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CAN PLANTING TREES MAKE IT RAIN?
- THE IMPACT OF TREES ON PRECIPITATION THROUGH LAND-ATMOSPHERE FEEDBACK

A thesis submitted in fulfilment of the requirements for the degree of Doctor of Philosophy in the Faculty of Agriculture and Environment at The University of Sydney

Chun Xia Liang

August 2013
DECLARATION

The text in this thesis contains no material that has been accepted as part of the requirements for any other degree or diploma in any university, or any material published by another person without due reference being made to the material.

Chun Xia Liang
ABSTRACT

Land surface atmosphere feedback is the interaction between land and atmosphere through the water cycle and energy exchange. Observational studies on this topic are limited due to the constraints on data availability and uncontrolled experiments. Currently the main stream of research on this topic is the development of large numerical models. These models require high computational capacity and large number of parameter calibrations. Despite these efforts, reported results on feedback are inconclusive, partially due to the difference between models. This research therefore aims to investigate the vegetation-rainfall feedback relationship using relatively simple methods as these will allow more accurate tracking of the different relationships.

The first study is to detect the signal of land cover change from rainfall data in two regions in the Murray Darling Basin (MDB) with known land surface intervention. One study region is in the south of central Queensland (QLD) and the other one is on the border of New South Wales and Victoria (NSW/VIC). The study is based on a 30-year gridded precipitation dataset. The changes in tree cover are obtained from the MODIS product Global Vegetation Continuous Field dataset. A semi-parametric regression model and a regression model with non-parametric step trend test are applied to identify changes in rainfall data. The regression model shows that the climatic indices of SOI and IOD, seasonality and the land cover variable explain around 13% of rainfall variations in the QLD region and 19% in the NSW/VIC region. The step trend test compares rainfall before and after the land cover intervention, when the effects of large scale factors and seasonality have been removed by the regression model. Both methods have identified some rainfall changes in the Snowy Mountain...
ranges in NSW/VIC after the 2003 severe bushfires in this area. The results from the regression model show that the 2003 land cover change was significant in rainfall prediction for an area larger than the burned area. The step trend test estimates the precipitation change is significant in the Snowy Mountain. The rainfall changes in QLD as a result of the 2002 - 2004 land clearing appear to be less significant.

The second study focuses on the development of the boundary layer which is highly sensitive to the land surface process and directly influences the rainfall generation. A simple model, “CLASS”, is used to simulate day time boundary layer dynamics and atmospheric conditions. The model is applied to two study regions in southern NSW, Kyeamba (short grasses) and Tumbarumba (Eucalyptus forest), during 09 - 11 November, 2006. The model can simulate temperature and heat fluxes over the two locations well. Assuming afforestation in Kyeamba and deforestation in Tumbarumba, sensitivity experiments show that the development of convective boundary layer (CBL) mainly depends on the sensible heat flux and temperature. With a higher net radiation, forests can experience lower sensible heat and temperature to compensate for the energy demand in the latent heat. The sensitivity experiments do not find a preferential land cover type for the lifting condensation level (LCL). Leaf area index, vegetation fraction and minimum surface resistance are found to have a strong impact on the latent heat fluxes but their impacts on the sensible heat fluxes are relatively small. On the other hand, sensible heat flux and temperature are more sensitive to the change of albedo and in turn change the CBL height. Due to the competition among the different factors, the combined effect of the surface parameters is small on the CBL height. With higher soil moisture or advective moisture convergence, there is higher potential for cloud formation and hence convective precipitation. The model shows that even though
latent heat fluxes and humidity are usually higher over forest, higher convective rainfall might not occur since the boundary layer development is not significantly stronger than the grassland.

In the last study, rainfall sensitivity to land surface conditions is assessed using a simple equilibrium model. The model is developed from the box models in D’Andrea et al. (2006) and Baudena et al. (2008). More vegetation properties are implemented in the model to provide better coupling with the atmospheric conditions and allow vegetation changes. A Sobol’s sensitivity analysis is applied on the set of boundary conditions, initial conditions and the vegetation parameters. The boundary conditions, especially the lateral moisture fluxes, have the highest impact on rainfall, followed by initial soil moisture and initial LAI. Rainfall is not sensitive to the vegetation parameters, including the canopy interception fraction, the vegetation height and the maximum stomatal conductance. Afforestation cannot enhance long term summer precipitation in the MDB, unless there is a constant supply of high lateral moisture. The amount of lateral moisture needed for the afforestation to affect rainfall depends on the precipitation efficiency. When the lateral moisture supply can compensate for the potential higher loss of water after afforestation, the high ET can be sustained and a higher rainfall feedback is maintained. Furthermore, the system is most sensitive to low lateral moisture flux which explains the long lasting drought after a dry episode.

The current study could not find a strong relationship between the forest cover and precipitation. Loss of tree cover might decrease precipitation in some areas of the MDB but this relation is not significant, especially in areas with low rainfall. On the other hand, a high moisture convergence is required for afforestation to affect rainfall. Although the vegetation-precipitation relationship is inconclusive, land managers still
need to be careful of any action taken, given the possible unwanted outcomes under unfavourable conditions.
ACKNOWLEDGEMENTS

Firstly, I am indebted to my supervisor, Associate Professor Willem Vervoort (Hydrology and Catchment Management, Faculty of Agriculture and Environment, The University of Sydney). Willem has given me enormous ideas and advice on the research topic. Besides offering guidance on the thesis work, he has always been supportive and encouraging, which continuously helped me to build up my confidence.

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I am very grateful to Ryan Teuling (Assistant Professor in Wageningen UR, the Netherlands) for his support during my visiting study in Wageningen and for his inspiration on the work of Chapter 4. I would also like to thank Dr. Jordi Vilà-Guerau de Arellano (Wageningen UR) for the provision of CLASS model and the explanation to help me understand the model.

I would like to acknowledge Professor Andy Pitman for useful discussion on the direction of the research at the very early stage of my PhD; Dr. Eva van Gorsel (CSIRO Marine and Atmospheric Research team) for the provision of the flux data in Tumbarumba; and Dr. Edith Lees for spending hours helping me to understand stomatal conductance.

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Finally I sincerely thank my in-law family. Without their help, especially in the last two months, I would not be able to go through the deep blue after giving birth to my baby Emily in March and finish this thesis.
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Chapter 1

General Introduction

1.1 Prologue

On the 4th June, 2006, a rally took place in Melbourne to protect Victoria’s old-growth forests and water catchments for fresh air, clean water and the threatened wildlife. About 15,000 people participated in the march from the State Library to Federation Square, Melbourne. The rally was challenged by the Forestry and Conservation Minister Eric Abetz. The Minister claimed that the scale of the timber harvesting in Australia was insignificant compared to such activities in Indonesia. (source: The Wilderness Society, [http://www.wilderness.org.au/regions/victoria/VFA-report](http://www.wilderness.org.au/regions/victoria/VFA-report), viewed on 10/07/2012)

From 23 November 2009 to 13 January 2010, Peter Spencer who is a NSW grazier has gone on a hunger strike for 52 days to protest against land-clearing laws. The widespread anger among farmers in this event was caused by a move of the NSW state government to a stricter native vegetation conservation act. The new act was to ban land clearing on private land in NSW. Farmers in NSW asked for compensation to be paid to cover their potential loss in production under the new law. (The Australian, 13 January 2010)
Later, on 20 March, 2011, more than 1,000 protesters gathered at Batman’s Bridge in Tasmania against a controversial pulp mill project. The protesters pled to stop the pulp mill fearing the loss of 570,000 ha of native forest. At the same time, a few smaller anti-mill rallies also took place around the state. However the Tasmania Liberal Senator Richard Colbeck and the forest industry argued that the mill could maintain a sustainable harvest. (source: Wikipedia, http://en.wikipedia.org/wiki/Bell_Bay_Pulp_Mill, viewed on 09/07/2012)

Recently, Federal Senator Barnaby Joyce, who had supported Peter Spencer’s hunger strike in 2010, repeatedly accused the Commonwealth and the state governments’ position in land clearing regulation. He claimed “the pendulum of public debate has swung too far towards the environment and away from social and economic considerations”. (The Sydney Morning Herald, 19 April 2012)

These are just a few of the incidents highlighting the conflicts between land clearing and anti-land clearing campaigners. For farmers like Peter Spencer and politicians like Senator Joyce, the matter might be about freedom, economics and politics. For many others who took part in the Melbourne and/or Tasmania protests, their concern was about the sustainability of the future environment. The debate can keep on going, but will only be guided in the right direction when the environmental value of the resources is better understood.

Besides regulations and laws, adequate knowledge on the potential outcome of land disturbance is important. The relationships between land use and land cover change (LULCC) and the local climate in both the short and the long term might be of great interest. Good scientific information, such as high quality measurement and unibias analysis, are urgently required for the public and the policy makers to deal
with environmental problems and to sustain a healthy environment for our future generations.

1.2 Motivation

1.2.1 Agricultural land clearing

Land clearing activities have a long history in Australia. Since the start of European settlement 200 years ago, native land including forests, savannah, woodlands and native grasslands have been cleared to make way for residential, agricultural and other industry uses. According to figures released by Australian Native Vegetation Assessment (Cofinas and Creighton, 2001), 21.2% of the pre-European extent of forest and woodland has been cleared by 2001 (table 1.1). Total cleared area was nearly 1 million km\(^2\), accounting for 13% of the pre-European native vegetation extent.

Table 1.1: Pre-European area (km\(^2\)) of selected vegetation groups and total area cleared by 2001. Percent cleared is calculated as the percentage (%) of pre-European area.

<table>
<thead>
<tr>
<th>Vegetation Group</th>
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<th>Area cleared (km(^2))</th>
<th>Percent cleared (%)</th>
</tr>
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<tr>
<td>Forest &amp; Woodland</td>
<td>3,325,491</td>
<td>705,865</td>
<td>21.2</td>
</tr>
<tr>
<td>Shrub lands</td>
<td>1,185,824</td>
<td>173,429</td>
<td>14.5</td>
</tr>
<tr>
<td>Grassland</td>
<td>2,446,678</td>
<td>63,053</td>
<td>2.6</td>
</tr>
</tbody>
</table>

Australia claims a position in the top 10 of high land clearing nations in 2001 (Australian Conservative Fundation, 2001; of the Environment Committee, 2001). Although broadscale land clearing took place during the early settlement, high rates of clearing continued to happen after the Second World War (WWII). According to the Australian State of the Environment 1996 report, “as much land was cleared...
over the last 50 years as over the previous 150 years before 1945”. In 1990 alone, approximately 6,500 km² of forest was cleared (this figure does not include illegal logging and clearing) \cite{Macintosh2007}, which is one percent of the total forests cleared in more than 200 years.

Queensland has the highest rates of land clearing within Australia. The annual clearing rate of bushland was 4,500 km² during 1995 - 2005 in Queensland \cite{WildernessSocietyInc2011}. The Australian State of the Environment 2001 report stated that, according to the Australian Greenhouse Office estimation, among the 4,690 km² of woody vegetation cleared nationally in 1999, Queensland and New South Wales cleared the largest areas which were 4,250 km² and 300 km² respectively. Hence, Queensland removed almost a 15-fold of the area of vegetation compared to New South Wales. The absolute figure of Queensland is striking. On the other hand, the other states have proportionally cleared more land than Queensland. For example, 60% of previous vegetated land has been cleared in Victoria compared to only 18% in Queensland \cite{CofinasandCreighton2001}.

As a result of enforcement of land clearing laws and rising awareness of the environmental problem of forest loss, land clearing has decreased since 1990. According to the figure released by the National Inventory Report 2008 \cite{DepartmentofClimateChangeandEnergyEfficiency2008}, the national annual rate of forest conversion and reclearing sharply decreased from 5,607 km² to 3,169 km² in the first half of 1990s, and further decreased to 2,165 km² by 2008. However, during the years in between, there was a temporary rise of clearing rate, from 2,975 km² in 2003 to 4,067 km² in 2004. The sudden increase reflected a change of policy. Interestingly, however, it was not the result of relaxing land clearing laws but it was due
to landowners’ panic over a new law that would be more restrictive. In the future, the land clearing rate can still change, mainly subject to economic drivers and government policies.

1.2.2 Fires

The other main cause of vegetation loss is fires. Being a dry continent, Australia has frequent bushfires. Many of the Australian ecosystems have adapted to such climate and environmental conditions or even promote fires as a way to reduce competition. In fact, according to CSIRO bushfire research, bushfires were part of the Australian history and helped to shape the continental landscape (Bushfires, Commonwealth Scientific and Industrial Research Organisation (CSIRO), 2003-2012).

Throughout Australian history, the locals learned to live with fires and benefited from fires. However, as population grew and changes were made to the environment such as agriculture and urbanisation, fires have become less frequent but become more severe when they occur (Bushfire in Australia, Commonwealth Scientific and Industrial Research Organisation (CSIRO), 2003-2012). Major bushfires occurred more often in the last decade than in the previous 10 to 30 years in Victoria (Bushfire history - Major bushfires in Victoria, Department of Sustainability and Environment, Victoria) and southern Australia (Bartlett et al., 2007). These large fires have claimed lives, damaged important construction and destroyed some of the rare bioecological systems.

Although bushfire contributes to today’s landscape of Australia, it is one of the causes making Australia a dry continent. Different from the current environment which is dominated by grassland, Eucalyptus woodlands or savannas, Australia was once wet and covered by rainforest 60 million years ago. Studies have found a close
relationship between fires and the distribution of woodlands/savannas and rainforest (Hopkins et al., 1993, Fensham et al., 2003, e.g.). It is possible that fires have created more space, by removing rainforest, for drought-resistant species such as Eucalyptus. The surviving species further encouraged fires. Over a long history of interaction between fires and vegetation, the local hydroecological system has eventually changed and likely the climate has been affected too. The recent severe bushfires were partially due to the long drought conditions since the beginning of this century (Taylor and Webb, 2005).

### 1.2.3 Climate change

Climate change is one of the major current issues, both globally and locally, in almost every aspect of our life. The term “climate change” is defined by the Intergovernmental Panel on Climate Change (IPCC) as:

“... a change in the state of the climate that can be identified (e.g. 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. It refers to any change in climate over time, whether due to natural variability or as a result of human activity.” (IPCC, 4th Assessment Report, 2007)

As the leading international body on climate change assessment, IPCC has confirmed the existence of global warming and related it to human activities in the 2007 report. In Australia, climate changes are shown by increasing impact of highly variable weather conditions, droughts and floods (Holper, 2011). In most parts of the country, temperature indices including minimum, maximum and mean temperatures
have been increasing at around 0.2°C per decade in the last century (Hughes, 2003; Alexander et al. 2007; Alexander and Arblaster 2009). The annual or seasonal rainfall has also declined in many areas, together with a decrease in rainfall frequency and intensity, based on rainfall records from the last century (Gallant et al., 2007). In Australia, warming is a widespread phenomenon, and while total rainfall has not changed much since 1990, rainfall patterns have changed significantly (Department of Climate Change and Energy Efficiency, 2012).

The 2007 IPCC report attributed most of the temperature increase to the higher greenhouse gas concentrations which were caused by human activities. Land cover change was also identified as one of the drivers of climate change, along with greenhouse gas emissions. Some recent research suggest that land use and land cover change should bear at least equal or more responsibility for the warming (e.g. McAlpine et al., 2007; van der Werf et al., 2009; Zhang et al., 2009a; Schmidt, 2010; Stott et al., 2010; Mahmood et al., 2010). One consequence of land use change is that it can alter the surface energy balance and heat distribution, demonstrating by the heat island effect of urbanisation as an example (Mahmood et al., 2010; Schmidt, 2010). Additionally, forest is an important carbon sink but deforestation causes the release of carbon dioxide (CO2) into the atmosphere (van der Werf et al., 2009). Zhang et al. (2009a) also pointed out that vegetation cover change can possibly affect El Niño-Southern Oscillation (ENSO) impacts.

It is important to know the direction of future climate change in order for the society to adapt to the change, or to mitigate its impact. The latest IPCC report (Climate Change 2013) projected that global mean temperatures will be likely 1.5 - 2 °C above the 1986 - 2005 level in 2081 - 2100 if greenhouse gas emissions continue
unabated. Extreme weather events, such as heat waves, floods and drought, are likely to occur more often and to intensify in strength. In order to slow down climate change, many countries are making an effort to reduce greenhouse gas emissions, including governmental and/or intergovernmental action and non-governmental approaches. Some of the well known strategies are renewable energy and energy saving, sustainable farming techniques etc. Among many other mitigation policies and sustainable development practices, reforestation and afforestation are suggested as a potential effective method to mitigate global warming (Bonan, 2008a; Nijnik et al., 2009; Lasco et al., 2010). The effectiveness of these strategies depends highly on the scientific understanding of the relationship between climate change and our human activities, such as land cover change. However, currently this understanding is at a low to medium level (Lawrence and Chase, 2010).

1.3 Research interest

As mentioned above, land surface has changed significantly in Australia and climate change is happening. At the continental scale, loss of natural vegetation cover might have contributed to the warming temperature. However the feedback relationship between vegetation cover and climate, in particular rainfall, is not clear. Management can be improved to minimise land practice impact if the impact of clearing on the catchment water cycle is better understood. Further development of the relevant information and knowledge is required to assist policy makers, land and water managers and other stakeholders.

The research reported in this PhD thesis focuses on the feedback of local rainfall to vegetation cover change in the east and southeast Australia. Our hypothesis is “the loss of forest cover will reduce local precipitation; while on the other hand, increase in
forest cover can reverse the process”. This project therefore aims to contribute to the understanding of land-atmosphere feedback, especially on the impact of a change in vegetation cover on local precipitation. In the next chapter, we will review the literature on the vegetation cover-precipitation feedback in different regions worldwide, in order to study what has been understood and identify thesis questions. Then in the following three chapters, observational studies and model experiments are conducted in areas in the Murray Darling Basin (MDB) to study the relationship between the vegetation cover changes and rainfall variations. The last chapter discusses and concludes general results and suggests future work.
Chapter 2

Literature Review

Climate and vegetation are closely related. A change in the climatic conditions, such as temperature increases and/or rainfall decreases, can have significant impact on vegetation growth (Barger et al., 2009; Giese et al., 2009; Lejju, 2009; Opdam et al., 2009; Ahmad et al., 2010; Yu et al., 2010; Salazar and Nobre, 2010). On the other hand, a change in land cover is expected to feed back to the climate. Land cover change has been identified by the IPCC report (2007) as one of the causes leading to global temperature increases. The effect of land cover change on the local climate, including precipitation, cloud development and heat exchange, has been, and is still, studied by many researchers (e.g. Chagnon et al., 2004; Lim et al., 2005; Seneviratne et al., 2006; Zhang et al., 2009b; Teuling et al., 2010; Ma et al., 2011; Garcia-Carreras and Parker, 2011). The interaction between climate and vegetation is a two-way relationship and is referred to as feedback (Seneviratne et al., 2010). In this research, the effect of land cover on local precipitation is the particular interest. The feedback relationship is also important when considering the long term effect of dynamic vegetation change.

The influence of climatic conditions on vegetation might be relatively straightforward; however the climate feedback to vegetation changes is not that obvious. In an early review of regional LULCC impacts on rainfall, Pielke et al.
documented a few observational studies indicating that afforestation activities in the tropics could lead to enhanced convective precipitation. However such a positive feedback is difficult to find in the midlatitude region due to the different precipitation types and the various air moisture sources. On the other hand, model simulations of LULCC impact on precipitation also depend on geographical location, regional atmospheric characteristics and spatial scale of land cover change. Recently \textcite{Pielke2011} again reviewed the changes of local, regional and global climate, including surface fluxes, surface and boundary layer dynamics, that due to LULCC from both observational evidence and modelling analysis. In their summary, they highlighted from the observational studies that “the temperature response to LULCC is multidirectional” while the spatial scale is important in the modelling experiments. They also pointed out that the current studies could not conclude that LULCC has a major impact on atmospheric conditions although strong evidence of LULCC effect were shown at local or regional scale. Reviewing the impacts of LULCC on climate, \textcite{Mahmood2010} pointed out that in observational studies using rain gauge data a spurious change is possible to be detected. For example, an increase of trees cover can reduce wind speed and hence decrease bias in rainfall measurements. Overall, the impact of LULCC on precipitation is recognised but the feedback mechanism is a more complicated problem and needs to be investigated case by case.

The purpose of this review is to provide a general overview of global studies on the impact of vegetation cover change on local or regional precipitation in the last 23 years (from 1990 to 2012)\footnote{Some cases studies in the early 1990s’ are closely related to this topic. They are included in this review.}. Building on the earlier reviews by \textcite{Pielke2007}, \textcite{Mahmood2010} and \textcite{Pielke2011}, this review focuses specifically on studies investigating precipitation feedback to vegetation changes and exploring the possible
explanations for the heterogeneity in their results. Hence it has a more specific focus on precipitation than the previous reviews. Different from the previous reviews, data and methods used in the studies are also reviewed and compared in order to provide a reference for future studies. Relevant evidence is extracted from observational studies and modelling experiments in the literature. Details of the literature review methodology are given in Appendix A.

Change in vegetation cover has a relatively strict definition in this review. Vegetation conversion related to trees, such as from native forest or woodland to plantation or to pasture and cropland (including irrigation), is essential in this review. Both deforestation and afforestation are considered. The cause of land cover change is not limited to human induced but includes natural causes, such as wildfires.

Precipitation changes caused by LULCC are the main interest. Cloud feedback is also included in this review, since cloud formation is a necessary precursor for precipitation and the amount of cloud determines the amount of precipitation. The specific questions addressed in the review are: (1) What is the observed/modelled change in precipitation due to LULCC (the feedback)? (2) What determines this local or regional feedback? Is it the type of land conversion, or the local climatic characteristics, or the scale of deforestation? (3) What are the important physical processes determining the feedback?

2.1 Trees and rains - Observational evidence

The observational studies spread widely around the world. Except Antarctica where land is almost all covered by an ice sheet, the vegetation-precipitation studies are conducted in all continents. Nevertheless, there is a bias towards certain geographical regions under study (see the pie charts in Figure 2.1). Asia and South America are
Figure 2.1: Distribution of observational studies in this review by continents (left) and by climate zone (right). Some of the studies cover multiple continents and/or multiple climate zones.

The two continents attracting most research attention. Within the 70 observational studies reviewed, 23 studies are based on areas in Asia, including five studies in the Middle East. There is a special focus on China and India. Increasing agricultural activities and urbanisation are expected to change the pattern and the strength of Indian monsoon (Wang et al., 2009, Douglas et al., 2009, Kishtawal et al., 2010). Similarly, China has undergone rapid economic and social development over the last 30 years or so. Agriculture and urban expansion continuously demand land resources which were previously occupied by natural vegetation. Under the pressure of severe land degradation after deforestation, a reforestation and afforestation campaign was run by the Chinese government in the late 1990s. Complicated outcomes of these policies have been reported by Cao (2008a), Cao et al. (2010), and FAO (2010). Twenty-two studies are conducted inside South America (e.g. Lin et al., 2006, Rollenbeck and Anhuf, 2007, Nunez et al., 2008, Wang et al., 2009, Molina et al., 2012). Furthermore, almost 90% of these South American studies are in countries and regions in the Amazon Basin, mostly in Brazil, while the rest are in areas either close to or influenced by the Amazon Basin. The Amazonian tropical rainforest has important ecosystem
values locally and globally. Deforestation in this region could lead to significant regional and global climate changes (Durieux et al., 2003; Chagnon and Bras, 2005).

There are only two relevant studies in Western Australia (Ray et al., 2001; Junkermann et al., 2009). Although land clearing has removed extensive natural forest cover in Australia (Cofinas and Creighton, 2001; Pielke et al., 2011), research on the climate impact of forest removal is still relatively limited.

Figure 2.2: The distribution of observational studies (blue bars) and modelling studies (red bars) on the precipitation feedback to vegetation cover change between 1990 and 2012 (up to June 2012). The number of publications generally increases over the two decades.

The shift in interest is also related to the time of the studies. In the early 1990s, three interesting articles addressed the rainfall change after land use change in Southern Israel. Southern Israel has improved its vegetation cover in Negev since conservation policies were implemented by the Israeli government in 1948, followed by the largest afforestation project in Beer Sheva in the central Negev in the early 1960s
As the purpose of the conservation policies was to reduce the increased aridity in this region (Otterman et al., 1990; Bengai et al., 1994), the effect of land use change on climate, especially rainfall became an important topic for Southern Israel. There are also quite a few studies in Africa in the 1990s, mostly in the Sahel region which is to the south of the Sahara desert. They are interested at the possible further desertification after the severe late 20th-century Sahel drought (Zeng, 2003; Dai et al., 2004; Lu and Delworth, 2005). There had been little research on investigating land cover-precipitation interaction from observational data in the United States and China until the 21st-century. In the U.S., the study areas are usually in the central cropping regions, such as the “corn belt” (Carleton et al., 2001; Matyas and Carleton, 2010) and the Great Plains (DeAngelis et al., 2010) where intensive irrigation has occurred. In terms of China, the rapid economic growth and rise in political attention in the last decade might explain the increased number of studies. Research on the Amazon region has steadily increased over the last two decades due to its important impact on the global climate change.

Deforestation appears to be the most common land intervention in the history around the world. The development of human society is associated with our modification of the environment. Agricultural activities are carried out on fertile land which had previously been covered by natural vegetations. Large-scale conversions of forest or other types of natural vegetation covers to croplands and pastures are common in Southeast Asia, east and west Australia and the Amazon in South America (Ray et al., 2001; Falk et al., 2005; Kirby et al., 2006; Department of Environment and Resource Management, 2010). Urban expansion also acquires lands from natural environment and agriculture, such as in Europe and the U.S. (Brown et al., 2005; Price...
There are 23 deforestation studies. In addition, within the 20 studies comparing land-atmosphere interaction between different land cover types, at least five articles involve study areas that at some point suffering from deforestation. On the other hand, there are only seven studies related to afforestation/reforestation. The regions which afforestation/reforestation has occurred are mainly in Asia, including south Israel and China.

2.1.1 Types of data in the studies

Generally the reviewed observational studies perform statistical analysis on climate data for an area of interest. The linkage between precipitation and land cover (change) has been established in the following way depending on the availability of land cover information:

1. detailed land surface data is available and it is combined with the climate data. This includes time series information of land cover change and/or spatial distribution of different land cover types (e.g. [Durieux et al., 2003; Diem and Mote, 2005; Kaufmann et al., 2007; Molina et al., 2012]).

2. no detailed land cover data is available but specific time or location of land cover change is known (e.g. [Otterman et al., 1990; Ray et al., 2001; Dubreuil et al., 2012]).

3. Change of local climate is attributed to the historical land cover change for which no specific information is given (e.g. [Kanae et al., 2001; Altmann et al., 2002]).

Detailed land cover data is used in 28 studies. Remote sensing data, including Landsat and Moderate Resolution Imaging Spectroradiometer (MODIS) are commonly applied to identify the location of vegetation cover or to classify different...
Chapter 2. Literature Review

land cover. A simple application of Landsat data is to visually differentiate the land cover based on the false colour in the images, such as in \textit{Lawton et al.} (2001). In the study of \textit{Fuller and Ottke} (2002), the authors overlayed categorical land cover data and continuous percent tree cover data, both from the NOAA Advanced Very High Resolution Radiometer (AVHRR) 1992-1993 Landsat data, to estimate the percentage of vegetation cover in each of the study sites. Although MODIS data series have lower spatial resolution, they are also popular in the land-atmosphere feedback studies. Some MODIS products, such as the tree cover product, are adopted by several studies (e.g. \textit{Chagnon et al.} 2004; \textit{Chagnon and Bras}, 2005; \textit{Spracklen et al.}, 2012). The Normalized Difference Vegetation Index (NDVI) calculated from MODIS data is most widely used to estimate vegetation changes and distribution (e.g. \textit{Zhang et al.} 2005; \textit{Sato et al.}, 2007; \textit{Sarkar et al.}, 2007; \textit{Sun et al.}, 2008; \textit{Mendez-Barroso and Vivoni}, 2010). Land cover spatial products from the literature or other agencies are also used in some studies. For example, \textit{Carleton et al.} (2001) compared albedo and precipitation from three types of land surface in the Midwest Corn Belt, U.S.: cropland, forest and boundary of cropland and forest. Their selection of the study areas is based on a land use/land cover map produced by \textit{Copeland et al.} (1996) from the U.S. Geological Survey (USGS) Landcover Database and potential vegetation from model results. Some studies use map products that were derived from remote sensing data (e.g. \textit{Negri et al.}, 2004; \textit{Lim et al.}, 2005). Aerial photograph is another source of land cover information (\textit{Molina et al.}, 2012). In most cases, land cover data only provides categorical information for the study. Such as in \textit{Fuller and Ottke} (2002), even when time series of land cover data is adopted, the continuous percentage tree cover is integrated to generate a mean value and is used in the analysis. The main
problem with land cover data is its continuity and temporal scale. It is expensive to conduct regular site surveys of the land surface types. Many land cover datasets are built on combined information of past surveys and model results, and are updated annually or multi-annually. Remote sensing technology largely reduces the cost of field measurement. However, errors could occur due to cloud cover and classification methods. Land surface dynamics at smaller time steps, such as hourly or daily, are useful in studying convection activities but again it is expensive to monitor.

Table 2.1: Commonly used remote sensing data.

<table>
<thead>
<tr>
<th>Remote sensing products</th>
<th>Source</th>
<th>Resolution</th>
</tr>
</thead>
<tbody>
<tr>
<td>NDVI</td>
<td>MODIS-Terra</td>
<td>16-day 250m - 1°</td>
</tr>
<tr>
<td>Land surface temperature</td>
<td>MODIS-Terra</td>
<td>8-day 1km</td>
</tr>
<tr>
<td>Albedo</td>
<td>MODIS-Terra &amp; Aqua</td>
<td>16-day 1km</td>
</tr>
<tr>
<td>Precipitation</td>
<td>TRMM</td>
<td>3-hour 0.25°× 0.25°</td>
</tr>
<tr>
<td>Land cover (incl. tree cover)</td>
<td>AVHRR Landsat</td>
<td>16-day 15m - 60m</td>
</tr>
<tr>
<td>Cloud field</td>
<td>Landsat</td>
<td>16-day 15m - 60m</td>
</tr>
</tbody>
</table>

There is a wide range of precipitation data sources. Long record weather station data are most useful. More than half (39 out of 70) of the observational studies are based on precipitation and other climate data from station measurements, mostly provided by the local weather service (e.g. Jauregui, 1991; Seleshi and Demaree, 1995; Sadhukhan et al., 2000; Altmann et al., 2002; Conradt et al., 2007; Lu et al., 2009). Many studies use precipitation data from multiple stations to ensure the representativeness of the region. Single rain gauge data, such as in Lohar and Pal (1995), Altmann et al. (2002) and Sun et al. (2008), might only provide limited information. In some situations, weather stations have not been established in some
locations hence a field experiment has to be carried out in the particular study site (e.g. Eltahir and Bras, 1996; Eltahir and Humphries, 1998; Rollenbeck and Anhuf, 2007). Usually, this kind of field experiments cannot represent the whole region but only the local site. Global climate network data which might have higher credibility but lower density has also been adopted by several studies (e.g. Durieux et al., 2003; Sarkar et al., 2007; Hossain, 2010). Mahmood et al. (2010) pointed out that there is a wide range of uncaught bias in precipitation gauge data and the impact of LULCC on this bias is unclear. Satellite-derived precipitation estimates based on the Tropical Rainfall Measuring Mission (TRMM) are popular in the studies after 2000 (e.g. Negri et al., 2004; Fisch et al., 2004; Chagnon and Bras, 2005; Lin et al., 2006; Spracklen et al., 2012). With the launch of this dataset, more information about precipitation, including intensity and distribution, together with detailed cloud information has become available. Some studies, such as Negri et al. (2004), Xu et al. (2007) and Kishtawal et al. (2010), combined station data and remote sensing data to provide a more comprehensive analysis. However this type of precipitation information is limited by its relatively coarse resolution compared to rain gauge data and weakness in detecting light rains (Ebert et al., 2007).

Long time series data is also important in land cover-climate interaction research, as the long term effect of land use land cover change is the main interest here. Without a sufficiently long historical data set, it is difficult to conclude that the observed changes in precipitation or temperature are “real”. In other word, it is unclear whether the changes are due to random variability or it would lead to a new equilibrium. Among the observational studies, some have managed to use long time series climate data, as shown in Table 2.2. On the other hand, short time observations are also often used,
especially in the case of cloud studies or studies comparing different land covers. For example, in Cutrim et al. (1995), only one cloud image of August 1988 was used to investigate cloud development over native and agricultural landscape. Since boundary layer change and cloud development can be very sensitive to land surface condition in a short time period, long time series data was not necessary in their study. However, single or individual event like the one in Cutrim et al. (1995) would be too limited to draw a firm conclusion.

Table 2.2: Studies that use long period of rainfall data. % is the percentage of the studies in the reviewed observational studies.

<table>
<thead>
<tr>
<th>Data length</th>
<th>%</th>
<th>Studies</th>
</tr>
</thead>
<tbody>
<tr>
<td>&gt;100 years</td>
<td>3</td>
<td>Schlunzen et al. (2010); DeAngelis et al. (2010)</td>
</tr>
<tr>
<td>≥50 years</td>
<td>13</td>
<td>van Rompaey (1995); Seleshi and Demaree (1995); Sadhukhan et al. (2000); Chagnon et al. (2004); Chagnon and Bras (2005); Diem and Mote (2005); Conradt et al. (2007); Lu et al. (2009); Hossain (2010)</td>
</tr>
<tr>
<td>≥40 years</td>
<td>10</td>
<td>Otterman et al. (1990); Bengai et al. (1994); Sanga-Ngoie and Fukuyama (1996); Hogg et al. (2000); Kanae et al. (2001); Nunez et al. (2008); Makarieva et al. (2009)</td>
</tr>
<tr>
<td>≥30 years</td>
<td>6</td>
<td>Bengai et al. (1993); Webb et al. (2005); Sun et al. (2008); Butt et al. (2011)</td>
</tr>
<tr>
<td>≥10 years</td>
<td>3</td>
<td>Chu et al. (1994); Sarkar et al. (2007)</td>
</tr>
</tbody>
</table>

Overall, multiple sources of data have been used in the reviewed studies. There is a large amount of precipitation data available but in general land surface information is limited. With the expanding of observational network, such as satellite and aircraft, and the availability of reanalysis data, more information on land surface type and land surface process can become accessible to future research. Apart from availability, standardisation is another data issue. The collection of precipitation data might
be varied from country to country, depending on equipments used and handling techniques. An international data hub is desired to provide standardised high quality data for studies.

2.1.2 Methods applied in the studies

The pattern of the precipitation data is often investigated by statistical methods. The methods used are one of the main differences between various studies. There are two major types of analysis:

1. trend analysis - the change of precipitation over time can be detected by this analysis. Popular methods include regression analysis and Mann-Kendall rank test. In this case long time series data is more important.

2. significance analysis - usually this analysis is applied to identify difference between study sites or study periods.

Linear regression analysis involves a least square fitting to time series data (e.g. Bengai et al. 1993, Almann et al. 2002). Non-parametric tests, like the Mann-Kendall test, has the advantage of being robust against outliers and free from assumptions about the data distribution (Chu et al. 1994, Depaiva and Clarke 1995, Conradt et al. 2007, Schlunzen et al. 2010). It is regarded as useful in precipitation analysis since rainfall distribution is generally skewed. Depaiva and Clarke (1995) used both linear regression and Mann-Kendall statistics to examine the monthly precipitation trend from 15 years station records in the Amazon region. In their study, the results from the two methods only differ slightly. Kanae et al. (2001) also applied both methods on more than 40 years of precipitation data in Thailand. But in their study the linear regression analysis was used as a supplemental method to the Mann-
Kendall rank test. On average the Mann-Kendall test had detected a significant trend at seven sites. The regression analysis, however, could show significant changes at an additional 15 to 25 sites. Mann-Kendall test is preferred over the linear regression method as it is less sensitive to data variation. But this test has difficulty with large autocorrelations so additional treatment might be required (Schlunzen et al., 2010). Furthermore, the Mann-Kendall test is suitable for long-term changes rather than short-term fluctuations (Sadhukhan et al., 2000).

The two sample Student’s t-test is commonly used as a significance test. It is often used in situations where before- and after-change periods are compared (DeAngelis et al., 2010) and values between two different locations are compared (Durieux et al., 2003; Negri et al., 2004; Chagnon and Bras, 2005; Schlunzen et al., 2010; Butt et al., 2011). Although the t-test is simple to use and easy to understand, without careful treatment of the data, its statistical assumptions (such as normality) could be easily violated. As a parametric test, it can be affected by the outliers in the data due to extreme events. Given this reason, many studies employ non-parametric tests, for example the Mann-Whitney U-test, to assess the significance of changes between study periods (i.e. Diem and Mote, 2005; Zhang et al., 2009a; Schlunzen et al., 2010). As non-parametric tests do not rely on the assumption of the data distribution and are free from the effect of outliers, they are popular in the study of climatic data.

In most cases the information of the land surface conditions or land cover is used to either indicate the time of change or to indicate the surface difference between study sites. There are also some studies that establish the link between land cover and precipitation through mathematical methods. One common method is to assess the correlation between land cover change and precipitation (Fuller and Ottke, 2002).
For example, Sarkar et al. (2007) presented a scatter plot of mean annual precipitation versus percentage of tree cover showing a positive relation between the two. In Zhang et al. (2003a) and Zhang et al. (2003b), the Pearson’s correlation coefficient between NDVI of previous season(s) and summer precipitation was calculated. Sarkar et al. (2007) correlated both the Fourier harmonics and empirical mode decomposition of the two data series, rather than comparing the NDVI and precipitation data. Another common method is to investigate the relationship between land cover and precipitation through regression models (Webb et al., 2005; Kaufmann et al., 2007; Sun et al., 2008). In this case, the land surface variables (such as tree cover percentage, soil moisture and roughness length) are used as predictor for precipitation. Liu et al. (2006b) and Zhang et al. (2008) constructed a feedback efficiency parameter to indicate the control of land surface on precipitation, expressed as the ratio of the lagged correlation coefficient between vegetation cover and precipitation and the cross-correlation coefficient of vegetation. In all cases, time series of land surface data is required.

In general, the statistical methods are useful in understanding data. Additional consideration needs to be taken in atmospheric information, especially precipitation. Rainfall is highly variable, which involves a large amount of randomness and is influenced by multiple factors. The likelihood of detecting a trend in random data might be small and the results could be unreliable. Furthermore, autocorrelation and local variation can invalidate the assumptions in standard tests. Efforts are needed to explore methods to address these issues.
2.1.3 Discussion of observational results

2.1.3.1 Deforestation

Deforestation studies are mainly conducted in the tropical/subtropical and some temperate regions across five continents (see Table 2.3). Some previous studies suggest that the land-atmosphere feedback is location dependent (e.g. Koster et al., 2004; Lawrence et al., 2007). However in this review, the location or climate classification does not explain much of the variation in the results between these studies. As illustrated in Figure 2.3, there is no consistent result of the deforestation studies in each climatic region in each continent.

Results vary widely among the three studies in Africa, although the sites are all in the equatorial tropics. In the west of Africa on the coast of the North Atlantic Ocean, the rainfall patterns is not different between a extensively deforested Côte d’Ivoire and a less deforested Liberia. Hence van Rompaey (1995) concluded that deforestation was not responsible for the droughts of the seventies and eighties. On the other side of the equatorial belt of Africa, Altmann et al. (2002) did not detect any particular trend in precipitation although rainfall variation increased in the Amboseli basin, Kenya. However, in the central south of Africa where large areas of rainforest remain, Sanga-Ngoie and Fukuyama (1996) found a decreasing rainfall trend which could be explained by deforestation. Nevertheless, they cannot prove a strong link between the change of precipitation patterns and deforestation.

In the Asian equatorial region, the Indochina Peninsula and the Bay of Bengal of India are both affected by the summer monsoon (June - September). Sadhukhan et al. (2000) studied the premonsoon precipitation (March - May) and found a decreasing trend in rainfall due to gradual deforestation in Bengal. Kanae et al. (2001) also
Table 2.3: Locations and the main climate characteristics of the observational deforestation studies.

<table>
<thead>
<tr>
<th>Continent</th>
<th>Region</th>
<th>Climate</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Africa</td>
<td>Zaire River Basin</td>
<td>tropical monsoonal, winter dry</td>
<td>Sanga-Ngoie and Fukuyama (1996)</td>
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<tr>
<td></td>
<td>Kenya</td>
<td>tropical semi-arid dry</td>
<td>Altmann et al. (2002)</td>
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<tr>
<td></td>
<td>west Africa</td>
<td>equatorial monsoonal, fully humid or</td>
<td>van Rompaey (1995)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>winter dry</td>
<td></td>
</tr>
<tr>
<td>Asia</td>
<td>Bengal region, India</td>
<td>tropical monsoonal, winter dry</td>
<td>Sadhukhan et al. (2000)</td>
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<td></td>
<td>India</td>
<td>tropical monsoonal</td>
<td>Kishtawal et al. (2010)</td>
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<td></td>
<td>Indochina Peninsula</td>
<td>tropical arid, winter dry</td>
<td>Kanae et al. (2001)</td>
</tr>
<tr>
<td></td>
<td>Indonesia</td>
<td>tropical monsoonal, fully humid</td>
<td>Falk et al. (2005)</td>
</tr>
<tr>
<td></td>
<td>China</td>
<td>warm temperate, winter dry &amp; hot summer</td>
<td>Zhang et al. (2005); Kaufmann et al. (2007); Xia et al. (2007); Sun et al. (2008); Zhang et al. (2009a)</td>
</tr>
<tr>
<td>South America</td>
<td>Amazon</td>
<td>tropical monsoonal, winter dry</td>
<td>Chu et al. (1994); Cutrim et al. (1995); Depaiva and Clarke (1995);</td>
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<td></td>
<td></td>
<td></td>
<td>Eltahir and Bras (1996); Eltahir and Humphries (1998); Durieux et al.</td>
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<tr>
<td>North America</td>
<td>South-eastern U.S.</td>
<td>warm temperate, hot summer, fully</td>
<td>Diem and Mote (2005)</td>
</tr>
<tr>
<td>Europe</td>
<td>Germany</td>
<td>humid</td>
<td>Schlunzen et al. (2010)</td>
</tr>
</tbody>
</table>
Figure 2.3: Map of 23 deforestation studies. The green up-arrows indicate the result of the study is rainfall increases after deforestation. The red down-arrows indicate rainfall decreases after deforestation. The blue circles represent the three studies that both increase and decrease effects are found, depending on the season. The grey circles indicate no particular trend is found in rainfall associated with deforestation. The symbols on the map are rough indications of the study locations.
found the after-monsoon (September) rainfall decreased in Thailand. But for India as a whole, urbanisation might have increased the frequency of heavy rainfall during the monsoon season (Kishtawal et al., 2010). The impact of deforestation in the east and north of China is neither confirmed in the studies by Zhang et al. (2005), Xia et al. (2007) or Sun et al. (2008).

