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Assessing the Impacts of Land Cover Change on Climate in non-Amazonian South America

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Abstract

Non-Amazonian South America has one of the highest rates of conversion of native ecosystems globally. Most of the studies investigating the climate impacts of these changes focus on the Amazon while the possible influences that these changes may have on climate of non-Amazonian regions have received less attention. The aim of this thesis was to evaluate the impacts of land use and land cover on the mean and extreme climate of non-Amazonian South America by conducting modelling experiments for pre-clearing (before year 1500) and present (year 2005) land covers. It develops new data sets of changes in land surface characteristics for this period and applies a high resolution regional climate model to simulate the potential impacts of changes in natural vegetation cover.

The thesis begins by providing a theoretical framework of land-atmosphere interactions. It then reviews the process of land use and land cover change and subsequent climatic consequences in non-Amazonian South America and identifies those ecosystems most affected and least studied. The review highlights that non-Amazonian regions have lost more than 3.4 million km^2 of natural vegetation since European colonization. Despite the magnitude of ecosystem loss, non-Amazonian South America accounts for significantly fewer studies addressing land surface-atmospheric processes compared to the Amazon region, highlighting the knowledge gap in relation to the main topic of this thesis. Based on these results, the following chapters address this knowledge gap by focusing on the climatic impacts of land use and land cover change in four main ecosystems: the Atlantic Forest (Brazil, Argentina and Paraguay), Cerrado (Brazil), Dry Chaco (Argentina, Bolivia and Paraguay) and the Chilean Matorral (Chile). I then apply the variable resolution CSIRO Conformal Cubic Atmospheric Model (CCAM) global climate model at a 25 km spatial resolution over the South American continent to quantify the seasonal climate impacts of each of historic land cover change.

The results of computed modelling experiments show significant changes in surface fluxes, temperature, precipitation and moisture in all ecosystems. For instance, simulated temperature changes were stronger in the Cerrado and the Chilean Matorral with an increase of between 0.7 and 1.4 °C. Changes in the hydrological cycle revealed high regional variability. Results also show that the loss of natural vegetation has significantly affected temperature extremes as a decrease in the number of warm days and an increase in the number of warm nights. Importantly, there is a strong dependence on both seasonality and the vegetation contrast inflicted by land use/cover change, with large roughness changes resulting in increasing wind speed and advection, while smaller roughness

changes result in feedbacks more reliant on the distinction between sensible and latent heat fluxes. This explains the dry season response in both temperature extremes and the increase in aridity according to land use/cover change, whereby regions with increased wind speed reduce warm day temperature extremes, despite increasing mean temperature trend, and have a greater impact on atmospheric water demand than those regions that mainly increase sensible heat fluxes. These results can explain the observed trends in temperature extremes in non-Amazonian South America and highlights the need to embed land use/cover change as a forcing within future climate change scenarios.

The main conclusions of the thesis are: a) non-Amazonian South America is one of the most impacted and least studied regions worldwide in terms of climatic impacts of land use and land cover change, b) modelled impacts of this process are expressed as significant changes in surface temperature and the hydrological cycle through changes in soil and atmospheric moisture, and have increased the aridity in all the examined ecosystems. The thesis highlights the importance of considering the influence of land use and land cover on the mean and extreme climate through changes in biophysical properties that significantly impact the surface-atmospheric coupling and therefore the hydrological cycle. This is critical consideration for national natural ecosystem management strategies as conserving and restoring natural vegetation cover may help mitigate the negative consequences of climate change and therefore have a direct influence on the welfare of the region’s 200 million inhabitants.

Declaration by author

This thesis is composed of my original work, and contains no material previously published or written by another person except where due reference has been made in the text. I have clearly stated the contribution by others to jointly-authored works that I have included in my thesis.

I have clearly stated the contribution of others to my thesis as a whole, including statistical assistance, survey design, data analysis, significant technical procedures, professional editorial advice, and any other original research work used or reported in my thesis. The content of my thesis is the result of work I have carried out since the commencement of my research higher degree candidature and does not include a substantial part of work that has been submitted to qualify for the award of any other degree or diploma in any university or other tertiary institution. I have clearly stated which parts of my thesis, if any, have been submitted to qualify for another award.

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Publications during candidature

Peer-reviewed papers

Salazar, A., Baldi, G., Hirota, M., Syktus, J. and McAlpine, C. (2015). Land use and land cover change impacts on the regional climate of non-Amazonian South America: A review. *Global and Planetary Change*, 128, 103-119. doi:10.1016/j.gloplacha.2015.02.009.

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Conference Abstracts

Salazar, A., McAlpine, C., Syktus, J. Assessing the impacts of land cover change on climate in Non-Amazonian South America. *Global Land Project Open Science Meeting. Land Transformations: Between Global Challenges and Local Realities*. Humboldt University, Berlin, 19th - 21st March, 2014.

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Contributor	Statement of contribution
A. Salazar (Candidate)	Literature review (100%) Generation and analysis of data to calculate areas of land use/cover change (100%) Wrote the paper (100%)
G. Baldi	Edited paper (30%)
M. Hirota	Edited paper (30%)
J. Syktus	Edited paper (20%)
C. McAlpine	Edited paper (20%)

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J. Katzfey	CCAM climate model setup (10%) Paper editing (5%)
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Chapter 1

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Keywords

land use change, land cover change, natural vegetation, non-amazonian south america, atlantic forest, cerrado, dry chaco, chilean matorral, land-atmosphere interactions, deforestation

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List of Abbreviations

LUCC	Land Use and Land Cover Change
CSIRO	Commonwealth Scientific and Industrial Research Organisation
CCAM	Conformal-Cubic Atmospheric Model
CABLE	CSIRO Atmosphere Biosphere Land Exchange
MODIS	Moderate Resolution Imaging Spectroradiometer
CONICYT	Comisión Nacional de Investigación Científica y Tecnológica (Chile)
BNU	Beijing National University

Glossary

Advection	Transport of water or air along with its properties (e.g., temperature, chemical tracers) by winds or currents. Regarding the general distinction between advection and convection, the former describes transport by large-scale motions of the atmosphere or ocean, while convection describes the predominantly vertical, locally induced motions (IPCC, 2013a).
Atmospheric (planetary) boundary layer	The atmospheric layer adjacent to the Earth's surface that is affected by friction against that boundary surface, and possibly by transport of heat and other variables across that surface (AMS, 2000). The lowest 100 m of the boundary layer (about 10% of the boundary layer thickness), where mechanical generation of turbulence is dominant, is called the surface boundary layer or surface layer (IPCC, 2013a).
Climate	Climate in a narrow sense is usually defined as the average weather, or more rigorously, as the statistical description in terms of the mean and variability of relevant quantities over a period of time ranging from months to thousands or millions of years. The classical period for averaging these variables is 30 years, as defined by the World Meteorological Organization. The relevant quantities are most often surface variables such as temperature, precipitation and wind. Climate in a wider sense is the state, including a statistical description, of the climate system (IPCC, 2013a).
Climate change	Climate change refers to a change in the state of the climate that can be identified (e.g., by using statistical tests) by changes in the mean and/or the variability of its properties, and that persists for an extended period, typically decades or longer. Climate change may be due to natural internal processes or external forcings such as modulations of the solar cycles, volcanic eruptions and persistent anthropogenic

	changes in the composition of the atmosphere or in land use. Note that the Framework Convention on Climate Change (UNFCCC), in its Article 1, defines climate change as: 'a change of climate which is attributed directly or indirectly to human activity that alters the composition of the global atmosphere and which is in addition to natural climate variability observed over comparable time periods'. The UNFCCC thus makes a distinction between climate change attributable to human activities altering the atmospheric composition, and climate variability attributable to natural causes. See also Climate change commitment, Detection and Attribution (IPCC, 2013a).
Climate extreme	The occurrence of a value of a weather or climate variable above (or below) a threshold value near the upper (or lower) ends of the range of observed values of the variable (IPCC, 2012).
Climate feedback	Climate feedback An interaction in which a perturbation in one climate quantity causes a change in a second, and the change in the second quantity ultimately leads to an additional change in the first. A negative feedback is one in which the initial perturbation is weakened by the changes it causes; a positive feedback is one in which the initial perturbation is enhanced. In this Assessment Report, a somewhat narrower definition is often used in which the climate quantity that is perturbed is the global mean surface temperature, which in turn causes changes in the global radiation budget. In either case, the initial perturbation can either be externally forced or arise as part of internal variability (IPCC, 2013a).
Climate model	A numerical representation of the climate system based on the physical, chemical and biological properties of its components, their interactions and feedback processes, and accounting for some of its known properties. The climate system can be represented by models of varying complexity, that is, for any one component or combination of components a spectrum or hierarchy of models can be identified, differing in such aspects as the number of spatial dimensions, the extent to which physical, chemical or biological processes are explicitly represented or the level at which empirical parametrizations are involved. Coupled Atmosphere–Ocean General Circulation Models (AOGCMs) provide a representation of the climate system that is near or at the most comprehensive end of the spectrum currently available. There is an evolution towards more complex models with interactive chemistry and biology. Climate models are applied as a research tool to study and simulate the climate, and for operational purposes, including monthly, seasonal and interannual climate predictions (IPCC, 2013a).
Deforestation	Conversion of forest to non-forest (IPCC, 2013a).

Drought	A significant deviation from the normal hydrologic conditions of an area (Palmer, 1968).
Ecosystem	Community of living organisms in conjunction with the non-living components of their environment, interacting as a system (Blew, 1996; Tansley, 1935).
Ecosystem services	The processes and conditions derived from ecosystems that sustain and enhance human wellbeing (Martinez-Harms et al., 2015).
Free atmosphere	The atmospheric layer that is negligibly affected by friction against the Earth's surface, and which is above the atmospheric boundary layer (IPCC, 2013a).
Hydrological cycle	The cycle in which water evaporates from the oceans and the land surface, is carried over the Earth in atmospheric circulation as water vapour, condenses to form clouds, precipitates over ocean and land as rain or snow, which on land can be intercepted by trees and vegetation, provides runoff on the land surface, infiltrates into soils, recharges groundwater, discharges into streams and ultimately flows out into the oceans, from which it will eventually evaporate again. The various systems involved in the hydrological cycle are usually referred to as hydrological systems (IPCC, 2013a).
Land use/cover change (LUCC)	Human-caused changes that affect the biophysics, biogeochemistry, and biogeography of the terrestrial surface and its effects on the atmosphere (Pielke et al., 2011).
	Atmospheric boundary layer The atmospheric layer adjacent to the Earth's surface that is affected by friction against that boundary surface, and possibly by transport of heat and other variables across that surface (AMS, 2000). The lowest 100 m of the boundary layer (about 10% of the boundary layer thickness), where mechanical generation of turbulence is dominant, is called the surface boundary layer or surface layer (IPCC, 2013a).
Latent heat flux	The turbulent flux of heat from the Earth's surface to the atmosphere that is associated with evaporation or condensation of water vapour at the surface; a component of the surface energy budget (IPCC, 2013a).
Reanalysis	Reanalyses are estimates of historical atmospheric temperature and wind or oceanographic temperature and current, and other quantities, created by processing past meteorological or oceanographic data using fixed state-of-the-art weather forecasting or ocean circulation models with data assimilation techniques. Using fixed data assimilation avoids effects from the changing analysis system that occur in operational analyses. Although continuity is improved, global reanalyses still suffer from changing coverage and biases in the observing systems (IPCC, 2013a).
Sensible heat flux	The turbulent or conductive flux of heat from the Earth's

	surface to the atmosphere that is not associated with phase changes of water; a component of the surface energy budget
Vegetation type	Vegetation community defined by their general structure (physiognomy) and seasonal activity pattern (Box and Fujiwara, 2013).

CHAPTER 1 INTRODUCTION

1.1 Influences of land cover change on near-surface climate: The need of a regional perspective

By the year 2000, approximately 55 percent of the Earth's biomes had been converted into pastures, croplands, settlements and other land uses (Ellis et al., 2010). These changes have impacted biotic components of ecosystems such as biodiversity, and also modified land-atmosphere interactions through changes in water and energy balances (Foley et al., 2003). Land Use and Cover Change (LUCC) can affect climate through absorption or emission of greenhouse gases (biogeochemical impacts) and by modifying the physical properties of land surface (biogeophysical effects) (e.g. Bonan, 2008b). Changes in vegetation features can lead to changes in surface fluxes of radiation, heat, moisture and momentum that can further impact the climate system (Pielke et al., 2002). In terms of radiation changes, LUCC can alter the surface albedo and thereby evapotranspiration processes and partitioning of sensible, latent and ground heat fluxes that can influence near-surface temperature and precipitation (Pielke et al., 2007). In addition, changes in land cover can transform vegetative attributes such as roughness, which influences the mixing of air close to the land surface affecting cooling processes (Foley et al., 2003). Cumulatively, these modifications of the land surface can impact the climate at a range of spatial and temporal scales and, owing to non-linear feedbacks between terrestrial ecosystems and climate, climate changes can exhibit threshold behaviour (Zehe and Sivapalan, 2009). This non-linear characteristic of the climate system can cause substantial qualitative differences in its dynamic due to small changes of some parameters (Rial et al., 2004) such as those determined by land cover change.

Increasing global evidence of the climate impacts of LUCC has underlined the need to understand the extent of which these changes affect the climate system and on the consequences for ecological and socio-economic systems. Many studies have shown that the conversion of forests, savannas, woodlands and grasslands (hereafter referred to as native vegetation) can have a significant impact on evapotranspiration rates which can enhance climate extremes such as droughts (Deo et al., 2009; Lee et al., 2011; Taylor et al., 2002). By way of illustration, changes in climate have been associated to vegetation changes in the Brazilian Amazon and regions of Australia (Costa et al., 2007; McAlpine et al., 2007; Pitman et al., 2004; Sampaio et al., 2007). In these regions, the replacement of forests and natural vegetation by other land uses has decreased evapotranspiration rates, leading to changes in heat fluxes and convection which in turn reduces rainfall and increases surfaces temperatures. Moreover, evidence from arid and semi-arid climates has shown that overgrazing, agriculture expansion and fuelwood collection result in increased surface temperature and changes in the hydrological cycle (Balling et al., 1998; Bryant et al., 1990; Otterman, 1989).

The response of climate to LUCC varies according to location, biogeophysical characteristics and spatial and temporal scales of the processes under study. Global climate models have shown variations in climatic responses to changes in the Earth's surface according to latitude and vegetation type (e.g. Pielke et al., 2007). Nonetheless, even though these global models have substantially increased the global understanding of land-atmosphere interactions, many questions remain unanswered concerning the impacts of LUCC on climate at the regional scale. In this regard, most regional studies have focused on temperate and boreal regions of the northern hemisphere and tropical deforestation (Betts, 2000; Mahmood et al., 2006; Marshall et al., 2003; Schneider et al., 2004). In general, these studies have found that climate response to changes in boreal and temperate vegetation is largely driven by changes in albedo (Betts et al., 2007), while LUCC in tropical regions affect climate through latent heat flux changes (Lee et al., 2011). Conversely, with the exception of Australia (McAlpine et al., 2007; McAlpine et al., 2009; Pitman et al., 2012; Pitman et al., 2004), evidence of LUCC impacts on climate from subtropical regions of the southern hemisphere is limited and land-atmosphere interactions are not well understood.

Many authors have highlighted the role of LUCC in climate science is not adequately assessed as a globally averaged measure because this process is a highly regionalized phenomenon and its impacts on climate can vary according to the region of study (e.g. Mahmood et al., 2010). Although the globally averaged surface temperature change may be close to zero in reaction to LUCC (Forster et al., 2007), regional changes in surface temperature and precipitation can be significant compared with those that result from the anthropogenic greenhouse gases (Pielke, 2005). As stated by Mahmood et al. (2010), a regional focus is appropriate to the understanding of human impacts on climate because it is the regional responses that produce droughts, floods and other climate impacts, and it is at a regional scale that people and ecosystems experience these consequences. As the climate trend (i.e. cooling or warming) and the strength of the changes depend on variables such as latitude, features of local vegetation and extension of the affected area (Beltrán-Przekurat et al., 2012b), generalizations of the impacts of LUCC on climate are difficult to make. Therefore, there is a need of more studies focusing on the relations between human alterations of Earth's surface and climate in order to better understand the mechanisms and biogeophysical impacts of these changes at a regional scale.

1.2 Evaluating the impacts of LUCC on the climate of non-Amazonian South America

The conversion rate of South American landscapes is one of the highest worldwide and is likely to continue in the future. In relation to surface-climate interactions, most of the attention has focused on the Amazon. Here, many studies have highlighted the consequences of deforestation on surface temperature and precipitation (Hirota et al., 2011; Sampaio et al., 2007; Xue et al., 2006). Nonetheless, recent research from non-Amazonian South America has shown potential links between changes in land cover and variations in climate (Beltrán-Przekurat et al., 2012b). However, major gaps remain in our understanding of land surface-climate interactions in this region, particularly in water-limited environments.

European colonization of non-Amazonian South America began in the 16th century. This region is dominated by a variety of ecosystems, most of which are drylands distributed largely in Argentina, Bolivia, southern Brazil, Peru, Paraguay and Chile. The expansion to agriculture and the building of new cities and the necessity of fuelwood changed the regions landscapes dramatically (Camus, 2006). Ecoregions such as the Brazilian Cerrado, Chilean Matorral, the Argentinean Chaco and the Atlantic forest of Argentina and Brazil have suffered substantial deforestation rates even higher than that of the Amazon. For example, in Central Chile, 42 percent of dryland forests disappeared between 1975 and 2008 (Schulz et al., 2010). Boletta et al. (2006) showed high deforestation rates in the Argentine Chaco. Here, from 1992 to 1999, more than 273,000 ha were deforested at an annual rate of 5 percent, with approximately 80 percent of the former forest extent now occupied by crops, pastures, and secondary scrub (Zak et al., 2008). Viglizzo et al. (2011) evaluated the impacts of agricultural expansion between 1960 and 2005 in over 1.5 million km^2 (63 percent of Argentina), and showed that the natural forests area suffered a significant reduction accounting for 42 percent, 28 percent and 16 percent of the initial area for the Atlantic, Chaco and Yungas forests, respectively. This process is also extensive in the grasslands of Argentina and Uruguay, the Dry Chaco ecoregion of Argentina, Bolivia and Paraguay (Clark et al., 2010b; Vega et al., 2009), and in the Cerrado savanna in central Brazil (Klink and Machado, 2005).

Recent research has shown potential links between these LUCC and changes in climate. According to Collins et al. (2009), increases in surface temperature identified from year 1948 over certain areas of tropical and subtropical South America cannot be explained by El Niño or La Niña events alone and may be the result of recent LUCC processes. In central Argentina, an area that yields most of the countries agricultural production, a cooling trend observed by Rusticucci and Barrucand (2004) has been associated by Beltrán-Przekurat et al. (2012b) to albedo changes resulting from conversions from natural grassland to agriculture (using a fully coupled atmospheric-biospheric regional climate model). These results also agree with estimations from Nunez et al. (2008) for the period 1961-2000 using the "observation minus reanalysis" (OMR) method to estimate the impacts of LUCC on surface temperature over Argentina. Nunez et al. (2008) also identified a warming trend in northern and western areas of Argentina which have experiences high LUCC rates during the last decades (Viglizzo et al., 2011). Collins et al. (2009) observed warming trends over Chile, and also central and northeast Brazil and recognize that only global and regional climate models including anthropogenic forcing can simulate or explain temperature warming due to human activities in South America.

In addition to these temperature variations, rainfall in other regions of South America has also experienced significant variability that could be partially linked with LUCC. Using the National Centre of Atmospheric Research (NCAR) CAM3 climate model, Lee et al. (2011) found that local land use changes were critical in simulating precipitation decreases and temperature increases in central Brazil. However, recent studies show high spatial variability in the trend of precipitation for the region (Magrin, 2014).

Potential relations between rainfall variability, increasing droughts and LUCC processes have not yet been investigated in non-Amazonian South America (Haylock et al., 2006; Minetti et al., 2010). Variations in temperature and rainfall across the sub-continent, in addition to the highly dynamic landscapes, make non-Amazonian South America an extremely fragile region to climate perturbations. This is of particular concern as large areas of non-Amazonian South America are water-limited environments and vulnerable to desertification. These areas are considered one of the extra-tropical regions most affected by El Niño and La Niña events and experience strong inter-annual precipitation variability associated with these events (Grimm and Tedeschi, 2009). Droughts have been a frequent phenomenon, and many countries have had to apply emergency programs to combat its impacts and subsequent desertification. Non-Amazonian South America is recognized as an important part of the desertification prone area of the world (Hellden and Tottrup, 2008), and the feedbacks between climate and LUCC dynamics accelerate this process with substantial ecological and socioeconomic impacts.

In spite of the concern about climate variability in the region and the need to improve water use efficiency, the effects of LUCC on the climate of non-Amazonian South America have not been considered in natural resources management and climate risk assessments. Although the historic and ongoing process of LUCC, increasing drought frequency and extreme climate variability, the biogeophysical influences of land surface modification on the climate of the sub-continent are poorly understood. Recent evidence has shown that changes in vegetation cover do have influences in temperature and heat fluxes, especially in semi-arid and arid regions (Beltrán-Przekurat et al., 2012b; Nuñez et al., 2008). These alterations can have significant impacts on rainfall regime and, therefore, on water availability for ecosystems and people. As in other regions, the process of LUCC in non-Amazonian South America can also influences changes of CO_2 concentrations and in the surface energy budgets that can enhance droughts and subsequent increase risk of desertification. The decrease in vegetation cover may result into a highly irreversible shift to a dry climate state, in which rainfall would be insufficient to allow for the recovery of vegetation (Brovkin et al., 1998).

Changes in land cover are directly related to land degradation and its impacts on both surface albedo and moisture fluxes can reduce water availability as has been shown in drylands of North America, Sahel and Nagev-Sinai Regions (Balling et al., 1998; Charney, 1975; Otterman, 1989). Therefore, the potential impacts of LUCC on the regional climate, both the mean climate and climate extremes, in non-Amazonian South America warrant further investigation.

1.2.1 The Problem Statement

Non-Amazonian South America has one of the highest rates of conversion of native ecosystems globally. However, most of the studies investigating the climate impacts of LUCC in South America focus on the Amazon. However, we do not currently know the potential biophysical influences these changes in native vegetation cover may have on the regional climate. Biophysical processes and feedbacks resulting from changes in land surface characteristics such as albedo, surface roughness and leaf area have an important role in regulating climate, especially at the regional scale. These changes can alter the energy balance by tens of Watts/m^2 compared to a radiative forcing of around three Watts/m^2 resulting from current increases in the concentration of greenhouse gases (Anderson et al., 2011; Betts et al., 2007). While the radiative forcing from greenhouse gases is distributed homogeneously over the global surface, the radiative forcing from changes in forest cover is regionally-specific. Changes in forest and tree cover alter the partitioning of energy into sensible heat (which warms the air), and latent heat (which cools the land surface through evaporative cooling). Forests and tree cover enhance moisture recycling, cloud formation and precipitation processes, which help stabilise the regional climate and influence weather. They evaporate up to 10 times more than herbaceous vegetation. This is an important problem in non-Amazonian South America because large areas of natural vegetation have been already transformed to other land uses, and any loss of forest and tree cover increases the risk of a shift to a drier climate and subsequent desertification.

1.3 Aim and Objectives

1.3.1 Overall Aim

The overall aim of this thesis was to evaluate the impacts of LUCC on the climate of non-Amazonian South America by conducting modelling experiments for pre-clearing (before year 1500) and present (year 2005) land covers. It develops new data sets of changes in land surface characteristics for this period and applies a high resolution regional climate model to simulate the potential impacts of changes in land cover from pre-clearing conditions to current conditions.

1.3.2 Objectives

The specific objectives are:

1. Review the existing literature about land use and land cover change processes and their impacts on the regional climate of non-Amazonian South America.

2. Conduct modelling simulations to evaluate climate impacts of historic land use and land cover change to present conditions in non-Amazonian South America with a focus on mean seasonal temperature and rainfall.

3. Evaluate the impacts of historical land use and land cover change to present conditions on climate extremes and aridity in non-Amazonian South America.

1.4 Approach

The project applies a high-resolution (~25 km) modelling approach to investigate the potential climatic impacts of historical changes of natural vegetation cover in non-Amazonian South

America. It focuses on four broad ecosystems: the Dry Chaco (Argentina, Bolivia and Paraguay), the Cerrado (Brazil), the Atlantic Forest (Brazil, Argentina and Paraguay) and the Chilean Matorral (Chile). However, in Chapter 4 two additional ecosystems are included: the Temperate Grasslands (Argentina and Uruguay) and Tropical Dry Forests (continental). A climate model developed by Commonwealth Scientific and Industrial Research Organisation (CSIRO) was used to detect for changes in the local and regional climate as a consequence of changes in vegetation. The preparation and analysis of land surface data was performed using MODIS satellite images of land cover and leaf area index combined in a Geographical Information System. This process required spatial modelling and interpolation approaches.

1.5 Structure of the thesis

The remainder of this thesis is presented as a series of logically ordered Chapters that address the main objectives.

Chapter 2 provides the conceptual basis of vegetation-climate interactions from which the following Chapters are based and applied. It describes the main vegetation-climate connections in terms of energy balance, surface and moisture fluxes, and surface hydrology that finally influence surface temperatures and precipitation.

In Chapter 3, I review the published work between years 1900 and 2013 in the field of LUCC impacts on the climate in order to identify the current knowledge gaps for non-Amazonian South America. This chapter is published in Global and Planetary Change as: Salazar, A., Baldi, G., Hirota, M., Syktus, J. and McAlpine, C., 2015. Land use and land cover change impacts on the regional climate of non-Amazonian South America: A review. Global and Planetary Change, 128(0), 103-119.

Chapter 4 is based on the main findings from the previous Chapter and addresses the climatic impacts of historical vegetation change for the least studied and most impacted ecosystems. It is focused on the impact of LUCC on seasonal temperature and rainfall in non-Amazonian South America.

Chapter 5 further examines the consequences of historical LUCC on selected climate extremes and aridity to account for the potential role that natural vegetation loss might have on extreme surface temperatures and precipitation events.

Finally, Chapter 6 synthesises the main project's findings, limitations and research recommendations.

CHAPTER 2 THEORETICAL BASIS OF VEGETATION-ATMOSPHERE INTERACTIONS

Abstract

The distribution of world's vegetation broadly depends on the climate, although underlying parental material and soil type, along with specific drainage conditions also determine the fine limits of vegetation distribution. However, the properties of the vegetation canopy influence the climate through changes in the net energy budget. This is in turn influenced by alterations of albedo and emissivity which respectively affect the dynamics of shortwave and longwave radiation at the surface. In addition, vegetation directly impacts the fluxes of momentum, latent and sensible heat within the surface boundary layer, affecting the energy partitioning and hence precipitation and temperature. These impacts are driven by changes in the turbulent mixing of air imposed by vegetation and by the exchange of CO_2, oxygen and water vapour between vegetation and the atmosphere. Moreover, vegetation affects overland flow by deep rooted transpiration of soil moisture through the leaves of the canopy, by changing soil structure affecting infiltration rates, and by intercepting and evaporating precipitation through the soil-plant-atmosphere continuum.

This Chapter addresses the theoretical basis of vegetation-atmosphere interactions through a review of the main physical concepts. It discusses the role of vegetation in the energy balance, fluxes of heat and hydrology that gives the conceptual framework of the following Chapters.

2.1 Surface energy budget and vegetation

Following Barry and Chorley (2010), the climate system is driven by the energy coming from the sun. According to the Planck's radiation law, the sun behaves as a black body absorbing all the energy that receives and radiates energy at the maximum possible rate at a given temperature (6,000 K). From the integration of Planck's law equation, we can obtain the total energy emitted by a black body across all wavelengths as described by the Stefan-Boltzmann law. In the case of the sun, the emitted radiation is composed of about 8 percent of wavelengths falling in ultraviolet, 48 percent in visible light and 52 percent in near-infrared, collectively named as shortwave or solar radiation (wavelength less than 4μm). Solar energy is the primary source of energy reaching the land surface and oceans, ice sheets and vegetation are key elements in determining the partitioning of this energy.

The net amount of energy available at the land surface from the incoming and reflected shortwave and the downwelling and emitted longwave radiation at the Earth's surface is called surface radiation balance and is expressed by:

$$R_n = S\downarrow(1-\propto) + L\downarrow - L\uparrow \tag{1}$$

where S↓ is the solar shortwave radiation, α is the surface albedo, L↓ is the incoming longwave radiation re-emitted back from the atmosphere down to the surface and L↑ is the longwave radiation emitted by the surface which depends on its emissivity and temperature according to Stefan–Boltzmann law (Barry & Chorley, 2010).

The properties of the vegetation canopy influence the net radiation budget by modifying the albedo and emissivity which respectively affect the dynamics of shortwave and longwave radiation at the surface. Earth's vegetation shows different absorbing and scattering surfaces to the incident radiation which are driven by the optical properties of the plant leaves, the spatial and angular distributions of the foliage and the architecture of the canopy (Rowe, 1993). These features are determinant in surface albedo and any change imposed on vegetation cover will eventually impact the reflected shortwave radiation and consequently the surface energy budget by reducing the available energy at the near-surface (Barry and Chorley, 2010). In terms of longwave radiation, the vegetation also emits longwave radiation downward and upward, and absorbs longwave radiation coming from the ground and the sky. A schematic illustration of the surface energy fluxes over a vegetated area is shown in Figure 2.1.

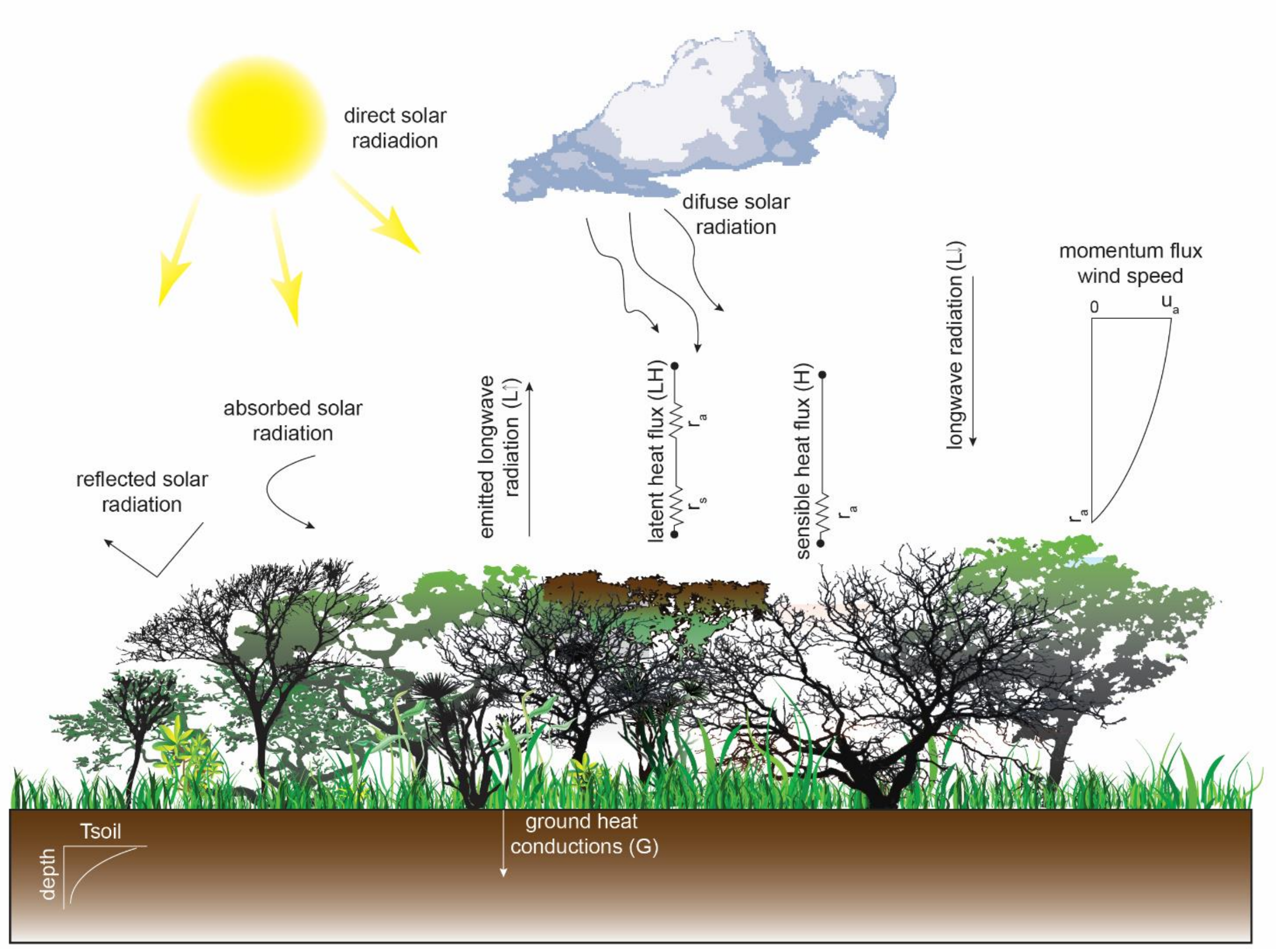

Figure 2.1 Illustration of the surface energy fluxes over a vegetated land. Adapted from Bonan (2008b, page 1446) and Pielke (2001, page 152). In the Figure, r_s and r_a are the resistances for surface and air, respectively. U_a is the wind speed.

2.2 Surface energy balance and vegetation

Considering the energy budget of the Earth's surface, we can appreciate the existence of a net approximated surplus of 98 W m^{-2} (Trenberth et al., 2009). This surplus energy is conveyed to the atmosphere by sensible heat flux (the transport of heat through convection of air) and latent heat flux (heat transfer through evapotranspiration). Heat fluxes are critical components of the energy budget and boundary layer climates, and are closely dependent on the properties of the land surface, atmosphere and substrate (Bonan, 2008a). Following Barry & Chorley (2010), the surface energy balance can be obtained from the net surface energy (R_n), the sensible heat flux (H), the latent heat flux (LH) and heat flux into the ground (G) as:

$$R_n = H + LH + G \tag{2}$$

In the climate system, the partitioning of available energy between sensible and latent heat fluxes is noteworthy since more latent heat raises the vapour water content in the atmosphere and can increase cloudiness and precipitation, while increases in sensible heat tends to warm the planetary boundary layer (Kabat et al., 2004). As shown by Pielke (2001), changes in vegetation cover can directly impact the heat and moisture fluxes within this boundary layer. For instance, landscape change, such as tropical deforestation, decreases latent heat fluxes and increases sensible heat fluxes, diminishing convective available potential energy due to the former, and increasing temperature of the boundary layer near the surface because of the later (Angelini et al., 2011).

In the climate system, the partitioning of available energy between sensible and latent heat fluxes is noteworthy since more latent heat rises the vapour water content in the atmosphere and can increase cloudiness and precipitation, while increases in sensible heat tends to warm the planetary boundary layer (Kabat et al., 2004). As shown by Pielke (2001), changes in vegetation cover can directly impact the heat and moisture fluxes within this boundary layer. For instance, landscape change, such as tropical deforestation, decreases latent heat fluxes and increases sensible heat fluxes, diminishing convective available potential energy due to the former, and increasing temperature of the boundary layer near the surface because of the later (Angelini et al., 2011).

