However, during the rainy season, the anomaly patterns reflect a weakening of the evapotranspiration over the Guinean region while the soil moisture anomaly is still present. It is only from October 1986 (day 150) that it increases again to evacuate the soil water excess so maintained by the rainy season. This explains the secondary maximum in figure 4.11. From February 1987 (day 270), maximum anomalies are representative of local and random variability. The spin-up time is estimated to be 250 days.

In the E* simulations, the behaviour of the evapotranspiration negative anomaly is similar to the F* simulations, but quicker due to an initially lower excess of soil moisture. For selected months, a positive anomaly (less evapotranspiration than in REF) develops in equatorial Africa. This happens in July 1986 (day 60 in E05), December 1986 (days 210, 90 and 30 for E05, E09 and E11 respectively), July 1987 (days 420, 300 and 240 resp.) and each time lasts for 2 to 3 months. The same appears in the F* simulations but is not as emphasized because of initially wetter soils than in the E* runs. All E* runs also show a positive anomaly toward the end of the simulation. This is directly related to a lack of moisture that develops from November 1987. This lack of soil moisture is not represented on the graphs of the difference in soil moisture because it does not reach the same intensity as the dry anomaly over the Sahara, which is triggered by the initial lack of moisture. The spin-up time appears to be 30 days for E05 and 90 days for E09 and E11.

The **surface temperature** anomaly is distinctive. This variable is also

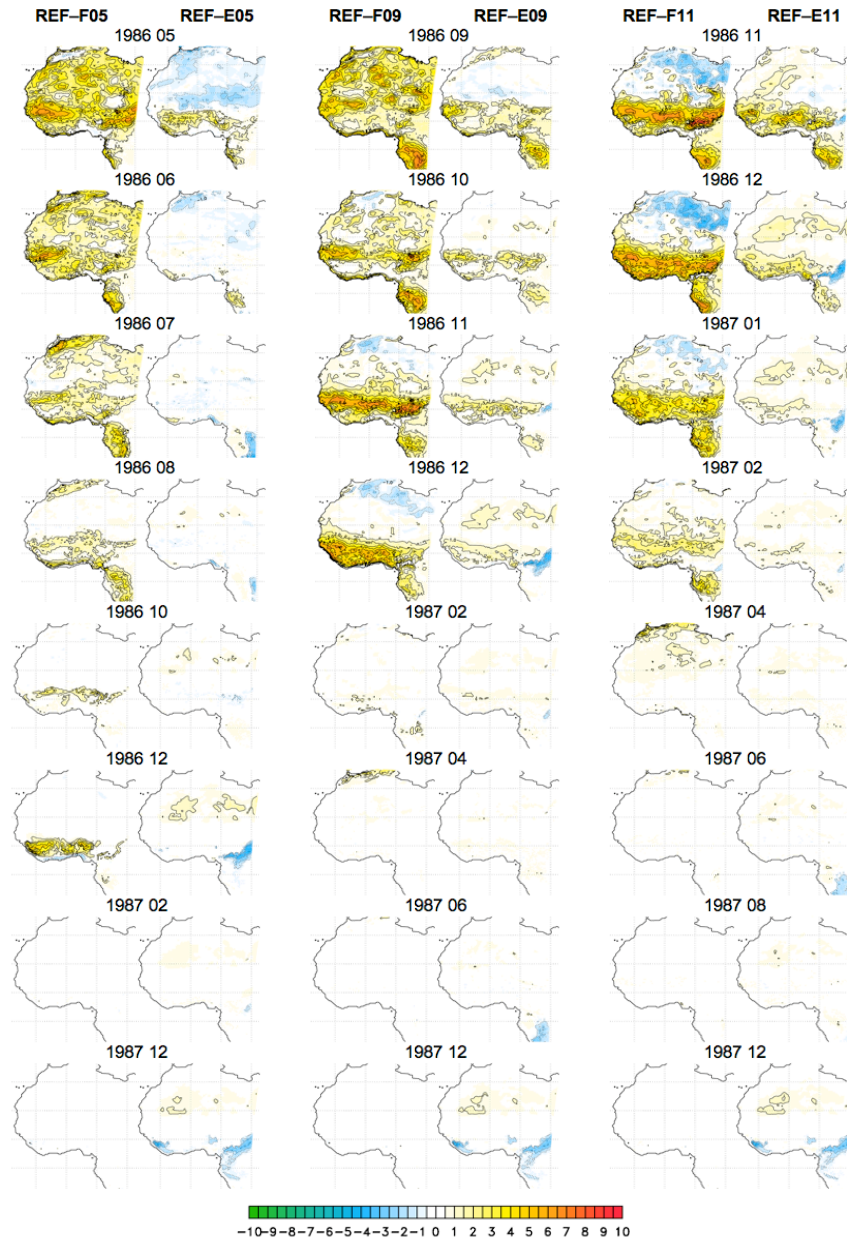


Figure 4.13 – Monthly anomalies of surface temperature (°C) between each simulation and REF. Irregular time scale.

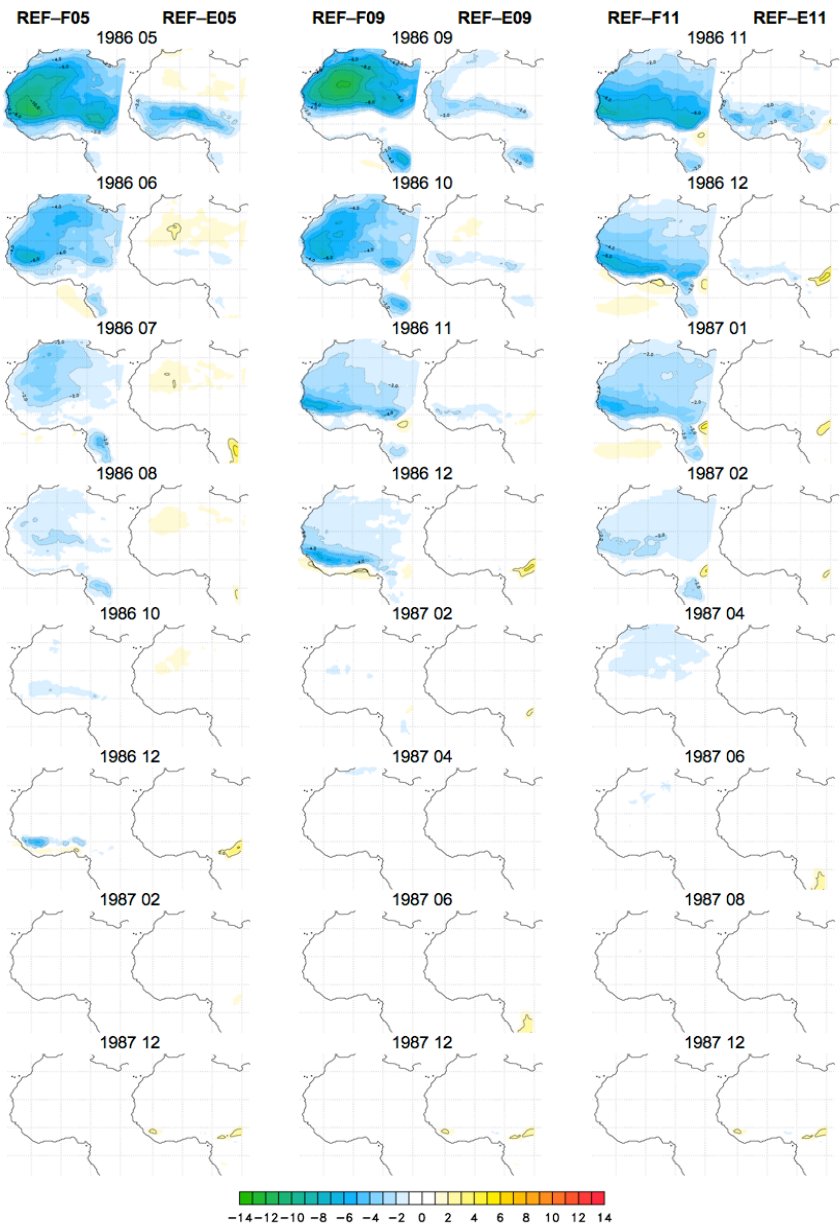


Figure 4.14 – Monthly anomalies of 10-m specific humidity (g/kg) between each simulation and REF. Irregular time scale.

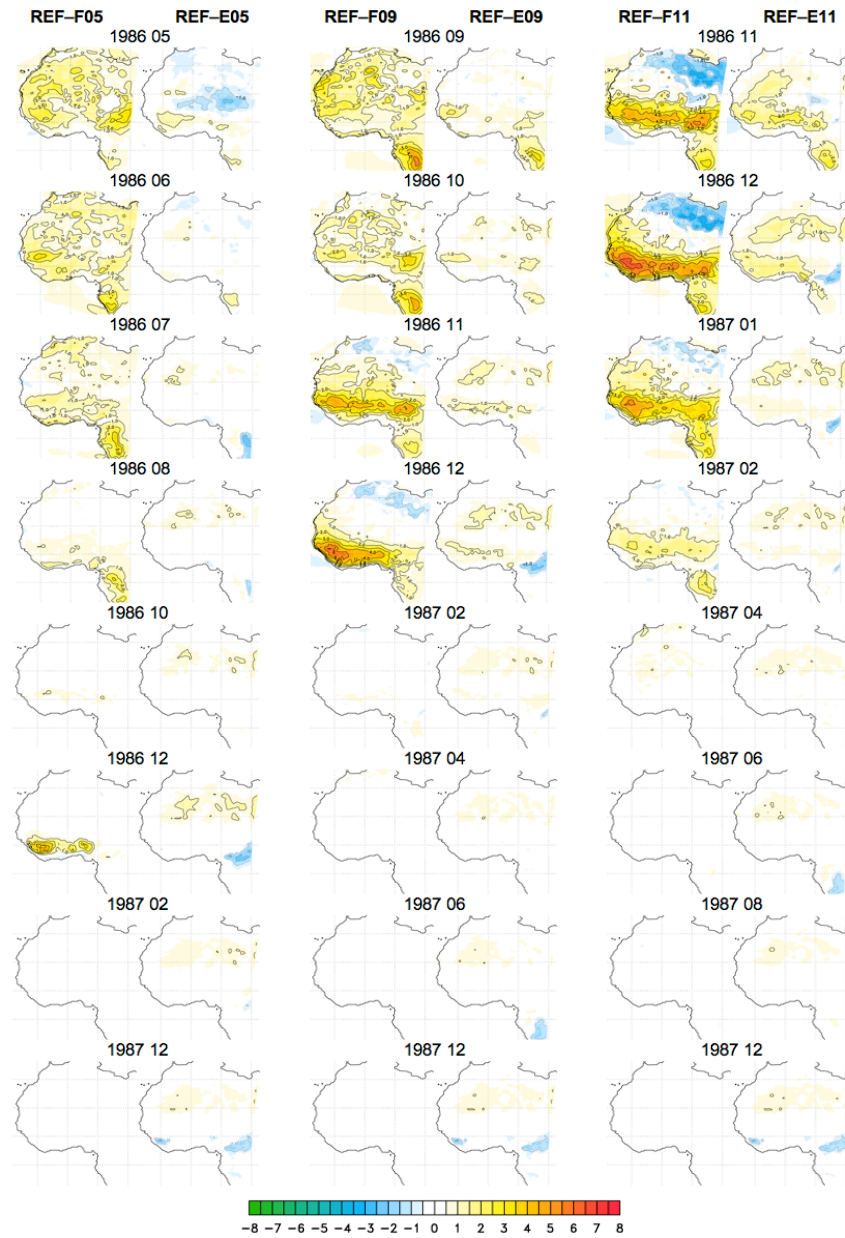


Figure 4.15 – Monthly anomalies of 10-m temperature (°C) between each simulation and REF. Irregular time scale.

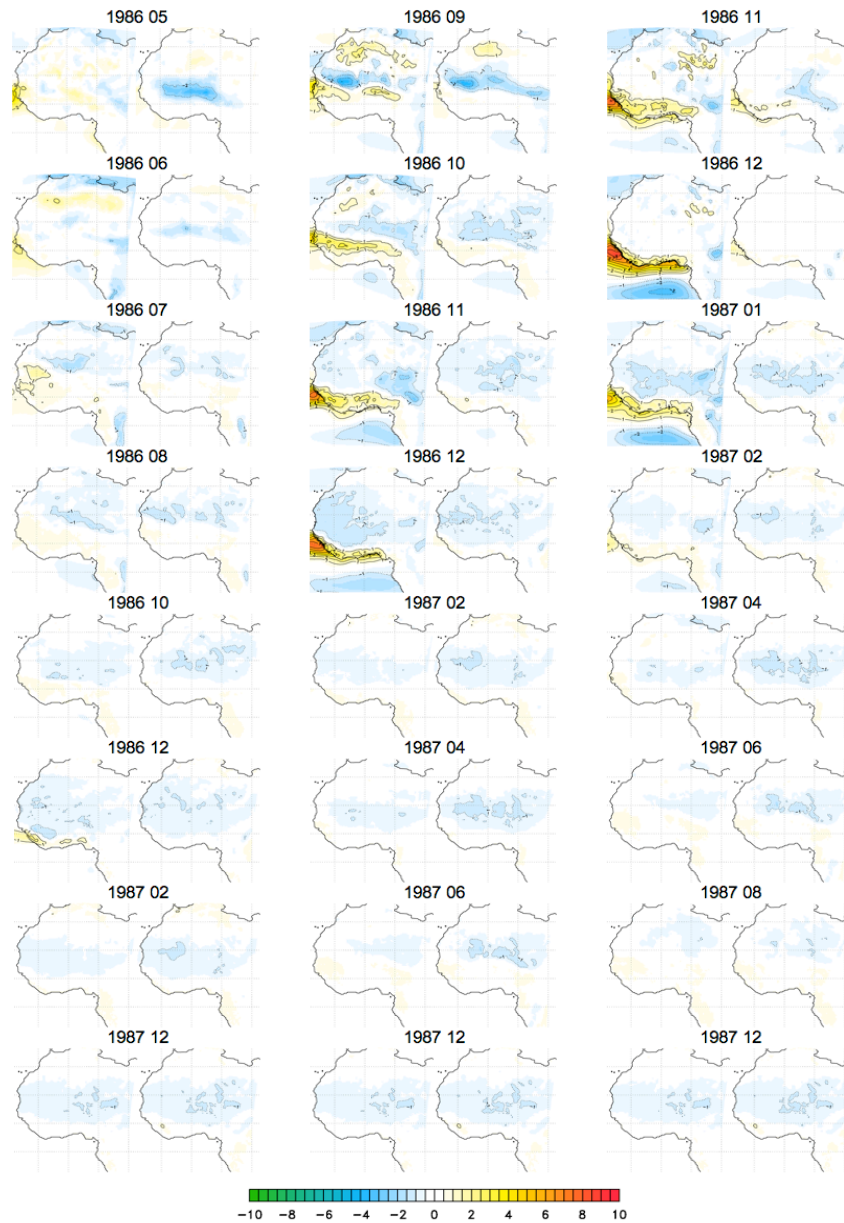


Figure 4.16 – Monthly anomalies of 10-m zonal wind (m/s) between each simulation and REF. Irregular time scale.

initialised by external data. Indeed, in terms of physical variables, the MAR needs the initial values of the zonal and meridional winds, pressure depth, air specific humidity at all atmospheric levels, surface temperature (sea and land), albedo, leaf area index, green leaf fraction, soil moisture content per layer and the soil layer temperatures. However, after the initial time step, the surface temperature is let free to evolve according to the physics of SISVAT. Since the soil moisture anomaly is severe at the initialisation and persists for months as previously shown, the monthly mean of the surface temperature anomaly is very likely to show variability caused mainly by the soil water anomalies.

In common with the other variables, the anomaly with respect to REF shows a decrease with time in figure 4.11. The E* simulations again present weaker departures than the F* simulations and several features confirm the link between the surface temperature and the evapotranspiration, hence the soil moisture anomalies. Figure 4.13 illustrates the monthly anomalies of surface temperature, and again will help to explain the shape of the curves in figure 4.11.

In the E* runs, higher surface temperatures (negative anomaly) are linked to a lack of soil moisture. That anomaly is weaker in absolute amplitude than the cooling (positive anomaly) of the surface in regions of soil moisture excess. In E05, two months are sufficient to remove the initial soil moisture anomalies. Further increases in the maximum differences shown in figure 4.11 are connected to particular and local events (linked to the evapotranspiration anomalies), especially over tropical Africa and the Cameroon region. These features also appear in E09 and E11 during the same periods as previously observed: July 1986 (day 30 in E05), December 1986 (day 210, 90 and 30 in E05, E09 and E11 respectively), July 1987 (day 420, 300, 240 respectively) and December 1987 (day 570, 450, 390 respectively). In E09 and E11, the excess of soil moisture takes longer to disappear than in E05 and maintains the colder surface temperatures. The spin-up time is around 40 days for E05, 120 for E09 and 90 for E11. For the first two, the time might appear shorter than expected in figure 4.11, but this is due to the fact that a secondary maximum biases the signal (this maximum

Chapter 4: Soil moisture and spin-up time

is not linked to the initial soil moisture anomalies).

The F* simulations experience stronger surface temperature anomalies and show a similar decrease over the first 5 months of simulation (150 days). In figure 4.11, some secondary maxima are observed: around day 210 in F05 (December 1986), linked to the excess of soil moisture still present over the Guinean band; around day 180 (April 1987) and day 240 (May 1987) in F09, linked to the excess of soil moisture over the Atlas mountains and to the lack of soil moisture in equatorial Africa respectively; and around day 180 in F11 (May 1987) for the same latter reason. The spin-up time is 260 days for F05, 200 for F09 and 150 for F11.