There are a large number of deforestation studies on the Amazonia but their results are also different between specific sites. Brazilian western Amazon has the highest deforestation rate due to active development (Nobre et al., 1991). In Reserva Foresta Ducke, Brazil, deforestation increases temperature hence decreases precipitation (Eltahir and Bras, 1996; Eltahir and Humphries, 1998). In Alta Floresta, Mato Grosso, rainfall decreases at the end of dry seasons and the beginning of rainy seasons (Dubreuil et al., 2012). In Rondonia, Brazil, Chagnon et al. (2004) and Chagnon and Bras (2005) found increased precipitation over a deforested area but Butt et al. (2011) found that inside the deforested area the rainy season is delayed by 11-18 days. However Depaiva and Clarke (1995) detected both significant increasing and decreasing trends of rainfall among stations across Amazonia.

The results can be inconsistent which might be due to different study set-ups. For example, Durieux et al. (2003) conducted an analysis on 10 years monthly precipitation data where they found less precipitation during the dry season in the Brazilian Amazon; on the other hand, Lin et al. (2006) showed increasing rainfall in the dry season in the same region based on 4 years of data. Apart from the difference in the length of data, the focus of the study could also be a reason for the different findings. While Durieux et al. (2003) compared cloud cover over deforested and forested area, Lin et al. (2006) investigated the correlation between aerosol and precipitation.
Chapter 2. Literature Review

The type of conversion might have some effects on the result. In Nopu Rahmat, Indonesia, tropical rainforest has been converted into plantation cacao trees. Falk et al. (2005) reported a lower annual precipitation in this region than in other tropical rainforest areas. There are also a few studies on urbanisation effects. Studies in Beijing and the Pearl River Delta, where rapid urbanisations have occurred, found decreased precipitation (Kaufmann et al., 2007; Zhang et al., 2009a). On the other hand, in Hamburg, Germany and Georgia, U.S., precipitation increases due to urbanisation (Diem and Mote, 2005; Schlunzen et al., 2010). Although the heat island effect is the main concern in urbanisation studies, industrial pollution might be more striking in cities in China, which has led to the lower precipitation.

2.1.3.2 Afforestation/reforestation

Afforestation/reforestation has been found to increase precipitation in the tropical or subtropical regions, including the Mediterranean (Otterman et al., 1990; Jauregui, 1991; Bengai et al., 1993, 1994; Zhang et al., 2003a). Three studies of semi-arid Southern Israel are based on rainfall data collected for more than 30 years in the Negev desert (Otterman et al., 1990; Bengai et al., 1993, 1994). By comparing the rainfall patterns before and after 1960, these studies found an increasing trend in rainfall about two decades after the large scale afforestations. Otterman et al. (1990) and Bengai et al. (1993) agree on the increase of October rainfall from the observations. Bengai et al. (1994) estimated annual average rainfall would increase by 30%.

On the other hand, afforestation/reforestation has a possibly negative relation with precipitation in temperate regions. In the arid/semi-arid Northeast Asia region, including western Mongolia and northern China, the satellite images of cloud patterns over mountain and agricultural lands on August 1994 were studied by Sato et al.
They found that above the vegetated area the convective activities were suppressed and cloud formation was weak as a result of a reduction of sensible heat, while the increase of moisture from evapotranspiration (ET) was not significant. Their results imply that afforestation on desertified lands would decrease cloud frequencies and potentially reduce precipitation.

The detection of rainfall change due to land cover changes is hard when large-scale effects present. Otterman et al. (1990) did not find much change in the November rainfall (and possibly in the other non-October months too) during the study period. Otterman et al. (1990) and Bengai et al. (1993) both suggested the strong feedback of rainfall to land cover changes observed in October only is probably due to a weaker large-scale effects in October than the other months. Bengai et al. (1994) did not eliminate the possibility of global climate change as a cause of the annual rainfall increase, although the October increase is more likely to be a local effect. The occurrence of climate change increases the rainfall variation and makes research on the cause of precipitation changes even harder.

2.1.3.3 Comparison between land covers

Precipitation over forested areas and non-forested areas are compared and various results are found. In Inner Mongolia, China, precipitation decreases the least over forest followed by grassland and desert areas (Lu et al. 2009). The deforested site in southeast Western Australia has smaller size cloud droplets than the neighbouring native vegetation site, and precipitation was found to decrease by 30% within a decade over the agricultural land (Junkermann et al. 2009). Investigating various regions around the world, Makarieva et al. (2009) suggested a forest-covered region could encourage maritime moisture to be carried further inland. On the other hand,
Davidowitz (2002) disagreed that a non-desert biome (forest, woodland or grassland) area has less precipitation variability than the desert area (mainly covered by scrubs) in south-western America. De Luis et al. (2000) and Negri et al. (2004) have even found lower rainfall over forested regions than non-forested or less forested regions in east Spain and Rondônia, Brazil.

The atmospheric circulation on the boundary of different land cover areas is another interesting topic. Carleton et al. (2001) found that on days with weak anticyclonicity, free convection was strong over the crop-forest boundary in the Midwest U.S. corn belt. Matyas and Carleton (2010) observed that turbulence on the boundary between cropland and forest actually enhance the development of convective precipitation in the same study region. Surface heterogeneity due to different land covers, which often shown by uneven wetness, can trigger mesoscale convection and can encourage the upward motion of moist air (Avissar and Liu, 1996). This is an important mechanism in generating precipitation.

Positive relationships are found between forest cover and convective clouds. This is usually observed in hot dry seasons. Unstable boundary layer and frequent cumulus clouds is more likely over forest and native vegetation, compared to the other types of land covers (Ray et al., 2001; Fisch et al., 2004). On the other hand, Wang et al. (2009) and Sato et al. (2007) suggested clouds are less frequent but deeper over forest. The study by Carleton et al. (2001) found greater convection over forest areas in moist days but in drier days convections were stronger over croplands. Similarly, Fisch et al. (2004) also suggested CBL is higher over pasture during dry seasons. Negri et al. (2004) found that deforestation could delay the onset of convective cloud during the day. Junkermann et al. (2009) pointed out that although there could be more cloud
droplets over agricultural land, the conditions for rainfall, such as the size of cloud droplets and dew point temperature, might not be favourable.

In Western Australia, land clearing activities have converted more than 10 million hectares of native forests into agricultural lands since the 1800s (Ray et al., 2001; Junkermann et al., 2009). The State Barrier Fence was initially built in the early 1900s to protect agricultural land on the west coast from animal damage (McKnight, 1969; Crawford et al., 2001). By 2001, the total fence has extended to 3,256 km (figure from the Department of Agriculture and Food, Government of Western Australia). Due to its massive size, a well documented long history of changes and the clear separation between the two types of land covers, this region provides a unique natural laboratory for land-atmosphere interaction studies (Junkermann et al., 2009).

The three Western Australian studies focus on the behaviour and different causes of cloud formation above agricultural land and native vegetation separated by the State Barrier Fence. Satellite imagery reveals possible preferential summer cloud formation over native vegetation on the east side (Ray et al., 2001; Lyons, 2002; Nair et al., 2011). Ray et al. (2001) obtained higher latent heat and lower sensible heat over the native perennial vegetation on the 16th January 2001 from the ASTER data (Advanced Spaceborne Thermal Emission and Reflection Radiometer). On the other hand, in the study by Junkermann et al. (2009), lower humidity and higher temperature were found over native vegetation compared to agricultural land during a winter cloudy day. As a result, there were a smaller number of cloud droplets over natural vegetation. Although more clouds were found above agricultural land, the droplet sizes were too small to generate rainfall. The difference in cloudiness in this region is visible on the satellite images (Nair et al., 2011), where summer cumulus clouds on the east were much
“heavier” than the winter clouds on the west of the fence. [Ray et al. (2001)] further explained that the summer cloud is a result of higher ET from perennial vegetation (a local effect), while the winter cloud is more likely influenced by the frontal rainfall as well as irrigation (mixture of maritime and local effect).

### 2.1.4 Summary

Based on the observational studies reviewed above, there is no conclusive relationship between forest cover and precipitation. Both positive and negative relations are found in the tropics and the mid-latitude. A general problem among these studies is that the link between land cover change and precipitation or cloud cannot be established by clear evidence or methods. I also notice there is few observational studies on the southeast Australia land cover change effects.

Finding evidence of precipitation feedback to forest cover change from data is difficult due to the unquantified noise in the data and the lack of detailed information, such as the amount and the evolution of soil moisture, the change of land cover and the amount of local recycled precipitation. (1) Many land cover changes are gradual and not well documented (e.g. [Sadhukhan et al. 2000]). Historical data of land use and land cover changes is lacking. (2) Although some efforts have been made to eliminate large-scale effects (e.g. [Bengai et al. 1993]), it is still unclear whether the changes observed can be attributed to the local land cover changes. (3) Furthermore, unlike rain water that can be directly used by vegetation, land surface conditions do not have a direct effect on rainfall. The change of land surface conditions has to change the other factors and processes, such as heat fluxes, to influence rainfall. Between land surface and precipitation, some processes are also affected by large scale effects which
further complicate the relationship. For example, the amount of ET that contributes to the local precipitation depends on the strength of the atmospheric circulation (Eltahir and Bras, 1996; Trenberth, 1999).

The mechanisms that are responsible for the interactions between forest cover and precipitation are many and probably location dependent (e.g. Ray et al., 2001; Lin et al., 2006; Junkermann et al., 2009; Makarieva et al., 2009). This question is not finalised yet. At this stage, scientists have proposed various hypotheses, such as energy balance, heat transport and aerosol production (Chagnon and Bras, 2005; Lin et al., 2006; Sato et al., 2007; Junkermann et al., 2009 etc.), and provided evidence drawn from observational data to support their hypotheses. However, there is no easy way to conduct a comprehensive assessment on the various causal relationships. It might take a long time to reach a consistent and robust conclusion from the observations.

2.2 LULCC and climate - results from model simulations

The number of modelling studies dominates the observational studies in the literature. It implies that model simulations are the main method in the study of land-atmosphere feedback. This again reflects the difficulties in conducting empirical experiments and seeking evidence from observational data.

2.2.1 Models in the studies

In the study of land-atmosphere interaction, coupled models that represent different components of the global or regional climate conditions and land surface are commonly used. The most important development might be the general circulation models (GCMs), which estimate general circulation of energy, mainly at the global scale, and simulate climate response to various forcings.
Simulated climate patterns are usually sensitive to the design of land surface scheme (Saha et al. 2006; Zeng et al. 2010, 2012; Gianotti et al. 2012; Zabel et al. 2012; Matthes et al. 2012). Hence, choosing or developing an appropriate land surface scheme is one of the main interests in the research of land-atmosphere interaction. Over the last twenty years, further understanding of important land surface properties has improved the construction of land surface schemes (e.g. Friend and Kiang 2005; Saha et al. 2006; Queguiner et al. 2011; Davin and Seneviratne 2012; Lawrence et al. 2012). This development is expected to increase the credibility of model results.

2.2.1.1 1990 - 1995

One of the widely used GCM models during this period was the National Center for Atmospheric Research (NCAR) Community Climate Model version 1 (CCM1) (e.g. Dickinson and Kennedy 1992; Henderson-Sellers et al. 1993; Pitman et al. 1993; McGuffie et al. 1995). The resolution of this model is 4.5° × 7.5° which is quite coarse. Compared to the previous models, this model includes seasonal and diurnal cycles (Dickinson and Kennedy 1992; Henderson-Sellers et al. 1993). The surface scheme implemented in CCM1 is the Biosphere-Atmosphere Transfer Scheme (BATS). BATS consists of three soil layers and one vegetation layer with 18 surface cover types (Yang and Dickinson 1996). There are seven main parameterisations in BATS: (1) drag coefficients, (2) soil surface evaporation, (3) wet canopy evaporation and dry canopy transpiration, (4) infiltration and surface runoff, (5) internal soil water fluxes, (6) base flow and saturation excess and (7) soil moisture rate equations. It covers a large range
of information of the land surface. One problem with BATS as recognised by Yang and Dickinson (1996) is that latent heat fluxes are underestimated on dry soil due to difficulties in root dynamics modelling.

The other popular model was the National Meteorological Center (NMC) global spectral model (e.g. Shukla et al., 1990; Nobre et al., 1991; Dirmeyer and Shukla, 1994) which similarly consists up to 18 vertical layers. This model has a finer resolution of 1.8° × 2.8°. Its land surface scheme was initially the Simple Biosphere Model (SiB) and later it was updated to a more simplified version (SSiB) (Xue et al., 1991). SiB is one of the early land surface parameterisations (LSPs) that provides detailed physical and morphological properties of vegetation (Walker et al., 1995) and describes the interaction between vegetation dynamics and the exchange of heat and momentum (Nobre et al., 1991). In a study to estimate heat fluxes from a temperate broad-leaved forest using SiB, Schelde et al. (1997) pointed out that this land surface model tends to overestimate below-canopy eddy diffusive resistance and hence underestimates soil evaporation. Otherwise SiB is able to produce good estimates of heat fluxes in a forest region.

During this period, increasing attention had been given to the hydrological functions of vegetation in the climate model. Rainfall interception was recognised as an important property of the vegetation cover. Water storage on the canopy foliage and ground cover foliage is one of the SiB variables (Nobre et al., 1991). In Dickinson and Kennedy (1992), it is suggested that the reduction in ET is mostly due to the reduction in interception after deforestation. Noticing the weakness of Meteorological Office GCM in modelling the interception of precipitation, Lean and Rowntree (1993) used a new interception formulation and found that the previous studies overestimated
the impact of deforestation. Eltahir and Bras (1994) pointed out that the interception formulation could be as simple as the one proposed in BATS but if spatial variability was ignored the interception loss might be overestimated.

2.2.1.2 1996 - 2000

The “second-generation” land surface models which include BATS, SiB and CLASS (Canadian Land Surface Scheme) were still used during this period (e.g. Sud et al., 1996; Verseghy, 1996; Xue and Shukla, 1996). BATS and SiB continue to be updated from the earlier versions. CLASS is written specifically for the Canadian GCM developed at the Atmospheric Environment Service, Canada (Verseghy, 1996). This model has three soil layers, a snow layer and a canopy layer. Canopy ET is influenced by intercepted water and the bulk stomatal resistance. The canopy component is a “big leaf” model for each grid cell (i.e. homogeneous canopy properties in each cell) (Verseghy, 1996).

The third version of CCM (CCM3) has been developed and implemented in studies during this time period. These models are configured at a higher resolution (2.8° × 2.8°) and a shorter time step (20 min) than the earlier versions (Bonan, 1999). The corresponding land surface scheme in this model is the NCAR LSM which is a third generation one-dimensional land surface model. It contains a detailed vegetation component which includes 12 plant types. Nine properties, which are leaf and stem areas, root profile, height, leaf dimension, optical properties, stomatal physiology, roughness length, displacement height and biomass, are used to distinguish the different types of plants (Bonan, 1999). This relatively high resolution coupled model, however, does not simulate precipitation well and its result is subject to spatial bias (Bonan, 1999; Costa and Foley, 2000).
During this period, multi-dimensional models were also applied in some studies. De Ridder and Gallee (1998) used a mesoscale atmospheric model, Modèle Atmosphérique Régional (MAR) to study land surface-induced regional climate change in Southern Israel. In their study, the two-dimensional mode of the model was adapted. The model employs the full continuity equation which allows solving for vertical and horizontal dimensions. Furthermore, it contains a cloud model which can extend up to three dimensions. In a rainfall study in India, Raman et al. (1998) also used a three-dimensional mesoscale model developed by the Naval Research Laboratory/North Carolina State University. These multi-dimensional models appear to have the ability to resolve convective and nonconvective clouds and boundary layer processes in details but might need to forfeit computation efficiency by introducing a large number of parameterisations (De Ridder and Gallee 1998; Raman et al. 1998; Sewall et al. 2000).

2.2.1.3 2001 - 2005

One important feature of the modelling studies after 2000 is the use of regional or mesoscale climate models (RCM). Compared to the global climate models which usually run at a coarse resolution, regional climate models have the capacity to downscale global-scale information and simulate local scale events more accurately and efficiently, as pointed out by Norman Miller, leader of the Regional Climate Center in Berkeley Lab’s Earth Science Division, during an interview in July, 1999 (Preuss 1999). Typical spatial resolution of the global climate models is 200 km or above, while for RCMs it can be as fine as 50 km or even smaller (Heck et al. 2001). Hence an RCM is a better model for predicting regional climate that is highly affected by mesoscale features (Gaertner et al. 2001; Gao et al. 2003). Some of the RCMs
used by the reviewed studies are listed in Table 2.4 and 2.5. The RCMs are mostly built on the same physics as the global model (Kanamitsu and Mo, 2003) so they can be integrated with the same set of land surface models, such as SSiB and BATS. However, boundary conditions are needed for models at a sub-global scale, such as regional models. This was and is still a problem for smaller scale modelling. In this case, the boundary inputs that are supplied by the global model are required (Gao et al., 2003; Kanamitsu and Mo, 2003), which means the regional models would be still dependent on global models.

The other feature of models in this period is the popular use of the Community Climate Models (CCMs) and Community Atmosphere Models (CAMs). These models were developed at NCAR. CAM is the current name for the model series after CCM3 to reflect the fully coupled climate system (Collins et al., 2004). The NCAR climate/atmosphere model series contains comprehensive information on the dynamics and physics of the atmosphere with three-dimensional capacity. Hence they were widely used at the time and the collaboration between users and developers were encouraged to improve the models (Collins et al., 2004).

Some new members of the land surface schemes had emerged in the use of CCMs/CAMs. Zhang et al. (2001) implemented CCM1-Oz, which is a modified version of CCM1, coupled to BATS1e to assess the effects of tropical deforestation on regional and global climate. The community models are flexible in terms of the choice of the land surface models. Zhao and Pitman (2002) coupled the CCM3 with BATS while Semazzi and Song (2001) and Hoffmann et al. (2003) integrated it with the NCAR Land Surface Model (LSM, v1) as described in Bonan (1996). As documented in Collins et al. (2004), BATS has become optional since CCM2 and the CCM3
Table 2.4: Information of different regional models or mesoscale models and associated land surface model (LSM) (part 1).

<table>
<thead>
<tr>
<th>RCM</th>
<th>Dim.(^1)</th>
<th>Resolution</th>
<th>Type(^2)</th>
<th>LSM</th>
<th>Applied region</th>
<th>Studies</th>
</tr>
</thead>
<tbody>
<tr>
<td>BRAMS</td>
<td>3D</td>
<td>8-64 km</td>
<td>1 d</td>
<td>N</td>
<td>Amazon</td>
<td>Saad et al. (2010)</td>
</tr>
<tr>
<td>HIRHAM</td>
<td>2D</td>
<td>0.5×0.5°</td>
<td>12 - 40 min</td>
<td>N</td>
<td>Whole Mediterranean</td>
<td>Gaertner et al. (2001)</td>
</tr>
<tr>
<td>NCAR RegCM2</td>
<td>3D</td>
<td>60 km</td>
<td>H</td>
<td>BATS 1e</td>
<td>China</td>
<td>Gao et al. (2003); Zeng et al. (2002)</td>
</tr>
<tr>
<td>NCAR RegCM3</td>
<td>3D</td>
<td>30 km</td>
<td>H</td>
<td>BATS 1e</td>
<td>U.S, China</td>
<td>Kueppers et al. (2007); Chen et al. (2010)</td>
</tr>
<tr>
<td>NCEP ETA</td>
<td>3D</td>
<td>40 km</td>
<td>H</td>
<td>SSiB</td>
<td>Amazon</td>
<td>Correia et al. (2008)</td>
</tr>
<tr>
<td>NCEP RSM(^3)</td>
<td>3D</td>
<td>20 or 50 km</td>
<td>1 d</td>
<td>SSiB</td>
<td>Southwestern U.S.</td>
<td>Kanamitsu and Mo (2003)</td>
</tr>
<tr>
<td>PROMES(^4)</td>
<td>2D</td>
<td>50×50 km</td>
<td>5 min - 15 d</td>
<td>H</td>
<td>western Mediterranean</td>
<td>Gaertner et al. (2001)</td>
</tr>
</tbody>
</table>

\(^1\) Dimensions.

\(^2\) Whether it is hydrostatic (H) or non-hydrostatic (N).

\(^3\) National Centers for Environmental Prediction Regional Spectral Model.

\(^4\) Spanish: “PROnóstico a MESoescala”.
<table>
<thead>
<tr>
<th>RCM</th>
<th>Dim.</th>
<th>Resolution</th>
<th>Type</th>
<th>LSM</th>
<th>Applied region</th>
<th>Studies</th>
</tr>
</thead>
<tbody>
<tr>
<td>PSU-NCAR MM5</td>
<td>3D</td>
<td>50 km 1 d</td>
<td>N</td>
<td>Pleim-Xiu LSM</td>
<td>Australia</td>
<td>Narisma and Pitman (2003); Pitman et al. (2004)</td>
</tr>
<tr>
<td>RAMS</td>
<td>3D</td>
<td>56 km 90 s - 1 d</td>
<td>N</td>
<td>LEAF6-2</td>
<td>U.S., southwest Western Aus., Puerto Rico</td>
<td>Roy et al. (2003); Pitman et al. (2004); Narisma and Pitman (2006); van der Molen et al. (2006); Kanae et al. (2001)</td>
</tr>
<tr>
<td>RAM v4.3</td>
<td>3D</td>
<td>2 - 80 km 3 s - 3 h</td>
<td>N</td>
<td>LEAF-2</td>
<td>U.S., India</td>
<td>Pitman et al. (2004); Georgescu et al. (2008); Lei et al. (2008); Wichansky et al. (2008); Nair et al. (2011)</td>
</tr>
<tr>
<td>RAM v6.0</td>
<td>3D</td>
<td>1 - 64 km</td>
<td>N</td>
<td>7 LEAF-2</td>
<td>southwest Western Aus.</td>
<td>Pitman et al. (2004); Georgescu et al. (2008); Lei et al. (2008); Wichansky et al. (2008); Nair et al. (2011)</td>
</tr>
<tr>
<td>RIEMS</td>
<td>3D</td>
<td>1°</td>
<td>N</td>
<td>BATS 1e</td>
<td>northern China, southern Mongolia</td>
<td>Zhang et al. (2005)</td>
</tr>
<tr>
<td>WRF</td>
<td>3D</td>
<td>5 - 36 km</td>
<td>N</td>
<td>Noah-LSM</td>
<td>Thailand, Turkey, South America</td>
<td>Takahashi et al. (2010); Sertel et al. (2011); Lee and Berbery (2012)</td>
</tr>
</tbody>
</table>

5 PSU: Pennsylvania State University. This is the fifth generation Mesoscale Model.
6 Land Ecosystem-Atmosphere Feedback model.
7 This model calculates surface temperature and ground temperature based on the surface energy balance and multiplier diffusion.
8 Regional Integrated Environmental Model System.
incorporated LSM given its comprehensive treatment of land surface processes. The use of BATS in Zhao and Pitman (2002) might be due to their focus on the global domain hence local land surface treatment is not important. Oleson et al. (2004) compared the results from two land models, NCAR LSM v1 and CLM2 (Community Land Model), implemented in CAM2 which is a successor to CCM3. Although CAM2 is not documented in Collins et al. (2004), Oleson et al. (2004) pointed out that it improves the prognostic cloud-water parameterisation and radiation package which have contributed to the latest version of CAM3 (Collins et al., 2004). In Gibbard et al. (2005), the CAM3 with finer resolution (2.0° × 2.5°) coupled with CLM land model is used to assess the global climate impact.

2.2.1.4 2006 - 2012

As research attention on land-atmosphere interaction increases, more climate models have been developed and implemented in the studies. During this last period, the number of modelling studies is almost equal to the sum of studies in the previous 15 years. Within the 44 studies under review, 28 atmosphere models are used and three of them have more than one version, such as CAM (CCM3 and CAM3), CCSM (CCSM3 and CCSM4) and ECHAM (ECHAM3 and ECHAM5).

Compared to the previous models, the new models generally have higher spatial resolution. The resolution of CAM3 is T42 (2.8° × 2.8°) (Swann et al., 2012). The CSIRO Mar3 AGCM has grid size of about 1.8° × 1.8° (McAlpine et al., 2007). As a regional climate model, PRECIS uses 0.5° × 0.5° (Williams and Kniveton, 2012). RegCM3 has a horizontal resolution of 30 km (Kueppers et al., 2007) and RAMS can perform simulation on a 1 km grid set (Nair et al., 2011). Precipitation estimates require a higher model resolution than temperature (Narisma and Pitman, 2003; Avila...
The new generation of models developed with high resolution make the prediction of precipitation after land cover changes become more reliable.

There was a high frequency of use of the Regional Atmospheric modelling System (RAMS). RAMS is a flexible meteorological modelling system (Narisma and Pitman, 2006) which can run at very fine spatial resolutions (Lei et al., 2008; Nair et al., 2011). The latest version, RAMS v4.3, is a non-hydrostatic model which can solve full non-linear equations of processes between soil, land surface and atmosphere (Georgescu et al., 2008; Wichansky et al., 2008). RAMS is found to simulate convective systems well as a result of the grid nesting design (van der Molen et al., 2006; Georgescu et al., 2008; Lei et al., 2008; Wichansky et al., 2008). Narisma and Pitman (2006) suggested that RAMS can produce good simulations for Australian atmospheric processes. In Narisma and Pitman (2006), the dynamic plant model GEMTM is used given its ability in allowing dynamic vegetation response to CO2 concentration and climate changes. But most commonly the Land Ecosystem-Atmosphere Feedback (LEAF-2) model is coupled to RAMS. LEAF-2 can be used with high-resolution land surface data and reflects land surface heterogeneity in the simulations (Roy et al., 2003).

The Weather Research and Forecasting (WRF) modelling system is another emerging regional climate model. It is also a community model which was developed by the collaboration of several institutions, such as the NCAR Mesoscale and Microscale Meteorology (MMM) Division, the National Oceanic and Atmospheric Administration’s (NOAA) National Centers for Environmental Prediction (NCEP), a number of university scientists (Skamarock et al., 2005), as well as potential future users and developers. This model is designed to apply in a broad research area, such as numerical weather prediction, downscaling climate simulation and land-ocean-
atmosphere coupling for scales from meters to thousands of kilometers (Skamarock et al., 2005). WRF is a fully compressible, non-hydrostatic and flexible model which also provides grid nesting facility (Skamarock et al., 2005; Takahashi et al., 2010; Sertel et al., 2011; Lee and Berbery, 2012). The Advanced Research WRF (ARW) is coupled with the Noah land surface model (Noah-LSM). The LSM contains four soil layers and a canopy layer in which interception, resistance and ET are estimated (Lee and Berbery, 2012).

2.2.1.5 Modelling experiments in the studies

In these modelling studies, model experiments are generally set up to compare different land cover scenarios (sensitivity analysis). Among the different scenarios, a “control” or “reference” case often reflects the current land cover (e.g. Williams and Kniveton, 2012) or undisturbed land cover (e.g. Nobre et al., 1991; Hirota et al., 2011) of the study region. In a few studies on the effect of heterogeneous surface, a control run refers to a homogeneous surface (e.g. Garcia-Carreras and Parker, 2011). The land surface data in the control scenario usually come from the following sources:

- Map of current/natural vegetation distribution. This is seen in the following studies: Clark et al. (2001), Semazzi and Song (2001), Narisma and Pitman (2003), Voldoire and Royer (2004), Pitman et al. (2004), Voldoire and Royer (2005), Hirota et al. (2011) and Lee and Berbery (2012).

- Data from the literature. Verseghy (1996) used four datasets, representing needle-leaf forest, continental grassland, temperate mixed forest and alpine tundra respectively, from four previous studies. Matthews et al. (2004) combined two historical land cover change datasets from the literature to represent the
change of croplands and pasture globally. Nobre et al. (2009) used the vegetation types described by another study using the same AGCM model.

- Remote sensing data. For example, Fairman et al. (2011) used the mean monthly MODIS-derived NDVI to specify the current vegetation cover. In Georgescu et al. (2008), the land cover data is derived from the US Geological Survey’s 30 m National Land Cover Dataset and the AVHRR products.

- Results from previous model simulations, such as in Taylor et al. (2002), Gao et al. (2003), Matthews et al. (2003) and Roy et al. (2003). For example, in Roy et al. (2003), a mechanistic terrestrial biosphere model is used to estimate the vegetation cover in the United States.

- Hypothesized full vegetation cover or zero vegetation cover. Wang et al. (2000) and Berbert and Costa (2003) assumed a uniform forest cover in the Amazonia in the control run, while Hoffmann et al. (2003) assumed that the surface was covered by 95% broadleaf evergreen trees and 5% pasture grass. One of the control run cases in Gibbard et al. (2005) assumed bare soil cover.

However, sometimes details of the control scenario are not clearly given (such as Gedney and Valdes, 2000; Gaertner et al., 2001; Zhang et al., 2001; Gibbard et al., 2005; Takahashi et al., 2010; Schneck and Mosbrugger, 2011).

The deforested or reforested cases could be based on observational data or hypothesized conditions. In the study of Wang et al. (2000), the observed deforestation data from literature is used in the deforestation run, compared to a control run that is defined by a uniform forest cover over the Amazon. Kanae et al. (2001) constructed the past vegetation cover by replacing part of short vegetation to forest. In the study of Takahashi et al. (2010), the control case (with current land cover) is compared to a
wetter scenario (dense rainforest with 90% green fraction) and a drier scenario (sparse
forest with 20% green fraction). Commonly a extreme scenario of land cover change
is set up, such as complete deforestation/afforestation or complete desertification (e.g.
Costa and Foley 2000; Gedney and Valdes 2000; Kleidon et al. 2000; Voldoire and
Royer 2004; Lee and Berbery 2012). Studies on complete land cover change might
provide a upper/lower bound of the predicted climate but the situation is unrealistic
and less useful (Sud et al. 1996; Clark et al. 2001; Semazzi and Song 2001; Snyder 2010). By comparing sensitivity experiments of deforestation scenario and
afforestation scenario, Gaertner et al. (2001) argued that deforestation case is easier to
specify in a consistent way when applied in different land surface schemes. This might
explain the higher number of deforestation studies than afforestation studies.

Some studies compare precipitation simulations with the observed climate data
to test the model’s performance, mostly before conducting the sensitivity analysis
(e.g. Xue and Shukla 1993; Polcher and Laval 1994b; Clark et al. 2001; Wichansky
et al. 2008; Fairman et al. 2011). Gaertner et al. (2001) conducted a seasonal
comparison between the ensemble averages (six members) of control runs and the
observed precipitation for the two models PROMES and HIRHAM. HIRHAM can
give a quite accurate estimate of the mean precipitation, while PROMES models the
spatial distribution of precipitation well. Correia et al. (2008) averaged precipitation
data from the literature and ground station measurements in Brazil to construct a
validation dataset. Clark et al. (2001) found the model simulation is poor in the
mountainous area compared to observations and reanalysis data. Reanalysis data is
quite often used in the model validation part, such as in Clark et al. (2001), Silva et al.
found that RAMS is able to reproduce the broad patterns of temperature distribution in North America through validation with observations. Such tests could increase the credibility of model application.

Overall, improvements have been continuously made and models have become more complex over the last two decades or so, as more details were added into the models and finer resolution could be achieved. These efforts tend to reduce the difference between the model and the real world. Nevertheless, complex models are still rough representation of the reality and cannot always represent all factors. Sensitivity experiment is a common method adapted by modelling studies. The difference in the knowledge of current land cover and in the expectation on land cover changes could contribute to the difference between studies.

2.2.2 Regions and vegetation-precipitation feedback

The land-atmosphere feedback relationship is stronger in some places but weaker in others. As pointed out by Pitman et al. (2009), the biogeophysical impacts of land cover change (LCC) on the climate are regionally significant. Music and Caya (2009) found that in model simulations the sensitivity of water budget components to land surface parameterisations and lateral boundary conditions varies from basin to basin. Hence it is worth to investigate the relationship between spatial location and the vegetation-precipitation feedback from the studies.

2.2.2.1 Tropics

The tropical rainforest area, especially the Amazon Basin, has clearly attracted most of the research attention in this field (e.g. Henderson-Sellers et al., 1993, Pitman et al., 1993, Dirmeyer and Shukla, 1994, Eltahir and Bras, 1994, Walker et al., 1995, Gedney 46
The Amazon basin covers 7 million km$^2$ land area and more than 78% (equivalent to 5.5 million km$^2$) is the moist broadleaf forest. Economic activities and infrastructure improvement in this region have triggered rapid land clearing and are threatening the sustainability of the rainforest (Kirby et al., 2006). According to the National Institute for Space Research (INPE) of Brazil, average annual deforestation rate in the 1990 - 2011 period was 16,062.8 km$^2$ which is about one fourth of the size of Tasmania. The large scale of deforestation could result in loss of carbon sink which might lead to climate change such as global warming (Gedney and Valdes, 2000; Cleal and Thomas, 2005; Bala et al., 2007; Le Quéré et al., 2009; Pan et al., 2011; Lewis et al., 2011). It is therefore no surprise that this region has been the center of land-atmosphere feedback studies. The world’s second largest tropical rainforest is in the Congo Basin in Africa, comprising an area of 3.4 million km$^2$. The tropical rainforest also extends into Southeast Asia, Indonesia and Papua New Guinea. According to the Köppen-Geiger climate classification, these regions are dominated by hot equatorial temperature and high annual rainfall under the effect of monsoon.

Studies generally agree that there is a significant land-atmosphere interaction in the tropics. More important, the local land-induced signal is strong in this region (Reale and Dirmeyer, 2002; Feddema et al., 2005; Pitman et al., 2009). In Amazon, most studies have found a decreased precipitation associated with deforestation, although the magnitudes of reduction are not the same. In terms of percentage, the range is approximately from 7% to as high as 40% (Nobre et al., 1991; Henderson-Sellers et al., 1993; Lean and Rowntree, 1993; Eltahir and Bras, 1994; Walker et al., 1995; Hoffmann et al., 2003; Silva et al., 2006; Nobre et al., 2009). There is a large difference even within the same study when comparing multiple models. Nobre et al. (2009)
estimated 26.1% and 42.2% rainfall reduction from AGCM and CGCM respectively, assuming a complete deforestation in the Amazonia.

Decreasing precipitation is not always the case after deforestation. If the albedo change, CO$_2$ concentration effect and deforestation patterns are considered, precipitation could increase as a result of deforestation (Dirmeyer and Shukla, 1994; Costa and Foley, 2000; Correia et al., 2008; Saad et al., 2010). Dirmeyer and Shukla (1994) found that if albedo increases by only a small amount (i.e. $< 0.03$) after deforestation, average precipitation would increase given a higher surface temperature. Similarly, Correia et al. (2008) modelled an increase in precipitation in the cases of relatively small scale deforestation. The thermal driven moisture convergence explains the increase in precipitation in these studies. While deforestation would decrease precipitation in the Amazon basin, increased CO$_2$ concentration can have the opposite effect (Costa and Foley, 2000). This implies that if deforestation leads to a significant increase of CO$_2$ concentration, the combined effect can possibly increase precipitation. Increasing precipitation is also found over strip-shape deforestation areas, which is parallel to the wind direction, due to the deforestation breeze (Saad et al., 2010).

The effect of deforestation is more variable or less obvious in the other tropical rainforest regions. Assuming the tropical rainforest in Africa is also replaced by grassland, precipitation change might have the same sign as in the Amazon but with a smaller magnitude (Polcher and Laval, 1994a; Hoffmann et al., 2003; Voldoire and Royer, 2004). Polcher and Laval (1994a) estimated that deforestation in the tropical belt could increase precipitation in Africa by 3.4%, compared to 15% in the Amazon. Hoffmann et al. (2003) modelled a 26% decrease in Africa and a 34% decrease in the Amazonia, assuming complete deforestation. It is common that a smaller
deforestation-precipitation signal can be found in Africa in comparison to the Amazon (McGuffie et al., 1995; Zhang et al., 2001; Voldoire and Royer, 2005; Semazzi and Song, 2001; Maynard and Royer, 2004). Similar results have been shown in Southeast Asia and Indonesia but with a slightly stronger effect (McGuffie et al., 1995; Zhang et al., 2001; Findell et al., 2007). McGuffie et al. (1995) pointed out that the different feedbacks in the tropical regions are due to the regional circulation patterns. Moisture convergence would decrease in the Amazon after deforestation; but it is usually found to increase in Southeast Asia and Africa, which might ease the impact of reduced ET. Furthermore, the monsoon influence can complicate the precipitation feedback in Southeast Asia and Indonesia (Polcher and Lavallée, 1994b).

The semiarid tropical region, Sahel, has also been studied extensively. In this region, the long lasting drought over the last 50 years and desertification are the major concerns (Xue and Shukla, 1993; Taylor et al., 2002). Sensitivity analyses were carried out by a degradation scenario in which current vegetation was replaced by desert shrubs with bare soil in Xue and Shukla (1993); Clark et al. (2001). Precipitation could significantly decrease by 14% - 33% if desert expands into this sub-Saharan region (Xue and Shukla, 1993; Clark et al., 2001). Both Xue and Shukla (1993) and Clark et al. (2001) agreed that the impact of land degradation is most severe in West Africa with a possible increase of precipitation in the southern part of Africa. A small decrease of rainfall (4.6% in 1996 and 8.7% in 2015 compared to the rainfall in 1961) is estimated when land use changed to cropland (Taylor et al., 2002) but larger decrease (40-50%) resulted from changing forest to short grass (Abiodun et al., 2010).
2.2.2.2 Mid-latitude temperate

Although tropical deforestation has attracted most attention, LCC mainly occurs in Europe, India, China and the U.S. (Zhao and Pitman [2002], van Dijk and Keenan [2007], Lawrence and Chase [2010]). Most of these regions are in the mid-latitude temperate area. By conducting a multi-model global simulation, Pitman et al. (2009) detected significant precipitation changes due to LCC are more likely in these regions, especially in East Asia and Central United States.

Many studies conducted in the North America have found no or unclear LULCC-precipitation feedback. The modern land use in the west and central U.S. is mainly agriculture activities, including intensive irrigation (Kueppers et al. [2007]). By comparing the simulated climate over potential natural land cover and present-day land cover, neither Kueppers et al. (2007) nor Wichansky et al. (2008) have detected any significant changes in precipitation. Lawrence and Chase (2010) found that the decrease of summer precipitation is outweighed by increase in the other seasons. Sud et al. (2001) agreed that increased vegetation cover could help to improve local rainfall through the effect of higher radiation absorption and ET; however such a relationship is greatly modified by the background circulation. Their study region is in the mid-western Great Plains which is affected by strong moisture advection from large-scale forcing. Vegetation-induced effects could be magnified by the active circulation during the wet seasons but this effect is negligible in a dry season. The unclear results are also partially due to model variations (Bonan, 1999, Oleson et al., 2004).

In the temperate region of Asia, most studies agree on a positive relationship between vegetation cover and precipitation. Land use change in this region, especially in China, is mainly from natural vegetation to farmland and urbanisation. Zhao and
Pitman (2002) found that, if broadleaf forest is replaced by grass and crops, the frequency of low to medium precipitation increases but high precipitation occurs less, which is due to the reduction in latent heat fluxes after the land cover change. There is a significant precipitation decrease in northwest China, where the climate is hot and dry (arid/semi-arid area), after expanding cropland and desertification followed by weaker monsoon circulation (Gao et al., 2003; Li and Xue, 2010). However such precipitation change is not seen in east China (Gao et al., 2003) which might be due to the maritime influence.

There are almost even distribution of different LULCC-precipitation feedback results found in Europe and the Mediterranean. The land use experiments in De Ridder and Gallee (1998) and Zhao and Pitman (2002) have shown positive effect on precipitation, one with intensified irrigation agriculture in Southern Israel and the other one with grasses and crops replacing forest over Europe. De Ridder and Gallee (1998) suggested the cooling effect of wet surface is more important than the reduction in roughness length due to LCC. Similarly, Zhao and Pitman (2002) argued the reduction in stomatal resistance after LCC in Europe could overcome the impact of changes of the other land surface properties, which eventually leads to higher latent heat flux and lower temperature. Furthermore, De Ridder and Gallee (1998) pointed out that wetting the coastal area of Southern Israel could reduce the downward transport of dry air and promote moist convective activity.

Gaertner et al. (2001) noticed that precipitation response is negligible in deforested areas under strong maritime influence. Advective moisture from remote locations is the main source of precipitation. This might also explain why no significant precipitation change is detected in the Marmara region, Turkey (Sertel et al., 2011) and a region
close to the Atlantic (Findell et al., 2007). Small magnitude of positive precipitation changes in response to vegetation changes are also found in inland Europe (Findell et al., 2007; Galos et al., 2011).