The exchange of sensible and latent heat between land and the atmosphere is due to the turbulence mixing of heat transporting moisture and momentum of air in the near-surface. The turbulence mixing has a chaotic behaviour with high momentum convection and variation of pressure and flow velocity in space and time. This process is regulated by the aerodynamic resistance, where wind speed and surface roughness are key factors.

2.3 Surface fluxes and vegetation

Turbulent mixing of air results in transport of momentum, heat and moisture (Figure 2.2). This process creates eddies that mix air and thus momentum, heat and water vapor between the different layers in a wind profile above the ground. Near the surface, the velocity of an air parcel flowing above it decreases due to friction exerted by the presence of rough elements such as trees, grass, and other objects, creating a profile of wind speeds that increases with distance from the surface to a level where wind speed is independent of surface friction. The reduction of wind speed at the surface creates eddies that transfer momentum from the atmosphere to the surface, and heat and water from the surface into the atmosphere as shown in Figure 2.2 and Figure 2.3.

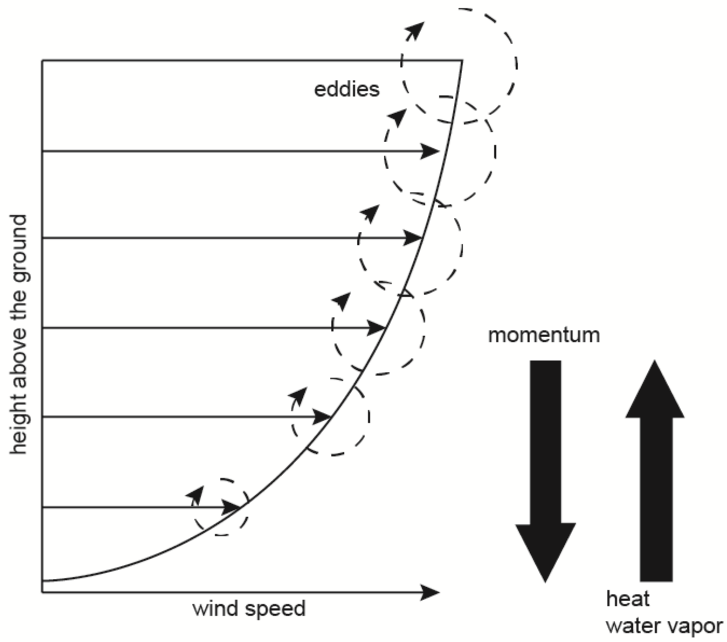

Figure 2.2 Schematic diagram of momentum, heat and water vapour fluxes. With increasing height, the wind speed becomes more independent on surface friction, creating eddies that transfer momentum from the atmosphere to the surface and generally heat and water vapour to the surface to the atmosphere. Source: Bonan (2008a, page 206).

Fluxes of momentum and heat can be obtained through the eddy diffusion model or the resistance model. With regard to the first approach, following Monteith & Unsworth (2007) and Bonan (2008a), any vertical flux of a scalar is the covariance between its fluctuating concentration and vertical velocity. Therefore, the exchange of momentum between the atmosphere and the land surface is represented by momentum flux or shearing stress (τ) and is the result of the product of the air density (ρ), wind velocity fluctuations in the horizontal component (u') and fluctuations of vertical wind velocity (w'). Under near-neutral atmospheric conditions the wind profile in the atmosphere over a homogeneous terrain increases logarithmically with height (Figure 2.3) and is related with momentum flux through the frictional velocity (u_*) as:

$$\tau = \rho u'w' = \rho u_*^2 \tag{3}$$

As wind speed profile is strongly influenced by surface roughness, the protrusion of elements above the surface such as trees or other vegetation types will display the turbulent flux of wind profile upwards, with mean wind speed profile becoming:

$$\bar{u}_z = \left(\frac{u_*}{k}\right) \ln[(z-d)/z_{0M}] \tag{4}$$

where $\bar{u}_z$ is the mean wind speed (ms^{-1}) at height z (m) above the ground, u_* is the frictional velocity for the wind profile, k is von Karman's constant (0.4), d is the zero-plane displacement height, and z_{0M} is the theoretical height at which wind speed is zero, also named roughness length. By way of illustration, Figure 2.3 shows how differences in surface roughness and height for short grass ($z_{0M} = 1\ mm; d = 7\ mm$) and tall crop ($z_{0M} = 1\ mm; d = 7\ mm$) influence the logarithmic wind profile. This change is produced by modifications in surface friction and the absorption of momentum by the vegetation and the land surface (Lawrence, 2004; Monteith and Unsworth, 2007).

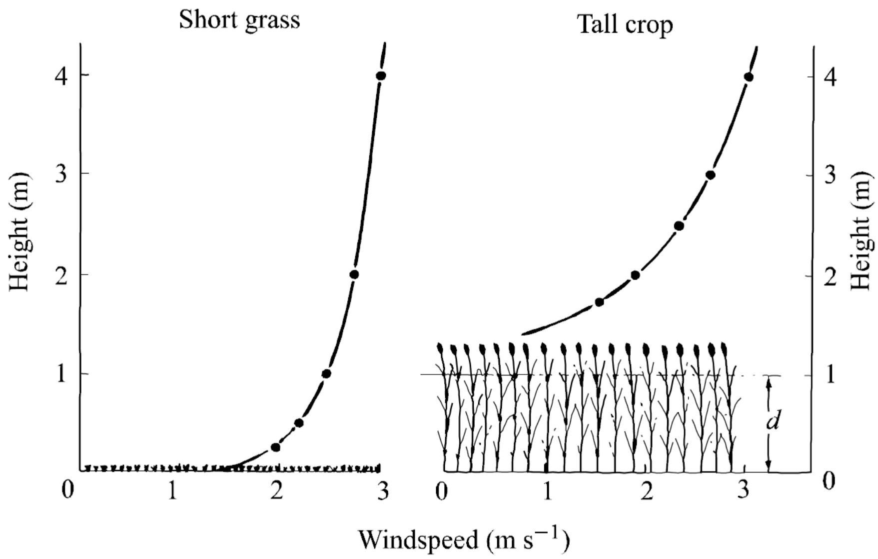

Figure 2.3 Differences in wind profiles over short grass and tall crop when wind speed at 4 m above the ground is 3 m s^{-1}. Taken from Monteith & Unsworth (2007, page 314).

The equations for fluxes of momentum, sensible heat and latent heat can also be expressed in relation to aerodynamic resistance. This is a turbulent diffusion term which impedes the transfer of momentum and heat away from the vegetation or soil surface to the free atmosphere, and it is inversely dependent on the wind speed and the logarithm of the surface roughness length (Kabat et al., 2004). Aerodynamic resistance decreases with increasing vegetation height because of greater roughness length, with the amount of foliage, conventionally specified as leaf area index, playing an important role (Monteith and Unsworth, 2007).

Considering the aerodynamic resistance (r_{aM}), the transfer of momentum (τ) between two different levels at different wind speeds above the ground (u and u_s) is limited by the resistance of the surface in the form of:

$$\tau = \rho(u - u_s)/r_{aM} \quad (5)$$

where the aerodynamic resistance term (s m^{-1}) evaluated under neutral conditions between the atmosphere at height *z* and the surface is defined by:

$$r_{aM} = \frac{1}{k^2 u}\left[ln\left(\frac{z-d}{z_{0M}}\right)\right]^2 \quad (6)$$

Equation (6) can be extended to obtain the fluxes of sensible heat and latent heat fluxes according to temperature and humidity profiles in different layers of the atmosphere near the surface. In a resistance notation, sensible and latent heat fluxes are expressed by:

$$\begin{gathered} H = \rho C_p(\theta - \theta_s)/r_{aH} \\ E = \rho(q - q_s)/r_{aW} \end{gathered} \quad (7)$$

where the sensible heat flux (H) is directly proportional to the temperature difference between boundary layer levels $(\theta - \theta_s)$ and the volumetric specific heat of air ρC_p, and inversely proportional to the aerodynamic resistance for sensible heat (r_{aH}), whilst the latent heat flux E is directly proportional to the specific humidity difference and inversely proportional to the aerodynamic resistance for latent heat (r_{aW}).

For non-neutral conditions momentum and heat fluxes can be explained following the similarity theory of Monin-Obukhov, in which aerodynamic resistance can be expressed as a function of buoyancy (Garratt, 1992).

Highlights

- Vegetation type and cover influence the energy budget by modifying the albedo and emissivity.
- Vegetation changes directly impact the fluxes of momentum, latent and sensible heat within the surface boundary layer affecting precipitation and temperature.
- These impacts are driven by changes in turbulent mixing of air imposed by vegetation, with increased surface roughness from vegetation structure.

2.4 Theoretical Basis of vegetation-hydrology interactions

At a global scale, 40 percent of the rainfall over land stems from oceanic moisture transported inland by wind and the remaining 60 percent stems from evaporative fluxes generated from terrestrial ecosystems (Falkenmark and Rockström, 2004). This means that the water flux between the surface and the atmosphere depends largely on evaporative processes occurring upon the land surface and any change in the sources of moisture can have significant impacts on sustainability and reliability of rainfall (Falkenmark and Rockström, 2004). The movement of water in the hydrological cycle occurs through the soil-plant-atmosphere continuum (SPAC) which represents the movement of water from soil, through roots and leaves, and out into the atmosphere (Eamus et al., 2006). The structure of the vegetation is therefore linked to the water content of the soil and the atmosphere, the climate, and the water status of the vegetation. Changes in any of these components will affect the others in a non-linear way.

2.5 Moisture fluxes in vegetation: The Canopy Model

In the plant phase of the SPAC, the moisture transport from canopies is regulated by the exchange of CO_2, oxygen and water vapour from leaves through photosynthesis. This exchange is defined by the vapour pressure deficit created by differences in water vapour concentration between the substomatal cavities and the surrounding atmosphere, and is controlled by the stomatal pore at the opening of the substomatal cavity (Figure 2.4).

The exchange of water vapour and CO_2 in the plant leaf is controlled by stomatal conductance, which is determined by photosynthetically active radiation (PAR), temperature, vapour pressure deficit, foliage water potential and ambient CO_2 concentration (Jarvis, 1976), and by stomatal and boundary layer resistances (Figure 2.4). Stomatal resistance is the reciprocal of the stomatal conductance and thus depends in the same variables. The gas exchange between the leaf and the surrounding atmosphere is also influenced by the resistance between the leaf surface and the surrounding atmosphere (Taiz and Zeiger, 2006). Factors such as wind speed and turbulence affects the transpiration process in the leaf by determining the thickness of the boundary layer and thus the transfer of moisture (Eamus et al., 2006).

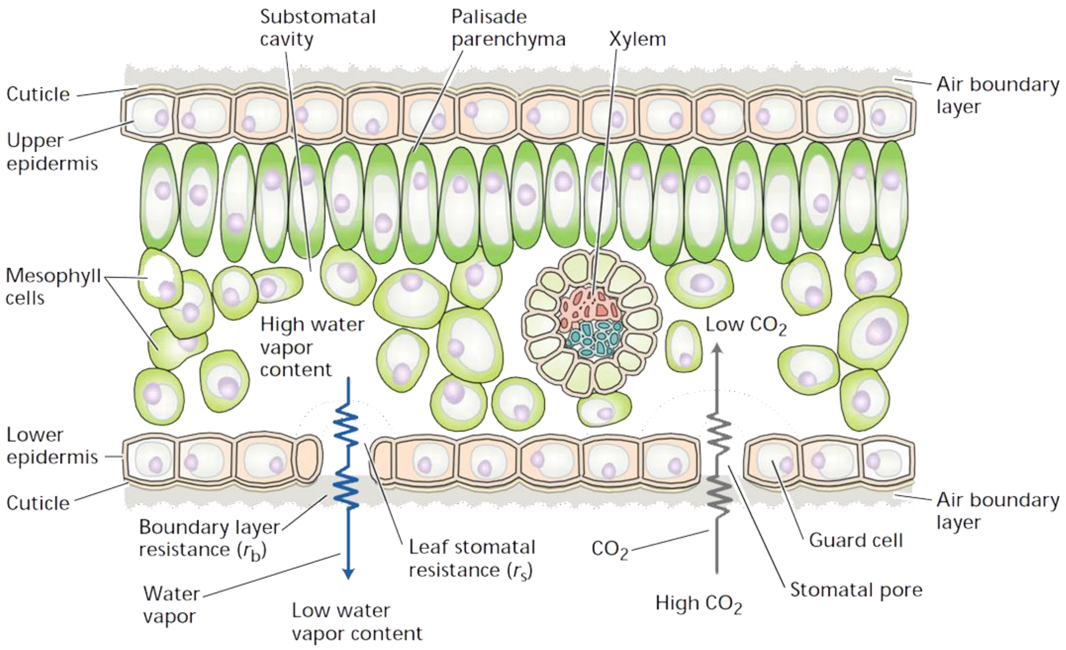

Figure 2.4 Water and CO2 flux through the leaf. From Taiz & Zeiger (2006, page 57)

Stomatal control of transpiration in leaves can be scaled to evaluate the canopy moisture fluxes. This is based on the fact that the principles determining the temperature and energy fluxes of a leaf also determine the temperature and energy fluxes of plant canopies (Bonan, 2008a). In this approach, canopies are treated as a single leaf scaled to represent a canopy and are referred as "Big leaf" canopy models. In these models, the canopy resistance represents the resistance of all the leaves in the canopy. Following Wang & Leuning (1998), transpiration from canopies can be obtained through a two-leaf model, which calculates transpiration considering sunlit and shaded leaves separately. In this model, the latent heat (λE) is calculated for sunlit and shaded leaves through the extended Penman-Monteith equation as:

$$\lambda E = \frac{s(R_n - G) + \rho C_p (e_*[T_a] - e_a)/r_{aH}}{s + \gamma (r_c + r_{aW})/r_{aH}} \tag{8}$$

where s is the slope of the curve relating saturation water vapour pressure to temperature, R_n is the net available energy, G is the total conductance, $e_*[T_a]$ is the saturation vapour pressure evaluated at the surface temperature, e_a is the vapour pressure of air, γ is the psychrometric constant (66.5 Pa^{o}C^{-1}) and r_c is the canopy resistance. This model states that latent heat exchange between the canopy and the atmosphere is regulated by canopy resistance (r_c) governing the process within the plant canopy, and the aerodynamic resistance for water vapour (r_{aW}) influencing the turbulent processes above the canopy (Bonan, 2008a).

As the leaf area determines the amount of transpirable surface from a canopy, the transfer of latent heat from the canopy to the atmosphere is closely related to the leaf area index. Greater leaf area index increases the moisture exchange with the atmosphere by increasing the vegetated surface from which moisture is lost (Naithani et al., 2013). However, canopy conductance is constant with high leaf area because low incident light levels on leaves deep in the canopy close stomata. For low leaf area index values (less than 3 m^2m^{-2}), surface conductance exceeds canopy conductance and soil evaporation is more important (Bonan, 2008a). Leaf area index is also used to estimate the evaporative fluxes of intercepted precipitation and dew in canopies (e.g. Bulcock and Jewitt, 2010).

2.6 Overland flow and vegetation

The soil phase in the SPAC is a vital process of the hydrological cycle. Soil moisture dynamics affect the water content in plant cells, photosynthesis and carbon assimilation and thus latent heat flux from terrestrial vegetation to the atmosphere with impacts on precipitation and water cycle (D'Odorico and Porporato, 2006). This is particularly noteworthy in water limited environments where respiration is the most relevant parameter for the loss of moisture, regulating the local soil water balance and determining plant growing conditions (Rodríguez-Iturbe and Porporato, 2004).

Infiltration is the water flux from the surface into the soil. This movement is governed by gravitational potential, which is the dominant force of water movement in wet soils, and by matric potential, resulting from the attraction of the water to the solid surface of mineral and organic particles in the soil (Eamus et al., 2006). Water moves downward when the gravitational potential exceeds matric potential and the opposite occurs when the absorptive force retaining the water in the soil matrix is stronger than the gravitational force, stopping water flux and determining the wilting point in which water is no longer available to plants (Bonan, 2008a).

The amount of water infiltrating into the soil, also named infiltration capacity, depends upon antecedent conditions that affect soil permeability such as texture, pore size, temperature, moisture content at the start of infiltration and the length of the rain event. If rainfall exceeds the infiltration capacity, the excess of water flows downhill as surface runoff or overland flow. This mechanism of overland flow generation is known as “Hortonian” or “infiltration excess” overland flow. However, water flow can also be generated when the soil is already at or near saturation due to the presence of an impervious surface (e.g. a bedrock, clay layer or the water table) in which case only a limited amount of water can enter the soil and overland flow can be produced by “saturation excess” (D'Odorico and Porporato, 2006).

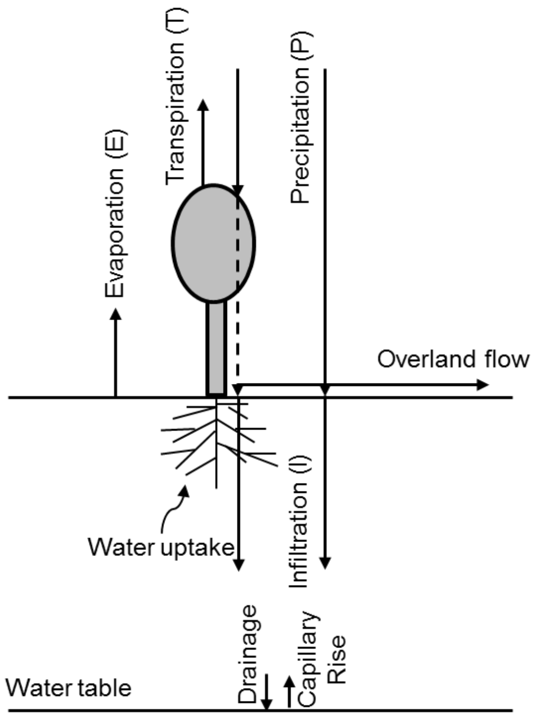

Figure 2.5 Water fluxes and overflow generation in the soil-vegetation system. Adapted from D'Odorico & Porporato (2006, page 110)

Figure 2.5 shows a schematic representation of the mechanisms of the soil water balance and its interaction with vegetation. Vegetation type, density and structure plays a prominent role in determining overland water fluxes, particularly in water limited environments. Vegetation controls the redistribution of runoff, water table levels and soil moisture by altering soil permeability (D'Odorico et al., 2010; Lubczynski, 2009; Saco et al., 2007). A number of field studies have shown that vegetation can enhance soil infiltration rates and water infiltration capacity through improved soil structure and aggregation by organic matter and plant litter production (Ludwig et al., 2005; Miller et al., 2010; Wilcox et al., 2003). These processes interact with physiography and regulate flow distribution of energy and materials within landscapes and help stabilise slopes, prevent water and wind erosion, and influence the transport of nutrients and sediments in them, maintaining landscape connectivity and affecting species movement and biodiversity (D'Odorico et al., 2010). Likewise, hydraulic redistribution by deep-rooted plants maintain soil moisture in dry seasons and prevent water table rises and subsequent soil salinization (Jayawickreme et al., 2011).

The improvement of soil infiltration by vegetation decreases peak flows following storms and increases the volume and duration of stored water in the landscape (Ryan et al., 2010) which in turn increments overall water availability, particularly in dry seasons.

Highlights

- Exchange of CO_2, oxygen and water vapour between vegetation and the atmosphere is controlled by stomatal resistance, boundary layer resistance, water vapor content and CO_2 concentration in the surrounding atmosphere and the substomatal cavity.
- Woody vegetation affects overland flow by deep rooted transpiration of soil moisture through the leaves of the canopy, by changing soil structure affecting infiltration rates, and by intercepting and evaporating precipitation through the soil-plant-atmosphere continuum.
- Vegetation controls the redistribution of runoff, water table levels and soil moisture by altering soil permeability.

2.7 Climate Response to Global Land Cover

Climate models with thorough representations of land surface processes are commonly used to identify the effects of vegetation on climate. This is due to the fact that vegetation-climate feedbacks at regional scales can hardly be established through direct observations (Bonan, 2008a). Climate models have been used to identify mesoscale circulations produced by discontinuities in surface heating associated with landscape patterns and patchiness (Kabat et al., 2004). It has been expected that, since contrasts between land and water generate mesoscale circulation breezes, landscape dissimilarities such as irrigated land in arid areas, deforestation, and afforestation would also produce mesoscale circulations of a similar magnitude (Lawton et al., 2001).

Since the extent of tropical deforestation, particularly in the Amazon (~414 million km^2 between 1988 and 2015 (INPE, 2013)), is one of the highest worldwide (Achard et al., 2002), climate models and land surface models imbedded in them have been applied to find climate variations due to changes in biophysical properties of the land surface in these regions. For instance, Costa & Pires (2010) using the National Centre for Atmospheric Research (NCAR) Community Climate Model performed experiments to assess the impacts of deforestation scenarios in the Amazon and central Brazil Cerrado on rainfall regime. They predict and increase from 5 months to 6 months of the dry season with changes in the biosphere-atmosphere equilibrium in the region. This results have been confirmed by flux tower measurements in eastern Amazonia (Rocha et al., 2004) and other model-based studies which also project a decrease in rainfall and increases in surface temperatures due to conversion from forest to pastures and soybean plantations in the Amazon (Costa and Pires, 2010; Sampaio et al., 2007). In tropical forests, the climate response to deforestation is driven by changes in evapotranspiration (Bonan, 2008b). Here, the presence of forest and deep-rooted vegetation produces a shallow, cool, and moist boundary layer that change to warmer and drier conditions after forest suppression and offset the cooling trend due to albedo increments when forest (low albedo) is replaced by croplands (higher albedo). However, it has been proposed that tropical deforestation could have contrasting climate impacts depending on the spatial scale and structure of the disturbance, and small scale, heterogeneous deforestation may produce mesoscale circulation that enhance clouds but with not enough energy to produce deep precipitation events (see review by D'Almeida et al., 2007).

In contrast to tropical forests, the climate response to deforestation in boreal forest seems to be driven by albedo changes. Using the Hadley Centre Atmosphere Model, Betts (2000) simulated the radiative forcing associated with albedo changes as a result of afforestation in temperate and boreal forest, where the presence of wooded area during the snow season warms the surface due to increases in energy availability at the near surface, offsetting the negative forcing that is expected from carbon sequestration (Betts, 2000). Flux Tower measurements have confirmed the warming trend due to the presence of forest compared to where there is an absence of trees (Baldocchi et al., 2000). It has been argued that the boreal forest, of all world biomes, has the greatest biophysical effect on mean annual global temperature (Snyder et al., 2004).

Climate responses to LUCC in other regions can have different signs depending on the type and direction of the conversion. For example, irrigation in arid and semi-arid climates of India and North America has altered local surface energy balance by decreasing surface albedo, increasing latent heat flux and decreasing sensible heat flux (Roy et al., 2007). This has generated mesoscale circulations that have decreased surface temperatures and increased rainfall (DeAngelis et al., 2010; Mahmood et al., 2006; Roy et al., 2007). Conversely, replacements from woody vegetation to non-irrigated agriculture could be related with warmer and drier boundary layer, as has been shown in semi-arid climates of Australia (McAlpine et al., 2007; Pitman et al., 2004). Similar results using climate models and observations have been found in semi-arid regions of Africa (Charney, 1975), Europe (Gates and Ließ, 2001) and South America (de Souza and Oyama, 2011) (also see review of Pielke et al., 2011).

Climate models have also been used to test the influence of specific biomes on climate. Using a coupled atmosphere-biosphere model, Snyder et al. (2004) performed six experimental simulations where tropical forest, boreal forest, temperate forest, savanna, grassland, and shrubland/tundra were individually replaced by bare soil to evaluate changes in surface climates. In their study, tropical forest creates a cooler and wetter climate, while boreal forest warms the surface climate. Temperate forest warms temperature in winter and spring be decreasing surface albedo of snow-covered areas, but it cools temperature in summer compared with bare soil because the lower albedo is offset by higher latent heat flux. Similarly to tropical forest, savanna and shrubland induces a cooler and wetter climate as a result of substantial increases in latent heat flux. Though generalist, and considering scaling and latitudinal dependences of the climatic response to changes land surface, these results could widen the understanding of the influence of specific biomes on temperature and precipitation at a regional and global scales.

Figure 2.6 shows a global view of the impacts of LUCC on surface temperature and precipitation based on studies using climate models and observation. Change direction is from natural vegetation (forest, shrubland and grassland) to different land uses (rainfed and irrigated agriculture, grazing, pastures or bare ground). As explained in the text, vegetation change in the tropics and subtropics is mainly associated with an increase (decrease) of surface temperature (rainfall) due to decreases (increase) in latent heat fluxes (sensible heat fluxes) (Costa et al., 2007; McAlpine et al., 2007), while in high-latitudes deforestation has led to decreases in temperature and precipitation due to increases in surface albedo and its associated changes in net radiation (e.g. Betts, 2000; Snyder et al., 2004). Figure 2.6 also shows rainfall (temperature) decreases (increases) in semi-arid environments of North Africa, where grazing and vegetation loss has decreases evapotranspiration and exacerbates droughts in the Sahel (Charney et al., 1975). The opposite response has been observed in the semi-arid shrublands of the Sinai region, where overgrazing and vegetation loss led to decreases in temperature and precipitation due to an albedo-driven climate response (Otterman, 1989). Regardless of broad patterns in the climate response to changes in land surface, regional and local change signs can be opposite compared with trends observed at continental or global scales (Mahmood et al., 2010; Pielke et al., 2011).

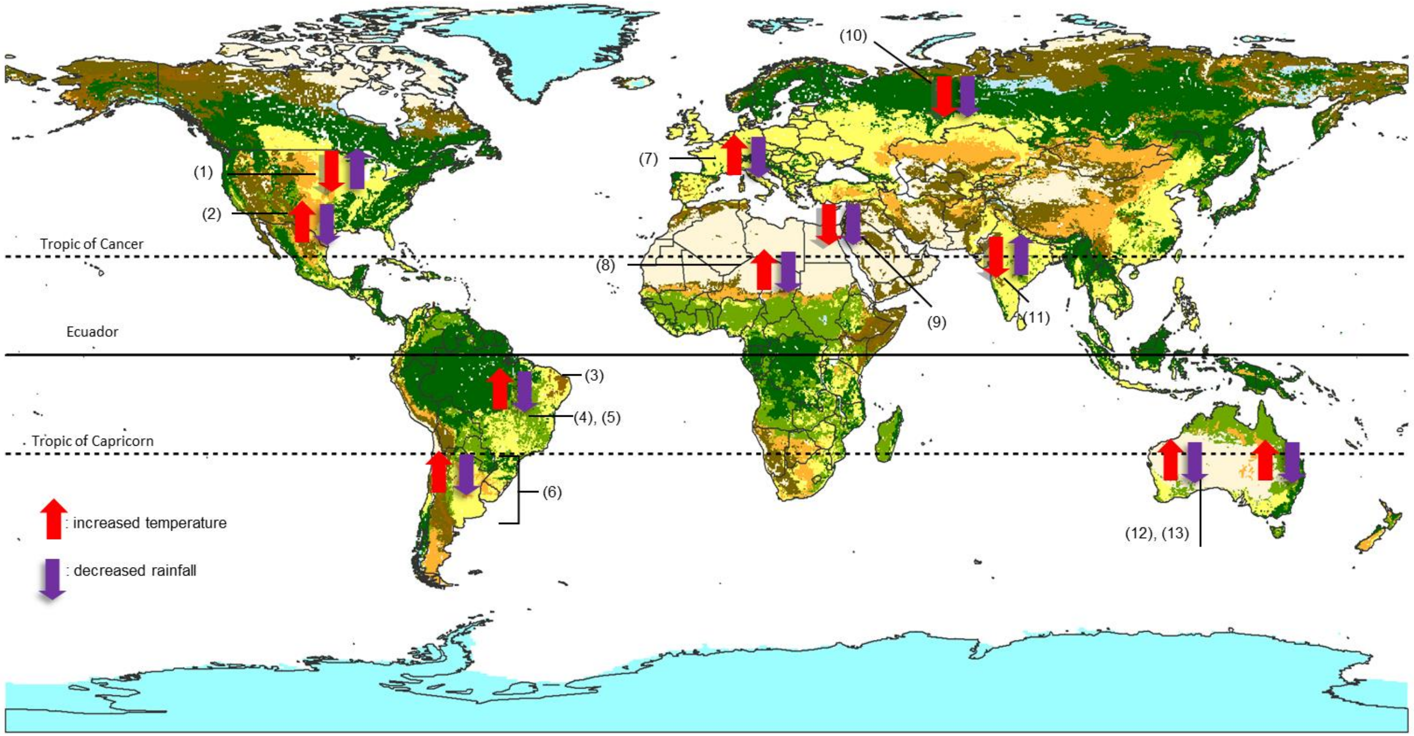

Figure 2.6 Geographic location of studies addressing LUCC impacts on temperature and rainfall using climate/surface models. Red arrow refers to changes in surface temperature. Blue arrow refers to changes in rainfall. Numbers correspond to the specific study as (1) Mahamud et al. 2004, (2) DeAngelis et al. 2010, (3) de Souza & Oyama, 2011, (4) Costa et al. 2007, (5) Sampaio et al. 2007, (6) Beltran-Przekurat et al. 2011, (7), Geist and Ließ, 2001, (8) Charney, 1975 (9) Otterman, 1989, (10) Betts, 2000, (11) Roy et al. 2007, (12) Pitman et al. 2004, (13) McAlpine et al. 2007. Land cover map from GlobCover2009.

CHAPTER 3 LAND USE AND LAND COVER CHANGE IMPACTS ON THE REGIONAL CLIMATE OF NON-AMAZONIAN SOUTH AMERICA: A REVIEW

Abstract

In relation to vegetation-climate interactions in South America, most of the attention has focused on the Amazon. However, even though land cover changes in non-Amazonian South America are higher than in the Amazon, we do not currently know the possible influences that these changes may have in climate and overall water availability. Non-Amazonian South America is recognized as an important part of the desertification prone area of the world, and the feedbacks between climate and land cover change dynamics can enhance this process with substantial ecological and socioeconomic impacts. This Chapter responds to the need of identify the level of scientific knowledge in relation to land use and land cover change-climate interactions in non-Amazonian South America and aims to identify the most impacted and least studied ecosystems in the region. This Chapter is composed by two sections. In the first, I use available remote sensing datasets in conjunction with regional studies to calculate natural vegetation loss from potential (pre-European) to present conditions. This analysis is focused on four main ecosystems: The Dry Chaco (Argentina, Bolivia and Paraguay), the Cerrado Savanna (Brazil), the Atlantic Forest (Brazil), Temperate Grasslands (Argentina and Uruguay), the Chilean Matorral (Chile) and Tropical Dry Forests (continental). The Chapter also considers the Amazon Biome for comparison proposes. In the second section, I review available studies for each one of non-Amazonian ecosystems and results are discussed in relation to climatic impacts of land use and land cover change, methodological approaches, risks and consequences, and research priorities.

This chapter has been published as Salazar, A., Baldi, G., Hirota, M., Syktus, J. and McAlpine, C. *Global and Planetary Change*, 128, 103-119. doi:10.1016/j.gloplacha.2015.02.009. For this reason, some information presented in Chapter 1 is repeated in this Chapter.

3.1 Introduction

Land use and land cover change (LUCC) affects climate through changes in moisture and energy budgets (IPCC, 2013b; Mahmood et al., 2014; Pielke et al., 2011). In South America, most of the focus on these impacts has been directed at deforestation of the Amazon forests (e.g. Pires and Costa, 2013). In contrast, non-Amazonian South America has received less attention despite experiencing the highest transformation rates in the tropics (Hansen et al., 2013; Marris, 2005). This is a significant problem because the loss of native ecosystems can modify the local and regional surface-climate coupling through feedback processes, and increase the risks imposed by climate extremes in an area that sustains a human population of over 200 million (Grimm and Tedeschi, 2009).

Non-Amazonian South America, also referred to as non-Amazonian ecosystems, covers an area of more than 12 million km^2 and is characterized by a high diversity of biomes including tropical rainforests, tropical savannas, grasslands, shrublands, deserts and a wide array of woodland formations that are distributed according to rainfall, temperature, soil properties and disturbance regimes. Precolonial pressures upon these biomes expressed through settlement, cultivation, grazing, hunting and burning by indigenous people (Knapp, 2007). However, these changes were temporary and therefore relatively rapidly reversed by ecological succession (Armesto et al., 2010). Since 1500 and especially since 1900, the expansion of European agriculture has resulted in widespread ecosystem transformations. Global demand for food commodities such as soybeans and beef has pushed the expansion of the agro-pastoral frontier into former natural and seminatural areas (Richards et al., 2012). Recent studies have shown high LUCC rates in tropical savannas of Brazil (hereafter referred as Cerrado) (Sano et al., 2010), grasslands in Argentina (Baldi et al., 2006), Atlantic forests in eastern Brazil (Joly et al., 2014) and the dry forests in the Paraguayan Chaco (Huang et al., 2009). Of the 542,000 km^2 of deforestation in South America between 2000 and 2012, 42% occurred in the Amazon region and 58% in the non-Amazonian region (Hansen et al., 2013).

Changes in land use and land cover can have profound impacts on land surface climate interactions by altering the exchange of heat, moisture, momentum, trace-gas fluxes and albedo (Bonan, 2008b). Cumulatively, they can impact the climate at a local (Hidalgo et al., 2010; Mohamed et al., 2011; Montecinos et al., 2008), regional (Fairman et al., 2011; Pitman et al., 2004; Roy et al., 2007) and even global scales (Avissar and Werth, 2005; Bounoua et al., 2002; Feddema et al., 2005; Lawrence et al., 2012; Snyder et al., 2004). Many of the studies addressing climatic impacts of LUCC focus on the tropical forest climates, particularly in the Amazon region. Results suggest that tree removal produces a drier and warmer climate due to reductions in evaporative cooling with implications to vegetation dynamics, river discharge and climate extremes (Costa and Pires, 2010; D'Almeida et al., 2007; McGuffie et al., 1995; Pires and Costa, 2013; Rocha et al., 2004; Sampaio et al., 2007). However, evidence relating climate and LUCC in other ecosystems of tropical and subtropical South America is scarce and dispersed. The high conversion rates of natural vegetation and the vulnerability of ecosystems to climate variability, create an increasing need to identify signals and patterns of the impacts of LUCC on the regional climate. This will better inform climate science and natural resource management. It’s been argued that climate impacts induced by LUCC are significantly comparable to those resulting from anthropogenic greenhouse gases (Pielke et al., 2002), particularly in local to regional scales, in which people and ecosystems are mostly affected (Mahmood et al., 2010). Though there is a good understanding of the major biogeophysical feedbacks of Amazon deforestation, land surface climate interactions and their consequences in non-Amazonian South America are much less understood.

In this paper, we review the modeling and empirical evidence that shows the climatic impacts of LUCC in non-Amazonian ecosystems of South America. First, we estimate the original and remaining amount of natural vegetation in the Amazon and in six non-Amazonian ecosystems. We then assess the impacts of LUCC on the climate of non-Amazonian South America and evaluate the implications and potential risks with regard to climate change and future research priorities.