We also chose three **dynamic variables**: specific humidity, air temperature and zonal wind. Again, their dependence on the lateral boundary forcings has to be noted here again. Indeed, after the initialisation, the model is fed every 6 hours with boundary conditions for the zonal and meridional winds, the air specific humidity, the potential temperature, the model pressure depth, the sea surface temperature, the leaf area index and the green leaf fraction. In order to distinguish this from the anomalies caused by the initialisation, we look at the closest level to the surface where these variables are likely to be mainly influenced by the surface state. Moreover, a main feature appearing in each pair of simulations (F05 and E05, F09 and E09, etc.) is likely to be due to the boundary conditions rather than to the initialisation.

For all three variables, the typical decrease with time is visible in figure 4.11. Local maxima in the specific humidity and air temperature maximum anomalies are obviously linked to anomalies previously identified for the soil moisture, the evapotranspiration and the surface temperatures. Figure 4.14, figure 4.15 and figure 4.16 give the monthly anomalies for these three variables. It is interesting to note the deficit in air moisture over the Sahara region.

Once the irrelevant secondary maxima are removed from figure 4.11 and the evolution of the monthly anomalies is taken into account, the spin-up time can be estimated for both air moisture and temperature for all simulations: 30, 90, 60, 270, 250, 250 days for E05, E09, E11, F05, F09 and

F11 respectively for the air moisture, and 30, 120, 120, 270, 240, 220 days respectively for the air temperature.

The case of the zonal wind is a bit more complex. Although maximum values range quickly within the internal variability, the spatial distribution of the zonal wind anomaly is different from all the previously analysed variables (see figure 4.16), but correlations can be made with the surface and air temperatures. In general, the E* simulations show weaker anomalies than the F* simulations. The anomalies are relatively constant in spatial distribution and tend to be located over the Sahara region, slightly to the south of the soil moisture dry anomaly. This link between both anomalies is probably false since it also appears in the F* simulations. The strongest anomalies are located out of Fouta Djallon, a region known for its high variability. The strongest anomalies do not occur in F05 and tend to suggest that their cause is linked to the start date of the simulations, i.e. towards the end or after the rainy season, a period which would favour their amplification (if existing), because E09 and E11 do not show such anomalies). The global spin-up time is estimated to be 140, 30 and 40 days for E05; E09 and E11 respectively, and 230, 140 and 130 days for F05, F09 and F11 respectively.

The spin-up values established here are summarised in table 4.2 and discussed in the next section.

4.5 Soil moisture influence

Table 4.2 summarises the spin-up lengths established on the basis of the results shown in figure 4.11 and adapted according to each variable specificity, together with the results of the statistical approach. Accuracy is to a maximum of 10 days on each number and even lower for the albedo values for the E* simulations (likely 50 days).

In terms of range of values, each group of three simulations is coherent within itself. For all surface variables except the soil moisture, the E* simulations balance faster with respect to their initial conditions than the F* simulations. They require around 3 months to remove the soil moisture

Chapter 4: Soil moisture and spin-up time

anomaly introduced at the initialisation. The F* simulations require at least double that time.

It is interesting to note that soil moisture is the unique surface variable to balance fastest. The water budget modifications at the surface take longer to reach the deepest soil layer. However, results provided by EXP2 indicate a higher internal variability which explains the quickest entrance of these results within the internal variability range. The weak values obtained for the sea-level pressure, while other atmospheric variables require a longer period, could be linked to the fact that the RMS errors were computed on a large area (the inner domain), thereby reducing the influence of local but spread anomalies.

	F05	F09	F11	E05	E09	E11	<i>mean E / mean F</i>
sea-level pressure	75	115	125	-	-	-	100/-
soil moisture (0-28.8 cm)	220	160	160	-	-	-	180/-
soil moisture (28.8-88.8 cm)	190	120	100	-	-	-	140/-
soil moisture (> 88.8 cm)	160	80	70	-	-	-	100/-
albedo	250	260	250	260	150	100	250/170
evapotranspiration	250	150	190	30	90	90	200/70
surface temperature	260	200	150	40	120	90	200/80
10-m specific humidity	270	250	250	30	90	60	260/60
10-m air temperature	270	240	220	30	120	120	240/90
10-m zonal wind	230	140	130	140	30	40	170/70
<i>mean per simulation</i>	220	170	170	90	100	80	
<i>start date of the simulation</i>	<i>May</i> 86	<i>Sep</i> 86	<i>Nov</i> 86	<i>May</i> 86	<i>Sep</i> 86	<i>Nov</i> 86	
<i>end of the global spin-up time</i>	<i>Jan</i> 87	<i>Mar</i> 87	<i>May</i> 87	<i>Aug</i> 86	<i>Dec</i> 86	<i>Feb</i> 87	

Table 4.2 – Estimations of the spin-up time in days for each simulation. Maximum accuracy of 10 days (likely 50 for the albedo in the E* runs). Mean 1 is a mean per variable and group of simulations (group divider = /), mean 2 is a mean per simulation (both are rounded to the closest ten).

Both F09 and F11 started towards the end of the rainy season and this particular condition significantly helped to remove the soil moisture anomaly

faster—they need almost 2 months less than F05. F11 even seems to be faster than F09 for a few variables, which is explained by the weaker initial soil moisture anomaly. Nevertheless, the results for F09 and F11 can not be analysed directly from January 1987, only from March or May of that year. If the point of interest is exclusively the rainy season of 1987, then these simulations are worth doing, with a preference for F09. If the interest is in the whole year 1987, only F05 is appropriate (or REF). Starting before the rainy season of 1986 helped to remove the anomalies in the northern part of the domain, but the rainy season postponed the evacuation of the soil moisture excess in the Guinean band until after the rainy season. This is why this simulation requires the longest spin-up time.

An apparently opposite behaviour is found for the E* simulations. E09 and E11 variables tend to require more time to balance than E05. This is linked to the strength of the wet soil anomaly. E09 had the strongest, followed by E11 and E05. However, compared to the F* simulations, these anomalies are weaker and disappear faster. The influence of the rainy season on E05 results is null as the anomaly has disappeared before rainfall arrives. Eventually, all three simulations are similar in terms of spin-up time, so that the rainy season of 1987 is relevant for analysis for all of them, but the whole year is only relevant for E05 and E09.

None of the variables tends to balance quicker than the other ones, apart from the soil moisture in the F* simulations. It should be remembered here that the EXP2 simulation could only provide us with a lower limit to the internal variability in the model. For some variables such as the albedo, air temperature and air moisture, the results seriously suggested that the internal variability is actually slightly greater. This taken into account, it is likely to reduce by one month the spin-up time estimated here.

Which method of soil moisture initialisation is then the most appropriate? All results point towards the method E using the ERA40 soil moisture fields to initialise the soil water content of the model MAR. This is only true if the dry soil anomaly initiated over the Sahara, and which maintains throughout the simulations as well as a wet soil anomaly deeper in the ground, does not

lead to other perturbations than those noted so far.

In terms of the atmospheric water budget, nothing was noticed apart from a weak dry anomaly of air moisture for a few months—not significant as it was inside the range of internal variability. In terms of radiative budget on the surface, the albedo is perturbed for a long period, but its anomaly is very weak across all the simulations. No severe anomalies were found either for the latent and sensible heat fluxes over the Sahara: the maximum anomaly is 10 W/m^2 for E09 and E11, and 20 W/m^2 for E05 over the first month only and is a negative (positive) anomaly for the sensible (latent) heat flux. Therefore, it is reasonable to conclude that the dry soil moisture anomaly does not cause any significant perturbation in the results even if it is maintained over the whole length of each simulation.

More generally, these results also suggest that starting the simulation in January of the year of interest with method F for the soil moisture initialisation (as has often been done) could lead to a spin-up time ending when the rainy season has already started, thereby introducing bias into the first months of results. On the other hand, starting a simulation in January of year Y-1 with initialisation of the soil moisture with method E could lead to double performance: both rainy seasons of years Y and Y-1 become relevant for analysis.

4.6 Conclusion

Seven simulations were performed in order to study the influence of soil moisture initialisation on the performance of the model and the length of the spin-up time, i.e. the time needed by the model to be independent of its initial conditions and to reach a dynamic equilibrium with respect to its driving fields.

Three simulations were realised for each of the two methods of soil moisture initialisation. Method F attributes a fixed value of soil moisture to all soil layers with spatial distribution influenced only by the soil texture. Method E attributes soil moisture values on the basis of the ERA40 soil moisture fields for the corresponding time of the year. The first method leads to global wet soil anomalies all over the domain, while the second leads to a

sustainable dry soil anomaly over the Sahara and a wet soil anomaly elsewhere.

The time taken by the model to succeed in removing the excess of soil moisture varies according to the time of the year and the initialisation method used. Starting the simulation before the rainy season permits quicker removal of most of the anomaly over the domain, but is not enough to remove the one over the Guinean band where the rainy season will considerably slow down the process of soil moisture evacuation and will eventually lead to the longest spin-up.

In terms of the dry soil anomaly presented by the E* simulations over the Sahara, the whole simulation length is insufficient to remove it. However, the results are acceptable very quickly—i.e. as soon as the wet soil anomaly has disappeared—because this anomaly does not lead to any substantial modification of the energy and water budgets of the surface and the lowest layer of the atmosphere.

In conclusion, method E eventually leads to a global spin-up time of 3 months, against 6 for method F. Ideally, a simulation with year of interest Y must be started during the year Y-1. If interest is focussed only on the rainy season of Y, it can be either initialised before the rainy season of Y-1 (i.e. before May) if using the method F for initialising the soil moisture, or at any month of Y-1 if using method E. If the interest is to study the whole year Y, then the same can be applied for method F, although the simulation must preferably be initialised in September the latest if using method E.

Some further studies could be carried on on how to improve both methods and reduce the spin-up time.

In the method F used in this chapter, we did not consider the possibility of changing of the percentage value of soil moisture used to calculate the soil moisture content per layer. 70 % leads to a wet anomaly across the whole domain. Increasing that value is aberrant since the soil moisture excess would take even longer to be evacuated. Decreasing this value seems more sensible. However, it is very likely to lead to dry soil anomalies (especially over the Guinean band and equatorial Africa): dry anomalies located there

Chapter 4: Soil moisture and spin-up time

could have a stronger influence than if located in the Sahara.

In the explanations of method E, we mentioned the use of the soil moisture of only two soil layers (the shallowest ones). The ERA40 actually provides soil moisture fields for four layers: 0-7 cm, 7-28 cm, 28-100 cm, 100-289 cm. Since these layers match closely the soil discretisation of SISVAT, it would be worth using the four of them to establish a correct correspondence between the proposed and modelled depths of soil. Perhaps doing so would avoid an overestimate of the soil moisture anomaly of the deepest soil layer of the model as it currently is.

Chapter 5

Validation of the new MAR

Purpose

The validation of a model is an essential step in its conception and use. In the context of climate simulations, it consists of checking how well (or badly) the model reproduces the climate of the studied region at different time and spatial scales. This step has to be achieved before any further study is carried out with the model in order to avoid an incorrect analysis of the results. The model MAR has been used to model the West African climate for more than five years. We used two different versions for this thesis, the MAR of 2002 and the MAR of 2006. While the former's climatology has been extensively validated for different climate situations, the latter's one has not. This is what we tackle in this chapter.

Contents

5.1	Introduction	114
5.2	Validation of the MAR2002	115
5.3	Validation of the MAR2006 (1987 and 1988)	119
5.3.1	Experimental setup and observational datasets	119
5.3.2	Rainfall	122
5.3.3	Dynamics	144
5.4	Conclusion: MAR2006 versus MAR2002	155

5.1 Introduction

In this thesis, we have used two different versions of the model. The first one, dated March 2002 (further referred to as the MAR2002 in this chapter), has undergone different modifications (see Moufouma-Okia, 2003, and Ramel, 2005, for example) and was used for the implementation of the new root formulation, the study of the internal variability and the soil moisture initialisation. This version has already been extensively validated. The second version is the latest and promising version dating from June 2006⁴³ (further referred to as the MAR2006 in this chapter). After all our modifications⁴⁴ were incorporated, it was used for obtaining the reference simulations of 1987 and 1988 as well as performing the sensitivity experiments to the vegetation forcings. Apart from the standard theoretical tests applied to any version, it has never been validated. This validation exercise inscribes itself in the continuity of the work previously achieved: improve the model and determine the optimal conditions as for its use (see previous chapters).

Many changes have been brought to the MAR2006 with respect to the MAR2002. These were initiated by the model results obtained over Antarctica. It is impossible to list them exhaustively, and it will be impossible to associate any result improvement/deterioration to one specific modification as they all interact. However, for the sake of completeness, there follows a list of the main modifications:

- The snow model has been coupled to the soil-vegetation-atmosphere transfer scheme in order to allow an efficient vectorisation of the code.
- The K-e scheme of Duynkerke used over Antarctica has been slightly modified to adopt the more classical scheme of Bintanja (2000) in order to be more coherent with the K-L scheme of Therrey and Lacarrère used over West Africa.
- The surface turbulence scheme has been recalibrated on basis of

⁴³ MARONH.0.1, 15/06/2006.

⁴⁴ See chapter 2.

observations in Dome C (Antarctica). In particular, the values of the stability parameters of the stability function have been reviewed.

- The conservation of water (all forms) has been reconsidered and improved.
- The microphysics of the icy clouds has been reviewed, notably the autoconversion scheme of Levkov *et al.* (1992).

The aim of this chapter is to see how the MAR2006 performs. The same approach as for the MAR2002 is applied, but with the possibility of improvement or correction of some of the systematic errors noticed in the MAR2002.

Section 5.2 sums up all statements of the validation of the MAR2002. Section 5.3 validates the MAR2006 in the context of the years 1987 and 1988. Section 5.4 concludes by comparing both versions. Particular attention will be given to the systematic model errors mentioned here above.

The West African climate has been explained in detail in the introduction of the thesis. Although most of its components and abbreviations are explained again in the text of this chapter, the reader is highly recommended to refer to the introduction for any unexplained item or need for more precision. The same suggestion applies to specific areas or spatial features of the West African region to which we shall refer continuously in the further analysis.

5.2 Validation of the MAR2002

The MAR2002 has already been carefully validated over West Africa in the last years, thanks to the work of the MAR team based in Grenoble, France.

Probably the pioneer for detailed studies of the West African monsoon with the MAR, Moufouma-Okia (2003) dedicated his PhD thesis to study the years 1987, 1988 and 1992. He looked in detail at the impact of the boundary forcings on the model results (ERA15 versus ERA40) and concluded that the MAR rainfall is very sensitive to the forcing large-scale data, but that the ERA40 tends to improve the results globally. He also

analysed the improvements of the convection scheme of Bechtold *et al.* (2001) with respect to that of Brasseur (1999), in terms of representation of the seasonal variability of the precipitation.

Then came Messenger *et al.* (2004) and Gallée *et al.* (2004). The former authors validated the model over 1983 and 1984 and then used the MAR to investigate the link between the sea-surface temperatures (SSTs) and the rainy season for both these years. The model could reproduce what is found in observations and the work helped to identify the physical mechanism at work with the SSTs: for example, warmer SSTs increase the evaporation flux and the meridional moisture transport in the lower atmosphere, leading to more abundant rainfall along the coast. The latter authors evaluated the model performance over the year 1992 and could propose, on basis of the model results, a mechanism to explain the oscillations of the rainy regimes (transition between two regimes of the Hadley cell).

Ramel (2005) went further into the validation process. During his PhD, he ran the model over 10 years (1983-1992) and validated very meticulously the rainy regime of the model. He also studied surface-atmosphere interactions over the year 1992, improving the representation of the surface properties in the model (albedo and vegetation). The work presented in the current thesis would not have been achievable without this important step. He finally looked into detail at the physics responsible of the abrupt northward shift of the monsoon late June and proposed a new mechanism for it (see Ramel *et al.*, 2006).

All of these authors meet on this conclusion: the MAR reproduces reasonably well the West African rainy regime. That is:

- The main **dynamical features** such as the Tropical Easterly Jet (TEJ), the African Easterly Jet (AEJ) and the westerly monsoon flow are modelled, are located reasonably well in latitude and altitude, and present behaviours in agreement with the observations (intensity, intra- and interannual variability). Both the thickness and the extension of the monsoon flow are captured.
- The main phases of the **progression of the monsoon** are correctly reproduced: 1) the onset in March on the coastal zone, 2) the installation

phase showing the reinforcement of rainfall until May, 3) the abrupt northward shift of the location of the maximum rainfall band, 4) the high rain period, and 5) the southward retreat from late September.

- The **rainfall spatial variability** exhibits maxima over the Fouta Djallon, the Cameroon highlands, the Jos and Bauchi Plateaus. With respect to observations, the model properly locates the rain band.
- The **rainfall temporal variability** is captured. The seasonal evolution is caught by the model (as mentioned here above) as well as the interannual variability (especially between dry and wet years). Ramel (2005) showed that rainfall produced by the MAR is in good agreement with observations for time scales larger than 3-5 days. At the mesh scale, the rainfall chronology is usually reproduced until a 10-15 day time scale.

In addition to these points, the model does not present any significant bias above 700 hPa. In terms of air humidity, this provides good confidence in the convection scheme which does not accumulate humidity at high altitudes like other RCMs do. The MAR also reproduces over the Guinean band a feature known as *dry corridors* (Poccard, 2001), i.e. alternation of rainfall and no-rainfall bands orientated south-north. As Messenger *et al.* (2004) explained, it is still unknown whether these are real or imagined by the model. Some sets of observations tend to show these corridors, but their intensity is strongly diminished by the usually coarse resolution of the data.

However, systematic errors were reported about the monsoon dynamics⁴⁵:

1. **Slight underestimation of the AEJ** (-1 to -2 m/s). Possible cause: the continental cold bias (see further) in the lower atmosphere leads to an underestimation of the north-south temperature gradient in the area of the AEJ.
2. **Too strong westerly monsoon flow** with respect to the ERA (+ a few m/s). Cause: a too strong deepening of the Saharian heat low associated with a more meridional localisation.
3. **Cold bias in the lower atmosphere** (under 700 hPa) over the continent

⁴⁵ Moufouma-Okia (2003) [1, 2]. Gallée *et al.* (2004) [1, 2, 3, 4].