Although observational studies are very limited in Australia, there has been high interest in modelling experiments since 2000. The regions subjected to major LULCC are southeast Australia and southwest of Western Australia. Both areas have experienced significant land cover changes since European settlement mainly due to agricultural activities (Ray et al., 2001; Junkermann et al., 2009; Deo et al., 2009a). By comparing pre-European land cover and modern day land cover, precipitation in eastern and southeast Australia is found to decrease by 4-8% (McAlpine et al., 2007; Syktus et al., 2007), associated with shifting of precipitation distribution, increasing number of hot days and dry spells and being more prone to drought (McAlpine et al., 2007; Syktus et al., 2007; Deo et al., 2009a,b). In southwest of Western Australia where natural land cover and cropland are separated by a rabbit-proof fence, simulations have also found conversion of previous natural vegetation causing a reduction of local precipitation (Pitman et al., 2004; Timbal and Arblaster, 2006; McAlpine et al., 2007; Nair et al., 2011). Both Pitman et al. (2004) and Nair et al. (2011) agreed that changes in roughness length is the major reason for the rainfall decline. On the other hand, increased rainfall associated with LULCC is suggested. McAlpine et al. (2007) simulated a significant increase in precipitation over southern and central Australia by 4-12%. Syktus et al. (2007) estimated a small increase (0.6%) of winter rainfall in southwest Western Australia. Furthermore, the studies by Narisma and Pitman (Narisma and Pitman, 2003, 2006) do not support the influence of LULCC on rainfall. Their deforestation study (Narisma and Pitman, 2003) detected no
significant change in the January rainfall but a small significant change in July rainfall. In another reforestation study, they found no clear change in rainfall patterns (Narisma and Pitman, 2006).

Generally, the response of precipitation to land cover change reduces from the tropics to the poles (Kleidon et al., 2000; Gibbard et al., 2005; Findell et al., 2007; Lawrence and Chase, 2010). There are a few possible explanations for this. (1) There is a larger amount of global radiation towards the tropics than in the high latitudes (Snyder, 2010; van der Molen et al., 2011). (2) The types of trees are different, i.e. boreal forest in the tropics and needle leaf forest in the high latitudes, so their influence on heat and water redistribution could also be different (Gibbard et al., 2005).

Australia covers a wide range of climate zones, from the equatorial in the north to the temperate to the south and deserts in the center. There are a large variety of natural vegetations, such as closed canopy rainforest, Eucalyptus open forest, woodland and shrubland (Geoscience Australia, 2009). Hence land cover changes are expected to have various impacts on this continent.

2.2.3 LULCC and seasonal precipitation changes

Although land cover-precipitation interaction could be region dependent, it is more pronounced if studied by season. In many regions, the change in annual precipitation is negligible but in particular seasons, especially dry summer seasons, researchers have found significant changes due to LULCC (e.g. Gaertner et al., 2001; Semazzi and Song, 2001; Oleson et al., 2004; Deo et al., 2009a). Therefore recent land-atmosphere studies tend to investigate seasonal precipitation in order to obtain more meaningful results (Kleidon et al., 2000; Wichansky et al., 2008; Lawrence and Chase, 2010).
Summer is an important season for land surface process to influence the atmosphere. Since thermal energy is higher in summer as a result of long and intense solar exposure, strong convection and turbulence are active (Kanamitsu and Mo, 2003; Georgescu et al., 2008). Land-atmosphere interactions are encouraged by the highly frequent energy exchange between the surface and the air above it. Hence land cover change can impose a significant impact on the interaction. Such change is even more significant if the summer is dry (Semazzi and Song, 2001; Klingaman et al., 2008; Georgescu et al., 2008; Fairman et al., 2011). Using biome-derived land surface parameters in the NCAR land surface model (LSM), Oleson et al. (2004) detected a significant increase in summer precipitation but no change in autumn precipitation. Similarly, Gaertner et al. (2001) found that precipitation decreases in late spring and summer but not in autumn and winter, based on simulations from two RCMs. In Lawrence and Chase (2010) and Klingaman et al. (2008), LULCC causes precipitation to decrease in all seasons, with the largest decrease in summer and smallest in winter. The strong relationship between land cover change and summer rainfall is also recognised by many other studies (e.g., Xue and Shukla, 1993; Semazzi and Song, 2001; Alessandri et al., 2007; Deo et al., 2009a).

In the tropical regions, seasonal difference is more distinct between the wet period and the dry period. Wet seasons are usually influenced by summer monsoon and hence featured by very high precipitation. The signal of land-atmosphere interaction can be weaker under the strong monsoon effect. For example in the observational studies in Southern Israel, significant changes in rainfall has been detected during October but not November (and possibly other months) (Otterman et al., 1990; Bengai et al., 1994). Similarly, modelling studies have found LULCC might only have a seasonal effect.
On the Indochina Peninsula, Kanae et al. (2001) reported from model experiments that precipitation decreases over deforested areas in September but not in August. Although both months have shown very high rainfall amounts, August is controlled by the summer monsoon while in September the monsoon westerlies disappear over the region. These studies imply that the Asian Monsoon suppress land-atmosphere interaction. However, large scale land cover forcing might be able to affect the strength and timing of monsoon circulation and indirectly influence precipitation (Feddema et al., 2005; Lei et al., 2008; Li and Xue, 2010; Dallmeyer and Claussen, 2011).

2.2.4 Type of LULCC and precipitation changes

Again, deforestation is the type of land cover change that researchers are most interested at. The following scenarios are usually compared in the models: (1) forest; (2) savannah; (3) crops, pastures or short grasses; (4) bare soil; (5) urbanisation. Pielke et al. (2007) gave a detailed review on rainfall feedback as affected by different types of land cover change. In this review the knowledge of the impact of LULCC type is updated from more recent studies.

Change from tropical rainforest to short grass or pasture is the most common case among the deforestation studies. However, there are also complications in relation to the types of land cover changes between studies. For example, in terms of forest, there are broadleaf trees versus needle-leaf trees and evergreen trees versus deciduous trees. The various types of trees could have different effects on the energy redistribution and the hydrological balance (Snyder, 2010; Lawrence and Chase, 2010). On the other hand, a conversion to cropland which is irrigated and harvested has a very different impact on the local climate, compared to a conversion to non-irrigated and
non-harvested grassland (Narisma and Pitman, 2003). Furthermore, variations are also found between complete or partial deforestation/reforestation (Sud et al., 1996; Fairman et al., 2011).

Studies show that afforestation or reforestation does not necessarily have the opposite effect of deforestation. Fairman et al. (2011) simulated the impact of a complete deforestation and a complete reforestation on the climate of Kilimanjaro Mountain, which is on the border of Tanzania and Kenya. On the windward area, both types of forest cover change have a similar effect (same direction but not the same magnitude) below elevations of 1300 m and between elevations of 2200-4000 m. More important, on the leeward areas rainfall decreases at higher elevations in the deforestation scenario but shows no significant change in the reforestation scenario. A similar experiment is run by Dallmeyer and Claussen (2011) over the Asian monsoon domain. The feedback on precipitation appears in different locations across the domain depending on the types of forest cover changes. Decreasing precipitation in the deforestation case is relatively large and concentrated on the southeast coast of China and Indochina peninsula, while precipitation response to afforestation is weaker and mainly occurs in inland regions. Narisma and Pitman (2006) found that if the current land cover in southeast and southwest Australia is recovered by 25% to 75%, there would be no clear change in the precipitation pattern. At the same time, many deforestation studies in the same region of Australia simulate decreasing precipitation (Pitman et al., 2004; Timbal and Arblaster, 2006; McAlpine et al., 2007; Syktus et al., 2007; Nair et al., 2011).

Narisma and Pitman (2003) argued that replacing native vegetation cover by grassland which is not irrigated or harvested would reduce ET and lead to a reduction in
precipitation. They suggested the characteristics of the replacement vegetation cover, such as access to water and growth patterns, can determine the way that the LCC influences the climate. Their argument is similar to [Klingaman et al. (2008)] in which evergreen forest has been converted to sparse vegetation. Supporting this opinion, [Zhao and Pitman (2002)] found that in Europe, latent heat fluxes and precipitation increases after crops with lower stomatal resistance replacing deciduous forest; while in China, the opposite impact occurs as broadleaf forest being converted to grasses and crops, in which case the lower LAI, roughness length and root depth have more important influence.

Although the type of conversion could possibly make a significant contribution to the study results, it has not attracted much attention from the researchers given the fact that many studies did not specify the details of deforestation or LULCC. Furthermore, among the studies which the type of land cover change can be classified, such as in [Polcher and Laval (1994b)], [Berbert and Costa (2003)], [Silva et al. (2006)], [Kueppers et al. (2007)], [Klingaman et al. (2008)] and [Lei et al. (2008)], there is no consistent pattern of land cover change effect either in directions or in magnitudes. This implies that the types of land cover changes might be only a secondary cause of the variation in the results.

### 2.2.5 Scale of LULCC and precipitation changes

The effect of land cover change might vary depending on the spatial scale of the study area. [Schneck and Mosbrugger (2011)] found decreasing precipitation at deforested grid cell but increasing precipitation if looking at the regional changes. In the study of [Xue and Shukla (1993)], precipitation decreases both at the deforested region and
the whole Amazon basin but the basin-wide decreasing is only about half of the local decrease. On the other hand, by comparing a variable-resolution grid configuration simulation with a uniform-resolution grid configuration simulation\(^2\) Medvigy et al. (2011) obtained a 1.5% higher precipitation reduction in the latter case. The largest seasonal decrease is completely different between the two simulations, which is in December - February in the variable-resolution configuration and in March - May in the uniform-resolution configuration. (Medvigy et al., 2011) also pointed out that if 40% of the Amazon is deforested the impact is generally weaker than a total deforestation. A similar conclusion has been drawn by Nobre et al. (2009). Sud et al. (1996) conducted an Amazon deforestation experiment on disjointed region. The result is a savannah landscape after land cover change. Their results show that significant precipitation changes are only found in some small regions as a consequence of a more realistic deforestation scenario.

Here local scale studies are limited. Modelling experiments are usually conducted at mesoscale or above but local scale studies are based on observations (i.e. van der Molen et al., 2006). At the local scale, convective activities are important in the land-atmosphere interaction (Garcia-Carreras and Parker, 2011; Nair et al., 2011). Heat fluxes at the surface, the coupling between the mixed layer and the surface as well as the development of the boundary layer are the basic elements that influence the convection (Stull, 1988). In order to predict rainfall and cloud cover, microphysical parameterisation is generally needed, such as the cloud-resolving models, which requires a large amount of computations, as demonstrated in Gao and Li (2010). It is a question whether a local scale convection model which is manageable can be used

\(^2\) A 25 km characteristic length scale was applied over South America and nearby oceans then gradually coarsening to a 200 km for the rest of the world in the variable-resolution grid configuration; the entire world was simulated at 200 km resolution in the uniform-resolution grid configuration (Medvigy et al., 2011).
to estimate the feedback between the land surface and the precipitation.

The remote effect of LULCC is debatable. A complete deforestation of South American tropical rainforest can decrease precipitation over the Amazonian region. But it might increase winter precipitation in eastern North Atlantic and western Europe through the modification of the upper level circulation (Gedney and Valdes 2000). The tropical deforestation impact on the Northern Hemisphere can be explained by its influence on the Rossby wave forcing (Gedney and Valdes 2000; Snyder 2010). Forest cover changes in the Asian monsoon region could also trigger a teleconnection effect (Dallmeyer and Claussen 2011). Deforestation in this region is estimated to enhance precipitation in North Africa, while afforestation suppress summer precipitation in the Middle East. The remote effect of Southeast Asia deforestation is also studied by Schneck and Mosbrugger (2011). Temperatures in high latitude northern Asia is found to decrease after land cover change in the southeast. In their study, precipitation does not necessarily change although cloud covers increase. In the Sahelian region, desertification might decrease precipitation locally but it can increases precipitation further south of 10°N (Xue and Shukla 1993). Swann et al. (2012) found that afforestation in the mid-latitude could lead to lower rainfall in the Amazon and higher rainfall in the Sahel due to the northward shift of Inter-Tropical Convergence Zone (ITCZ). However, Voldoire and Royer (2005) and Pitman et al. (2009) did not find an agreement between models on the remote teleconnection patterns due to LULCC, nor any statistical significance in a single model simulation. The various results shown in the literature may be due to (1) the size of LULCC estimated, or (2) the level of complexity of LULCC effect, (3) simulation with a fixed/dynamic SSTs, or (4) the length of simulation, or (5) model variability (Pitman et al. 2009).
2.2.6 Summary

Currently model experiments can provide more information on the vegetation-precipitation feedback relation than observations. Sensitivity analysis of land cover change in model has the flexibility to control certain conditions, such as background forcing, soil properties or some land cover characteristics. Hence it is possible to isolate the influence of a single change as well as compare single effect and combined effects as in Costa and Foley (2000) and Zhao and Pitman (2002).

However even with stronger controls, the simulation results presented in the reviewed studies do not show a clear pattern by region, season, type of conversion or spatial scale. It is possible that the land-atmosphere feedback is under the combined effect of various factors, including the above four. Therefore studies should be treated on a case by case base and results cannot be transferred from one situation to another. It would have an even more important implication on management. Although a certain level of prediction can be made from previous studies, detailed research is still required before taking actions.

Furthermore, variation between models would also contribute to the diversity in results. Fundamental differences between models need to be better understood. The models differ in terms of the following aspects: (1) assumptions on the boundary conditions and influence; (2) spatial resolution; (3) detail of the land surface properties and processes; (4) assumptions on the mechanism controlling the interaction. Especially in the third point, a question could be asked: how much details are needed to obtain a reasonably good estimate? Before going even further on details, coming back to simple might be necessary to find out what has been achieved by adding complexity.
2.3 Feedback mechanism

Overall, the climate feedback to the vegetation surface changes is due to the effects of some surface physical processes and characteristics. Some of these important processes and characteristics include the light reflecting ability of land surface or surface albedo, the degree of roughness and heterogeneity of the surface. Vegetation cover changes can modify these physical processes and characteristics, which would change the heat and moisture exchange between the land surface and the atmosphere. They are the fundamental causes for the climate feedback. Some of the important feedback mechanisms are shown in Figure 2.4.

![Figure 2.4: The possible land-atmosphere feedback mechanisms. The impact of the vegetation surface on the atmosphere is through the effect of some land surface processes and characteristics.](image)

For example, the mechanism of the albedo effect lies in its controls on the energy budget. Deforestation leads to higher albedo (Chagnon and Bras, 2005) which means more solar radiation is reflected by the surface and less is absorbed by the ground.
Subsequently, the total energy available for sensible heat and latent heat reduces. If there is a dramatic decrease of sensible heat, convective energy is reduced so cloud development is suppressed ([Junkermann et al.,] 2009). On the other hand, if the reduction in net radiation accrues to latent heat, less water is supplied to the atmosphere hence water available for precipitation is lower too ([Kaufmann et al.,] 2007). In both cases, less clouds allow more solar radiation to reach the ground to compensate the loss ([Dirmeyer and Shukla,] 1994). But at the same time less clouds also significantly reduce the amount of downward longwave radiation ([Charney et al.,] 1977). Eventually the net radiation is likely lower and convective clouds as well as precipitation are further reduced ([Falk et al.,] 2005; [Charney et al.,] 1977).

On the other hand, the feedback explains by the effect of surface roughness is shown by changes of boundary layer depth and cross-isobaric moisture convergence ([Sud et al.,] 2001). The surface roughness can affect the vertical mixing of the boundary layer and hence it can determine the stability of this layer ([Gash and Nobre,] 1997). More or less moisture convergence can occur as the patterns of the uprising motion changes ([Pitman et al.,] 2004; [Evans et al.,] 2011). Therefore local climate would change as the surface roughness changes.

The different mechanisms tend to link to certain climate characteristics. The dependence on soil moisture, especially on deep soil moisture is prominent in arid/semi-arid regions ([Kleidon et al.,] 2000; [D’Odorico et al.,] 2007). For regions that oceanic influence is strong, surface roughness plays an important role ([Dickinson and Kennedy,] 1992; [Narisma and Pitman,] 2003; [Pitman et al.,] 2004). Similarly, the albedo effect is likely to be regional, which could affect the temperature difference between ocean and continent and lead to stronger monsoon ([Charney,] 1975; [Dallmeyer and]
In the relationship between the land surface and the atmosphere, multiple mechanisms coexist. Hence the feedback is often complicated. In some cases one or two mechanisms are in the dominant role over the others. For this reason, local situation is important in determining the types of feedback.

2.4 Thesis questions and layout

The aim of this thesis is to investigate the climate impact of vegetation cover change in regional Australia, with a specific interest on areas inside the MDB. In particular, the main focus is on the local precipitation feedback. The thesis is set out to address the following questions:

Q1 Is the feedback relationship related to the type of land cover change, e.g. vegetation clearing versus revegetation?

Q2 Does the feedback relationship have different behaviours between short term and long term?

Q3 What mechanisms can explain the existence/non-existence of the vegetation-precipitation feedback?

These questions will be answered using simple methods instead of large numerical models, although the latter was the main stream of the current land-atmosphere interaction research. The purpose of applying simple methods here is to allow for more accurate tracking and analysis of the basic relationship ([Wainwright and Mulligan](#) 2005). Since there is few observational studies on the vegetation-precipitation feedback in Australia, especially east and southeast Australia, in Chapter 3 an empirical study is carried out to detect step changes in rainfall and related such changes to observed vegetation cover changes. In Chapter 4 sensitivity experiments, which
grassland is replaced by forest and forest is replaced by shortgrass, are conducted using a one-day soil-land surface-atmosphere coupled model and the changes on the convective boundary layer and cloud formation are analysed. In Chapter 5, an equilibrium land-atmosphere interaction model for vegetation changes study is developed and the rainfall sensitivity to various factors, including vegetation properties and boundary conditions, are tested. Discussion and conclusions of this work and suggestions of future work would be given in the final chapter.
Chapter 3

Empirical study of land surface effects on local rainfall

3.1 Introduction

Land use and land cover changes can lead to changes in the local climate. Both empirical and modelling studies have found cloud types and rainfall are correlated to large scale vegetation cover changes, such as deforestation in Amazon (Chagnon and Bras, 2005; Pinto et al., 2009; Wang et al., 2009; Mei and Wang, 2010) and afforestation in south Israel (Otterman et al., 1990; Ben-Gai et al., 1998). Using airborne measurement in Western Australia, Junkermann et al. (2009) showed a significantly higher level of aerosols over an agricultural area compared to adjacent natural vegetation. They suggested a modification of aerosol concentrations due to deforestation could have contributed to the reduction of local rainfall, as there are more but smaller rain droplets. Nair et al. (2011) reported from the Bunny Fence Experiment that the local land use changes have altered the west coast trough dynamics and surface roughness, and this has resulted in the observed rainfall decrease. Climate sensitivity to land cover change is also found in eastern Australia (McAlpine et al., 2007).

Rainfall over land is generally influenced by multiple factors. Locally, there are two main sources which are: moisture from advective atmospheric transport; and local
evapotranspiration (Eltahir and Bras, 1996; Bosilovich and Chern, 2006; Dirmeyer et al., 2009; Gimeno et al., 2010). According to Trenberth (1999), the contribution of advective moisture partially depends on the availability of external moisture and atmospheric transport. On the longer time scale, such as monthly and annually, the large scale atmospheric dynamics are affected by large scale climate drivers. Many studies have found significant relationships between rainfall in large parts of Australia and the El Niño-Southern Oscillation (ENSO) (Verdon et al., 2004; Risbey et al., 2009; Speer et al., 2011). ENSO can be used to represent longer term cycles in rainfall data such as drought. On the other hand, local ET is determined by local land surface characteristics. Local land surface characteristics further influence the local scale atmospheric dynamics and hence the amount of rainfall, including contribution from both sources. Therefore land surface plays an important role in local rainfall.

Conventionally, large scale climate drivers are used as predictors for seasonal rainfall forecasts. According to the Australian Bureau of Meteorology (BoM), seasonal outlooks are “based on the statistics of chance, taken from Australian rainfall/temperatures and sea surface temperature records for the tropical Pacific and Indian Oceans” (BoM, 2012c). Although climate drivers demonstrate some capabilities in predicting Australian rainfall, there is still a large amount of unexplained variance. Westra and Sharma (2010) pointed out that models based on global sea surface temperature anomalies can only predict up to 14.7% of precipitation variance. In addition to random factors, local impacts are expected to explain some of the remaining variance. Pitman et al. (2004) found a good match between observations and simulated rainfall changes in southwest Western Australia forced by land cover change. Timbal and Arblaster (2006) were able to reproduce the rainfall decline in the
south west of Australia by including land cover influence. Local land use change might not be a primary, but is likely to be a secondary cause of rainfall change (Nicholls 2006). Therefore, land surface modification has, at least partially, contributed to local rainfall variability.

The aim of this study is to investigate the cause and effect relationship between land cover change and local rainfall using empirical evidence. I hypothesize that a step change on the land surface will cause a step change in the rainfall. The rainfall data is studied to detect such step change in the time series. Two locations of land cover changes due to land clearing and bushfires are identified based on tree cover data. Two statistical methods are then applied to detect step changes. The rainfall change, if any, is then associated with land cover changes through spatial comparison.

3.2 Data

Several land surface data sets are used in this study. The main one is the MOD44B product Global Vegetation Continuous Field dataset (version 5). This dataset provides estimates of percent tree cover (percentage of ground surface covered by trees) at a resolution of 250 m (Townshend et al. 2011). The dataset is available on an annual basis for the period of 2000 - 2010. The tree cover data is produced from 16-day Terra MODIS Land Surface Reflectance data and Land Surface Temperature (Townshend et al. 2011). The National Dynamic Land Cover Dataset (DLCD) (Lymburner et al. 2010) from the Australian Collaborative Land Use Mapping Program (ACLUMP) is used to verify the trend of vegetation cover change calculated from the previous dataset. This dataset, developed by Geoscience Australia and Australian Bureau of Agricultural and Resource Economics and Sciences (ABARES), is the first nationally consistent and thematically comprehensive land cover reference for Australia. The
DLCD is based on the 16-day Enhanced Vegetation Index (EVI), again from the MODIS satellite, between April 2000 and April 2008. It also has a high resolution of 250 m. The dataset provides information on the final land cover types (as in 2008) and estimated trend of EVI statistics (annual mean, maximum and minimum).

Rainfall data for Australia (Jones et al., 2009) is obtained from BoM. The data is projected onto a national 0.05° × 0.05° grid (approximately 5 km × 5 km). This gridded dataset is generated from station observations using an optimised Barnes successive correction technique. The Barnes technique combines a weighted averaging process and defined topographical information to estimate rainfall values between spatial points (BoM, 2009). The resulting dataset provides additional information for data-sparse areas like central Australia but reduces information in the data-rich areas, such as southeast Australia where station density is up to 20 per 100 km². The data is available on a monthly basis from 1900 to current. Here a subset of 30 years (1979 - 2008) is used. The study is conducted on the monthly data, as land cover change effect on annual rainfall might be negligible but it is often found significant in particular months or seasons (e.g. Otterman et al., 1990; Gaertner et al., 2001; Semazzi and Song, 2001; Oleson et al., 2004; Deo et al., 2009a).

Large scale climate drivers are represented by various climatic indices. The Southern Oscillation Index (SOI) is generally regarded as a good predictor for Australian rainfall (Risbey et al., 2009; Chowdhury and Beecham, 2010; Westra and Sharma, 2010), but its skill is weaker in some parts of Australia. For example the Southern Annular Mode (SAM) is found to be more important than ENSO in south Western Australia (Meneghini et al., 2007). The suitability of each index for the regions of interest is tested in section 3.4.1. The following climate indices are used
as candidate predictor for local rainfall.

- **Southern Oscillation Index (SOI).** The Troup version of the monthly SOI series used in this study is obtained from BoM (available online at [http://www.bom.gov.au/climate/current/soihtm1.shtml](http://www.bom.gov.au/climate/current/soihtm1.shtml)).

- **Eastern, East Central and Central Tropical Pacific Sea Surface Temperatures (NINO 3, NINO 3.4 and NINO 4).** Monthly SST anomalies are available from IRI/LDEO data library and the extended NINO dataset is used (available online at [http://iridl.ldeo.columbia.edu/SOURCES/.Indices/.nino.EXTENDED/](http://iridl.ldeo.columbia.edu/SOURCES/.Indices/.nino.EXTENDED/)).

- **Pacific Decadal Oscillation (PDO)** The monthly PDO series is provided by JISAO (Joint Institute for the Study of the Atmosphere and Ocean, University of Washington) (available online at [http://jisao.washington.edu/pdo/PDO.latest](http://jisao.washington.edu/pdo/PDO.latest)).

- **The interaction of PDO and SOI (PDO×SOI) ([Kamruzzaman et al.](#) 2011).**

- **Indian Ocean Dipole (IOD)** Monthly IOD is obtained from JAMSTEC (the Japan Agency for Marine-Earth Science and Technology) (available online at [http://www.jamstec.go.jp/frcgc/research/d1/iod/DATA/dmi.monthly.txt](http://www.jamstec.go.jp/frcgc/research/d1/iod/DATA/dmi.monthly.txt)).

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1 Pacific Decadal Oscillation is the leading principal component of monthly SST anomaly in the North Pacific Ocean.  
2 Indian Ocean dipole is commonly measured by the difference between SST anomaly in the western (50 - 70°E and 10°S-10°N) and eastern (90 - 110°E and 0 - 10°S) equatorial India Ocean ([Saji et al.](#) 1999).
### Table 3.1: Summary of data used in this chapter.

<table>
<thead>
<tr>
<th>Data</th>
<th>Source</th>
<th>Resolution</th>
<th>Analysis period</th>
</tr>
</thead>
<tbody>
<tr>
<td>Percent tree cover</td>
<td>MOD44B</td>
<td>Annual</td>
<td>2000-2010</td>
</tr>
<tr>
<td>Rainfall</td>
<td>AWAP gridded rainfall data</td>
<td>Monthly</td>
<td>Jan 1979 - Dec 2008</td>
</tr>
<tr>
<td>SOI</td>
<td>BoM</td>
<td>Monthly</td>
<td>Jan 1979 - Dec 2008</td>
</tr>
<tr>
<td>NINO 3, 3.4, 4</td>
<td>IRI/LDEO data library</td>
<td>Monthly</td>
<td>Jan 1979 - Dec 2008</td>
</tr>
<tr>
<td>PDO</td>
<td>NOAA</td>
<td>Monthly</td>
<td>Jan 1979 - Dec 2008</td>
</tr>
<tr>
<td>IOD</td>
<td>POAMA-2 dataset</td>
<td>Monthly</td>
<td>Jan 1979 - Dec 2008</td>
</tr>
</tbody>
</table>
3.3 Study regions and tree cover change

In Australia, significant tree cover change has mainly occurred in the north east of the continent and on the southeast coast, as well as in the southwest of Western Australia. According to DLCD data, most of these areas have experienced decreasing EVI over 2000 - 2008. As an index for vegetation greenness, the decreasing values indicate lower biomass over time in the tree cover regions. The possible EVI reduction might be due to land clearings, bushfires or droughts.

![Map showing study regions](image)

Figure 3.1: Selected study regions are highlighted by red rectangles in the main map (the red rectangle in the insert map indicates the location of the main map). The types of tree cover in 2008 from the DLCD product is shown at the background. In site 1 (the QLD region), the tree cover is mostly sparse. In site 2 (the NSW/VIC region), many areas have open or close forest in which tree cover is denser.

Two regions are selected where significant tree cover change is presented. The first region is located in south central Queensland to the north of the Murray Darling Basin (MDB) (site 1 in Figure 3.1). High rates of land clearing have been reported...
Chapter 3. Empirical study of land surface effects on local rainfall in this region during the early 2000s (Department of Natural Resources and Water 2007). The second study region is located at the border of New South Wales and Victoria, and includes the Snowy Mountains ranges (site 2 in Figure 3.1). Severe bushfires occurred in this area and the surroundings in early 2003 (see Figure 3.2). The 2003 bushfires have been the largest and the worst in this area in the last 60 years (The State Government of Victoria 2011). Two thirds of Kosciuszko national park has been heavily burned out and regrowth is reported to be slow due to drought and cold conditions (ABC News 2003). In both study regions, significant tree cover loss has happened within the last decade, either permanently or temporarily. It makes the selected area suitable for this land cover change study.

Figure 3.2: Location of bushfires occurring in January 2003, in and around the NSW/VIC study region, as shown by the red pixels. The map shows large area in the Kosciuszko national park has been burned. Some locations in the southwest of ACT have also experienced intensive bushfires.

The two regions have different climate characteristics. The QLD region is partially grassland and partially subtropical, while the NSW/VIC region is mainly within the temperate zone, under the Köppen classification. According to BoM, the NSW/VIC
region receives 1000 - 2000 mm rainfall annually, which is more than double of the rainfall in the QLD region. Evapotranspiration is similar in both regions. Marine moisture and orographic effects are likely the main contributors to rainfall in the southeast mountain areas of the NSW/VIC region.

The land use and land cover characteristics in the two regions are also different. In the Queensland region, the tree cover is sparse in most area. The MODIS tree cover data shows that tree cover in this region is mostly below 20% of total ground area. Grazing is the main activity in this region, with over 90% of land used by the grazing industry (ABARES 2010). I assume that EVI has decreased over large part of the region mainly due to land clearing. Tree cover has been cleared at a massive scale over the last decade, especially during 2002 - 2004. The Kosciuszko national park is within the NSW/VIC region. Here tree cover is denser with open or even closed forest (tree cover distribution is bimodal at 10 - 20% and 60 - 70%). The dominant species in the alpine is Snow Gum and large stand species such as Alpine Ash and Mountain Gum in the sub-alpine area. These trees can reach a great height but they take long time to grow. For example, Alpine Ash would need about 20 years to mature. Although land clearing is not the major issue in this region, it is vulnerable to fires and droughts.

Therefore two types of land cover changes are studied. The reports from the Queensland Statewide Land Cover and Trees Study (SLATS) (e.g. Department of Natural Resources and Mines 2005, 2006) are used to investigate the time and location of land clearing in the QLD region. The MODIS burned area product, MCD45A1 (Roy et al. 2002, 2005, 2008), is used to locate bushfires areas in the NSW/VIC region. With a coarser resolution of 500 m compared to MOD44B, MCD45A1 provides monthly burning information on the pixels, which helps to pinpoint an abrupt event. Due to the
nature of the different land cover change, the post-change vegetation status in the two regions are expected to be different (see Figure 3.3).

![Figure 3.3](image)

Figure 3.3: The expected evolution of the land surface after trees have been removed in (a) the QLD region and (b) the NSW/VIC region.

This study focuses on the effect of 2003 - 2004 land clearings in the QLD region and the effect of 2003 bushfires in the NSW/VIC region. These events are expected to cause a step change in the local rainfall. The actual tree cover change at the pixel level during this time is derived from the 11-year MODIS data. The difference between tree cover before and after the land disturbance is tested using a Student’s t-test. As the length of the tree cover data is shorter than the length of the rainfall data, earlier land clearings in the QLD region cannot be identified spatially hence they are excluded from the analysis.

### 3.4 Statistical method

As shown in Figure 3.4 a step change is not obvious in the time series data, even though the data is deseasonalised and detrended. Hence the step changes on rainfall are analysed by two statistical methods to provide a comparison to each other. Both
methods make use of a regression model to remove variability in rainfall due to factors other than vegetation cover changes. In the first method, the tree cover change is implemented as a factor variable in the regression model. In the second method, a rank sum test (step trend test), is applied on the model residuals after effects of other major factors have been removed, assuming that vegetation cover change is the only factor explaining the non-random pattern in the rainfall residuals.

(a) QLD

(b) NSW/VIC

Figure 3.4: The deseasonalised and detrended rainfall over the 30 years period in (a) the QLD region and (b) the NSW/VIC region. The vertical red lines indicate the year of 2003, in which the studied land cover changes occurred. A change in the time series data is not obvious before and after the land cover changes.

3.4.1 Regression model

Rainfall, even a single event, is generally affected by several factors. Many studies suggest the large-scale climate drivers, which are related to the global or regional atmospheric air mass movement, are the dominant controls on the overall rainfall ([Maynard and Polcher 2003, DeAngelis et al. 2010, Holper 2011, Smith and Timbal 2012]). Long term trends are shown in Australian rainfall, although whether they are due to global warming is still under discussion ([Wardle and Smith 2004, Nicholls])
The difference in available radiation and hence ET in different seasons increase the variability in rainfall. The land surface effect on rainfall can be confused by changes in these factors. Here a regression model is used to estimate the amount of variability in rainfall that is due to these important factors and to isolate changes resulting from vegetation cover change.

The Australian climate is influenced by sea surface temperature in the tropical Pacific and Indian Oceans, as well as pressure systems in the Southern Ocean \(\text{(BoM, } 2012\text{a)}\). Different parts of the continent are more or less influenced by each of these climatic drivers \(\text{(BoM, } 2008\text{)}\). Using both SOI and PDO in a prediction model, \(\text{Kamruzzaman }\text{et al., } 2011\) reported that PDO is rarely significant for rainfall stations in the MDB and on the southeast coast of Australia, while SOI is at least significant at the 1% level. In addition to SOI, \(\text{Speer }\text{et al., } 2011\) found that the observed rainfall decrease in the southeast of NSW was linked to an increasingly positive SAM in 1976 - 1998. In southwest WA, the influence of any oceanic indices is small \(\text{(Smith and Timbal, } 2012\text{)}\). \(\text{Risbey }\text{et al., } 2009\) compared five large-scale drivers, including ENSO (measured by SOI and the Tropical Pacific SSTs), IOD, SAM, MJO\(^3\) (Madden-Julian oscillation) and blocking, in relation to Australian rainfall variability. They identified SOI as the most important index among all indices tested for broad parts of Australia (including QLD and NSW/VIC) in almost any season. In this study, up to two important indices from the seven climatic indicators (see section \(3.2\)) are used as the explanatory variables in the model for each study region.

The correlations between rainfall and each climatic index are compared. Rainfall in each study region is first deseasonalised and detrended using the seasonal decomposition function “stl” in R \(\text{ (R Development Core Team, } 2011\text{)}\). The

---

3 MJO is a large scale eastward-propagating wave-like disturbance in equatorial latitudes \(\text{(Risbey }\text{et al., } 2009\text{)}\).
appropriateness of using detrended data has been discussed in Smith and Timbal (2012). The cross-correlations between the deseasonalised and detrended rainfall and the climatic indices are tested using the Pearson’s product moment correlation method, assuming the relationships are linear. Although the optimal technique for exploring the correlation with each index could be different as described in Risbey et al. (2009), the Pearson’s method is applied to all for consistency. Here as PDO describes the multi-decadal SST with lower frequency MacDonald and Case 2005, Zanchettin et al. 2008, Kamruzzaman et al. 2011, instead of 30-year rainfall data, a longer period (108 years, from 1900 to 2008) is used to estimate the correlation with PDO up to lag 24. For the other indices, the 30-year data is used.

Based on the correlation between the climatic indices and rainfall (as shown in Figure 3.5), the following results are found:

- In QLD, the correlation between rainfall and SOI at zero time lags is the highest across all indices, outweighing the other ENSO indicators.
- In NSW/VIC, again the SOI has the highest correlation with rainfall, followed by the IOD. Both occur at the zero time lags.

The above findings are consistent with previous studies. Although some indices are serially correlated with rainfall up to several months, the lag zero events have the most significant correlation coefficients. Concurrent climatic index series are generally found most useful in rainfall prediction (e.g. Risbey et al. 2009, Kamruzzaman et al. 2011). The correlations between the climatic indices and rainfall for each individual season have also been tested. The results are similar as above. So SOI and IOD are selected for the study regions as I am only interested at no more than two indices.

Rainfall in Australia shows strong seasonal patterns Holper 2011, Australian
Figure 3.5: Cross-correlation of the six indices and rainfall in each study region: (a) QLD and (b) NSW/VIC. Where the correlations with PDO are sought, 108-year rainfall data (1900 - 2008) are used. Otherwise, 30-year rainfall data are used. The correlation with NINO 3 is not shown as it is very similar to but weaker than the case of NINO 3.4. The blue dashed lines indicate the 95% confidence interval.
Chapter 3. Empirical study of land surface effects on local rainfall

Bureau of Statistics (2012). For example, the north part of the country is summer rainfall dominant with dry winter, while most of the southern part has a winter rainfall regime. This character is given by the movements of subtropical high pressure system which dominates the Australian climate (BoM 2012b). The seasonal component of rainfall has a periodic pattern so it is better modelled as a smooth term. A spline function is applied on the months series to define a smooth seasonal pattern.

Long term trends in the regional rainfall in some parts of Australia are significant (Hughes 2003; Gallant et al. 2007; Chowdhury and Beecham 2010). In the northern and eastern parts of the continent, increasing rainfall is generally found during the last century (Hughes 2003). The presence of such long term trends may be confused with the outcome of a step change in rainfall. A linear trend term is implemented in the model to remove the long term trend effect.

I assume all the factors are additive components in determining rainfall as in Kamruzzaman et al. (2011). Generally monthly rainfall has skewed distribution so the normality assumption in general linear model could be violated. In this case, the rainfall model is expressed as a generalised additive model (GAM) (Hastie and Tibshirani 1986) with a log link function \( g() \), assuming the data has a gamma distribution (see Figure 3.6).

\[
g(E(R_r)) = \beta_0 + s_1(D_{1,r,SOI}) + s_2(D_{2,r,IOD}) + s_3(Season) + \beta_1Trend + \epsilon_r \quad (3.1)
\]

The bold letters represent the time series vectors. The subscript \( r \) denotes the region, \( r = \text{qld} \) or \( \text{nswvic} \). \( \beta_u \ (u=0, 1) \) are the fitted coefficients in the model. \( s_v \ (v=1, 2, 3) \) are the smooth functions on the climatic indices and the seasons. \( D_{1,r} \) and \( D_{2,r} \) switch on/off the corresponding climatic index in the model as discussed previously.

\[
D_{1,r} = 1 \quad (3.2)
\]
as SOI is used in both regions as rainfall predictor.

\[
D_{2,r} = \begin{cases} 
1 & \text{for NSW/VIC} \\
0 & \text{for QLD} 
\end{cases}
\]  

(3.3)

The linear long term trend in the rainfall data is modelled by \textbf{Trend}=1,2,3...n, where \( n \) is the total number of months in the time series. \textbf{Season} is the seasonal component which is represented by applying the smooth spline function on months. The monthly rainfalls in both regions appear to have a lag 1 autocorrelation, with 0.26 for QLD and 0.33 for NSW/VIC. The SOI and IOD terms are also modelled by the spline function. As the effect of large scale drivers on Australian rainfalls is more likely to be seasonal \cite{Murphy2008, Schepen2012}, the spline function can closely reproduce the high and low impacts of the climatic indices.

(a) QLD

(b) NSW/VIC

Figure 3.6: Distribution of monthly rainfall in the two study regions. By using a Kolmogorov-Smirnov test with shape = 1 and 2.4 respectively, rainfalls in both regions are shown to have a gamma distribution.
3.4.2 Tree cover change as factor variable

Recent studies suggest that land surface processes are important for predicting local rainfall (e.g. [Ma et al., 2011; Zeng et al., 2012]). However, they are mostly based on modelling experiments and little evidence is reported from the observations. One of the main difficulties in observational studies is the lack of continuous monitoring of the land surface variable(s) or even no defined variable that can clearly represent the land surface process. Given the lack of a full picture of the land surface process, a factor variable is used in this study to represent the abrupt land surface change (see Equation 3.5). The change could be a result of either land clearing or bushfires as long as it is permanent or it takes a long time to recover.

In the first method, the tree cover change is used as a predictor in the regression model, representing by a factor variable $LC$. The significance of the coefficient of $LC$, denoted as $\beta'$ in Equation 3.5, can be determined by a ratio test.

$$LC = \begin{cases} 
\text{Trees} \\
\text{Removed}
\end{cases}$$  \tag{3.4}

Therefore in both regions, land cover is “trees” for the period before land cover change and “removed” for the period after the change. Here I simply assume that vegetation cover change has occurred on every pixel. The remaining term $\epsilon_r$ is the amount of rainfall that is attributed to other unspecified factors and random errors. Hence the regression model becomes

$$g(E(R_r)) = \beta'_0 + s'_1(D_{1,r} \text{SOI}) + s'_2(D_{2,r} \text{IOD}) + s'_3\text{Season} + \beta'_1\text{Trend} + \beta'_2 LC + \epsilon'_r \tag{3.5}$$

Vegetation cover changes have occurred at different times in the two regions. In the QLD region, there is no exact timing of the land clearings. Clearings have occurred
within 2003 - 2004 according to SLATS reports. The information on the type of land cover during this time period is missing. Therefore, four scenarios are tested in the analysis, which the after change period is set to start from (1) June 2003, (2) January 2004, (3) June 2004 and (4) January 2005. In the NSW/VIC region, severe bushfires were reported in early January 2003. Hence the “tree” cover state was up to December 2002 then it has changed to “removed” state since January 2003. The regression model is run from 1979 in both regions.

3.4.3 Step trend test

A step trend test is used to detect changes in rainfall after the vegetation cover change. This nonparametric statistical test is modified from the Mann-Whitney Rank-Sum test by Hirsch and Gilroy (1985). The test is developed to look for a step change in the data which possesses cross-correlation. The gridded rainfall dataset is built by an interpolation method and there is high correlation between neighbouring pixels. Hence the step trend test is suitable for the analysis of such dataset. The advantages of this test are: (1) it does not depend on assumptions of the data distribution; (2) it is not restricted to datasets with no missing data; (3) it is robust and not as easily influenced by outliers and negative numbers (Hirsch and Gilroy, 1985).