3.2 Methods

3.2.1 Delimiting the Amazon and non-Amazonian South America

We focused on six broad ecosystems, collectively referred as non-Amazonian South America. We also considered the Amazon biome as defined by (WWF, 2010) to compare surface climate feedbacks studies between Amazonian and non-Amazonian South America. We defined non-Amazonian ecosystems in South America based on two criteria: 1) they must be located outside the area covered by the Amazon biome and 2) they must exhibit at least one peer-reviewed study describing impacts of land use and land cover change (LUCC) on local or regional climate (see 2.4). We geographically delimited them using Olson et al. (2001), MMA/IBAMA (2011b) and MMA/IBAMA (2011a). The final selection covered an area of about 6.3 million km^2 and included: 1) Dry Chaco, 2) Cerrado, 3) Temperate Grasslands, 4) Chilean Matorral, 5) Tropical Dry Forests and 6) Atlantic Forest (Figure 3.1). These ecosystems represent a variety of functional groups including moist forests, dry broadleaf forests, grasslands, savannas, shrublands, mediterranean forests, and xeric shrublands. All of them have been subjected to extensive anthropogenic modification (Friedl et al., 2010; Olson et al., 2001).

3.2.2 Estimating potential and current natural vegetation cover

LUCC information in South America is highly fragmented and localized. For this reason, we estimated potential and current natural vegetation extent for both regions using different sources: 1) peer-reviewed publications, 2) technical reports and 3) the Collection 5 MODIS Global Land Cover Type for year 2012 (Friedl et al., 2010). We first defined potential forest cover (natural) in the Amazon region as the total area described in WWF (2010) without considering savanna ecoregions as classified by Olson et al. (2001). Then we extracted areas covered by evergreen broadleaf forests in these savannas according to Collection 5 MODIS Global Land Cover Type for year 2012. This procedure added those forests (e.g., gallery forests) distributed in areas dominated by savanna vegetation (e.g., Beni Savannas in Figure 3.1) inside the Amazon region and gave us the approximate potential extent of dense moist tropical forest in the Amazon.

We obtained the potential historical natural vegetation extent in non-Amazonian South America from regional and local studies. We used Olson et al. (2001) classification for the Dry Chaco, Temperate Grasslands and the Atlantic Forest; MMA/IBAMA (2011b) for the Cerrado; MMA/IBAMA (2011a) for the Caatinga; Portillo-Quintero and Sánchez-Azofeifa (2010) for the Tropical Dry Forests; and Luebert and Pliscoff (2006) for the Chilean Matorral. We included Caatinga into Tropical Dry Forests as suggested by Portillo-Quintero and Sánchez-Azofeifa (2010). However we presented vegetation change results separately since we used two different sources for calculating the vegetation cover change for these regions (Table 3.1).

We obtained current natural vegetation cover from peer-reviewed studies and Collection 5 MODIS Global Land Cover Type for the year 2012. We calculated current vegetation area in the Amazon biome and the Atlantic Forests as that covered by evergreen broadleaf forest in the MODIS product. We adopted the same procedure to define the current extent of Temperate Grasslands (grasslands in the MODIS product). For the Dry Chaco ecosystem we used the work of Clark et al. (2010a) who classified forest cover at 250 m resolution using MODIS. We used IBAMA's (Brazilian Institute of Environment and Renewable Natural Resources) estimation of remnant natural vegetation for the Cerrado (MMA/IBAMA, 2011b) and the Caatinga (MMA/IBAMA, 2011a). Current extent of forests and shrublands in the Chilean Matorral was obtained from (Conaf, 1999). For Tropical Dry

Forests, we used the forest cover area calculated by Portillo-Quintero and Sánchez-Azofeifa (2010). The maps used to calculate current natural vegetation area were the most complete and accurate we had access to (Table 3.1). We used the Albers Equal Area projection and South American 1969 Datum for all maps.

Table 3.1 Methods for potential and current natural vegetation cover estimation in the Amazon Biome and non-Amazonian South America

Ecosystem	Reference used to estimate potential vegetation extent	Methods for potential vegetation extent	Reference used to estimate current vegetation extent	Methods for current vegetation extent as described in the references used
Amazon Biome	Olson et al. (2001); Friedl et al. (2010); WWF (2010)	WWF (2010) biome boundary with substracted savanna areas as defined by Olson et al. (2001) not covered by evergreen broadleaf forests according to Collection 5 MODIS for year 2012	Friedl et al. (2010)	Classified as evergreen broadleaf forest in Collection 5 MODIS for year 2012
Chaco	Olson et al. (2001)	Covered by dry forest according to Olson et al. (2001)	Clark et al. (2010)	MODIS 250 m vegetation index product (MOD13Q1) for year 2006
Cerrado	MMA/IBAMA (2011b)	Covered by savanna according to MMA/IBAMA (2011b)	MMA/IBAMA (2011b)	Classification of Landsat TM images for year 2009
Temperate Grasslands	Olson et al. (2001)	Covered by grasslands according to Olson et al. (2001)	Friedl et al. (2010)	Classified as grasslands in Collection 5 MODIS for year 2012
Chilean Matorral	Luebert and Pliscoff (2006)	Covered by forest and shrubland according to Luebert and Pliscoff (2006)	Conaf (1999)	Classification based on aerial photo interpretation
Tropical Dry Forests	Portillo-Quintero and Sánchez-Azofeifa (2010)	Olson et al. (2001) ecoregions defined as Tropical Dry Forests.	Portillo-Quintero and Sánchez-Azofeifa (2010)	Supervised classification of MODIS surface reflectance imagery at 500-m resolution for year 2004
Tropical Dry Forest-Caatinga	MMA/IBAMA (2011a) and IBGE (2012)	Savanna and forests given MMA/IBAMA (2011a)	MMA/IBAMA (2011a)	Classification of Landsat TM and CBERS - 2B CCD images for year 2009
Atlantic Forest	Olson et al. (2001)	Covered by forest according to Olson et al. (2001)	Friedl et al. (2010)	Classified as evergreen broadleaf forest in Collection 5 MODIS for year 2012

3.2.3 Recent Change in Forest Cover

Based on Hansen et al. (2013), we explored the recent changes in forest cover for the Amazon Biome and non-Amazonian South America. Hansen et al. (2013) used Landsat time-series data to quantify global forest loss and gain at a spatial resolution of ca. 30 m for the period 2000-2012. However, this dataset does not discriminate between native forest and exotic tree plantations. Since some regions of South America are now used for intensive forestry practices with high rates of forest loss and gain (Jobbágy et al., 2012), we discuss the Hansen et al. (2013)'s results in conjunction with local studies in order to better understand forest cover dynamics in non-Amazonian South America.

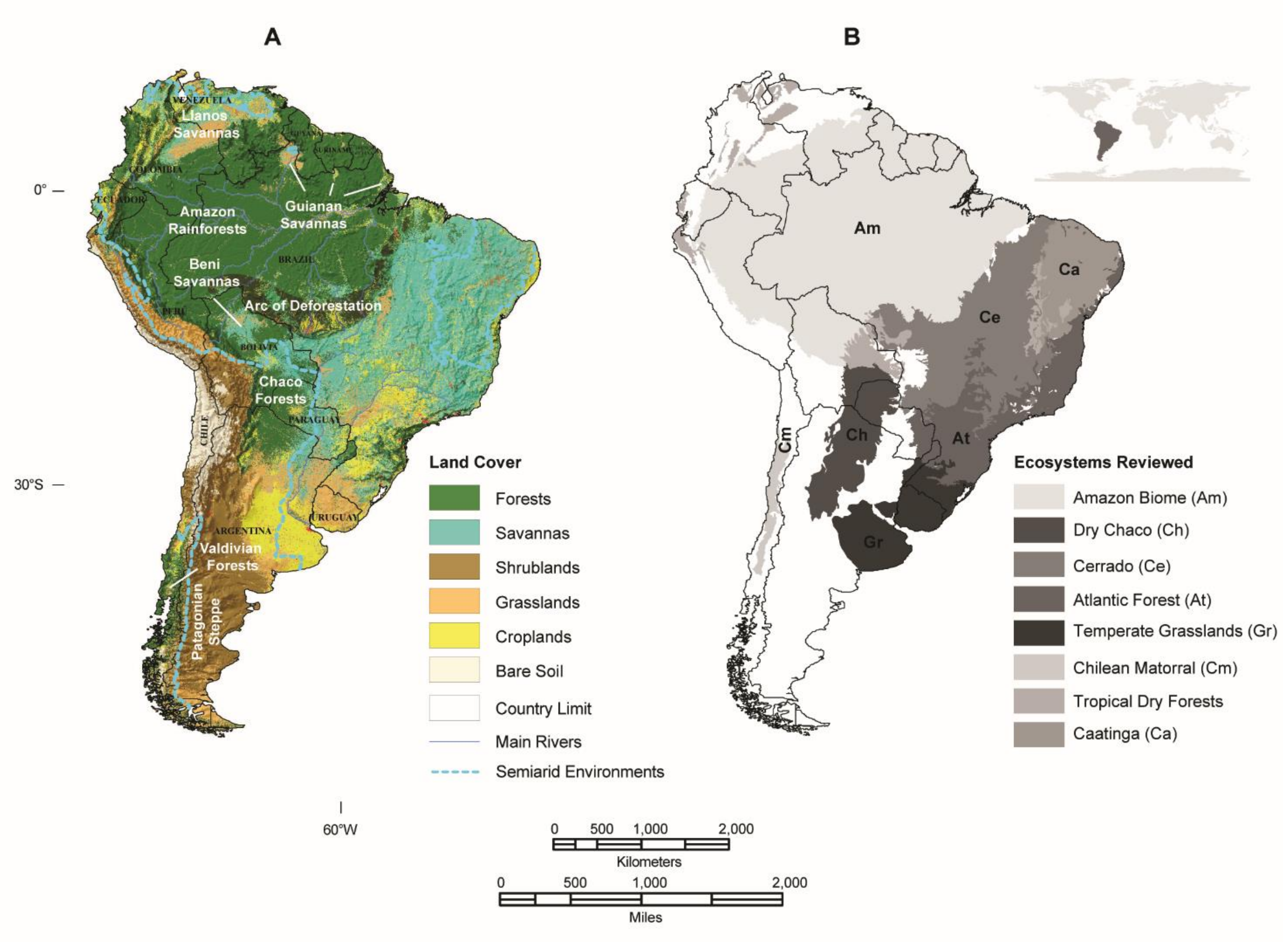

Figure 3.1 (A) Land cover map of continental South America based on Collection 5 MODIS Land Cover Type product for year 2012 and (B) ecosystems of non-Amazonian South America reviewed in this article (Amazon biome also shown). Other non-Amazonian ecosystems not included in this study are displayed as white in map (A) (e.g., Llanos Savannas in Colombia and Venezuela). Semiarid environments (Aridity Index <0.65) are derived from Trabucco and Zomer (2009). The southern Amazon arc of deforestation is also shown. Ecosystems limits in map (B) were obtained from Olson et al. (2001) for Dry Chaco, Atlantic Forest, Temperate Grasslands and Chilean Matorral. The maps from MMA/IBAMA (2011a) and MMA/IBAMA (2011b) were used to define the Cerrado and the Caatinga ecosystems, whereas Portillo-Quintero and Sánchez-Azofeifa (2010) was used for Tropical Dry Forests.

3.2.4 Review of LUCC impacts on climate

We searched peer-reviewed literature in Web of Science database from year 1900 to 2013 using a combination of the following key words: "climate", "land cover change", "South America", "land use change", "deforestation" and "ecosystems' names". Due to the spatial extent of atmospheric studies, the same work might be cited for two or more ecosystems of non-Amazonian South America. We also searched Web of Science for LUCC climate feedbacks in the Amazon using the key words: "deforestation", "Amazon", "climate", "impact", "land cover change" and "land use change" for the period 1993-2013. We focused the search only on biophysical impacts.

From the bibliographic lists of these articles and from previous literature searches conducted by the authors, we included supplementary articles referring to LUCC and climate feedbacks for non-Amazonian South America. We considered articles only if they met the following criteria:

1) The article must be published, peer reviewed and written in English.
2) The article must have focused on a geographic area outside the Amazon Biome.
3) The article must have explicitly referred to LUCC processes (e.g. conversion from forest to crops).
4) The article must have contained information of LUCC impacts for at least one of the following climatic components: temperature, precipitation and albedo.

Moreover, LUCC articles were also searched to identify the immediate and underlying causes of these dynamics in non-Amazonian South America. The key words used in this search were: "land cover", "South America", "land use", "ecosystem's name", "land cover change", "land use change" and "deforestation". Relevant studies identified from the bibliographic lists of the articles were also included.

Based on the aforementioned criteria, we included a total of 19 LUCC climate feedback related studies for non-Amazonian South America. For each article, we recorded location, study type (modelling or observational), study period, LUCC direction and impacts on temperature, precipitation and albedo (Table 3.3; an expanded version of this Table is shown in Appendix A). The selection of articles in this review was used to choose the regions classified as non-Amazonian South America. This implied that other non-Amazonian ecosystems were not included because we did not find LUCC climate feedbacks studies for these ecosystems. Examples are represented by the Llanos savanna of Venezuela and Colombia (Etter et al., 2008; Portillo-Quintero et al., 2012; Romero-Ruiz et al., 2012), the Patagonian Steppe in southern Argentina (Bisigato and Laphitz, 2009; Paruelo et al., 2001), the Valdivian forests in southern Chile (Farley, 2007; Huber et al., 2008) and Ecuadorian Páramo (Farley, 2007), amongst others.

We acknowledge that land cover datasets and models have varying levels of accuracy and methodologies which limit the ability to make comparisons. We recognize that these differences provide bounds of uncertainty on the major findings, yet they do not invalidate the major conclusions presented here. The main focus of the paper is on the available evidence of LUCC impacts on climate in non-Amazonian South America and is not possible to cover in detail data inaccuracies of the varying approaches.

3.3 Results

3.3.1 General Trends in LUCC

Historically, the non-Amazonian ecosystems of South America have lost more than 3.6 million km^2 (58% of their potential natural vegetation). This is equivalent to about 4 times the historic Amazon deforestation (918,473 million km^2). Recent forest loss (period 2000-2012) in non-Amazonian ecosystems (excluding Temperate Grasslands) accounts for 45% of total forest loss in South America (241,551 km^2), compared to the loss of rainforest in the Amazon Biome which represents 42% (227,249 km^2) of total South American deforestation (541,887 km^2).

The ecosystem relatively most impacted by LUCC was the Chilean Matorral, where 83% of its potential natural vegetation had been transformed to other land uses by 1999 (date of the current vegetation map). This ecosystem also showed a high loss and gain in forest area for period 2000-2012 (Table 3.2 and Figure 3.3), indicating the presence of exotic tree plantations as reported by different studies (e.g. Niklitschek, 2007). The second most relatively impacted was the Atlantic Forest, with 81% (978,031 km^2) of its potential extent lost by 2012. It also experienced high rates of forest loss and gain between years 2000 and 2012 (Table 3.2). Together with the Chilean Matorral and the Cerrado, the Atlantic Forest shows the greatest area of exotic tree plantations in South America (Jobbágy et al., 2012). Conversely, the Dry Chaco exhibited the lowest relative extent of historic transformation (34%). Yet, it showed the highest deforestation rate for tropical forests between the years 2000 and 2012 (Hansen et al., 2013) and a very low forest gain in the same period (Table 3.2). The Dry Tropical Forests have also undergone high historic deforestation and presently cover approximately 40% of their former extension (Sánchez-Azofeifa and Portillo-Quintero, 2011). The most studied areas are the Caatinga in northeast Brazil and the Chiquitano forests in Bolivia, with limited references found for the Tropical Dry Forests of Colombia, Ecuador, Venezuela and Peru, where the remaining forest area is less than 6% of its potential extent (Portillo-Quintero and Sánchez-Azofeifa, 2010). In the case of the Cerrado, 52% has been converted into crops and pastures over an area of about 1 million km^2 (MMA/IBAMA, 2011b). Interestingly, even though the Temperate Grasslands were formerly composed by grassland vegetation, only 30% of which remained by year 2012, they showed high dynamic forest area between 2000 and 2012, presumably due to an increased area of exotic tree plantations (Hansen et al., 2013; Jobbagy and Jackson, 2007; Nosetto et al., 2012).

Table 3.2 Changes in natural vegetation cover for the Amazon Biome and non-Amazonian South America. Ecosystems borders were taken from Olson et al. (2001).

Region	Main vegetation type	Potential historic area (km²)	Present area (km²)	Converted area (km²)	Converted area (%)	Recent forest cover change for period 2000-2012 based on Hansen et al. (2013)	
						Loss (km²)	Gain (km²)
Amazon Biome	Forest	6,546,242	5,627,769	918,473	14	227,249	15,972
Dry Chaco	Forest	786,790	516,011	270,779	34	62,815	695
Cerrado	Savanna	2,039,386	983,348	1,056,038	52	87,274	21,691
Atlantic Forest	Forest	1,204,467	226,436	978,031	81	44,658	33,056
Temperate Grasslands	Grasslands	777,571	236,240	541,331	70	5,562	12,181
Chilean Matorral	Forest and shrublands	62,935	10,751	52,184	83	2,127	3,065
Tropical Dry Forests	Forest	664,191	268,877	395,316	60	27,661	2,404
Caatinga	Forest and shrublands	787,968	431,877	356,091	45	17,016	1,634
Total Amazon Biome	-	**6,546,242**	**5,627,769**	**918,473**	**14**	**227,249**	**15,972**
Total non-Amazonian	-	**6,323,308**	**2,673,538**	**3,649,770**	**58**	**247,113**	**74,726**
Total South America	-	**-**	**-**	**-**	**-**	**541,887**	**118,532**

In terms of vegetation climate feedbacks and associated publications, results showed clear differences between the Amazon and non-Amazonian South America. The Amazon presents an historic land cover change area (loss of rainforests) of about 920,000 km^2 with 54 publications addressing the associated climate impacts. In contrast, historic LUCC in non-Amazonian South America totalled 3.6 million km^2, and its climatic effects were addressed by 19 publications (Figure 3.2). Of these studies, 70% focused on the Cerrado and the Tropical Dry Forests including Caatinga, whereas just one publication was found for the Atlantic Forest, where 978,031 million km^2 of the estimated original forest cover has been cleared. Most of the studies in non-Amazonian ecosystems were accomplished using climate or surface models and only 4 observational-based publications were conducted using remote sensing and weather stations.

Table 3.3 Summary of LUCC impacts on temperature, rainfall and albedo in non-Amazonian South America for 19 studies reviewed. Numbers represent the amount of peer-reviewed publications and signs represent the direction of change in temperature, rainfall and albedo (e.g., 2+: two publications reporting an increase in temperature for the specific LUCC direction, e.g., from woody to crops in the Dry Chaco). Woody refers to woody vegetation (forests, savannas or shrublands).

Ecosystems	LUCC		Temperature	Rainfall	Albedo	Reference
	From	To				
Dry Chaco	Woody	Crops	2+;1-	1+;1-	2+	1.Houspanossian et al. (2013); 2.Loarie et al. (2011a); 3.Canziani and Carbajal Benitez (2012); 4.Lee and Berbery (2011); 5.Beltrán-Przekurat et al. (2012b)
	Grassland	Crops	1+	1-	1+	Beltrán-Przekurat et al. (2012)
	Crops	Woody	1-	1+	No impact	Beltrán-Przekurat et al. (2012)
Cerrado	Woody	Crops	2+	2-	NE[a]	6. Costa and Pires (2010); 7. Lee et al. (2011); 8.Pongratz et al. (2006); 9. Mendes et al. (2010)
	Woody/crops	Sugarcane	1+; 1-	1-	1+	10.Georgescu et al. (2013)
	Woody	Crop/Pasture	1+	NE	1+	11.Loarie et al. (2011b)
	Crop/Pasture	Sugarcane	1-	NE	1+	Loarie et al. (2011b)
	Woody	Pasture	1+	NE	NE	Pongratz et al. (2006)
Temperate Grasslands	Grassland	Crops	2-; 1+	2-; 3+	3+; 1-	Beltrán-Przekurat et al. (2012); Lee and Berbery (2011); Loarie et al. (2011a)
	Grassland	Woody	1-	1+	1+	Beltrán-Przekurat et al. (2012)
Chilean Matorral	Woody	Crops	1-	NE	1-	12.Montecinos et al. (2008)
	Grasslands	Crops	1+	No impact	1+	Beltrán-Przekurat et al. (2012)
Tropical Dry Forests-Chiquitano	Woody	Crops	1+	NE	1+	13.Bounoua et al. (2004)
	Grasslands	Crops	1+	NE	No Impact	Bounoua et al. (2004)
Tropical Dry Forests-Caatinga	Woody	Desert	2+	3-	2+	14. Oyama and Nobre (2004); 15.Castilho de Souza and Oyama (2011); 16.Hirota et al. (2011)
	Savanna albedo	Desert albedo	NE	1-	NE	17.Sud and Fennessy (1982)
	Normal conditions by 1979	Evaporation supressed	NE	1-	NE	18. Sud and Fennessy (1984)
Atlantic Forest	Woody	Crops/pastures	NE	1-	NE	19. Webb et al. (2005)

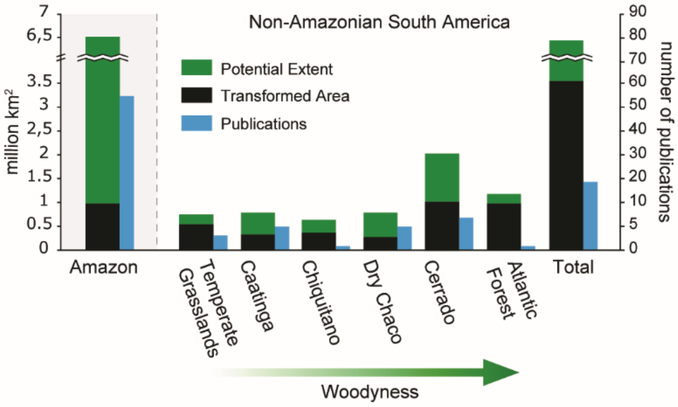

Figure 3.2 Comparison of the geographic extent of LUCC and the number of publications documenting the climatic impacts of LUCC for the Amazon and non-Amazonian ecosystems (Chilean Matorral not shown). Dark green and back bars represent potential natural vegetation extent and transformed area, respectively. Blue bars indicate number of publications.

In the following sections, we present the results for specific ecosystems. For each non-Amazonian ecosystem, the results are organized based on patterns of LUCC and evidence of climatic impacts according to modeling and observational studies.

3.3.2 LUCC and its Climate Impacts in non-Amazonian South America

3.3.2.1 Dry Chaco

LUCC

Former land use of the Dry Chaco was influenced by aboriginal people who used fire as a management tool to modify vegetation for hunting, communication and war (Gordillo, 2010). These activities were mainly conducted in grasslands occurring on sandy soils near rivers and in ancient river beds (Morello and Adamoli, 1974). Following European settlement and especially during the first half of the 20th century, extensive cattle ranching, logging, firewood and charcoal extraction led to changes in herbaceous/woody vegetation dynamics and in the forest cover with agriculture occurring in the foothills of the Andes and humid valleys (Adamoli et al., 1990; Bucher and Huszar, 1999).

LUCC accelerated during the second half of the 20th century. From the 1990s, a synergistic combination of global food demand, technology and climatic factors increased the rate of land cover change in the Chaco to those comparable with the Amazon deforestation (Boletta et al., 2006; Zak et al., 2008). Methods for monitoring Dry Chaco deforestation vary from visual interpretations of aerial photos to digital classifications of multi-temporal satellite imagery (Clark et al., 2010a; Gasparri and Grau, 2009; Gasparri et al., 2008; Paruelo et al., 2006; UMSEF, 2007). All studies document high rates of deforestation. In the Argentinian Chaco, LUCC intensified after the 1980s predominantly over flat terrains where rainfall supported rainfed agriculture (Gasparri and Grau, 2009; Gasparri et al., 2008; Grau et al., 2005a; Paruelo et al., 2006). Most studies from the Chaco show consistent pattern of the replacements of native forests by pastures and croplands, particularly soybean plantations (Clark et al., 2010a; Hoyos et al., 2013). In some areas, such as the Cordoba and Santiago del Estero provinces, deforestation rates of 2% to 5% per year have been reported (Boletta et al., 2006; Zak et al., 2004), which is higher than deforestation in some of the world's humid tropical forests (Achard et al., 2002). Recent observations using Landsat imagery show that between 2010 and 2011, more than 600,000 ha of the Dry Chaco ecosystem was deforested, 86% of which occurred in Paraguay, 12% in Argentina and about 2% in Bolivia (Rodas et al., 2012). This is corroborated by the results of Hansen et al. (2013), who described the Dry Chaco as registering one of the highest rates of tropical forest loss between the years 2000 and 2012, with a total of 62,800 km^2. The continued growth in the global demand for soybean, technological advances and the use of transgenic crop varieties to overcome climate limitations, are expected to increase deforestation in the Dry Chaco, particularly in those areas with more fertile and moister soils (Grau et al., 2005b).

Impacts on climate

Modeling studies. In a study covering southern South America and including the Dry Chaco, Beltrán-Przekurat et al. (2012) applied a regional climate model to evaluate near-surface changes resulting from conversions of pre-European vegetation to present day land cover and under future afforestation scenarios. The conversion of wooded vegetation to soybean plantations decreased surface parameters such us roughness length, leaf area index and rooting depth, and decreased latent and sensible heat fluxes. These changes resulted in an increase in the 2 m surface temperature up to 0.6 °C during dry years, with uncertain effects on rainfall. Similarly, the study of Canziani and Carbajal Benitez (2012) reported a temperature increase in 1 °C during austral winter (dry season) and spring over the deforested areas of the Chaco and beyond, yet without clear impacts on precipitation. By contrast, Lee and Berbery (2011) using the Weather Research and Forecasting Model (WRF), found less surface temperatures and rainfall when crops replaced savanna and evergreen broadleaf forests. According to the authors, the decrease in temperature was triggered by a significant increase in surface albedo and subsequent decrease in sensible heat fluxes, while the decrease in precipitation was related to a reduction in moisture convergence because of stronger low level winds that favoured the advection of large amounts of moisture out of the deforested areas.

Observational studies. In the Dry Chaco, remote sensing approaches have been used to identify effects of land cover change in surface temperatures and albedo. For instance, Houspanossian et al. (2013) combined the remote sensing and modeling techniques to calculate differences in temperature and albedo between dry forests and crops. Based on satellite images, the authors report a black-sky albedo ca. 50% higher in croplands (mainly soybean but also corn, sunflower, wheat, and rye) compared to dry forests. These results agree with those described by Loarie et al. (2011a), who found that forest–agriculture conversions in the Chaco are responsible for about 7% of albedo increases in South America between 2000 and 2008. Houspanossian et al. (2013) also reported temperatures 1.6 to 5 °C higher in croplands than in dry forests, which they attribute to the cooling effect of the higher evapotranspiration rates of dry forests compared to rainfed croplands.

Other studies not included in this review have also highlighted the potential influences that land surface processes likely associated to LUCC, specifically rainfed agricultural practices, may exert in the Chaco's surface climate. According to Collins et al. (2009), observed increases in surface temperature from 1948 over specific areas of tropical and subtropical South America cannot be explained solely by El Niño or La Niña events and might be the result of human activities such as land use change and/or increased levels of anthropogenic greenhouse gases. Moreover, Nuñez et al. (2008), for the period 1961–2000, applied an "observation minus reanalysis" method to estimate the potential links between land cover change and surface temperature change over Argentina, identifying a warming trend of minimum and maximum temperatures in northern and northeastern areas of the country, which have experienced high land conversion rates during recent decades (Viglizzo et al., 2011). However, in the study of Nuñez et al. (2008) there was no clear link between changes in temperature, precipitation, and changes in land cover.

Overall, studies from the Dry Chaco suggest that the dry forests may induce a cooler and wetter climate as a result of presenting higher latent heat fluxes. Here, the presence of deep-rooted forest and woodland vegetation can produce a shallower, cooler, and moister boundary layer that shifts to warmer and drier conditions after conversion to croplands. This offsets the cooling trend associated with albedo increases when forest (low albedo) is replaced by croplands (higher albedo) (Beltrán-Przekurat et al., 2012; Houspanossian et al., 2013). Yet, the impact on precipitation is not as clear as with surface temperature, showing a positive trend during dry years and negative during wet years (Beltrán-Przekurat et al., 2012). In this regard, a complementary study conducted by Saulo et al. (2010) identified that local feedback effects occur between land and precipitation in subtropical Argentina, and that these are expressed by variations in soil moisture (which is partially controlled by photosynthetic respiration) and consequently potential influences on evaporation, convective available potential energy, and hence, precipitation.

The modeling and observational studies for the Chaco generally agree in the positive trend of surface temperature without clear impacts on precipitation after deforestation. Comparison of the results is limited however, due to differences in modeling settings, period of analysis, spatial resolution and input datasets. In this regard, modeling studies using global climate models better account for land–atmosphere feedbacks and interactions with neighbouring areas, and therefore are the most suitable approach for LUCC–climate interactions.

3.3.2.2 Cerrado

LUCC

During the last 30 years, the Brazilian Cerrado has been rapidly transformed through large-scale agriculture into one of the world's most threatened ecoregions (Machado et al., 2004). Before the 1970s, land use in the Cerrado was dominated by low-impact cattle ranching on native vegetation (Sano et al., 2010). However, in recent years, planted pastures and the introduction of extensive and mechanized agriculture including soybean production has transformed the Cerrado savanna into a commercial agro-pastoral landscape (Brannstrom et al., 2008). Brazil is the world's second-largest (after U.S.A.) soybean producer with a production that increased from 1.5 million tonnes in 1970 to 74.8 million tonnes by 2011, 60% of which is concentrated in the Cerrado (FAO, 2013; Jepson et al., 2010; Smaling et al., 2008).

According to Machado et al. (2004), 55% of the Cerrado natural vegetation was cleared by 2002 at rates that would remove all natural vegetation in the region by 2030. However, differences in methodological approaches such as geographic boundaries, mapping scale and remote sensing approaches, have reported land conversion areas between 40% and 80% (e.g. Alho and Martins, 1995; Sano et al., 2010). The southern Cerrado has experienced the highest transformation rate with a clearing frontier expanding north where most of the natural vegetation remains (Diniz-Filho et al., 2009) (Figure 3.1). According to MMA/IBAMA (2011b), more than 980,000 km^2 (54%) of the Cerrado's natural vegetation has been converted into other land uses. Hansen et al. (2013)'s datasets show highly dynamic forest area in the ecosystem: from year 2000 to 2012, 87,274 km^2 of the forested area was removed. However, woody cover increased over almost 22,000 km^2 within the same period, probably because of the expansion of exotic tree plantations (Sano et al., 2010). The main causes of Cerrado's LUCC are linked to explicit state development policies (Klink and Moreira, 2002), global soybean demand, and increasing forestry and sugarcane for biofuel production (Jobbágy et al., 2012; Loarie et al., 2011b).

Impacts on climate

Modeling studies. The few modeling studies for the Cerrado have addressed land surface climate feedbacks in the transitional area between the tropical Amazon rainforests and the Cerrado semi-deciduous forests, the zone known as the “arc of deforestation” (Pongratz et al., 2006; Costa and Pires, 2010; Mendes et al., 2010) (Figure 3.1). Studies using climate models generally agree in the temperature and rainfall response when woody vegetation is replaced by crops. For instance, Georgescu et al. (2013) simulated the replacement of Cerrado vegetation by sugarcane using the WRF model, reporting a surface cooling up to ca. 1.0 °C during the growing season, and a warming of similar magnitude after harvesting. In their study, the cooling was due to an increased albedo, the warming was influenced by a decline in evapotranspiration and increased sensible heating, while total rainfall also decreased. A drying trend was also described by Costa and Pires (2010) who report a decrease in moisture fluxes and consequently increase the duration of the dry season from the current 5 to 6 months when the cumulative influence of the Amazon deforestation is considered, and by Lee et al. (2011), who project a reduction in total rainfall during the dry season and a warming that increases the risk of more frequent and severe droughts.

In agreement with the warming trend shown by the climate models, land surface modeling also projects an increase in surface temperatures following deforestation in the Cerrado. A study by Pongratz et al. (2006) used a land surface model to evaluate the effects of vegetation changes on the local energy and water balances in the north-central state of Mato Grosso. Here, the conversion of transitional forests, composed by both Amazon and Cerrado vegetation, to cropland resulted in an increased canopy temperature of up to 0.7 °C at midday. Similarly, the conversion of transitional forests to pasture caused an increase in maximum temperature of ca. 0.5 °C, driven by a reduced roughness length and increased aerodynamic resistance. These compensated the cooling trend associated with higher physiological activity of pastures (C4 photosynthetic pathway) compared to transitional forests (C3 photosynthetic pathway). In addition, temperature response is intensified when transitional forest are cleared and converted into bare soils, resulting in a temperature anomaly of 1.2 °C in the dry season (Pongratz et al., 2006). Interestingly, these modelled impacts induced by the loss of woody vegetation in the Cerrado, can potentially affect neighbouring areas of the Amazon biome and can enhance the transition from rainforest to Cerrado type vegetation in the next 40 years due to a drier climate associated with Cerrado deforestation (Mendes et al., 2010).

Observational studies. Evidence from observations of the climatic effects of vegetation loss in the Cerrado generally agrees with those described by modeling studies. Satellite images were used by Loarie et al. (2011b) to evaluate the climate effects of crop/pasture and sugarcane expansion in the Brazilian Cerrado. In their study, transformations from natural vegetation to crop/pasture triggered a decrease in evapotranspiration and an increase in average surface temperature of 1.6 °C. On the other hand, conversions from crop/pasture to sugarcane plantations lead to a mean cooling of 0.93 °C due to an increase in evapotranspiration and in the albedo, with the former exerting the greatest influence on the surface temperature response.