(-2°C). Maximum biases found (i) to the east of the Fouta Djallon and (ii) to the south of the Aïr mountains. Causes: (i) a rainfall anomaly associated with a maximum of latent heat flux, (ii) a maximum optical depth modelled in that region as well as a minimum in the surface outgoing long-wave radiation. Both suggest an overestimation of the hydrological cycle in these areas.

4. **Cold and wet bias in the lower atmosphere** (under 700 hPa, -2°C and +0.5 g/kg) found when comparing MAR to radiosounding data in Niamey and Bamako in August. Possible cause: a too weak sensible heat flux and a too strong latent heat flux respectively.

... and errors in connection with the rainfall spatial and temporal distributions included⁴⁶:

5. **Slight general underestimation of the rainfall amounts.**
6. **Isohyet 2 mm/day too far north (20°N) and underestimation of the northward extent of the high rainfall area**, especially around the Senegalese coasts. **South-north rainfall gradient too strong** (this leads to underestimated rainfall in the Sahelian region).
7. **Overestimation of the frequency of light rain events and underestimation of the intensity of convective events.**
8. **Bad representation of the rainfall daily structure.** Main cause: item 7 above, since the number of rainy days and the length of the rainy season is fairly well reproduced by the MAR.
9. **Overestimation of rainfall during the retreat phase, belated rainy events** (over the Guinean band). Sometimes, the retreat rate is too slow. Possible cause: an overestimation of the latent heat flux partially due to an overestimation of the leaf area index at these latitudes.
10. **Underestimation of rainfall over the Atlantic Ocean.**
11. **No simulation of rainfall maxima over the coastal area during the installation phase** (May-June especially).
12. **Underestimation of rainfall maxima over the Fouta Djallon and the**

⁴⁶ Moufouma-Okia (2003) [3, 10]. Messenger *et al.* (2004) [6, 10]. Gallée *et al.* (2004) [1, 2, 4, 6, 8, 9, 10]. Ramel (2005) [4, 5]. Ramel *et al.* (2006) [3, 7, 9, 10].

Cameroon highlands. Maximum over Fouta Djallon extended further into the continent.

13. **Simulated maxima over the Jos and Bauchi Plateaus not found in observation.** Possible cause: lack of measurements in this area where mesoscale convective system events often occur.

5.3 Validation of the MAR2006 (1987 and 1988)

The MAR2002 has been proved to perform reasonably well over the West African region. The question of whether or not the MAR2006 has improved the results of the MAR2002 is addressed in this section.

5.3.1 Experimental setup and observational datasets

Two 2-year long simulations were performed with the MAR2006. The standard characteristics described in chapter 1, section 1.4 were used. In addition, a few parameters specific to these two simulations are summarised in table 5.1.

	a01	a02
period of simulation	01/01/1986 → 31/12/1987	01/01/1987 → 31/12/1988
boundary forcings	every 6 hours, ERA40	every 6 hours, ERA40
soil moisture initialisation	ERA40	ERA40
spin-up	1986	1987
NDVI for LAI	as observed	as observed

Table 5.1 – Additional information regarding the two simulations.

Different observational datasets are used to evaluate the model's performance. The dynamics is essentially compared to the ERA40⁴⁷ driving fields. Rainfall is compared to the dataset briefly described hereafter⁴⁸:

⁴⁷ The reader can refer to the chapter 1 for the description of the ERA40.

⁴⁸ Indications for download: CMAP available from www.cdc.noaa.gov (provided by the NOAA/OAR/ESRL PSD, Boulder, Colorado, USA); CRU from www.cru.uea.ac.uk/cru/data/hrg.htm; GSWP from www.monsoondata.org:9090/dods/gswp/grid; GTS from www.monsoondata.org:9090/dods/precip (created in the CPC of NOAA by P. Xie and J. E. Janowiak); IRD available upon request; NCEP from www.cgd.noaa.gov.

- The CPC⁴⁹ Merger Analysis of Precipitation (CMAP, Xie and Arkin (1996)) are 5-day and monthly analyses of global precipitation. They merge rainfall observations from rain gauges with estimates from several satellite-based algorithms. Their horizontal resolution is of 2.5° longitude/latitude. They cover the year period 1979-2001.
- The Climate Research Unit (CRU, New *et al.* (2000)) provides a dataset of monthly surface variables (precipitation, mean temperature, etc.) over global land areas, obtained from station observations. They cover the period 1901-1995 with a 0.5° longitude/latitude resolution.
- The Global Soil Wetness Project (GSWP) provides 3-hourly near-surface meteorological data based on atmospheric reanalyses (NCEP, GPCC⁴⁹ and GPCP⁴⁹ for rainfall) mixed with observational data. For rainfall, they cover the period 1982-1995 over land worldwide from 60°S to 90°N with an average 1° longitude/latitude resolution.
- The Global Telecommunications System (GTS) provides a dataset of global estimates of daily rainfall over land derived from rain gauge measurements. These data are closely related to the ISLSCP⁴⁹ Initiative II. They cover the period 1977-2001 with an average resolution of 0.5°.
- The French Institut de la Recherche et du Développement (IRD), the Agence pour la Sécurité de la Navigation Aérienne en Afrique et à Madagascar (ASECNA) and the Comité Inter-africain d'Etudes Hydrauliques (CIEH) have measured daily rainfall since 1950. They are available for the period 1950-1990 and include more than 1300 stations (between 700 and 800 during the 80's) over 3-20°N and 18°W-25°E. These data were krigged onto the MAR grid. They probably constitute the dataset of best quality available over West Africa. An additional dataset was obtained with a 10-day period resolution and extending further eastwards (courtesy of Véronique Chaffard). It was not krigged.
- The NCEP/NCAR⁴⁹ reanalysis project furnishes reanalyses of rainfall that we here used at a monthly resolution. They cover the whole world with

monsoondata.org:9090/dods/rean2d; soundings of the University of Wyoming from weather.uwyo.edu/upperair/sounding.html.

⁴⁹ See the list of abbreviations.

an averaged resolution 2° latitude/longitude.

- The Department of Atmospheric Science of the [University of Wyoming](#) provides radiosounding data for referenced stations worldwide. Over West Africa, only a few stations provide such soundings. The maximum time resolution is 12 hours and varies according to the availability of the measurements.

The CMAP, ERA40 and NCEP data are the only datasets covering the oceanic regions. The CMAP, GSWP, GTS and IRD datasets are particularly interesting for the high temporal resolution.

The scarcity of stations over West Africa is a major problem for the assessment of model results such as ours. Figure 5.1 illustrates this problem by showing the relative errors on rainfall in the CMAP dataset, the number of stations available for the GTS datasets and the localisation of the IRD stations. In the CMAP data, relative errors less than 20 % are only obtained for regions where the number of stations is not too small. The GTS plot clearly indicates the scarcity of stations over West Africa. The localisation of the IRD stations shows where data are not provided and where comparison with the model will be difficult.

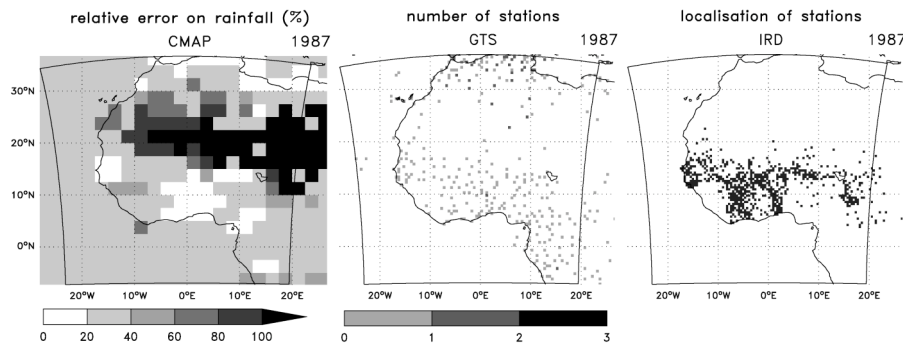


Figure 5.1 – Precision of rainfall datasets for 1987. The arrow-like bar indicates values may go beyond the indicated maximum.

All the datasets relying mainly on gauge measurements will be subject to unreasonable patterns of rainfall over regions with few gauges. The GTS indicates that Africa contains roughly 1300 stations while approximately

7500 stations exist worldwide—out of these 1300 stations, only 800 to 1200 report daily. This too small number of stations will be a serious handicap to the detection of local, regional, or even some large-scale hydrological and climatic phenomena.

5.3.2 Rainfall

a. Annual scale (spatial variability)

Cumulative annual rainfall for 1987 and 1988 is shown in figure 5.2 and figure 5.3 respectively. For both years, the spatial distribution of annual rainfall of the MAR is in rather good agreement with the observations: the continental rainfall band extends until $\sim 12^\circ\text{W}$ and presents the typical south-north decrease, and the maxima are located over the Fouta Djallon (10°W , 7°N), the Niger delta (7°E , 6°N) and the equatorial Africa (east of 10°E , south of 6°N).

Out of both years, **1987** is the best represented in the model (dry year over the Sahel). With respect to observations over land, rainfall sits in the correct range of values. The maximum over the Fouta Djallon reaches 2 m/year, in accord with the CRU and GSWP. The Ivorian-Ghanaian coasts receive more than 1.5 m of rain as in the CRU, GSWP and IRD. The maximum over the Niger delta is correctly located but not as extended (Cameroonian coast). Rainfall reaches 2.5 m versus 2 m in the CRU, GSWP and GTS. The MAR simulates more spatial variability over equatorial Africa than found in the observations. However, the coherence of the observations is less strong over this region and there is no consensus with respect to the localisation of the 1.5-m isohyet (for example). Nevertheless, the intensity of rainfall simulated by the MAR is in the same range as the observations.

In terms of the northern limit of the rain band, the MAR represents it reasonably well. As it is observed, the 1-m isohyet for example extends until $12\text{--}13^\circ\text{N}$ on the western part—albeit it is too far north over the southern Mali (8°W , 12°N)—and at approximately 10°N further eastwards. On the eastern part of the domain (eastern Nigeria, northern Cameroon), low rainfall is simulated. Although the IRD data only shows a narrow band of observations

at this location, it confirms rainfall is also under 1 m/year to the north of 9°N, a feature not visible in the other datasets with coarser resolutions.

1988 was a wet year for the Sahel. At first sight, this is missed by the model which tends to suggest that 1987 was wetter. The maxima are found at similar locations as in 1987 (Fouta Djallon, Niger delta, equatorial Africa). While those over the Ivorian-Ghanaian coast and the Niger delta agree with observations (same remarks as before for the latter one), the other maxima are underestimated. This is particularly clear for the Fouta Djallon region (1.5 m/year versus 2 to 2.5 m/year).

The northern limit of the rain band is too far south. The 1-m isohyet is similar to observations over the Mali, but the agreement is much worse in the Guinean band in general. It is found at around 8°N over the Ghana and Benin while observations locate it around at least 10°N. The rain band seems not to have moved northwards enough, a feature also noticeable above the oceanic regions. Versus the CMAP, the 1-m isohyet in the MAR extends a few latitude degrees more southwards.

The GTS dataset shows a *dry corridor* over the eastern boundary of Ghana (0°E) for both 1987 and 1988. The MAR shows an attempt of reproducing this feature each year. These corridors appear at a location which corresponds to Lake Volta, situated in a valley, at the Ghana-Togo boundary. Does this explain why MAR captures them? Probably not. The lake is not modelled, it only appears through the topography and a vegetation anomaly on a small region. Are they real in the observations? They are not located in a region with a good spatial coverage of gauges (see IRD on figure 5.1). As we mentioned it, the observations are likely to be biased by the grid interpolation algorithms used to construct them. Hence, artefact or not? The question stays open. A possible direction of investigation would be to run the model over this region with a finer resolution and see whether the corridor appears again.

With respect to MAR2002, the most striking detail is probably the presence of rainfall over the oceanic regions, a feature which was almost totally absent in the MAR2002. A westward rain band is located off Fouta Djallon for both years and is found in the NCEP, CMAP and ERA40. Its

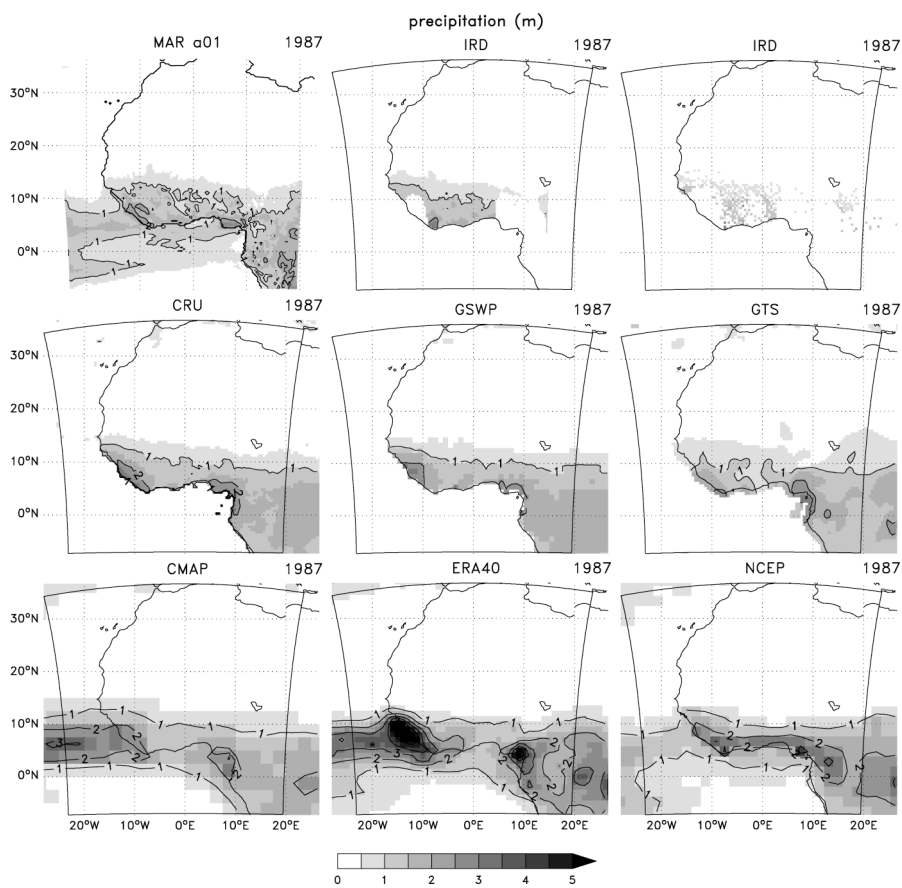


Figure 5.2 – Annual rainfall of the MAR and observations for 1987 (m). The muffin-shaped polygon around each plot represents the MAR domain. The arrow-like bar indicates values may go beyond the indicated maximum.

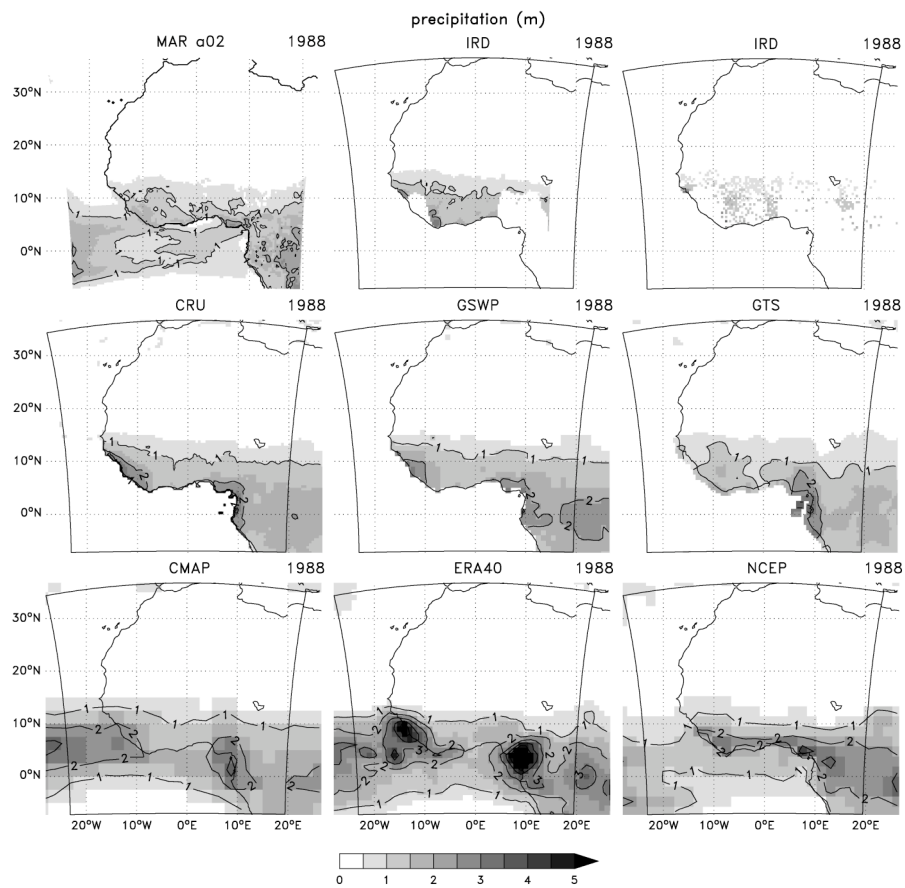


Figure 5.3 – Annual rainfall of the MAR and observations for 1988 (m). The muffin-shaped polygon around each plot represents the MAR domain. The arrow-like bar indicates values may go beyond the indicated maximum.

intensity is underestimated with respect to the latter two datasets (1 to 2 m/year) but in the same range of values with respect to the NCEP. As in the ERA40 and NCEP, another rain band is present around 3°S. The coarse resolution of these three datasets is likely to prevent assessment of the MAR rainfall low in the Bight of Benin (2°E-5°N).