3.4.3.1 Rainfall residuals

The rainfall residuals from the regression model in Equation 3.1 are used in the test. According to Hirsch and Gilroy (1985), deseasonalised and detrended data is important in a test to detect step change. Furthermore, since rainfall is only partially attributed to local source and conditions, noise can be introduced by large scale dynamics and the changes in other climatic factors. The regression model has also removed such noise.
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Hence the model residual $\epsilon_r'$ excludes effects due to climate drivers, seasonality and the long term trend. Local effects are assumed to at least partially explain the variation in the residuals. The following test associates possible changes in rainfall trends with the tree cover changes.

### 3.4.3.2 Mann-Whitney rank-sum statistic

The step trend test is a modified version of the Mann-Whitney rank-sum statistic ([Hirsch and Gilroy](#1985)). As a nonparametric rank-based test, the Mann-Whitney test does not use the exact values of rainfall but depends on the ranks of the data. For each month, rainfall residuals of each year are ranked in an ascending order. The ranking of January rainfall in a sample pixel k in QLD is illustrated below:

<table>
<thead>
<tr>
<th>Year</th>
<th>Rainfall Residuals</th>
<th>Rank ($R'_{1k}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1998</td>
<td>-0.3</td>
<td>6</td>
</tr>
<tr>
<td>1999</td>
<td>-60.9</td>
<td>2</td>
</tr>
<tr>
<td>2000</td>
<td>-16.1</td>
<td>4</td>
</tr>
<tr>
<td>2001</td>
<td>-71.7</td>
<td>1</td>
</tr>
<tr>
<td>2002</td>
<td>111.1</td>
<td>7</td>
</tr>
<tr>
<td>2005</td>
<td>-7.2</td>
<td>5</td>
</tr>
<tr>
<td>2006</td>
<td>-60.5</td>
<td>3</td>
</tr>
</tbody>
</table>

Therefore, the smallest or most negative value has rank 1 and the largest value has the maximum rank.

The before period and the after period form two groups of samples. The split point of the two periods is defined based on the timing of the vegetation cover changes. In the QLD region, changes occurred anytime during 2003 and 2004. Different from the previous method, the time period of changing is excluded here, as the nonparametric test allows missing data. ([Hirsch and Gilroy](#1985)) also pointed out that the power of the test is higher if the data of the change implementation period is ignored. Hence 2003 and 2004 are excluded from the analysis. The after-change period is 2005 - 2007.

In the case of NSW/VIC, the bushfires had broken out from early January 2003. The change was within a relatively short period of the year. Therefore the after-change
period in this region is still from January 2003. Same as in \cite{Hirsch and Gilroy 1985}, the before period is set to five years (1998 - 2002) in both regions.

The rank of rainfall in month $j$ year $i$ in pixel $k$ is denoted as $R'_{ijk}$. The sum of ranks of rainfall in month $j$ in pixel $k$ before the known intervention is:

$$W_{jk} = \sum_{i=1}^{n_1} R'_{ijk}.$$  

(3.6)

$n_1$ is the number of years before the land cover change. The expected value of $W_{jk}$ is

$$\mu_w = \frac{n_1(n_1 + n_2 + 1)}{2}$$  

(3.7)

$n_2$ is the number of years after the change. Hence the expected value of the rank sum before the intervention is the same for all months and all pixels. The sum of ranks for the whole time period is fixed, as $(n_1 + n_2)(n_1 + n_2 + 1)/2$. In this study, since there are only two groups (before and after), knowing the rank-sum of one group is the same as knowing the rank-sum of the other group. If the rainfall data is temporally and spatially independent, the variance of $W_{jk}$ is

$$\sigma^2_w = \frac{n_1 \cdot n_2(n_1 + n_2 + 1)}{m}$$  

(3.8)

where $m$ is the number of months which is 12 in the case of a full year.

3.4.3.3 Step trend test

Instead of completing the Mann-Whitney U-test, \cite{Hirsch and Gilroy 1985} applied the rank-sum statistics in a standard normal $Z$ test. The modified test can be used to detect step change and it accounts for serial and cross-correlation in the data. In the case here, the deseasonalised and detrended data shows little autocorrelation in the time series but possesses strong cross-correlation between neighbouring pixels, i.e. $R > 0.99$. Hence the covariance between pixels would need to be considered in the test.
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The sum of $W_{jk}$ for a block of $ns$ pixels over the whole year, $\sum_{j=1}^{12} \sum_{k=1}^{ns} W_{jk}$, has mean value of

$$E\left( \sum_{j=1}^{12} \sum_{k=1}^{ns} W_{jk} \right) = 12 \cdot ns \cdot \mu_W$$

(3.9)

and variance of

$$Var\left( \sum_{j=1}^{12} \sum_{k=1}^{ns} W_{jk} \right) = \sum_{j=1}^{12} \sum_{k=1}^{ns} \sum_{h=1}^{ns} C(W_{jk}, W_{jh}).$$

(3.10)

$C(W_{jk}, W_{jh})$ is the covariance of the W statistics between pixel k and pixel h in month j. When $k = h$, $C(W_{jk}, W_{jh}) = \sigma^2_w$. When $k \neq h$,

$$C(W_{jk}, W_{jh}) = \sigma^2_w r(R_k, R_h)$$

(3.11)

where $r(R_k, R_h)$ is the product moment correlation coefficient of the concurrent ranks in pixel k and h. Here $r$ is calculated on the full time series in each pixel. In this analysis, the test is applied on a square block of four pixels each time. As argued by [Hirsch and Gilroy (1985)], $ns = 4$ is the most optimal solution to balance the cost and the gain in the test power, as discussed in [Hirsch and Gilroy (1985)].

The statistic of the step trend test is then defined as

$$Z' = \frac{\sum_{j=1}^{12} \sum_{k=1}^{ns} W_{jk} - 12 \cdot ns \cdot \mu_W}{\sqrt{Var\left( \sum_{j=1}^{12} \sum_{k=1}^{ns} W_{jk} \right)}}.$$  

(3.12)

The above statistic is written for a 12 months period. By changing the value 12, it can also be used to test seasonal rainfall changes or other customized periods.

The null hypothesis ($H_0$) in this study is that there is no change in rainfall due to land surface intervention. The results of the step trend test can be interpreted according to the sign of $Z'$ score (see Table 3.2). $Z'$ is normally distributed as the standard normal statistics $Z$. Hence it can be compared to a standard normal distribution to determine the $p$ value.
Table 3.2: The interpretation of Z’ score in the step trend test. Following Hipel and McLeod (1994, Chapter 23, P887).

<table>
<thead>
<tr>
<th>Z’</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Z’ &gt; 0</td>
<td>rainfall decreases after change</td>
</tr>
<tr>
<td>Z’ &lt; 0</td>
<td>rainfall increases after change</td>
</tr>
<tr>
<td>Z’ = 0</td>
<td>rainfall does not change</td>
</tr>
</tbody>
</table>

3.5 Result

3.5.1 Tree cover changes

The pixels with significant tree cover change in each study region are shown in Figure 3.7 at the 0.05 significance level. Given the limitation of the length of this data series, it still indicates a large change in the NSW/VIC region. In the NSW/VIC region, much of the tree loss between 2002 and 2003 was concentrated in the Snowy Mountains which are at the border of NSW and VIC, as expected. Tree cover loss occurred in large parts of the QLD region between 2002 and 2005. Most of the clearings are shown at the center of this region. The tree cover change map is consistent with the annual mean EVI trend map (based on DLCD data, map not shown here), which confirms these changes are significant and persistent over the study period.

3.5.2 Regression Model & Significance of Vegetation Cover Changes

The regression model does not explain much of the rainfall variability. The model in Equation 3.5 accounts for around 13% of the rainfall variations in the QLD region and 19% in the NSW/VIC region on average. The residual analysis shows that the assumptions of the regression model are generally met. The standardised residual plots, however, show some funnelling for the NSW/VIC regions, suggesting non-

\(^4\) Here the adjusted \(R^2\) was reported. Adjusted \(R^2\) is the coefficient of determination, a measurement of the amount of variability predicted by the model adjusting for the number of explanatory terms.
Figure 3.7: The maps show the areas with significant changes in tree covers in (a) the QLD region and (b) the NSW/VIC region. The amount of changes is calculated as the difference between tree covers before and after the specified land cover intervention and it is shown as the percentage of the ground area. Green colour indicates an increase in tree cover, while red colour indicates a decrease in tree cover.
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constancy of variance. The residual analysis on one sample pixel from each region in Figure 3.8 to illustrate these. The residual patterns are consistent within each region.

The model confirms the importance of the climate drivers and the seasonality in Australian rainfall. Even at the grid level, the seasons and the climatic indices are significant ($\alpha = 0.05$) everywhere in both regions. The explaining power of the model is mostly due to these variables. The climate drivers (at lag zero) alone account for on average 6% of the rainfall variability in the two regions (see Figure 3.9 for the distribution of $R^2$ in these two regions). These figures are within the upper bound of seasonal rainfall predictability by SST anomaly field reported by Westra and Sharma (2010).

Statistically significant long term trend is not observed in these two regions. However it might not disapprove the importance of long term trend in rainfall. More pixels in NSW/VIC would have significant step change if the long term trend effect is not removed by the model. As trend free is an important property of the step trend test, the trend term is kept in the regression model to ensure the detection of step change is not due to a possible long term trend.

The land cover variable implies a step change with different values before and after the land cover intervention. This variable is only significant ($\alpha = 0.05$) for the rainfall estimates in some areas in NSW/VIC, as shown in Figure 3.10(b). The effect of tree removal in the NSW/VIC region is highly significant in the area from the Snowy mountains ranges in the south to the west side of Australian Capital Territory (ACT), highlighted by the red colour in Figure 3.10(b). However no significant step change due to the land cover changes is found in rainfall of the QLD region in any of the four scenarios.
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(a) QLD: analysis of residuals

(b) NSW/VIC: analysis of residuals

Figure 3.8: The residual analysis of a sample pixel in each region.
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Figure 3.9: The distribution of $R^2$ when rainfall is only modelled by the climate drivers. It shows the percentage of rainfall variability that can be explained by the climate drivers. In the case of (a) the QLD region, SOI is the only climatic index considered. In (b) the NSW/VIC region, SOI and IOD are used.

Figure 3.10: The spatial distribution of significance of vegetation cover changes in the two study regions. The significance of vegetation cover changes is assessed in the regression model. The figures report the p values of the coefficient of the land cover variable in the model at different significance level. The plot of the QLD region is for the scenario that the after change period started from January 2005, as the other scenarios have no p value less than 0.2.
In the NSW/VIC region, the locations of rainfall change show some agreements with the locations of vegetation cover change. Compared to Figure 3.7(e), the alpine area with massive tree cover loss is located inside this highlighted region. The results show that step change in rainfall also occurs in the Cotter river catchment which was heavily burned in 2003 bushfires (refer to Figure 3.2). However, Figure 3.10(b) also shows that significant step change in rainfall is found in an area larger than where the bushfires has occurred. From this point of view, the results might be showing a large scale drought effect instead of the vegetation cover change.

The model also shows that the tree cover has a positive impact on rainfall. The fitted coefficients for the reference state “trees” are consistently positive for the pixels whose step changes of rainfall are significant. It implies that rainfall was higher when the surface was covered by trees. This is confirmed by applying a two sample Student’s t-test on the annual rainfall of the before-change and after-change periods.

### 3.5.3 Step Trend Test

The step trend test $Z'$ scores of each individual pixel are shown in Figure 3.11. The figure provides two types of information: the sign and the significance level. The sign indicates the direction of the step change, as listed in Table 3.2. In each region, there is a broad area of positive $Z'$ values which implies a decrease in rainfall. In QLD, the areas of positive $Z'$, especially when $\alpha$ is smaller than 0.1, almost align with the locations reported heavy land clearing (see Figure 3.12). However, the vegetation cover change map in Figure 3.7(d) does not clearly show a similar pattern. In the Snowy Mt area and the west of ACT, where severe bushfires had occurred in 2003, positive $Z'$ values are found. It indicates that loss of tree cover might be related to
Figure 3.11: Spatial distribution of the step trend test $Z'$ statistics in the two study sites. The test is conducted on the 30-year rainfall data from 1979 to 2009. Warm colours (yellow, orange and red) are for positive $Z'$ values which indicate decreasing rainfall trend due to the land surface intervention. Cold colours (light blue to blue) are for negative $Z'$ values which indicate increasing rainfall trend. The deeper the colour, the more significant the statistic.
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rainfall decrease. Nevertheless the step change in rainfall is not strong. Changes at the 5% significance level are found in a number of pixels in the alpine area where might be the place that the tree recovery is slow. In the QLD region, 35 pixels obtained a $Z'$ score with $p < 5\%$. In the NSW/VIC region, there are 18 pixels in the alpine area having the $Z'$ score with $p < 5\%$. They are a very small proportion of the whole study regions.

Figure 3.12: Woody vegetation clearing rate in QLD. The maps and figures are obtained from [Department of Natural Resources and Water (2007)](#). The red rectangle is the boundary of the QLD study region.

The rainfall residuals of the two periods (before-change since 1998 and after-change) are also compared (Figure 3.13). The annual values of the 11-year (1998 - 2008) are calculated for each pixel. The boxplots indicate that there are different mean values between the “before” and “after” periods in each region. There are small outliers in the after period in NSW/VIC. A close look at the rainfall data reveals that rainfall was consistently very low in 2006 for the whole region. The low rainfall in 2006 was due to a weak El Niño and unrecovered conditions from the previous drought. While
Figure 3.13: Boxplots of annual rainfall (estimated based on rainfall residuals from Equation 3.1) before and after the land cover interventions during 1998 - 2008 in the study regions. On average, the after period has a lower annual rainfall amount and with outliers of small values.

the regression model has removed the effect of ENSO, the previous drought effect may still remain in the residuals. This can also explain the significant results in the previous method. To remove the outlier effects, the two periods are compared excluding the 2006 rainfall. An F test confirms that the variances are different between the before and after periods. Hence an unpaired unequal variance two sample Student’s t-test was applied to the group of pixels showing a negative step change in rainfall (\( \alpha = 0.05 \)) and the group of pixels with no change, respectively. The rainfall in the after period is lower than the before period (\( p < 0.05 \)) for pixels showing a negative step change. For those pixels without a step change, there is no statistical evidence that the two periods
have different mean values. Hence the t-test results are consistent with the step trend test results.

The choice of \( n_s \) has some impacts on the test results, as shown by Hirsch and Gilroy (1985). The cases of \( n_s = 1 \) and \( n_s = 9 \) are also tested. When \( n_s = 1 \), 18 pixels in the QLD region obtain a Z’ score at the 5% significance level. In the NSW/VIC region, the detection of negative step change (\( \alpha = 0.05 \)) reduces to 16 pixels. On the other hand, when \( n_s = 9 \), the results are similar to the case of \( n_s = 4 \). The changes (\( \alpha = 0.05 \)) are detected in 35 pixels in the QLD region and 18 pixels in the NSW/VIC region. The power of the test does not change much after \( n_s = 4 \), as shown by Hirsch and Gilroy (1985). The cause of the highly inconsistent results in the QLD region between \( n_s = 1 \) and \( n_s = 4 \) was unclear.

The “field significance” of the test is considered to make inferences about the step change at regional scales from multiple local tests (Wilks, 2006; Westra et al., 2013). Here I adopt the bootstrapping resampling method from Westra et al. (2013) to evaluate the field significance. The spatial structure of the pixels is maintained, while the orders of years and months have been changed by random resampling. The test statistics are the percentage of the pixels with significant step change, positive and negative respectively, for the step trend test. The test statistics on 1000 resampled replicates are used to draw the distribution of these percentage values under the local null hypothesis that there is no step change.

The bootstrapping resampling technique is applied on both \( n_s = 1 \) and \( n_s = 4 \) cases. In the \( n_s = 4 \) case, a spatial moving block bootstrapping is used, in which the change of time sequence within the \( 2 \times 2 \) block is consistent. The distribution of the test statistics is highly concentrated at zero with a skew to the right. In all four outputs,
positive step changes and negative step changes in the case of $n_s = 1$ and $n_s = 4$, the test statistics on the observed time series fall within the null hypothesis distribution (as shown in Figure 3.14 and Figure 3.15). The results show that the chance of detecting a rainfall change is small. Again there is no strong evidence to support the hypothesis that vegetation cover change can affect local rainfall.

Figure 3.14: Percentage of pixels showing statistically significant positive $Z'$ values (left) and negative $Z'$ values (right) in QLD. The histograms show the distribution of results from 1000 bootstrap resampling of the rainfall time series. The red dots represent the results from the observed data.
Figure 3.15: Percentage of pixels showing statistically significant positive $Z'$ values (left) and negative $Z'$ values (right) in NSW/VIC. The histograms show the distribution of results from 1000 bootstrap resampling of the rainfall time series. The red dots represent the results from the observed data.

3.6 Summary and Discussion

Generally, empirical studies on LCC-precipitation interaction are conducted within the area with known land surface intervention (e.g. Otterman et al. [1990]; Durieux et al. 2003; Negri et al. 2004; Sato et al. 2007). In this study I test a broad region rather than specific locations where known changes have occurred. The advantage of this approach is that it does not require a long time series of land cover data which is usually
unavailable. Furthermore, it does not assume a certain relationship between vegetation cover change and rainfall but allows the data to show this relationship, by applying the analysis to a broader area outside the boundary of the vegetation cover change. This approach is expected to provide a way to reduce the risk of false positive paradox, by comparing results between areas with and without vegetation cover change.

Parametric tests are generally more powerful than nonparametric test in detecting a trend, when the data is normally distributed (Onoz and Bayazit, 2003; Kundzewicz and Robson, 2004). As a non-parametric test, the step trend test has the advantages of distribution free and no restriction on missing data (Hirsch and Gilroy, 1985). This is particularly useful in rainfall analysis since rainfall data is usually skewed. On the other hand, the disadvantages of a non-parametric test, such as limited to hypothesis testing and weaker in power, are also hold for the step trend test (Whitley and Ball, 2002).

The regression model used here is a very simple model. Only the important effects of the historical trend, seasonality and climate drivers are considered. Furthermore, no more than two large scale climatic indices are used, in order to avoid the problems of over-parameterisation and multiple cross-correlations between climatic indices. A purpose of the regression model is to remove the variability in rainfall that is due to these known important factors. The model shows that seasonality, ENSO and IOD together explain no more than 20% of the rainfall variability, with around 6 - 8% on average attributing to the climate drivers. This is consistent with the literature (e.g. Westra and Sharma, 2010). A large amount of variation is left in the model residuals, which is likely due to some random factors. Rainfall is generally considered as a stochastic process (e.g. Fowler et al., 2005; Cowpertwait et al., 2009; Burton et al., 1989).
The high variability increases the difficulty in detecting a change in rainfall.

There is some but no strong evidence that a step change has occurred in the QLD region. The semi-parametric model cannot identify any step change in the rainfall data. On the other hand, the step trend test suggests a couple of locations (300 - 500 km²) might have experienced rainfall change ($\alpha = 0.05$). The results from the step trend test also indicate a possible widespread change, but the spatial pattern of change is not consistent with the tree cover change map. Although land clearings in QLD have occurred at a high rate and broad scale (Department of Natural Resources and Water, 2008), this study cannot confirm its impact on the local rainfall. Similarly, LCC is not found to change the mean rainfall in this location significantly in some other studies (e.g. Narisma and Pitman, 2003; McAlpine et al., 2007). The characteristics of vegetation cover changes in QLD might increase the difficulties in detecting a step change. QLD has a long history of land clearing. According to the series of SLATS reports on land cover changes in QLD released by the Queensland government, land clearing has continued in and around the study region within period of 1988 - 2008. Major broad scale and high rate clearings occurred in 1999 - 2000 and 2002 - 2004. It is difficult to define a clear cut change in this region. The ongoing land clearing could have reduced the significance of a step change.

The results from the two methods are quite different on the level of changes in the NSW/VIC region. The semi-parametric model shows that a large area in the NSW/VIC region has experienced significant step change ($\alpha = 0.05$) in rainfall after 2003. But this result is likely to be affected by the low rainfall in 2006 as the study area is larger than the bushfires region. The step trend test is able to detect significant changes ($\alpha = 0.05$) in about 400 km² areas in the Snowy Mt. and some possible changes along
the bushfires region ($\alpha = 0.1$). This might be due to the possible small size of the step change and/or a short after-fire period \cite{Hirsch and Gilroy, 1985}. The vegetation cover change due to bushfires might have possibly changed the rainfall but the evidence is not strong either.

The bushfires locations highlighted in the analysis results is an interesting outcome. The results might be due to a drought effect. The 2002 - 2003 drought has affected rainfall in a large part of Australia, particularly the MDB \cite{Nicholls, 2004}. The severe bushfires in 2003 are also triggered by the extreme drought conditions. Although the dry episode effect on rainfall has been accounted for in part by the SOI, further impact of drought could be passed through the local land-atmosphere interaction. However, this impact might not be statistically significant, even under an extreme event such as bushfires. Overall, the rainfall feedback to the vegetation cover change could be weak under the dry conditions. I have found similar results in the later chapters.

The different causes of vegetation cover change in these two regions could lead to different post-change characteristics. The magnitude of EVI decreasing trends in the QLD region are less than in NSW region, as reported in the DLCD data. This is due to the lower tree density in the QLD region than in the NSW/VIC region before land surface interventions. Wild fires might have totally damaged the vegetation cover and recovery is very slow in some areas within the Snowy Mt. The persistent drought in the 2000s \cite{Howden, 2012} has delayed the regrowth of trees. On the other hand, replacing tree cover with pasture and crops might have a relatively subtle impact on the EVI. The regrowth of trees can be expected after fires or as drought condition is relieved; but clearings for agricultural purpose impose permanent or semi-permanent changes to the land surface.
Chapter 3. Empirical study of land surface effects on local rainfall

The rainfall data used in this study is a gridded data set. This data set is robust and consistent over a long time series (from 1900 to current) and has a broad national wide coverage which can provide more spatially information. However, high cross-correlation between pixels due to the interpolation method generating this data set can also introduce spatial noise. Here the cross-correlation has been accounted for in the step trend test. Some other methods are also available which can be used to perform a comparative trial. For example, Narisma et al. (2007) applied a spatial Gaussian filter on a similar data set and used wavelet analysis to detect step change in rainfall. High quality station data is another option to test whether the observed spatial pattern in the step trend test results is not due to the gridded data itself. Resampling methods, such as bootstrapping and permutation (Wilks, 1997; Kundzewicz and Robson, 2004; Westra et al., 2013), can also be used to further assess the strength of significance of results and incorporate spatial and temporal patterns in the analysis. I am also aware that the gridded data set is most useful in regions with sparse rain gauge networks but it actually reduces information where the rain gauge density is high (Jones et al., 2009). In the Snowy Mt. areas, the coverage of rainfall stations is intensive but they are mainly located in the valleys. The interpolated data might not best represent the local rainfall.

The current study cannot provide strong evidence to reject the null hypothesis (no step change in rainfall is due to the tree cover loss). Limited by the available data, the time frame under study has been chosen within a long lasting drought period (Holper, 2011). The strong impact of this prolong drought might have suppressed the land-atmosphere interaction and confused the cause-effect relation between rainfall and vegetation cover changes. This could be one of the reasons that the LCC effects
found in other studies (Görgen et al., 2006; McAlpine et al., 2007, e.g.) are not found to be significant here. So possible future work can be carried out for non-drought period, when longer series of land cover data become available. This approach should be further assessed in different scenarios of wet periods and/or afforestation situations. Continuous monitoring of land surface conditions is important for future research. The power of the test can also be improved with the longer length of the after-intervention period (Hirsch and Gilroy, 1985). But then the question is what is an appropriate length for land cover change study, when ongoing change is possible, such as tree regrowth or cropping. This will require further research.

3.7 Conclusion

In this study, I have found some evidence, although not strong, that vegetation cover change has changed local rainfall. The semi-parametric method and non-parametric method did not reach an agreement on detecting significant step change in rainfall in the hot dry QLD region where land clearing has occurred. On the other hand, the bushfires in the humid temperate mountain range in the NSW/VIC region could have reduced rainfall. But the dry spell also plays an important role in the results.

Drought has had a pronounced impact on the land surface conditions during the study period, leading to significant reduction in vegetation and extreme events such as bushfires. The lack of rainfall together with high temperature may cause a step change in the vegetation. Hence, the signal of LCC feedback on rainfall is probably weaker under such regional dry condition, as the impact of LCC on rainfall is mainly through changes in moisture convergence (Görgen et al., 2006; Pitman and Hesse, 2007).
Chapter 4

Comparison of boundary layer development over forest and shortgrass

4.1 Introduction

4.1.1 Rainfall formation

Rainfall formation is a complex process. It includes several stages: rising of moist air, saturation, condensation, forming of droplets and descending of droplets (Shelton, 2009, Chapter 4, P76). These form a cycle of water transportation, for which the fundamental element of rainfall is water. Although local rainfall is sourced from both the external moisture influx and local ET (Eltahir and Bras, 1996; Bosilovich and Chern, 2006; Dirmeyer et al., 2009; Gimeno et al., 2010), the actual amount of moisture that eventually condenses and rains out is determined by the atmospheric dynamics (Trenberth, 1999) and mainly the vertical atmospheric motion.

The first stage of rainfall formation, which is the rising of moist air, is part of the vertical atmospheric motion. According to Shelton (2009, Chapter 4, P77), there are mainly three types of mechanisms to trigger the rising of air. They are: convective lifting; frontal lifting; and orographic lifting. Among the three, convective rainfall
has the highest potential to be impacted by land surface conditions; hence it is most interesting to us. Furthermore, land surface driven warming/cooling could affect cloud cover and water recycling in continental regions, but mainly alter sea breezes under maritime conditions (van der Molen et al., 2011).

Convection describes the process of rising air parcels which are less dense than the surroundings (Dingman, 2008, Chapter 4, P98). During the day, the ground surface is heated up by solar radiation. This heating is uneven across the land surface. Hence lower layers in contact with the relatively hotter surface are warmer and have lower density than the surroundings and, more important, than the air above it. As a result, the lighter air rises and cools. The rate of air temperature changes with height is called “lapse rate”. When no condensation occurs as part of the cooling, it is classified as the “dry adiabatic lapse rate”. The warm air parcels keep rising until its temperature is reduced to the same as the surroundings. This uplift process is frequent during the day time and forms a well mixed layer above the ground. Sometimes, if an air parcel contains sufficient moisture, water vapour might start to condense before it reaches the top of the mixed-layer. After condensation, the air parcel still rises but follows the “moist adiabatic lapse rate” which is the same as the environmental lapse rate (Dingman, 2008, Chapter 4, figure 4-7, P103). In the latter case, the air parcel would have enough energy to finish the last four stages of rainfall formation.

Convection occurs within the lowest part of the atmosphere which is the planetary boundary layer (PBL). Hence this is also called a convective boundary layer (CBL). The height of PBL is usually 1 - 2 km during day time hours depending on the latitude and falls to a couple of hundred meters in the evening. The structure of diurnal PBL is illustrated in Stull (1988, Figure 1.1). The lowest part of PBL is the surface layer which
Chapter 4. Comparison of boundary layer development over forest and shortgrass

is usually considered to be 10% of the PBL. This layer is in direct contact with the land surface. The air in this layer can be quickly warmed up or cooled down depending on the land surface temperature. Therefore the air temperature in this layer is generally higher than the upper layer during day time and lower during night time which can be seen in atmospheric sounding profiles.

Above the surface layer is the mixed layer. This is the main part of the PBL, which accounts for about 70 - 80% of the PBL. As mentioned above, during the day when the ground surface is heated by solar radiation, thermal activities in the form of convection and turbulence are active, which is the reason for the mixed layer to form (Kim and Entekhabi, 1998a; Stevens, 2004). In the evening, the mixed layer is replaced by a residual layer which might possess the properties of the mixed layer, such as temperature and humidity, but lacks convection and turbulence.

At the top of the mixed layer, the atmosphere forms a shallow entrainment layer or cloud layer. It is also called capping inversion where it marks a distinct difference between the mixed layer below and the free atmosphere above (Kim and Entekhabi, 1998a). This layer prevents heat and humidity in the PBL escaping to the free atmosphere. In the free atmosphere, conditions are stable and vertical motion is rare.

Since convection is a process that transports heat and moisture from the surface to the upper atmosphere and contributes to rainfall generation, surface conditions are expected to affect the convection and the rainfall process. In this study a one-dimensional land-atmosphere coupled model (van Heerwaarden, 2010) is applied to study the impact of land surface properties on local convection and boundary layer development. In the later part of this chapter, the term CBL is used in order to emphasize the daytime boundary layer. The model was developed and tested under
Chapter 4. Comparison of boundary layer development over forest and shortgrass

the European conditions and has been applied in the Netherlands and Niger \cite{vanHeerwaarden et al. 2010}. Here the model is applied to Australia using local initial and boundary conditions, and then a sensitivity experiment is conducted to test the effect of vegetation change on atmospheric conditions. Data from eddy covariance flux sites and weather stations in the Murray Darling Basin (MDB) are used to validate the model under Australian conditions. The main strength of this model is that it can simulate the convection and entrainment processes well with relatively simple formulations. This will result in a better understanding of the surface contribution to the mixed boundary layer and the upper atmosphere where clouds and rain form.

4.2 CLASS Model

The CLASS model (Chemistry Land-surface Atmosphere Soil Slab model) was initially developed by \cite{vanHeerwaarden 2010} in his PhD thesis and further developed by Jordi Vilà-Guerau de Arellano \cite{de Arellano and van Heerwaarden 2012}. The current model couples the atmosphere, land surface, soil and chemical reactions together with options to switch off particular components. The benefit of using this model is the local convection can be studied without a high demand on computation. In this study, I am interested in the heat and moisture transfer in the land-atmosphere interaction, which was the main part developed in \cite{van Heerwaarden 2010}. Hence the chemical reaction component is not used.

According to \cite{van Heerwaarden 2010}, the coupled model consists of four sub-models:

- radiation model: includes a set of state equations to calculate energy fluxes and the energy balance;
• convective boundary-layer model: includes a set of prognostic equations to simulate the mixed layer development, such as change of boundary layer height, air temperature and humidity;

• surface model: in this case, surface refers to the lowest part of PBL, i.e. the surface layer as mentioned previously. Surface layer processes such as turbulent fluxes and ET are estimated using state equations, except the canopy water depth used in the calculation of canopy evaporation;

• force-restore soil model: the soil is separated into two layers. The upper soil properties such as temperature and water content are estimated by prognostic equations. The deeper soil layer can be initialized and is assumed to be constant afterwards.

The main variables in each model and the relationship between models in terms of the variables used by other models are shown in Figure 4.1.

The model is designed for the simulation of daytime dry conditions. The model is not coupled to any rainfall model or neither does it use a rainfall input. There is no runoff and no infiltration process in the surface layer model or soil model. Hence the water balance decreases gradually in the system due to ET, convection and dry air entrainment from the free atmosphere. Furthermore, although a cloud factor can be predefined and low level shallow clouds can be simulated when the chemical component is switched on, cloud feedback is not fully included in the model. Therefore the model is more suitable for a cloudless dry day (van Heerwaarden et al., 2010). Conditions at night time are used to initialize the model. However, since the planetary boundary layer is shallow and stable in the evening, land-atmosphere interaction is weak at night. Therefore, evening dynamics are not of interest.
Chapter 4. Comparison of boundary layer development over forest and shortgrass

Figure 4.1: A simple structure of the CLASS model. Each large box represents one sub-model. The main variables estimated by each sub-model are shown in the inserted boxes. See Table 4.1 for description of the variable symbols. The arrows indicate which variables are passed from one sub-model to another and hence the four sub-model components are linked.

The model provides some information on the cloud prediction. It contains a vertical component to estimate the atmosphere profile at specified time steps. Based on the knowledge of the CBL temperature and humidity, the position and timing of the lifting condensation level (LCL) within the CBL can be calculated to infer cloud formation \cite{Nair2011} and convective precipitation. However, this estimation is based on a strong assumption that the vertical temperature changes are purely dependent on height with very limited information on the entrainment layer. The actual formation of cloud and especially precipitation is a more complex problem which involves microphysical processes \cite{Gao2006, Strangeways2007}. Interpretation of model results on LCL and cloud formation therefore requires careful considerations.
Table 4.1: List of parameters in CLASS model.

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>$R_n$</td>
<td>Net radiation which is the difference between incoming and outgoing radiation</td>
</tr>
<tr>
<td>$S_{in}$</td>
<td>Incoming solar (shortwave) radiation</td>
</tr>
<tr>
<td>$S_{out}$</td>
<td>Outgoing solar (shortwave) radiation</td>
</tr>
<tr>
<td>$L_{in}$</td>
<td>Downward longwave radiation</td>
</tr>
<tr>
<td>$L_{out}$</td>
<td>Upward longwave radiation</td>
</tr>
<tr>
<td>$\theta$</td>
<td>Potential temperature in the convective boundary layer</td>
</tr>
<tr>
<td>$\Delta \theta$</td>
<td>Potential temperature jump at the entrainment layer</td>
</tr>
<tr>
<td>$q$</td>
<td>Humidity in the convective boundary layer</td>
</tr>
<tr>
<td>$\Delta q$</td>
<td>Humidity jump at the entrainment layer</td>
</tr>
<tr>
<td>$h$</td>
<td>Boundary layer height</td>
</tr>
<tr>
<td>$T_a$</td>
<td>Specific air temperature in the convective boundary layer</td>
</tr>
<tr>
<td>SH</td>
<td>Sensible heat flux</td>
</tr>
<tr>
<td>LE</td>
<td>Latent heat flux</td>
</tr>
<tr>
<td>$T_s$</td>
<td>Surface layer specific air temperature</td>
</tr>
<tr>
<td>$T_{soil}$</td>
<td>Soil temperature (upper layer: $T_{soil1}$; deeper layer: $T_{soil2}$)</td>
</tr>
<tr>
<td>$w_1$, $w_2$</td>
<td>Soil water content at upper soil and deeper soil respectively</td>
</tr>
</tbody>
</table>

### 4.3 Study area and data

In this study, eddy covariance flux measurements from a grassland site and a forest site are used. The sites are both inside the NSW MDB region. The grassland site is in the Kyeamba catchment about 20 km to the southeast of Wagga Wagga, NSW. In 2006, the National Airborne Field Experiment (NAFE’06) was run in the Murrumbidgee catchment, including Kyeamba, from 29 October to 20 November. The objective of this experiment is to provide mapping for near-surface soil moisture using remote sensing techniques ([Walker et al.](#) 2006). Various other measurements, including heat

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1 Land use in this area is dominated by cattle grazing ([Merlin et al.](#) 2008).
fluxes, temperature and humidity, are also obtained from Kyeamba (Merlin et al. 2008). The forest site is located approximately 70 km to the southeast of Kyeamba, within the Bago State forest in Tumbarumba. This site is part of the Australian and New Zealand Flux Research and Monitoring project. The forest is dominated by Eucalyptus delegatensis with an average height of 40 m. This site was first established in March 2000. Continuous monitoring is managed by CSIRO. In this study, the flux data is provided by Dr Eva van Gorsel from the CSIRO Marine and Atmospheric Research team. Further descriptions of the two sites are provided in Table 4.2.

The flux measurements provide information on the time dimension, for example air temperature at the measurement height every one hour or even half hour. On the other hand, the mixed layer profile at a particular point in time can be constructed from the radiosonde measurement. Radiosonde soundings provide information on the vertical dimension, for example, air temperature at different heights from the surface up into the free atmosphere. However, radiosonde sounding data is not widely available due to the high cost of collecting this information. There are limited radiosonde launch sites in Australia. Currently only 46 stations are available national wide (personal communication with Larry Oolman, University of Wyoming, April 2012). The Wagga Wagga meteorologic station is the closest point of radiosonde measurements to the Kyeamba eddy covariance flux site. Given the two locations have similar land covers (see Figure 4.2) and elevation (232 m and 212 m respectively), their atmospheric conditions are assumed to be the same. There is no sounding data available for the forest site in Tumbarumba. The atmospheric sounding data can be downloaded free from the website of the Department of Atmospheric Science, University of Wyoming (http://weather.uwyo.edu/upperair/sounding.html).

<table>
<thead>
<tr>
<th>Site</th>
<th>Kyeamba$^a$</th>
<th>Tumbarumba</th>
</tr>
</thead>
<tbody>
<tr>
<td>Location</td>
<td>147.53°E, 35.32°S</td>
<td>148.15°E, 35.66°S</td>
</tr>
<tr>
<td>Elevation [m]</td>
<td>232</td>
<td>1200</td>
</tr>
<tr>
<td>Surface cover</td>
<td>short grass</td>
<td>open forest</td>
</tr>
<tr>
<td>Köppen-Geiger climate classification</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Main</td>
<td>warm temperate</td>
<td>warm temperate</td>
</tr>
<tr>
<td>Precipitation</td>
<td>fully humid</td>
<td>fully humid</td>
</tr>
<tr>
<td>Temperature</td>
<td>hot summer</td>
<td>warm summer</td>
</tr>
<tr>
<td>Rainfall [mm/year]</td>
<td>650</td>
<td>1000</td>
</tr>
<tr>
<td>Potential evaporation [mm/year]$^b$</td>
<td>1500</td>
<td>1200</td>
</tr>
<tr>
<td>Measurement height [m]</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Air temperature</td>
<td>2</td>
<td>70</td>
</tr>
<tr>
<td>Humidity</td>
<td>2</td>
<td>70</td>
</tr>
<tr>
<td>Radiation</td>
<td>1</td>
<td>70</td>
</tr>
<tr>
<td>Soil heat flux &amp; temperature</td>
<td>-0.08</td>
<td>70</td>
</tr>
<tr>
<td>Soil moisture</td>
<td>2 depths</td>
<td>8 depths</td>
</tr>
<tr>
<td>Measurement frequency [hour]</td>
<td>0.5</td>
<td>1</td>
</tr>
</tbody>
</table>

$^a$ Site K10 in NAFE’06 ([Walker et al., 2006](#)).

$^b$ [Guerschman et al., 2008](#) Table 2-2
Figure 4.2: The locations of the radiosonde station and the eddy covariance flux sites as indicated by blue dot and purple triangles respectively. The background map shows the surface cover in these and surrounding regions. The land cover is classified by ACLUMP. The land cover shown is the 2008 extent from DLCD data \cite{Lymburner2010}. The Tumbarumba flux tower is within the open forest. The Kyeamba site is mainly covered by rainfed pasture or low density savannas. The radiosonde station in Wagga Wagga, with a relatively small distance to Kyeamba, has the same land surface features as the eddy covariance flux site. These two locations have similar elevation; hence I assume the atmospheric conditions are the same in these two sites.

4.3.1 Sounding profile

The radiosonde measurements are taken once or twice per day, at UTC 00Z\textsuperscript{2} and 12Z. The first snapshot corresponds to a local daytime profile and the latter one is for the evening at the local time. In most cases, only the sounding data at UTC 00Z is available. For the purpose of this study, which is to investigate the mixed layer

\textsuperscript{2} “UTC” is the abbreviation for Coordinated Universal Time. The Australian Eastern Standard Time (AEST) is 10 hours ahead of UTC. For example, 01 July 0000 (or 00Z) UTC is local time 01 July 10:00 AEST in Wagga Wagga. During summer time, the difference between UTC and local time in Wagga Wagga increases to 11 hours.
Chapter 4. Comparison of boundary layer development over forest and shortgrass development due to local convection forcing, the daytime snapshot is used. Although there is only one atmosphere profile per day, the sounding data provides sufficient information of the upper atmosphere to initialize the model. As stated previously, the free atmosphere is relatively stable over time. Hence its properties can be estimated from information obtained anytime during the day. Furthermore, the single mixed layer profile can still be used to validate the model.

The CLASS model is ideal for a dry day situation which is classified as “cloudless” (van Heerwaarden et al., 2010) and no rainfall during the day. This is because no cloud factor is included in the radiation component of the model. The existence of cloud would greatly reduce incident solar radiation and hence increase errors in the estimated energy balance. Furthermore, soil moisture is initialized without additional input (no rainfall or recharge from ground water) during the simulation. According to BoM’s records, no rain was reported in Wagga Wagga and Tarcutta between 26 October and 10 November, 2006, except from 02 November to 4 November. At both stations, 10 mm rain was recorded on 03 November and it rained almost every day during 12 - 16 November. Tarcutta is another weather station which is less than 10 km from the Kyeamba site. The same rain days are reported from Tumbarumba. It appears that the summer rains on these days were due to large scale climate phenomena. Nevertheless according to BoM, these areas received very much below average rainfall between May and December in 2006 due to a weak El Niño.

The atmospheric profiles of potential temperature and humidity in Wagga Wagga during 30 Oct - 11 Nov are shown in Figure 4.3. These days were mainly dry except 03 November on which rainfall was recorded at 9am. During these days, potential air temperatures at 11am local time were between 290 - 300 K and mainly around

\footnote{No rain was recorded at 9am on 11 November.}
Chapter 4. Comparison of boundary layer development over forest and shortgrass

Figure 4.3: The vertical profiles of (a) potential temperature and (b) humidity within the lower 4 km of the atmosphere. The radiosonde soundings are collected in Wagga Wagga at 11am local summer time.

294 K. Most days featured a boundary layer height of 1,000 - 1,500 m and a free atmosphere with steadily increasing potential temperature. At the higher level of the free atmosphere, the variation between potential temperatures on different days was smaller and the positive lapse rates of temperature were similar, indicated by the slope of the lines in the figure. These observations support the assumption of a stable free atmosphere. A comparison between the humidity profiles (Figure 4.3(b)) implies the atmosphere was wetter after the rain on 03 November (the blue spectrum) than before the rain days (the red spectrum). Above the convective boundary layer, humidity was decreasing with similar trends for all days.