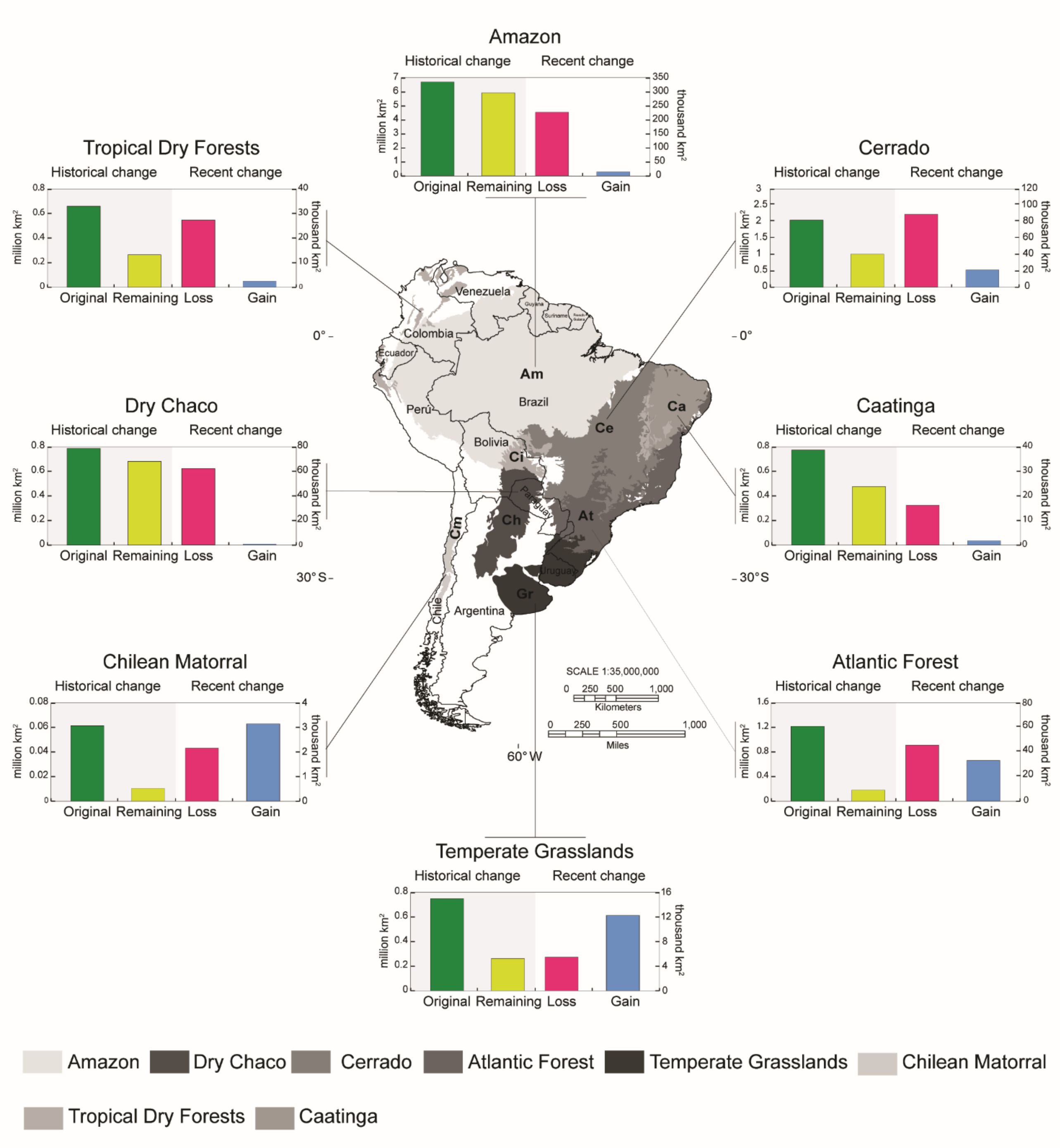

Figure 3.3 Distribution and land cover change in the Amazon biome and non-Amazonian South America: Amazon (Am), Dry Chaco (Ch), Cerrado (Ce), Atlantic Forest (At), Temperate Grasslands (Gr), Chilean Matorral (Cm) and Tropical Dry Forests including Caatinga (Ca) and Chiquitano (Ci). For non-Amazonian South America, original and remaining vegetation area (green and yellow bars) was obtained using ecoregions of Olson et al. (2001) and literature. For the Amazon, the original forest cover was calculated as the total biome area without considering savannas according to Olson et al. (2001), and remaining forest area was calculated from Collection 5 MODIS for year 2012 at 500 m resolution. Recent change (red and blue bars) represents changes in forest cover for period 2000-2012 and was taken from Hansen et al. (2013) datasets. Forest gain, particularly in the Atlantic Forest, Chilean Matorral and Temperate Grasslands are linked to exotic tree plantations (see main text).

As in the Chaco, the climatic response after land cover change in the Cerrado depends largely on changes in energy fluxes rather than albedo changes. Water uptake by deep-rooted vegetation is released to the lower atmosphere through evapotranspiration and contributes significantly to Cerrado's water balance. Therefore, it is expected that the replacement of woody vegetation by crops and pastures change the hydrological cycle of the Cerrado (Oliveira et al., 2005).

3.3.2.3 Atlantic Forest

LUCC

Even though the Atlantic Forest is recognized as one of the most biodiverse ecoregions in the world, with more than 20,000 plant species, over 1360 vertebrate species and high levels of endemism (Myers et al., 2000), it remains as one of the most threatened tropical forests. Almost 1 million km^2 or 81% of its original extent has been converted with increasing deforestation in Paraguay and Argentina (Table 3.2).

In Brazil, several studies have estimated a remaining forest cover of between 1% and 12% (Câmara, 2003; Morellato and Haddad, 2000; Oliveira-Filho and Fontes, 2000; Saatchi et al., 2001). Recently, Ribeiro et al. (2009) suggested that the Atlantic Forest extends over between 11% and 16% of its original cover in Brazil, most of which is distributed in small patches in a highly fragmented state. According to MMA/IBAMA (2012), the remaining forest by 2002 was 22% of its former extent.

Deforestation of the Atlantic Forest began with the arrival of European colonizers who exploited the commercially-valuable Brazilwood (Caesalpinia echinata) and cleared the rainforests for cropping and human settlements (Câmara, 2003). In the 18th century, the introduction of sugar cane plantations triggered rapid deforestation on fertile soils of the northeastern coast, while the introduction of coffee plantations added further pressures to the forest, particularly during the 19th century (Frickmann, 2003). The expansion of cattle pasture, gold mining and hydroelectric projects is also recognized as an important immediate cause of forest loss, with the former continuing to be an important driver of deforestation (Dean, 1997; Metzger, 2009). More recently, the expansion of urban areas and exotic tree plantations are replacing the remaining forest patches (Metzger, 2009).

Atlantic Forest deforestation in Paraguay began after the 1940s when the establishment of settlements, expansion of the agricultural frontier and the introduction of African grasses for pasture occurred, driving deforestation rates to about 2000 km^2 per year in the 1980s and continuing at 1000 km^2 per year through the 1990s (Cartes, 2003; Catterson and Fragano, 2004). Approximately 25% of the original Paraguayan Atlantic Forest remains (Huang et al., 2007), and soybean plantations are an important factor of recent forest loss, particularly since the 1990s (Richards, 2011). In northern Argentina, the Atlantic Forest is located in the Misiones province and represents the largest remnant of continuous forest (Izquierdo et al., 2008). Covering a former area of about 29,800 km^2 (Chebez and Hilgert, 2003), the Argentinean Atlantic Forest has been affected by soybean plantations, cotton, sugar cane, coffee, and more recently exotic tree plantations represented by *Eucalyptus spp*. and *Pinus spp*. (Plací and Di Bitetti, 2006). Currently, approximately 34% of the original forest extent remains (ca. 10,000 km^2) (Chebez and Hilgert, 2003). Between 1973 and 2006, most of the land cover change (2702 km^2) was characterized by an expansion of exotic tree plantations (Izquierdo et al., 2008), representing, along with human population growth, one of the central threats to the future of the Atlantic Forest in Argentina (Izquierdo et al., 2008; 2011).

Exotic tree plantation expansion could explain the 33,056 km^2 of new forests observed by Hansen et al. (2013) for the period 2000 to 2012, the largest absolute increase in forest cover in non-Amazonian South America (Table 3.2). Considering the decrease of 44,658 km^2 of all forest cover for the same period, the Atlantic Forest of Argentina, Brazil and Paraguay could be considered as the ecosystem's most affected by intense forestry practices in South America.

Impacts on climate

Despite the extent of change in the Atlantic Forest, studies addressing related climatic impacts are extremely rare. The only study found was conducted by Webb et al. (2005), who analysed weather station data to evaluate the potential effect of rainforest clearance on rainfall in the State of São Paulo in Brazil. Although no strong relationships were observed between forest cover and total rainfall, tree cover was significantly correlated with the number of rainy days and with interannual rainfall variability, with more fragmented forests associated to fewer rain days. Webb et al. (2005) argue that large scale factors independent of vegetation cover, such as coastal weather fronts, control the total amount of rainfall in the study area. However, local geographical features (e.g., topography) together with tree cover explain the number of days over which rain falls.

3.3.2.4 Temperate Grasslands

LUCC

Since the arrival of Europeans in the early 16th century (Báez, 1944), large areas of grasslands in Argentina and Uruguay have been converted into crops and pastures. In the last decades, technological improvements, global food, timber and energy demand and climate changes have intensified LUCC in the remaining native grasslands, which have been converted to annual crops such as soybean, maize, sunflowers, wheat and oats at increasing rates in response to demand from Asia (Zak et al., 2008) and more recently to tree plantations (Nosetto et al., 2012). This trend is partially explained by an increase in rainfall with subsequent replacement of natural grassland located in more humid areas (Pérez and Sierra, 2012).

Conversion of Temperate Grasslands to fast growing *Pinus* and *Eucalyptus* plantations in Argentina and Uruguay has increased rapidly during recent decades, expanding from 23,000 ha in 1992 to 125,000 ha in 2001 (Nosetto et al., 2012; Paruelo et al., 2006; Silveira and Alonso, 2009). Between 2000 and 2012, 13,859 km^2 of new forested area was added to the Temperate Grasslands (Hansen et al., 2013).

Impacts on climate

Modeling studies. LUCC in the Temperate Grasslands involves conversion of natural grasslands (C3 and C4 photosynthetic pathways) to croplands and exotic tree plantations (Baldi et al., 2008a; Baldi et al., 2008b). Climatic consequences of these changes have been mainly addressed through regional climate models. In central Argentina, the nation's most important agricultural region, the cooling trend observed by Rusticucci and Barrucand (2004) has been linked to albedo changes as a result of the conversions of natural grasslands by croplands (Beltrán-Przekurat et al., 2012b). However, the temperature response depends on whether C3 or C4 grasslands are converted. A cooling effect results from converting C3 grasslands and a warming from converting C4 grasslands, arguably explained by differences in evapotranspiration rates. In addition, changes in precipitation were related with those areas where land cover change occurred, particularly during dry years (Beltrán-Przekurat et al., 2012b). The climatic response of LUCC in the Temperate Grasslands seems to be sensitive to how vegetation is described in the land surface component of the climate model used. Though this makes modelling comparison more difficult, it gives insights of land surface feedbacks under different LUCC scenarios in terms of changes in the water and the energy budget. For instance, Lee & Berbery (2011) modelled increases in the near-surface temperature when grasslands were replaced by dry croplands in lower La Plata Basin, with the effects extending beyond the areas where the changes occurred. These changes were associated with alterations in heat fluxes after slight reductions of roughness length and low level wind acceleration that determine net positive effects over precipitation. As the authors discuss, their results are not directly comparable with those from Beltrán-Przekurat et al. (2012) because of differences in the vegetation cover inputs of the regional climate models which affect the resulting biophysical processes.

Observational studies. Observations of albedo changes in the Temperate Grasslands using remote sensing techniques have reported albedo increases up to 16% between years 2000 and 2008, with agricultural expansion and reduced surface water recognized as the main drivers of these increments (Loarie et al., 2011a). The link between changes in albedo and observed net cooling for central Argentina (Nuñez et al., 2008; Rusticucci and Barrucand, 2004) has been recently proposed in the literature (e.g. Beltrán-Przekurat et al., 2012). However, further research is necessary to relate changes in albedo induced by LUCC and observed temperature trends in the region, especially to separate out the effects of global warming and LUCC.

3.3.2.5 Chilean Matorral

LUCC

Patterns and rates of land cover change in the Chilean Matorral are similar to those in the other areas of non-Amazonian South America. Before European settlement, human populations were restricted to coastal areas and river basins with limited impacts on natural vegetation (represented by localized fire) (Armesto et al., 2010). After the arrival of European colonizers, but intensified after the country's independence, extensive loss of forests occurred due to a massive demand of timber extraction for mining, agriculture and cattle grazing (Armesto et al., 2010). From the 1970s, government subsidies for agriculture and exotic forest plantations were responsible for the loss of 42% of native forests between 1975 and 2008 (Niklitschek, 2007; Schulz et al., 2010). These subsidies particularly impacted the sclerophyllous and temperate forests of central and southern Chile (Echeverria et al., 2006; 2008). In the same period, a large proportion of forests were converted into a savanna dominated by the invasive species *Acacia caven*, which is now the most common land cover type in Central Chile (Schulz et al., 2010; Van de Wouw et al., 2011). Recently, the Chilean Matorral, as in the Atlantic Forest and the Temperate Grasslands, has shown a high dynamic forest cover. Between 2000 and 2012, 2127 km^2 and 3065 km^2 of forest were loss and gain, respectively.

Impacts on climate

Modeling studies. LUCC climate interaction studies in the Chilean Matorral are almost absent. Most of these derive from modeling approaches. In Central Chile, the work of Beltrán-Przekurat et al. (2012b) found a warmer and drier climate after the conversion of wooded grassland to croplands (wheat), with increased Bowen ratio in spring. However, former vegetation used by Beltrán-Przekurat et al. (2012b) does not agree with vegetation classifications in the Chilean Matorral, previously composed by sclerophyllous forests and shrublands with different physiological features and vegetative characteristics (e.g., leaf area index). Most of this vegetation change has been for irrigated crops (Schulz et al., 2010) and only a few available studies report the atmospheric effects resulting from this change. For instance, in the northern Chilean Matorral Montecinos et al. (2008) used a mesoscale climate model to evaluate the impacts of irrigated agriculture on the local meteorological variables in the semiarid Elqui valley. In this area, the increased soil moisture along the valley's floor due to irrigation facilitates the transport of moist air through advection into the surrounding areas. Moreover, evapotranspirative changes in specific humidity and temperature on the valley's floor increase the relative humidity which in turn can induce fog formation both in early morning and late afternoon. These impacts are not restricted to the irrigated areas alone, but also influence the energy balance components in the surrounding hillsides by modifying thermally-induced winds (Montecinos et al., 2008). To date, there is no evidence of the potential climatic impacts of LUCC in other areas of the Chilean Matorral.

Observational studies. The modeling study of Montecinos et al. (2008) agrees with observations from eddy covariance stations in the same valley, where the Bowen ratio of irrigated fields is more than ten times lower than the surrounding semiarid natural vegetation. This is related to the strong differences in the radiation and energy balance between the two land cover types as well as in the increased evapotranspiration caused by irrigation (Kalthoff et al., 2006).

3.3.2.6 Tropical Dry Forests

LUCC

Tropical Dry Forests are considered as one of the most threatened ecosystems in the Neotropics (Pennington et al., 2006). Originally comprising a large and contiguous forest from Mexico to Bolivia, current Tropical Dry Forests in South America are distributed in small patches covering approximately 34% of their former extent (Sánchez-Azofeifa and Portillo-Quintero, 2011). This vegetation type is associated with fertile soils and therefore is one of the most impacted by crop and livestock production (Pennington et al., 2000). Today, cattle ranching, cropping, timber plantations and fuel-wood extraction are important drivers of forest loss (Miles et al., 2006). In Venezuela, for example, only 15% of the original Tropical Dry Forests remains after cattle ranching and agriculture development, with urbanization and fire that are also representing important factors of deforestation (Fajardo et al., 2005). In Colombia, Tropical Dry Forests are one of the most historically impacted ecosystems (Etter et al., 2008). Currently, <1.5% of the original Colombian Tropical Dry Forests remains, although some degree of recovery has been recently observed (Sánchez-Cuervo et al., 2012). In Bolivia, the Chiquitano dry forests, the largest extant areas of dry forests in South America, are considered as one of the most endangered ecoregions in the Neotropics (Dinerstein et al., 1995) with deforestation rates reaching 80,000 ha per year near the city of Santa Cruz as a result of agriculture expansion, highway construction, gas pipelines and mining (Killeen et al., 1998). Deforestation rates between 3 and 5% per year have been reported for the Chiquitano area of Bolivia (Mertens et al., 2004; Steininger et al., 2001). In the Brazilian Caatinga, 30–52% of dry forests have been altered by human activities, ranking third as the most degraded and destroyed ecosystem in Brazil after the Atlantic Forest and the Cerrado (Leal et al., 2005). In southwestern Ecuador, b1% of the formerly Tropical Dry Forests area (about 28,000 km^2) remains in an undisturbed state (Dodson and Gentry, 1991). Similarly, it is estimated that 95% of the original Tropical Dry Forests in Peru has been converted to human land uses such as those mentioned before (Sánchez-Azofeifa and Portillo-Quintero, 2011).

Impacts on climate

Modeling studies. Despite the extensive conversion of Dry Tropical Forests, evidences of LUCC influences on surface climate are restricted to the Chiquitano dry forest in Bolivia and the Caatinga in north-eastern Brazil, all of them presented through modeling studies. In the Caatinga of northeastern Brazil, early studies using climate models have shown changes in the land surface moisture budget and albedo impacts on near atmosphere, particularly regarding to adiabatic heating and precipitation (Sud and Fennessy, 1982). More recently, some studies have found potential climatic influences of desertification in the Caatinga. According to Oyama and Nobre (2004), desertification (change from xerophytic vegetation to bare soil) may weaken the hydrological cycle, with a strong decrease in precipitation, evapotranspiration, atmospheric moisture convergence and runoff, and an additional increase in surface temperature. In agreement with Oyama and Nobre (2004), Hirota et al. (2011) modelled negative precipitation anomalies as a result of Caatinga desertification, affecting also neighbouring northwestern Amazon. Recently, Castilho de Souza and Oyama (2011), using a regional climate atmospheric model, assessed progressive desertification influences on climate in the Caatinga, with similar results as Oyama and Nobre (2004). It has been stated that the expansion of desert areas could feedback upon their selves through radiative and heat alterations (Adams, 2007) such as those reported for the Caatinga. In the Chiquitano dry forest near to the city of Santa Cruz, Bounoua et al. (2004) applied a land surface model to evaluate the sensitivity of local climate to recent vegetation change using the Simple Biosphere Model (SiB2) (Sellers et al., 1996). They found an increase of 0.6 °C in surface temperature when broadleaf dry forest was converted to cropland. This warming was associated to morphological changes such as a decreased surface roughness, increased aerodynamic resistance and decreased stomatal conductance (Table 1 in Supplementary Information). These variations reduced latent heat flux and increased canopy sensible heat flux, and consequently temperature. Similarly, conversions from wooded grasslands to croplands in the Chiquitano produced an increase in mean temperature of 1.5 °C. This warming was driven by physiological changes when C4 wooded grasslands were replaced by C3 croplands, reducing canopy conductance by approximately 50% (Bounoua et al., 2004).

3.4 Discussion

3.4.1 Amazon bias

The total loss of natural vegetation in non-Amazonian South America is estimated in 3.6 million km^2. This area is 4 times greater than the historic Amazon deforestation and equivalent to 37% of U.S. land mass or 3 times the total surface area of Germany, France and United Kingdom. The region has experienced consistent LUCC pressures since European colonization, which are expected to increase in the coming years due to advances in technology, access to former remote areas and an increasing global demand of food commodities and biofuels (Liverman and Vilas, 2006).

It is important to note that present vegetation areas estimated here are not accurate because of assumptions made in calculating areas for potential historic natural vegetation that, for example, did not consider vegetation heterogeneity (e.g. grasslands distributed in the Dry Chaco) and imprecisions of satellite images used to obtain the area of the vegetation types. For the last, errors depend, among other factors, on the training data used and/or the ability of the algorithm to differentiate between two classes with similar spectral signatures. For example, the image from Friedl et al. (2010) used to calculate the current evergreen broadleaf forest extent in the Amazon and Atlantic Forest has a producer and user accuracy of 93% and 83%, respectively, which is high compared to the accuracy of all categories in the MODIS product (75%). However, for grasslands (used to calculate vegetation in Temperate Grasslands) the classification from Friedl et al. (2010)

shows a moderate producer accuracy (74%) and low user accuracy (60%), which reflects the difficulties to distinguish this class from others, particularly open shrublands and croplands (Friedl et al., 2010). Notwithstanding, Friedl et al. (2010) was the only available dataset that allowed us to calculate present vegetation cover for the Amazon, Atlantic Forest and Temperate Grasslands. In the case of the Dry Chaco, Clark et al. (2010a) developed a robust methodology to classify forest (woody vegetation) at 250 m resolution using training datasets taken from high-resolution Quick Bird imagery in Google Earth. Their producer and user accuracy was 96% and 85%, respectively, which is higher than that obtained by Friedl et al. (2010) for deciduous broadleaf forest (69% and 76%) described for the Dry Chaco at 500 m resolution. High accuracy was obtained for the Caatinga and Cerrado with overall classification accuracy between 92% and 97%, respectively (MMA/IBAMA, 2011a; 2011b).

Despite the large magnitude of the natural vegetation loss showed by the various remote sensing approaches for South America, the number of publications addressing later modifications of surface atmospheric feedbacks is relatively very low. Compared to the Amazon region, non-Amazonian South America registers far less peer-reviewed publications in the field of surface atmospheric processes and feedbacks. Although there is some observational and model-based evidence of the LUCC effects on surface temperature and precipitation, changes in atmospheric circulation and links with climate extremes are barely known. In terms of the impacts of LUCC on climate, non-Amazonian South America remains as one of the least studied regions worldwide.

3.4.2 Patterns and processes of change

By changing the land cover we are modifying land surface attributes that are important in the exchange of heat and momentum between earth's surface and the atmosphere. These alterations ultimately modify moisture and energy budgets and with them surface temperature and precipitation (Bonan, 2008b; Mahmood et al., 2014; Pielke et al., 2011). In non-Amazonian South America, evidence suggests that LUCC, expressed as replacement of native forests, savannas and grasslands by crops and pastures, is associated to changes in surface temperature and precipitation (Figure 3.4. 4 summarizes the main impacts of LUCC on climate in non-Amazonian South America). These responses are mainly driven by a decline in evapotranspirative cooling due to differences in morphological attributes that influence evapotranspiration and land–atmosphere coupling, including leaf area index, roughness length and rooting depth between natural and non-natural vegetation, particularly in the Dry Chaco and the Cerrado. Other modeling and observational studies have reported similar patterns of temperature and rainfall response after land cover change in dry forests and savanna type biomes. In Australia, replacement of woody vegetation by agriculture was related to increments in surface temperature between 0.4 °C to 2 °C and lower summer rainfall (McAlpine et al., 2007). As in the Dry Chaco dry forests, the replacement of dry forest to crops and grasslands in southwest Western Australia also explains 50% of the observed warming (Pitman et al., 2004). Similar to the Cerrado (Figure 3.4-b), in the savannas of Australia, southern Africa and northern South America, Hoffmann and Jackson (2000) modelled increases in surface temperature of 0.5 °C and 10% decreases in rainfall when natural vegetation was converted to grasslands, primarily because of reductions in surface roughness length. This pattern was also described by Snyder et al. (2004), who verified that the removal of the world's savannas caused large reductions in precipitation and surface warming due to reduced latent heat.

In non-Amazonian South America, significant albedo variations associated to LUCC were reported (e.g., Loarie et al., 2011a). These variations are of particular interest in semiarid environments, where albedo enhancement has been linked to precipitation suppression via subsidence anomalies (Otterman, 1989). In the Caatinga of northeastern Brazil, desertification increases albedo and therefore diminishes moisture convergence and precipitation, creating a positive feedback that limits moisture recycling in desertified areas (Oyama and Nobre, 2004; Sud and Fennessy, 1982; 1984) (Figure 3.4-e). This mechanism has been described in the early study of Charney (1975) for the Sahel region, where strong increments in albedo after degradation of vegetation produced a sinking motion and an additional drying that would perpetuate the arid conditions. Conversely, insertion of irrigated agriculture in semiarid environments would increase moisture convergence that can enhance cloudiness and precipitation. In the arid extremity of the Chilean Matorral, Montecinos et al. (2008) report lower albedo in irrigated cultivated areas compared to the surrounding semiarid vegetation (Figure 3.4-d). This increased the net radiation over irrigated areas and consequently the available energy to be transferred into the lower atmosphere through latent heat flux. This process leads to net cooling by up to 2 °C for the irrigated valley compared to semiarid vegetation. Similar examples are found in semiarid environments of India and North America, where the effect of irrigation increases soil moisture levels and consequently latent heat flux, cooling the boundary layer over irrigated areas and increasing atmospheric moisture and cloudiness (Kharol et al., 2013; Kueppers et al., 2007; Roy et al., 2007). In California's Central valley, a region with many biogeographical and climatological similarities to the Chilean Matorral (Zedler et al., 1995), increasing evapotranspiration after irrigation significantly impacts the atmospheric circulation and strengthens the hydrological cycle over southwestern United States (Lo and Famiglietti, 2013). Changes in surface climate after introduction of irrigated agriculture have also been reported for the great plains of North America, where observed maximum daytime temperatures over wheat fields are 2.3 °C cooler than the surrounding grasslands in the growing season but 1.61 °C warmer after its harvest (Ge, 2010). These alterations do not only depend on the direction of land cover change but also on crop phenology, atmospheric moisture content and synoptic scale atmospheric circulation (Mahmood et al., 2014).

In the case of the Atlantic Forest, despite showing very similar vegetation features to the Amazon rainforests (D'Almeida et al., 2007), the only observational study found in this review did not show a strong relationship between the amount of total rainfall and deforestation. However, the strong link between deforestation and the amount of rain days (Webb et al., 2005) (Figure 3.4-f) suggests that hydrological impacts of deforestation need to be addressed by future research. Since the original distribution of the Atlantic Forest covered an area that presently sustains the highest population density in South America and one of the most populated cities in the world (São Paulo), further studies are required to investigate the influence of historic deforestation on patterns of precipitation and temperature, and the potential role of remaining forest patches in the hydrological cycle and water availability in the region, particularly during climate extremes like droughts.

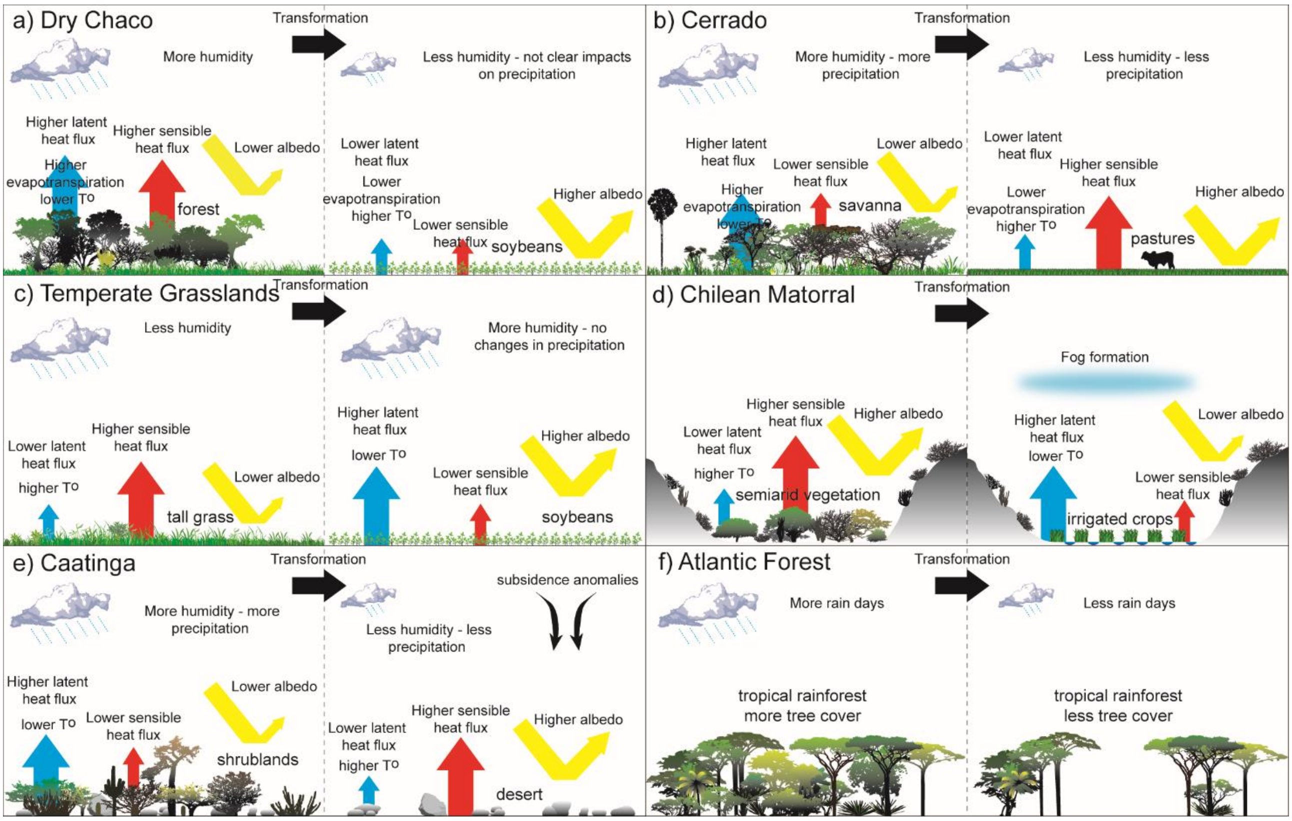

Figure 3.4 Summary of the main climatic impacts of LUCC in non-Amazonian South America based on the literature reviewed. Impacts are represented by variations in heat fluxes, temperature, precipitation and albedo, following transformation of natural vegetation to other land uses. The studies of Beltrán-Przekurat et al. (2012b) and Houspanossian et al. (2013) were used for the Dry

Chaco, Loarie et al. (2011b) and Costa and Pires (2010) for the Cerrado, Beltrán-Przekurat et al. (2012) for the Temperate Grasslands, Montecinos et al. (2008) for the Chilean Matorral, Castilho de Souza and Oyama (2011) for semiarid Caatinga, and Webb et al. (2005) for the Atlantic Forest. In the Chilean Matorral, changes in precipitation are not shown because they were not evaluated by Montecinos et al. (2008). Similarly, heat fluxes and albedo are not displayed for the Atlantic Forest because Webb et al. (2005) evaluated changes in precipitation only. The trends in climatic responses shown here (heat exchange, temperature, precipitation and albedo) vary greatly between studies and model experiments, and therefore must to be taken carefully because they not necessarily represent unidirectional climatic changes after LUCC.

How local changes can affect the climate of nearby regions and influence neighbouring and remote areas through teleconnections are still under discussion in the literature (e.g., Pielke et al., 2011). Most of the studies evaluated here assess only the local climate impacts associated with LUCC processes. However, the impacts should be accounted in a concurrent manner as they occur in reality. This is important because results could differ markedly, and synergistic non-linear processes could occur in order to alter the climatic effects of the combination among LUCC processes in a regional perspective (Costa and Pires, 2010; Hirota et al., 2011).

3.4.3 Modelling vs observational studies

Most of LUCC impacts on climate in non-Amazonian South America have been addressed through modeling approaches including land surface (2 studies) and climate models (13 studies), while few studies were based on observations (4 studies) (Figure 3.5). Land surface models can overestimate the impact of deforestation because they do not take into account land surface atmospheric feedbacks (Pielke et al., 2011) and hence climate models are more robust as they incorporate land surface atmosphere interactions and feedbacks. However, since most of modeling experiments use just one climate or surface model it is very hard to identify whether the results are model dependent or are representative of the climate responses to land cover change. In addition, models differ in their settings and description of land surface processes and atmospheric physics. An example is the difference in the characterization of the land surface properties. Climate models use land cover classifications derived from satellite images with varying classification systems that sometimes are not comparable with others. For instance, in the Temperate Grasslands Beltrán-Przekurat et al. (2012) used a classification that differentiate between C3 and C4 grasses, while Lee and Berbery (2011) applied a climate model with a land cover classification that does not distinguish between photosynthetic pathways, which makes it difficult to compare their results and could explain the difference in the sign of change between the two climate models applied for the same region.

Another aspect that dampens the reliability of LUCC experiments is the testing for statistical difference. Most of the studies use the Student's t-test that does not take into account autocorrelation issues that affect the independence of climate observations. This problem affects model results because the t-test can overestimate the climate impact of LUCC through Type-I errors or false positives (Zwiers and von Storch, 1995). For non-Amazonian South America, only 3 of 19 studies considered autocorrelation in the statistical analysis and therefore it can be argued that most of the climatic impacts shown could have been overestimated.

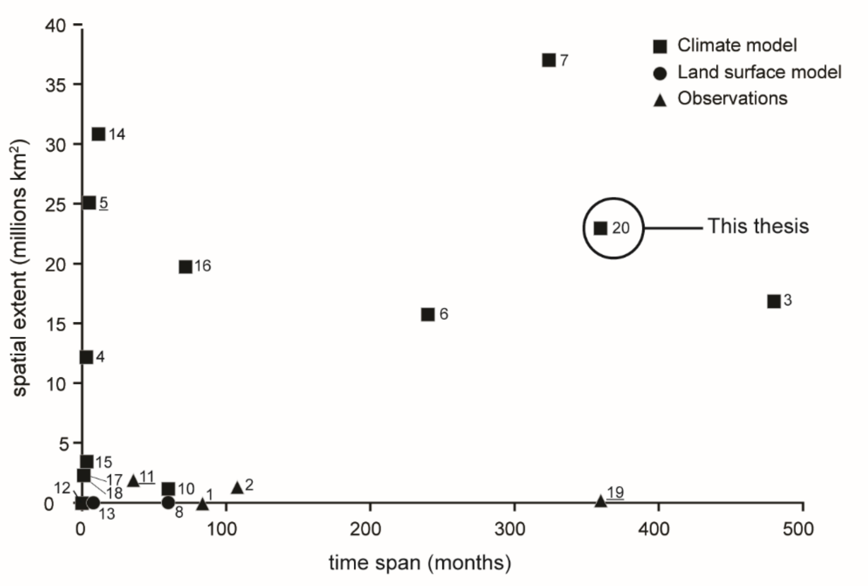

Figure 3.5 Spatial and temporal extent of the studies reviewed. Studies approached are categorized as climate models, land surface models and observational studies. Squares represent climate models, circles refer to land surface models and triangles to observations using either satellite images or weather station data. The numbers relate to those shown in the reference column of **Table 3.3** (Mendes et al., 2010 not shown). Underlined numbers correspond to those studies that included autocorrelation in statistical testing. The work described in this thesis is also shown.

3.4.4 Risks and consequences

According to observations, South America shows different trends in precipitation and temperature depending on the geographic area under analysis. Skansi et al. (2013) describe a wetting trend in many areas of the continent since the mid-20th onwards, mostly in southeastern South America, northern Peru and Ecuador. On the other hand, negative tendencies in evapotranspiration and soil moisture have also been observed between 1982 and 2008 over much of South America (Jung et al., 2010). In southern South America, and contrasting the global tendency, specific humidity has decreased in recent decades (IPCC, 2013b). Central-south Chile and Argentina have registered significant reductions in precipitation and increased surface temperatures (IPCC, 2014a). Increased warmer days, decreased cold days and more extreme rainfall events have also been observed in many parts of the continent (Alexander et al., 2006; Skansi et al., 2013). These climate extremes (length of drought and/or extreme precipitation events) are projected to increase in future climate change scenarios (Giorgi et al., 2011).