To sum up, the annual repartition of rainfall modelled by the MAR is in good agreement with the observations. The extension of the rainfall band is well reproduced and the maxima are located where expected, and this is for both years. The intensity of rainfall is well caught for the year 1987, but strongly underestimated for the year 1988.

b. Seasonal scale

In order to investigate the evolution of rainfall at a finer temporal resolution, we consider three different kinds of graph. The first kind is the spatial distribution of rainfall averaged per month (figure 5.4 to figure 5.7), which will help to determine whether cumulative annual rainfall previously analysed is constructed on basis of the expected seasonal evolution of rainfall. The second kind is a Hovmöller graph (figure 5.9). This represents the 10-day averaged rainfall over 10°W-10°E. It allows the observer to go under the monthly time scale and look into more detail at features of the monsoon such as the abrupt northward shift. The third kind is the time series of daily rainfall averaged per region of interest (figure 5.8). The time scale is even finer and it will help to characterise and discriminate where the model performs well (or badly).

Monthly rainfall

Monthly rainfall modelled by the MAR is compared to the GSWP (cf. figure 5.4 to figure 5.7). For both years, the progression of the rain band northwards then southwards is clearly visible. The year starts with rainfall centred around the equator. The band progressively moves northwards and reaches the continent (*installation phase*) before settling in the Guinean band in April-May (*first rainy season*). In June, it *shifts abruptly northwards* and extends then much further into the continent (*high rain season*). This is synonymous with a *short dry season* for the regions along the coast. The

band remains at this location until September and finally *retreats southwards* until the end of the year (*dry season*).

During the year **1987**, the agreement in band position and rainfall intensity is reasonably good. In January, the rain band is located over equatorial Africa but seems a few degrees too far north (see the 2-mm isohyet). Low rainfall is unexpectedly present on the coasts. During the *installation phase* and *first rainy season* (February to May), the model matches the observation. The rainfall intensity is in very good agreement with the GSWP, except for a slight overestimation in February on the coasts, more spatial features over equatorial Africa in general, and an underestimation by a factor 2 in May in the Guinean band. According to observations, the *northward shift* (June) has pushed the location of the maximum rainfall band over 8-10°N, rainfall intensity has increased and the 2-mm isohyet reaches 16°N (which is slightly too far north). Noteworthy is the improved agreement over equatorial Africa. During the *high rain period* (July to August), the 2-mm isohyet is well located at 16-17°N, maxima are found over the Fouta Djallon, Niger delta, Bauchi and Jos plateaus (see remark at section 5.2 for the latter two), and over eastern Nigeria with intensities close to observations in July but slightly weaker in August. The *short rainy season* is visible along the coast. The *retreat phase* (September to December) is slower (or later) in the model (for example, rainfall over equatorial Africa is already more spread southwards in the GSWP in August than in the MAR). There is a lag of at least one month over the Guinean band while agreement is better to the east of 10°E. Rainfall is hence maintained on the coast for too long a time.

The same conclusions can be drawn at first sight for **1988**. However, the global lack of rainfall is striking: while the rain band progresses according to the GSWP (the 1-m isohyet is well positioned each month), rainfall is underestimated almost everywhere. This is likely to be responsible for disappearance of unexpected rainfall on the coasts at the beginning of the year (see February for example) and for the attenuation of the excess of rainfall during the retreat phase. The region suffering this problem the least is equatorial Africa to the east of 10°E, where the agreement with the

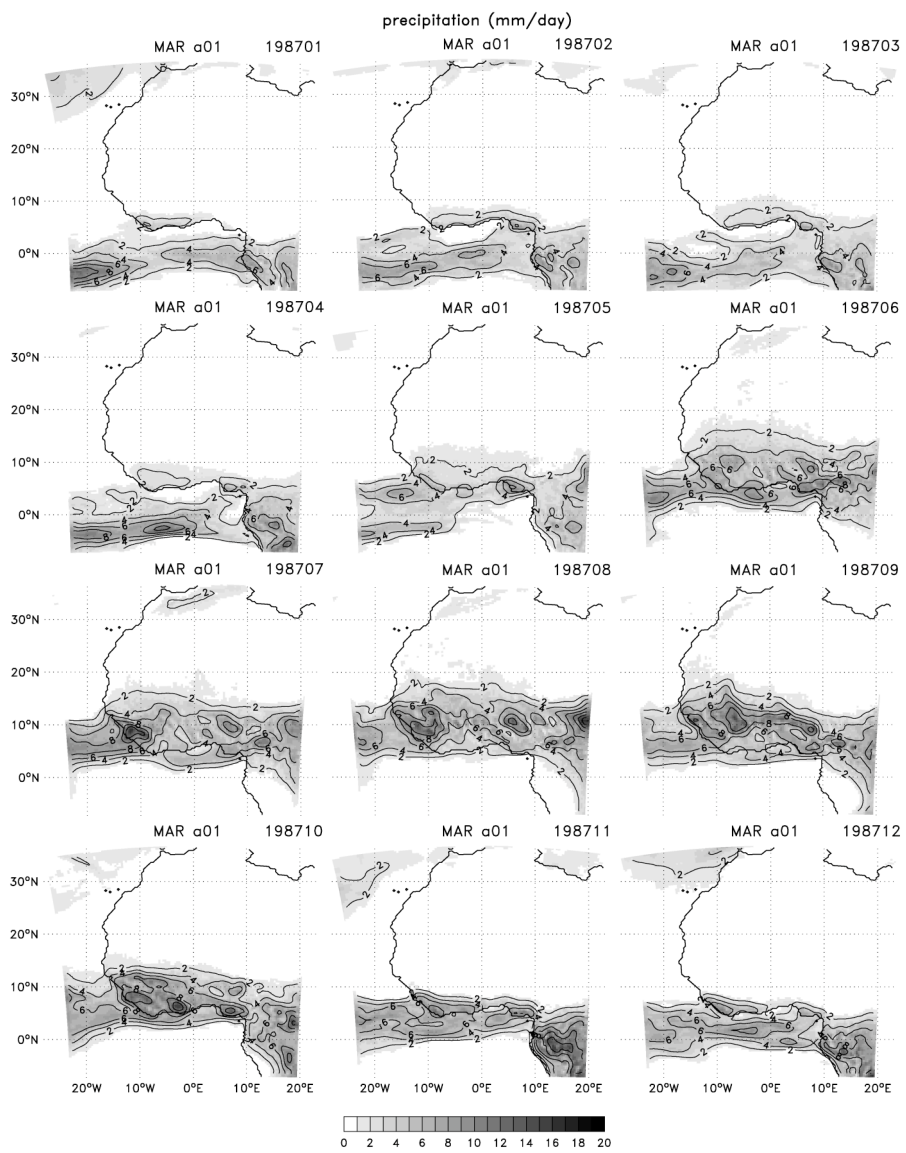


Figure 5.4 – MAR monthly rainfall for 1987 (mm/day).

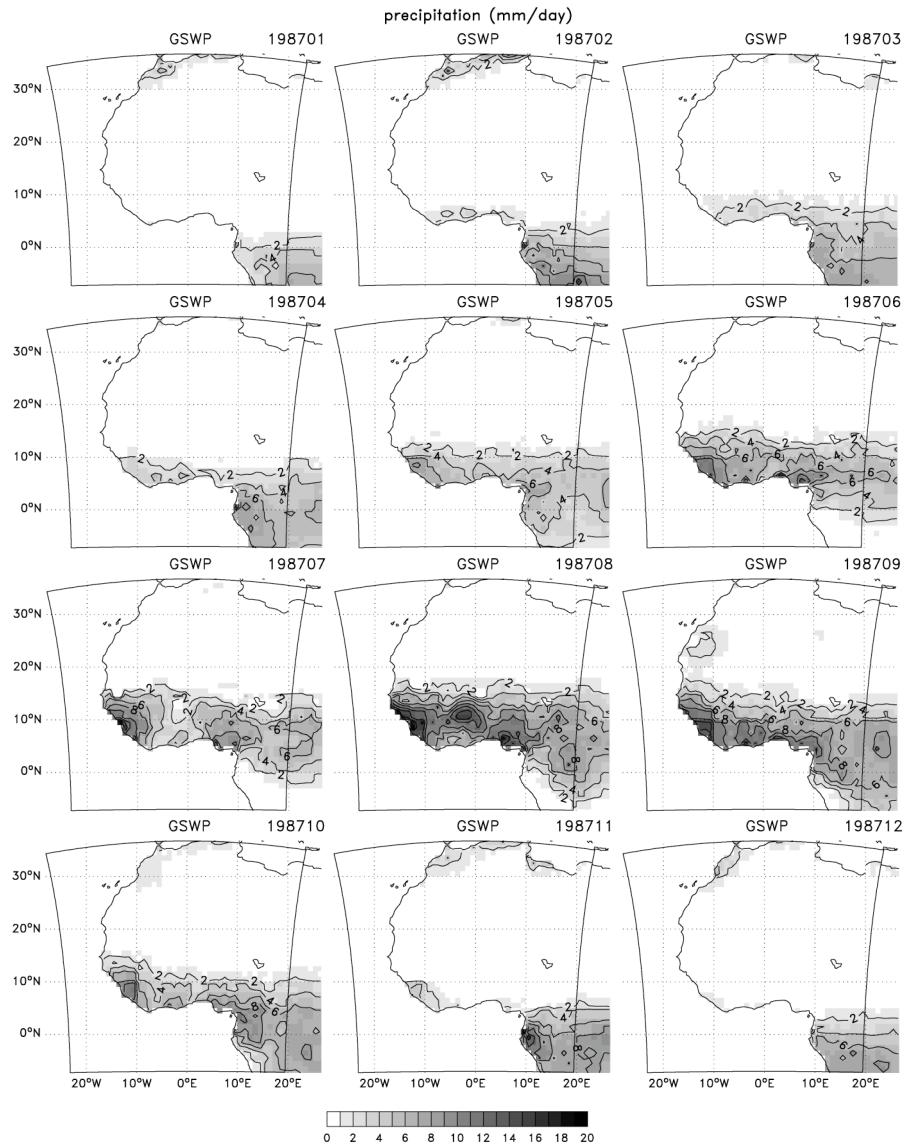


Figure 5.5 – GSWP monthly rainfall for 1987 (mm/day). The muffin-shaped polygon around each plot represents the MAR domain.

Chapter 5: Validation of the new MAR

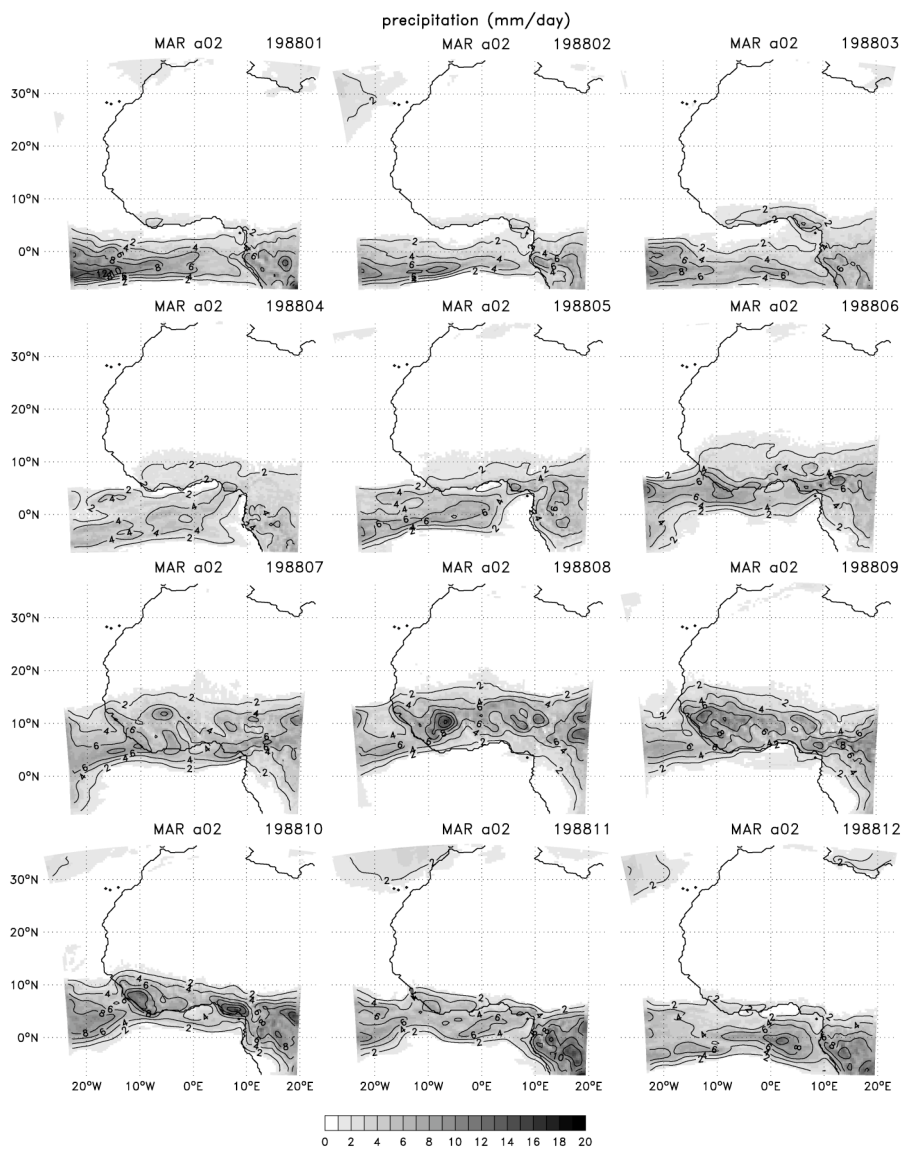


Figure 5.6 – MAR monthly rainfall for 1988 (mm/day).

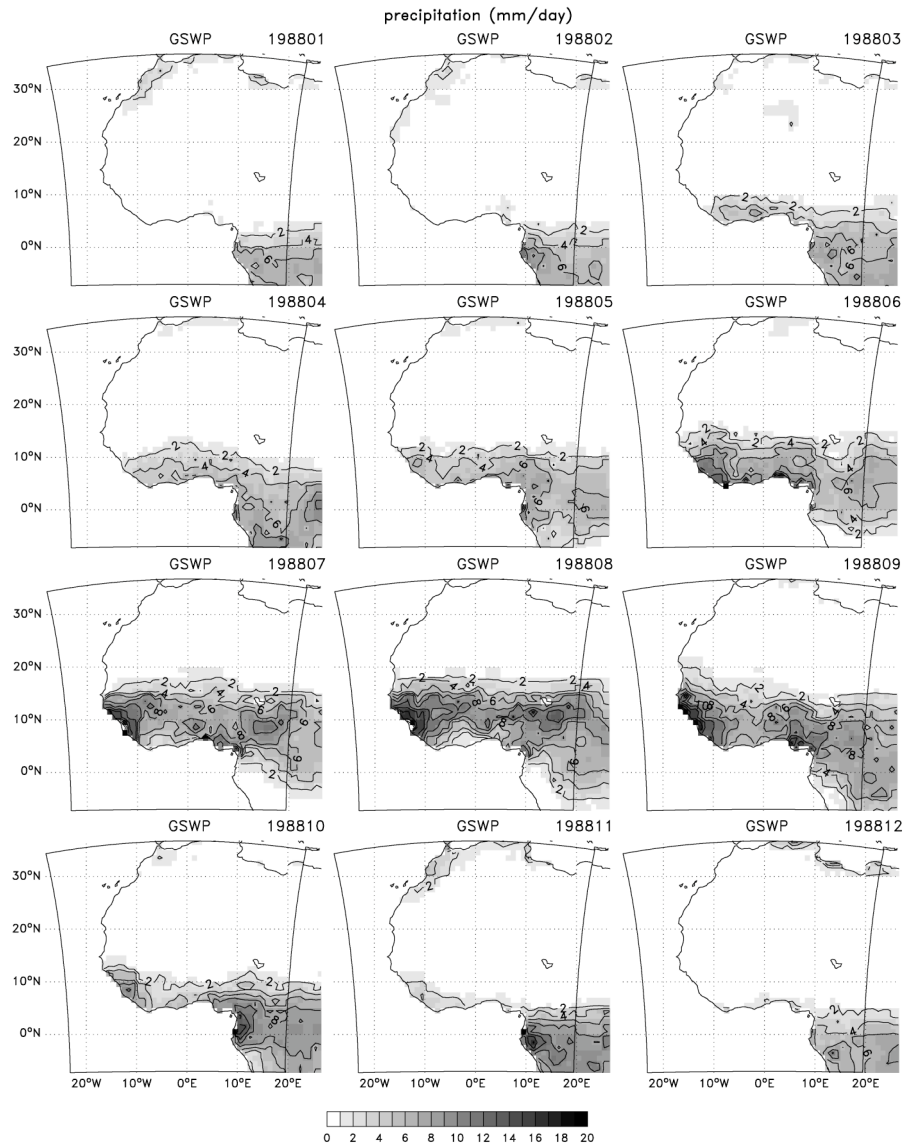


Figure 5.7 – GSWP monthly rainfall for 1988 (mm/day). The muffin-shaped polygon around each plot represents the MAR domain.

observations is good in terms of rainfall intensity.

To sum up, the temporal and spatial progression of rainfall is caught by the MAR and confirms that rainfall leading to good cumulative annual rainfall fell generally at the right period of the year. Points of particular interest, apart from the lack of rainfall in 1988, include: the slightly too early start of rainfall on the coast, the good coherence afterwards, the northward shift reproduced in June as observed, the good position of the 2-mm isohyet in general, and the southward retreat starting on time but appearing too slow and leading to an overestimation of rainfall on the coast.

10-day averaged rainfall

The Hovmöller graph of 10-day averaged rainfall, averaged over 10°W-10°E is shown in figure 5.9 for 1987 and 1988. Its interest is that it easily shows the temporal evolution of rainfall together with its latitudinal extension. The comparison is made with daily IRD rainfall regridded onto the MAR grid. In order to relativise the quality of the MAR results with respect to the IRD data, rainfall produced by the ERA40 is shown but will not be discussed.