4.3.2 Flux data

The measurement of CO$_2$, water vapour and energy exchange between the terrestrial biosphere and atmosphere using the eddy covariance method had not started until the
1980s, when the necessary technology became available \cite{Baldocchi2001}. Eddy flux towers equipped with devices using new technology have been built in various locations in Europe and America since the end of the last century. Soon a global network of these micrometeorological flux tower sites was established to incorporate activities and integrate data which provides easy access to the research society and even general public \cite{Baldocchi2001}. Today, flux data are widely used in the land-atmosphere interaction studies \cite{Hasler2007, Ma2009, Cho2012}. It provides frequent (usually half hourly or hourly) and continuous information of soil, land surface and local atmosphere conditions for research use.

Sensible heat fluxes and latent heat fluxes are measured as part of the primary flux data collection. Figure 4.4 shows the measured sensible heat fluxes and latent heat fluxes during 04 - 14 November 2006 at the two sites. However, heat flux measurements from eddy covariance are usually underestimated and the estimated energy balance is not close. This is still an unresolved problem in the eddy flux measurements \cite{Wilson2002, Barr2006, Foken2008, Sanchez2009}. Here the heat flux time series used in the model and analysis are adjusted for the heat balance using a statistical method following \cite{Panin1998}. Theoretically, the heat balance equation is expressed as

\begin{equation}
R_n = SH + LE + G
\end{equation}

$G$ is ground heat flux. $R_n$ and $G$ are also measured at the flux station and assumed to be reliable. \cite{Panin1998} pointed out that the imbalance relationship in measured fluxes indicates the existence of a “systematic error” and a “statistical error”, denoted by $\Delta$ and $\delta$ respectively. With the error terms, Equation \eqref{eq:heat_balance} becomes

\begin{equation}
R_n = SH + LE + G + \Delta + \delta
\end{equation}
Figure 4.4: The hourly heat fluxes measured at the two flux tower sites in (a) Kyeamba and (b) Tumbarumba. The circles denote the sensible heat and the triangles denote the latent heat. The flux data shows abnormal latent heat on Julian day 316 - 318 in Kyeamba and on Julian day 316 - 317 (Julian day 316 is 12 November 2006) in Tumbarumba, corresponding to the rain days in BoM’s rainfall record. The crosses indicate the residuals in the heat balance (see Equation 4.1). The residuals are sometimes larger than the latent heat and the sensible heat as one has been recorded as negative value. In this study, I am interested in the daytime conditions when the net radiation is positive so some large negative fluxes are not shown.
The relationship between the residuals in the energy balance and $R_n$ is approximately linear (see Figure 4.5).

Figure 4.5: The scatter plot of energy balance residuals versus the net radiation at the Kyeamba site. The relationship appears to be linear, as indicated by the regression line (red). The residuals are in general about 28% of the net radiation which is within the range for grassland reported in [Foken 2008].

Here in order to preserve the ratio of $SH$ and $LE$ ([Barr et al., 2006], i.e. the Bowen ratio, I assume the systematic error has the same impact on both sensible heat and latent heat and it is denoted by $k$. Hence

$$R_n - G = k \cdot (SH + LE) + \delta$$  \hspace{1cm} (4.3)

where $k$ is the correction factor to minimise the statistical error $\delta$. For the energy balance in Kyeamba, $k$ is found to be 1.32 by using a generalised linear regression method on the data excluding a small number of outliers. The reduced residual is
about 8% of $R_n$ (compared to 28% in the raw data). By using the same method, $k$ is estimated at 1.20 for Tumbarumba and the reduced residual is even smaller (0.7%).

The statistical method used here to deal with the energy balance closure problem is only for the purpose of this study. The model is expected to simulate the range and variation of the heat fluxes in an environment where advective moisture is negligible. The adjusted heat fluxes can provide a reference for the results.

4.4 Experiments

To study the effect of vegetation cover change on the atmospheric conditions, a sensitivity experiment has been performed. For each studied location, two different land cover scenarios are set up. The control run is based on the current vegetation properties (CTL) and the sensitivity run is based on a different land cover type (SEN). The current vegetation type in Kyeamba is short grasses so it is assumed to change to Eucalyptus forest in the second scenario as Eucalyptus is the most common tree type in this region. In the case of Tumbarumba, the forest is assumed to be replaced by short grasses in the sensitivity experiment. The CLASS model is run for each site on the selected days. The simulation starts from sunrise and stops at sunset. The exact starting hour and end hour are determined by the solar radiation from the flux data. The available observational data is used to estimate some of the initial and boundary values. Furthermore, simulations from the control run are compared to the measurements at the site to evaluate the model performance.

The model has been run separately for three consecutive days from 09 to 11 November. This was the last three days of a dry spell after the last rain events on 03 November (Tarcutta/Wagga Wagga: 10 mm; Tumbarumba: 11.4 mm). Hence the weather was hot and dry. The daily maximum temperatures recorded by BoM at the
Chapter 4. Comparison of boundary layer development over forest and shortgrass

Wagga Wagga station were 25.5°C, 30.4°C and 32.2°C on the three days. Relative humidity (RH) at 11am were 29%, 31% and 11%, respectively. Most of the rain water has been evaporated on the previous days so the soil water content was very low. The three-day average soil moisture at the top 30 cm was only 0.1 m$^3$ m$^{-3}$, which is far below the wilting point. Daily wind speed was below 7 m s$^{-1}$. Temperature is generally lower in Tumbarumba. Over the three days, the maximum temperature at this site was between 23°C and 27.5°C. RH was over 95% at 14:00 on the first two days but very low on the last day. The daily wind speed is generally quite low in Tumbarumba which might be due to the high-rising trees. The three-day average soil moisture at the top 30 cm was about 0.16 m$^3$ m$^{-3}$. Hence the soil was much wetter at this forest site than the grassland site in Kyeamba, although they both received about the same amount of rainfall from the previous event. The initial and boundary values used in the model for Kyeamba and Tumbarumba on 09 November are shown in Table 4.3 as an example. The model is initialised using the condition at 5am (1900 UTC) before sunrise, which is indicated by the positive incoming solar radiation (> 2 W m$^2$).

Since the simulations of the three days are separated, an individual set of initial and boundary conditions is needed for each day. As the days were successive, most of the conditions could be similar so set to the same values over the three days. They are shown in Table 4.3. The set of variables with different values for each day is listed in Table 4.4. For example, the initial mixed layer potential temperatures ($\theta$) on each day are based on the flux data. This value has decreased over the three days. The initial soil moisture values at the upper layer are also estimated from the field data. The potential temperature lapse rate ($\gamma_\theta$) and humidity lapse rate ($\gamma_q$) are estimated

---

4 Here the wilting point is set to 0.14 m$^3$ m$^{-3}$ for the silty loam soil at K10 (Kyeamba, from personal communication with Jeffrey Walker, Dept. of Civil Eng., Monash University, VIC, Australia)
from the sounding profile as mentioned previously. The difference between the three
days is quite small.

Table 4.3: Initial and boundary conditions in Kyeamba and Tumbarumba on 09 -
11 November (Julian day: 313 - 315). The wind speed and soil moisture values are
calculated from the site data provided by NAFE’06 [Walker et al., 2006].

<table>
<thead>
<tr>
<th>Variables</th>
<th>Description &amp; Units</th>
<th>Kyeamba</th>
<th>Tumbarumba</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Basic</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Start hour (UTC)</td>
<td>1900</td>
<td>1900</td>
<td></td>
</tr>
<tr>
<td>sunshine hours</td>
<td>13</td>
<td>13</td>
<td></td>
</tr>
<tr>
<td><strong>CBL &amp; entrainment layer</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$S_0$</td>
<td>Constant solar irradiance at the top of atmosphere [W m$^{-2}$]</td>
<td>1368</td>
<td>1368</td>
</tr>
<tr>
<td>$u_0$</td>
<td>Initial wind speed in x direction [m s$^{-1}$]</td>
<td>6</td>
<td>4</td>
</tr>
<tr>
<td>$v_0$</td>
<td>Initial wind speed in y direction [m s$^{-1}$]</td>
<td>-4</td>
<td>-2</td>
</tr>
<tr>
<td>$u_g$</td>
<td>Geostrophic wind in x direction [m s$^{-1}$]</td>
<td>20</td>
<td>20</td>
</tr>
<tr>
<td><strong>Surface layer</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$P_{hPa}$</td>
<td>Surface pressure [hPa]</td>
<td>1013</td>
<td>1013</td>
</tr>
<tr>
<td>$z_0$</td>
<td>Height of surface layer as proportion of boundary layer height [-]</td>
<td>0.1</td>
<td>0.1</td>
</tr>
<tr>
<td>$g_D$</td>
<td>vapour pressure deficit correction factor for surface resistance [-]</td>
<td>0.0</td>
<td>0.03</td>
</tr>
<tr>
<td><strong>Soil layer</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$w_{wilt}$</td>
<td>Wilting point volumetric water content [m$^3$ m$^{-3}$]</td>
<td>0.14</td>
<td>0.17</td>
</tr>
</tbody>
</table>

Continued on next page
Later in the sensitivity experiment, when short grasses are replaced by Eucalyptus trees or vice versa, the change is reflected in some of the surface parameters. They include the albedo \((\alpha)\), vegetation fraction \(C_{\text{veg}}\), leaf area index (LAI), minimum leaf surface resistance \(r_{s,\text{min}}\) and the roughness length for momentum, heat and moisture \((z_{0m}, z_{0h})\). These parameters are set constant in the model and no change is assumed between days. Their values in the two experiments CTL and SEN are shown in Table 4.5. The values used in the control experiments are selected from the literature.
Table 4.4: Different initial and boundary conditions in Kyeamba and Tumbarumba on 09-11 November. The subscript \(_0\) denotes the initial value.

<table>
<thead>
<tr>
<th>Variables</th>
<th>Description &amp; Units</th>
<th>Kyeamba</th>
<th>Tumbarumba</th>
</tr>
</thead>
<tbody>
<tr>
<td>( h_0 )</td>
<td>Mixed layer height [m]</td>
<td>600 450 200</td>
<td>600 600 600</td>
</tr>
<tr>
<td>( \theta_0 )</td>
<td>Mixed-layer potential temperature [K]</td>
<td>285.26 279.56 278.96</td>
<td>277.94 282.95 288.55</td>
</tr>
<tr>
<td>( \Delta \theta_0 )</td>
<td>Initial temperature jump at entrainment layer [K]</td>
<td>3 9 6</td>
<td>3 2 0.5</td>
</tr>
<tr>
<td>( \gamma \theta )</td>
<td>Potential temperature lapse rate [K m(^{-1})]</td>
<td>0.0049 0.0025 0.0042</td>
<td>0.0049 0.0025 0.0035</td>
</tr>
<tr>
<td>( q_0 )</td>
<td>Mixed-layer specific humidity [g kg(^{-1})]</td>
<td>4.877 4.419 4.06</td>
<td>3.86 5.48 3.65</td>
</tr>
<tr>
<td>( \Delta q_0 )</td>
<td>Humidity jump at entrainment layer [g kg(^{-1})]</td>
<td>-0.05 -0.20 -2.00</td>
<td>-0.20 -0.20 -1.00</td>
</tr>
<tr>
<td>( \gamma \theta )</td>
<td>Humidity lapse rate [g kg(^{-1}) m(^{-1})]</td>
<td>-0.0015 -0.0020 -0.0003</td>
<td>-0.0015 -0.0017 -0.0003</td>
</tr>
<tr>
<td>( \beta )</td>
<td>Ratio of entrainment flux to surface flux [-]</td>
<td>0.15 0.05 0.15</td>
<td>0.15 0.20 0.05</td>
</tr>
<tr>
<td>( T_{s,0} )</td>
<td>Surface layer temperature [°C]</td>
<td>9.60 2.90 2.30</td>
<td>4.78 7.34 9.34</td>
</tr>
<tr>
<td>( T_{soil1,0} )</td>
<td>Upper soil layer temperature [°C]</td>
<td>16.34 16.34 16.34</td>
<td>8.34 8.34 10.34</td>
</tr>
<tr>
<td>( T_{soil2,0} )</td>
<td>Deeper soil layer temperature [°C]</td>
<td>18.84 18.84 18.84</td>
<td>11.84 11.84 11.84</td>
</tr>
<tr>
<td>( w_{1,0} )</td>
<td>Upper soil layer volumetric water content [m(^3) m(^{-3})]</td>
<td>0.10 0.10 0.10</td>
<td>0.25 0.23 0.17</td>
</tr>
<tr>
<td>( w_{2,0} )</td>
<td>Deeper soil layer volumetric water content [m(^3) m(^{-3})]</td>
<td>0.21 0.19 0.18</td>
<td>0.27 0.24 0.24</td>
</tr>
</tbody>
</table>
(e.g. [De Ridder and Gallee] 1998, [Narisma and Pitman] 2003, [Whitehead and Beadle] 2004; [Leuning et al.] 2005). Then these values are again used exchangeably in the sensitivity experiments of the other study site. There is no change on the other conditions between the control experiment and the sensitivity experiment under my assumptions.

Table 4.5: Surface parameters for the control experiment and the sensitivity experiment in Kyeamba and Tumbarumba. $\alpha$: albedo; $C_{veg}$: vegetation fraction; LAI: leaf area index; $r_{s,min}$: minimum leaf surface resistance; $z_0m$: roughness length for momentum; $z_0h$: roughness length for heat and moisture. The values in CTL for Kyeamba are based on [De Ridder and Gallee] (1998), [Hagemann et al.] (1999), [Kleidon et al.] (2000), [Narisma and Pitman] (2003) and [Whitehead and Beadle] (2004). The values in CTL for Tumbarumba are from [Leuning et al.] (2005).

<table>
<thead>
<tr>
<th>parameters</th>
<th>Kyeamba CTL</th>
<th>Kyeamba SEN</th>
<th>Tumbarumba CTL</th>
<th>Tumbarumba SEN</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\alpha$ [-]</td>
<td>0.25</td>
<td>0.10</td>
<td>0.10</td>
<td>0.25</td>
</tr>
<tr>
<td>$C_{veg}$ [-]</td>
<td>0.30</td>
<td>0.85</td>
<td>0.85</td>
<td>0.30</td>
</tr>
<tr>
<td>LAI [-]</td>
<td>0.9</td>
<td>2.5</td>
<td>2.5</td>
<td>0.9</td>
</tr>
<tr>
<td>$r_{s,min}$ [s m$^{-1}$]</td>
<td>50</td>
<td>70</td>
<td>70</td>
<td>50</td>
</tr>
<tr>
<td>$z_0m$ [m]</td>
<td>0.03</td>
<td>2.00</td>
<td>2.00</td>
<td>0.03</td>
</tr>
<tr>
<td>$z_0h$ [m]</td>
<td>0.003</td>
<td>0.800</td>
<td>0.800</td>
<td>0.003</td>
</tr>
</tbody>
</table>

4.5 Simulations of Convective Boundary Layer Dynamics

The model simulations of the convective boundary layer dynamics under current vegetation cover are presented. In both sites, the day time changes of heat fluxes, air temperature and humidity from the model simulations are compared to the flux tower measured data. Furthermore, based on the information provided by the simulations, the chance of cloud and convective precipitation are also analysed. In this study, convection is measured by the CBL height. A high CBL could indicate that the
convection is strong. If the LCL is also within the CBL, it implies the CBL is wet and there is a chance of cloud and even rainfall.

4.5.1 Kyeamba: 09 - 11 November, 2006

The heat fluxes in Kyeamba on 09 - 11 November are quite well represented by the model. The amount of latent heat fluxes is within a similar range over the three days. Slightly higher sensible heat is estimated on the third day. As shown by the circles in the top row of Figure 4.6, the sensible heat fluxes appeared to rise slowly in the morning but fell steeply after a peak in the afternoon. The model simulations could roughly reproduce the daily range and perform well in the falling stage. However they all tend to rise up too fast and peak too early. On the other hand, the simulations of latent heat generally miss out the afternoon fall. A similar difference is also seen in the simulations for Cabauw, the Netherlands in van Heerwaarden et al. (2010). It is possible that a different mechanism for energy redistribution exists in the afternoon which has not been included in the model. For example, the data indicates that wind speeds were higher around noon and in the early afternoon, especially on the second and the third days. This might have removed part of the latent heat under the measurement point hence the observed latent heat flux measurements were lower In the model the wind speed is assumed to be constant during the day, which should affect the simulation quality.

The model is clearly able to simulate atmospheric temperature. The temperatures at the measurement height of 2 m are very well reproduced by the model for all days (see

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5 The sensible heat fluxes were not affected in the same way. This could be due to the fact that the sensible heat is estimated from the balance between the observed net radiation, ground heat fluxes and latent heat fluxes (from communication with Eva van Gorsel, CSIRO Marine and Atmospheric Research, Canberra, Australia). While wind can reduce both net radiation and latent heat fluxes, the impact on their balance is small.
Figure 4.6: Simulated and observed day time dynamics of atmospheric conditions in Kyeamba during 09 - 11 November. The observational data are shown by circles and the model simulations are shown by lines. Top row: Sensible heat flux (red) and latent heat flux (blue). Middle row: temperature at 2 m height (thin lines) and CBL potential temperature (thick lines). The site measurement is also taken at 2 m height. Bottom row: Specific humidity.
thin solid lines and circles in the middle row of Figure 4.6) with a root mean squared error (RMSE) of 0.88 (Table 4.6). The mixed layer potential temperatures were lower than the temperature at 2 m after the day started. Their difference increased as the day proceeded; but the surface layer temperature dropped quickly at sunset, while the CBL potential temperature still increased or remained at a high level. On the first day, the temperature increased steadily which was different from the other two days. The air temperatures rose quickly during the morning of day two and three, which were mainly due to the high temperature jump at the entrainment layer and were partially due to the dry top soil (as listed in Table 4.4).

Table 4.6: Root mean squared error (RMSE) between measurements and simulations. RMSEs are provided for the sensible heat flux (SH), latent heat flux (LE), temperature at 2 m ($T_{2m}$) and humidity ($q$) in Kyeamba. RMSE is not calculated for temperature in Tumbarumba, as the temperature is measured at 70 m. The model does not simulate temperature at this height. The model performs better at the grassland site, as RMSEs of SH, LE and q are 15%, 80% and 10%, respectively, which are lower than the forest site.

<table>
<thead>
<tr>
<th></th>
<th>SH (W m$^{-1}$)</th>
<th>LE (W m$^{-1}$)</th>
<th>$T_{2m}$ (°C)</th>
<th>$q$ (g kg$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kyeamba</td>
<td>117.77</td>
<td>26.71</td>
<td>0.88</td>
<td>0.51</td>
</tr>
<tr>
<td>Tumbarumba</td>
<td>138.64</td>
<td>132.50</td>
<td>-</td>
<td>0.57</td>
</tr>
</tbody>
</table>

The simulations are not as good in terms of the specific humidity. On the first day, the modelled humidity did not fall as far as the observed values but appeared to be lower in the afternoon. A deep fall on day two and day three at different time, 19:00 and 12:00 respectively, again could not be picked up by the model. With the setting of the advective moisture input, the model is able to reproduce the sharp rise and the peak but again the fall is smaller than the observations. The changes in the model, such as soil moisture and radiation input, only account for half of the drop in the observations on 10 November, which decreased more than 65% within 10 hours. If assuming a
small value of the humidity lapse rate, the proper range could be reproduced but it would lead to negative humidity above the CBL. Furthermore the slow decreasing rate in the afternoon could not be mimicked by the model. Although the latent heat fluxes in the afternoon were slightly overestimated, the magnitude was not comparable to this large fall. It is again possibly due to some external forcings which are not considered in the model. Investigation of the wind directions and speeds recorded at the location during the three days does not suggest the dry westerlies as an explanation for this big fall of the humidity. The simulation cannot reproduce the slow decreasing rate either. Such a slow decrease in the afternoon hours is not shown in the simulations in \cite{vanHeerwaarden2010}. An explanation could be slow drying of the CBL before 15:00 and then the reduced latent heat has made the situation even worse. Although later of the day air dryness might encourage a higher latent heat flux, as net radiation significantly reduced, this recovery was very small.

The mixed layer increased from several hundred meters to over 2,000 m on all three days. The strongest boundary layer development occurred on the second day, 10 November, when the initial mixed layer height was 450 m and the highest height at the end of the day was nearly 2,800 m which was also the highest over the three days. There was not much difference in the maximum CBL height between day one and day three. But the growth on day three was larger with a lower initial height at 200 m, while the initial height on day one was 600 m. Hence the boundary layer development on the first day was the weakest. The difference in the CBL development could be due to the mixed layer temperature (as shown in Figure \ref{fig:4.6}). The mixed layer height is mainly determined by sensible heat and temperature \cite{Fisch2004}. However the sensible heat fluxes of day three were underestimated. It is therefore possible that the
CBL height on day three was even higher than day two, given its high sensible heat and high temperature compared to the previous two days.

High energy and moisture supply as well as strong boundary layer growth are necessary conditions for convective clouds and precipitation (Stull, 1988; Ahrens, 1991; Strangeways, 2007). Here precipitation is not modelled by CLASS. In order to get some ideas of the chance of precipitation, the LCL is used as an indicator for the onset of cloud and convective precipitation. The LCL is determined by the height where the mixed layer specific humidity is equal to the saturation humidity in the model. If the LCL is reached within the convective boundary layer (i.e. $LCL < h$), then it is an indication of the onset of clouds and a chance of rainfall (Nair et al., 2011; de Arellano and van Heerwaarden, 2012). No LCL is simulated within the CBL over these three days. There was generally a difference of 1 - 6 g kg$^{-1}$ between the mixed layer humidity and the saturation level at the top of the CBL. The average difference was the smallest on day one, due to a lower CBL potential temperature and relatively stable medium range humidity throughout the day. On day two, although the specific humidity is underestimated in the simulation, there was little chance that the observed humidity has exceeded the saturation level as the error is still much smaller than the difference between the specific humidity and the saturation humidity. The results from the model do not imply the upcoming rainfall event. However, it is possible that the rain event was caused by the frontal system movement rather than local convection.

### 4.5.2 Tumbarumba: 09 - 11 November, 2006

The heat flux patterns are quite different in Tumbarumba. The major difference between the two sites is that the summer latent heat fluxes are much higher in
Chapter 4. Comparison of boundary layer development over forest and shortgrass

Tumbarumba. During the three days which were in late spring and early summer, the latent heat fluxes could reach 400 W m$^{-2}$ over the forest, but was generally lower than 100 W m$^{-2}$ on the grassland. Both heat fluxes increased rapidly in the morning hour, which was not seen in Kyeamba. There was much more variation in the heat fluxes in Tumbarumba, as shown by the points in Figure 4.7 (top row). It increases the difficulty in reproducing the observed heat fluxes with the model simulations.

![Figure 4.7: Simulated and observed day time dynamics of atmospheric conditions in Tumbarumba during 09 - 11 November for sensible heat fluxes and latent heat fluxes (top row), air temperature ($\theta$, middle row) and specific humidity ($q$, bottom row). The site measurement of air temperature is at 70 m height. The simulated temperature at 2 m (thin line) and CBL potential temperature (thick line) are both shown for comparison. The colour scheme and symbols are the same as Figure 4.6.](image_url)
One common problem in the simulations over the three days is that they fail to reproduce the active latent heat fluxes in the morning hours. The observed latent heat fluxes increased at almost the same time as the sensible heat fluxes. However the simulated latent heat fluxes tended to rise at a later time due to the energy competition between the two types of heat fluxes. Furthermore, temperature is a driver of ET in the model, so low temperature suppressed ET in the early morning. On the other hand, the decrease of the simulated latent heat fluxes also lagged behind the observed flux. In the model it was due to the high temperature in the afternoon, which is also seen in van Heerwaarden et al. (2010) and Figure 4.6. In terms of magnitude, on day two and three, the observed latent heat was as high as the sensible heat, or even higher. But the simulated maximum latent heat did not reach the same level without compensation from the sensible heat fluxes. The energy used in the high latent heat fluxes at noon, on day two, was possible from the sudden drop of sensible heat fluxes. However, the source of the high latent heat fluxes around noon on day three is unknown.

The overall heat fluxes observed suggests an additional amount of energy available at the forest site in the morning. But this could not be modelled under the current assumptions. For example, it is possible that the presence of light cloud can increase the downward longwave radiation, hence provide extra heat energy to the surface (NASA, 2013). Furthermore, heavy fogs in the forest could be a source of evaporated water (Scholl et al., 2011; Giambelluca et al., 2011; Tobón et al., 2011) which utilise the extra heat and increases the morning latent heat and air humidity. These are not included in the model.

Compared with the heat fluxes and the air humidity, the simulated air temperatures are relatively good. However, the match between the simulations and the observations
is not as good as in Kyeamba. As pointed out in Table 4.2, the air temperatures as well as radiations are measured at 70 m at this forest site. While the model outputs temperature of soil layer, ground level, 2 m height level and CBL potential temperature, these simulations are expected to be different from the observations. Here the simulated temperatures at 2 m and the potential temperature are used as a comparison to the observed temperatures. In fact, the measured air temperature at 70 m is only partially matched by the modelled temperature. Over the three days, the 2 m temperatures followed a similar pattern as the observations. During day time, especially on day two and three, the potential temperature was of the same magnitude as the measured temperatures which might indicate a well mixed layer. In the later afternoon, the measured temperatures fell and their behaviour was generally the same as the 2 m temperatures due to the reduced sensible heat.

The variation in the specific humidity in Tumbarumba was less than Kyeamba. In the first two days, the model could not simulate the maximum $q$ which is probably due to the same reasons as the latent heat fluxes. Therefore the underestimated latent heat fluxes are consistent with the underestimated humidity. Again the model cannot reproduce the slow decrease of humidity during noon of 10 November. This scenario is different to that in Kyeamba. The observed latent heat fluxes were much higher than the simulated result before 3pm. Hence the specific humidity might have a small decrease given the rising boundary layer. The simulations of the third day specific humidity are very close to the observed data, although the measured latent heat fluxes indicate a higher value is possible.

The boundary layer simulations were generally higher over the forest site than the grassland site. On the second day, the maximum boundary layer was the highest at
3,656 m. The rapid boundary layer development in the morning was due to a small initial temperature jump at the entrainment layer and a small lapse rate at the free atmosphere. The LCL is estimated to reach within the boundary layer on the first two days in the simulations. On 09 November, a cloud level was estimated after 10:30 at the top of the boundary layer of about 1,430 m. As the boundary layer grew, the LCL also rose but with a different growth rate from the CBL. At 18:00, the LCL was almost 520 m below the top of the boundary layer. On the next day, the estimated cloud level was two hours earlier than in the previous day. By the end of the day, the LCL was nearly 1,500 m below the top of the CBL, which indicates a possible deep convection. The LCL is not simulated within the CBL for day three, although boundary layer development was similar to day one. Higher temperature and lower humidity (as shown in Figure 4.7) are the reasons that the LCL is not simulated within the CBL for day three, as the saturation humidity is determined by the air temperature. Nevertheless, the CBL height in Tumbarumba was still slightly higher than that in Kyeamba. Overall, this suggests that cloud formation is more likely over the forested site than over the grassland site.

4.6 CBL sensitivity to land surface change

4.6.1 Kyeamba: 09 - 11 November, 2006

In the sensitivity experiment in Kyeamba, the surface configuration has been changed from short grasses to forest, while the other initial and boundary conditions remain the same. Consequently, there were a series of changes in the convective boundary layer. The heat fluxes, temperatures and the air humidity in the forest scenario are shown as dashed lines in Figure 4.6. The changes over the three days are consistent in Kyeamba.
Figure 4.8: Compare the convective boundary layer development between control experiments and sensitivity experiments in Kyeamba and Tumbarumba. (a) potential temperature changes and CBL heights; (b) specific humidity changes and CBL heights. In Kyeamba, temperature increased rapidly on 10 and 11 Nov, due to a large initial jump. The development of the boundary layer on 10 Nov in Tumbarumba was the highest, as a result of high $q_0$, low $\theta_0$ and low $\gamma_\theta$. 
There was a clear increase in the latent heat fluxes, which is due to a much higher ET from the forest. The difference between the two scenarios was smaller in the morning and larger in the afternoon to early evening. In the morning, it was cooler and not too dry (Figure 4.8). The solar radiation gradually increased as the day proceeded. In addition to soil moisture, solar radiation, temperature and air vapour deficit are all constraints of transpiration (Jarvis, 1976; Rodriguez-Iturbe and Porporato, 2004).

On the other hand, soil evaporation is mainly limited by soil moisture in the model. Hence in the morning the evapotranspiration was dominated by soil evaporation which is the reason that the difference in latent heat fluxes was small between the two types of land surface. The latent heat fluxes in the afternoon over the forest surface were more than four-fold the fluxes over the grasses. Moisture supply from the deeper soil and demands from the still high temperature allow the forest to maintain a high transpiration rate in the afternoon, while the solar radiation started to decrease. These explain the lower sensible heat fluxes in the afternoon compared to the grassland scenario. Consequently, the afternoon air was moister in the forest scenario. This is consistent with the findings over the native vegetation areas in the southwest Western Australia (Ray et al., 2001).

The overall heat balance was higher in the forest scenario, due to a lower albedo. However, the distribution of the energy gain towards the two types of heat fluxes varied during the day. In the morning, there was hardly any difference in the sensible heat fluxes over forest and short grasses. The extra net radiation was almost fully utilised by the latent heat fluxes. In the afternoon, the increase of latent heat fluxes have been compensated by a relatively small decrease in the sensible heat fluxes. Corresponding to the small change in the sensible heat fluxes, the change in the daily mean potential
temperature between the two scenarios was less than a 0.2 K decrease. The difference is relatively higher in the early evening (Figure 4.6). Hence there was almost no change in the CBL height before the evening (Figure 4.8). The lower CBL height over the forest in the early evening is a result of the lower sensible heat fluxes and reduced potential temperature.

Above the forest cover, the LCL was again not able to form within the CBL during the day, similar to the grassland conditions. As mentioned above, there was hardly any difference in the CBL height until the late afternoon or early evening. Increased latent heat fluxes have resulted in higher specific humidity. The potential temperature changed little so the levels of saturation humidity at the top of CBL did not change much. Higher specific humidity means the difference with the saturation humidity has been reduced. This reduction was more noticeable in the afternoon. However the maximum increase of specific humidity after vegetation cover change was not large enough to reach the saturation level. Based on this information, it cannot be concluded that the chance of precipitation is higher over a particular vegetation cover.

The changes in the output variables are different in magnitude after vegetation cover changes. Among all the variables, the relative difference\(^6\) in latent heat was the largest. The three-day average shows that the latent heat flux was greatest at about 15:00, associated with the maximum difference between the control run and the sensitivity run. This is also illustrated in Figure 4.6. The maximum difference in the sensible heat fluxes before and after LCC was at around 18:00, and it was much smaller than the difference in the latent heat fluxes. While the net radiation was higher during most of the day in the reforestation scenario as a result of lower albedo, it was most significant after midday and encouraged the latent heat fluxes to increase more

\(^6\) The output difference between the two experiments for each variable is normalised by the average value of the CTL run.
than the gain in the net radiation. Figure 4.6 shows that the latent heat fluxes rose to the daily maximum while the sensible heat fluxes have to reduce further to offset this. The high sensitivity of saturation vapour pressure to temperature changes as described in the Clausius-Clapeyron relation (D’Andrea et al., 2006; van Heerwaarden, 2010) could cause an increase in the demand of ET through vapour pressure deficit. Overall, the nonlinear relationship between variables has a leverage effect on the latent heat fluxes and leads to the unbalanced redistribution of the available energy.

**4.6.2 Tumbarumba: 09 - 11 November, 2006**

In Tumbarumba, the sensitivity experiment is performed by changing the forest surface configuration to the grassland configuration. In this case, the grassland configuration in Kyeamba is used (see Table 4.5). The heat fluxes, temperatures, specific humidity and the boundary layer development are again the main variables of interest. The results are shown in Figure 4.7 and 4.8.

The total heat fluxes have been reduced in the sensitivity experiment (top row in Figure 4.7). This is due to a higher surface albedo which was almost double in the grassland scenario. The major changes were again in the latent heat fluxes. The differences in the latent heat fluxes between the two experiments in Tumbarumba were consistent among the three days. In the first two days, the model estimates that the daily maximum latent heat flux over the grass surface could drop down to almost one third of the amount over the forest surface. Soil water content is depleted from the upper soil layer, but is assumed in the model to be constant at the deeper soil layer. As the vegetation coverage on the grassland is small (C_veg = 0.3), soil evaporation from the upper soil is a main part of ET. The latent heat fluxes dropped as the water content
in the upper soil fell below the wilting point. Therefore, the new latent heat fluxes on
day three were relatively steady during the day as the initial soil moisture in the upper
layer was at wilting point.

The other variables do not show much difference from the control experiment. The
overall changes in the air potential temperature due to deforestation are insignificant,
as a result of the small changes in the sensible heat fluxes. The daily humidity
after deforestation did not change much either, although the latent heat fluxes were
significantly reduced. The outputs also show that the boundary layer heights were
generally lower on the first two days over the grassland, particularly before 15:00 as
a result of the lower sensible heat fluxes. The mean temperature was actually lower
over the grassland. The LCL was again predicted during these two days, but delayed
by about half hour. A slightly larger difference is seen in the air humidity between
the two experiments on day three, which was associated with a 0.5 K increase in the
mean potential temperature. Higher sensible heat and temperatures lead to higher CBL
height and consequently a more noticeable reduction in the mixed layer humidity.

Although the distribution of energy has an important impact on the boundary layer
conditions, the results also reveal the important control of boundary layer height on
temperatures and air humidity. As mentioned above, because the CBL height increased
on day three, lower latent heat fluxes have led to lower specific humidity in the
deforestation experiment. However in the previous two days, the boundary layer had a
weaker development during the day, compared to the control case (Figure 4.8). Hence
the decreases in latent heat fluxes were offset, leaving little change in the humidity.
Through the regulation of the boundary layer height, changes in the sensible heat fluxes
have resulted in even smaller changes in the potential temperatures.
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The water content in the top soil layer plays an important role in the grassland scenario. In Tumbarumba, the initial soil moistures in the upper layer were at least equal to the wilting point level. In addition to plant transpiration, soil evaporation has been active until the soil moisture dropped below the wilting point. Therefore, the morning sensible heat fluxes on the first two days were lower in the grassland scenario than in the forest scenario, while this was not the case on day three. The lower sensible heat flux and hence a lower air temperature in the morning have delayed the boundary layer development. So the CBL height above the short grasses was clearly lower than over the forest on the first two days. Unlike in Kyeamba, the grassland scenario here has a lower boundary layer in the early evening due to the weaker boundary layer development during the day.

In summary, the sensitivity experiments show that when vegetation cover changes, there are some corresponding changes in the boundary layer conditions. The largest effect is shown in the latent heat fluxes as transpiration from trees is much more than grasses in this case. However the overall impact is not strong enough to significantly affect the convective boundary layer development and the likelihood of precipitation. It appears that there is strong resistance within the system which could reduce the impact of vegetation cover change. Hence I need to look into details of vegetation cover changes and the possible feedback relationships involved.

4.7 Significance of land surface parameters

In this study, the vegetation cover changes are represented by six parameters. They are the albedo ($\alpha$), vegetation fraction ($C_{veg}$), leaf area index (LAI), minimum leaf surface resistance ($r_{s,min}$), the roughness length for momentum ($z_{0m}$) and the roughness length for heat and moisture ($z_{0h}$). Each of the land surface parameters has different effect on
the boundary conditions. Some of these effects are opposite to each other. As the land surface changes from one type to the other, the overall impact on the boundary layer conditions would depend on the significance of each parameters and their possible influence on the effects of the other parameters. In the model the influence of $z_{0h}$ is similar with $z_{0m}$, but with smaller magnitude. Hence only $z_{0m}$ is discussed here.

The $\alpha$ and $z_{0m}$ can affect the sensible heat fluxes and the latent heat fluxes, as shown in Figure 4.9. As mentioned previously, lower $\alpha$ increases the net radiation. Through heating up the surface, it is possible that more energy would be used in each of the heat fluxes. Higher $z_{0m}$ decreases the aerodynamic resistance $r_a$ which in turn has a negative impact on both heat fluxes. There are some complications. The $r_a$ has a positive impact on the surface temperature $T_s$. In this case, higher $z_{0m}$ can decrease the heat fluxes too. As $r_a$ is also determined by air temperature and humidity, the impact of $z_{0m}$ depends on the initial conditions of the CBL and the feedback between heat fluxes and the boundary layer conditions. Furthermore, $z_{0m}$ also affects the canopy surface resistance $r_s$ through a few more processes and its impact again depends on the boundary layer conditions.

The $C_{veg}$ can also affect both heat fluxes. It influences the latent heat fluxes directly but the sign of influence would depend on some other factors. For example, if the soil surface resistance is low, higher $C_{veg}$ could possibly lead to lower latent heat fluxes. In the model, since vegetation has access to a constant water supply from the deep soil layer, a higher vegetation fraction means higher transpiration as long as the deep soil layer is above wilting point. Higher vegetation fraction also reduces the impact of the water content in the top soil layer. When the soil surface resistance is low, evaporation from the top soil layer can be a bigger contributor to the latent heat fluxes.
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Figure 4.9: The core relationship between the land surface parameters and the main output variables. The land surface parameters are: $\alpha$ - albedo; $C_{\text{veg}}$ - vegetation fraction; LAI - leaf area index; $r_{s,\text{min}}$ - minimum leaf surface resistance; $z_0$ - roughness length. The other variables are: $C_{\text{liq}}$ - equivalent water layer depth for vegetation; $h$ - boundary layer height; LE - latent heat flux; $q$ - air humidity; $R_n$ - net radiation; $r_a$ - aerodynamic resistance; $r_s$ - canopy surface resistance; SH - sensible heat flux; $T_s$ - surface temperature; $\theta$ - potential temperature. The red lines represent positive relations and the blue lines represent negative relations. Grey lines indicate the relationship can be either positive or negative, depending on the other values. The bottom panels with number 1, 2 and 3 indicate the shortest, second shortest and the longest paths between LAI and SH.
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Therefore, higher vegetation cover does not necessarily cause higher evaporation if the other vegetation properties are not considered. The $C_{\text{veg}}$ can also affect $T_s$ which then has a positive influence on the heat fluxes. Similarly, the sign of the effect of $C_{\text{veg}}$ on $T_s$ is determined by the other factors, such as the resistance terms. But the effects of $C_{\text{veg}}$ on $T_s$ and on the latent heat are expected to have an opposite sign.

The LAI and $r_{s,\text{min}}$ mainly affect ET or the latent heat fluxes (see the right hand side of Figure 4.9). Both LAI and $r_{s,\text{min}}$ influence the $r_s$ directly which then has a negative impact on the latent heat. The difference between LAI and $r_{s,\text{min}}$ is that higher LAI decreases $r_s$, while higher $r_{s,\text{min}}$ increases $r_s$. When the land cover changes from grasses to forest, both LAI and $r_{s,\text{min}}$ are expected to increase. The $r_{s,\text{min}}$ is assumed to be higher over forest than over short grasses to reflect the stronger control of trees on water use (Teuling et al., 2010). The LAI also negatively relates to the equivalent water layer depth for vegetation ($C_{\text{liq}}$), which is a term for the intercepted water that can be used for evaporation from the canopy surface (de Arellano and van Heerwaarden, 2012). As higher $C_{\text{liq}}$ increases latent heat but decreases $T_s$, the effect of LAI is yet to be determined.

Figure 4.9 illustrates the relationships between factors but the magnitude of change cannot be shown. As a simple example, the “shortest distance” effect of $z_{0m}$ on SH is positive while the shortest distance effect of $\alpha$ on SH is negative. Dependent on the significance of each land surface parameter on SH, SH might increase, decrease or have no change. When LE is increased by either LAI or $C_{\text{veg}}$, it can influence SH through different paths, as indicated by the number 1, 2 and 3 in Figure 4.9. Again, each path imposes either a positive or negative impact on SH. It is also useful to know

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7 Here the shortest distance effect between A and B is defined as the control of A on B through the least processes. For example, change of roughness length ($z_{0m}$) can affect LE through change in aerodynamic resistance ($r_a$) or through a chain change in $r_a$, SH, $\theta$. So the shortest distance effect of $z_{0m}$ on LE is ($z_{0m} \rightarrow r_a \rightarrow LE$).
the level of impact of each surface parameter on the output variables.

Alternatively, the significance of the land surface parameters can be shown by the changes of the output variables after varying the surface parameters. Based on the grassland configuration, each of the five land surface parameters has increased by 10%, 20%, 30%, 40% or 50% in each new simulation. The grassland scenario results from the two sites are used as the reference. The relative changes\[^8\](in %) of the heat fluxes and $h$ are calculated.

The $\alpha$ has a negative impact on the heat fluxes. When there is no change on the other surface parameters, a 50% decrease in $\alpha$ can increase both heat fluxes by more than 10%, as shown in Figure 4.10(a). The relative changes in the sensible heat fluxes and the latent heat fluxes due to a varying $\alpha$ are similar. The other land surface parameters would have a leverage effect on the impact of $\alpha$. For example, on 11 November in Kyeamba, when there was no change on the other parameters, the difference between a 10% and a 50% decrease of $\alpha$ on the latent heat fluxes was 8.8%. When LAI increased by 50%, this difference increased to 15.4%. Similarly, when $C_{veg}$ increased by 50%, this difference increased to 15.8%. On the other hand, as LAI and $C_{veg}$ increase, the difference between a 10% and a 50% decrease of $\alpha$ on the sensible heat fluxes decreased from 21% to 20%, which was much smaller than in the latent heat fluxes.