LUCC can exacerbate these regional changes in climate. The removal of native forests and savannas decreases evapotranspiration and moisture flux, enhancing the dryness produced by other drivers of climate change such as increased concentrations of anthropogenic greenhouse gases, with potential increases in extreme events such as droughts and floods. This relationship has been shown in Australia, where LUCC can significantly raise the decile-based drought duration index, particularly during El Niño years (Deo et al., 2009). LUCC can also enhance the negative effects of climate change through alterations in surface hydrology because natural vegetation controls the redistribution of runoff, water table levels and soil moisture by altering soil permeability (Asbjornsen et al., 2011; D'Odorico et al., 2010), which in turn affect water supply for cities, hydropower generation and agriculture (IPCC, 2014a). This is particularly relevant for semiarid environments of non-Amazonian South America, where drying trends have been observed (Masiokas et al., 2008; Quintana and Aceituno, 2012). Presently, about 200 million people live in these ecosystems (Verbist et al., 2010; WB, 2014), which are already experiencing water stress and reduced agricultural productivity (IPCC, 2014a). For example, in the Caatinga, current decreased precipitation and river discharge will intensify in the future with strong impacts on crop yields and water security that will force the migration of population from rural to urban areas (Krol and Bronstert, 2007). It is expected that LUCC will intensify these impacts (Montenegro and Ragab, 2010). In the Dry Chaco, increasing precipitation has stimulated the advent of large-scale rainfed crops into areas formerly covered by dry forests (Clark et al., 2010a; Gasparri and Grau, 2009). Since this land transformation has locally raised the dryness and surface temperatures (Houspanossian et al., 2013), it could potentially create a negative feedback that will revert the favourable climatic conditions, with significant socioeconomical consequences. In other semiarid environments of non-Amazonian South America such as the central Andes and Chilean Matorral, a weaker hydrological cycle is projected with an associated increased risk of lower water availability (Fiebig-Wittmaack et al., 2012; Vicuña et al., 2012). It is recognized that climate change and LUCC combined with water governance structures, institutional arrangements, societal values and development pathways are the major threats to water security in semiarid South America (Scott et al., 2013).

LUCC climate feedbacks can negatively affect both urban and rural areas. At present, 79% of the South American population live in cities (WB, 2014), many of which are subjected to risks associated with impacts of LUCC, climate change and their feedbacks: increasing flooding and landslides (Andrade and Scarpati, 2007; Marengo et al., 2013), intensification of the heat island effect (Nobre et al., 2011), urban expansion over areas with increased climate risks (e.g. flat terrain), increased food insecurity (IPCC, 2014a), increase in diseases (CEPAL, 2014) and less water availability (Le Quesne et al., 2009; Little et al., 2009; Winchester and Szalachman, 2009). Compared to urban populations, rural populations show higher poverty levels and therefore are more vulnerable to the adverse impacts of environmental change (IPCC, 2014b). However, increasing evidence shows that both urban and rural population are highly exposed to the negative consequences of climate extreme events and alterations in the moisture budget influenced by LUCC in developing countries (see IPCC, 2014a and references therein). Most of the natural disasters in South America between 1972 and 2011 had a hydroclimatic origin, 57% of which were associated with floods, droughts and extreme temperatures (CEPAL, 2014). The potential influences that LUCC could exert upon these events are still not clear.

LUCC in non-Amazonian South America has increased environmental stress and threatens ecosystem resilience. Because of the non-linear relationship between terrestrial ecosystems and climate, changes can exhibit threshold behaviour (Zehe and Sivapalan, 2009). These changes may result into an irreversible shift to a drier climate state, in which rainfall would be insufficient to allow for the recovery of ecosystems and their services (Brovkin et al., 1998; Folke et al., 2004; Scheffer et al., 2001).

3.4.5 Research priorities

Below we outline five key research priories arising out of the findings of the review.

1. Expand the focus towards understudied regional-scale processes and impacts. Despite that the high LUCC rates observed in non- Amazonian ecosystems, this paper highlights the need to extend research-oriented activities to quantify the magnitude, climatic consequences and implications of such changes. Examples of these are the Tropical Dry Forests of northern South America, the Cerrado, the Atlantic Forest and the Chilean Matorral. Additional research efforts are required to measure the spatial extent and rate of LUCC and the detection of resulting climatic impacts and risks in these ecosystems. This research effort also needs to be expanded to other regions not included in this review, where evidence of high land cover transformation rates has been reported. Examples of these are the Valdivian Forests, the Llanos Savannas and the Ecuadorian Páramo.

2. Linking land atmosphere interactions with climate extremes. There is a need to increase the focus towards the relative contribution of LUCC in regional climate change and its interaction with other forcings such as greenhouse gases. In addition, although climatic extremes are recognized as major threads in South America, there is insufficient evidence of how changes in land cover interact with these climatic phenomena. Though some research has related LUCC with dry/wet El Niño Southern Oscillation conditions (e.g. Beltrán- Przekurat et al., 2012), further research needs to be conducted in order to understand feedbacks and potential societal consequences.

3. Increasing the surface climate and hydrological observation platform. It has been recognized that one of the major problems in South America is the lack of long-term homogeneous and continuous climate and hydrological records (IPCC, 2014a). This makes very difficult to identify historical patterns and trends in local and regional mean climate and in extremes, and hence address hypothesis in relation to the impacts of LUCC over the hydrological cycle. A major investment of resources is required to increase the number and distribution of meteorological and gauge stations, and widen current networks through partnerships between governments, universities, research institutes and programs.

4. Improving land surface descriptions for regional climate models. Many of the land surface characterizations used in regional climate models can be improved through the incorporation of more accurate representations of land cover such as different crop varieties, irrigated agriculture, and descriptions of different biomes. In the case of non-Amazonian South America, land surface models embedded in climate models are usually calibrated in regions where the models were developed and do not accurately represent the conditions where such models are applied. Upgrading surface features to local/regional conditions (e.g. leaf area index, vegetation fraction, roughness length, and albedo) will make modeling results more robust.

5. Statistical testing for LUCC experiments. There are problems related to the use of discredited statistics to test for differences in LUCC experiments. For example, most of the studies for non-Amazonian South America reviewed used the classical Student's t-test for calculation of differences without considering autocorrelation. Problems associated with it relate to an over-estimation of LUCC impacts on climate, which make results less reliable. The modified

Student's t-test (Zwiers and von Storch, 1995) is one of the many available options to overcome this issue.

3.5 Conclusions

In this study, we reviewed the main patterns of land use and land cover change and subsequent climatic impacts for non-Amazonian South America. Our major findings can be summarized as follows:

- Non-Amazonian South America has been subjected to a consistent historic process of LUCC. Overall, 3.6 million km^2 (58% of the former area) of potential natural vegetation has been converted into anthropogenic land use practices, representing more than 4 times the area of Amazon deforestation, which has lost ~920,000 km^2 or 14% of its former area. The most affected ecosystems are the Chilean Matorral and the Atlantic Forest with 83% (52,000 km^2) and 81% (978,000 km^2) of their former natural vegetation transformed by year 1999 and 2012, respectively. LUCC also affected other ecosystems such as the Cerrado, Temperate Grasslands and Tropical Dry Forests where at least 52% of the original natural vegetation has been converted to anthropogenic land uses. The main drivers behind the conversion of natural vegetation are the expansion of croplands (soybean) and cattle pastures to meet global food demand, technological advances, climatic factors and governmental subsidies to increase production of food commodities.

- Based on the datasets of Hansen et al. (2013) for non-Amazonian South America, the Dry Chaco and the Atlantic Forest showed the highest relative amount of forest loss for the period 2000–2012, followed by the Cerrado and Tropical Dry Forests. While the Dry Chaco deforestation is related to native forest loss, in the other ecosystems, particularly in the Atlantic Forest, the Chilean Matorral and the Temperate Grasslands, forest loss is accompanied by a high proportion of forest gain, suggesting intensive forestry practices (mostly Eucalyptus and Pinus spp. plantations) as described by local studies.

- Climatic consequences of LUCC based on the studies reviewed are mainly related to an increase in surface temperature and a decrease in precipitation and cloudiness. Even though significant albedo variations are reported, the net change in temperature and precipitation after LUCC is mostly driven by shifts in latent and sensible heat fluxes. However, in semiarid areas albedo seems to play a significant role in reducing precipitation via subsidence anomalies. These impacts can manifest beyond the regions where land cover changes occur and could affect neighbouring regions such as the Amazon or even teleconnect beyond South America.

- More studies need to be conducted in order to estimate the magnitude of LUCC in non-Amazonian South America and its related climatic impacts, particularly in the most disturbed and understudied ecosystems. It is also necessary to understand the influence of LUCC on the duration an intensity of climate extremes such as droughts using climate model results supported by increased hydrological and climatic observations. LUCC experiments using such models should be parameterized according to local/regional surface characteristics and appropriate statistical tests need to be applied to make results more robust.

3.6 Acknowledgments

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CHAPTER 4 DEFORESTATION CHANGES LAND-ATMOSPHERE INTERACTIONS ACROSS SOUTH AMERICAN BIOMES

Abstract

In the previous Chapter I demonstrated that the natural ecosystems of non-Amazonian South America are one of the most affected by LUCC worldwide and with major uncertainties in terms of surface-climate interactions. In this Chapter, I address this knowledge gap by modelling vegetation-climate feedbacks in the region with a focus on those ecosystems identified in Chapter 3 as being the most impacted and least studied: the Atlantic Forest, the Cerrado, The Dry Chaco and the Chilean Matorral. I applied a 3 member ensemble regional climate model simulation for the period 1981-2010 (30 years) at 25 km resolution to quantify the changes in the regional climate resulting from historical land cover change. The results of computed modelling experiments show significant changes in surface fluxes, temperature, precipitation and moisture in all ecosystems. For instance, simulated temperature changes were stronger in the Cerrado and the Chilean Matorral with an increase of between 0.7 and 1.4 °C, respectively. Changes in the hydrological cycle revealed high regional variability. Yet, results showed consistent significant decreases in relative humidity and soil moisture, and increases in potential evapotranspiration across all sub-regions. These impacts are mostly strong during the dry season, which is more extreme and drier after natural vegetation clearing. This Chapter underlines the potential importance that natural vegetation can have on the biogeophysical components of the surface climate. In terms of surface temperature change, for instance, based on the results shown in this Chapter, vegetation change can be at least as important as other climate forcings such as the increase of greenhouse gases.

This chapter has been submitted as a journal paper to *Global and Planetary Change* as Salazar, A., Katzfey, J., Thatcher, M., Syktus, J., Wong, K. and McAlpine, C. Land use and land cover change impacts on the climate of non-Amazonian South America.

4.1 Introduction

By the year 2000, approximately 55 percent of the Earth's biomes had been converted into pastures, croplands, settlements and other land uses (Ellis et al., 2010). These changes have impacted biotic components of ecosystems such as biodiversity, and also modified land-atmosphere interactions through changes in the water and energy balance (Foley et al., 2003). Understanding these processes is important because it can enhance or dampen anthropogenic climate change and therefore increase vulnerability of ecosystems and people to climate variability.

It is well recognised that land use/cover change (LUCC) can affect climate through the absorption or emission of greenhouse gases (biogeochemical impacts) and by modifying the physical properties of land surface (biogeophysical effects). Changes in land use and land cover can lead to changes in surface fluxes of radiation, heat, moisture and momentum that can further impact the climate at local to regional scales (Pielke et al., 2002). In terms of radiation changes, LUCC can alter the surface albedo and thereby evapotranspiration processes and partitioning of sensible, latent and ground heat fluxes that can influence near-surface temperature and precipitation (Pielke et al., 2007). In addition, changes in land use/cover can transform vegetative attributes such as roughness, which influences the mixing of air in the boundary layer and surface temperature (Foley et al., 2003). Cumulatively, these modifications of land surface characteristics can impact the climate at a range of spatial and temporal scales (Mahmood et al., 2014; Pielke et al., 2011).

Non-Amazonian South America is considered as one of the least studied regions worldwide in terms of LUCC impacts on the surface climate (Salazar et al., 2015). During the last 500 years, LUCC has resulted in a massive transformation of its biomes, with more than 1 million km^2 (52%) of the Brazilian Cerrado savanna converted to crops and pastures (MMA/IBAMA, 2011b) and ~980.000 km^2 (81%) of the Atlantic moist forests transformed into crops, pasture and urbanization (Ribeiro et al., 2009; Salazar et al., 2015). High deforestation rates have also been reported for the forests of the Chilean Matorral and the Dry Chaco of Paraguay, Argentina and southern Brazil, with the later registering the highest global rate of tropical deforestation (Hansen et al., 2013). Increasing evidence from observational and modelling studies shows that this transformation of natural native forests significantly affects the flux of moisture, heat and momentum with subsequent impacts on surface temperature and precipitation (e.g. Beltrán-Przekurat et al., 2012a; Loarie et al., 2011b; Pongratz et al., 2006). However, high levels of uncertainty remain in relation to the mechanisms and consequences of the influence of deforestation on the climate of non-Amazonian South America.

In this paper we addressed the question “what are the potential impacts of historic deforestation on the mean climate of non-Amazonian South America?”. We present results of simulations of a three model ensemble for the period 1981-2010 (30 years) to quantify the changes in the regional climate produced by deforestation of different biomes distributed in four key regions: the Atlantic Forest (tropical and subtropical moist broadleaf forest), Cerrado (tropical and subtropical savanna), Dry Chaco (tropical and subtropical dry broadleaf forest) and the Chilean Matorral (Mediterranean forest), which are considered as the regions most affected by LUCC and where subsequent climate impacts are not well understood.

4.2 Methods

4.2.1 Climate model: CCAM

In this study, we used the Conformal-Cubic Atmospheric Model (CCAM) developed by CSIRO (McGregor, 1996; 2003; 2005a; 2005b; McGregor and Dix, 2008) to model the influence of historic deforestation on the climate of southern South America. CCAM is an hydrostatic full atmospheric Global Circulation Model formulated on a quasi-uniform grid derived by projecting the panels of a cube onto the surface of the Earth (McGregor and Dix, 2008). It can also be employed in a stretched mode by utilising the Schmidt (1977) transformation. In this study, the model was run in stretched mode with about 25 km horizontal resolution from latitude 10-45 S and longitude 30-90 W. This allows for dynamical downscaling where the computational grid is denser over the region of interest, but sparser elsewhere. CCAM contains a comprehensive set of physical parameterizations. It employs the diurnal varying Geophysical Fluid Dynamics Laboratory (GFDL) parameterization for long wave and short wave radiation (Freidenreich and Ramaswamy, 1999; Schwarzkopf and Fels, 1991), with interactive cloud distribution derived in combination with the liquid and water-ice scheme of Rotstayn and Lohmann (2002). In addition, it employs a stability-dependent boundary layer scheme according to Monin-Obukhov similarity theory (McGregor, 1993), the mass-flux cumulus convection (McGregor (2003) and the gravity-wave drag scheme over mountainous terrain (Chouinard et al., 1986) to reduce orographically-related systematic errors.

4.2.2 Land surface model: CABLE

The coupling between the land surface and the atmosphere is an important component of climate variability. In CCAM, feedbacks between the Earth’s surface and the climate are described through the Community Atmosphere Biosphere Land Exchange version 2.0 (CABLE) land surface model.

CABLE simulates the exchange of CO_2, radiation, heat, water and momentum fluxes between the land surface and atmosphere, and is composed of five main sub-models: (1) the radiation sub-model estimates the radiation transfer and absorption by both sunlit and shaded leaves and by soil surface in the visible, near infrared and thermal radiation, and also the surface albedo for visible and near infrared radiation (Wang and Leuning, 1998); (2) the surface flux sub-model estimates the coupled transpiration, stomatal conductance, photosynthesis and partitioning of net available energy between latent and sensible heat of sunlit and shaded leaves (Wang and Leuning, 1998). Photosynthesis is calculated for both C3 and C4 plants; (3) the canopy micrometeorology sub-model describes the surface roughness length, zero plane displacement height, and aerodynamic resistance from the reference height to the air within the canopy or to the soil surface (Raupach, 1994); (4) the soil and snow sub-models compute the heat and water fluxes within each of the six soil layers and three snowpack layers, snow age, snow density and snow depth, and snow covered surface albedo. Soil moisture is calculated with the Richards' equation and the heat conduction equation is used to obtain soil temperature (Kowalczyk et al., 2006; Wang et al., 2011); and (5) the ecosystem carbon module, which estimates respiration of stem, root and soil organic carbon decomposition (Dickinson et al., 1998).

The fluxes of heat, water and momentum depend on the mean properties of the flow through the use of aerodynamic resistances. Surface temperature is calculated through the energy balance equation and may consist in a combination of surface elements such as vegetation, bare ground, snow and ice. In CABLE, the vegetation is placed above the ground and hence allows for full aerodynamic and radiative interaction between the vegetation and the ground. The total surface fluxes are therefore the sum of the fluxes from the soil to the canopy air space and from the canopy to the atmosphere. This vertical flux is calculated using the Monin-Obukov similarity theory (Kowalczyk et al., 2006; Raupach et al., 1997), where surface roughness is an important factor influencing the friction velocity. A complete description of CABLE including its development history, major features and physics is given by Kowalczyk et al. (2006).

4.2.3 Experimental design

4.2.3.1 Land surface datasets

In order to evaluate the historic impacts of deforestation on the climate of southern South America, we completed two sets of model simulations (3 ensembles each) for the period 1981-2010. The only difference between the simulations was the description of land surface datasets. The first scenario had present (CNTRL) land cover characteristics and the second simulation had natural (NAT) land surface characteristics. For the CNTRL scenario, we upgraded the default land cover map integrated in CABLE (Loveland et al., 2000) by that developed by Friedl et al. (2010) for 2005, also known as Collection 5 MODIS Global Land Cover product or MODIS MCD12Q1 product. This represents a more actualized and accurate description of the land cover in South America. In the NAT scenario, we recreated the original vegetation in four main regions representative of distinct biomes: 1) the Atlantic Forest (tropical and subtropical moist broadleaf forest), 2) Cerrado (tropical and subtropical savanna), 3) Dry Chaco (tropical and subtropical dry broadleaf forest) and 4) Chilean Matorral (mediterranean forest) according to literature (see Table 4.1) and Olson et al. (2001) biomes and ecoregions boundaries. These regions are considered the most affected by deforestation in non-Amazonian South America (Salazar et al., 2015) (Figure 4.2). From the CNTRL land cover characteristics, we identified the most complex vegetation types in the MODIS image that agreed with descriptions from literature of natural vegetation types for each one of the regions. We then extrapolated historic native vegetation by replacing current (e.g. crops) by natural (e.g. forest) vegetation types (Table 4.1). Leaf area index (LAI) for modern land cover was based on that developed by Beijing National University (BNU) for the period 2000-2009 (Yuan et al., 2011). Finally, we inferred the leaf area index of the original vegetation by interpolating the BNU leaf area index of remaining natural vegetation in the MODIS image using a nearest neighbour rule. Outside the study area, land cover and land surface characteristic were set for modern day conditions for all simulations (Figure 4.1). For each simulation, we calculated the seasonal averages of surface temperature, precipitation, heat fluxes, evaporation, moisture and wind speed and analysed the statistical difference using bootstrapping at 95% confidence level (*$p < 0.05$*) where:

$X = \{X_1, X_2, \ldots, X_{n_X}\}$ is the sample of a climate variable from the NAT experiment during period 1981-2010, and $Y = \{Y_1, Y_2, \ldots, Y_{n_Y}\}$ the sample of the same climate variable taken from the CNTRL experiment in the same period.

The null hypothesis (H_o) states that there is no significant difference between the means $\overline{X}$ and $\overline{Y}$ (i.e. deforestation has no significant impact on the selected climate variable).

The alternative hypothesis (H_1) states that there is a significant difference between the means $\overline{X}$ and $\overline{Y}$ (i.e. deforestation has a significant impact on the selected climate variable).

The observed *t*-statistic is:

$$t_{obs} = \frac{\bar{Y} - \bar{X}}{\sqrt{\frac{\sigma_X^2}{n_X} + \frac{\sigma_Y^2}{n_Y}}}$$

Where ($\bar{X}$, σ_X , n_X) and ($\bar{Y}$, σ_Y , n_Y) are the mean, standard deviation and sample size of the NAT and CNTRL samples, respectively.

The bootstrap statistic for each sample is computed as:

$$t^* = \frac{\bar{Y}^* - \bar{X}^*}{\sqrt{\frac{\sigma_{X^*}^2}{n_{X^*}} + \frac{\sigma_{Y^*}^2}{n_{Y^*}}}}$$

Where ($X^*, \sigma_{X^*}, n_{X^*}$) and ($Y^*, \sigma_{Y^*}, n_{Y^*}$) are the mean, standard deviation and sample size of randomly selected bootstrapped samples, respectively. The Achieved Significance Level (ASL) will be the proportion of samples where $|t^*| \geq |t_{obs}|$. The *p* value is calculated as $p = 1 - \mathrm{ASL}$. The null hypothesis is rejected if $p < 0.05$, indicating a significant change in the climate variable across the two scenarios for the 30-year period. The sample sizes for *X* and *Y* were both 90 (3 ensembles over 30 years), with $N = 500$ bootstrap samples conducted to test for statistical significance.

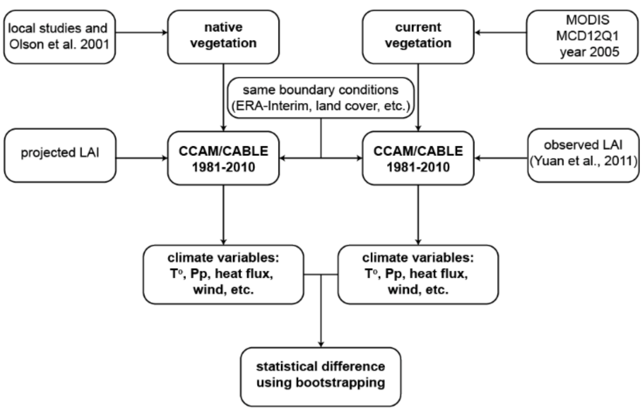

Figure 4.1 Conceptual model of methods used to identify the impacts of historic deforestation over the climate of non-Amazonian South America (for details see section 4.2.3.1).

4.2.3.2 Model experiments

For the modelling experiments, CCAM was driven by ERA-Interim reanalysis (Dee et al., 2011). This reanalysis was chosen in order to reduce the noise and maximize the local signal of deforestation. A recent evaluation of ERA-Interim reanalysis over the central Andes for winter temperature and precipitation showed a good correspondence with the gridded observations from Climate Research Unit (CRU) in terms of interannual variability (Rusticucci et al., 2014). Evaluations of ERA-Interim over different regions have resulted in good agreements with observed surface temperature (Mooney et al., 2011; Wang et al., 2014).

The dynamical downscaling was performed using the scale-selective filter proposed by Thatcher and McGregor (2009). The scale-selective downscaling approach allows consistency between the host model and model simulation at larger scales by replacing atmospheric fields at a spectral domain greater than a particular length scale (Radu et al., 2008; Thatcher and McGregor, 2009). This technique allows the model to freely develop regional-scale features consistent with the large-scale ones driven by the reanalysis and results are independent of the domain size (Thatcher and McGregor, 2009). In this study, CCAM used a scale cut-off configuration of about 24°, corrected every 6 hours above 850 hPa (about 1.5 km above the surface), allowing to assess the impact of land cover change on surface climate both near the surface and at scales less than the cut-off.

In this study, we focused on the climate effects on regions representing a variety of biomes considered the most historically impacted by deforestation (Salazar et al., 2015) and for which there are still little research in relation to LUCC impacts on the regional climate and how these vary spatially and seasonally. We did not focus on the effects of specific conversions between different land uses since these interactions have been addressed by other studies in South America (e.g. Beltrán-Przekurat et al., 2012a; Pongratz et al., 2006).

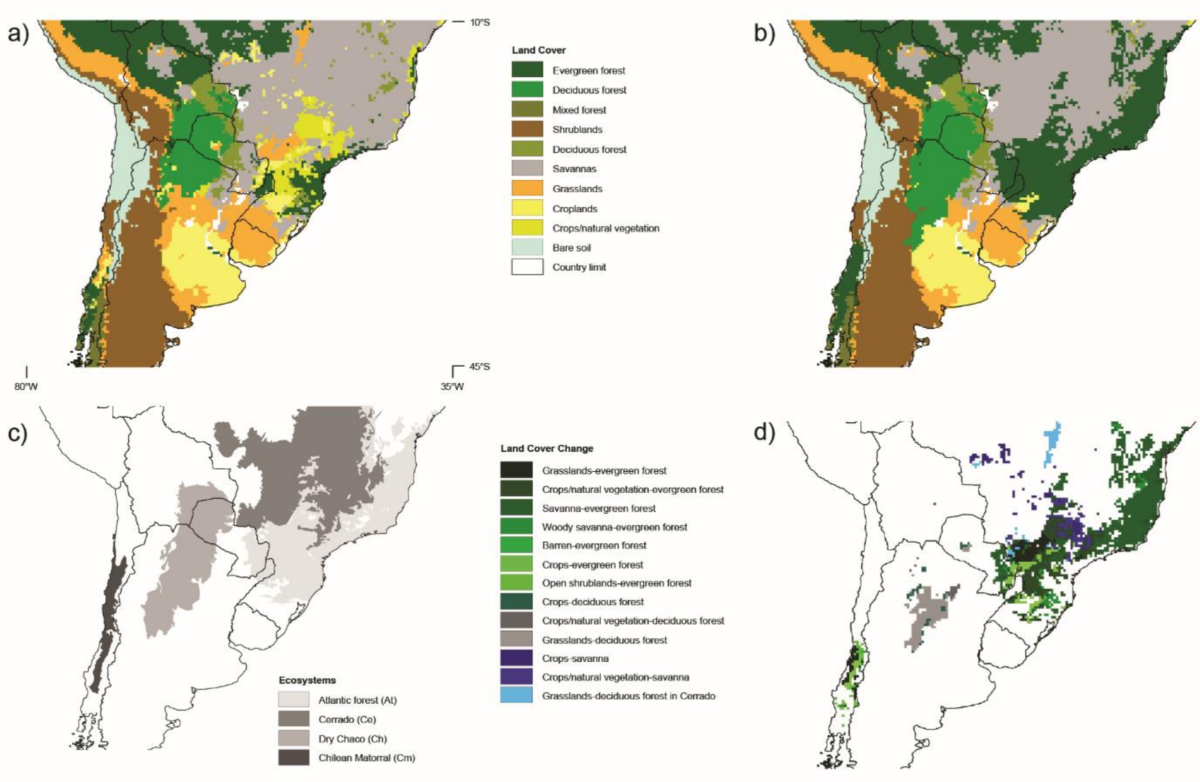

Figure 4.2 Land cover maps used in modelling experiments. a) Land cover map for CONTROL (CNTRL) experiment taken from MODIS MCD12Q1 for 2005; b) land cover map for NATURAL (NAT) experiment projected according to literature and the Olson et al. (2001) classification for non-Amazonian South America. Cerrado land cover is taken from MMA/IBAMA (2011b); c) boundaries of the regions considered in this study; and d) main changes (CNTRL-NAT) in vegetation between experiments. The green colour scale in map d) shows conversions of evergreen forest to other land uses, the grey scale shows conversion from deciduous forest, and the blue colour scale shows conversion of savanna vegetation and deciduous forest in the Cerrado.

Table 4.1 Main vegetation types for CNTRL and NAT model simulations, and equivalent MODIS classification scheme for four regions of non-Amazonian South America.

Region	CNTRL vegetation	Biome	NAT vegetation according to MODIS MCD12Q1
Atlantic Forest	Crops, grassland, savanna	Tropical and subtropical moist broadleaf forest (Câmara, 2003; Oliveira-Filho and Fontes, 2000)	Evergreen broadleaf forest
Cerrado	Crops, grassland	Tropical and subtropical savanna (Eiten, 1972)	Savanna
Dry Chaco	Crops, grassland	Tropical and subtropical broadleaf dry forest (Pennington et al., 2000; Prado, 1993b).	Deciduous forest
Chilean Matorral	Crops, grassland, open shrubland	Mediterranean forest (Luebert and Pliscoff, 2006).	Evergreen broadleaf forest

4.3 Results

In this section, we describe the changes in the land surface characteristics, heat fluxes, temperature, precipitation and moisture for present day conditions relative to the natural scenario. Mean changes correspond to the variables averaged across ensemble members for each experiment (CNTRL minus NAT). We present results according to dry and wet seasons. In eastern southern South America (Atlantic Forest, Cerrado and Dry Chaco), the dry season corresponds to austral winter (JJA) while in the Chilean Matorral it corresponds to austral summer (DJF). This pattern reverses in the wet season. Complementary variables are shown in Table B.1 of the Appendix B.

4.3.1 Change in surface characteristics

The changes in natural vegetation cover were related to changes in LAI, roughness length and albedo. All regions showed a decrease in the LAI for both the dry and wet seasons (Figure 4.3). The Atlantic Forest presented the greatest mean decrease with 1.54±1.01 and 1.28±0.99 in the dry and wet season, respectively (Table 4.2). Here, reductions in LAI were largest when crops and savannas replaced evergreen broadleaf forest (down to 4.5 in the dry season and 4.6 in the wet season (Figure 4.3)). The Chilean Matorral registered the second largest decrease in LAI in the dry season associated with the conversion of evergreen broadleaf forest to crops and grasslands used for pastures (1.04±0.58). Similarly, in the Cerrado, the main differences in LAI were in the dry season, particularly in those areas where savannas were replaced by crops (0.84±0.74, Table 4.2). The Dry Chaco represented the smallest change in mean LAI during the dry season with 0.04±0.07 and only few pixels showed a change (Figure 4.3).

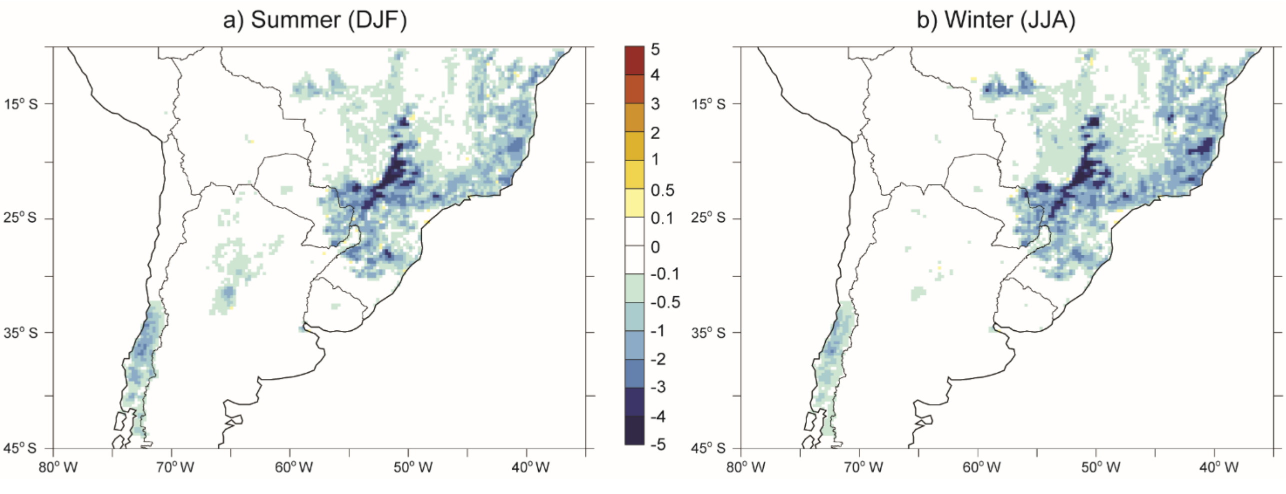

Figure 4.3 Seasonal changes (CNTRL-NAT) in Leaf Area Index (LAI, dimensionless) for non-Amazonian South America.

Table 4.2 Mean differences ± standard deviation (CNTRL – NAT) for seasonal leaf area index (LAI), roughness length (Z_o) and albedo (α, dimensionless) in four regions of non-Amazonian South America.

Region	Dry Season			Wet Season		
	LAI	Z_o (m)	α	LAI	Z_o (m)	α
Atlantic Forest	-1.54 ± 1.01	-2.00 ± 0.43	-0.0038 ± 0.01	-1.28 ± 0.99	-1.98 ± 0.40	0.0003 ± 0.01
Cerrado	-0.84 ± 0.74	-0.32 ± 0.47	-0.0073 ± 0.01	-0.60 ± 0.45	-0.30 ± 0.46	-0.003 ± 0.01
Dry Chaco	-0.04 ± 0.07	-1.85 ± 0.51	0.0014 ± 0.002	-0.24 ± 0.29	-1.29 ± 0.34	0.0012 ± 0.003
Chilean Matorral	-1.04 ± 0.58	-1.75 ± 0.72	-0.0078 ± 0.01	-0.64 ± 0.43	-2.09 ± 0.87	0.0143 ± 0.03

The change in LAI was related to a general decrease in roughness length. In CABLE the roughness length depends on the canopy height (35 m for evergreen broadleaf forest and 0.6 m for grass and crops) and LAI. Therefore, changes in vegetation cover were associated with changes in low level winds and heat fluxes (Figure 4.7-d and 4.4, $p < 0.05$). Again, the Atlantic Forest showed the greatest decrease in roughness length with an average of 2±0.43 m in the dry season and 1.98±0.4 m during the wet season. A large reduction in roughness length also occurred in the Chilean Matorral with 1.75±0.72 m in the dry season and 2.09±0.87 m in the wet season. Likewise, in the Dry Chaco the roughness length decreased -1.85 ± 0.51 and -1.29 ± 0.34 in the dry and wet season, respectively. The Cerrado registered the smallest mean reduction in roughness length in both the dry (-0.32 ± 0.47) and wet (0.30±0.46 m) seasons (Table 4.2). Changes in albedo between the experiments showed contrasting results. While in the Cerrado albedo respectively decreased between 7 and 15% in the dry and wet season, in the Dry Chaco it increased between 6 and 6%. Decreases between 1 and 8% were recorded for the Atlantic Forest and the Chilean Matorral in the dry season, while in the wet season albedo decreased between 1 and 16% in the same regions.

4.3.2 Change in heat fluxes

4.3.2.1 Dry season

CCAM shows a high contrast in heat fluxes response after deforestation. Because these are closely influenced by variations in the LAI, the geographical extent of changes was concurrent. However, orography and distance from the coast seem to be also important factors in the heat flux response that requires further research. Figure 4 shows the significant changes ($p < 0.05$) in sensible (H) and latent (LH) heat fluxes for non-Amazonian South America. During the dry season, the Cerrado and the Dry Chaco recorded a strong increase in H of about 12 and 15%, respectively. In the Cerrado, significant increments in H were found in western areas where crops and grasslands replaced savanna vegetation, while in the Dry Chaco the main changes occurred where herbaceous vegetation replaced deciduous forests (Figure 4.4-a, Table 4.3). The Atlantic Forest and the Chilean Matorral registered a decrease in H by about 1% in areas previously covered by evergreen forest (Figure 4.2-b, Figure 4.4-a).

Reductions in LH were more consistent than H across all regions, with changes ranging between 3 and 8%. In the Cerrado, changes in LH were conspicuous. These corresponded with reduced LAI and conversion of natural savannas to crops and grasslands (Figure 4.3). Likewise, significant decreases in LH were observed in the Chilean Matorral (5%) in areas near the Pacific coast (Figure 4.4). Less negative changes in LH were observed for the Atlantic Forest and the Dry Chaco with 3 and 5%, respectively. In the Atlantic Forest, the main reduction was located in the western areas, also concordant with the greatest decrease in LAI and in the second largest changes located in the central region of the Atlantic Forests (Figure 4.4-b).