For both years, MAR rainfall agrees reasonably well with the IRD data. The *installation phase* extends from February (March) to early May in 1987 (1988) on the coastal areas (6-8°N) with slightly weaker rainfall amounts (no peak to 4 mm/day as in the IRD). From May to late June, rainfall increases (*first rainy season*) at these latitudes, reaching 6 mm/day (still drier than the IRD showing until 10 mm/day). The rain band *jumps northwards* in June and stabilises at 10°N approximately (12°N in 1988) during July and August. Maxima of rainfall are closer to observations, the model simulating until 12 mm/day as the IRD. The *southward retreat phase* starts in September, but is characterised by rainfall maxima absent in the observations.

Looking at the maximal latitudinal extension of the rain band, the MAR is satisfying as it represents correctly the month-to-month variation, except during the retreat phase. For 1987, the 1-mm isohyet also oscillates between 8 and 10°N from February to April, then advances progressively until 16°N (June) and further (IRD data not available), and is back to 14°N in late September. From that moment, there is a lag of two degrees in latitude

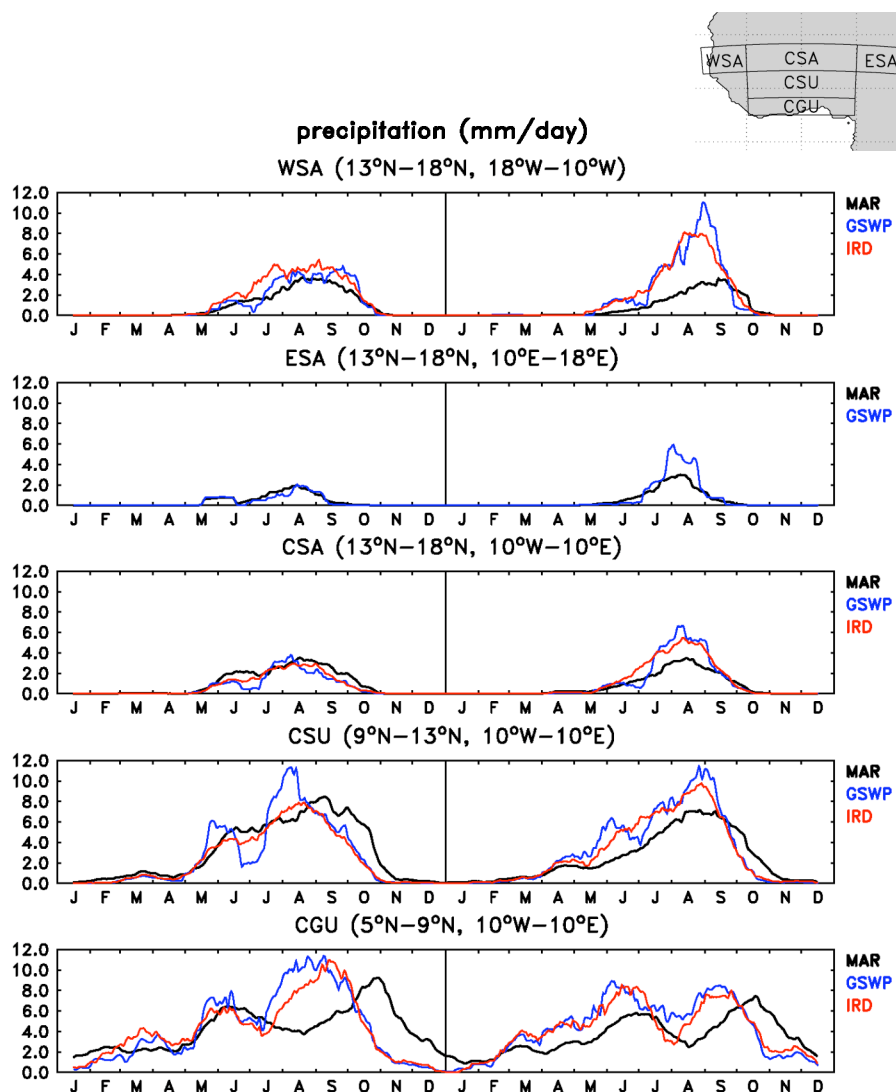


Figure 5.8 – Temporal series of daily rainfall per region (mm/day) : MAR (black), GSWP (blue) and IRD (red) when available. Running average with a window of 30 days.

Chapter 5: Validation of the new MAR

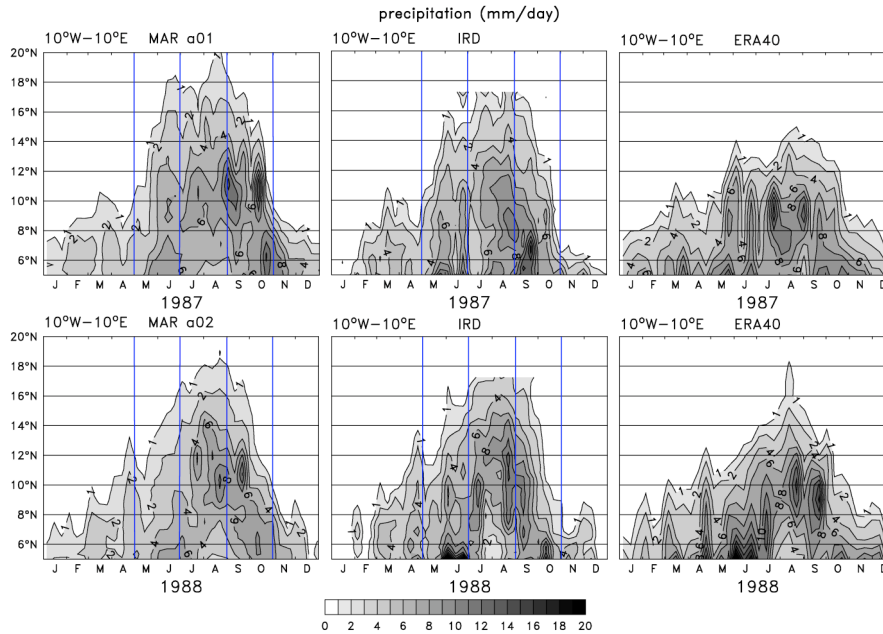


Figure 5.9 – Time-latitude graphic of the 10-day averaged rainfall (mm/day) averaged over 10°W-10°E. MAR versus IRD and ERA40.

between the MAR and the IRD, a lag which the model can not subsequently catch up. Indeed, in November-December, the retreat slows down similarly to the IRD, but the 1-mm isohyet is still located too far north (8°N instead of 6°N). The lag seems to be introduced by a strengthening of rainfall in September-October at 11°N. Although the width of the rain band has started to reduce (see the 1-mm isohyet), its hard core of rainfall has not. For 1988, the same statements can be made with respect to the period extending between January and September—notice this time that although the 1-mm isohyet is correctly located, the 2-mm isohyet is systematically too far south due to the general lack of rainfall that year in the model. The lag in the retreat is reduced (1 degree of latitude), for the same reason as before: the hard core of the rainfall band (11°N) only starts to retreat from the end of September. The lag being less strong that year, the model is actually able to catch up with it and reproduces correctly the last two months of the year.

The contrast between both years—1987 was a dry year over the Sahel

while 1988 was a wet year—is also apparent: in July-August, the maximum rain band is located further north in 1988 (12°N) than in 1987 (10°N). The 2-mm isohyet of the model is even nicely located at 16°N during that period in 1987.

The analysis of rainfall averaged per 10 days has helped to refine the previous statements obtained with the annual and monthly time scales. The model correctly reproduces the temporal and spatial evolution of rainfall as well its northward extension. Apart from the delay in the retreat phase—the hard core of rainfall does not start to retreat southwards on time—, the installation phase, first rainy season, abrupt shift northwards and high rainy season are in good agreement with observations. The correct reproduction of these features for the year 1988 suggests that the systematic lack of rainfall that year is linked to a problem with the hydrological cycle rather than the dynamics. This will be investigated further in this chapter.

Regional scale

The figure 5.8 helps to illustrate the MAR results from another perspective. Daily rainfall is shown for five regions of interest (detailed hereafter). A running mean with a 30-day window was applied to remove the high frequency signal. Observations shown are the IRD and GSWP datasets.

The model agrees with the main features of the rainfall cycle. From south to north, the rainy season is divided into two rain periods to the south of 9°N. Further north, the rainy season is unique and shortens in length as the latitude is more northern. From west to east, the rainfall cycle is progressively less strong.

Over the Western Sahel (WSA, ~Senegal), the agreement is good in 1987 with rainfall starting on time and showing the same intensity as the observations. In 1988, both intensity and cycle are poorly caught. Rainfall peaks one month later than observed and the MAR misses abundant rainfall (factor 3). The same statements apply to the Eastern Sahel (ESA, around Lake Chad) and the Central Sahel (CSA, ~southern Mali and Niger). Notice here the much shorter rain period over ESA. The rainfall underestimation in 1988 is not as marked as over WSA (factor 1.5-2).

Over the Central Sudan (CSU, Burkina Faso, northern Benin and

Nigeria), the agreement between the model and the observations is less good. The rain period starts on time but ends later as expected both years. Albeit the rain peak is 1-month lagged in 1987 (September instead of August), it is appropriately situated in 1988 (end of August). The rain amount matches observations (IRD) for 1987 but is underestimated in 1988 (factor 1.5).

Over Central Guinea (CGU, coastal regions of Ivory Coast, Ghana, Togo, Benin, Nigeria), the same particularities are found. The rainy season starts on time and progresses similarly to the observations until June—disregarding the known underestimation of rainfall in 1988. The evolution is then progressively delayed. After the first rain period, rainfall continues to decrease and reaches an amount very close to the observations, but the process lasts until mid-August instead of mid-July. Rainfall then again increases for around two months as in the observations and reaches a maximum in good agreement with the amount observed. The 1-month lag is still present but not enlarged. Rainfall finally decreases. As we observed on the Hovmöller graphs, the retreat speed is similar in the MAR and the observations.

Some additional information is provided by this regional analysis. Over the Sahel, the rainfall evolution is well reproduced in the model for the year 1987 with cycle and intensities matching the observations. Over the Sudanian and Guinean band, the intensities are quite well estimated both years, even if the cycle suffers from the retreat phase problem. Indeed, these regions suffer less than the Sahelian ones from the lack of rainfall in 1988.

c. Daily scale

We investigate here daily rainfall simulated by the MAR in order to gather more information on the possible causes of biases previously found.

The table 5.2 summarises information about the rainy season of the five regions we are considering (see previous subsection for details). For each of the regions are indicated cumulative annual rainfall (C), the beginning date (B), end date (E) and length (L) of the rainy season. These are estimated on the basis of daily rainfall. The beginning and end dates are calculated following the method used by Ramel (2005) who investigates with much

detail the daily precipitation produced by the MAR2002. The rain period is said to start when cumulated rainfall reaches 2.5 % of the annual cumulation (of the region of interest). It is said to finish when 97.5 % of the annual cumulation is reached. The length is the difference between both dates.

Cumulative annual rainfall confirms what we previously found. The year 1987 is not badly estimated and ends up to be globally wetter than 1988, unlike the observations. In 1987, the model is found to be in very good agreement with the IRD over the CGU, then wetter than observed over the CSU, CSA, and ESA. Over the WSA, this tendency is reversed. In 1988, the annual amount of rainfall is underestimated over every region.

The *length of the rain period* in the model is usually in good agreement with the observations (except over the CGU and CSU where it is 1.5 times longer) with around 100 days over the Sahelian regions and from 200 to 300 days over more southern regions. Even if it rains less in 1988, we have here another confirmation this is not due to a shorter rain period. The MAR actually tends to simulate a rain period longer than observed (the difference ranges between 5 and 57 days with respect to the IRD). The largest differences are obtained over the CGU and CSU. These regions are characterised by too early rainfall and a retreat phase delayed by one month, as stated previously. The large differences obtained for CSA (and ESA) in 1988 are explainable by the shape of rainfall that year (see first months on figure 5.8). Indeed, rainfall in the model is very close to observations, but constant and slightly greater than observed.

The modelled rain period *begins* earlier than observed. Over the Sahelian regions, the difference ranges between 4 and 7 days with respect to the IRD observations—except the CSA and ESA in 1988, for the reason given previously. Over the CGU and CSU, it ranges between 10 and 28 days. Indeed, we saw that rainfall appears in January and February along the coast, unlike the observations.

The modelled rain period *ends* later than observed. The difference between the model and the IRD ranges between 4 and 17 days over the Sahelian regions and between 7 and 36 days over the CGU and CSU. The greatest differences occur over the CGU as expected.

		1987				1988			
		C	B	E	L	C	B	E	L
WSA	MAR	337	3 Jun	17 Oct	137	259	10 Jun	12 Oct	125
	IRD	511	7 Jun	12 Oct	128	567	3 Jun	30 Sep	120
	GSWP	409	4 Jun	11 Oct	130	574	4 Jun	30 Sep	119
ESA	MAR	125	31 May	19 Sep	112	162	5 Jun	14 Sep	102
	GSWP	107	30 May	2 Sep	96	229	19 Jun	4 Sep	78
CSA	MAR	343	26 May	12 Oct	140	271	28 Apr	30 Sep	156
	IRD	263	31 May	8 Oct	131	400	3 Jun	24 Sep	114
	GSWP	239	24 May	2 Oct	132	385	2 Jun	25 Sep	116
CSU	MAR	1083	17 Mar	28 Oct	226	865	16 Mar	28 Oct	227
	IRD	865	27 Mar	13 Oct	201	1030	3 Apr	8 Oct	189
	GSWP	959	2 Apr	10 Oct	192	1110	13 Apr	2 Oct	173
CGU	MAR	1523	27 Jan	15 Dec	323	1256	30 Jan	6 Dec	312
	IRD	1490	17 Feb	9 Nov	266	1461	27 Feb	29 Nov	277
	GSWP	1543	17 Feb	2 Nov	259	1529	29 Feb	21 Nov	267

Table 5.2 – Cumulative annual rainfall (C, mm), beginning (B), end (E) and length (L, days) of the rainy season per region, computed on the basis of daily rainfall of the MAR, IRD (when available) and GSWP.

Hence, the model MAR simulates a rainfall cycle usually in good agreement with the observations. This gives confidence in its dynamics. In order to investigate more deeply the lack of rainfall usually observed (especially for 1988), we need to look at the types of rainfall the model generates.

In table 5.3 are given the rainfall classes for each of the five regions and for both years: number of days with no rainfall, with rainfall less than 1 mm/day (not null), with rainfall between 1 and 10 mm/day, with rainfall between 10 and 20 mm/day, and with rainfall above 20 mm/day. Data are restricted to the rainy season as determined in table 5.2, and the model is compared to the same observations as before. For each region and each class, the rainfall anomaly (MAR versus IRD) is indicated (*italic numbers in the Δ rows*). The total anomaly (i.e. all classes gathered) is indicated in the first cell of these rows, and the percentage each class anomaly contributes to

the total (absolute) anomaly per region is given in the % rows. The figure 5.10 presents these classes expressed in percentage with respect to the length of the rainy season as well as the average rainfall associated with each class (numbers above the bars). Because of the general weak agreement between the GSWP and IRD data at these time scales (except in terms of average rain amount per class), the GSWP data seem not very reliable for our current purpose. For this reason, the ESA region is shown for information only since its comparison with observations is probably biased and will not bring any reliable information. We shall here consider each class separately.

Null rainfall occurs in the model for none of the regions. This is due to the spatial average performed on the data before calculating the classes. This average weakens the extrema (null/high rainfall). The model results are very close to the IRD data. The GSWP shows numerous days for this class, unlike the model and the IRD.

Not null rainfall (weaker than 1 mm/day) are simulated with less precision. The number of days and percentage with respect to the rainy season are good only for the WSA in 1987 and the CSU for both years. However, the average rain amount for this class is usually in good agreement with both IRD and GSWP, and the absolute total anomaly with respect to the IRD does not exceed 8 mm (18) for the whole rainy season in 1987 (1988).

Rainfall between 1 and 10 mm/day shows the strongest departure to the observations, along with the next class. The number of days is overestimated almost everywhere, the percentage of the rainy season that this represents is variable, and the average rainfall amount for the class with respect to the IRD observations is only good in 1987 (except over the WSA—too weak rainfall). In 1988, the latter is generally underestimated for all regions. These all lead to strong rainfall biases ranging between -109 and +463 mm over the rainy season. The strongest bias is observed over the CGU.

Rainfall between 10 and 20 mm/day behaves similarly to the previous class, i.e. the very strong departures with respect to the observations are found. The model underestimates the number of days over all regions, the percentage that this represents is also too weak. However, the intensity of the

		1987						1988					
		C	0	<1	1-10	10-20	>20	C	0	<1	1-10	10-20	>20
WSA	MAR	320	0	40	96	1	0	247	0	57	65	3	0
	IRD	493	1	43	73	11	0	541	4	26	72	17	1
	GSWP	402	44	33	41	8	4	552	33	27	41	11	7
	Δ	-173		8	-49	-132		-294		18	-109	-182	-21
	%			4	26	70				6	33	55	6
ESA	MAR	121	0	77	35	0	0	155	0	54	47	1	0
	GSWP	104	50	24	20	2	0	218	20	30	21	4	3
CSA	MAR	328	0	27	113	0	0	258	0	61	95	0	0
	IRD	251	0	45	86	0	0	382	0	21	90	3	0
	GSWP	230	32	53	41	6	0	370	16	34	58	6	2
	Δ	76		-5	81			-124		11	-100	-35	
	%			6	94					8	68	24	
CSU	MAR	1029	0	33	179	14	0	823	0	34	185	8	0
	IRD	823	3	40	144	14	0	981	0	26	134	28	1
	GSWP	913	9	64	94	18	7	1052	2	39	95	29	8
	Δ	205		1	203	1		-159		8	94	-239	-22
	%			0.5	99	0.5				2	26	66	6
CGU	MAR	1449	0	10	303	10	0	1195	0	16	292	4	0
	IRD	1415	0	25	203	36	2	1390	0	42	197	33	5
	GSWP	1477	0	61	149	37	12	1458	0	43	187	29	8
	Δ	34		-8	463	-368	-53	-195		-10	330	-391	-124
	%			1	52	41	6			1	39	46	14

Table 5.3 – Rainfall classes per region, expressed in number of days of the rainy season for the MAR, IRD (when available) and GSWP. The columns C give cumulative annual rainfall for that period. The Δ rows give the MAR rainfall anomaly⁵⁰ per class with respect to IRD (mm); the total anomaly is indicated in the first cell. The % rows give the percentage the class anomaly contributes to the whole (absolute) anomaly. See text for details on the regions.