Higher $C_{veg}$ is found to increase the latent heat fluxes but decrease the sensible heat fluxes. Its impact on the latent heat fluxes is generally much higher than on the sensible heat fluxes (Figure 4.10(b)), which might be due to its direct relationship with the latent heat fluxes. Furthermore, the direct relationship is likely to be a positive one as: (1) $T_s$ has a positive relationship with both heat fluxes and the sensible heat

\[^8\] This is the difference between the new simulated outputs and the reference results, normalized by the reference results.
Figure 4.10: Relative changes (in %) of the heat fluxes at midday due to changes in (a) albedo, (b) vegetation fraction, (c) LAI, (d) minimum leaf surface resistance and (e) roughness length, with reference to the results from the grassland scenario in Kyeamba on 11 November.
Chapter 4. Comparison of boundary layer development over forest and shortgrass decreases with higher $C_{veg}$; and (2) the direct relationship with the latent heat fluxes is expected to have opposite sign to the relationship between $C_{veg}$ and $T_s$. In Kyeamba on 11 November, a 10\% increase in $C_{veg}$ could lead to almost 10\% increase in the latent heat fluxes but less than a 2\% decrease in the sensible heat fluxes. LAI and $r_{s,\text{min}}$ are shown to have an impact on the effect of $C_{veg}$. Higher LAI enhances the effect of $C_{veg}$, while higher $r_{s,\text{min}}$ suppresses the effect of $C_{veg}$ (see Figure 4.11).

Figure 4.11: Relative changes of the latent heat fluxes (LE) at midday due to changes in $C_{veg}$, with reference to the results from the grassland scenario in Kyeamba on 11 November. The $C_{veg}$ has a positive impact on LE. The figures also show the leverage effects of LAI and $r_{s,\text{min}}$ on the impact of $C_{veg}$.

In terms of magnitude, the three parameters, $C_{veg}$, LAI and $r_{s,\text{min}}$, have similar effects on the heat fluxes when they are changed by a small amount (comparing Figure 4.10(b), 4.10(c) and 4.10(d)). But the effects of $r_{s,\text{min}}$ have opposite sign to the other two. Compared to $C_{veg}$ and LAI, $r_{s,\text{min}}$ shows a clear nonlinear relationship with the heat fluxes for the range tested (see Figure 4.10(d)). As $r_{s,\text{min}}$ increases, the rate of change in the latent heat fluxes decreases. Hence the effect of $C_{veg}$ or LAI is expected to outgrow the effect of $r_{s,\text{min}}$, if they change by a large amount. At the same time, $r_{s,\text{min}}$ can suppress the effects of both $C_{veg}$ and LAI.
Figure 4.12: Relative changes of (a) the sensible heat fluxes (SH) and (b) the latent heat fluxes (LE) at midday due to changes in $z_{0m}$, with reference to the results from the grassland scenario in both study sites. The $z_{0m}$ changes by 10% to 50%. The other surface parameters are the same as in the grassland configuration. The black symbols are for Kyeamba and the red symbols are for Tumbarumba. If the relationships between the change of $z_{0m}$ and the changes of the heat fluxes are linear, the black circles should fall on the lines. It shows that the change of $z_{0m}$ has a quasi-linear positive effect on SH. There is no clear pattern of the relationship between the change of $z_{0m}$ and LE, which is possibly due to the impact of $r_s$. 
Chapter 4. Comparison of boundary layer development over forest and shortgrass

The impact of $z_{0m}$ on the heat fluxes is shown in Figure 4.10(e). The change of $z_{0m}$ has a positive effect on the sensible heat fluxes. A 10% increase in $z_{0m}$ could lead to 0.28 - 0.38% increase in the sensible heat fluxes in all cases. There is a small decrease in the change of the sensible heat fluxes as $z_{0m}$ increases further. Hence the change of $z_{0m}$ and the change of the sensible heat fluxes have a quasi-linear relationship (Figure 4.12(a)). On the other hand, although higher $z_{0m}$ generally results in lower latent heat fluxes, there is no consistent pattern of the effect. As shown in Figure 4.12(b), for every 10% increase in $z_{0m}$ the latent heat fluxes could decrease or does not change. The effect of $z_{0m}$ on $r_s$ might have triggered more feedback processes which can be eliminated by the impact of a small change on $z_{0m}$. Nevertheless, the impact of $z_{0m}$ on the boundary layer conditions is quite small. Figure 4.12 confirms that the initial and boundary conditions affect the impact of $z_{0m}$ on the model outputs. The changes in the other surface parameters have no influence on the impact of $z_{0m}$.

In summary, the five land surface parameters mainly affect the latent heat fluxes. A 50% increase in $C_{veg}$, LAI or $r_{s,min}$ can result in more than 30% change in the latent heat fluxes but only around 5% in the sensible heat fluxes. The effects of $z_{0m}$ on both heat fluxes are very small, as a 50% increase of $z_{0m}$ can only lead to less than 1.5% change in the heat fluxes. Hence the surface roughness effects are usually less important (Evans et al., 2011). The impact of $\alpha$ on the sensible heat fluxes might double its impact on the latent heat fluxes, giving more than 20% increase in SH when it decreases by 50%. The signs of impacts of each parameter on the heat fluxes are listed in Table 4.7. Overall, $\alpha$ has the largest impact on the boundary layer height (Figure 4.13). Although LAI and $C_{veg}$ have small negative impacts on the CBL height, they could offset the impact of $\alpha$ depending on the magnitude of the change.
Table 4.7: The relationship of the land surface parameters and the heat fluxes. SH is sensible heat flux. LE is latent heat flux. Positive relationship is indicated by “+” and negative relationship is indicated by “−”.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>SH</th>
<th>LE</th>
</tr>
</thead>
<tbody>
<tr>
<td>$z_0$ m</td>
<td>+</td>
<td>−</td>
</tr>
<tr>
<td>LAI</td>
<td>−</td>
<td>+</td>
</tr>
<tr>
<td>$C_{veg}$</td>
<td>−</td>
<td>+</td>
</tr>
<tr>
<td>$\alpha$</td>
<td>−</td>
<td>−</td>
</tr>
<tr>
<td>$T_{s,min}$</td>
<td>+</td>
<td>−</td>
</tr>
</tbody>
</table>

Figure 4.13: Relative changes (in %) of the midday CBL height ($h$) with reference to the results from the grassland scenario in Kyeamba on 11 November, due to changes in each of the land surface parameters. $\alpha$ is showing as a decreasing series and labelled by the top axis (in red). $\alpha$ has the highest impact on $h$ due to its effect on the sensible heat fluxes. The parameters, which have positive relationships with the sensible heat fluxes (see Table 4.7), have a positive relationship with $h$; and vice versa.
In our sensitivity experiment, the magnitude of change on each surface parameter varies. The change in $z_0$ was the largest, with the forest roughness being more than 60 times of the grassland roughness. The sole impact of $z_0$ on the midday heat fluxes was only 13% (SH) and 17% (LE). Although there was 60% change in $\alpha$, it almost doubled the effect on the sensible heat fluxes. As expected, the change of $\alpha$ has the largest impact on the boundary layer height, followed by $z_0$ given its significant change. Their combined effect was about the same as the sum of the individual effects. For example, in the Kyeamba sensitivity experiment on 11 November, $h$ increased by 9% and 17% due to the change of $z_0$ and $\alpha$, respectively. When both changed, there was 26% increase in $h$. On the other hand, $C_{veg}$ and LAI each contributed to 11% and 9% decrease in $h$ but the combined effect is -32%. The overall impact of the changes of the six surface parameters was actually 2% increase of $h$, which is probably due to the damping effect of $r_{s,min}$ on $C_{veg}$ and LAI. The nonlinear feedback relationships between the parameters and the outputs tend to balance the individual impact. Hence in our sensitivity experiment, the boundary layer growth did not change much.

4.8 Sensitivity to advective moisture convergence

In the previous experiments, only the local changes of the land surface have been considered. However, changes of vegetation can also modify the boundary conditions. The advective moisture convergence can be affected by the vegetation due to the changes in the roughness length. This is considered to be the main reason of rainfall changes \cite{Pitman2004,Pitman2007}. Hence the effect of advective moisture is also tested in the case of Kyeamba afforestation. The CBL is assumed to be moistened by this external moisture supply, at a rate of 0.5 kg m$^{-2}$ hr$^{-1}$ for 2.5 hours in the morning.
The chance of reaching the LCL within the CBL is higher after including the advective moisture convergence in the forest scenario. In the previous experiments, the LCL did not form within the CBL in either the control experiment (grassland) or the sensitivity experiment (forest). In the new experiment with a positive advective moisture term, the LCL was reached in the afternoon of the first two days (Figure 4.14). On 09 November, the LCL was well below the top of the boundary layer, which implies possible cloud formation. The reason that the LCL was still above the maximum CBL height on 11 November could be the high air temperature in the afternoon.

### 4.9 Sensitivity to initial soil moisture

The model sensitivity to the initial soil moisture is also tested. Given there are two soil layers in the model, two cases are tested: (1) the initial water content in the deeper soil is assumed to be the same as that in the top soil; (2) the initial water content in
the deeper soil was above wilting point and did not change in all the scenarios tested. The first case allows test on the sensitivity of the deeper soil, which is expected to have large impact on the forest. The second case guarantees water supply to vegetation. The tested range covers the dry end to the wet end, from below wilting point to above field capacity (refer to Table 4.3). The same range has been applied on both sites, with the volumetric soil moisture changing from 0.1 to 0.4 m$^3$ m$^{-3}$.

The difference in daily ET between the two sites as well as the difference between the control and the sensitivity experiment are shown in Figure 4.15. In the case that the initial deeper soil moisture is the same as the upper soil moisture (Figure 4.15(a)), daily ET increases with higher initial soil moisture in both forest and grassland scenarios. In the control experiments, when the soil moisture is below wilting points or above the field capacity, the difference in ET between the two sites is small. However in between, their difference can be up to almost 2 mm day$^{-1}$. The main reason is that ET from the grassland and that from the forest have different response functions to the initial soil moisture. Over the grassland, with a lower vegetation cover, ET increases exponentially with soil moisture. On the other hand, the forest greatly utilises soil moisture when it is relatively dry but the increase of ET slows down as the soil gets wetter. The simulated ETs from the sensitivity experiments also confirm the ET responses to the initial soil moisture are different for different vegetation types. The ET-soil moisture relationships for different surface types are consistent with Entekhabi et al. (1992).

The relation between ET and initial upper soil moisture is quite different from the first case, as shown in Figure 4.15(b). In this case, the initial deeper soil moisture is kept constant at 0.25 for all different initial upper soil moisture values. The response of
Figure 4.15: Simulated daily ET as a function of initial soil moisture for (a) initial deeper soil moisture changing together with upper soil moisture and (b) constant initial deeper soil moisture at 0.25 m$^3$ m$^{-3}$. The simulations are based on the model configurations on 09 November. The daily value of ET is computed as integration of the model outputs (per minute time step). The solid lines are for the control experiments in the two study site, with black colour for the Kyeamba grassland site and green colour for the Tumbarumba forest site. The dashed lines are for the sensitivity experiments.

ET to initial upper soil moisture is similar between forest and grassland. However, ET from the forest surface is less sensitive to soil moisture availability than ET from the grassland surface in both sites. This is in agreement with Nair et al. (2011), although due to a different mechanism. In Nair et al. (2011), the forest ET is less sensitive to soil moisture availability since the trees have access to ground water with the deep root system. In the CLASS model, although both types of vegetation have access to the deep soil layer water, the water demand from the upper soil layer in forest is less due to the higher $C_{veg}$. Overall, ET still has a positive relationship with the initial soil moisture in the upper layer.

Higher soil moistures can increase the chance of the LCL within the CBL. In Kyeamba, the CBL could grow higher than the LCL when the initial water content of the whole soil profile reaches the field capacity. The initial wet and cool conditions
encourage the LCL to occur in Tumbarumba. The effect of initial soil moisture is shown by the amount of time that the CBL takes to reach the LCL. Generally less time is required with higher initial soil moisture. Compared to the grassland surface, the LCL also occurs earlier over the forest surface. For all cases, the LCL is slightly lower over the forest than the grassland, which was consistent with Fisch et al. (2004).

4.10 Summary and discussion

The CLASS model simulates the day time convective boundary layer development, computing the time series of change in the boundary layer conditions and the vertical boundary layer profiles. The model has been applied to two study locations in southern NSW, Kyeamba and Tumbarumba, given their close proximity and similar atmospheric boundary conditions. The two sites are covered by different vegetation types, one with short grasses and the other with Eucalyptus trees. Model simulations are better over the grassland site than the forest site. As shown in Table 4.6, the RMSE of latent heat flux simulation in Kyeamba is 80% lower than in Tumbarumba. It is partially due to the hotter and drier CBL conditions above the short grasses.

In the sensitivity experiments, changes of land surface cover have led to distinct difference in heat distribution. Much higher latent heat fluxes have been found over the forest covers. In addition to heat redistribution, total heat fluxes have also changed due to changes in the surface albedo. For example deforestation has been found to decrease net radiation (Junkermann et al., 2009; Zeng and Yoon, 2009). However the changes in the sensible heat fluxes are not consistent during the day. In the model, transpiration is from the deep soil layer which has constant water content. When transpiration is not active in the cool wet morning, the sensible heat fluxes over the forest are higher as a result of the higher net radiation. Higher sensible heat fluxes over forest (Teuling
et al., 2010) could lead to faster growth of the CBL, which was also seen in Silva Dias et al. (2002). After transpiration becomes active in the afternoon with a wet deep soil, the gain in the net radiation could not satisfy the energy requirement of ET in the forest so the sensible heat flux has to decrease. This knowledge of the evolution of the boundary layer over different land surfaces has important implication for rainfall event forecasting.

The boundary conditions at the top of the CBL control the boundary layer development. In the model, there are two conditions: (1) initial temperature jump at the entrainment layer; (2) potential temperature lapse rate. The initial temperature jump is the difference between the CBL potential temperature and the free atmosphere temperature. The temperature difference prevents entrainment into the upper atmosphere so it can delay the CBL growth (Fisch et al., 2004). On the other hand, without a quick boundary layer growth, air temperature and humidity can have a large increase in the morning. This situation was seen on 10 - 11 November in Kyeamba. The potential temperature lapse rate at the free atmosphere affects the rate of CBL growth. With a small lapse rate, a small increase in the potential temperature can enable the boundary layer to reach a much higher level. But usually as the boundary layer grows fast, temperature increases would be small and humidity decrease can be large. A small lapse rate and a small initial temperature jump are the main reason that the boundary layer growth was strong and associated with a cooler temperature in Tumbarumba on 10 November.

In the case that initial and boundary layer conditions remain the same but the vegetation cover changes, the combination of land cover type and soil moisture availability can also affect the boundary layer growth. The development of the
boundary layer is supported by high energy fluxes, mainly the sensible heat fluxes \cite{Fisch2004, Konings2010}. High air temperature is also important for the convective activity. If the top soil is dry, the CBL is likely to be higher over the short grasses than over the forest, especially in the late afternoon or early evening, due to the higher sensible heat fluxes over the short grasses \cite{Fisch2004}. I also notice this result is only valid when the trees have access to water supply from the deep soil. In the model, the soil evaporation is small over both land surface types when the top soil is dry, but transpiration is higher in forest than in grassland. Higher transpiration in native forest during dry seasons has also been found in southwest Western Australia \cite{Ray2001, Nair2011}. ET consumes a large amount of energy. Even though with a higher net radiation due to a lower albedo, less energy has been contributed to the sensible heat in the forest scenario \cite{SilvaDias2002}. However, the high CBL is not usually sufficient for clouds to form. With a low latent heat flux, deeper convection is required to trigger rainfall \cite{Butt2011}.

The lifting condensation level is used as an indicator of the onset of clouds. The chance of reaching the LCL within the CBL is higher with wetter soil, especially the upper soil. As a wet and cool boundary layer is preferred for the LCL to form \cite{Konings2010, Hauck2011}, high soil moisture could affect the energy partition to trigger moist convective instability \cite{Eltahir1998}. Forest surface can have a similar impact as the wet soil, given it has access to sufficient water supply. The LCL is generally lower above forest \cite{Fisch2004}. Furthermore, turbulent heat exchange and convective cooling above the rough forest surface is more efficient \cite{Teuling2010}, which could also support a high CBL. It implies a higher potential for deep convective cloud formation \cite{Nair2011}.
Previous studies suggest changes in moisture convergence is the primary cause that land cover change affect rainfall (Esau and Lyons 2002; Pitman et al. 2004; Pitman and Hesse 2007). Our study also supports this mechanism. In the case that the dry Kyeamba grassland is forested, when positive advective moisture has been introduced in the model, the chance of reaching the LCL within the convective boundary layer is higher. The important role of roughness length in modifying the moisture convergence has been highlighted in the literature (Pitman et al. 2004; Lee and Berbery 2012). Without considering the influence on the boundary conditions, the local effect of roughness length is very small.

Similarly, the combined effect of changes of surface characteristics on the convective boundary layer growth is possibly small, if there is no change on the boundary. This is due to the various interacting relationships between the vegetation properties and the environment, which in many cases compete with each other and reduce the impact of changes due to individual factors (e.g. Pitman et al. 1993; Narisma and Pitman 2003; Findell et al. 2007; Strengers et al. 2010). These complex feedbacks could be the basis of longer term ecohydrological equilibrium (Hatton et al. 1997). As a result, the difference of the maximum CBL height between the forest and the grassland is usually smaller than 100 m.

The model can produce relatively good estimate of the temperature evolution. The difference between the observed and simulated maximum temperature in Kyeamba was around 0.5°C for all three days. This is probably better than some of the numerical models (e.g. Pitman et al. 2004); but as the CLASS model needs to be initialised daily, the error in the estimate is expected to be smaller. In the afforestation experiment, decrease in the mean surface temperature were approximately 0.2°C which is slightly
less than the range estimated by [McAlpine et al. (2007)] for summer in eastern Australia (0.2 - 2.0°C). But again as the model was initialised daily with the same initial and boundary conditions, the difference between the two experiments could be smaller.

In the deforestation experiment, temperatures were lower over the grassland than the forest on the first two days when the soil was wetter. This situation might be similar to the stage I drying in [Teuling et al. (2010)], in which more energy can be distributed towards latent heat fluxes than sensible heat fluxes when soil moisture is available over the grassland. As the soil became drier on day three, the boundary layer was hotter above the grasses. It implies that LCC can show different impacts on the regional temperature, due to the availability of soil moisture. This might also explain the different temperature changes in summer and winter time found by [McAlpine et al. (2007)].

The availability of the “right” data is again an important issue. As noticed in the current study, flux data in the forest site is measured at a different height from the grassland site, which is also inconsistent with the model simulations. Although inference can be made based on assumptions on the relations between the different parameters, results need to be validated using corresponding measurements. Secondly, information on the vertical profile of the atmosphere during the day is not available since radiosonde measurements are expensive. As the initial and boundary conditions at the top of the CBL are important in predicting boundary layer development, this information is necessary to improve the quality of the model outputs. Furthermore, the vertical profile might also allow the detection of cloud hence the model assumption can be checked.


4.11 Conclusion

In this study, a relatively simple convective boundary layer model, CLASS, is used to simulate boundary layer conditions over three days before a November rain event in a grassland site, Kyeamba and a forest site, Tumbarumba. The model performs relatively well, especially in simulating air temperature in the grassland site. However, the information given by the model, such as the CBL heights and the LCLs, does not imply a possible following rain event. The model estimated stronger boundary layer development on day two rather than the day before the rainfall in both sites. The LCL has been predicted in Tumbarumba in the first two days, but not on the day before the rainfall. It is also possible that the rain event on 12 November was due to frontal system movement rather than the convective activity.

The physical properties of vegetation cover, for example albedo, LAI and roughness length, have different impacts on the boundary layer conditions. Sometimes these impacts are opposite to each other. It is also possible that the final effect of an individual property on the boundary layer conditions is completely different under the interaction with other factors. Furthermore, some parameters can either strengthen or suppress the other parameters’ effect. The overall effect of vegetation cover change would depend on the magnitude of change in each land surface characteristic, as well as the initial conditions of the system. In general, the competition between the different characteristics reduces the impact of individual factors. To evaluate the land surface change effect, it is important to consider the proper set of physical properties that represent the LCC.

The results from the sensitivity experiments in the two study sites do not support the hypothesis that the LCC have a significant impact on the convective boundary layer
Chapter 4. Comparison of boundary layer development over forest and shortgrass growth and the cloud formation. Hence the change in rainfall is likely to be small. This could indicate a weaker land-atmosphere coupling in the semi-arid MDB (Evans et al., 2011). However clouds and rainfall might be affected in regions where moisture convergence can be modified by LCC. Furthermore, the different combinations of land cover and availability of soil moisture can also produce different effects. These regional conditions have important policy implication in land management.
Chapter 5

Vegetation-precipitation feedback in a simple land-atmosphere model

5.1 Introduction

The interactive feedback between land surface and atmosphere is a complicated process. Current studies focussing on land cover change (LCC) effects on climate, especially on rainfall, have shown a large variation in results, as reported in Chapter 2 and previous literature (Pielke et al., 2007, 2011). Multiple factors are suggested contributing to this variation, such as atmospheric characteristics (Pitman et al., 2004, Pielke et al., 2007, Konings et al., 2011), types of land cover change (Narisma and Pitman, 2003, Pielke et al., 2007, Bonan, 2008b), as well as model resolutions and scales (D’Almeida et al., 2007). However, from the previous review, there are still unexplained variations when the studies are grouped based on each factor.

Another possible reason could be the conditional nonlinear response of climatology to land surface changes. As discussed in Chapter 4, the physical parameters that change in the land cover conversion have one-to-many relationships in the land-atmosphere system. The positive and/or negative effect(s) in each relation and their combination and interaction would most likely result in a nonlinear impact on the
climate conditions (Herbert et al., 2010; Teuling et al., 2010; Seneviratne et al., 2010; Lo and Famiglietti, 2011). The nonlinearity is also conditional which reflects the influence of other conditions on the relationship. As an example, the relation between soil moisture and ET is nonlinear if the soil moisture vary within a broad range from almost zero to saturation; however if the soil is always wet (e.g., by irrigation), the soil moisture-ET relationship might be linear as long as the soil moisture is kept within a small range. The conditional response implies that both the type of change and the magnitude of the changes are important in studying LCC effects.

As the problem is highly complex, the current main stream of the land-atmosphere interaction research is based on the complex numerical climate models. With large numerical models, such as RAMS and WRF, the influence of a specific land cover change scenario can be predicted and quantified (e.g. Pitman et al., 2004; Narisma and Pitman, 2006; Sertel et al., 2011; Lee and Berbery, 2012). However, due to the complex structure and large number of parameters in these models, the user controls on the model become weaker so it is difficult to trace down how the changes have affected the results. These types of models tend to become more detailed and complex (Wainwright and Mulligan, 2005) and might limit the understanding of LCC-precipitation feedback relationship. In this case, simple models are desired to investigate the physical mechanisms in controlling vegetation-precipitation interaction. Simple models have been often used to understand behaviour and relationships in the environment system and are found to have good performance (e.g. Guswa et al., 2002; Teuling et al., 2006; Chang et al., 2009).

The aim of this study is to understand the underlying processes in the vegetation-precipitation feedback through investigating atmospheric conditions response to
vegetation cover changes. A simple one-dimensional land-atmosphere model from D’Andrea et al. (2006) and Baudena et al. (2008) is used. Some modifications are made to this model to increase its capacity to represent vegetation. Then a sensitivity analysis is performed to assess the behaviour of the model after changes in the land surface parameters. Finally, an application of the model in a case study in the MDB, Australia, is presented to demonstrate the knowledge gained from the sensitivity analysis.

5.2 A simple 1D model

5.2.1 D’Andrea’s model

The simple 1-D land-atmosphere model in D’Andrea et al. (2006) (DA afterwards) is “an idealized box model of the continental water cycle”, representing the interaction between the planetary boundary layer (PBL) and the surface hydrology. The model assumes constant boundary conditions, such as radiation input, advective heat fluxes and moisture fluxes. Dynamic changes only occur inside the model boundary which is defined as a “box”. Although this model also simulates vertical processes as a one-dimensional model, it is different from the previous CLASS model in terms of its scale. The current land-atmosphere model is for a large spatial scale area such as continental, while the CLASS model can be used at a small local scale. Hence the DA model assumes homogeneous horizontal conditions and estimates average values. Other assumptions include a well-mixed boundary layer and no cloud.

The DA model is a two-layer model which keeps track of the heat and water exchanges between the PBL and the soil. The model consists of a 50 cm deep soil layer and a 1000 m high planetary boundary layer. There is no subdivision within
each layer so this is a “bulk dynamic” system \cite{Baudena et al., 2008}. As in many other models, the heat and water exchanges are driven by temperature differences and air vapour deficit. The evolution of air temperature, humidity, soil temperature and moisture are expressed by differential equations (DEs). At each time step, the changes in temperatures in the air and the soil are the balance of heat input and output. This is similar with the moisture content terms. The four prognostic variables are represented by a set of first-order DEs. These DEs can be solved using an improved Euler method \cite{Dalziel, 1998} with given initial conditions and a time step that is small, i.e. hourly. The other input/output variables, such as ET and heat fluxes, mostly have defined solutions in the model. This means that the input/output terms can be estimated exactly from the current temperature and moisture content rather than approaching an approximation from the previous values.

The advantage of the DA model is that it considers the most important processes but it is simple and easy to understand. As a two-layer model, it simulates heat and water fluxes between the PBL and soil, including sensible heat and latent heat, longwave radiation and ET. In addition to these, the model also simulates convection as a cooling and drying process due to equivalent potential temperature difference between the PBL and the free atmosphere. Precipitation rate is estimated based on the moisture updraft. After rainfall, the processes of runoff, infiltration and leakage occur on/in the soil layer. With all these interactions, the model is a simple complete system of the environment. There are 34 parameters, four DEs for air temperature, soil temperature, air humidity and soil wetness, as well as 14 solved equations for the other variables in the model. Hence this model can be easily managed and controlled. A flow chart of the model structure is shown in Figure 5.1.
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The limitations of the DA model are also obvious when studying vegetation feedback. There is no vegetation layer in the model. The existing vegetation is implicitly represented by the constraints of plants on ET, which are the wilting point and maximum plant efficiency as shown in Figure 5.2. Except ET, vegetation does not influence any other factors in the model system. Hence the function of vegetation cover is oversimplified in this model.
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Figure 5.2: The nonlinear relationship between ET and soil moisture level. ET is evapotranspiration in DA model. $Tr$ and $E_s$ are transpiration and soil evaporation, respectively, in BA model. No ET occurs as soil is below the hygroscopic point $\theta_{s,h}$. Plants start to transpire when soil is above wilting point $\theta_{s,w}$. $\theta_{*}$ is the maximum plant efficiency point at which plants can reach maximum rate of transpiration. $\theta_{s,fc}$ is field capacity and soil evaporation is maximum beyond this point. Here the maximum evaporation rates for plants and soil are assumed to be the same. In DA model, the existing of vegetation has eliminated the effect of field capacity ($\theta_{s,fc}$) constraint. This diagram shows that plant transpiration possibly requires a higher soil moisture level and achieves a higher rate when soil is relatively wet.

5.2.2 Baudena’s model

Following the work of DA, Baudena et al. (2008) (BA afterwards) incorporated vegetation dynamics into the two-layer model. The vegetation cover can increase or decrease as plants grow or die. Its change depends on the soil moisture and the previous vegetation state. The change of vegetation cover also influences the ET and albedo in the land-atmosphere system which will be explained later. The dynamic vegetation cover is implemented as the fifth prognostic variables and it is again modelled by a first-order DE. Compared to the DA model, there are one extra DE for vegetation growth and two more functions which are for ET and albedo respectively.
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The dynamic vegetation model is developed from [Levins (1969)](cited by [Baudena et al., 2007]). This is a site occupancy model with implicit-space. The vegetation state is represented by the fraction of the area that is covered by vegetation. When the environmental conditions are in favour of vegetation growth, the vegetation cover can expand but it will be capped by the total area. The total area is the carrying capacity or maximum capacity for plants. Growth is hence limited and slows down when the area is close to full cover.

\[
\frac{db}{dt} = gb(1 - b) - \mu b \quad (5.1)
\]

The growth part of the vegetation dynamics (first component on the right hand side of Equation [5.1], \(b\) is the vegetation fraction and \(g\) is the growth rate) is modelled by the Verhulst’s logistic growth function ([Tsoularis and Wallace, 2002]). On the other hand, if the environmental conditions are not favourable, vegetations can no longer colonize and they will die. In BA, this occurs when the available soil water is below wilting point. The plants generally have a small natural mortality rate.

The dynamics of vegetation cover is then coupled to the model and affects the atmosphere. In this case, vegetation does not only affect ET but also changes the surface albedo (\(\alpha\)) as shown in Equation [5.2] below.

\[
\alpha = \alpha_b b + \alpha_0 (1 - b) \quad (5.2)
\]

If the surface types are known and the surface albedo values are set (here \(\alpha_b\) is the vegetation surface albedo and \(\alpha_0\) is the bare soil surface albedo), the average surface albedo can be calculated from their fraction of the land surface. With the effect of surface albedo, net solar radiation and the total heat balance change as vegetation...
cover changes. ET is also modelled differently from DA. Here transpiration and soil evaporation have a different response to soil moisture as shown by the green and brown lines in Figure 5.2. According to this model, soil evaporation can occur at a lower soil moisture level than transpiration but the rate of transpiration increase is higher than soil evaporation. When the soil is relatively wet, transpiration rate can exceed soil evaporation rate. The total ET is the weighted sum of transpiration and soil evaporation based on the fractions of each land surface types.

The BA model is an advancer of the DA model with a clearer vegetation representation. The feedback relationship between the vegetation cover and the atmosphere above can be shown in this model. Furthermore, it still contains a low level of complexity in the relationship: increasing vegetation cover can increase both ET and net radiation; but each of them has an opposite impact on the temperature. However the vegetation layer should be well represented by considering its major biophysical properties to study the effect of vegetation on climate.

5.2.3 Further development

Vegetation cover changes can affect the atmosphere condition through albedo (Chagnon and Bras 2005; Falk et al. 2005; Kaufmann et al. 2007; Junkermann et al. 2009), interception (Dickinson and Kennedy 1992; Mao et al. 2011; Carlyle-Moses and Gash 2011) and roughness length (Pitman et al. 2004; Nair et al. 2011), as well as through its influence on ET (e.g. Zhao and Pitman 2002; Narisma and Pitman 2003) and soil moisture (e.g. Andréassian 2004; Junkermann et al. 2009; Mohammad and Adam 2010; Jayawickreme et al. 2011). The models developed by DA and BA have implemented the vegetation effect on soil moisture and albedo, hence little
change is required. Although ET is already included in their model, it is too limited for vegetation cover change study and a higher degree of coupling is necessary. In order to properly represent the vegetation cover without a significant increase in the model complexity, the following modifications are made.

5.2.3.1 Evapotranspiration (ET)

Potential ET

A major limitation in the DA model is that the maximum ET is predefined as a constant. Here the maximum ET tends to be the same as potential ET which is the maximum rate of evaporation over a large area that is completely and uniformly covered by growing vegetation, while the vegetation has access to an unlimited water supply and there is no advection or heat storage effects (Brutsaert, 2005; Dingman, 2008). Although in the model the maximum ET is later regulated by the air vapour deficit, it ignores the commonly accepted fact that the potential evaporation is primarily determined by the availability of energy, either radiation-based or temperature-based (Weiß and Menzel, 2008; Dingman, 2008). If the constant value is estimated using observational data which usually does not reach the potential condition, this value is unlikely to represent the maximum level (Brutsaert, 2005). Hence it would be more appropriate to adapt to operational methods such as the well-known Penman-Monteith equation and the Priestley-Taylor equation.

By definition, the potential ET is only affected by climatic parameters. Although the definition of potential ET is ambiguous and the use of “reference crop ET” is encouraged (Allen et al., 1998; Brutsaert, 2005; Dingman, 2008), potential ET is used here to indicate a maximum rate. It is not supposed to be affected by vegetation characteristics and soil factors. The difference between the Penman-Monteith equation
and the Priestley-Taylor equation for potential ET is that whether a mass transfer term should be included. The details of each method are shown in Box 5.1 and Box 5.2. In both methods, the potential ET is driven by a radiation term. However, the potential ET is also affected by the dryness of the air which is shown as a mass transfer term in Penman-Monteith (the second additive term in the right hand side numerator in Equation 5.3). In Priestley-Taylor, this term is ignored. Their argument is that over a uniform surface with unlimited water supply, the air should have become saturated (Dingman, 2008).

The definition of potential ET reveals that both calculations might underestimate the maximum ET. In the FAO Penman-Monteith method, the reference surface does not account for all different vegetation heights, which can cause the mass transfer term to be underestimated (Wallace and Shuttleworth, 2009). Wallace and Shuttleworth (2009) suggested a way to improve the FAO equation by using a blending height to adjust for different crop heights and measurement heights. The use of the blending height eliminates the need of potential ET estimation. However, although this method is interesting, it does not serve the purpose of using the maximum value. In Priestley-Taylor, without considering the air dryness effect, the method is not able to estimate high ET rates which can occur in dry regions. But using an $\alpha'$ value of 1.74 for arid areas, Weiß and Menzel (2008) found that the Priestley-Taylor method produces the best results in the arid Jordan River basin compared to the other methods, including the Penman-Monteith. Hence the Priestley-Taylor method with $\alpha' = 1.74$ is chosen for the study on the arid/semi-arid MDB.

Here the model requires five more parameters: $\alpha'$, $\Delta$, $\gamma$, $R_n$ and $G$. Except $\alpha'$, the other four parameters can be defined using parameters already in the model.
Box 5.1: Compute potential ET using Penman-Monteith method

The FAO Penman-Monteith equation ([Allen et al., 1998]):

\[
ET_p = \frac{0.017 \cdot \Delta(R_n - G) + \gamma \cdot \frac{37.5}{\theta_a} \cdot u \cdot (e_s - e_a)}{\Delta + \gamma \cdot (1 + 0.34 \cdot u)}
\]  

(5.3)

\( ET_p \): potential ET [mm hr\(^{-1}\)];
\( \Delta \): slope of the saturation vapour pressure temperature curve [kPa K\(^{-1}\)];
\( R_n \): net radiation [J hr\(^{-1}\)];
\( G \): ground heat flux [J hr\(^{-1}\)];
\( \gamma \): psychrometric constant [kPa K\(^{-1}\)];
\( \theta_a \): potential temperature of the PBL [K]. In the FAO 56 paper, this is the temperature at 2 m height. Since it is close to the surface, it is assumed to be the same as the temperature at the reference pressure.
\( e_s \): saturation vapour pressure [kPa];
\( e_a \): actual air vapour pressure [kPa];
\( u \): wind speed at 2 m height [m hr\(^{-1}\)].

The above equation is derived from a more general Penman-Monteith equation, based on a reference surface with an assumed crop height of 0.12 m, a fixed surface resistance of 70 s m\(^{-1}\) and an albedo of 0.23 ([Allen et al., 1998]). Hence this equation is referred to as the “reference crop evapotranspiration” and is increasingly used instead of the ambiguous potential ET ([Allen et al., 1998; Brutsaert, 2005; Dingman, 2008; Weiß and Menzel, 2008]).

The saturation vapour pressure is calculated using the Clausius-Clapeyron law. Here \(e_0 = 0.6108\) is the reference saturation vapour pressure at the reference temperature \(T_0 = 273.15\) K. \(L_e\) is the specific latent heat of evaporation [2.45 \(e\) J Kg\(^{-1}\)] and \(R\) is the dry air gas constant [287 J kg\(^{-1}\)K\(^{-1}\)].

\[
e_s = e_0 \cdot \exp \left(-0.622 \cdot \frac{L_e}{R} \cdot \left(\frac{1}{\theta_a} - \frac{1}{T_0}\right)\right)
\]

\(\Delta\) is the derivative of Clausius-Clapeyron law with respect to the temperature.

\[
\Delta = 0.622 \cdot e_0 \cdot \frac{L_e}{R \cdot \theta_a^2} \cdot \exp \left(-0.622 \cdot \frac{L_e}{R} \cdot \left(\frac{1}{\theta_a} - \frac{1}{T_0}\right)\right)
\]

\(\gamma\) depends on the atmospheric pressure \(P_{hPa}\) [kPa], such that

\[
\gamma = \frac{C_p \cdot P_{hPa}}{0.622 \cdot L_e}
\]

where \(C_p\) is air specific heat. The atmospheric pressure above the plant is calculated as

\[
P_{hPa} = 101.3 \cdot \left(1 - \frac{0.0065 \cdot h_{veg}}{293}\right)^{5.26}.
\]
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Box 5.2: Compute potential ET using Priestley-Taylor method

The Priestley-Taylor equation ([Weiß and Menzel, 2008]):

\[
ET_p = \alpha' \cdot \frac{\Delta}{\Delta + \gamma} \cdot (R_n - G)
\] (5.4)

\(\alpha'\): a factor accounts for the aerodynamic component. In the humid regions, its value is 1.26 ([Dingman, 2008; Weiß and Menzel, 2008]). In the arid regions, its value is 1.7-1.75. The higher values account for advection which is likely to occur in the arid regions ([Weiß and Menzel, 2008]).

\(\Delta, \gamma, R_n\) and \(G\) are the same definitions as in Penman-Monteith.

Partitioning ET

There are two components in ET: soil evaporation and plant transpiration. In the model, each of them is assumed to be capped by an amount, which are summed up to the total potential ET. The maximum contributions of soil evaporation and plant transpiration are determined by the amount of radiation that each type of surface can access ([Porter et al., 2000; Teuling et al., 2005; Vervoort and van der Zee, 2009]). Here a light extinction coefficient \(\kappa\) is used together with LAI to calculate the proportion of moisture evaporated from different surfaces, based on Beer’s law ([Hatton et al., 1993; Leuning et al., 2005]).

The proportion of radiation that reaches the soil surface is

\[e^{-\kappa \text{LAI}}\]

and the rest is intercepted by vegetation as

\[1 - e^{-\kappa \text{LAI}}.\]

Therefore the partitions of potential ET into soil evaporation (\(E_s\)) and plant
transpiration ($Tr$) are

\[ E_{sp} = ET_p e^{-\kappa \text{LAI}} \]  \hspace{1cm} (5.5)

\[ Tr_p = ET_p (1 - e^{-\kappa \text{LAI}}). \]  \hspace{1cm} (5.6)

As shown, LAI is a controlling factor in the evapotranspiration partitions. More vegetation (higher density and/or larger cover) can result in potentially more plant transpiration than soil evaporation.

In summary, the partitioning only requires one additional parameter in the model which is $\kappa$.

**Transpiration**

Plant transpiration in BA is purely based on the availability of soil moisture. This is reasonable for short-rooted grasses in a water-limited environment for a relatively long timescale (Laio et al., 2001). However, a forest could have a different water use mechanism under the same environment (Teuling et al., 2010), which is probably because trees are able to access deep ground water and they have high degree of coupling with the atmosphere (Jarvis and McNaughton, 1986; Vervoort and van der Zee, 2008; Maxwell and Kollet, 2008; Duursma et al., 2011; Nair et al., 2011). In the land-atmosphere model used in this study, the soil layer is shallow and the groundwater component is not included. In order to keep the simple structure of this model, no change is made to the soil layer. Alternatively, I adapt the stomatal conductance approach to reflect the difference between short grasses and tall trees in terms of stomatal control and effect of roughness length.

Stomatal conductance is the amount of water vapour that can move through the leaf stomata pores in a given time. It is limited by the following factors (Jarvis, 1976; Rodriguez-Iturbe and Porporato, 2004):
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- solar radiation ($R_{\text{sin}}$);
- ambient temperature ($\theta$);
- soil-leaf water potential ($\Psi_l$);
- air vapour deficit ($D$);
- CO₂ concentration.

The last factor CO₂ concentration can be assumed to be constant during the day (Rodriguez-Iturbe and Porporato, 2004). So its effect is ignored. Using Jarvis’ method, the stomatal conductance is the product of the functions related to the above factors (Equation 5.7).

$$g_s = \frac{g_{\text{smax}}}{f_{R_{\text{sin}}}(R_{\text{sin}}) f_\theta(\theta) f_{\Psi_l}(\Psi_l) f_D(D)}$$

$g_{\text{smax}}$ is the maximum stomatal conductance. In order to obtain canopy conductance ($G_s$), one method is to multiply $g_s$ by LAI to achieve the areal estimate (Waring and Landsberg, 2011). Without considering the distribution and structure of leaves, this is a big-leaf model.

Each $f()$ is a weighting function evaluating the impact of individual factor respectively, as proportion to the optimal/sufficient value. Hence their values range between zero and one.

$$f_{R_{\text{sin}}}(R_{\text{sin}}) = \frac{1}{1 - e^{a \frac{R_{\text{sin}}}{2}}}$$

$$f_\theta(\theta) = \frac{1}{1 - c(\theta - \theta_{\text{opt}})^2}$$

$$f_D(D) = 1 + \frac{q_{\text{sat}} - q}{q_{\text{sat}}}$$

where $a = -0.005$ m² W⁻¹ and $c = 0.0016$ K⁻² as given in Daly et al. (2004). $R_{\text{sin}}$ is the incoming shortwave radiation. The available energy to leaf, i.e. the photosynthetically active radiation, is assumed to be half of $R_{\text{sin}}$ (Kim and Entekhabi).
\[ \theta_{\text{opt}} \text{ is the optimal temperature at which } g_s \text{ can reach a maximum value.} \]

\[ q_{\text{sat}} \text{ and } q \text{ are the saturation humidity and the specific humidity, respectively. Here } (q_{\text{sat}} - q) \text{ is used to indicate the water vapour deficit as in } \text{Daly et al. (2004).} \]

Instead of using leaf water potential \( \Psi_l \) in the function \( f_{\Psi_l}(\Psi_l) \), the soil moisture \( \theta_s \) is used. Since the function \( f_{\Psi_l}(\Psi_l) \) describes a relative relationship, replacing \( \Psi_l \) by \( \theta_s \) does not change the value of this function (de Arellano and van Heerwaarden, 2012, unpublished). \( f_{\theta_s}(\theta_s) \) is a piecewise linear function (Equation 5.11) which is adapted from the transpiration model in BA.

\[
f_{\theta_s}(\theta_s) = \begin{cases} 
0 & \text{for } \theta_s \leq \theta_{s,w} \\
\frac{\theta^*_s - \theta_{s,w}}{\theta_s - \theta_{s,w}} & \text{for } \theta_{s,w} < \theta_s \leq \theta^*_s \\
1 & \text{for } \theta_s > \theta^*_s
\end{cases} \tag{5.11}
\]

The actual transpiration over a defined area is regulated by the water vapour deficit (Jarvis and McNaughton, 1986; Farquhar and Sharkey, 1982; Schymanski et al., 2007; Whitley et al., 2008) as well as canopy resistance and aerodynamic resistance (Seneviratne et al., 2010; van Heerwaarden et al., 2010). The canopy resistance \( r_s \) is the reciprocal of canopy conductance, i.e. \( r_s = 1/G_s \). However, in some cases \( f_{\theta_s}(\theta_s) \) might be zero value, which will cause \( g_s \) in Equation 5.7 to be undefined. This can be avoided by calculating \( r_s \) using minimum canopy resistance \( r_{s,\text{min}} = 1/(\text{LAI} \cdot g_{s,\text{max}}) \).