These changes in the partitioning of sensible and latent heat are represented by changes in the Bowen ratio, which showed consistent increments across all regions. They indicate that, despite small reductions of H in the Atlantic Forest and the Chilean Matorral, declines in LH were greater in magnitude and hence, with current vegetation, more energy is used in heating the surface rather than in evapo-transpirative cooling. The last recorded the largest increase in the Bowen ration during the dry season with 0.35±0.54, followed by the Cerrado with 0.19± 0.24 and the Dry Chaco with 0.06±0.29. For the Atlantic Forest, the Bowen ratio increased by 0.06±0.31, mainly due to the reductions in LH.

Table 4.3 Mean differences ± standard deviation (CNTRL – NAT) for seasonal sensible (H) and latent (LH) heat fluxes, and proportional changes for Bowen ratio (β=H/LH, dimensionless) averaged for each region of non-Amazonian South America.

Region	Dry Season			Wet Season		
	H (w/m^2)	LH (w/m^2)	β	H (w/m^2)	LH (w/m^2)	β
Atlantic Forest	-0.18 ± 1.2	-1.45 ± 1.4	0.06± 0.31	-2.09 ± 1.5	0.56 ± 1.6	-0.04 ± 0.08
Cerrado	3.82 ± 1.3	-4.21 ± 1.7	0.19± 0.24	-0.83 ± 0.6	1.91 ± 1.4	-0.01 ± 0.02
Dry Chaco	1.48 ± 1.1	-1.13 ± 1.3	0.06± 0.29	-0.41 ± 2.4	1.30 ± 2.5	-0.01 ± 0.13
Chilean Matorral	-1.83 ± 2.3	-2.82 ± 2.1	0.35 ± 0.54	2.94 ± 0.7	-4.31 ± 0.8	-0.004 ± 0.11

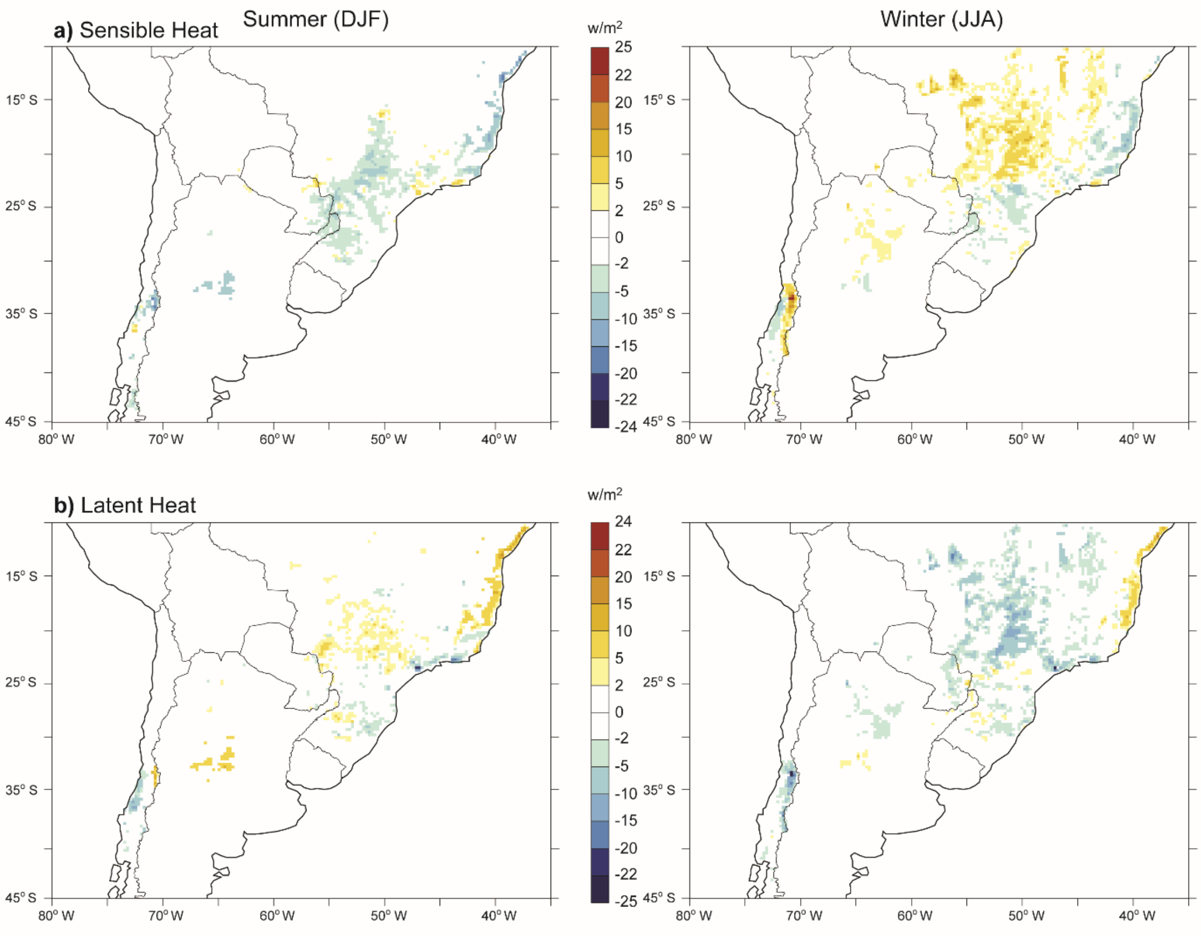

Figure 4.4 Differences in the seasonal means (CNTRL-NAT) for sensible (H) and latent heat flux (LH) across non-Amazonian South America. Differences are expressed as w/m2. In the Chilean Matorral, the summer and winter correspond to the dry and wet season, respectively. This pattern is opposite for the Atlantic Forest, Cerrado and Dry Chaco. Only pixels that are statistically significant at a 95% confidence level are shown.

4.3.2.2 Wet season

During the wet season, changes in heat fluxes generally showed the opposite trend to the dry season and were less in magnitude and more variable (expressed by the larger standard deviation). The only exception was the Chilean Matorral (in winter) which recorded a significant increase in sensible heat (37%) and the greatest seasonal mean reduction in latent heat (12%). In eastern South America, results showed a decrease in sensible heat (H), with the Atlantic Forest and the Cerrado showing the greatest reductions with 7 and 6%, respectively. These reductions corresponded with areas experiencing the greatest changes in LAI where herbaceous vegetation replaced evergreen forest and natural savannas. In the Atlantic Forest, a reduction in sensible heat was recorded near the coast, while in the Cerrado, changes located in the easternmost areas (Figure 4.4-a, Summer). The Dry Chaco registered the smallest decrease in sensible heat for the wet season with significant changes located in the southernmost areas of the region (Figure 4.4-a, Summer).

In eastern South America (Atlantic Forest, Cerrado and Dry Chaco), changes in latent heat flux (LH) showed the opposite pattern of H in areas where natural vegetation was converted to herbaceous vegetation (crops/grasslands). Yet the increments were less in magnitude compared to H, ranging between 1 and 2%. The Cerrado showed the greatest increment in LH, while in both the Atlantic Forest and the Dry Chaco, positive mean changes of +1% were recorded, particularly in coastal areas in the first and in the southernmost tip of the Dry Chaco (Figure 4.4-b, Summer). For these regions, the Bowen ratio tends to decrease during the wet season. These decreases indicate that more energy was used for evapo-transpirative cooling rather than heating the planetary boundary layer. The Atlantic Forest registered the greatest reduction trend in the Bowen ratio with 0.04±0.08, followed by the Cerrado and the Dry Chaco with 0.01±0.02 and 0.01±0.13 w/m^2, respectively. By contrast, during the wet season (winter), the Chilean Matorral showed a positive change in the Bowen ratio by about 0.004±0.11. Yet this change was highly variable (Table 4.3).

4.3.3 Change in surface temperature

4.3.3.1 Dry season

According to CCAM, deforestation resulted in significant differences ($p < 0.05$) in average surface temperatures in all regions (Figure 4.5). These changes were prominent in the dry season for both eastern and western southern South America. The Chilean Matorral showed the greatest mean increase in surface temperature in the dry season with 1.42±0.17 °C (Figure 4.5). A significant warming of up to 4 °C occurred in central areas, particularly where crops and grasslands replaced evergreen forest. The conversion of natural savannas in the Cerrado resulted in a warming increase of 0.68±0.17 °C that was also coincident with strong reductions in the LAI. The Atlantic Forest also showed significant increase in surface temperature corresponding to areas with the highest reductions in LAI. A mean temperature increase of 0.52±0.16 °C was observed when evergreen broadleaf forest was replaced with herbaceous (crops, grasslands and savannas) vegetation. There was a warming reaching up to 2 °C in the westernmost areas of the Atlantic Forest, and coincident with the greatest reductions in LAI (Figure 4.3 and Figure 4.5). On the contrary, the Dry Chaco showed a cooling in surface temperature by 0.01±0.24 °C in the hilly areas of Sierras de Córdoba, although this was less significant and highly variable compared to other regions.

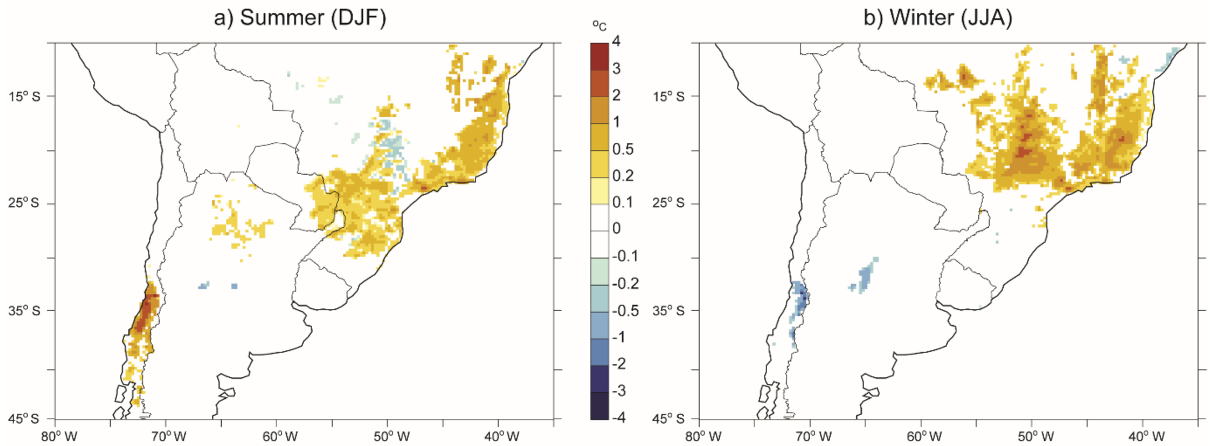

Figure 4.5 Significant differences of seasonal mean (CNTRL-NAT) for surface temperature (°C) across non-Amazonian South America. In eastern southern South America (Atlantic Forest, Cerrado and Dry Chaco) the dry season corresponds to austral winter (JJA) while in the Chilean Matorral corresponds to austral summer (DJF). This pattern reverses in the wet season. Only pixels that are statistically significant at a 95% confidence level are shown.

4.3.3.2 Wet season

During the wet season, the Atlantic Forest showed a significant warming of 0.5±0.12 °C in surface temperature which coincided with areas of highest deforestation (Figure 4.5-a). Similarly, the Dry Chaco experienced a significant warming of 0.2±0.19 °C, particularly in the northern areas (Figure 4.5). However, it also showed a decrease in surface temperature in southern hilly areas where deciduous forests were replaced by grasslands and crops (Figure 4.5). Likewise, for the Cerrado a slight cooling in surface temperature of 0.02±0.09 °C was recorded, particularly in those areas adjacent to the Atlantic Forest and was coincident with a replacement of savanna vegetation by crops. The Chilean Matorral also experienced a cooling in surface temperature of 0.37±0.18 °C, particularly near the Andes foothills.

Table 4.4 Mean differences ± standard deviation (CNTRL – NAT) for seasonal surface temperature and precipitation in four regions of non-Amazonian South America.

Region	Dry Season		Wet Season	
	Temperature (C°)	Precipitation (mm/month)	Temperature (C°)	Precipitation (mm/month)
Atlantic Forest	0.52 ± 0.16	-1.75 ± 3.39	0.5 ± 0.12	5.69 ± 6.40
Cerrado	0.68 ± 0.17	-0.49 ± 1.39	-0.02 ± 0.09	-0.35 ± 6.40
Dry Chaco	-0.01 ± 0.24	-0.96 ± 1.32	0.2 ± 0.19	1.55 ± 8.62
Chilean Matorral	1.42 ± 0.17	-0.47 ± 1.52	-0.37 ± 0.18	-2.58 ± 11.16

4.3.4 Change in precipitation and moisture

4.3.4.1 Precipitation

Figure 4.6 shows the mean differences in total precipitation for each of non-Amazonian South America. In general, modelled changes in precipitation showed a low significance due to the high variability of the rainfall response, particularly during the wet season (Figure 4.6-a).

Dry Season

All regions showed a mean decrease in total precipitation during the dry season (Figure 4.6-a). The Dry Chaco had the greatest reduction in precipitation with the current land cover conditions having 10% less rain relative to the natural vegetation conditions. These reductions were concentrated in the northern area of deforestation (Figure 4.6-a). Similarly, both the Atlantic Forest and the Cerrado showed a rainfall reduction of approximately 5%. However, only a small number of pixels were significant. Positive significant changes were observed in the northern coastal areas of Atlantic Forest, whilst the Cerrado did not record major significant changes (Figure 4.6-a). The same pattern was present in the Chilean Matorral which showed a mean reduction in rainfall of 3%, but this was not significant.

A somewhat similar pattern was observed for convective precipitation, with the Atlantic Forest showing a significant increase of 27% in coastal and hilly areas (Table 4.4, Figure 4.6-b). The Cerrado and the Dry Chaco showed a reduction of 6% while the Chilean Matorral showed a reduction of 2%. These changes were not significant.

Wet Season

Changes in rainfall were more significant during the wet season (Figure 4.6). The Atlantic Forest showed a significant increase in rainfall of 4%. A small but significant increase of 1% was observed in the southern extremity of the Dry Chaco. Similarly, a small significant reduction of 0.2% was observed in parts of the Cerrado. The Chilean Matorral did not experience significant changes during the wet season.

Changes in convective precipitation during the wet season followed a similar pattern of total rainfall, but over a larger area (Figure 4.6-b). The increase in convective precipitation was the greatest in the Atlantic Forest with a mean increase of 10% after deforestation. Smaller both positive and negative changes were recorded for the Cerrado and the Dry Chaco (Figure 4.6-b).

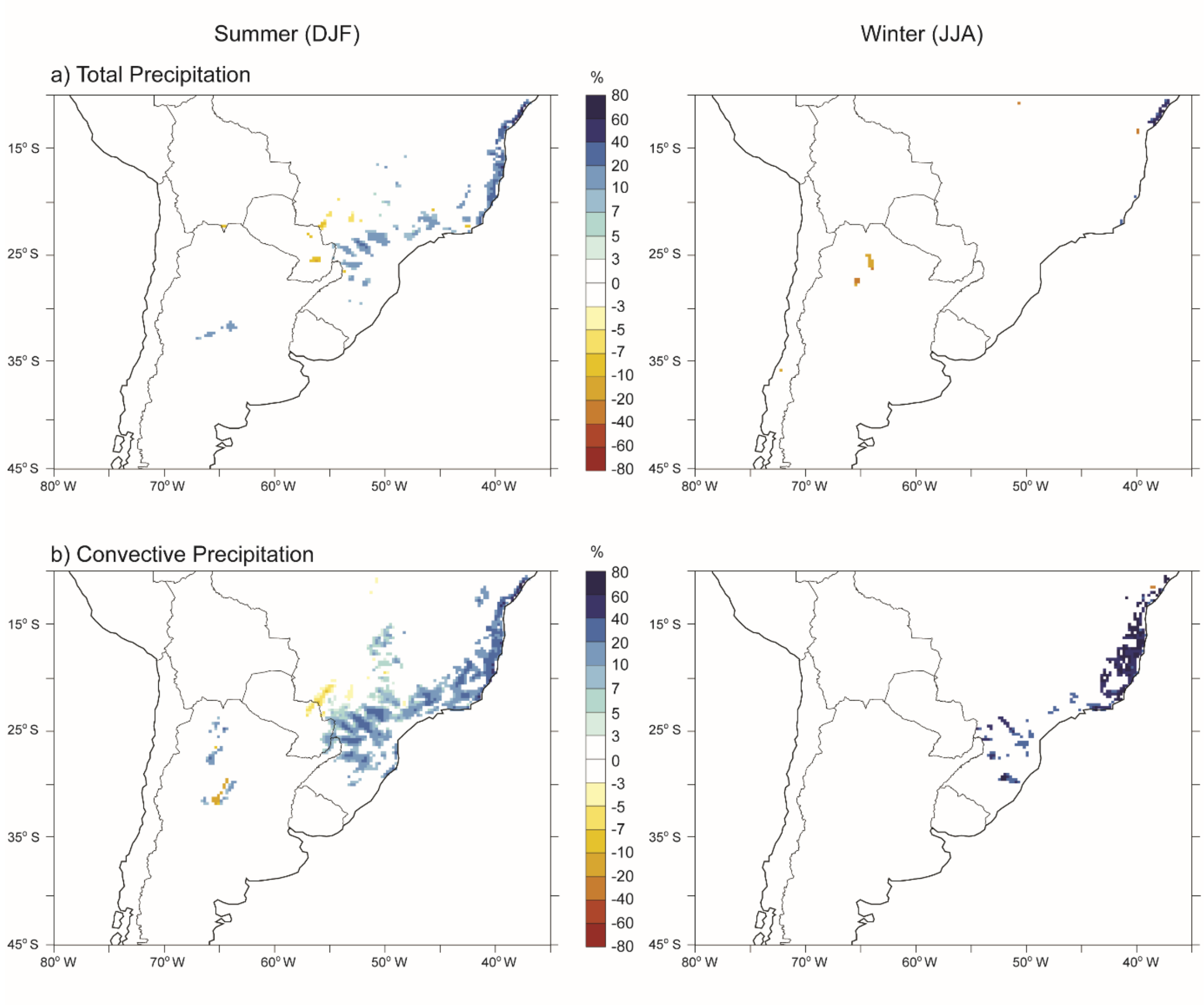

Figure 4.6 Mean changes in: a) total precipitation (%) and b) convective precipitation (%) between CNTRL and NAT experiments for non-Amazonian South America. The areas shown are those significant at 95% confidence level ($p < 0.05$).

4.3.5 Changes in atmospheric and soil moisture

4.3.5.1 Dry Season

Changes in precipitation were accompanied by a significant decrease in atmospheric moisture and were concurrent to areas of deforestation. These reductions were strongest where forests were converted to herbaceous vegetation. The Cerrado showed the greatest reductions of 8% in evapotranspiration, which located in those areas of reduced precipitation (Figure 4.7-a). Evapotranspiration in the Chilean Matorral decreased by 5%, mostly in areas were herbaceous vegetation replaced evergreen forest (Figure 4.7-a). Smaller reductions were recorded for the Atlantic Forest (3%) and the Dry Chaco (4%). With the exception of the Atlantic Forest, which recorded an increase in average plant respiration CO_2 flux, all regions registered a reduction in respiration of between 3 and 30%, indicating that decreased evapotranspiration is strongly influence by plant respiration during the dry season (Figure 4.7-f). Changes in evapotranspiration were observed following the same pattern of changes in latent heat.

Spatial and temporal variations in evapotranspiration were accompanied by a reduction in relative humidity. CCAM showed significant decreases in humidity throughout non-Amazonian South America (Figure 4.7-b). These were evident in areas with significant changes in surface temperature. The Atlantic Forest and the Cerrado registered the greatest mean reductions of 6%, with maximum reaching 22% coinciding with areas experiencing the greatest reductions in LAI (Figure 4.3 and Figure 4.7-b). The Chilean Matorral experienced a mean 2% mean reduction in relative humidity after conversion of forest to herbaceous vegetation. The Dry Chaco experienced a 2% reduction in relative humidity although southern hilly areas showed a small increase (Figure 4.7-b).

The Atlantic Forest showed the greatest increase in potential evaporation (Figure 4.7-a). Mean increases of 202% were recorded, though with positive changes up to fifteen fold in its eastern areas. Significant changes were also observed in the Chilean Matorral, which registered mean growths of 99% in potential evaporation during the dry season. Mean increases of 94% were recorded in the Cerrado, while the Dry Chaco registered a registered a smaller increase compared to the other regions. These changes in potential evaporation are associated to significant changes soil moisture, which showed reductions across all regions (Figure 4.7-d, e).

The Cerrado registered the largest reduction in soil moisture of 11%, followed by the Atlantic Forest (6%) and the Chilean Matorral (4%), whilst a reduction of 1% was observed in the Dry Chaco during the dry season. These changes in soil moisture were associated with a significant reduction in evapotranspiration (Figure 4.7-a, Table B.1 in Appendix B), which registered strong negative changes after deforestation. The changes were significant for all regions and as expected were correlated with variations in latent heat.

4.3.5.2 Wet Season

Increments in evapotranspiration were observed across eastern South America, though these ranged from 1-2%. For the Cerrado and the Dry Chaco, this change matched significant reductions of 26-32% in average plant respiration CO_2 flux, which indicates that most of increased evapotranspiration came from evaporation rather than transpiration (Figure 4.7-a, d). The Chilean

Matorral recorded a mean reduction of 12% in evapotranspiration without mean changes in average plant respiration CO_2 flux (Appendix B, Table B.1)

The Atlantic Forest experienced significant reductions of about 6% in relative humidity. By contrast, humidity did not show a significant change for the Cerrado (Figure 4.7-b). Changes in relative humidity in the Dry Chaco were larger than in the dry season, with mean changes of -5% corresponding with deforested areas (Figure 4.7-b). The Chilean Matorral showed a reduction of 3% with a slight increase observed in the northern Matorral.

Similar to the dry season, the wet season potential evaporation significantly increased over all regions with the Atlantic Forest showing the greatest mean increase of 88%, followed by the Cerrado (42%) and the Dry Chaco (32%). The Chilean Matorral recorded the smallest increase of 3% in potential evaporation during the wet season. As for the dry season, soil moisture decreased by 1-3% in most regions. By contrast, the Chilean Matorral showed an increase of 1% in soil moisture corresponding with areas of decreased potential evaporation and increased evapotranspiration, mostly in the Andes foothills (Figure 4.7-a).

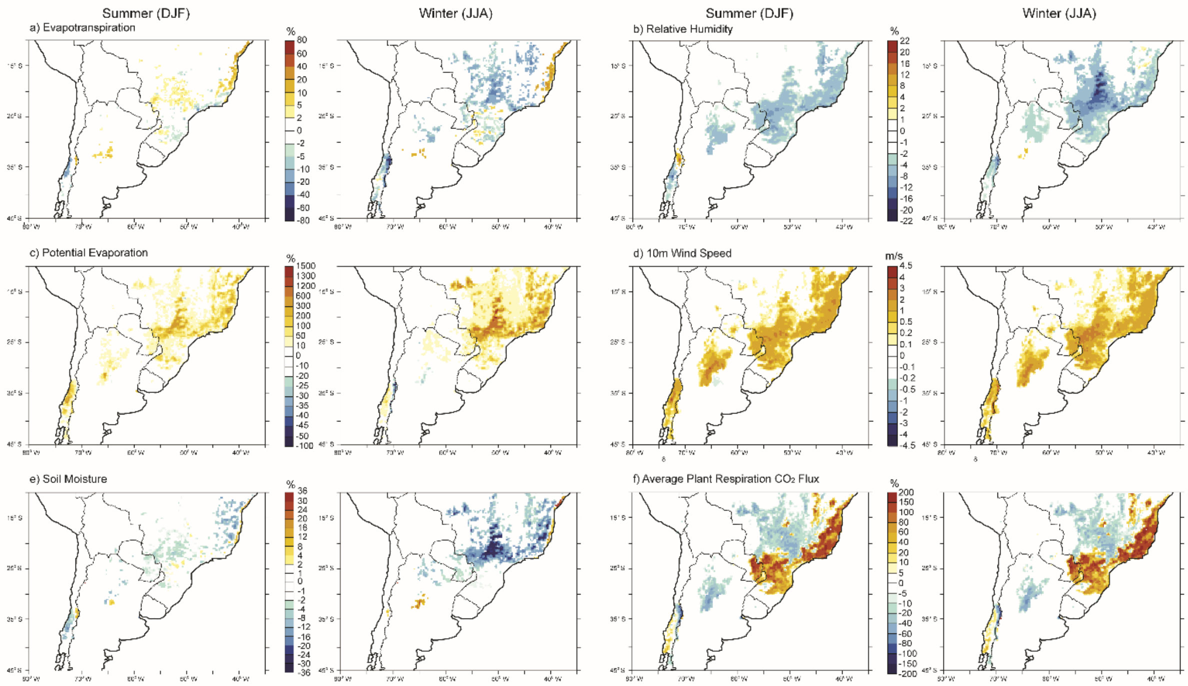

Figure 4.7 Seasonal significant changes for a) evapotranspiration (%), b) surface relative humidity (%), c) potential evaporation (%), d) 10 m wind speed (m/s), e) top level soil moisture (%); and f) average plant respiration CO_2 flux (%). Summer (DJF) corresponds to the dry season in the Chilean Matorral and the wet season in eastern South America (Atlantic Forest, Cerrado and Dry Chaco). This pattern changes in winter (JJA). Only pixels that are statistically significant at a 95% confidence level are shown.

4.4 Discussion

4.4.1 Key findings

Overall, the results support the alternative hypothesis that deforestation has significant impacts on the land–atmosphere interactions in non-Amazonian South America. However, there was considerable spatial and seasonal variability of the climate response, with changes in surface temperature and precipitation varying across regions and seasons in the 30 years of analysis. The key findings of our study are: i) deforestation has modified key characteristics of the non-Amazonia land surface that affect land-atmosphere coupling. For different biomes distributed in four key regions, we estimated reductions in the leaf area index of between 6 and 48% and reductions in roughness length of between 70 and 89%; ii) effects on heat fluxes and surface temperature show high intra-regional and seasonal variability. In the Atlantic Forest and Chilean Matorral sensible and latent fluxes tend to decrease during the dry season. Temperature variation manifests through a change of between -0.01 °C and +1.42 °C during the dry season and between -0.37 °C and +0.5 °C during the wet season. Temperature increase in the Atlantic Forest, Cerrado and Chilean Matorral were significant; iii) changes in precipitation showed low significance and high variability, indicating that other factors independent of forest cover control total precipitation. Though the combination of changes in roughness length and hilly areas seems to play a role in precipitation; and iv) deforestation has a significant impact on atmospheric and soil moisture in both seasons as expressed by a decrease of between 2 and 6% in relative humidity, an increase of 1 to 202% in potential evapotranspiration and a reduction of 1 to 6% in soil moisture. All these changes are more significant during the dry season, which tends to be drier and warmer after deforestation.

4.4.2 Contribution

This study differs from previous studies in three main aspects. First, we estimate natural vegetation for regions whose boundaries approach the original extent of natural communities prior to major LUCC (Olson et al., 2001). This compares to the modelled historical vegetation maps that, though they describe the global distribution of main vegetation types, they do not accurately describe natural vegetation in South America (e.g. Ramankutty and Foley, 1999). Second, we include those areas recognized as the most impacted by deforestation and least studied in terms of related climatic impacts. Finally, we used the CCAM model at 25 km spatial resolution using the ERA-Interim reanalysis (Dee et al., 2011) in combination with changes in vegetation cover and land surface characteristics. This approach allowed us to show significant regional and seasonal differences in the climate response to deforestation. In particular, we found variations across biomes, the potential influence of topography on the effect of deforestation for the Atlantic Forest and the Chilean Matorral, and the less sensitivity of rainfall compared to surface temperature for both the dry and wet season. In this regard, natural vegetation cover is more important to the hydrological cycle when the water is limited. Results from this study are in line with observations that show changes in vegetation cover have an important influence on the surface temperature and is particularly relevant in semi-arid regions where small changes in vegetation can have significant impacts on surface temperature and hence the water cycle (Mildrexler et al., 2011; Oyama and Nobre, 2004).

The results of this study are conditional on a number of limitations. First, the land cover map that shows present vegetation cover is developed for 2005 and does not capture recent LUCC processes such as those described by Hansen et al. (2013). Another limitation is the ability of surface models such as CABLE to accurately describe the diversity of functional vegetation types and land uses at regional scales. In addition, the lack of observation of fluxes and other prognosis variables at the same spatial and temporal scale of the model predictions makes it difficult to validate the results (Wang et al., 2011). This is particularly important for South America, which has been consistently described as a region that lacks of enough observation platforms to validate climate models projections (Magrin, 2014). In the following sections, we discuss the main findings and explain the underlying mechanisms for each of the regions addressed in this work.

4.4.3 Atlantic Forest

Changes in land-atmosphere interactions for the Atlantic Forest were spatially concurrent with deforestation. In general, results in the dry season were more robust and less variable compared to the wet season. During the dry season, variations in surface temperature can be explained by a reduced cooling capacity through sensible and latent heat fluxes, which finally warms the surface. Increased surface and soil temperature could also have strong influences on modelled moisture fluxes in the Atlantic Forest for both seasons. In the dry season, deforestation significantly decreases atmospheric and soil moisture. Strong increments in 10 m wind speed also influence moisture dynamics by transporting the moisture from the evaporating surfaces, which in turn increases potential evaporation.

The only previous study relating deforestation and the surface climate was conducted by Webb et al. (2005). The authors used weather station data in the state of São Paulo to identify links between deforestation and precipitation. Though they did not find significant relationships between forest cover and total rainfall, forest cover was positively related to the number of rain days. In our study, associations between deforestation and precipitation were more evident. However, the link is rather complex and it appears that distance from the coast and orography are key factors explaining the modelled precipitation patterns after deforestation. During the wet season (summer), the change in precipitation occurs in pixels located mainly near the Atlantic coast and at high elevations. Our results agree with those from Webb et al. (2005) indicating that distance to the coast in combination with elevation may exert strong influences on the precipitation pattern of the Atlantic Forest. As shown in Figure 4.8, there is a high proportion of significant pixels presenting an increased precipitation in the first 100 km from the coastline, with the next peak occurring at the highest elevation level (700 m) and decreasing towards the interior where the number of significant pixels showing a decreased precipitation seems to increase. These factors may act in combination with deforestation to significantly decrease surface roughness and increases 10 m easterly winds speed (Figure 4.7-d), which boosts the moisture transport from the coast to the Atlantic Plateau where increased condensation occurs. The complex interactions between orography, deforestation and precipitation still remain as an understudied field in vegetation climate feedbacks studies and more work needs to be done to elucidate these interactions in the Atlantic Forest.

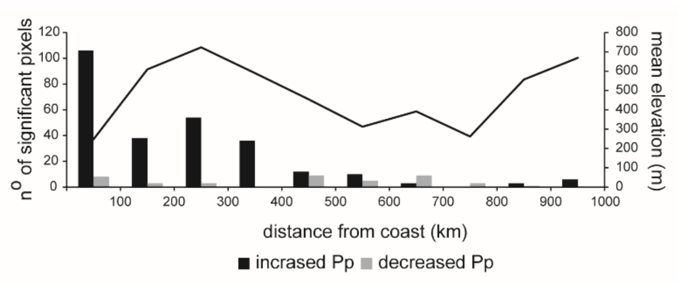

Figure 4.8 Number of significant pixels ($p < 0.05$) in relation to distance from the coast and elevation in the Atlantic Forest during the wet season (DJF). Bars represent number of significant pixels for increased (black) and decreased (grey) precipitation (Pp) after deforestation. The line refers to elevation taken from the CCAM model output expressed as the mean elevation for each distance range.

4.4.4 Cerrado

Small differences in surface roughness in the Cerrado are explained by small differences in canopy height between crops/grassland and savannas or the low proportion (10%) of forests in savanna type biomes considered in CABLE. Consequently, small significant differences emerge in heat and moisture fluxes, particularly during the wet season, though these differences are more conspicuous during the dry season. Here, the increase in temperature and heat fluxes was accompanied by a decrease in total precipitation and changes in soil moisture, humidity, convection, and a general decrease in atmospheric moisture content (Fig 7). Compared to the Atlantic Forest, coastal influences are less important and even low changes in the leaf area index can trigger significant atmospheric responses on the surface. This suggests that changes in latent heat fluxes can drive main atmospheric alterations to deforestation in the Cerrado. The overall result is a drier atmosphere and a more intense dry season.

These results agree with others studies in the Cerrado using other climate and surface models. For instance, Georgescu et al. (2013) modelled a warming after conversion of natural vegetation to sugarcane and a decrease in evapotranspiration and rainfall. Also, Lee et al. (2011), using an atmospheric model, found that climatic consequences of LUCC in the Cerrado occurs primarily during the dry season and are expressed through mean changes in surface temperature of about +0.7 °C during the dry season (in agreement with our results) and a drier atmosphere that according to the authors would stimulate more intense and frequent droughts in the Cerrado. This trend in temperature and moisture change from modelling approaches is confirmed by observations conducted by Loarie et al. (2011b) using satellite images over the entire Cerrado. The authors measured a mean increase of 1.55 °C and an evapotranspiration change of -0.6 mm/day when natural vegetation was converted to crops/pastures. This temperature response was mostly influenced by evapotranspiration rather than albedo changes. In our study, though mean albedo decreases for the Cerrado in both seasons, the mean increment in surface temperature during the dry season confirms the results from modelling and observations in terms of the higher relative importance of evapotranspiration on the surface temperature. As shown here, small changes in land cover can significantly affect the surface climate in the Cerrado.

4.4.5 Dry Chaco

In the Dry Chaco, the magnitude and spatial extent of the changes in surface characteristics is not as high as the Cerrado and Atlantic Forest. Differences in LAI are almost absent during the dry season and very low during the wet season. Therefore, variations in surface temperature in both periods are the lowest in non-Amazonian South America. The temperature during the wet season appears to be more sensitive to changes in vegetation than the dry season, possible because in the Dry Chaco deciduousness occurs during the last and no major changes in the leaf area index (Figure 4.3). Under the current land cover conditions, the difference between soil and surface temperature was 0° C implying that surface temperature approaches soil temperature after deforestation, affecting evapotranspiration rates and relative humidity. Though differences in latent and sensible heat fluxes were not significant, average plant respiration CO_2 flux for the Dry Chaco decreased 28% during the wet season, which indicates that increased evapotranspiration is mostly driven by increased soil evaporation than plant transpiration. During the dry season, differences between the sensible and latent heat fluxes become significant but absolute differences are less than 5 w/m^2 and no major increments in surface temperature were observed despite a 14% increase in the Bowen ratio. By contrast, surface temperature decreases in the southern tip of the Dry Chaco, which is consistent with increments in latent heat flux, relative humidity and soil moisture. The possible mechanism that explains these increments in moisture and ultimate precipitation in this area could be related to increased northerly and westerly 10 m wind speed carrying moisture from deforested areas (Figure 4.7-d) that finally condenses in the hilly areas of the southern Dry Chaco.