⁵⁰ Per class: number of days (MAR) * average rainfall (MAR) – number of days (IRD) * average rainfall (IRD).

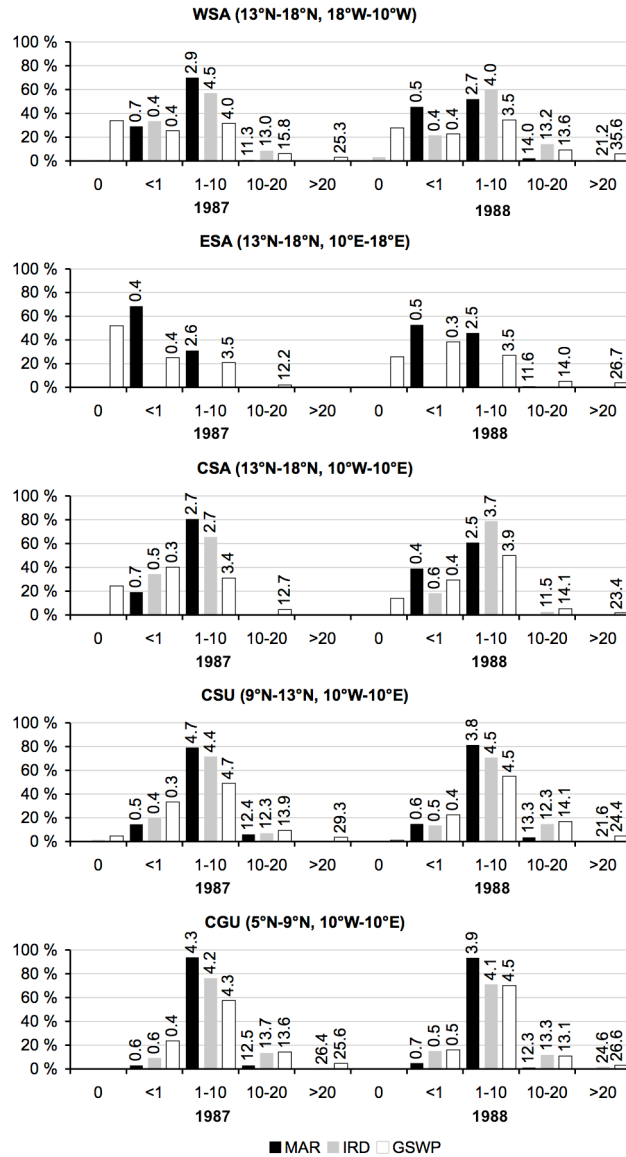


Figure 5.10 – Rainfall classes expressed as a percentage with respect to the length of the rainy season, for the MAR, IRD (when available) and GSWP. Classes are 0, < 1, 1-10, 10-20 and > 20mm/day. Numbers above bars indicate the average rainfall per class (mm/day). See text for details on the regions.

rainy events (when occurring) is not badly captured. Albeit this class represents a lower percentage of the rainy season than the previous class, the final bias obtained is as strong: the class anomaly ranges between -391 and +1 mm, values sometimes correcting the positive bias due to the previous class.

Rainfall above 20 mm/day is not present in our results (same remark as for the null rainfall: extrema are weaker). The agreement with the IRD is less good than for the null rainfall. For these observations, there are more days with such high rainfall (until 26.4 mm/day), so the bias introduced is likely to be strong. The anomalies range between -21 and -124 mm over the rainy season.

The *total anomaly for each region* can be great (see Δ and % rows in table 5.3). Even if the class anomalies may compensate each other to lead to less strong bias of rainfall, the total rainfall anomaly per region is mainly caused by the mismodelling of the light rainfall (1-10 mm/day) and high rainfall (10-20 mm/day). In 1987, the first class dominates the second one over all regions except the WSA: 52 % of the anomaly over the CGU, 99 % over the CSU, 94 % over the CSA. This leads to positive rainfall anomalies. Over the WSA, the second class dominates (70 %) and strengthens the negative anomaly of rainfall started by the first class. In 1988, neither of the classes really takes over the other one and the global rainfall anomaly is negative over each region. The first class is only more important over the CSA (68 %) and is overtaken by the second in the other regions (39 versus 46 % over the CGU, 26 versus 66 % over the CSU and 33 versus 55 % over the WSA).

d. Summary

The MAR is able to reproduce the main features of the rainy regime both in time and space for both years 1987 and 1988.

The spatial repartition of rainfall is indeed in very good agreement with the observations. The maxima are found over the Fouta Djallon, the Cameroon highlands, the Jos/Bauchi plateaus, the delta of the Niger river, and over equatorial Africa, with intensities generally close to observations

(disregarding the general lack of rainfall for the year 1988). The northward extension of the rain band is especially well modelled regarding the isohyet 1 mm/day.

The temporal evolution of rainfall is also captured by the model. From January to September, the model performs respectably well, representing the installation phase of the monsoon, the first rainy season, the abrupt shift northwards and the short dry season, the second and high rainy season, and finally the retreat phase. Points to notice include: 1) a too early start of rainfall (light) over the coastal area in January and February, 2) an underestimation of the rainfall over the Guinean band later in the first rainy season, 3) a retreat phase marked by a too late southward move of the hard core of the rain band (it persists in September-October), although lower rainfall retreats on time and similarly to observations. This introduces a lag of 2 (1) degrees of latitude in 1987 (1988) in the position of the rain band, so much so that the model can hardly catch up. This occurs essentially over the Guinean band west of 10°E. Further eastwards, the problem does not appear and further south, over equatorial Africa, the retreat phase simply starts later.

All of this leads to good or bad representations of the rainfall cycle when considered by regions. Over the Sahelian band, the cycle and intensities are well represented, while the cycle shows significant time lag (due to the latitude lag) over the Sudanian and Guinean regions. The intensity of rainfall over the latter two regions is, however, quite well estimated. This is valid essentially for 1987, since the intensities are systematically underestimated in 1988 (with the problem actually more marked over the Sahelian regions).

The mis-evaluation of rainfall is not due to a problem in simulating the length of the rainy season. Indeed, the length is correct and even tends to be longer than observed. It usually starts earlier and ends later—around one week over the Sahel, and between one week and one month over the Guinean and Sudanian regions (because of the retreat phase problem).

Actually, rainfall anomalies are essentially due to two classes of rainfall: light rainfall (1-10 mm/day) and high rainfall (10-20 mm/day). The highest rainfall class (over 20 mm/day) also plays a role, but the global rainfall negative anomaly so introduced is weaker than the previous two classes.

In 1987, the number of days with light rainfall is overestimated. As the average rainfall amount for this class is well represented by the model, this leads to strong positive rainfall anomalies. Conversely, the number of days with high rainfall is underestimated. The average rainfall amount is again well represented by the model. Although high rainfall occurs less often, this leads to strong negative rainfall anomalies that sometimes compensate those caused by the class of light rainfall, or reverse the trend. In 1988, the same occurs except for light rainfall, for which the number of days is indeed underestimated, leading to a global negative anomaly everywhere.

All this suggests a problem with the generation of rainfall. The first hypotheses so far are a lack of humidity in the low atmosphere (not enough moisture brought by the monsoon flow? not enough latent heat?) and a problem in the triggering of the convection. The dynamics could also be a possibility, although this has not been strongly suggested so far by our analysis. In order to perfect the validation of the MAR2006, the dynamics of the monsoon as modelled is evaluated in the next section.

5.3.3 Dynamics

This section looks at different dynamic variables such as wind, air temperature and air moisture. The model is compared to the driving fields (ERA40) through cross sections of these variables, and to observations through soundings (temperature and humidity only).

a. Zonal and meridional winds

The vertical distributions of the zonal and meridional winds averaged over 10°W to 10°E are illustrated in figure 5.11 to figure 5.12 per season⁵¹ for the years 1987 and 1988 (JFM not shown as similar to OND). The model is compared to the driving fields (ERA40). These figures confirm that the model reproduced the main features of the circulation in the lower, mid and upper troposphere.

⁵¹ JFM: January-February-March, AMJ: April-May-June, JAS: July-August-September, OND: October-November-December (northern hemisphere) seasons.

The most striking dynamic feature is probably the *Subtropical Westerly Jet* (SWJ) located in the upper troposphere with a zonal component peaking at 32 m/s. As expected, it dominates the wintertime (OND, JFM) and is located around 30°N. The modelled zonal component compares well to the ERA40. It is best developed in autumn (reaching 30 to 32 m/s as observed) but much weaker in summer (16 to 20 m/s). It is also stronger in autumn 1988 than 1987 and less southwards extended in summer. Both years in spring, the maximum southward extension is at 400 hPa instead of 100-200 hPa. The northward component of the SWJ agrees less with the observations. In 1987, it is overestimated by almost a factor 2 for summer, and the agreement of the jet latitudinal extension in summer is poor. In 1988, the agreement is slightly better in summer but the jet velocity is still too strong for the two other seasons.

The other westerly flow is the *monsoon flow* in the low troposphere. From spring to summer, the monsoon extends west-, north- and upwards until its maximum development is reached (summer). The model reproduces well the meridional component of the flow with velocities in a good agreement with the ERA40 (4 to 6 m/s according to the season). However, the model systematically overestimates the zonal component by approximately 2 m/s leading, for example, to a westward flow of 6 m/s in summer instead of 4 m/s.

The northern position of the zero isotach for both components is in very good agreement with the ERA40, except in spring 1987 (1988) where the meridional (zonal) zero isotach does not extend northward enough (14°N instead of 17°N). The thickness of the monsoon flow is well captured for the zonal component. It reaches the 800 hPa in summer and is greater in 1988 than 1987. However, the meridional component is confined to a too thin atmospheric layer for all seasons and both years. The model also shows a southern tail of the zonal component situated around the equator close to 800 hPa, a feature that the ERA40 does not present at all.

The *African Easterly Jet* (AEJ) is quite well reproduced. Its altitude and latitudinal location match what is found in the ERA40. It is present all year long (around 600 hPa) but only gets a jet status during spring (see 10°N) and

Chapter 5: Validation of the new MAR

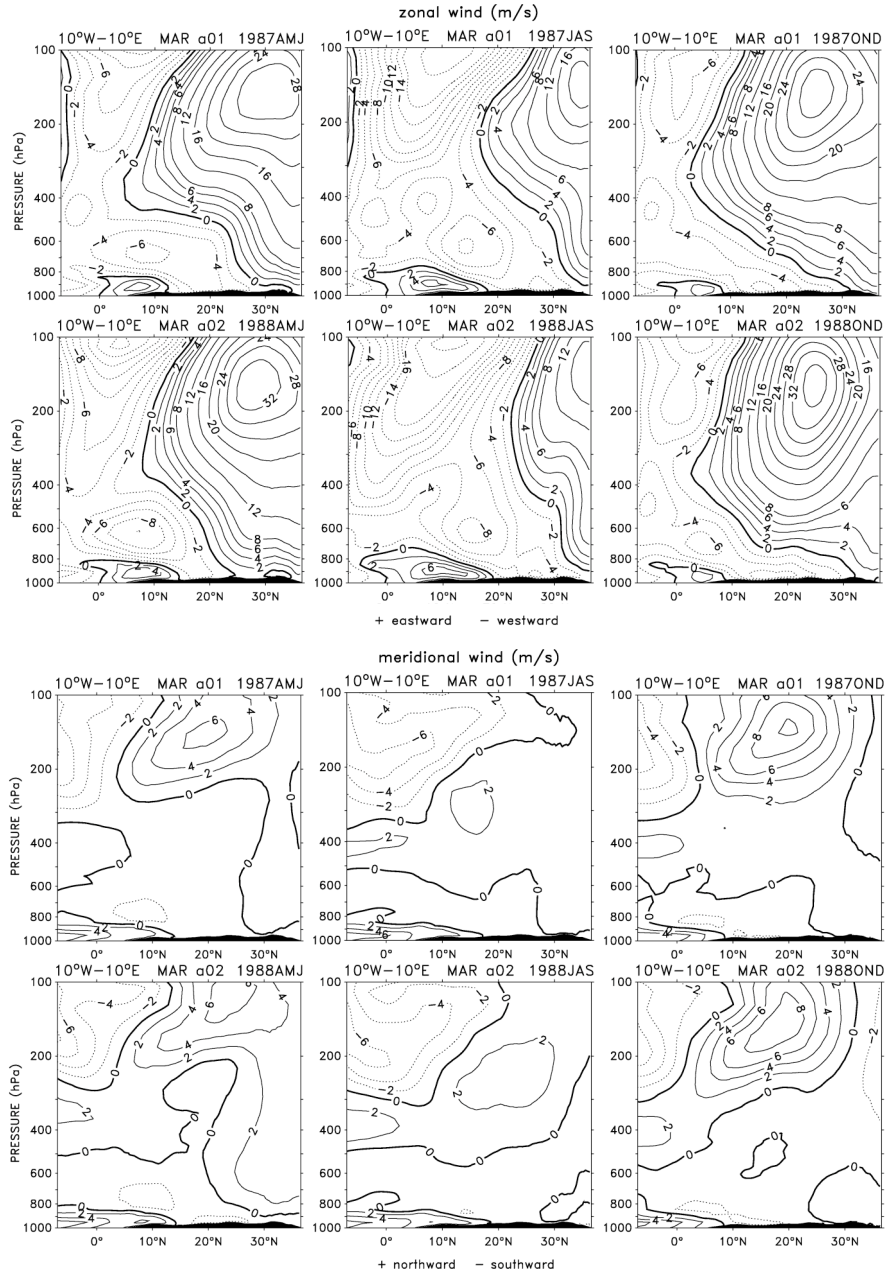


Figure 5.11 – Cross sections of the MAR zonal (top) and meridional (bottom) winds for spring, summer, autumn 1987 and 1988.

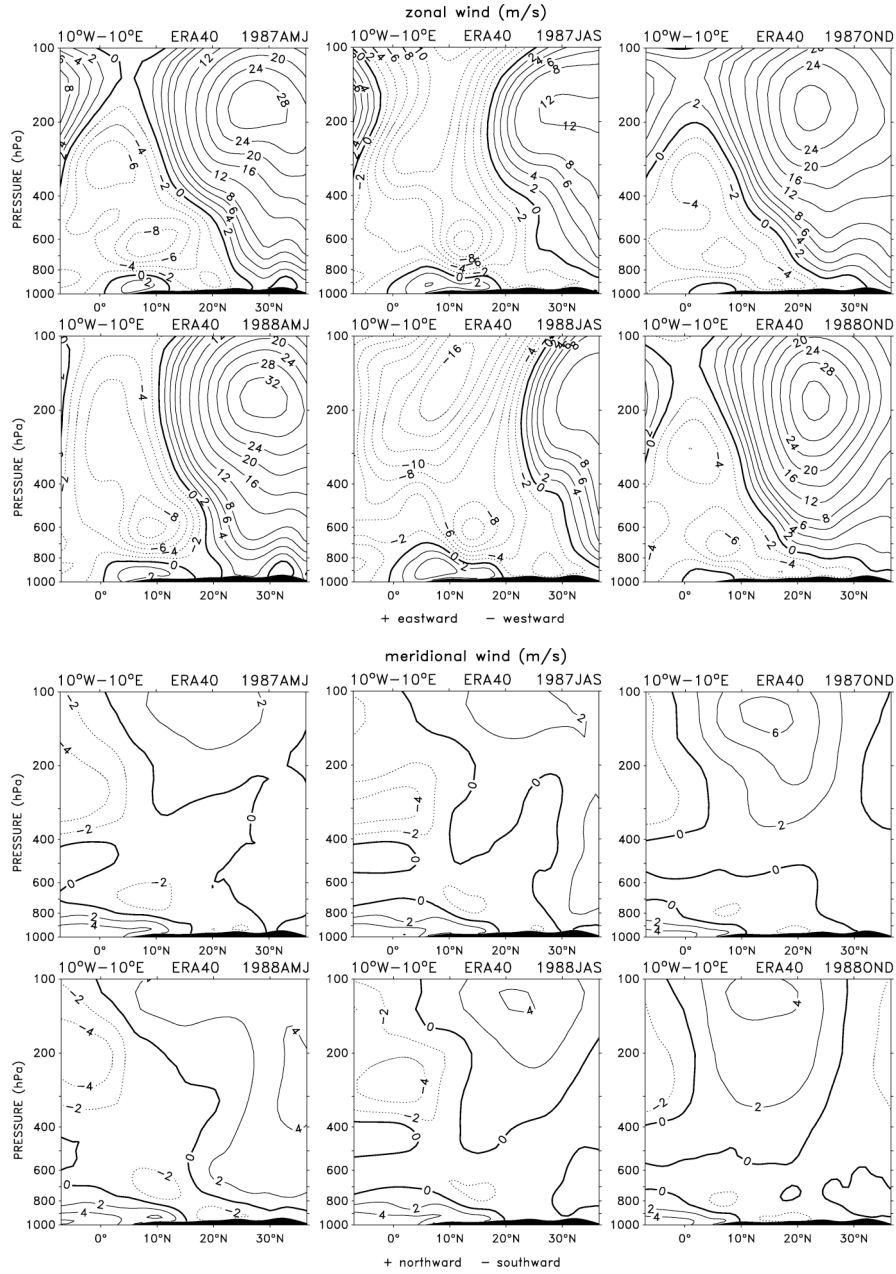


Figure 5.12 – Cross sections of the ERA40 zonal (top) and meridional (bottom) winds for spring, summer, autumn 1987 and 1988.

summer (see 15°N). However, its zonal velocity is somewhat underestimated: in summer 1987 (1988) it peaks at 8 m/s instead of 12 m/s (10 m/s). The jet is orientated essentially in the east-west direction. Indeed, its meridional component (southwards) is weak (1 to 3 m/s).