Hence

\[
r_s = r_{s,\text{min}} f_{R_{\text{sin}}}(R_{\text{sin}}) f_{\varphi}(\varphi) f_{\Psi_l}(\Psi_l) f_{D}(D) \tag{5.12}
\]

and

\[
Tr = \rho_a L_e \left( \frac{q_{\text{sat}} - q}{r_s + r_a} \right) \tag{5.13}
\]

\( f_{\varphi}(\varphi) \) is applied to situations that \( \varphi \) and \( \theta_{\text{opt}} \) are in a reasonable range, for example, 273-313 K. Furthermore, high \( \varphi \) can only be less than 25 K above \( \theta_{\text{opt}} \).
where $\rho_a$ is the air density. The right hand sides of the last equation are combined into an energy unit. Hence proper conversion will need to be applied to obtain the transpiration which is usually in millimeters.

The advantage of stomatal conductance approach is that transpiration is closely related to both the atmospheric conditions and the soil moisture level. It can describe the active response of plants to their environmental changes relatively well. It provides a way to distinguish different plants through the maximum stomatal conductance (de Arellano and van Heerwaarden, 2012, unpublished). This transpiration model increases the land-atmosphere model’s capacity to compare various land surface types.

Here five more parameters are needed to compute transpiration: $g_{\text{max}}$, $a$, $c$, $\theta_{\text{opt}}$ and $r_a$. $r_a$ can be estimated within the model, as explained in the later section (see “aerodynamic resistance”).

**Soil evaporation**

Accordingly, the soil evaporation model has been changed to include atmospheric conditions. Similar to the transpiration model, the air dryness and aerodynamic resistance can be important in controlling the soil evaporation. On the other hand, the soil evaporation is assumed to be dominated by the soil moisture content as in DA and BA. The soil water availability creates a soil surface resistance for soil evaporation (de Arellano and van Heerwaarden, 2012, unpublished). Therefore, the soil surface resistance can be expressed as

$$r_{\text{soil}} = r_{\text{soil,min}} \cdot g_{\theta_s}(\theta_s)$$  (5.14)
$g_{\theta_s}(\theta_s)$ is again a piecewise linear function of the soil moisture level $\theta_s$. 

\[
g_{\theta_s}(\theta_s) = \begin{cases} 
0 & \text{for } \theta_s \leq \theta_{s,h} \\
\frac{\theta_{s,fc} - \theta_{s,h}}{\theta_{s} - \theta_{s,h}} & \text{for } \theta_{s,h} < \theta_s \leq \theta_{s,fc} \\
1 & \text{for } \theta_s > \theta_{s,fc}
\end{cases}
\]

(5.15)

Combining the soil surface resistance and the aerodynamic resistance, soil evaporation can be modelled as 

\[
E_s = \rho_a L_e \left( \frac{q_{sat} - q}{r_{soil} + r_a} \right)
\]

(5.16)

Only one additional parameter is required to compute soil evaporation: $r_{soil,\min}$.

5.2.3.2 Interception

Vegetation cover can intercept part of incident rainfall and hence it would reduce the actual amount of rain water that reaching the soil. Due to the differences in the canopy structure as well as the precipitation characteristics, the amount of rainfall intercepted varies (Brutsaert, 2005, P100-105). In high rainfall areas such as Chile ($P \sim 2,100$ mm), interception loss in evergreen forest is about 30% compared with 1% loss from shrub cover (Diaz et al., 2007). Canopy interception in sparse Eucalyptus Capillosa (where LAI is around 0.66) is about 8-15% in Western Australia, calculated from the difference between field measurements on precipitation, throughflow and stem flow (Mitchell et al., 2009). In the rainforest in Northern Queensland where canopy cover is dense and local rainfall is quite high ($>2000$ m), annual rainfall interception, again calculated from the field measurements, is 22% while dry season interception is much higher than in the wet season (Wallace and McJannet, 2008). In the floodplains of the semi-arid lower River Murray ($P \sim 250$ mm) within South Australia where a mixture
of temperate evergreen forest coexists, rainfall interception is found to be in the range of 11 - 22% of annual rainfall (Holland et al., 2011, p.12).

Although rainfall interception can be calculated from models such as the Gash interception model with information on incident rainfall, canopy cover and canopy storage (Diaz et al., 2007; Wallace and McJannet, 2008), here a simple method is used to indicate the average values. A parameter $f_c$ is defined as the fraction of rainfall that is intercepted on the vegetated area. Similar with the radiation interception (Lawrence et al., 2007), the water storage on the canopy ($I_P$) can be described as

$$\frac{dI_P}{dt} = f_c \cdot P \cdot (1 - e^{-\kappa \text{LAI}}) - E_I$$

which takes into account of the amount of rainfall intercepted and leaf surface evaporation at any time step. The amount of water on the canopy is also capped by a maximum canopy water storage $I_s$. Wallace and McJannet (2008) estimated that the mean canopy storage capacity in Australian rainforest is between 2.0 and 3.6 mm, while lower values in tropical forests are reported in other literature (Köhler et al., 2007). The canopy water capacity used in this model is assumed to be related to the LAI with a defined maximum capacity ($I_{s,max}$). Hence

$$I_s = I_{s,max} \cdot (1 - e^{-\kappa \text{LAI}}).$$

The evaporation rate from the canopy ($E_I$) is estimated based on the Penman-Monteith method,

$$E_I = \rho_a L_e \left( \frac{q_{sat} - q}{T_a} \right).$$

In this model, the sum of evaporation from canopy interception and transpiration is limited by the potential transpiration partitioning. It is assumed that transpiration only occurs after water on the canopy has completely evaporated since this is a big-
leaf model and assuming homogeneous canopy condition. In order to combine the two
different rates, the following method is used.

\[
E_{tr} = \begin{cases} 
E_I & \text{if } I_P \geq T_{r_p} \\
E_I + \left(1 - \frac{I_P}{T_{r_p}}\right)T_r & \text{if } I_P < T_{r_p}
\end{cases}
\] (5.20)

As a new component in the model, the interception term requires two more
parameters: \(f_c\) and \(I_{s,max}\).

5.2.3.3 Aerodynamic resistance

Roughness length is another important property of vegetation which can influence the
atmospheric conditions when land cover changes (Zhao and Pitman 2002; Pitman et al. 2004; Nair et al. 2011). Its effect is mainly reflected in the aerodynamic resistance. As shown in the calculations of ET (Equation 5.13, 5.16 and 5.19), higher
aerodynamic resistance can reduce ET. Roughness length is different between various
vegetation types mainly due to the vegetation height. Roughness length is considered
to be higher over the native vegetation (mostly trees) to allow for energy transfer (Díaz et al. 2007; Pielke et al. 2011).

The Thom model (Wallace and McJannet 2006; Liu et al. 2006a; Diaz et al. 2007) is used here to estimate the aerodynamic resistance. This model is based on
the information of roughness length and wind speed. Unlike the method used in FAO
56 (Allen et al. 1998), the Thom’s method does not separate the roughness length for
momentum transfer and scalar transfer (heat and vapour). An integrated parameter \(z_0\)
is used. Its expression is shown in Equation 5.21.

\[
r_a = \frac{\ln^2[(z - d)/z_0]}{k^2 \cdot u}
\] (5.21)
where $r_a$ is aerodynamic resistance; $z$ is the reference height at 2 m above the vegetation height $h_{veg}$; $d$ is the zero plane displacement height as $\frac{2}{3}$ of $h_{veg}$; and $k$ is the von Karman’s constant (0.41). The roughness length $z_0$ is taken as 0.123$h$. $u$ is average wind speed within PBL.

To estimate $r_a$, the model needs another four parameters: $z$, $z_0$, $d$ and $k$. The first three can all be estimated from $h_{veg}$.

5.2.3.4 Other changes

Some other changes are necessary for the model to implement new features. Net radiation $R_n$ and leaf area index LAI required in the calculation of ET are not used as variables in either DA model or BA model. The model is modified to include these variables as shown in the following sections.

Radiation balance

In DA and BA, the change of soil temperature is due to the balance of net solar radiation, outgoing longwave radiation, sensible heat and latent heat (Equation (3) in [D’Andrea et al. (2006)]). In another word, the energy balance is

$$ G = R_s - R_l - SH - LE $$

(5.22)

where $G$ is the ground heat flux and $R_l$ is outgoing longwave radiation. Based on this equation, the net radiation is $R_n = R_s - R_l$.

According to [Trenberth et al. (2009)], the downward longwave radiation ($R_{ld}$) accounts for a significant amount of energy input to the land surface. The amount of downward longwave radiation is dependent on the mixed layer temperature using the Stefan-Boltzmann law ([Kim and Entekhabi, 1998a] (see Equation 5.23). If the downward longwave radiation is ignored, it could lead to underestimate of the sensible...
heat fluxes as shown in D’Andrea et al. (2006). Hence I assume on average of 63% of longwave radiation emitted by the PBL is towards and absorbed by the land surface, based on the figures given in Trenberth et al. (2009).

\[ R_{ld} = 0.63\sigma\epsilon_a\theta^4 \] (5.23)

where \( \sigma \) is the Stefan-Boltzmann constant and \( \epsilon_a \) is the emissivity of the atmosphere. According to the Kirchhoff’s law, a substance’s emissivity and its absorptivity are the same (Oke, 1987, p.12). Hence the value of \( \epsilon_a \) is already given in DA and BA.

The inclusion of the downward radiation as in Equation 5.23 in the BA model, without any other change, has shown to increase the sensible heat estimate. I found that the new sensible heat has a higher value when initial soil moisture is lower, compared to D’Andrea et al. (Figure 3 in 2006). When an \( \alpha = 0.25 \) is introduced in DA, results from DA become all negative. The high (low) sensible heat estimate in the new method is more than 70 W m\(^{-2}\) (30 W m\(^{-2}\)) higher than the results from BA. Overall, the new method produces a more reasonable range of heat fluxes and can be adapted to the new features of the model.

**Vegetation dynamics**

The vegetation dynamics model in Baudena et al. (2008) is based on Levins’ model (Levins, 1969). The vegetation status is represented by a horizontal site occupancy \( b \). In order to calculate transpiration, LAI is used instead of \( b \) in the new model. LAI measures the total leaf area (one side) above a unit area. Hence LAI can be greater than the ground surface while the value of \( b \) is strictly less than one. In this case the maximum space that the vegetation can grow can determine the maximum value of LAI.
In the Levin’s model, plant expansion is limited by a carrying capacity following a logistic growth. Based on the same idea as in Levin’s model, the new leaves occupy the available spaces and the falling leaves give up some spaces. Instead of defining the carrying capacity as the total ground area, here it is the whole space capacity to allow for layers of leaves. This is defined as the maximum LAI. The determinant factor of the space is the availability of radiation. It is assumed that leaves at the lowest layer will drop if there is no access to sunlight. Furthermore, the bottom layer of leaves is assumed to have the same access to sunlight as the soil surface underneath. If no light can be intercepted at this layer, LAI reaches its carrying capacity, or maximum LAI. Here I also assume nutrition is not a limited factor of growth as in Baudena et al. (2007). Therefore

\[ e^{-\kappa \text{LAI}_c} \rightarrow 0 \]

where \( \text{LAI}_c \) is the maximum value. The decay of \( e^{-\kappa \text{LAI}} \) with LAI for different \( \kappa \) values are shown in Figure 5.3.

In this study, the maximum LAI is the value that results in

\[ e^{-\kappa \text{LAI}_c} = \epsilon_l \]  \hspace{1cm} (5.24)

where \( \epsilon_l = 0.01 \). Therefore,

\[ \text{LAI}_c = \frac{\ln(0.01)}{-\kappa}. \]  \hspace{1cm} (5.25)

The maximum LAI calculated from this method is about 10. This value is higher than the LAI in the rainforest and Eucalyptus forest reported in Australia or most other global measurements of LAI (Whitehead and Beadle, 2004; Wallace and McJannet, 2008; Garrigues et al., 2008; Schiffman et al., 2008). But it is acceptable to use as a potential value.
The dynamic vegetation dynamic model then becomes

\[
\frac{d\text{LAI}}{dt} = gr(\theta_s) \cdot \text{LAI}(1 - \frac{\text{LAI}}{\text{LAI}_c}) - \mu(\theta_s) \cdot \text{LAI}
\]

\[
= gr(\theta_s) \cdot \text{LAI}(1 - \frac{\text{LAI}}{ln(0.01)/(-\kappa)}) - \mu(\theta_s)\text{LAI} \tag{5.26}
\]

where \(gr(\theta_s)\) is the growth rate and \(\mu(\theta_s)\) is the extinction rate as functions of \(\theta_s\), as defined in Levins (1969).

### 5.3 Sensitivity analysis

#### 5.3.1 Method

In the new model, there are 10 more parameters which require additional information compared to the DA model. In addition to those, initial values are needed for six internal variables. One simple terminology of “factor” is used to represent both parameters and variables in the sensitivity analysis, as in Campolongo and Saltelli.
Table 5.1: Factors in the new model and their values used in the sensitivity analysis.

<table>
<thead>
<tr>
<th>Group</th>
<th>Variable</th>
<th>Description</th>
<th>Min</th>
<th>Max</th>
</tr>
</thead>
<tbody>
<tr>
<td>Atmosphere</td>
<td>$R_{sin}$</td>
<td>Incoming radiation [W m$^{-2}$]</td>
<td>350</td>
<td>800</td>
</tr>
<tr>
<td></td>
<td>$F_q$</td>
<td>Lateral flux convergence [mm hr$^{-1}$]</td>
<td>-0.01</td>
<td>0.1</td>
</tr>
<tr>
<td>Vegetation</td>
<td>LAI0</td>
<td>Initial vegetation cover</td>
<td>0.1</td>
<td>10</td>
</tr>
<tr>
<td></td>
<td>$h_{veg}$</td>
<td>Vegetation height [m]</td>
<td>0.01</td>
<td>20</td>
</tr>
<tr>
<td></td>
<td>$g_{smax}$</td>
<td>Maximum stomatal conductance [m s$^{-1}$]</td>
<td>0.018</td>
<td>0.040</td>
</tr>
<tr>
<td></td>
<td>$f_c$</td>
<td>Fraction of rainfall intercepted</td>
<td>0.05</td>
<td>0.2</td>
</tr>
<tr>
<td>Soil</td>
<td>$\theta_s0$</td>
<td>Initial soil wetness</td>
<td>0.12</td>
<td>0.6</td>
</tr>
</tbody>
</table>

(1997). Here seven factors are my main interest, which are shown in Table 5.1. The main focus is on the factors in the vegetation group, while other important factors are included as well. These factors are assumed to be independent from each other. Each of them might or might not have an individual impact on the model outputs. However, their influence could also be constrained by the values of the other factors. The sensitivity analysis method should be able to assess the single effect and combined effects.

The sensitivity methods can be classified into local methods and global methods. For a model $F$ with multiple factors denoted by $x_1, x_2, x_3, ..., x_k$, its output changes, with respect to a particular factor $x_i$ ($i \in [1,k]$) while all the other factors are held constants, is studied as the local sensitivity of the model $F$ (Sobol', 2001; van Griensven et al., 2006). The local sensitivity can be represented by a sensitivity index ($S$) as the partial derivative of $F$. Here the relative form from van Griensven et al. (2006) is used with the advantage that it can be used to compare different model outputs and different factors (see Equation 5.27).
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\[ SI = \frac{F(x_1, \ldots, x_i + \Delta x_i, \ldots, x_k) - F(x_1, \ldots, x_i, \ldots, x_k)}{F(x_1, \ldots, x_i, \ldots, x_k)} \frac{\Delta x_i}{x_i} \]

(5.27)

Local methods are usually easy to apply and interpret hence they are popular in many studies. The results presented in D’Andrea et al. (2006) are a local sensitivity analysis of model outputs to initial soil moisture changes. On the other hand, global methods deal with changes not limited to one factor but a group of factors. For example, the study of Baudena et al. (2008) is on the combined effect of initial soil moisture and initial vegetation cover on the model outputs. This is a simple type of global sensitivity analysis estimating the second-order sensitivity. The computation cost of global sensitivity analysis can be much higher than local sensitivity analysis, depending on the number of parameters of interest. Hence efficiency is a main problem in the global sensitivity analysis (Campolongo and Saltelli, 1997; van Griensven et al., 2006).

In this study, a simple global method similar to the one in BA is first used to compare results from the new model, DA’s model and BA’s model. In the simple method, 25 initial soil wetness and 24 initial LAI values are used with 25×24 model runs to obtain the equilibrium results. The initial soil wetness ranges from 0.12 to 0.6, which is slightly larger than the distance between the hygroscopic point and the saturation point. The initial LAI ranges from 0.1 to 10 with a constant vegetation height of 0.12 m, assuming grasses cover. The mean sensitivity index for each factor of the initial soil wetness and the initial LAI is constructed by averaging the \( SI \) for one parameter given a fixed value of the other, as in Equation (5.27) over the range of...
other factors. Hence,

\[ \overline{ST}_{x_i} = \int_{x_j} SI_{x_i|x_j} \, dx_j \]  

(5.28)

In order to provide information on the combined effects of multiple factors, a more advanced global method is also used. Here the Sobol’ sensitivity index is adopted. This sensitivity index is developed by Sobol’ (1990) and followed by a series further studies by Saltelli, Sobol’ and some other mathematicians in the early 1990s (Archer et al., 1997). The Sobol’ sensitivity analysis is becoming more popular and widely used in today’s complex environmental models due to its quantitative qualities and model-free properties (Dimov and Georgieva, 2010; Nossent et al., 2011).

As a variance-based sensitivity analysis method, the Sobol’ method quantifies the amount of variance contributed by an individual factor to the unconditional variance of the model and represents this variance as a fraction of the total variance (Nossent et al., 2011). In this case, the unconditional variance refers to total variance under all possible situations given the changes in the other factors. The Sobol’ sensitivity analysis mainly consists of three parts: decomposition, calculation of variance and sampling. Details and new developments for each components are given in the literature (e.g. Archer et al., 1997; Campolongo and Saltelli, 1997; Sobol’, 2001; Saltelli et al., 2010; Dimov and Georgieva, 2010; Nossent et al., 2011).

In this study, the program scheme written in an R package is used (“sobol2007” Pujol et al., 2012). The function “sobol2007()” is based on the improved formula in Sobol’ et al. (2007) and Saltelli et al. (2010), implementing the Monte Carlo integrals (Nossent et al., 2011) and calculating first-order index and total index at the same time. The first-order index accounts for the partial effect of individual factor \( x_i \) (van Griensven et al., 2006; Nossent et al., 2011). The total sensitivity index measures the
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total effects (including first-order and higher order interactive effects) of a factor $x_i$. In this study, two sets of values for the seven factors, each of which is consisted of 12,000 random combinations, are used to run the model. According to [Nossent et al. (2011)], this is the sample size necessary to assure convergence of the Sobol’ sensitivity index $S_i$. There are in total of 108,000 ($= 12,000 \cdot (7 + 2)$) model runs. In order to sample values within the ranges shown in Table 5.1 without bias, the values of each factor in each model run are selected using the random number generator in R ([R Development Core Team, 2012]).

5.3.2 Results

5.3.2.1 Compare results from different studies

The land-atmosphere system converges to a stable condition after running the model over multiple steps. [D’Andrea et al. (2006)] obtained two equilibrium states by varying the initial soil moisture from 0 to 1 while the other conditions were constants. When the initial soil moisture is below a certain value (in DA’s case, below 0.33), the land-atmosphere system is in a hot and dry state. When the initial soil moisture is above that value, the system switches to a wet and cool state. Further in [Baudena et al. (2008)], with the inclusion of dynamic vegetation, they found the equilibrium state of the system is sensitive to both the initial soil moisture and initial vegetation cover. While the initial vegetation cover is low, there is only one dry state even when the initial soil is saturated (see Figure 1 in [Baudena et al. (2008)]). The temperature is in the high range but is not the highest when the vegetation cover is low. This is probably due to the negative albedo effect in the heat balance.

2 The soil hydraulic properties in DA: $\theta_{s,h} = 0.14$, $\theta_{s,w} = 0.18$, $\theta_{s}^* = 0.46$, $\theta_{s,fc} = 0.56$. 
With relatively consistent initial and boundary conditions, the new model can also produce multiple equilibrium states but the outputs are different from the previous literature. As shown in Table 5.2, for example, the equilibrium potential temperature in the PBL has two stable states at 288.5 K and 300.5 K, which are similar to DA and BA\(^3\) but the dry state soil moisture is much lower. While in DA the dry state soil moisture is above wilting point and in BA it is slightly below, in the new model this value is just above the hygroscopic soil moisture level. Furthermore, in the new model the initial vegetation cover has limited impact on the equilibrium states (see Figure 5.4). There is no dry state caused by the low initial vegetation cover when initial soil wetness is higher than 0.2. Although as initial vegetation cover increases, the system tends to achieve a wet state with lower initial soil moisture level. This behaviour is not as obvious as in BA.

![Figure 5.4](image-url)

Figure 5.4: The bi-equilibria outputs of the air potential temperature (K) and the soil wetness (as proportion to saturation) from the new model, when initial soil wetness changes from 0.12 to 0.60 and initial LAI changes from 0.1 to 10. The soil hydraulic properties are the same as in DA.

\(^3\) The initial conditions and boundary conditions used to run the model in DA and BA might not be the same as those in their papers. Secondly the definition of equilibrium here is different from DA. Therefore the results achieved are generally higher than the results presented in their papers.
Table 5.2: Compare equilibrium results from the models of DA, BA and the new model. The last two columns report the sensitivity indices with respect to initial soil wetness and initial LAI. The sensitivity indices are calculated using Equation 5.28.

| Variables | Dry state | Wet state | | | |
|-----------|-----------|-----------|-----------|-----------|
|           | DA        | BA        | New       | DA        | BA        | New       | $SI_{\theta,s}$ | $SI_{\text{LAI}}$ |
| $\theta$ [K] | 301.9     | 296.1     | 300.5     | 292.4     | 290.8     | 288.5     | -0.01        | 0.00         |
| $T_s$ [K]  | 303.1     | 295.2     | 300.9     | 292.9     | 290.7     | 289.2     | -0.01        | 0.00         |
| $q$ [g/kg] | 1.5       | 3.7       | 2.1       | 5.4       | 6.0       | 7.1       | 0.90         | 0.07         |
| $\theta_s$ [-] | 0.19     | 0.17      | 0.14      | 0.56      | 0.58      | 0.58      | 1.13         | 0.08         |
| $b$ [/max] | na        | 0.0       | 0.0       | na        | 0.8       | 0.8 $2.08 \times 10^{34}$ | $1.51 \times 10^{33}$ |
| ET [mm d$^{-1}$] | 0.3       | 0.3       | 0.3       | 3.1       | 2.7       | 5.2       | 6.49         | 0.47         |
| $P$ [mm d$^{-1}$] | 0.3       | 0.3       | 0.3       | 3.5       | 2.9       | 5.4       | 6.83         | 0.50         |
The new model generates higher ET than the previous models, as shown in the wet state (Table 5.2). This is due to a combination of reasons. The evaporation from interception which is not a component in the previous models contributed to about 5% of the total ET. The sum of transpiration and soil evaporation in the new model is determined by three more factors than that in BA, which are the aerodynamic resistance, solar radiation and ambient temperature. Hence the new ET model is more sensitive to the atmospheric conditions. Furthermore, the values of \( g_{\text{smax}} \) and \( r_{\text{soil,min}} \) also have a strong influence on ET. The results shown in Table 5.2 are based on the maximum stomatal conductance value given in Daly et al. (2004). In our experiment, similar results with the previous models can be obtained with much smaller conductance. But such values of conductance are outside the range reported in the literature (e.g. Hatton et al., 1993; Schulze et al., 1994; Hales et al., 2004; Daly et al., 2004). Nevertheless, the amount of ET and precipitation estimated by the new model might be more reasonable, for a wet state equilibrium for summer (equivalent to about 500 mm for the wet summer season).

The small impact of initial vegetation cover in the new model is due to the different function used. In BA the effect of vegetation cover is modelled linearly. For example, the albedo is estimated using Equation 5.2, hence the change of vegetation cover \( b \) has a linear effect on the surface albedo. In the new model this has changed to

\[
\alpha = \alpha_k(1 - e^{-\kappa \text{LAI}}) + \alpha_0 e^{-\kappa \text{LAI}}
\]  

(5.29)

where LAI has an exponential effect on albedo. For small LAI values, the weight of vegetation has a large increase even though LAI only changes a little. But on the other hand, as LAI becomes larger, the weight of vegetation approaches to a constant value. This explains the higher chance in the wet state compared to BA.
D’Andrea et al. (2006) also pointed out the effect of the flux convergence ($F_q$) on model output. The multiple equilibrium states occur within a certain range of $F_q$. Outside this range, the system can only achieve a single state, either dry/hot or wet/cool. In the new model, it is also tested for different $F_q$ and $R_{sin}$ values. Similar behaviour is found with changing $F_q$ over a large range. In terms of radiation, when $R_{sin}$ is low, the land-atmosphere system would result in a hot and dry state. This is reasonable, since convection is the major process for heat and water exchange in this model. Without enough heat input, convective activities and rainfall are suppressed. So there is only one steady state which is hot and dry. On the other hand, if $R_{sin}$ is extremely high, the higher initial vegetation cover could have a negative impact on the model equilibrium. As discussed in van Griensven et al. (2006), the model’s sensitivity to a particular parameter is usually relative, which means its influence on the model output may also depend on the other parameters. Therefore it is necessary to conduct a global sensitivity analysis when investigating the influence of some initial or boundary conditions of the model.

5.3.2.2 Sobol’ sensitivity analysis

Results from the Sobol’ analysis are presented in Table 5.3. The factors are listed by the order of the first order sensitivity index $S_j$. The total sensitivity index is denoted by $S_{Tj}$. The 95% bootstrap confidence intervals are also given for each index. The bootstrap averages of each of the sensitivity indices are not shown; but the bias between the Sobol’ values and the bootstrap values are all below $10^{-3}$. Hence the bootstrapping is regarded as unbiased (Nossent et al., 2011).

There is a very distinct difference between the factors in terms of their impacts on the precipitation. The Sobol’ indices reveal that the effect of the lateral moisture
flux $F_q$ is the highest among all the factors under tested, for both first order index and the total index. The critical role of $F_q$ in the multiple states of the model has been highlighted in [D’Andrea et al. (2006)]. This parameter represents a large scale effect in the model. It is consistent with many other studies that the large-scale circulation plays an important role, maybe the most important, in the quantity of precipitation ([Bengai et al. 1993; DeAngelis et al. 2010]). The second most important factor is the shortwave radiation $R_{sin}$. For precipitation, the higher heat energy enhances the convective activities and therefore it can improve precipitation. The Sobol’ analysis suggests water limitation is more important than energy limitation in this environment. Initial soil moisture and initial LAI both have a small influence on the output, with the sensitivity index of LAI0 is about 10% of $\theta_s0$. This is also consistent with the previous result that LAI0 is not as important as $\theta_s0$.

The last three factors are almost insignificant in the model outputs. These are shown by their 95% CI where zero is within the range. The first order index of the rainfall interception fraction, $f_c$, is slightly higher than the other two while the total index of the vegetation height, $h_{veg}$, is higher. Furthermore, there is an unexpected

### Table 5.3: Results from the Sobol’ analysis for the seven factors.

<table>
<thead>
<tr>
<th>Factor</th>
<th>$S_i$</th>
<th>95% CI ($S_i$)</th>
<th>$S_{Ti}$</th>
<th>95% CI ($S_{Ti}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$F_q$</td>
<td>0.658</td>
<td>(0.628, 0.688)</td>
<td>0.806</td>
<td>(0.774, 0.838)</td>
</tr>
<tr>
<td>$R_{sin}$</td>
<td>0.142</td>
<td>(0.124, 0.161)</td>
<td>0.235</td>
<td>(0.215, 0.255)</td>
</tr>
<tr>
<td>$\theta_s0$</td>
<td>0.021</td>
<td>(0.010, 0.032)</td>
<td>0.083</td>
<td>(0.074, 0.093)</td>
</tr>
<tr>
<td>LAI0</td>
<td>0.002</td>
<td>(-0.002, 0.006)</td>
<td>0.010</td>
<td>(0.007, 0.012)</td>
</tr>
<tr>
<td>$f_c$</td>
<td>0.0004</td>
<td>(-0.0010, 0.0019)</td>
<td>0.0002</td>
<td>(-0.0018, 0.0022)</td>
</tr>
<tr>
<td>$h_{veg}$</td>
<td>-0.0002</td>
<td>(-0.0010, 0.0006)</td>
<td>0.0004</td>
<td>(-0.0006, 0.0014)</td>
</tr>
<tr>
<td>$g_{smax}$</td>
<td>0.000</td>
<td>(0.000, 0.000)</td>
<td>0.000</td>
<td>(0.000, 0.000)</td>
</tr>
</tbody>
</table>
negative value of $S_i$ for $h_{\text{veg}}$. As the Sobol’ indices indicate the contribution of each factor to the total variance of the model outputs, the negative value might seem to be impossible. According to [Archer et al.] (1997), although it is theoretically impossible, a negative value can occur due to the Monte Carlo approach and again it does not affect the results of the important factors. The maximum stomatal conductance, $g_{\text{smax}}$, does not show any impact on the precipitation output. Although $g_{\text{smax}}$ is an important factor that physiologically controls transpiration ([Farquhar and Sharkey] 1982; [Schulze et al.] 1994), its effect is balanced out by soil moisture depletion in the model. Increasing $g_{\text{smax}}$ could decrease canopy resistance $r_s$, followed by a rise in transpiration and ET. An immediate effect is that soil moisture is depleted. Hence $f_{\theta_s} (\theta_s)$ (Equation 5.11) increases and $r_s$ increases (see Equation 5.12). Both of these influences are nonlinear, therefore the final effect of $g_{\text{smax}}$ is almost eliminated.

5.3.2.3 Effects of vegetation characteristics

The model represents a simple land-atmosphere system. It reveals that, even in this simple model, there are combinations of linear and nonlinear effects which can result in complicated relationships. A relationship diagram could best describe the causes and the effects, although it might still be difficult to demonstrate the complete picture. As shown in Figure 5.5 - 5.6, the effect of vegetation changes via LAI, vegetation height ($h_{\text{veg}}$), canopy interception fraction ($f_c$) and maximum stomatal conductance ($g_{\text{smax}}$) on precipitation is passed through multiple intermediate processes.

Here LAI, $h_{\text{veg}}$, $g_{\text{smax}}$ and $f_c$ are assumed to increase in an afforestation scenario. Higher $h_{\text{veg}}$ and $g_{\text{smax}}$ can increase ET through their nonlinear effects on the aerodynamic resistance $r_a$ and canopy resistance $r_s$ (Figure 5.6). However, more intense canopy interception ($I_p$) either by higher LAI and/or higher $f_c$ can reduce
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Figure 5.5: The flowchart demonstrates the series of changes due to the increase in LAI. See Table 5.4 for the definition of symbols. The blue arrows indicate positive coupling. The red arrows indicate negative coupling. Associated with the connection, the small coordinates show the types of relationships: linear or nonlinear. A dashed line with arrow indicates there is a potential effect which depends on the other conditions. For example, higher potential transpiration can increase actual transpiration if the actual transpiration is equal to the potential value. The increase of LAI can have opposite effect on some processes which are highlighted by thick red arrows: ET, SH, $T_s$, $\Delta q_s$. It also has a negative effect on itself through rainfall interception (shown in Figure 5.6). Hence the final effect is unclear. If $\Delta q_s$ increases eventually, precipitation increases. This relationship is positive as shown in the dashed line box. The key messages in this figure are: 1. Although most relationships are linear, there are some nonlinear relationships; 2. There are many small feedback loops between LAI and precipitation. Some feedbacks strengthen the effect and some weaken the effect. This figure demonstrates the complexity in determining the effect of LAI on precipitation.
Table 5.4: Definition of the symbols in the flow chart.

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\alpha$</td>
<td>Surface albedo</td>
</tr>
<tr>
<td>$\delta q_{sat}$</td>
<td>Derivative of saturation humidity with respect to temperature</td>
</tr>
<tr>
<td>$\Delta q_a$</td>
<td>Convective moisture</td>
</tr>
<tr>
<td>$\Delta \theta_a$</td>
<td>Temperature change associated with convective cooling</td>
</tr>
<tr>
<td>$E_i$</td>
<td>Evaporation from canopy interception</td>
</tr>
<tr>
<td>$ET_p$</td>
<td>Potential evapotranspiration (ET)</td>
</tr>
<tr>
<td>$E_{sp}$</td>
<td>Maximum soil evaporation</td>
</tr>
<tr>
<td>$E_s$</td>
<td>Soil evaporation</td>
</tr>
<tr>
<td>$I_p$</td>
<td>Canopy interception</td>
</tr>
<tr>
<td>$P$</td>
<td>Precipitation</td>
</tr>
<tr>
<td>$P_{eff}$</td>
<td>Effective precipitation</td>
</tr>
<tr>
<td>$q$</td>
<td>Air humidity</td>
</tr>
<tr>
<td>$q_{sat}$</td>
<td>Saturation humidity</td>
</tr>
<tr>
<td>RH</td>
<td>Relative humidity</td>
</tr>
<tr>
<td>$R_n$</td>
<td>Net radiation</td>
</tr>
<tr>
<td>$r_a$</td>
<td>Aerodynamic resistance</td>
</tr>
<tr>
<td>$r_s$</td>
<td>Canopy resistance</td>
</tr>
<tr>
<td>SH</td>
<td>Sensible heat</td>
</tr>
<tr>
<td>$Tr_p$</td>
<td>Maximum transpiration</td>
</tr>
<tr>
<td>$Tr$</td>
<td>Transpiration</td>
</tr>
<tr>
<td>$T_s$</td>
<td>Surface temperature</td>
</tr>
<tr>
<td>$\theta$</td>
<td>Potential temperature</td>
</tr>
<tr>
<td>$\theta_e$</td>
<td>Equivalent potential temperature</td>
</tr>
</tbody>
</table>
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Figure 5.6: The flowchart demonstrates the series of changes due to the increase in $g_{s\text{max}}$, $h_{\text{veg}}$ and $f_c$. The symbols are the same as in Figure 5.5. The effects after changes in ET and LAI can be found in Figure 5.5.

effective rainfall. Hence less water is available for plant transpiration. Eventually LAI itself could decrease. LAI has an exponential effect on the canopy interception (Figure 5.5), while $f_c$ has a linear effect on $I_p$ (Figure 5.6). Further processes influence $\theta_s$ linearly, but the effects of $\theta_s$ on ET and LAI are nonlinear again. Although ET might increase within a short period of time as a result of larger LAI, $g_{s\text{max}}$ and $h_{\text{veg}}$, the effect of vegetation on soil moisture and air humidity can decrease ET. The concurrent opposite effects are also seen on surface temperature ($T_s$), sensible heat flux (SH) and the convective drying rate $\Delta q_{\text{a}}$. In the model, the amount of $\Delta q_{\text{a}}$ determines the precipitation rate $P$. Higher $\Delta q_{\text{a}}$ leads to higher $P$ and this relationship is also nonlinear.

The final effect of vegetation changes on precipitation is therefore not clear. The model’s response to a particular vegetation change might be strengthened by some
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intermediate processes. Especially due to the nonlinear relationships, the initial and boundary conditions, the stages of the changes and the feedback relationship with the other processes could all make a difference on the final results. Hence a conclusion cannot be drawn on the theoretical framework. In the next section, the model is applied in a case study to evaluate the effect of vegetation changes on precipitation.

5.4 Case study

5.4.1 MDB

The MDB has an area of more than 1M km$^2$, about 14% of the total area of Australia ([Murray-Darling Basin Authority](#), 2008). The basin is important for the Australian agriculture industry, which contributes more than 30% of food products in Australia and covers 65% of the Australia’s irrigated land. Hence various water issues, such as water sources, storage, distribution and usage are critical in the basin.

Here the first issue of water sources is most interest in this study. The basin is mostly located in the arid/semi-arid climate zone, according to the Köppen-Geiger climate classification. Precipitation is the main water source. Based on the Murray-Darling Basin Water Resources Fact Sheet (released in November 2003), average annual rainfall in the basin is 480 mm, while the potential ET is about 1,968 mm year$^{-1}$. The potential ET is much higher than the precipitation.

In this case study, the model is applied to the MDB under summer conditions. The model has run for 5,000 steps (step size: hour) as a warm up process which is sufficient to reach a steady state. In the no change scenario, the model has run continuously until equilibrium is reached. In the complete vegetation cover change scenario, the value of LAI, $h_{veg}$, $g_{smax}$ and $f_c$ are changed to a higher value then the model has again run
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until equilibrium is reached. The purpose of this case study is to predict the potential stable state of the MDB climate with and without afforestation.

Values of some of the factors in the model have changed from DA and BA according to a theoretical summer daytime situation in the MDB. These values are shown in Table 5.5. Most of the changes are in the soil hydraulic properties. A mixture of different soil types is present in the MDB. Here it is assumed that the average soil type is sandy clay loam with a high clay content for the whole region (Kingham, 1998). According to the information provided by CSIRO Land and Water on the Australian Soil Resource Information System (ASRIS, http://www.asris.csiro.au/mapping/viewer.htm), in the northeast and south of the basin the clay content can be more than 45% in the upper 0 - 30 cm soil. Contributions from the maritime lateral flux is expected to be small due to the inland position of MDB blocked by the Great Dividing Range on the east and bounded by a desert area on the west. Here the lateral flux is assumed to be 0.4 mm day$^{-1}$.

In the MDB scenario, dry condition is expected over the long run with and without the afforestation according to the model simulations. Soil moisture falls from initial condition to just above the hygroscopic point. The air becomes extremely dry with only 1% saturation. These situations result in, and are influenced by, a low precipitation of about 0.1 mm day$^{-1}$. The model estimates higher ET than precipitation at the beginning, which is one of the main reasons leading to the dry equilibrium. Interestingly the vegetation cover changes affect the soil moisture, humidity and precipitation over much a shorter period than the case of the temperatures (e.g. Figure 5.7(a) and 5.7(c)). But all the variables approach the equilibrium as the scenario of no vegetation changes. It appears that the surface change might have a

4 In the model, after the warm up process, this value gives a daily rainfall of 3.3 mm which is equivalent to about 300 mm for summer in the MDB.
Table 5.5: List of factors used in the Soil-Vegetation-Atmosphere Box Model for the MDB region (summer daytime). The soil moisture values $\theta_{s,h}$, $\theta_{s,w}$, $\theta^*$ and $\theta_{s,fc}$ are calculated based on soil hydraulic properties for sandy clay loam in [Vervoort and van der Zee (2008)], supposing the soil water potential of -10 MPa, -3 MPa, -0.03 MPa and -0.01 MPa respectively.

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\theta_0$</td>
<td>Initial PBL potential temperature [K]</td>
<td>303</td>
</tr>
<tr>
<td>$T_{s,0}$</td>
<td>Initial surface layer average temperature [K]</td>
<td>305</td>
</tr>
<tr>
<td>$q_0$</td>
<td>Initial PBL specific humidity [g kg$^{-1}$]</td>
<td>5</td>
</tr>
<tr>
<td>$\theta_{s,0}$</td>
<td>Initial soil moisture [-]</td>
<td>0.4</td>
</tr>
<tr>
<td>LAI0</td>
<td>Initial LAI [m$^2$m$^{-2}$]</td>
<td>0.5</td>
</tr>
<tr>
<td>$R_{\text{sin}}$</td>
<td>Incoming radiation [W m$^{-2}$]</td>
<td>830</td>
</tr>
<tr>
<td>$F_q$</td>
<td>Lateral flux convergence [mm hr$^{-1}$]</td>
<td>0.018</td>
</tr>
<tr>
<td>$u$</td>
<td>PBL average wind speed [m s$^{-1}$]</td>
<td>4</td>
</tr>
<tr>
<td>$\theta_{a,*}$</td>
<td>Relaxation temperature for PBL [K]</td>
<td>300</td>
</tr>
<tr>
<td>$\rho_a$</td>
<td>Air density [kg m$^{-3}$]</td>
<td>1.15</td>
</tr>
<tr>
<td>$\rho_s$</td>
<td>Soil density [kg m$^{-3}$]</td>
<td>1400</td>
</tr>
<tr>
<td>$K_s$</td>
<td>Saturated hydraulic conductivity [kg m$^{-2}$ hr$^{-1}$]</td>
<td>22</td>
</tr>
<tr>
<td>$\phi$</td>
<td>Soil porosity [-]</td>
<td>0.37</td>
</tr>
<tr>
<td>$b$</td>
<td>Retention curve parameter [-]</td>
<td>6.41</td>
</tr>
<tr>
<td>$\theta_{s,h}$</td>
<td>Hygrosopic point soil moisture [-]</td>
<td>0.24</td>
</tr>
<tr>
<td>$\theta_{s,w}$</td>
<td>Wilting point soil moisture [-]</td>
<td>0.30</td>
</tr>
<tr>
<td>$\theta^*$</td>
<td>Soil moisture for maximum plant efficiency [-]</td>
<td>0.61</td>
</tr>
<tr>
<td>$\theta_{s,fc}$</td>
<td>Field capacity [-]</td>
<td>0.72</td>
</tr>
</tbody>
</table>
Figure 5.7: The evolution of (a) potential air temperature $\theta$, (b) LAI and (c) precipitation $P$ in the model in the first 50,000 steps (after warm-up). The solid lines are simulation outputs with no vegetation changes. The dashed lines are simulation outputs with vegetation changes. In (c), the inserted plot shows the difference between the two scenarios in the first 50 steps. The three variables show a change as new vegetation parameters are used. Both air temperature and precipitation are higher as soon as the change occurs; but as LAI cannot remain at the higher value, they all return to the same values and no change at the equilibrium state.
Table 5.6: List of factors with different values in the two runs in the Soil-Vegetation-Atmosphere Box Model for the MDB region.