Other studies from the Dry Chaco describe similar patterns of temperature and precipitation response after deforestation. For example, Houspanossian et al. (2013), using satellite data, portray a mean diurnal temperature difference of +2.5 °C between dry forests and crops in the Dry Chaco, despite a 50% higher albedo of crops compared to forests. Similarly, the modelling study of Canziani and Carbajal Benitez (2012) found increments in surface temperature < 1 °C in deforested areas and beyond, without significant changes in precipitation. Even though LUCC climate feedbacks studies in the Dry Chaco are very scarce, the available evidence suggests that the near surface atmosphere is sensitive to the loss of natural vegetation with main changes expressed through modifications of surface characteristics and consequently partitioning of the available energy between sensible and latent heat fluxes, which together affect surface temperature and moisture recycling, yet without impacts on local precipitation. Because the extensive ongoing process of deforestation of the Dry Chaco, more research needs to be conducted to account for the potential climatic impacts of historic and future deforestation.

4.4.6 Chilean Matorral

Differences in surface characteristics in the Chilean Matorral are the second largest in non-Amazonian South America (after the Atlantic Forest) and the consequences for the surface climate are also high in magnitude, especially during the dry season. The region registers the greatest increment in surface temperature and the greatest decrease in mean precipitation (yet not significant). For the first, since heat fluxes decrease, in combination with a strong reduction in roughness length, a weaker coupling between the canopy and air temperature arises. This reduces the surface capacity to cool down through sensible and latent heat, which finally increases surface temperature significantly.

Although seasonal precipitation change after deforestation is not significant in the Chilean Matorral, the increase in moisture flux seems to be enforced in the foothills of the Andes Mountains. The mechanism related to this could be the same as the Atlantic Forest in terms of the importance of orography in the moisture transport. Precipitation in Chile is strongly influenced by zonal winds carrying moisture from the Pacific Ocean (Garreaud et al., 2009; Viale and Garreaud, 2015) and deforestation weakens the local influence of vegetation on the climate by increasing surface wind speed (because of a reduced roughness length) strengths the ocean influence that transport moisture up to the Andes and could explain the increased latent heat flux, increased soil moisture, and decreased temperature in those areas.

As in other regions of non-Amazonian South America, the process of deforestation in the Chilean Matorral increase overall dryness, predominantly during the dry season. Because this region is semi-arid and therefore sensitive to droughts and desertification, it can be argued that the loss of natural vegetation has increased the drying and potentially desertification process through significant changes in the hydrological cycle including surface temperature, evaporation rates, soil moisture content and atmospheric humidity. The strong increments in surface temperature projected by CCAM in the Chilean Matorral highlights the potential importance of surface vegetation over this scalar, which is higher compared to the increments without considering deforestation processes (Rosenbüth et al., 1997). Recent studies have described a drying trend in south-central Chile. For instance, Falvey and Garreaud (2009) analysed observational data to characterize the spatial pattern of surface temperature change in coastal and continental Chile. In the Central Valley, a region coincident with the Chilean Matorral, the authors report a temperature change of 0.18±0.14 °C/decade in the period 1979-2006. We report a temperature change of 1.42±0.17 °C considering only land cover change. Because deforestation can be a rapid and extensive process, the subsequent effect over surface temperature can be noticeable within a decade. This is important because it suggests that vegetation management in the Chilean Matorral could be a key factor that could dampen or enhance the local effects of global warming. However, the only previous study relating vegetation change with subsequent changes on the lower atmosphere focused on the impacts of irrigated agriculture in the northernmost extremity of Central Chile (Montecinos et al., 2008). Our study is the first attempting to identify the climatic impacts of historic deforestation at a local/regional scale in the Chilean Matorral and hence more research needs to be conducted to confirm the results presented here.

4.5 Summary and Conclusions

We have addressed the main impacts of deforestation on the surface climate of four regions of non-Amazonian South America representative of distinct biomes. These regions, particularly the Atlantic Forest and the Chilean Matorral, have been subjected to intensive pressures since the arrival of the first European settlers and are currently the most affected in relation to their original extent (Salazar et al., 2015). The replacement of natural vegetation by grasslands and crops has affected surface characteristics, the exchange of heat and ultimately the surface temperature and precipitation. The main findings can be summarized as follows:

1. Historic deforestation influences the partitioning of surface energy into sensible and latent heat flux. Changes in these fluxes are significant and more pronounced during the dry season in all biomes and are characterised by strong increases in sensible heat flux. During the wet season deforestation results in a decrease of the Bowen ratio because available energy is transferred to the near surface atmosphere through latent heat flux more than sensible heat flux. However, this water vapour transfer comes mainly from soil evaporation rather than photosynthetic transpiration, as shown by a general decrease in average plant CO_2 flux for the Cerrado, Dry Chaco and Chilean Matorral. In addition, increase in evapotranspiration during the wet season indicates that soil moisture is rapidly transferred into the atmosphere through latent heat flux, yet this moisture flux is still less than increased potential evaporation. After deforestation, there is a decrease in the vegetation control of water infiltration into the soils that decreases overall soil water retention. Hence, in the wet season there is less moisture on the soil after vegetation change. This pattern is many levels of magnitude greater during the dry season. Here, the Bowen ratio increases in the Cerrado, the Dry Chaco and the Chilean Matorral, with the first and the second showing the greatest increments.

2. Deforestation-climate interactions in non-Amazonian South America are complex. This complexity is added by the variety of vegetation types, the orographic effect of the Brazilian Plateau and the Andes, which is important in atmospheric circulation at a continental level. According to this study, surface climate interactions in the Atlantic Forest seem to be strongly influenced by orography and Atlantic Ocean vicinity. Similarly, our results suggest that the Chilean Matorral is strongly influenced by the proximity of both the Pacific Ocean and the Andes Mountains. However, the climatic changes in the Cerrado and the Dry Chaco depend more in evapotranspirative factors, due to their continentally. However, the impacts of historic

deforestation on precipitation, as shown by the results, indicate that precipitation is partially influenced but not governed by changes in forest cover.

4.6 Acknowledgments

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CHAPTER 5 IMPACTS OF HISTORIC LAND COVER CHANGE ON CLIMATE EXTREMES AND ARIDITY IN NON-AMAZONIAN SOUTH AMERICA

Abstract

Humans have made global changes to natural vegetation cover, modifying energy and water balances (Salazar et al., 2015). However, these changes are generally not considered a significant forcing under current climate change scenarios, especially at regional scales (Seneviratne, 2012). In this Chapter I use the results from regional climate model applied in Chapter 4 (same model set up, experimental design and running period) to explore these effects on climate extremes and aridity in non-Amazonian South America. Results show that the loss of natural vegetation has significantly affected temperature extremes as a decrease in the number of warm days and an increase in the number of warm nights. Importantly, there is a strong dependence on both seasonality and the vegetation contrast inflicted by land use/cover change (LUCC), with large roughness changes resulting in increasing wind speed and advection, while smaller roughness changes producing a larger distinction between sensible and latent heat fluxes. This explains the dry season response in both temperature extremes and the increase in aridity according to LUCC, whereby regions with increased wind speed have a greater impact on atmospheric water demand than those that mainly increase sensible heat fluxes. These results can explain the observed trends in temperature extremes in non-Amazonian South America (Alexander et al., 2006) and highlights the need to embed LUCC as a forcing within future climate change scenarios and potential adaptations to changes in climate extremes and aridity.

This chapter has been submitted as a journal paper to *Nature Geoscience Letters* as Salazar, A. Larsen, J., Callow, N., Wong, K., Syktus, J. and McAlpine, C. Impacts of historic land cover change on climate extremes and aridity in non-Amazonian South America. In order to keep the thesis coherent, this Chapter was slightly modified from the version submitted to *Nature Geoscience Letters*. Some elements of Chapter 4 are repeated in this Chapter, particularly those referred to climate model setup and experimental design.

5.1 Introduction

Humans have made global changes to natural vegetation cover, modifying energy and water balances (Salazar et al., 2015). However, these changes are generally not considered a significant forcing under current climate change scenarios, especially at regional scales (Seneviratne, 2012). In this study we use a regional climate model to explore these effects on climate extremes and aridity in non-Amazonian South America. We find that the loss of natural vegetation has significantly affected temperature extremes as a decrease in the number of warm days and an increase in the number of warm nights. Importantly, there is a strong dependence on both seasonality and the vegetation contrast inflicted by land use/cover change (LUCC), with large roughness changes resulting in increasing wind speed and advection, while smaller roughness changes result in feedbacks more reliant on the distinction between sensible and latent heat fluxes. This explains the dry season response in both temperature extremes and the increase in aridity according to LUCC, whereby regions with increased wind speed reduce warm day temperature extremes, despite increasing mean temperature trend, and have a greater impact on atmospheric water demand than those regions that mainly increase sensible heat fluxes. These results can explain the observed trends in temperature extremes in non-Amazonian South America (Alexander et al., 2006) and highlights the need to embed LUCC as a forcing within future climate change scenarios.

Climate change has demonstrable impacts in the frequency, duration, intensity, spatial scale and timing of climate extremes (Seneviratne, 2012). Observational data indicates a significant global change in extremes associated with maximum and minimum temperatures, extreme precipitation events and more intense and longer droughts (Alexander et al., 2006; Dai, 2013; Seneviratne, 2012). To date, most studies have linked variations in these extremes with anthropogenic forcings (Seneviratne, 2012), principally through increasing atmospheric CO_2 concentrations (Seneviratne, 2012). However, because LUCC affects the surface temperature and moisture recycling, the climate feedbacks resulting from conversions of natural vegetation to other land uses may produce climate changes similar to those imposed by anthropogenic CO_2 (Pitman et al., 2012), especially at regional scales. Being able to accurately separate these different human impacts on climate is therefore necessary for future climate change scenarios, especially in attributing cause and effect to changes to climate extremes. However, the forcing contribution of LUCC on regional changes in extremes and aridity has been difficult to test explicitly (Christidis et al., 2013; Deo et al., 2009). Therefore, the potential influence that LUCC could exert on these processes remains poorly understood, particularly in the tropics and sub-tropics where LUCC have seen large scale conversion of forests and savanna and there are large contrasts in the water and energy balances between the summer wet, and winter dry seasons (Hansen et al., 2013; Kim et al., 2015).

LUCC can alter the surface climate through changes in biophysical attributes such as roughness that then affect wind speed, the partitioning of sensible, latent and ground heat fluxes, and convective cooling (Foley et al., 2003). Cumulatively, these land surface alterations can impact the climate at a range of spatial and temporal scales (Bonan, 2008b; Fairman et al., 2011; Lawrence et al., 2012; Mahmood et al., 2011). Modelling studies applied to tropical forests suggest that deforestation produces a drier and warmer climate due to reductions in evaporative cooling, roughness length and increases in surface temperature and wind speed (Hoffmann et al., 2003; Oyama, 2003). Nonetheless, despite the scale of the loss of natural ecosystems, there are still many uncertainties related to the role LUCC may play in explaining the observed and modelled changes in temperature extremes and drying/wetting trends. The relevance of these vegetation changes lies in the fact that biophysical alterations imposed by LUCC can be significant compared to changes in CO_2 forcing at local and regional scales, the scale at which people and ecosystems are mostly affected (Mahmood et al., 2010).

LUCC feedbacks on the land surface – climate system are particularly relevant in non-Amazonian South America, where extensive areas of former forests, savannas and shrublands have been converted to industrial agriculture and cattle production systems, affecting 3.6 million km^2 of the former natural vegetation cover, which is four times greater than the scale of Amazon deforestation (Gasparri and le Polain de Waroux, 2014; Hansen et al., 2013; Kim et al., 2015; Salazar et al., 2015). As such, this region contains many of the key LUCC conversions observed globally and also registers the highest current rates of deforestation in the tropics and sub-tropics (Hansen et al., 2013; Kim et al., 2015). In this paper, we estimate the seasonal impacts of historic LUCC on percentiles-based temperature and precipitation climate extremes (Appendix C, Table C.1), as well as the aridity index (precipitation (P) /potential evapotranspiration (PET)) using a 3 ensemble climate model for present day land cover (CNTRL) and natural vegetation (NAT) across non-Amazonian South America and four main ecosystems therein: Atlantic Forest, Dry Chaco, Cerrado and the Chilean Matorral, all together comprising an area of about 3 million km^2 (Figure 5.1-a). The regional climate model setup is deliberately conservative in terms of land-atmosphere feedbacks by nudging each time step with ERA-Interim reanalysis (Dee et al., 2011) inputs, and using realistic historical land cover projections to implement the difference between current and historical land use in the model (Figure 5.1-b). Most LUCC are expressed as a conversion from forest to grasses (crops and pastures), particularly in the Atlantic Forest, while conversion from savanna to grasses dominates in the Cerrado.

5.2 Methods

5.2.1 Model setup

We used the Conformal-Cubic Atmospheric Model (CCAM) coupled with the Community Atmosphere Biosphere Land Exchange (CABLE) surface model (Kowalczyk et al., 2006; McGregor, 1996; 2003; 2005a; 2005b). CCAM was driven by ERA-Interim reanalysis (Dee et al., 2011) in order to reduce the noise produced by climate oscillations while preventing possible model biases. Model domain was set between latitude 10-45° S and longitude 30-90° W at about 25 km horizontal resolution. We completed 3 ensembles for two sets of model simulations for the period 1982 and 2009 (28 years). The only difference between the simulations was the description of land surface datasets. The first scenario had present land surface characteristics and the second simulation had pre-European impact surface characteristics. In order to implement these pre-impact conditions, we modified land surface in four main ecosystems: 1) the Atlantic Forest, 2) Cerrado, 3) Dry Chaco and 4) Chilean Matorral. To extrapolate historic natural vegetation we first identified remnant natural vegetation in the MODIS image that agreed with descriptions from literature of natural vegetation types for each ecosystem (e.g. Savannas in the Cerrado or deciduous broadleaf forest in the Dry Chaco). We then projected historic natural vegetation by replacing current (e.g. crops) vegetation according to ecosystem types. Present land surface characteristics were set for modern day conditions in all locations outside the focus ecosystems for all simulations. More information is provided in Appendix C.

5.2.2 Climate extremes

For each simulation we extracted daily data of surface temperature to calculate percentile-base seasonal (summer and winter) extreme indices (Donat et al., 2013). For temperature extremes we used warm days, cold days, warm spell duration, warm nights, cold nights and cold spell duration (Appendix C). The statistical difference between simulations was assessed using bootstrapping at 99% confidence level ($p < 0.01$) where $X = \{X_1, X_2, ..., X_{n_x}\}$ is the sample of a climate extreme or aridity index from the natural vegetation experiment during period 1982-2009, and $Y = \{Y_1, Y_2, ..., Y_{n_y}\}$ the sample of the same variable taken from the unmodified control experiment in the same period. The null hypothesis (H_o) states that there is no significant difference between the means $\overline{X}$ and $\overline{Y}$ (i.e. LUCC has no significant impact on the selected climate extreme). The alternative hypothesis (H_1) states that there is a significant difference between the means $\overline{X}$ and $\overline{Y}$ (i.e. LUCC has a significant impact on the selected climate extreme). The observed *t*-statistic is $t_{obs} = \left((\overline{Y} - \overline{X}) / \sqrt{(\sigma_x^2 / n_x) + (\sigma_y^2 / n_y)}\right)$, where $\left(\overline{X}, \sigma_x n_x\right)$ and $\left(\overline{Y}, \sigma_y, n_y\right)$ are the mean, standard deviation and sample size of the natural and control samples, respectively. The bootstrap statistic for each sample is computed as $t^* = \left((\overline{Y}^* - \overline{X}^*) / \sqrt{(\sigma_{x^*}^2 / n_{x^*}) + (\sigma_{y^*}^2 / n_{y^*})}\right)$, where $\left(\overline{X}^*, \sigma_{x^*}^2, n_{x^*}\right)$ and $\left(\overline{Y}^*, \sigma_{y^*}^2, n_{y^*}\right)$ are the mean, standard deviation and sample size of randomly selected bootstrapped samples, respectively. The Achieved Significance Level (ASL) is the proportion of samples where $\left|t^*\right| \geq \left|t_{obs}\right|$. The p value is calculated as $p = 1 - \text{ASL}$. The null hypothesis is rejected if $p < 0.01$, indicating a significant change in the climate variable across the two scenarios for the 28-year period. The sample sizes for X and Y were both 84 (3 ensembles over 28 years), with $N = 500$ bootstrap samples conducted to test for statistical significance. Additional details are provided in the Supplementary Information.

5.3 Results and discussion

Seasonal indices for changes in temperature extremes show high statistical significance, whereas we find no significant response in the precipitation extremes (Appendix C, Table C.1). This result is consistent with previous studies which have also suggested a limited effect of LUCC on precipitation in non-Amazonian South America (Beltrán-Przekurat et al., 2012b). In terms of temperature extremes, the number of warm days and warm nights exhibit the greatest significant difference as a result of LUCC (Figure 5.2-c, d). Interestingly, this generally agrees with observational trends for the same region (Alexander et al., 2006; Donat et al., 2013). Warm days tend to decrease following LUCC in both wet and dry seasons with the exception of the Cerrado, which shows an increase of 3.63±2.82 extreme warm days in winter (dry season). This contrast is likely due to a sharp increase in the Bowen ratio, which was far greater in the Cerrado compared to all the other ecosystems studied (Appendix C, Fig. C.1, Table C.4). The negligible difference in roughness length following LUCC in the Cerrado (savanna to gasses) results in an increase in the surface-atmospheric coupling due to the strong increase in the sensible heat flux (3.82±1.3 wm^{-2}) and little change in wind speed or advection, which together increases vertical mixing and the dry season planetary boundary layer depth by up to 100 m (Appendix C, Fig. C.l).

An opposing trend occurs for wet and dry season extreme warm days in the other ecosystems. Here, the change from forest to grasses has strong impacts on the surface roughness and stimulates an increase in wind speed of between 75 and 106% in the dry season and between 78 and 104% in the wet season (Appendix C, Fig. C.1-h). This increase in wind allows greater advection of heat away from deforested areas, reducing vertical mixing and the planetary boundary layer depth (Appendix C, Fig. C.1), and can therefore account for the diminishing number of dry season extreme warm days by 2.60±1.57 in the Dry Chaco to 17.53±1.55 days in the Chilean Matorral (Figure 5.1-c). Importantly, these daytime feedbacks on extreme temperatures can occur despite an otherwise increase in mean surface temperature in most of the ecosystems. This indicates that large changes in vegetative roughness induce new (more regional) influences on local climate feedbacks that cause the response in maximum temperature extremes to differ from the overall warming trends in mean surface temperature.

In contrast to the general increase (except for the Chaco) in extreme warm days, LUCC was found to drive an increase in extreme warm nights for all ecosystems examined, with the exception of the northern Chilean Matorral and minor areas of the Dry Chaco in winter due to localised topographic effects (Figure 5.1-d). The number of extreme warm nights increased between 1.49±1.21 in the Dry Chaco and 5.46±1.89 nights in the Atlantic Forest during summer, most likely driven by a significant increase in soil temperature, which is most conspicuous during summer in the Atlantic Forest (wet season) and the Chilean Matorral (dry season), and during the dry season in the Cerrado and the Dry Chaco (Appendix C, Fig. C.1). This rise in soil temperature stimulates the release of stored heat during night time surface cooling, raising minimum temperatures and hence also extremes in warm nights.

In order to explain the temperature extreme results within the context of overall land-atmosphere feedbacks due to LUCC in non-Amazonian South America, we also examine the effects of LUCC on aridity. In this regard, there is a significant increase in aridity following LUCC that varies across ecosystems, with the wet season exhibiting greater changes in aerial extent, and the dry season much greater changes in magnitude (Figure 5.2). Interestingly however, much of the Cerrado exhibits only small or negligible changes, particularly in dry season aridity. In terms of attributing these changes in aridity to P or PET, we found no significant changes in P across most of the study area, thus the overall increase in aridity can be attributed to PET, which increases during the dry season from 1% in the Dry Chaco to 202% in the Atlantic Forest. This is caused by the increased wind speed and a 3 – 12% decrease in dry season latent heat flux across much of non-Amazonian

South America following LUCC, except the Cerrado where small roughness and wind speed changes ensure only small increases in PET.

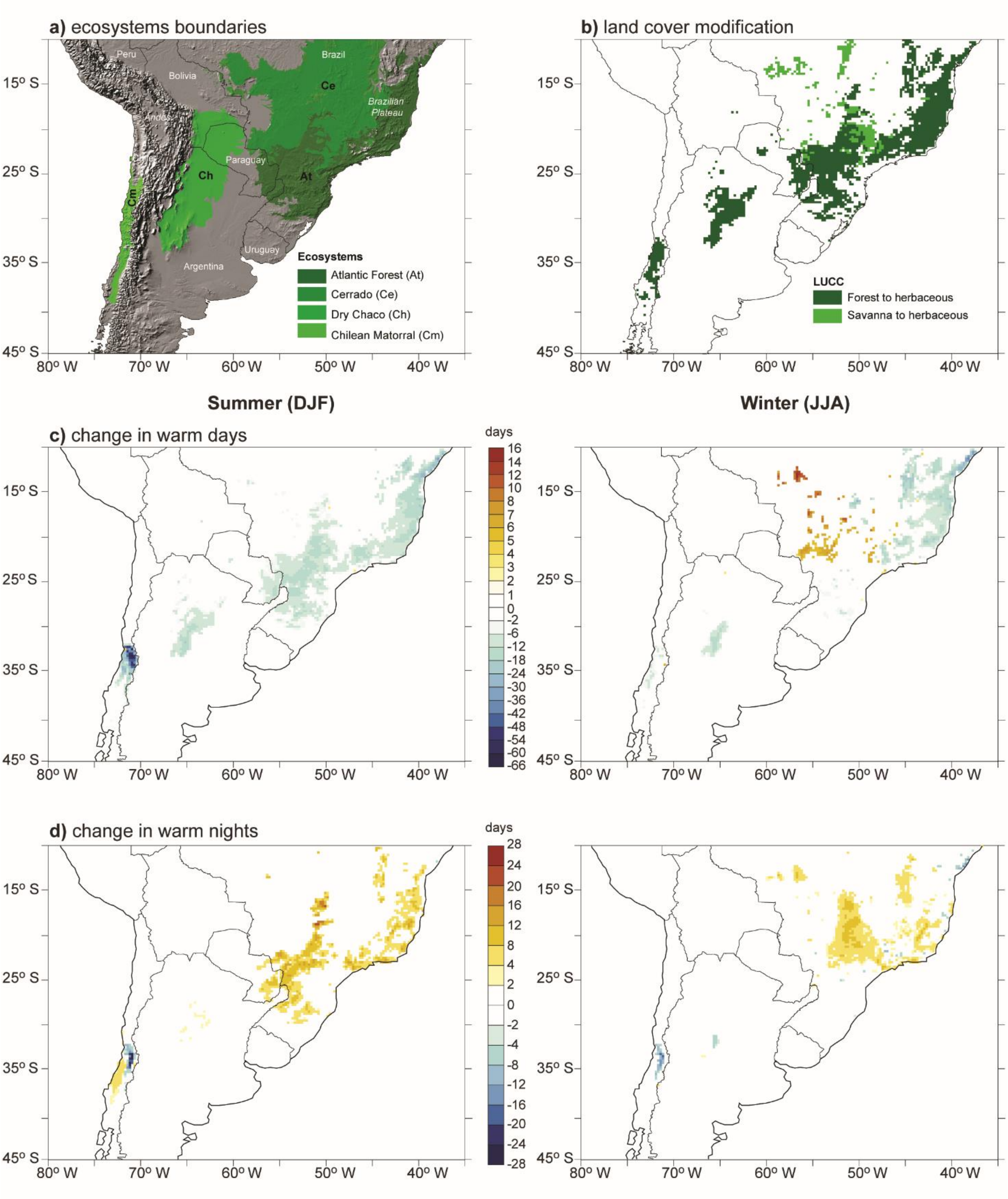

Figure 5.1 Study area and land cover modification (maps a) and b)) and maps showing significant differences (CNTRL-NAT, $p < 0.01$) of c) the number of warm days and d) the number of warm nights across non-Amazonian South America as modelled by CCAM climate model for period 1982-2009. For both extreme indices units are in days

per season. In southeastern South America (Atlantic Forest, Dry Chaco and Cerrado), summer corresponds to the wet season, while in the Chilean Matorral the dry season is in winter.

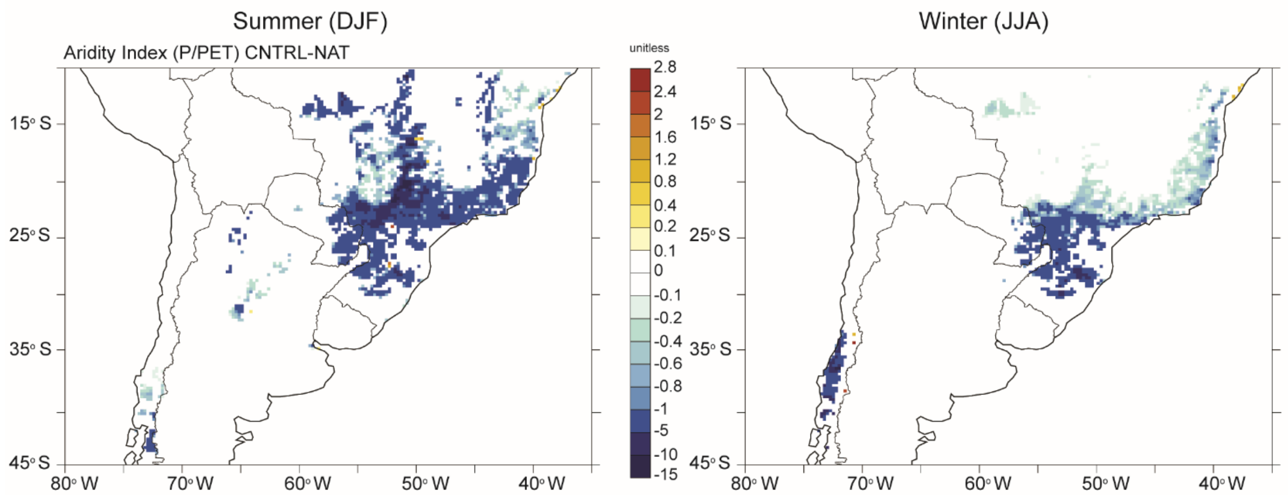

Figure 5.2 Significant seasonal differences (CNTRL-NAT, $p < 0.01$) in the Aridity Index (P/PET) across non-Amazonian South America calculated from CCAM model output 1982-2009. A decrease in the Aridity Index represents an increase in aridity. Country borders are shown.

Based on these results, a conceptual model for the feedbacks associated with LUCC that can account for changes in temperature extremes and aridity are shown in Figure 5.3. The LUCC feedbacks are differentiated in terms of total roughness change (forest to grass, or savanna to grass). An important result of this work is that although there is an increase in mean surface temperature following LUCC, changes in soil heat storage tends to increase warm night extremes, whereas changes in warm day extremes are linked to roughness length and wind speed change, leading to a redistribution of heat through advection that may differ from the mean temperature trend. This same set of feedbacks also explains the overall increase in PET and aridity, thus we expect that areas with increased atmospheric moisture demand following LUCC should also be linked with decreased daytime extreme temperatures. It is important to highlight that these results are model dependant, however the use of multi-model ensembles (IPCC, 2013b) is generally not feasible for LUCC experiments because of the lack of consistency in surface model parametrizations (Christidis et al., 2013; Pitman et al., 2012; Pitman et al., 2009). In our regional climate model, we replaced the default land surface data with more accurate high resolution surface descriptions of land cover and leaf area index. There are opportunities however, to further improve these adjustments by incorporating, for example, irrigated croplands, which are also important in land-atmosphere interactions in semiarid environments and yet are also not currently accounted for in climate change scenarios (Kharol et al., 2013; Lo and Famiglietti, 2013). Nonetheless, our work provides a clear mechanistic explanation for the impact of LUCC on climate extremes and aridity. Given the degree of change in temperature extremes that can be ascribed to LUCC alone, it is imperative that future climate change scenarios address the evolving modifications to land surfaces across the globe in order to improve our understanding of human influences on climate.

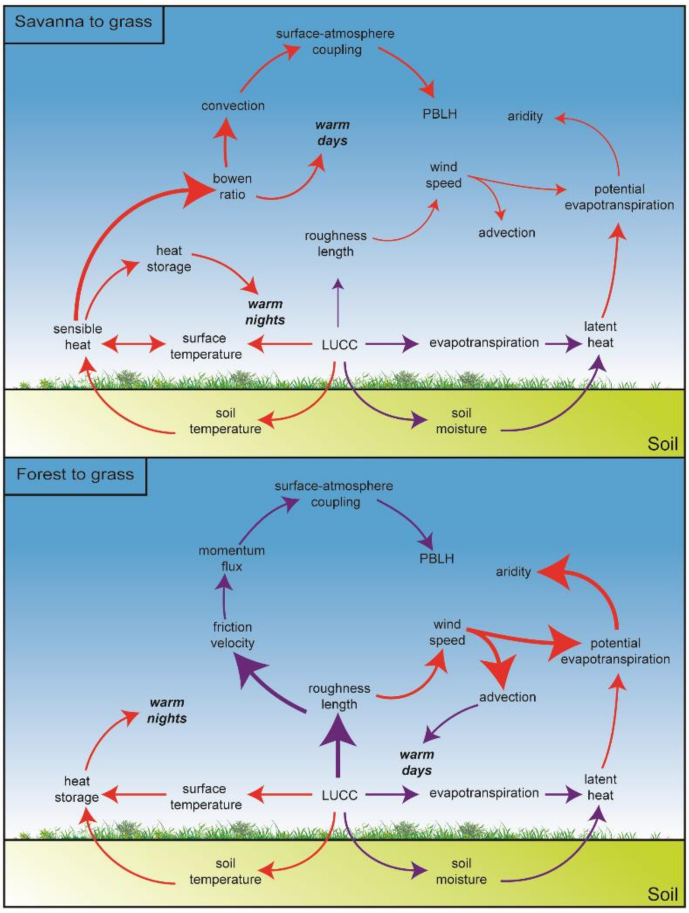

Figure 5.3 Schematic representation of the main dry season changes in land-atmosphere interactions after LUCC and subsequent impacts on warm days and nights based on the results presented in this study. Interactions are expressed for the dry season only. Changes are shown according to conversions from savanna and forest to crops/pastures. Red and blue arrows respectively represent positive and negative interactions. Arrows width indicates strength of interactions. PBLH refers to planetary boundary layer depth. For changes from savanna to grasses, the variation in the number of warm days is dominated by increases in the Bowen ratio, while from changes from forest to grasses warm days response is influenced by increases in wind speed and advection. In both cases the increase in the number of warm nights is affected by the release of an increased stored heat during cooling at night time. Aridity is strongly affected with changes in roughness length through changes in wind speed and potential evapotranspiration, especially when forest is converted to grasses.

CHAPTER 6 CONCLUSIONS

This chapter provides an overview of the key findings of the thesis. It also describes the major outcomes in relation to each specific objective, and challenges and limitations encountered along the way. It briefly outlines the most important contributions to the theory and practice of surface-atmospheric interactions, specially referred to non-Amazonian South America, and with future research priorities arising from the main results.

6.1 Major findings

This thesis investigated the changes in land-atmosphere interactions because of LUCC in non-Amazonian South America and, in particular, in the Atlantic Forest, the Cerrado savannas, the Dry Chaco and the Chilean Matorral. For the first time, it describes the most important mechanisms behind the changes in the energy and water balance imposed by the loss of natural vegetation and stresses the need of including LUCC in studies focused on anthropogenic climate change. It also identifies the areas that need to be developed in order to broaden the understanding of the atmospheric changes after LUCC and will help to propose new management practices that can potentially dampens the negative effects of climate change.

Objective 1

The first objective was to review the existing literature about LUCC processes and their impacts of the regions climate of non-Amazonian South America (Chapter 3, published in *Global and Planetary Change*). The main argument of this objective was that, despite the region has experienced high rates of conversions of its natural vegetation to other land uses, conversion extent and subsequent climatic impacts are not well understood because most of the attention has been directed to the Amazon deforestation. The work related to this objective addresses these differences and identifies current knowledge gaps that need to be tackled by future research. For the first time in the literature directed to South America, it gives a general view of the state of the natural ecosystems in relation to their historical extent and the current knowledge with regards to subsequent changes in the surface climate. The principal outcomes of this objective are specified as follows:

Outcome 1

I demonstrated that the historic loss of natural vegetation in non-Amazonian South America is about 4 times greater than the historic Amazon deforestation (3.6 million km^2). The Atlantic Forest and the Chilean Matorral show the greatest relative historical conversion of natural vegetation (81 and 83%, respectively), and along with the Cerrado and the Dry Chaco also show high conversion rates after year 2000 (Table 3.2).

Outcome 2

LUCC effects on the surface climate in non-Amazonian South America are very low compared to the Amazon. The most impacted are the least studied ecosystems: The Atlantic Forest and the Chilean Matorral, both with only one study. This outcome shows that, despite the high current and historic conversion of the natural ecosystems of non-Amazonian South America, studies focused on related changes in the surface climate are scarce and hence poorly understood.

Outcome 3

The few published work addressing changes in the surface climate because of LUCC show that these are expressed through alterations of heat fluxes and radiation, the last very important in arid and semiarid environments. Though changes tend towards a reduction on moisture and increase in surface temperature, the climate response depends also on factors such as physiology, latitude and topography, and large scale atmospheric circulation.

Objective 2

In the second objective I used a regional climate model along with realistic representations of historical natural vegetation cover to investigate the impacts of historic LUCC on the climate of non-Amazonian South America (Chapter 3, in review with *Global and Planetary Change*). The work focused on those ecosystems considered most affected and least studied arising from the findings of the previous objective: The Atlantic Forest, the Cerrado, the Dry Chaco and the Chilean Matorral. The results revealed significant changes in the surface temperature, moisture recycling and heat exchange. The changes were particularly conspicuous during the dry season, when the temperature increase reached a maximum of 1.42 °C in the Chilean Matorral. The principal outcome of this objective are specified as follows:

Outcome 1

I demonstrated that LUCC significantly affects the surface climate in non-Amazonian South America. These changes are mainly expressed through alterations in surface fluxes, temperature, precipitation and moisture on the ecosystems examined. The Chilean Matorral and the Cerrado registered the greatest increase in surface temperature during the dry season. On the contrary, precipitation is not sensitive to LUCC, except in those areas where topography is complex (Brazilian Plateau and the Andes Mountains). In addition, moisture recycling is strongly affected by LUCC. High increments in potential evapotranspiration were calculated, which were accompanied by significant reductions in soil moisture and relative humidity.

Objective 3

In this objective, the LUCC impacts on the mean climate were further analysed through an examination of variations in selected climate extremes and the aridity index in the same ecosystems worked in Objective 2 (Chapter 4, in review with *Nature Geoscience Letters*). In this objective I propose a conceptual model between LUCC and climate extremes and aridity, which is a contribution for the study of these interactions regionally and globally. The outcome of this objective revealed potential mechanisms that explain the interaction between changes in vegetative properties and the water and energy balance. Specifically, the importance of roughness length on wind speed and potential evapotranspiration affecting aridity, and the importance of heat fluxes on surface temperature and changes in moisture affecting the number of warm days and nights. The main outcome resulting from this objective is specified as follows:

Outcome 1

LUCC affects climate extremes and aridity in non-Amazonian South America. The mechanisms behind these changes depend on LUCC direction. For changes from savanna to grasses, the variation in the number of warm days is dominated by increases in the Bowen ratio, while from changes from forest to grasses warm days response is influenced by increases in wind speed and advection. In both cases the increase in the number of warm nights is affected by the release of an increased stored heat during cooling at night time. Aridity is strongly affected with changes in roughness length through interactions with wind speed and potential evapotranspiration, especially from changes from forest to grasses.