The *Tropical Easterly Jet* (TEJ) is wider, higher and more to the south than the AEJ, and is limited to the summer circulation. It is found in the model between 100 and 200 hPa at approximately 5-10°N. However, its core does not appear as clearly as for the AEJ, except in 1988. The model overestimates its zonal velocity by 2 m/s at 200 hPa for both years.

b. Air temperature and humidity

Similarly to the zonal and meridional winds, figure 5.13 and figure 5.14 give the vertical distributions of air temperature and air humidity per season for the years 1987 and 1988, averaged over 10°W to 10°E (JFM not shown as similar to OND⁵¹). The model is compared to the ERA40 data.

The general behaviour of the modelled *isotherms* is very similar to those of the ERA40. In altitude, the isotherm 0°C is located at 600 hPa and slopes northwards in spring. The heated surfaces to the north of 20°N redress it in summer but let it slope again when winter arrives over northern Africa. From 600 hPa to the top of the modelled atmosphere, the agreement is very good. From 600 hPa to the surface, each season is usually characterised by an adequate position of each isotherm at the surface, with maximum temperatures observed above the continent (see the isotherm 28°C in spring/summer and 24°C in autumn). Nevertheless, albeit the maximum temperatures have the same magnitude as in the reanalyses (except in autumn), they are confined to a too thin atmospheric layer close to the surface and do not extend upwards enough: this is the case of the isotherm 32°C in spring and summer. In autumn, the model is too cold and does not reach more than 24°C, unlike the 28°C of the reanalyses.

Looking at the *air humidity*, the model also compares well globally to the ERA40. The maximum humidity is concentrated in the low atmosphere. In the ITCZ, the moisture is pumped from the low atmospheric layers and sent upwards. The isoline 2 g/kg is close to 400 hPa at the latitude where the

convection is most active. The displacement of this zone with time relates the temporal northward-then-southward motion of the ITCZ and the rainfall band. Further north, the position of the heat low and dry air over the Sahara is easily identifiable. Close to the Mediterranean Sea, the agreement between model and ERA40 is very good. However, a few differences can be found. The same maximum humidities are reached, but like the air temperature, the moisture seems confined to the surface (see for example the position of the isoline 12 g/kg). To be noted too, the air above the Sahara is slightly drier.

The CRU data set also provides information on the air humidity close to the surface (vapour pressure). Figure 5.15 compares the model and the ERA40 to these observations. The mixing ratio r provided by the model and the reanalyses is converted into a vapour pressure e with $e = pr/(0.622 + r)$ where r is taken at the first level (10 m in height) and p is the corresponding atmospheric pressure. The figure confirms previous findings. Over the Sahara, the MAR is drier than the ERA40 and both seem drier than the CRU. Spatial details are respected (effect of the Atlas mountains for example) as well as the temporal evolution. It has to be specified that the CRU data are likely to correspond to a lower altitude. This partially explains their wetter character. In the regions to the south of the Sahara, the agreement between the MAR, the ERA40 and the CRU is better: notice the isobar 28 hPa over the coastal regions in spring, the position of the isobar 20 hPa in summer or the repartition of the vapour pressure over equatorial Africa in autumn.

A last comparison of the MAR with observations is performed thanks to radiosounding data in Niamey (Station 61052 DRRN Niamey-Aero), Niger (13.47°N, 2.16°E, 227 m). The vertical profiles of air temperature and humidity are shown for the middle month of each season both years (cf. figure 5.16). In altitude, no bias appears between the model and the observations, for both variables. It gives confidence in the convection scheme as there is no accumulation of moisture in the mid and upper troposphere. In the lower troposphere, biases appear. The temperature is colder by a few degrees for all three months and both years. Out of summer, the air is drier in Niamey in the model (see May and November), but wetter in August.

Chapter 5: Validation of the new MAR

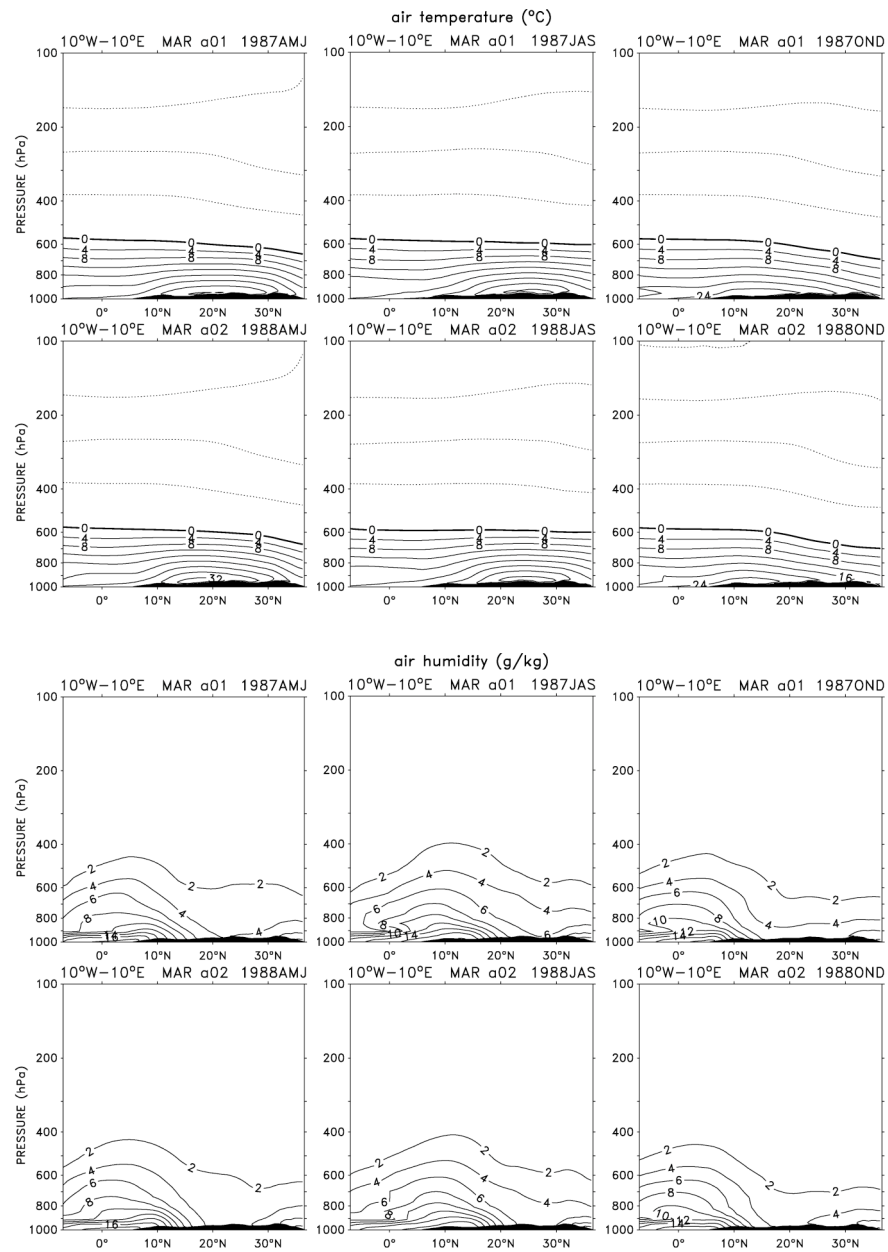


Figure 5.13 – Cross sections of the MAR air temperature (top) and humidity (bottom) for spring, summer, autumn 1987 and 1988.

Chapter 5: Validation of the new MAR

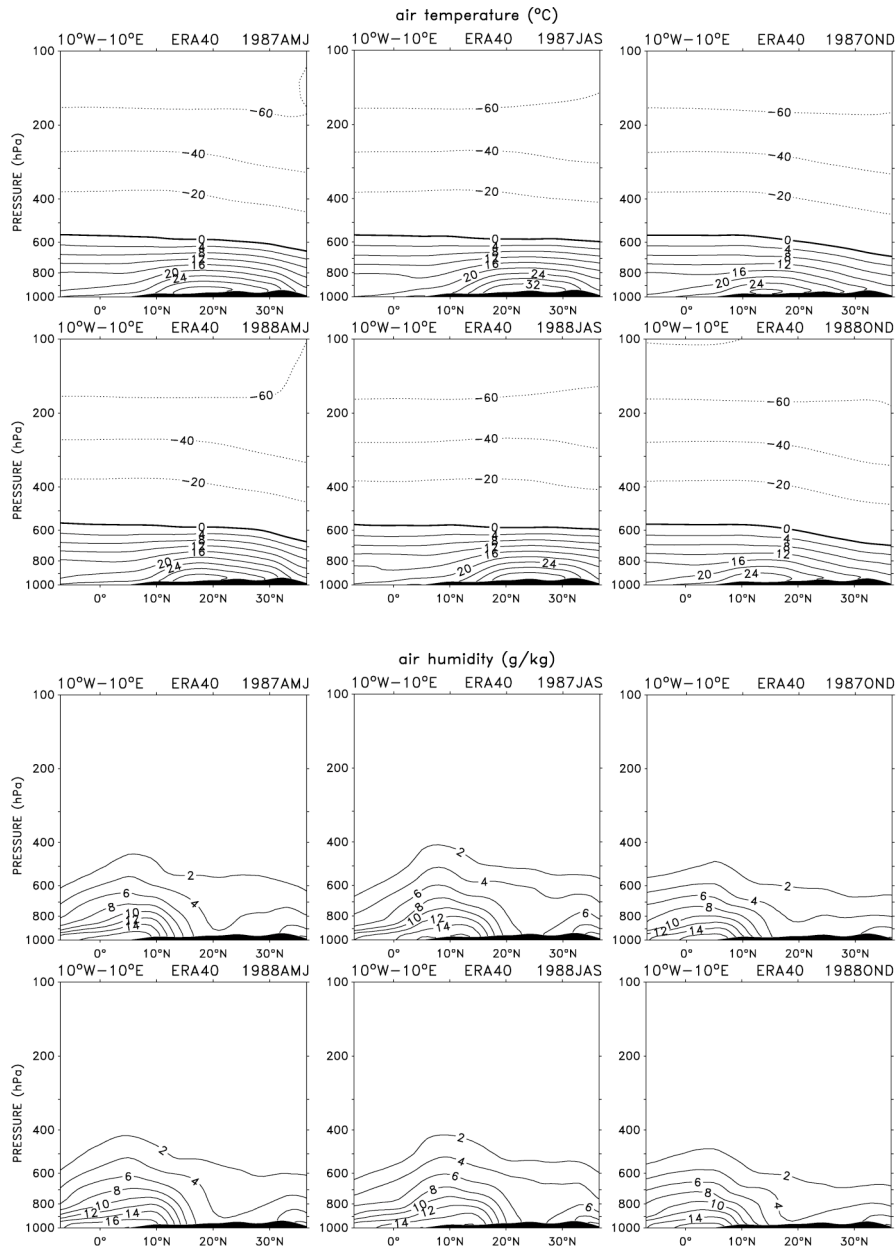


Figure 5.14 – Cross sections of the ERA40 air temperature (top) and humidity (bottom) for spring, summer, autumn 1987 and 1988.

Chapter 5: Validation of the new MAR

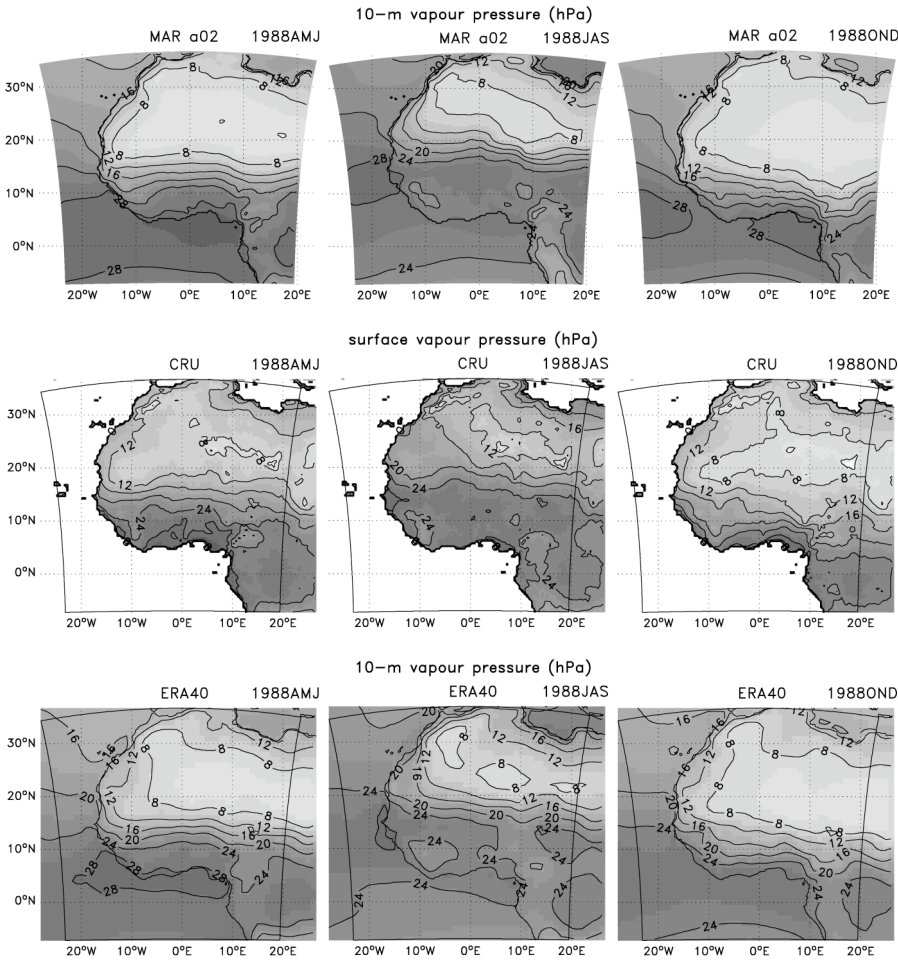


Figure 5.15 – Spatial distribution of vapour pressure per season (1988) close to the surface (for 1988), for the MAR (top), CRU (middle) and ERA40 (bottom).

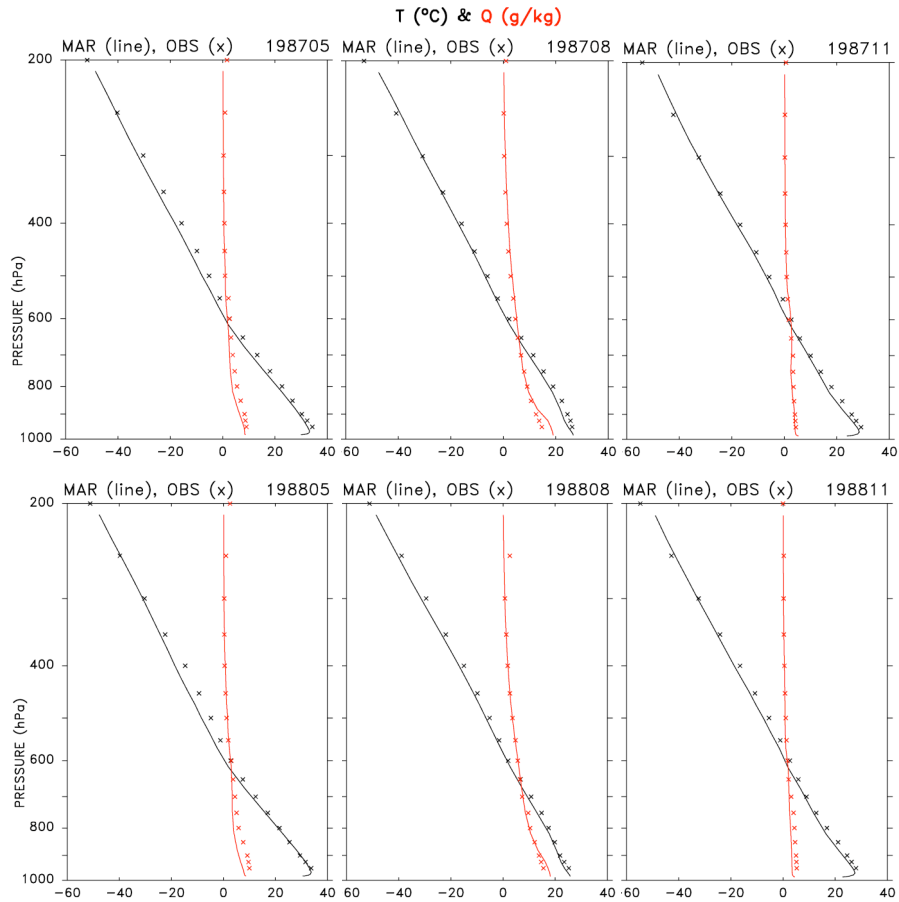


Figure 5.16 – Vertical profiles of the air temperature (T, black, in °C on the abscissa) and humidity (Q, red, in g/kg on the abscissa) in Niamey, Niger, for May, August and November 1987 and 1988. The MAR (solid lines) is compared to a local sounding (crosses).

c. Summary

The dynamics of the West African climate has been analysed for the years 1987 and 1988 and the model was found to reproduce its features reasonably well.

In the *upper troposphere*, the Subtropical Westerly Jet is correctly located at 30°N, dominates in wintertime and has an excellent zonal velocity (but an overestimated meridional one). The Tropical Easterly Jet is modelled at correct latitudinal and altitude locations (5-10°N, 100-200 hPa) and limited to the summer. Nevertheless, it seems spatially more diffuse and its zonal velocity is overestimated by approximately 2 m/s;

On the *mid troposphere*, the African Easterly Jet is correctly located in latitude (10-15°N) and altitude (600 hPa) and present all year long. Its zonal component is underestimated by 2 to 4 m/s in summer.

The model is in very good agreement with the reanalyses above 600 hPa where no bias of air temperature and air humidity is found.

In the *lower troposphere*, the monsoon flow is also reasonably well modelled by the MAR. It evolves accordingly with time and the zero isotach is very well located all year long (apart from spring with a lag of 3 degrees of latitude). Its thickness is correct for its zonal component but too thin for its meridional one. Its velocity is overestimated by 2 m/s zonally but correct meridionally. Associated to the monsoon flow, the maximum of air humidity is in good agreement with the reanalyses both in magnitude and location, but is confined to too thin a layer. The temperatures above the oceanic surface are correct. Above the continent, the situation is different. The model is slightly drier above the Sahara. The temperatures reach the same maxima as the reanalyses in spring and summer (28° to 32°C) but are underestimated in autumn (no isotherm 28°C). Moreover, for all seasons, the vertical position of the isotherms is underestimated and the model ends up to be globally colder. The soundings in Niamey revealed a model drier in spring and autumn and wetter in summer.

5.4 Conclusion: MAR2006 versus MAR2002

In section 5.2 we gave a list of the systematic problems experienced with the MAR2002. In section 5.3 we looked in detail at the performance of the MAR2006 (conclusions were drawn in subsections 5.3.2d and 5.3.3c). We shall now compare both versions in order to identify the extent to which the identified problems have been improved or not. In the following list⁵², a '+' sign denotes an improvement of the problem, a '-' sign denotes a deterioration, a '=' sign denotes no particular improvement or deterioration of the item.

- 1– **Underestimation of the AEJ:** still present and a bit stronger (-2 to -4 m/s).
- 2+ **Too strong westerly monsoon flow:** still present but reduced to 2 m/s.
- 3= **Cold bias in the lower atmosphere** over the continent: still present but the temperatures close to the surface are correct all year long except in winter time.
- 4= **Cold and wet bias in the lower atmosphere** found when comparing MAR to radiosounding data in Niamey in August: still present. It was noticed that in May (spring) and November (autumn), the bias becomes cold and dry.
- 5= **Slight general underestimation of the rainfall amounts:** still present.
- 6+ **Isohyet 2 mm/day too far north (20°N) and underestimation of the northward extent of the high rainfall area**, especially around the Senegalese coasts: modified. The isohyet 1 mm/day is properly located, the high rainfall area extends northwards enough on the Senegalese coasts but too far inland. **South-north rainfall gradient too strong:** modified. The gradient is better reproduced in the MAR2006.
- 7+– **Overestimation of the frequency of light rain events:** still present. **Underestimation of the intensity of convective events:** modified. The intensity of the convective events is now much better reproduced, but the frequency is underestimated.
- 8= **Bad representation of the rainfall daily structure:** still present. For a

⁵² The same numbering as in section 5.2 is used in this comparison list.

Chapter 5: Validation of the new MAR

year correctly modelled (for example 1987), the length of the rainy season is fairly well reproduced by the MAR and the magnitude of the rainy events is correctly caught too. However, the light rainfall class contains too many days and the high rainfall class not enough.

- 9= **Overestimation of rainfall during the retreat phase, belated rainy events** (over the Guinean band): still present. The hard core of the rainfall band maintains longer and introduces a latitude lag in the retreat phase (of which the other characteristics are correctly caught by the model).
- 10+ **Underestimation of rainfall over the Atlantic Ocean:** corrected. Rainfall now does occur above oceanic regions. This is a major improvement even if the rainfall pattern is not yet perfect.
- 11+ **No simulation of rainfall maxima over the coastal area during the installation phase** (May-June especially): modified. Maxima occur but are slightly underestimated.
- 12+ **Underestimation of rainfall maxima over the Fouta Djallon and the Cameroon highlands:** modified. The maximum over the Fouta Djallon is close to observations in magnitude, the rainfall pattern over the Cameroon highlands is close to observations too.
- 13= **Simulated maxima over the Jos and Bauchi Plateaus not found in observation:** still present and very likely due to the lack of observations.

Out of the 13 proposed items, 6 were improved, 2 got worse and 6 showed no noticeable/remarkable modification. The MAR2006 has shown its ability to model correctly the West African climate with a few exceptions we shall now discuss. Indeed, the validation of the MAR2006 has revealed five major problems: a too early start of rainfall in January-February (on the coastal areas), a too thin monsoon flow, a cold bias above the continental surfaces, a delayed retreat phase, and an underestimation of rainfall in 1988.

Possible explanations include the surface model SISVAT and the surface temperature. Indeed, recent simulations (H. Gallée, personal communication) have shown that the thickness of the shallowest soil layer impacts upon the

surface temperature. A thinner layer leads to increased sensible heat (and dry convection potentially able to thicken the monsoon layer) and latent heat. The global result is an increased moist static energy (MSE) and enhanced ocean-continent MSE gradient to which monsoon rainfall is known to be proportional.

In January and February, the ocean-continent surface and 10-m air temperatures gradients are stronger in the MAR than in the ERA40 which also show too early rainfall (with a weaker spatial amplitude). The surface temperatures are colder from the Sahel northwards (a few degrees) as well as the 10-m air temperatures. Potentially, this potentially could increase the MSE gradient (ocean warmer than continent at this time of the year) and enhance the rainfall activity in the rain band that is located south of the coast.

The SISVAT could also be responsible for the lag noticed in the southward retreat phase of the rainfall band. The whole rainfall band envelope behaves in accord with the observations, but the hard core of rainfall (between 10 and 12°N) reinforces in September-October when it should diminish. Moufouma-Okia (2003) suggested this could be linked to the leaf area index. Indeed, recent vegetation datasets have shown that the vegetation dataset currently used by the MAR overestimates the extent of the tropical forests in West Africa. This could potentially lead to an excess of moisture in the atmosphere through evapotranspiration, especially at the end of the rainy season, when the vegetation is at its growth peak and the soils are wet.

Conclusion and perspectives

West Africa is the westernmost region of Africa. The subcontinent is characterised by its high rate of natural population increase and an economy based mainly on rural agriculture, and the pressure on food production and water resources is huge. A monsoon system brings rainfall from July to October which is heterogeneous both spatially and temporally. This emphasises the need for a thorough understanding of the climate system of the region in order to better forecast rainfall and potentially better mitigate its potential impacts.

In this thesis, we have presented work conducted using the regional climate model MAR (Modèle Atmosphérique Régional). This model is composed of an atmospheric part, describing the dynamics and physics of the atmosphere, and a land part, representing the exchange of heat and moisture between the surface of the Earth (soil, vegetation) and the atmosphere. The model has the ability to represent the West African monsoon at high spatial and temporal resolutions—typically around 40 km horizontally and 30 minutes temporally.

After having described the model and presented some common configuration features for the simulations for which results are discussed, we have presented two modifications made to the model in order to improve its **representation of the biosphere** over West Africa. A new dataset of the Normalised Difference Vegetation Index (NDVI) has been implemented. Such data are very useful in representing aspects of the vegetation characteristics within the model. The new dataset is superior in quality to its predecessor, and now covers a period from 1983 up to the present day (instead of mid-1994). As a consequence of adopting the new NDVI dataset, the spatial and temporal properties of the representation of vegetation by the model have been improved. The second modification is the correction and improvement of the computation of the vegetation root fraction within each

Conclusion and perspectives

soil layer. The newly-implemented formulation of Gale and Grigal (1987) has enabled the computation of the root distribution at any depth in the ground. Results have shown coherent and realistic plant root structure that has impacted upon the heat fluxes computed by the model. Indeed, this was to be expected as the plant roots constitute the only link between the soil water and atmospheric water vapour in the land representation of the MAR. Fluxes such as latent heat and evapotranspiration have increased, which is expected to help to improve the representation of rainfall over West Africa.

► *In 2000 a new land cover data set of the Earth was published (GLC, 2000). For Africa, the resolution is 1 km which is higher than that of any datasets in existence at the time (Mayaux et al., 2004). Taking advantage of improved technology (instruments dedicated to vegetation observation onboard satellites), it provides a detailed map of the vegetation cover over West Africa. The most significant change with respect to the IGBP dataset currently used in the model to describe the vegetation types is the detection of croplands in the Guinean band rather than tropical forests which are reduced to isolated islands on the subcontinent. Such a dataset would be of major interest in the MAR to model the West African climate for which land-atmosphere interactions are important.*

The next chapter focussed on quantifying the uncertainty associated with any simulation performed with the model, uncertainty commonly referred to as **internal variability**. Such a measure is necessary for experiments where the influence of some change in physical processes or boundary conditions is under investigation, and for simulations of climate change and for impact studies, which can amplify uncertainties in small scale input data. This study (Vanvyve *et al.*, 2007), which had not previously been attempted over West Africa using a regional climate model has provided a practical measure of random uncertainty suitable for both applications. In essence it involved performing an initial condition perturbation experiment (perturbation on temperature, humidity, wind speed and pressure) with a regional climate model, using the fact that the regional climate model simulations converge because they are controlled by identical boundary conditions. We have shown that an equilibrium state is reached, in which any further difference

signal is internally generated and self-sustaining, and is independent of the initial condition. Non-identical pairs of experiments performed to evaluate a sensitivity or a response must differ by more than this amount for the conclusions to be reliable. We have found that the daily signal for precipitation at the 40 km scale can reach values of about 25 mm. The rainfall signal has a great deal of small-scale structure associated with essentially random variations in convective triggering. Temporal or spatial averaging thus always reduces this signal, and the reduction occurs relatively rapidly as we moved to longer periods and larger scales. Daily temperature and wind variations at the 40 km scale can reach values of around 3°C and 4 m/s. These values persist in time and space to a greater extent than the precipitation signal, particularly over the ocean where the convective influence is weaker and large scale dynamically coherent structures appear in the difference field.

➤ *It would be interesting to prolong this work with further perturbed simulations in order to strengthen the results (especially with the new version of the MAR). Although previous studies with other regional climate models on other regions have shown that the size of the initial perturbation is relatively insignificant, this should be confirmed for the MAR.*

The results obtained for the internal variability have then been applied to study the **influence of soil moisture initialisation on the performance of the model and the length of the spin-up time**—the spin-up time is the time needed by the model to be independent of its initial conditions and to reach a dynamic equilibrium with respect to its driving fields. More than one year long sensitivity simulations were started at different time of the year and divided into two groups in order to test two methods of soil moisture initialisation in the model: either fixed values were attributed to all soil layers, or more realistic values were taken from ERA40 data. The "re-analysis 40" are large-scale data used for driving the model's atmospheric module. Both methods led to wet or dry soil anomalies at the initialisation. The time taken by the model to successfully remove excess soil moisture has been shown to vary according to the time of the year and the initialisation method used. For example, starting the simulation before

Conclusion and perspectives

the rainy season permits quicker removal of most of a wet anomaly over the domain, but is not enough to remove the wet anomaly over the Guinean band where the rainy season will considerably slow down the process of soil moisture evacuation and will eventually lead to the longest spin-up. In particular, one dry anomaly induced by the second method was found not to disappear as quickly as the wet anomalies and maintained throughout the simulations. However, in general the results reach acceptable levels very quickly—i.e. as soon as the wet soil anomaly has disappeared—because this anomaly does not lead to any substantial modification of the energy and water budgets of the surface and the lowest layer of the atmosphere.

In conclusion, it was found that when using the ERA40 data, the global spin-up time was 3 months, compared to 6 for the fixed values method. Ideally, a simulation with year of interest *Y* must be started during the year *Y*-1. If interest is focussed only on the rainy season of *Y*, it can either be initialised before the rainy season of *Y*-1 (i.e. before May) if using the fixed values method for initialising the soil moisture, or at any month of *Y*-1 if using the ERA40 data. If the interest is to study the whole of year *Y*, then the same can be applied for the fixed values method, although the simulation must preferably be initialised by September at the latest if using the ERA40 data.

► *Some further studies could be carried out to investigate how to improve both methods and reduce the spin-up time. In the first method, we did not consider the possibility of changing the fixed value of soil moisture used to calculate the soil moisture content per layer. The value used leads to a wet anomaly across the whole domain. Increasing it is not logical as the soil moisture excess would take even longer to be removed. Decreasing it seems more sensible. However, it is very likely to lead to dry soil anomalies (especially over the Guinean band and equatorial Africa): dry anomalies located there could have a stronger influence than if located in the Sahara. In the second method, the ERA40 values used were not fully matching the soil description of the MAR—we mentioned the use of the soil moisture of only two soil layers (the shallowest ones). The ERA40 actually provides soil moisture fields for four layers: 0-7 cm, 7-28 cm, 28-100 cm, 100-289 cm.*

Since these layers match closely the soil discretisation of the land model of the MAR, it would be worth using the four of them to establish a correct correspondence between the proposed and modelled depths of soil. Perhaps doing so would avoid an overestimate of the soil moisture anomaly of the deepest soil layer of the model as it currently encountered.

The last chapter presented the **validation of the last version of the model MAR** (dated 2006). We looked in detail at the performance of the model over the years 1987 and 1988, and compared it to the previous version.

Concerning the rainy regime, the MAR has been shown to reproduce the main features of the regime both in time and space for both years 1987 and 1988. The spatial representation of rainfall is indeed in very good agreement with the observations. Maxima are found where expected—Fouta Djallon, Cameroon highlands, Jos/Bauchi plateaus, the delta of the Niger river, and equatorial Africa—with intensities generally close to observations (disregarding the general lack of rainfall for the year 1988). The northward extension of the rain band is especially well modelled regarding the isohyet 1 mm/day. The temporal evolution of rainfall is also captured by the model. From January to September, the model performs respectably well, representing the installation phase of the monsoon, the first rainy season, the abrupt shift northwards and the short dry season, the second and high rainy season, and finally the retreat phase. Points to notice include: 1) a premature start for the rainfall (light) over the coastal area in January and February, 2) an underestimation of the rainfall over the Guinean band later in the first rainy season, 3) a retreat phase marked by a late southward movement of the hard core of the rain band (it persists in September-October), although lower rainfall retreats on time and is similar to observations. The length of the rainy season is correctly modelled, although it tends to be slightly longer than observed. Rainfall anomalies are essentially due to two classes of rainfall, light rainfall (1-10 mm/day) and high rainfall (10-20 mm/day), of which the number of days is respectively overestimated (underestimated in 1988) and underestimated.

Concerning the dynamics of the West African climate, the model was

Conclusion and perspectives

found to reproduce its features reasonably well. In the *upper troposphere*, the Subtropical Westerly Jet is correctly located in latitude, dominates in wintertime and has an excellent zonal velocity. The Tropical Easterly Jet is modelled at correct latitudinal and altitude locations, and limited to the summer. Nevertheless, it seems somewhat faded and its zonal velocity is overestimated by approximately 2 m/s. In the *mid troposphere*, the African Easterly Jet is correctly located in latitude and altitude, and present all year long—it was noticed that its zonal component is underestimated by 2 to 4 m/s in summer. The model was also in very good agreement with the reanalyses above 600 hPa where no bias of air temperature and air humidity was found. In the *lower troposphere*, the monsoon flow is also reasonably well modelled by the MAR. It evolves accordingly with time and the zero isotach is very well located all year long (apart from spring). Associated to the monsoon flow, the maximum in air humidity is in good agreement with the reanalyses both in magnitude and location, but is confined to too thin a layer. The temperatures above the ocean surface are correct. Above the continent, the situation is different. The model is slightly drier above the Sahara. The temperatures reach the same maxima as the reanalyses in spring and summer but are underestimated in autumn. Moreover, for all seasons, the altitude of the isotherms is underestimated and the model ends up to be globally colder. The soundings in Niamey revealed a model drier in spring and autumn and wetter in summer.

Among the common 13 problems exhibited by the previous versions, most of them were found again in this version—a few less than half were improved, a few less than half showed no significant modification and two got worse. The too strong westerly monsoon flow is a bias still present but reduced in magnitude. The isohyet 2 mm/day is no longer located too far northwards and the high rainfall area extends northwards enough on the Senegalese coasts. The south-north rainfall gradient is also better reproduced. The intensity of the convective events is now much better reproduced—although the frequency of these events is underestimated. The strong underestimation of rainfall over the Atlantic Ocean has been significantly corrected: rainfall now does occur above oceanic regions. This

is a major improvement even if the rainfall pattern is not yet perfect. Previously, no simulation of rainfall maxima over the coastal area during the installation phase (May-June especially) has been correct, although the maxima are still slightly underestimated. The maximum over the Fouta Djallon is now close to observations in magnitude as well as the rainfall pattern over the Cameroon highlands.

In summary, in this thesis, we have described the modification and application of a regional climate model (MAR) for modelling the climate of West Africa. The MAR has been shown to model the climate of this region well. Indeed, the work in this thesis has illustrated the suitability of the MAR for further land-atmosphere sensitivity studies in West Africa. Many 2-dimensional models, including the 2-dimensional version of the MAR, have shown high sensitivity of monsoon rainfall to vegetation changes (deforestation, desertification), ranging from decreases in rainfall to total collapse of the monsoon (Charney, 1975; Xue *et al.*, 1990; Zheng and Eltahir, 1997; Wang and Eltahir, 2000a,b). As highlighted by these authors, the link between land-cover anomalies and the rainfall/monsoon response is strongly dependent on the model physics (surface, parameterisations of convection, etc.). These 2-dimensional models have a rather simple description of the surface (vegetation, soil) and, by definition, do not include the zonal dimension. In the Sahel, most of rainfall is produced by systems moving from east to west (squall lines, east waves). Hence, the full mechanism underlying rainfall changes may not be caught by such models. The MAR has a detailed description of both physics and dynamics of the atmosphere, and is usually employed at finer spatial scales than other RCMs. Basic experiments have been attempted with the MAR but, to date, the model has not shown a clear response. This surely is an area worthy of further attention.

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