6.2 Challenges and limitations

The outcomes of this research were constrained by many challenges and limitations. These were treated carefully to increase the robustness of results and hence maximize the contribution of this study to the theory and practice of land-atmosphere interactions research. The principal limitations identified in this thesis were: a) the lack of accurate and high resolution descriptions of land cover and natural vegetation distribution in South America, b) the lack of studies focused on LUCC impacts on the regional climate and therefore the absence of enough evidence to sustain a theoretical framework of surface-climate interactions in the region, and c) the accuracy of surface descriptions incorporated in the climate model that does not include key land cover types (e.g. irrigated croplands).

One strong challenge in conducting this research was the scarcity of current land cover descriptions for South America and the absence of reliable and accurate descriptions of historic natural vegetation. Some of the most used maps of natural vegetation in climate studies were discarded because their low accuracy in describing land cover types for the region under analysis. To deal with this limitation, I conducted a thorough literature review focused on those studies describing the distribution of the original natural vegetation types in South America, and particularly in non-Amazonian South America. I also identified the best datasets and studies describing current land cover conditions. This process allowed me to construct the most realistic maps of natural and current vegetation based on the literature and available datasets.

The lack of studies focused on climatic impacts of LUCC in non-Amazonian South America limited the ability to compare or validate the main results of this research. Despite that this represent an important contribution of the thesis, it also challenged the identification of patterns due to the size of the area investigated, its highly complex topography and diversity of climatic features. In this regards, I relied on the few but excellent descriptions of the climate of South America and the 19 studies describing LUCC climate interactions in the region identified in Chapter 3. For the last, the results between studies are not comparable because differences in vegetation descriptions used in surface models and the lack of consistency in model parametrizations.

6.3 Contribution

This thesis provides a comprehensive analysis of LUCC impacts on the climate of non-Amazonian South America. It represents a novel approach of land-atmosphere interactions in a region highly impacted by the loss of its natural ecosystems and increasingly vulnerable to changes in the mean and extreme climate. The contribution of the study can be summarizes in 3 key aspects:

Contribution 1: Quantifies the magnitude of natural vegetation change and the state of knowledge referred to the subsequent impacts on the mean climate and in climate extremes. The outcomes stress the need to consider land use and land cover change as an important forcing of surface climate, which can be substantial at regional scales.

Contribution 2: Provides new evidence that the response of climate extremes can differ from that of mean climate because of feedback processes, which change according to the type of vegetation involved. This represents the first evidence of these interactions in South America and the first mechanistic explanation of the interactions associated, which is of global interest.

Contribution 3: Provides a novel and realistic systematic analysis of the historical processes of land use and land cover change in South America and evaluates the climatic consequences. It gives evidence toward a change of paradigm in relation to the relevance of land use and land cover change as an important regional forcing of anthropogenic climate change.

6.4 Research priorities

The main outcomes of this research highlight the importance of land use and land cover change on the regional climate. Given the rate of historical and current loss of native vegetation in many regions of the globe, there is an urgent need of considering the relevance of the alteration of surface processes on the regional climate. This urgency connects with many other scientific disciplines and complements the increasing global need to propose and implement management practices to stop or decrease the consequences of global change. This thesis stresses the need of a regional focus because this is the scale at which people and ecosystems experience these consequences.

I highlight the importance of land use and land cover on the mean and extreme climate through changes in biophysical properties that significantly impact the surface-atmospheric coupling and therefore the hydrological cycle. This is of great relevance for water-limited environments and hence there is a need to direct the attention to these types of environments. This is of great importance for South America because currently more than 200 million inhabitants live in these areas and are already experiencing the consequences derived from native vegetation loss and desertification, which are expected to increase in the future. Though modelling approaches are currently the only tool to understand the feedbacks in the climate system, it needs to be complemented with other tools such as those provided by remote sensing and other observation platforms.

One main difficulty in the process of this thesis was the lack of coherent spatial information at regional scale and the lack of climatic information to understand patterns and the magnitude of changes in climatic variables. This is a common limitation mentioned in all IPCC assessment reports. There is a need, therefore, to increase the observation network in the continent, particularly for those non-coastal areas that are currently experiencing the greatest rates of vegetation change. This will also help to validate climate models and increase their applicability at finer scales. In this regard, the surface models embedded in regional climate models must to be calibrated according to the surface characteristics of the area of interest. This is relevant for South America because its high diversity of vegetation types, complex topography and climatic features.

Finally, I highlight the importance of management of natural ecosystems for the regional climate. This is of great importance in South America because, as shown in this research, the contribution of the loss of natural vegetation to the increasing surface temperature is comparable to those from increasing greenhouse gasses. The management implications rely on the fact that the way countries manage their natural vegetation can have great impacts on surface temperature and the hydrological cycle. This means that, through natural ecosystem management, countries are able to dampen of exacerbate the negative consequences of climate change and therefore have a direct influence on their future and the welfare of their societies.

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Appendix A

Short Biogeography of non-Amazonian South America

Dry Chaco

The Dry Chaco is the second largest forested biome of South America after the Amazon and covers ~790,000 km^2. It extends from northern Argentina, western Paraguay and south-eastern Bolivia to Brazil. The region is mainly composed by deciduous trees with the presence of a discontinuous shrub layer and sparse natural grasslands occurring in areas of sandy soil (Pennington et al., 2000). It has a flat terrain with a strong climatic seasonality with summer maxima of up to 49 ºC (the highest in South America), severe winter frosts, and mean annual precipitation varying from 500 mm in central areas to 1000 mm in the eastern and western extremes (Minetti, 1999). Regional environmental gradients determine a variety of vegetation patterns including arboreal and savanna-like communities, with few endemic genera but numerous endemic species (Prado, 1993a; Prado, 1993b; Werneck, 2011).

Cerrado

The Cerrado (~2 million km^2) is distributed in the central Brazilian Plateau as a result of climatic, topographic, and edaphic interactions that have moulded vegetation communities since about 7,000 years before present (Ledru et al., 1998; Werneck, 2011). It borders the Amazon rainforest to the north, the Atlantic forest to the south and southeast, Caatinga to the northeast, and Chaco and Pantanal to the southwest (Olson et al., 2001). The Cerrado is characterised by an undulating topography with elevations ranging from 300 to 1800 m and mean annual precipitation of 900-1800 mm, with an average monthly maximum temperature ranging from 22 º to 27 ºC. It has a severe 3-5 months duration dry season extending from May to September. The Cerrado's vegetation communities are shaped by climate, fire regimes, water availability, soil fertility and topography (Jepson et al., 2010). The physiognomy of the Cerrado vegetation varies along these environmental factors, ranging from open grasslands (campos cerrado), shrubland or savanna (cerrado sensu stricto) with trees 2-8 m in height, woodlands or forests (cerradão) with trees 12-15 m in height, and gallery forests (Eiten, 1972; Klink and Machado, 2005). It is the richest tropical savanna biome in the world, with more than 7,000 vascular plant species, 44% of which are endemic (Klink and

Machado, 2005). For a review of Cerrado ecology and natural history see Oliveira and Marquis (2002).

Atlantic Forest

The Atlantic forests originally covered an area of about 1.3 to 1.5 million km^2 extending along the coast of Brazil up to 700 km inward in southeast Brazil (91% of total area) (Oliveira-Filho and Fontes, 2000; Olson et al., 2001; Ribeiro et al., 2009) also reaching northern Argentina (2% of total area) (Giraudo et al., 2003) and eastern Paraguay (7% of total area) (Huang et al., 2007). These forests are comprised of: a) the coastal rain forests found lengthwise the Brazilian coast, where rainfall is locally boosted by moisture-laden air masses blowing from the Atlantic oceanic and seaside mountain ranges (Câmara, 2003; Oliveira-Filho and Fontes, 2000); b) Araucaria mixed forests, subtropical rain forests distributed in south-eastern Brazil (Ledru et al., 1998) and c) semi-deciduous forests of the interior of Brazil, eastern Paraguay and north-eastern Argentina (Giraudo et al., 2003; Huang et al., 2009; Oliveira-Filho and Fontes, 2000). Rainfall varies between 4,000 mm/year at the coastal ranges of Brazil where tropical rain forests and Araucaria mixed forests are found to about 1,000 mm/year for the inland forests dominated by deciduous and semi-deciduous trees (Câmara, 2003).

Temperate Grasslands

Temperate grasslands of southern South America, also named Rio de la Plata Grasslands, cover 750,000 km^2 in plains of central-east Argentina, southern Brazil and Uruguay (Baldi and Paruelo, 2008). These landscapes are composed of different phytogeographic units dominated by grasses distributed according to geomorphology, soils, drainage and physiography in an area where annual precipitation varies from 1200 mm in the Northeast to 600 mm in the Southwest, and temperatures from 17 ºC in the North to 13 ºC in the South (Baldi et al., 2006).

Chilean Matorral

The Chilean Matorral is a Mediterranean-type ecosystem located in the western margin of South America between 30 ºS and 36 ºS in central Chile. To the northern extreme limits with the Atacama Desert and with temperate rainforests to the south comprising a variety of environments which define a highly heterogeneous vegetation mosaic (Armesto et al., 2007). Shrublands and succulents dominate in more xeric environments while evergreen sclerophyllous and winter deciduous forests occur in more humid southern environments of the ecoregion (see review of Moreira-Muñoz, 2011).

Tropical Dry Forests

Definitions of tropical dry forest ecosystems are variable (Sánchez-Azofeifa et al., 2005), making difficult to establish their distribution. Yet, some efforts have been made to evaluate the current distribution of this biome though areas change depending on the classification systems adopted. According to Pennington *et al*. (2000), tropical dry forests occur where rainfall is less than 1600 mm/year with a dry season of 5-6 months. The vegetation occurring under these climatic conditions has a similar structure and physiognomy often including thorny species, deciduous and semi deciduous trees of lower height and basal area compared to tropical rain forests (Miles et al., 2006; Pennington et al., 2000). Portillo-Quintero and Sánchez-Azofeifa (2010), who take into account a broad definition, delimit tropical dry forests based on bioclimatic and phenological characteristics using the geographical boundaries established by Olson et al. (2001), and estimating a total forest cover of about 664,000 km^2 with the largest areas in South America occurring in the Caatinga region of north-eastern Brazil, the Chiquitano dry forest of Bolivia and northern Venezuela and Colombia. Collectively, these ecosystems comprise more than 50% of the world tropical dry forest (Linares-Palomino and Ponce-Alvarez, 2009). Dry forests also occur in dry valleys of the Andes in Bolivia, Peru, Ecuador, and in coastal Ecuador and northern Peru (Pennington et al., 2000; Sánchez-Azofeifa and Portillo-Quintero, 2011).

Table A.1. Summary of reviewed studies of land use/cover change (LUCC) impacts on temperature, rainfall and albedo in non-Amazonian South America

Ecoregion	Reference	Data creation tool	Time span (months)	Land use/cover change		Temperature	Rainfall	Albedo	Key Findings
				From	To				
Dry Chaco	Beltrán-Przekurat et al. (2011)	Climate Model	5	Wooded grasslands	Soybean plantations	+	-	+	Decreased roughness length, rooting depth and leaf area index. Increased temperature, less rainfall, increased albedo.
	Beltrán-Przekurat et al. (2011)	Climate Model	5	Crops	Evergreen broadleaf tree plantations	-	+	No impact	Decreased temperature, no average changes in rainfall, less sensible heat, more latent heat, less albedo.
	Beltrán-Przekurat et al. (2011)	Climate Model	5	Evergreen broadleaf forest	Crops	+	-(+)	+	Increase in 2m temperature up to 0.6 1°C in wet and dry years. Less rainfall during wet years and more rainfall during dry years.
	Houspanossian et al. (2013)	Remote sensing and column radiation model	84	Dry forest	Crops	+	NE	+	Increased temperature and 50% more albedo in crops than forests. Mean diurnal temperature 2.5 °C greater in crops
	Loarie et al. (2011a)	Remote sensing	108	Shrubby vegetation	Crops	NE	NE	+	LUCC in Chaco responsible of 7% of albedo increases in South America.
	Canziani and Carbajal Benitez (2012)	Climate Model	480	Forest/Savanna	Crops	+	No impact	NE	No impact on precipitation, increased temperature in < 1°C. Temperature changes found in deforested areas and beyond.
	Lee and Berbery (2011)	Climate Model	3	Savanna, evergreen broadleaf forest	Rainfed Crops	-	-	+	Decreased sensible heat flux. Decreased precipitation where crops replaced forests and savanna. Opposite effect where grasslands and savanna where replaced by crops in southern La Plata basin.
Brazilian Cerrado	Costa and Pires (2010)	Climate Model	240	Forested	Deforested	NE	-	NE	Dry season increased from 5 to 6 months if deforestation of Amazon and Cerrado occurs.
	Georgescu et al. (2013)	Climate Model	60	Cerrado and crops	Sugarcane	+-	-	+	Cooling in growing season due to increased albedo and warming impact after harvest because of more sensible heating. Decreased total evapotranspiration and rainfall
	Lee et al. (2011)	Climate Model	324	Natural vegetation	Non Natural Vegetation	+	-	NE	Increased temperature, decreased rainfall during dry season. Vegetation change could be critical in explaining observed rainfall decreases. LUCC plus global warming favour more frequent and intensive droughts.
	Loarie et al. (2011b)	Remote sensing	36	Natural Vegetation	Crop/Pasture	+	NE	+	Increased temperature of 1.55°C. Less evapotranspiration. More albedo.
				Crop/Pasture	Sugarcane	-	NE	+	Mean Cooling of 0.93°C. More evapotranspiration. More albedo.
	Pongratz et al. (2006)	Land surface model and remote sensing	60	Broadleaf deciduous forest	Crops	+	NE	NE	Reduced conductance, roughness and increased canopy temperature in 0.7 °C at midday.
				Broadleaf deciduous forest	Pasture				Decreased canopy transpiration, increased aerodynamic resistance,

Ecoregion	Reference	Data creation tool	Time span (months)	Land use/cover change		Temperature	Rainfall	Albedo	Key Findings
									0.5°C increase in maximum temperature.
	Mendes et al. (2010)	Results from Malhi et al. (2009) and Costa and Pires (2010)	-	Forested	Deforested	NE	NE	NE	Deforestation of the Cerrado could increase savannization of southern Amazon.
Temperate Grasslands	Beltrán-Przekurat et al. (2011)	Climate Model	5	Tall grass	Wheat	-	- (+)	+	Increased roughness length, LAI[a] and rooting depth. Decreased Bowen Ratio in spring, increased in summer. More albedo. Less temperature (-1°C), less rainfall during El Nino, more rainfall during La Nina.
					Soybean	-	- (+)	+	Increased albedo and LAI. Less sensible and more latent heat flux, less temperature (~-1), less rainfall during El Niño (~5mm on average), more rainfall during La Niña (~6mm on average).
					Broadleaf forest	-	+	+	More albedo, more LAI, decreased Bowen Ratio, less temperature (-0.7) and more rainfall (2mm) in summer.
	Lee and Berbery (2011)	Climate Model	3	Grasslands	Rainfed Crops	+	+	+	Small decrease in albedo and surface roughness. Increased both latent and sensible heat fluxes. Slight Warming. Increased rainfall. Change in 10 m winds.
	Loarie et al. (2011a)	Remote sensing	108	Grasslands	Crops	NE	NE	+	16% of albedo increases in South America.
Chilean Matorral	Beltrán-Przekurat et al. (2011)	Climate Model	5	Wooded grasslands	Wheat	+	No impact	+	Decreased roughness length and rooting depth. Increased Bowen Ratio in spring, decreased in summer. Increased surface temperature.
	Montecinos et al. (2008)	Coupled mesoscale model	0.07	Natural sparse semi-arid vegetation	Irrigated crops	-	NE	-	More evapotranspiration, differences in surface temperature of about -2 ºC. More net radiation over cultivated areas due to decreased albedo.
Tropical Dry Forests-Chiquitano	Bounoua et al. (2004)	Land surface model	5	Forest	Crops	+	NE	+	Increased albedo (2%), increased aerodynamic resistance, less canopy conductance, mean warming of about 0.6 ºC. Increased diurnal temperature range.
				Wooded grasslands		+	NE	No Impact	Decreased canopy conductance, mean warming of 1.2 ºC.
Tropical Dry Forests-Caatinga	Sud and Fennessy (1982)	Climate Model	1.6	Savanna albedo	Desert albedo	NE	-	NE	20% reduction in cloudiness. Less rainfall (0.53 mm/day), less sensible and latent heat flux. Reduced total diabatic heating.
	Sud and Fennessy (1984)	Climate Model	1.6	Normal conditions by 1979	Evaporation supressed	NE	-	NE	Small reduction in cloudiness. Increased sensible heat flux, increased diabatic heating, less total precipitation (0.31 mm/day)
	Oyama and Nobre (2004)	Climate Model	12	Xeromorphic vegetation	Desert (bare soil)	+		+	Increased temperature, less rainfall, less evapotranspiration, less atmospheric moisture convergence due to subsidence anomalies.
	Castilho de Souza and Oyama (2011)	Climate Model	3.5	Shrublands	Desert (barren or sparsely vegetated)	+	-	+	Increased temperature (+2.4K), less rainfall, less evapotranspiration and runoff, increased sensible heat flux
	Hirota et al. (2011)	Climate Model	72	Caatinga	Desert (bare soil)	NE	-	NE	Negative precipitation anomalies not restricted to Caatinga but also

Ecoregion	Reference	Data creation tool	Time span (months)	Land use/cover change		Temperature	Rainfall	Albedo	Key Findings
									extending to the Amazon. Less mean annual precipitation (-3.91 mm/day)
Atlantic Forest	Webb et al. (2005)	Weather stations	360	Forested	Non-forested	NE	-	NE	Tree cover positively correlated to mean annual rain days

NE: Not Evaluated

Appendix B

Table B.1. Mean differences (CNTRL-NAT, 3 ensembles each for the period 1981-2010) of key variables in non-Amazonian South America. Values are expressed in original units as mean ± standard error. Soil moisture was converted to litre (m^3x1000) to facilitate representation. Means and standard errors of sensible and latent heat fluxes are shown in Table 4.2 of the main text.

Ecosystems/variables	Atlantic Forest		Cerrado		Dry Chaco		Chilean Matorral	
	Summer	Winter	Summer	Winter	Summer	Winter	Summer	Winter
Evapotranspiration (mm/year)	7.08±20.02	-18.22±17.11	23.93±16.97	-52.79±21.08	16.31±31.36	-14.16±15.94	-35.42±26.75	-54.08±10.01
Potential evaporation (w/m^2)	31.13±3.10	98.87±4.60	13.47±1.62	132.12±9.19	48.88±13.62	1.18±8.68	340.94±10.35	0.96±1.75
Relative humidity (pp*)	-5±0.47	-4.81±0.75	-1.35±0.27	-4.56±1.19	-3.68±1.09	-1.31±1.00	-1.29±0.72	-3.13±0.44
Soil temperature (°C)	1.35±0.13	0.80±0.17	-0.03±0.10	0.76±0.19	0.93±0.21	0.04±0.24	1.87±0.18	-0.22±0.17
Soil moisture (lt/lt)	-4.94±4.21	-14.17±5.10	-3.61±2.54	-20.27±5.39	-5.97±6.56	-2.06±7.78	-6.19±2.56	2.37±5.11
10m wind speed (m/s)	1.40±0.03	1.45±0.03	0.33±0.03	0.44±0.03	1.50±0.03	1.40±0.02	1.30±0.02	1.20±0.04
Net radiation (w/m²)	-1.29±1.44	-1.71±2.04	1.49±1.59	0.48±2.16	1.89±0.98	0.10±1.14	-3.99±1.49	-2.95±1.58
Ground storage (w/m²)	0.01±0.43	-0.07±0.46	-0.19±0.08	0.31±0.08	0.10±1.24	-0.05±0.3	0.29±1.21	-0.75±0.62
Bowen ratio (w/m^2)	-0.03±0.06	0.05±0.18	-0.01±0.03	0.16±0.13	-0.01±0.02	0.05±0.04	0.27±0.44	0.03±0.05
Plant respiration CO2 flux (gC/m^2/s)	1.2E-05±8.9E-07	8.8E-06±7.2E-07	-1.2E-05±1.2E-06	-9.4E-06±1.2E-06	-5.7E-06±7E-07	-1.7E-06±4.2E-07	-3.4E-07±8.4E-07	1.8E-08±3.3E-07

*pp=percentage points

Appendix C

Experimental design

In this study, we used the Conformal-Cubic Atmospheric Model (CCAM) coupled with the Community Atmosphere Biosphere Land Exchange (CABLE) surface model (Kowalczyk et al., 2006; McGregor, 1996; 2003; 2005a; 2005b; McGregor and Dix, 2008) to explore the influences of historic LUCC on the climate of southern South America. For modelling experiments, CCAM was driven by ERA-Interim reanalysis (Dee et al., 2011) in order to reduce the noise produced by climate oscillations while preventing possible model biases. Model domain was set between latitude 10-45S and longitude 30-90W at about 25 km horizontal resolution. We completed 3 ensembles for two sets of model simulations for the period 1982 and 2009 (28 years). The only difference between the simulations was the description of land surface datasets. The first scenario had present (CNTRL) land surface characteristics and the second simulation had pre-European surface characteristics (NAT). In the CNTRL scenario we upgraded the default land cover map in CCAM using the Collection 5 MODIS global land cover product for year 2005 (Friedl et al., 2010). In the NAT scenario, we modified land surface in four main ecosystems: 1) the Atlantic Forest, 2) Cerrado, 3) Dry Chaco and 4) Chilean Matorral using available literature and boundaries that approach the original extent of natural communities prior to major European LUCC (Olson et al., 2001). To extrapolate historic natural vegetation we first identified remnant natural vegetation in the MODIS image that agreed with descriptions from literature of natural vegetation types for each ecosystem (e.g. Savannas in the Cerrado or deciduous broadleaf forest in the Dry Chaco). We then projected historic natural vegetation by replacing current (e.g. crops) by natural vegetation types. Leaf area index (LAI) for the CNTRL scenario was based on that developed by Beijing National University (BNU) for the period 2000-2009 (Yuan et al., 2011). We inferred the LAI of natural vegetation by interpolating the BNU LAI of remaining natural vegetation in the MODIS image using a nearest neighbour rule. Present land surface characteristics were set for modern day conditions in all locations outside the focus ecosystems for all simulations.

Climate extremes and Aridity

For each simulation we extracted daily data of surface temperature and precipitation to calculate percentile-base seasonal (summer and winter) extreme indices (Donat et al., 2013). For temperature extremes we used warm days, cold days, warm spell duration, warm nights, cold nights and cold spell duration. For precipitation extremes we used heavy precipitation days and very heavy precipitation days (Table S1). We also tested the potential impacts of historic LUCC on the annual Aridity Index (P/PET) that relates precipitation (P) with potential evapotranspiration (PET) (UNEP, 1997).

Table S1 shows the extreme indices calculated. We use extreme indices that are based on percentile thresholds rather than absolute thresholds. Percentile-based indices are much less sensitive to climate model biases. Climate extreme indices are presented as follows: The extreme indices used in our analysis are shown in Table 1.

Name	Description	Units
Warm days (TX90p)	Count of days with daily maximum temperature > 90th percentile	day
Warm nights (TN90p)	Count of days with daily minimum temperature > 90th percentile	day
Cold days (TX10p)	Count of days with daily maximum temperature < 10th percentile	day
Warm spell duration (WSD)	Count of days with at least 4 consecutive days when maximum temperature > 90th percentile	day
Cold nights (TN10p)	Count of days with daily minimum temperature < 10th percentile	day
Cold spell duration (WSD)	Count of days with at least 4 consecutive days when minimum temperature < 10th percentile	day
Very wet day precipitation (R95pTOT)	Total precipitation on days where precipitation > 95th percentile	mm
Extremely wet day precipitation (R99pTOT)	Total precipitation on days where precipitation > 99th percentile	mm

Table C.1 Extremes indices used in this study.

Statistical Analysis

We analyzed the statistical difference between simulations using bootstrapping at 99% confidence level (*p* < 0.01) where $X = \{X_1, X_2, ..., X_{n_x}\}$ is the sample of a climate extreme or aridity index from the NAT experiment during period 1982-2009, and $Y = \{Y_1, Y_2, ..., Y_{n_y}\}$ the sample of the same variable taken from the CNTRL experiment in the same period. The null hypothesis (H_0) states that there is no significant difference between the means $\bar{X}$ and $\bar{Y}$ (i.e. LUCC has no significant impact on the selected climate extreme). The alternative hypothesis (H_1) states that there is a significant difference between the means $\bar{X}$ and $\bar{Y}$ (i.e. LUCC has a significant impact on the selected climate extreme). The observed *t*-statistic is $t_{obs} = \left((\bar{Y} - \bar{X}) / \sqrt{(\sigma_x^2 / n_x) + (\sigma_y^2 / n_y)}\right)$, where

$(\bar{X}, \sigma_x n_x)$ and $(\bar{Y}, \sigma_y, n_y)$ are the mean, standard deviation and sample size of the NAT and CNTRL samples, respectively. The bootstrap statistic for each sample is computed as $t^* = \left((\bar{Y}^* - \bar{X}^*) / \sqrt{(\sigma^2_{x^*} / n_{x^*}) + (\sigma^2_{y^*} / n_{y^*})}\right)$, where $(\bar{X}^*, \sigma^2_{x^*}, n_{x^*})$ and $(\bar{Y}^*, \sigma^2_{y^*}, n_{y^*})$ are the mean, standard deviation and sample size of randomly selected bootstrapped samples, respectively. The Achieved Significance Level (ASL) will be the proportion of samples where $|t^*| \geq |t_{obs}|$. The p value is calculated as $p = 1 - \text{ASL}$. The null hypothesis is rejected if $p < 0.01$, indicating a significant change in the climate variable across the two scenarios for the 28-year period. The sample sizes for X and Y were both 84 (3 ensembles over 28 years), with $N = 500$ bootstrap samples conducted to test for statistical significance.

Supplementary Results

Ecosystem	Summer							
	TX90p	TX10p	WSD	TN90p	TN10p	CSD	R95pTOT	R99pTOT
Atlantic Forest	**-9.95±1.93**	**1.26±0.82**	**-4.02±1.11**	**5.46±1.89**	**-1.61±0.87**	-0.26±0.17	0.37±0.85	-0.13±0.64
Cerrado	**-3.10±1.6**	**0.34±0.8**	**-1.02±0.81**	**1.61±2.39**	**-0.60±0.72**	-0.07±0.12	0.02±0.91	-0.15±0.69
Dry Chaco	**-7.58±1.97**	**1.10±0.74**	**-2.11±0.69**	**1.49±1.21**	**-0.47±0.69**	-0.07±0.11	0.22±1.29	-0.03±1.08
Chilean Matorral	**-17.53±1.55**	**2.89±0.64**	**-9.39±0.76**	**-1.93±1.24**	**-1.31±0.64**	-0.09±0.09	-0.04±0.22	-0.01±0.2
Ecosystem	**Winter**							
	TX90p	TX10p	WSD	TN90p	TN10p	CSD	R95pTOT	R99pTOT
Atlantic Forest	**-5.05±2.1**	**1.09±0.84**	**-2.39±0.99**	**1.63±1.42**	**-0.97±0.94**	-0.27±0.28	-0.17±0.48	-0.12±0.36
Cerrado	**3.63±2.82**	**-0.80±1.01**	**1.15±1.54**	**3.74±1.43**	**-2.29±1.04**	-0.44±0.35	-0.04±0.19	-0.02±0.15
Dry Chaco	**-2.60±1.57**	**1.06±1.12**	**-0.63±0.56**	**-0.57±1.42**	0.57±0.81	0.07±0.16	-0.12±0.2	-0.08±0.18
Chilean Matorral	**-3.51±1.58**	**1.70±0.95**	**-1.17±0.58**	**-2.61±1.55**	**-0.52±1.02**	-0.05±0.22	0.08±1.35	0.02±0.84

Table C.2 Averaged ± standard deviation seasonal trends in percentile-based extremes indices for the period 1982-2009 in four ecosystems of non-Amazonian South America. Bold indicates significant differences at 99% level. Temperature extremes are in days and are represented by: warm days (wmd); cold days (cld); warm spell duration (wsd); warm nights (wmn); cold nights (cln); and cold spell duration (csd). Precipitation-based indices are in millimetres and correspond to very wet day precipitation (vwdp) and extremely wet day precipitation (ewdp). For eastern South America (Atlantic Forest, Cerrado and the Dry Chaco) summer represents the wet season whereas for the Chilean represents the dry season. This pattern reverses in winter.

Aridity Index (P/PET)			
Ecosystem	**Summer**	**Winter**	**Annual**
Atlantic Forest	-2.3±0.32	-0.84±0.19	-1.55±0.16
Cerrado	-1.9±0.32	-0.11±0.03	-0.89±0.09
Dry Chaco	-0.3±0.16	-0.02±0.03	-0.10±0.07
Chilean Matorral	-0.07±0.02	-0.32±0.87	-0.49±0.06

Table C.3 Mean differences ± standard deviation in the Aridity Index (dimensionless) calculated from CCAM model output for period 1982-2009 in four ecosystems of non-Amazonian South America.

Supplementary Discussion

Ecosystem	Summer			Winter		
	H (w/m^2)	LH (w/m^2)	β	H (w/m^2)	LH (w/m^2)	β
Atlantic Forest	-2.09 ± 1.5	0.56 ± 1.6	-0.03 ± 0.01	-0.18 ± 1.2	-1.45 ± 1.4	+0.05 ± 0.18
Cerrado	-0.83 ± 0.6	1.91 ± 1.4	-0.01 ± 0.02	3.82 ± 1.3	-4.21 ± 1.7	+0.16 ± 0.13
Dry Chaco	-0.41 ± 2.4	1.30 ± 2.5	-0.01 ± 0.03	1.48 ± 1.1	-1.13 ± 1.3	+0.05 ± 0.04
Chilean Matorral	-1.83 ± 2.3	-2.82 ± 2.1	+0.27 ± 0.5	2.94 ± 0.7	-4.31 ± 0.8	+0.03 ± 0.05

Table C.4 Mean differences ± standard deviation in sensible heat flux (H), latent heat flux (LH) and Bowen ratio (β) taken from Chapter 4.

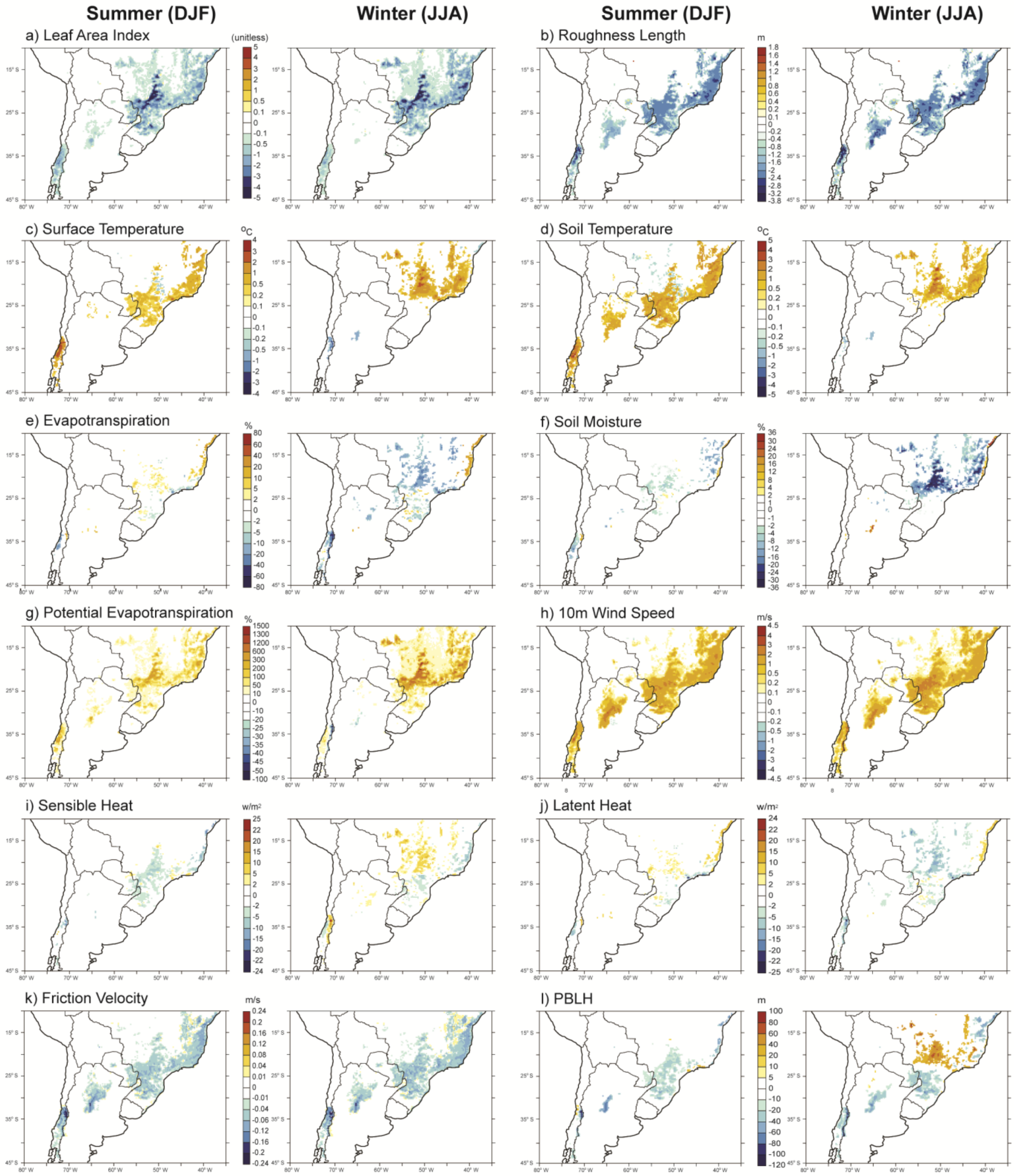

Figure C.1 Seasonal mean differences (CNTRL-NAT) during period 1982-2009 for key variables in non-Amazonian South America. With the exception of a) the leaf area index and b) roughness length, maps display only those areas with differences at 99% confidence level ($p < 0.01$). These variables are c) surface temperature (°C), d) top soil temperature, e) evapotranspiration (%), f) top level soil moisture (%), g) potential evapotranspiration (%), h) 10 m wind speed (m/s), i) sensible heat flux (w/m^2), j) latent heat flux (w/m^2), k) friction velocity (m/s) and l) planetary boundary depth (PBLH, m). Summer (DJF) corresponds to the dry season in the Chilean Matorral and the wet season in eastern South America (Atlantic Forest, Cerrado and Dry Chaco). This pattern changes in winter (JJA).