<table>
<thead>
<tr>
<th>Factor</th>
<th>No changes</th>
<th>With changes</th>
</tr>
</thead>
<tbody>
<tr>
<td>LAI</td>
<td>-</td>
<td>+1.5</td>
</tr>
<tr>
<td>$h_{\text{veg}}$</td>
<td>0.1</td>
<td>5</td>
</tr>
<tr>
<td>$g_{\text{max}}$</td>
<td>0.012</td>
<td>0.03</td>
</tr>
<tr>
<td>$f_c$</td>
<td>0.08</td>
<td>0.15</td>
</tr>
</tbody>
</table>

more noticeable effect on the heat balance rather than the water balance. As shown in Figure 5.5 and 5.6, the resistance effect in the moisture variables is quite strong. The changes in the vegetation parameters tend to balance each other in ET. Under the dry condition, although LAI is forced to increase, the vegetation eventually demises since no water is available. This situation is similar to the afforestation in the arid/semi-arid regions of China (Cao, 2008a; Cao et al., 2010). My results confirm that Australia is and will probably remain as a dry continent with sparse vegetation given the current climate conditions.

5.4.2 Variation

According to the Sobol’ analysis (see Table 5.3), $F_q$ is the most important influence on the precipitation. In the previous MDB case study, a consistent small flux value has been used to represent the current average dry condition. Alternatively, variations of $F_q$ are also applied to test the vegetation effect in the MDB under different large-scale circulation patterns.

**Variation case 1: higher $F_q$**

While all the other parameters are unchanged, the $F_q$ value increases by 50%, 100% and 200%. When $F_q$ increases by 50% to 0.6 mm day$^{-1}$, the results from
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the two scenarios, with vegetation change and no vegetation change, still approach one equilibrium. Compared to the previous drier case ($F_q < 0.3 \text{ mm day}^{-1}$), air temperature drops by more than 10 K and rainfall increases almost 500 times (Figure 5.8(a)). It again responses to the Sobol’ analysis that $F_q$ has a significant elevation effect on precipitation. Different equilibrium states appear when $F_q$ increases to 0.9 mm day$^{-1}$ and 1.3 mm day$^{-1}$. At 0.9 mm day$^{-1}$, there are two equilibrium states in the temperature fields and in LAI (Figure 5.8(b)). The difference between the air temperatures of the two scenarios is about 2 K. While the change in vegetation has a cooling effect immediately after the change and in the steady state, it cannot sustain itself hence the final LAI cover is lower than the LAI in the case that the vegetation is left to change naturally. It is explained by a decrease in soil moisture as LAI suddenly increases, partially due to a higher ET and partially due to more rain being intercepted by the vegetation cover. In the last case as $F_q$ increases by 200%, there are also distinct differences between the moisture fields as shown in Figure 5.8(c). With the changes in vegetation, rainfall is $0.1 \text{ mm hr}^{-1}$ higher than the case when there is no vegetation change. With such a high lateral flux, soil moisture is over the field capacity after the warm-up stage and it is able to support the water requirement of the vegetation. When $F_q$ is high, the equilibrium states after the vegetation changes do not divert much from the starting state, compared to the other cases. On the other hand, there are relatively larger difference between the starting states and the final states when the vegetation is not forced to change.

According to the model results, afforestation effect can take place when the area has access to an abundant moisture supply. The cooling effect of vegetation is more easily picked up than its effect on precipitation. Lower temperature in the afforestation
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Figure 5.8: The evolution of potential air temperature $\theta$, LAI and precipitation $P$ when $F_q$ value increases by 50%, 100% and 200%. The solid lines are simulation outputs with no vegetation changes. The dashed lines are simulation outputs with vegetation changes. (a): When $F_q$ increase by 50%, results approach one equilibrium. (b): When $F_q$ increases by 100%, $\theta$ and LAI reach different equilibrium but there is still one equilibrium in $P$. (c): When $F_q$ increases by 200%, there are two equilibrium in all outputs.
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region is also found in other studies, including studies on Australian regions (e.g. Kanamitsu and Mo, 2003; Narisma and Pitman, 2006; Wichansky et al., 2008). In this model, ET is higher with larger vegetation cover and it can produce a cooling effect (De Ridder and Gallee, 1998; Zhao and Pitman, 2002). Hence local convection which requires adequate heat and moisture inputs cannot be promoted. There is less change in precipitation until the supply of moisture can overcome the pressure on soil moisture. In the latter case, air moisture continues to rise or it is able to be maintained at a high level. So the high precipitation can be sustained as shown in Figure 5.8(c). This might explain that a positive feedback between vegetation cover and precipitation can be usually found in the tropics where large amounts of maritime moisture have been carried into land due to intensive heating (e.g. Fisch et al., 2004; Correia et al., 2008; Wang et al., 2009; Butt et al., 2011; Molina et al., 2012). The importance of the lateral flux in the vegetation cover-precipitation relationship is also supported by some other global or regional modelling studies (e.g. Kanamitsu and Mo, 2003; Pitman et al., 2004; van der Molen et al., 2006; Correia et al., 2008; Lee and Berbery, 2012).

I also test the deforestation case which the vegetation parameters are reduced by half after the warm-up. As expected, for a relatively small $F_q$ the steady state of the system does not change. Again when $F_q$ increases to 0.054 mm hr$^{-1}$ the precipitation equilibrium is different from the no change scenario. However as the current vegetation cover is already quite low, with an initial LAI = 0.5, further reduction of the vegetation cover can only have a small impact on the precipitation.

Variation case 2: dynamic $F_q$

Here $F_q$ has an annual change from low to high. In the first case a simple step change was assumed. $F_q$ switched between 0.018 mm hr$^{-1}$ and 0.054 mm hr$^{-1}$ for
every 4,320 steps (equivalent to 180 days). In the second case, the change of $F_q$ was assumed to follow a sine function with minimum value of 0.018 mm hr$^{-1}$ and maximum value of 0.054 mm hr$^{-1}$. Again the function has a half period of 4,320 steps. In both cases, the results approach to a steady cyclic pattern over the long run (see Figure 5.9). In the step change scenario, temperature and precipitation have sharp turns at the high values as the lateral flux switches from the high to the low. When $F_q$ changes smoothly, these variables also have smooth changes. The other effect of the smooth changes of $F_q$ is that the system reaches the steady states earlier than the case that $F_q$ changes abruptly. Although the amplitude of the $F_q$ oscillations are the same in these two cases, the outputs have a larger amplitude when $F_q$ has step changes. This is expected as $F_q$ stays at the two ends for longer time, under the step change scenario.

The effect of afforestation is the largest soon after the changes occur. With a much larger oscillation at the beginning, air temperatures in the afforestation scenario is eventually lower than in the no change scenario. There is a larger difference in the minimum temperature than in the maximum temperature. While the minimum temperature is corresponding to the high $F_q$, it again confirms that the vegetation effect can only be shown when the lateral flux is high. In the previous case when $F_q$ is constant at 0.054 mm hr$^{-1}$, there is a different precipitation equilibrium between the afforestation and no change scenarios. However here the vegetation change effect only affects precipitation up to a certain period of time, but has no impact in the long run despite a clear difference in the LAI between the two scenarios (see Figure 5.9). It appears that the precipitation is more sensitive to a low lateral flux than a high lateral flux. This might explain the persistence of drought. While a small decrease in moisture flux rapidly results in a drier environment, it requires a significant increase
Figure 5.9: The evolution of potential air temperature $\theta$, LAI and precipitation $P$ when $F_q$ has a step change and a sine change annually. The solid lines are simulation outputs with no vegetation changes. The dashed lines are simulation outputs with vegetation changes. (a): When $F_q$ has a step change between 0.018 mm hr$^{-1}$ and 0.054 mm hr$^{-1}$ for every 4,320 steps (equivalent to 180 days). (b): When $F_q$ as a step change between 0.018 mm hr$^{-1}$ and 0.054 mm hr$^{-1}$ with a half period of 4,320 steps.
in moisture flux over a period of time for the environment to be wet again. In order to increase precipitation equilibrium, a much stronger large-scale circulation impact would be needed to combine with the local vegetation changes.

5.5 Discussion

Compared to the models in D’Andrea et al. (2006) and Baudena et al. (2008), with the extra parameters the present model inevitably becomes more complex. The present model requires approximately less than double computation times of the previous models. On the other hand, the present model still remains the simple structure of a 1-D model with three layers vertically. The model is not highly sensitive to the new parameters, hence there is no strict requirement for the calibrations. The adding complexity provides flexibility to change the land surface configuration. It serves our purpose to study the effect of vegetation cover. Nevertheless, it is still a relatively simple model. The simple structure allows the model to be easily understood. Relationships between factors are traceable and manageable.

The model is a static model. It provides information for the long term behaviour of the land-atmosphere system. The model shows that the environment has a strong resilience. Unless the afforestation is associated with a high lateral moisture flux, a change on the surface can lead to short term instability but the system is able to eliminate the impact of the change over time. There is an “ecohydrological equilibrium” (Hatton et al., 1997) built into the model. The level of optimum vegetation that can be sustained is based on the available moisture. A shift in the vegetation without introducing an adequate shift in the moisture convergence could soon exhaust the water storage in the environment.

The model is set up for a regional scale interaction (D’Andrea et al., 2006), where
fine spatial heterogeneity is ignored. As a 1-D model, only the vertical direction is considered. It has a general application for large area but the horizontal direction is assumed to be homogeneous over the MDB. For example, the vegetation cover is spatially homogeneous in the model. Localised convective activities and boundary layer instability cannot be produced in the model. The estimated basin wide updraft motion tends to be weaker than expected, partially due to the lacking of some other elements that support the forming of strong turbulence, such as heterogeneities in land cover (Avissar and Liu, 1996; Zeng et al., 2002; Wang et al., 2009; Taylor et al., 2010; Garcia-Carreras and Parker, 2011). A detailed convective parameterisation is needed to use this model to local scale studies.

Using the Sobol’ method, this study has identified a few important factors to which the model is highly sensitive, under assumptions of no other time series changes at the boundary. GCM studies conducted in eastern Australia (Deo et al., 2009a,b) found that loss of natural vegetation has pronounced effects on local climate during strong El Niño events. This results in this study that drying episodes have a stronger effect on precipitation confirm their study. Nair et al. (2011) showed that precipitation changes after clearing in southwest Western Australia occur in areas having the same moisture convergence changes. Their study supports moisture convergence being the most important factor in precipitation changes, which is consistent with results from Sobol’ analysis in this study. The Sobol’ sensitivity analysis also indicates that the influence of soil moisture on precipitation is higher than vegetation parameters, such as LAI. The importance of soil moisture on climate has been highlighted in a large amount of studies (e.g. Koster et al. 2004, Fischer et al. 2007, D’Odorico et al. 2007, Seneviratne et al. 2010, Hauck et al. 2011, Taylor et al. 2012). The presence
of vegetation is essential for the soil moisture impact but can also complicate soil moisture-precipitation relations and influence soil moisture (McAlpine et al., 2007; Seneviratne et al., 2010; Hauck et al., 2011).

The effect of afforestation on precipitation can be strengthened in an environment with sufficient lateral moisture convergence. The amount of external moisture required depends on the relation between ET and precipitation. In the model, the precipitation efficiency, which is the fraction of convective moisture that is precipitated, increases sigmoidally with the convective moisture (D’Andrea et al., 2006). Under this assumption, increasing ET can lead to higher precipitation efficiencies. However if the convective moisture is at the high range in the precipitation function, in which the precipitation approaching to a steady state, increasing ET would cause more moisture divergence. When $F_q$ is 0.018, 0.027 and 0.036 mm hr$^{-1}$, the loss of moisture after afforestation is larger than the supply of external moisture. Therefore the water storage in the system is reduced. In the case that $F_q$ increases to 0.054 mm hr$^{-1}$, moisture divergence can be compensated by moisture convergence, hence there is a net gain of water on the land surface.

In this study the effect of afforestation on rainfall is found to be small and might be attributed to the following reasons:

- The model does not have a deep groundwater component. D’Andrea et al. (2006) and Baudena et al. (2008) have only tested the importance of soil moisture. In my experiment the system is dry, if the soil moisture is exhausted without sufficient recharge from rainfall over a period of time. However trees can have access to deep groundwater so they are less sensitive to soil moisture availability (Teuling et al., 2010; Nair et al., 2011). Furthermore, soil moisture
can also be recharged from groundwater and enhance precipitation \cite{Maxwell2008, Lo2011}. In such situations, afforestation can be expected to have a higher impact than that shown in the model in this study.

- In the model, the relationship between the vegetation cover and the lateral moisture convergence is not built. Potentially the changes on the landscape can influence the lateral moisture convergence. \cite{Makarieva2007} have argued that there is a biotic moisture pump over the forest, which attracts moist air from the ocean and further drives the hydrological cycle on land. The maritime moisture transportation is not affected in a forested landscape but decreases significantly over non-forested areas \cite{Makarieva2006, Makarieva2009}.

Some other modelling studies also found that land cover change could locally affect the magnitude of large-scale circulation \cite{Correia2008, Nobre2009, Li2010, Dallmeyer2011, Schneck2011}. \cite{Nair2011} suggested LCC might have changed the surface convergence and winter time west coast trough, and in turn rainfall decreased to the west of the Western Australian rabbit fence. Hence the ability to model the influence of forest on moisture convergence is the next important topic to study in vegetation-precipitation feedback research.

5.6 Conclusion

In this study the simple soil-vegetation-atmosphere model in \cite{DAndrea2006} and \cite{Baudena2008} has been further developed to represent more features of vegetation cover. These include LAI, vegetation height, stomatal conductance, canopy interception and separation of ET. The additional vegetation settings enable the model’s capability to test for surface cover changes in more details. However, the
model remains as a relatively simple physical deterministic model. System equilibrium are the expected output of the model under the conditions of no change in the boundary values.

By including interception and considering the impact of atmospheric conditions on transpiration, the estimated wet state ET is higher than in the previous models of [D’Andrea et al. (2006)] and [Baudena et al. (2008)], but still within a reasonable range. The new components increase the complexity of the feedback relationships in the model and decrease the impact of initial vegetation state compared to the results in BA. The Sobol’ sensitivity analysis indicates the system equilibrium is almost insensitive to the vegetation changes, but highly sensitive to the boundary conditions of lateral flux and radiation input. In terms of initial soil moisture, its effect is still strong in the outputs of the new model and this is consistent with the previous models. The Sobol’ sensitivity analysis shows that the initial soil moisture is the most important local factor, followed by initial LAI. This suggests that the feedback between the soil and the precipitation is more significant than the feedback between the vegetation and the precipitation.

In the current study, planting more trees cannot impact rainfall in a dry environment over the long run. Afforestation increases transpiration and evaporation from leaf surface. Higher latent heat flux can contribute to larger rainfall over a short period of time. However this will not be sustained as soil moisture is depleted quicker than groundwater recharge. Over time the impact of afforestation is eliminated due to the negative feedback from the soil moisture. The environment has strong ability to maintain an ecohydrological equilibrium with a complicated feedback system. My model shows that no fundamental change can occur unless there is a external
force. The results indicate that there could be some thresholds in the $F_q$ value which determine the outcome of afforestation. When $F_q$ increases by 50% and 100%, precipitation equilibrium does not change between the no change scenario and the afforestation scenarios. When $F_q$ increases by 200%, afforestation can lead to a higher precipitation steady state. Only in the case that the supply of external moisture overcome the loss of internal water storage caused by moisture divergence, afforestation can increase the long term precipitation.

The study shows that a drying episode has higher impact on the regional climate than a wetting episode. When the high moisture convergence is long lasting, afforestation can further increase rainfall. When the environment experiences alternative wet and dry episodes with equal length, afforestation cannot shift the precipitation to a different steady state. Hence small scale land management effort to increase precipitation by planting more trees during dry years or with frequent dry years might not be effective.
Chapter 6

General discussion, conclusions and future research

6.1 General discussion and conclusion

Australia is a dry continent. Rainfall is highly variable and tends to be very low in large part of the country (Holper, 2011). Hence rainfall changes have important policy implications. Annual rainfall declines significantly in the southwest, southeast, eastern coastal and inland regions (Nicholls, 2006; Gallant et al., 2007; Timbal et al., 2007; Timbal, 2009). Although these rainfall changes are mainly attributed to large-scale drivers, such as ENSO and IOD (Ummenhofer et al., 2008; Chowdhury and Beecham, 2010; Smith and Timbal, 2012), land use/land cover change is suggested as an additional contributor (e.g. Narisma and Pitman, 2003; Pitman et al., 2004; Timbal and Arblaster, 2006; McAlpine et al., 2007). However these conclusions are mainly based on complex numerical model results. There are very few observational studies on the land-atmosphere interaction, given the difficulties in attributing a cause to any change in a complex feedback system (Nicholls, 2006). There are, however, some advantages of estimating the impact of vegetation changes using models that are relatively simple, which allow for feedback analysis and uncertainties assessment (Shackley et al., 1998).
Chapter 6. General discussion, conclusions and future research

The studies presented in this thesis aim at answering the main question “can vegetation cover changes affect local rainfall?”. None of the three studies could confirm a strong impact of vegetation cover change on rainfall. The impact of LCC appears to be conditional and has a weak signal, as described by Nicholls (2006) that it is, at best, a “secondary cause” of the rainfall changes. From this point of view, what has been found in the current studies is consistent with the existing literature.

6.1.1 Answers to thesis questions

The results from the studies are used to answer the thesis questions raised in Chapter 2.

Q1 Is the feedback relationship related to the type of land cover change?

Overall, the effect of the type of LCC is not obvious, as no strong evidence on vegetation-rainfall feedback is found in these studies. On the other hand, this thesis has found that different types of land surface and their changes have some unique characteristics.

Various types of land surface change have been studied. Significant tree cover loss has occurred in the two study regions in Chapter 3 but the causes are different. In the QLD region, widespread land clearing for agriculture purpose was responsible for the tree cover loss in 2003 - 2004. In the NSW/VIC region, the tree cover loss in 2003 in the Snowy Mountains and the west border of ACT was due to severe wild fires. In Chapter 4 the current land cover in Kyeamba is short grasses, while in Tumbarumba is open forest. In the sensitivity experiments, the land covers in these two sites are perturbed: the grassland is replaced by forest and vice versa.

Significant vegetation-precipitation relationship is not found in these vegetation cover change cases. In Chapter 3, the feedback is expected to show as a step change in
the rainfall data as a result of the land surface disturbance. In the QLD study region, significant step change ($\alpha=0.05$) is not detected in the rainfall data corresponding to the LCC. In the NSW/VIC region the statistical analyses on the rainfall data do not reach an agreement on the significance of the step change. In Chapter 4 the feedback is defined as the changes in the boundary layer development and more important, the chance of reaching a LCL within the boundary layer, which indicates the potential for convective precipitation (Snyder, 2010; Konings et al., 2010; Nair et al., 2011). No LCL has been simulated in Kyeamba in either the control experiment or the sensitivity experiment. On the other hand, the LCL was simulated in 2 out of 3 days in both scenarios in Tumbarumba. Furthermore, the change of the boundary layer development was not consistent within a day and across the three days.

However, there are some differences between the types of LCC. Although the rainfall in the QLD region shows no significant step change at all in the semi-parametric regression model, in the NSW/VIC region significant step changes ($\alpha=0.05$) have occurred in a large area covering the bushfires locations. Part of the bushfire areas also shows up in the step trend test result with a step change significant at 10% level. As the bushfires were also the result of prolonged drought (Taylor and Webb, 2005), it is possible that the spatial matching reflects the feedback of bushfires to the dry conditions.

Comparing Kyeamba and Tumbarumba, the LCL development shows a preference over the forest sites. In the control experiments, the LCL has been simulated in the forest site but not the grassland site. Given the two sites are about 70 km apart, they are expected to be controlled by the same synoptic conditions. Hence, the difference is likely due to the local conditions, such as the soil moisture, which are partially
influenced by the vegetation cover (e.g. [Teuling and Troch, 2005; van Dijk et al., 2007; McAlpine et al., 2007; Mohammad and Adam, 2010; Zhang et al., 2011]).

Q2 Does the feedback relationship have different behaviours between short term and long term?

This thesis has found that there are two differences between short term and long term equilibrium effects in the feedback relationship: (1) over a longer term, vegetation responds to the changes of climate, especially rainfall; (2) in the equilibrium situation the overall moisture in the system becomes a factor.

When the vegetation cover change effect is investigated using the CLASS model, which is a single day model, there are some short term fluctuations in the boundary layer conditions. For example, afforestation decreases albedo, hence the amount of net radiation increases (Silva et al., 2006; Betts, 2007; Junkermann et al., 2009; Zeng and Yoon, 2009). As a result, the heat fluxes can increase, hence the boundary layer could be higher and the LCL could be lower over the forest cover (Fisch et al., 2004; Konings et al., 2010).

However, changes of the land surface properties have different and sometimes opposite relationships with the boundary layer conditions. Their effects compete with each other as the boundary layer grows. For example, afforestation is also associated with large increase of surface roughness which in turn reduces latent heat flux. In this case, the chance of LCL to occur within the boundary layer could depend on the magnitude of changes in each land surface property.

Due to the competition between the different land surface properties, their combined effect tends to be smaller than individual effects. Commonly only the effects of one or two properties are highlighted (e.g. De Ridder and Gallee, 1998).
Chapter 6. General discussion, conclusions and future research

As pointed out by Zhao and Pitman (2002), this is determined by the characteristics of vegetations before and after changes. The variation in land surface properties used to represent the land cover change could be one of the main reasons contributing to the variation in the results.

The land-atmosphere system tends to approach equilibrium over the long term. Different from the short term relationship where the vegetation does not feedback to the changes in the environment, vegetation dynamics after LCC are considered over a long term. In Chapter 5, the vegetation growth/decline is determined by the soil moisture availability and the relation between the current vegetation and the carrying capacity.

In the hypothetical case that the dry MDB is fully reforested, rainfall could increase due to higher ET. However moisture divergence also occurs, especially when rainfall efficiency is low. As the simple equilibrium model does not have a groundwater component, soil water could only be recharged by rainfall. When ET is continuously higher than rainfall, soil moisture is reduced and in turn the vegetation cover is also under pressure to decline. The model shows that there is not enough moisture in the system to maintain a high vegetation cover, if groundwater is not a source of water. The initial soil moisture and the lateral moisture convergence basically determine the equilibrium state of the system (D’Andrea et al., 2006; Cao, 2008a).

Q3 What mechanism can explain the existence/non-existence of the vegetation-precipitation feedback?

This thesis has found that there are in fact several mechanisms and factors that explain the above findings. The main driving forces of the land-atmosphere interaction,
especially the rainfall feedback, are the moisture and the radiation energy, as shown by the Sobol’ analysis in Chapter 5. More vegetation can increase net radiation due to the albedo effect (e.g. Gibbard et al. 2005, Correia et al. 2008, Lawrence et al. 2012). Vegetation cover change alone does not generate additional moisture. The source of moisture and how the moisture can be used are important questions in the vegetation-rainfall feedback.

The external moisture supply, such as the maritime moisture inflow, is important to fuel the local land-atmosphere interaction. If the initial soil moisture is dry and reforestation is used to change the local rainfall regime, it might lead to management failure as the vegetation cannot be sustained without sufficient water (Cao 2008b, Cao et al. 2010). With an adequate supply of external moisture, afforestation can be effective and result in a fundamental change in precipitation. This might explain why a strong positive rainfall feedback is often seen in the tropics (e.g. Correia et al. 2008, Wang et al. 2009, Butt et al. 2011, Molina et al. 2012). The importance of maritime conditions is also supported by van der Molen et al. (2006).

One of the important properties of the forest is its rough surface. The high roughness length can reduce horizontal wind speed, produce effective surface turbulence and possibly increase the lateral moisture convergence (Pitman et al. 2004, Pitman and Hesse 2007, McAlpine et al. 2007, Teuling et al. 2010, Lee and Berbery 2012). The change in the moisture convergence is the main reason for the change in precipitation (Pitman et al. 2004, Correia et al. 2008, Schneek and Mosbrugger 2011, Lee and Berbery 2012). In the afforestation scenario in Kyeamba, when a positive advective moisture term is included, the chance of a CBL that is higher than the LCL has increased. However, when there is a long dry episode due to the impact
of El Niño, the sensitivity of moisture convergence to land surface change could be small. This might also explain the weak signal of rainfall feedback to vegetation cover change in the study in Chapter 3.

Soil moisture is another important source for the vegetation-precipitation interaction. It itself has a strong feedback on the local climate (e.g. Koster et al., 2003; D’Odorico and Porporato, 2004; Fischer et al., 2007; Seneviratne et al., 2010). Plant status and transpiration are maintained by soil moisture availability (Levins, 1969; D’Odorico et al., 2007; Baudena et al., 2008; Cavanaugh et al., 2011). Furthermore, ET as a function of soil moisture is dependent on the types of land surface (Entekhabi et al., 1992), but the various functions are different only for a certain range of soil moisture. Hence the vegetation-precipitation feedback is conditional, given a soil moisture level that is in favour of this feedback relationship.

6.1.2 Uncertainty

The studies presented in this thesis are limited by a number of factors.

- In the empirical study, although effort has been taken to remove rainfall variability attributed to the concurrent ENSO, the effect of the previous drought could still remain in the time series and intensify the next El Niño effect. Such influences could increase the difficulty in detecting a step trend due to the vegetation cover change effect.

- The cross-correlation in the gridded data might not have been fully accounted for by taking a block size $n_s = 4$, i.e. the number of pixels in a block is four. However the results does not change much with $n_s = 9$, while the computation cost is much higher. It is uncertain that whether it is necessary to keep a square
block or if the cross-correlation in one direction is higher than in the other direction.

- In the CLASS model, vegetation uses water from the deeper soil. It does not distinguish between deep rooted trees and short rooted grass. Although the settings of vegetation fraction can compensate this shortage on ET difference, it cannot properly represent the effect of dynamic soil moisture and ecological niche separation.

- In the equilibrium land-atmosphere model, the dynamic vegetation feature is only represented by LAI. The other surface properties, such as vegetation height and maximum stomatal conductance, are constant with the vegetation type. This might be the reason that their effects are insignificant. It is uncertain whether these simple assumptions in the model have limited the effect of vegetation or they can be left out to maintain the simplicity.

- Although our findings of the important role of external moisture in the vegetation-precipitation feedback are supported by the literature, some other studies suggest precipitation feedback is suppressed during the monsoon season (e.g. Otterman et al., 1990; Bengai et al., 1994; Kanae et al., 2001). This conflict needs to be further addressed. It is unclear how large scale forcings could influence the local feedback.

6.2 Future work

Despite the progress made in this thesis, many aspects of the vegetation-precipitation feedback are still unclear. The following future work is suggested to expand the scope of this thesis.
• As mentioned before, cross-correlation in the gridded data is a problem in the empirical analysis. Besides using other methods to handle the cross-correlation as suggested in Chapter 3, the analyses can also be applied on high quality station data to provide a comparison. The criteria for the rainfall data from the station are: (1) at least 4 stations available at the local scale, for example, of the Snowy Mountain region; (2) more than 10 years time series available, with more than 5 years data after the LCC; (3) no missing data.

• Comparison experiments can be conducted on regions with known LCC and regions without LCC. The selection of the study sites should avoid being parallel to the major wind direction, to minimise the remote effect of LCC. However in order to reduce the noise due to other differences, the sites should be reasonably close to each other and under the same climatic controls. This would require careful investigation of the vegetation cover change data and the atmospheric data.

• Soil moisture models that include a deeper soil layer with water only available to deep rooted plants can be used in place of the original soil model in CLASS to study different vegetation effects. The model can also be applied to southwest Western Australia where distinctly different land surface characteristics are presented in the adjacent areas. The availability of cloud information and the large number of land-atmosphere studies in this region can provide validation data for the model results.

• The relationship between external forcings and local land-atmosphere interaction requires thorough understanding. Although the climate impact of large scale drivers or local land atmosphere interaction has been widely studied,
there are few studies on their possible combined impact. Two opinions, which
might not be completely opposite to each other, exist: one, strong synoptic
conditions suppress local land-atmosphere interaction; two, the vegetation
impact on climate requires supports from the external forcing. How to address
this question is still a challenge, but it is definitely worthy of more research
efforts.
Appendix A

Systematic Review Protocol: Does the existence of forest affect local rainfall?

Background

Forest cover has decreased along the development of human history. Human activities, such as cropping and grazing following deforestation, are regarded as the main cause for forest loss (Darbyshire et al. 2003; Sharma and Sharma 2005; Elmqvist et al. 2007; Elliott et al. 2010). Massive reductions of forest are found around the world, especially in the tropical regions (see Figure A.1). For example in Brazilian Amazon, 0.6M km$^2$ was deforested during the 50 years before 2001 (Chagnon et al. 2004). In Central America, Mexico alone lost 70% vegetation cover between 1989-1999 (Alcantara-Ayala et al. 2006). Zak et al. (2004) also found that 85% of total forest and woodlands in Gran Chaco, Argentina (South America) had been cleared out within just 30 years after 1969.

Meanwhile, there is increasing consensus on the existence of climate change caused by anthropogenic forcing in the last 30 years. Although remaining uncertainties were addressed, the 2007 Intergovernmental Panel on Climate Change (IPCC) fourth assessment report stated that substantial anthropogenic contribution has likely
Figure A.1: Annual change of forest cover for the periods of 1990-2000 (blue) and 2000-2010 (red) by regions. The graph was reproduced using figures in *FAO*(2010).
increased land surface temperature since mid 20th century except Antarctica. The opinion of global warming is shared by many other recent scientific assessment (Weber and Stern, 2011). There are two streams of opinion towards the type of human activities as the cause of climate change. One believes that burning of fossil fuel is the most important single source of increasing greenhouse gas (van der Werf et al., 2009; Weber and Stern, 2011; Reynolds et al., 2010). The other set of opinion is that global climate change is largely attributed to land use change (e.g. Schmidt, 2010). The latter has attracted rising attention. Human intervention on land cover might have changed the impact of large-scale factors, such as El Niño-Southern Oscillation (ENSO), on the local climate (Zhang et al., 2009c) or isolated local effect like the “urban heat island” (Dessler and Parson, 2010; Schmidt, 2010). On one hand, the connection of large-scale climate drivers and continental or regional climate are known (Bruijnzeel, 2004; Bosilovich and Chern, 2006; Hoerling et al., 2006; Findell et al., 2009; Kamruzzaman et al., 2011); on the other hand great uncertainties still exist at the local scale (Zhang et al., 2009c; Hurrell, 2009; Findell et al., 2009). This could be a proof of the partial influence of the local land use change impacts. When dealing with climate changes, it could be more effective to tackle the local land use problems first then gradually approach a global solution.

Land surface condition is considered as a control factor of the variations in local climate. Given the current level of climate change and possibly continuous land use and land cover change (LULCC), how will local climate behave in the future? Can reforestation/afforestation provide climate control benefits, such as increasing the local rainfall and water balance in dry regions? Currently the full hydrological function of forest is still unclear since there was unbalanced work on various aspects of the forestry
hydrologic cycle (e.g. runoff and stream flow versus rainfall). For example, reviews on the hydrologic impact of forest exclude rainfall (e.g. Andréassian, 2004; Cosandey et al., 2005), or reflect a smaller number of studies on rainfall than other hydrologic impacts (e.g. van Dijk and Keenan, 2007). It is necessary to increase effort in the research of meteorologic part of water cycle under land use and land cover change.

**Objective of the review**

*Primary question*

Does the permanent change of forest cover affect local rainfall?

Here we are interested at the permanent change where natural restoration is prevented. Hence areas with tree loss due to bushfire and regrowth afterward are outside the scope of this study. The primary question can be expanded into several specific questions. This step is necessary in exploring the possible causalities and clarify the matter of interest (Nichols et al., 2011).

- Does deforestation decrease (or increase) local rainfall?
- Does deforestation suppress (or enhance) the development of convective clouds?
- Does deforestation change the seasonal rainfall distribution?
- Does deforestation have higher impact on a particular season or month than the other time of a year?
- Does reforestation have the opposite effects to the above?
- Do more (or less) rains fall on forested areas than pasture/grasslands/crops annually?
- Do more (or less) rains fall on forested areas than urban areas?
• Do more (or less) rains fall on the boundary between forest and other land cover type, e.g. crops?

Methods

Search strategy

This systematic review aims to search and analyse observational evidence of the impact of forest cover change on local rainfall. The purpose of conducting a systematic review is to ensure the review is based on broad and unbiased evidence. Hence the search is carried out within multiple electronic databases and internet sources for journal articles, proceedings, theses and/or government reports. The relevant literature will be limited to the last two decades for most current research. When there are more than 600 hits from one single source, only the first 500 sorted by relevance will be imported into reference manage software for review. Here Endnote is used.

All articles retrieved from the search will be stored in an Endnote library for further processing. Firstly a quick examination will go through the titles to filter out articles that are obviously irrelevant. The abstract is the second section to be assessed and important information will be extracted for analysis. In the case that detailed information are needed from the full article, the section regarding studied sites and the conclusion will be read in order to identify the nature of land use and land cover changes and the impact on rainfall.

Search terms

Subject terms:

• Forest
• Woodland
• Trees
• Plant* forest*
• Production forest*
• Land cover
• Land use
• Precipitation frequency
• Precipitation amount
• Rain* frequency
• Rain* amount
• Storm
• Convective cloud
• Cloud formation
• Cloudness
• Climate

Intervention terms:

• Afforestation
• Deforestation
• Reforestation
• Forest regeneration
• Forest restoration
• Forest fragment*
• Land clearing

1 (*) denotes a wildcard for possible alternative word endings.
Appendix A. Protocol

• Land cover change
• Land use change
• Land surface change
• LULCC (land use and land cover change)
• Compare (or equivalent terms)

Outcome terms:

• Cloud* change (increas*, decreas*, higher, lower, enhance*, suppress*, frequent, etc)
• Rainfall/precipitation change (increas*, decreas*, higher, lower, enhance*, suppress*, frequent, more, etc)

Source

The search will be performed in the following online sources.

Database:

• ISI Web of Science
• ProQuest Research Library
• ScienceDirect

Study inclusion criteria

We look for evidence of forest cover change impact on local rainfall or at least cloud formation. All studies returned from the search will be further assessed at the title and the abstract levels for relevance according to the analysis of primary question (Table A.1).

The following scopes will also be applied on search.
Appendix A. Protocol

- **Language scope**: English language only.

- **Type of study**: observational studies, experimental studies, statistical analysis which are NOT modelling or simulations or reviews.

- **Availability of article**: At least title and abstract are electronically available.

A trial has conducted on about 200 articles and 26 of them were found more or less related to the review question. Then an analysis was performed on the title and the abstract to look for common pattern and/or other common terms. The following characteristics are found:

- Most of the relevant articles have one of the subject terms, especially rainfall or precipitation, in the title;
- A word indicating the relationship, such as impact, effect or following, generally appears in the title;
- In most cases, the terms “rain” and “forest” or their alternatives occur very often in the abstract;
- In many cases the study is on rainfall or climate changes analysis without much details on the land cover changes;
- In the conclusion remark, the words “possible” or “likely” using on the impact of land use changes indicate the seeking relationship is not a part of the analysis outcome.

In order to speed up the selection process, the following filters are applied before read into the content of the studies.

- One of the subject terms must appear in the title;
• If the title reveals the causality, rainfall/precipitation must appear at the effect part. For example in “An increase of early rains in southern Israel following land-use change” ([Otterman et al., 1990], the word “rains” appears in front of “following”.

• If “rainforest” (or “rain forest”), “cloud forest” or “seed rain” appears in the title, the article will be excluded unless “rain” (or equivalent) and “cloud” also appears anywhere in the title or abstract without the prefix “seed” or postfix “forest”.

Table A.1: Research question components analysis

<table>
<thead>
<tr>
<th>PICO/PECO</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Relevant subjects</strong></td>
<td>The term “forest” is used for both areas with dense tree covers (e.g. tropical rainforest) and low density forest (e.g. Mediterranean woodland). Plantation forest is also included. The use of this term here does not violate the definition given by the National Forest Inventory (NFI).</td>
</tr>
<tr>
<td></td>
<td>The review is interested at forest cover changes or comparison of different land covers in any geographic location. The size of land cover changes should be relatively large, i.e. 2-200 km$^2$. The effect due to forest composition, e.g. the types of tree species, will not be reviewed.</td>
</tr>
</tbody>
</table>
The review is interested at corresponding changes in local rainfall. The changes in rainfall equilibrium are sought. The level of changes can vary between seasons.

In the case that cloud formation is studied rather than rainfall, the article is included if the causal relation between cloud formation and the land surface, especially the forest cover, is supported by evidence.

<table>
<thead>
<tr>
<th>Type of intervention</th>
<th>Afforestation, reforestation or deforestation. These are the main types of land use land cover changes where grasslands and/or croplands are converted to forest, or vice versa. Land surface changes by other causes is excluded, such as by wildfire or natural regeneration. Both situations are included: rainfall or cloud changes as a result of the forest cover changes; or rainfall difference between land cover types. Studies on the impact of rainfall on forest growth are excluded.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Type of outcomes</td>
<td>Local rainfall might increase, decrease or even no change. The change on rainfall amount is of primary interest. However, shifts in rainfall patterns and/or types of rainfall are important outcomes too.</td>
</tr>
</tbody>
</table>
Appendix A. Protocol

Similarly, increase/decrease/no change of cloud coverage as well as changes on cloud height and type of clouds are sought.

Exclude studies on flood, runoff and streamflows.

<table>
<thead>
<tr>
<th>Examples of comparators</th>
<th>Potential comparison studies include:</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>• pre-land clearing versus post-land clearing;</td>
</tr>
<tr>
<td></td>
<td>• Forest versus grassland;</td>
</tr>
<tr>
<td></td>
<td>• Tropical versus mid-latitude.</td>
</tr>
</tbody>
</table>

Potential reasons for heterogeneity

Studies are likely to vary due to the characteristics related to the land cover and characteristics that are not. Information on these characteristics from the articles will be extracted to allow evaluation on the relationship between the desired cause and effect. The potential characteristics are shown as a list of variables in Table A.2.

Data extraction and synthesis

The quantitative information will be sought in the relevant articles regarding the outcomes and the explanatory variables (refer to Table A.2). However, due to the

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2 The National Forest Inventory’s definition of forest is:
“An area, incorporating all living and non-living components, that is dominated by trees having usually a single stem and a mature or potentially mature stand height exceeding 2 metres and with existing or potential crown cover of overstorey strata about equal to or greater than 20 per cent. This definition includes Australia’s diverse native forests and plantations, regardless of age. It is also sufficiently broad to encompass areas of trees that are sometimes described as woodlands.” (Hnatiuk et al., 2003, P180)

3 This size corresponds to the atmospheric mesoscale in some popular scale definitions (Thunis and Bornstein, 1996).
Appendix A. Protocol

Table A.2: Explanatory variables extracted from the relevant studies. These variables potentially contribute to the heterogeneity of the studies.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Data type</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Climate</td>
<td>text</td>
<td>Dominant climate of the studied sites (e.g. tropical, sub-tropical)</td>
</tr>
<tr>
<td>Region</td>
<td>text</td>
<td>Geographic region of the studied sites (e.g. Africa, Asia, South America)</td>
</tr>
<tr>
<td>Season</td>
<td>text</td>
<td>Summer, autumn, winter or spring</td>
</tr>
<tr>
<td>LULCC</td>
<td>text</td>
<td>Type of land use &amp; land cover changes (e.g. afforestation, deforestation)</td>
</tr>
<tr>
<td>Size</td>
<td>numeric</td>
<td>Spatial scale of LULCC or studied sites in km²</td>
</tr>
<tr>
<td>Large scale effect</td>
<td>boolean</td>
<td>Whether the large scale effects have been eliminated from the data</td>
</tr>
<tr>
<td>Data</td>
<td>text</td>
<td>Data sets used in the studies, incl. the sources and the variables</td>
</tr>
<tr>
<td>Method</td>
<td>text</td>
<td>Statistical methods used in the studies</td>
</tr>
</tbody>
</table>

complex nature of the matter, numeric value of the outcomes might not be available. When the amount of change is not available, indicative information of the sign of changes (e.g. increase or decrease) will be used. Data on the explanatory variables will be extracted where possible or searched from other sources such as internet. For example, many studies might not provide information on the climate zone of the sites, while this information is usually available online.

The retrieved data are used to confirm the hypothesized relationship between forest cover and local rainfall. The sensitivity of this relationship on the other factors can also be tested using available information on the explanatory variables. The quality of the studies can serve to weight each result in the data synthesis.
Database search

Search interface in each database is different. Hence it is necessary to document the settings and selection in each database during the search process. This documentation provides a reference for similar search for any future work.

Web of Science

Search: (forest* or woodland or trees or plant* forest* or production forest* or land clearing or land change) and (precipitation or rain or storm or cloud*) in Topic AND forest or woodland or trees or precipitation or rain or cloud in Title

Timespan: 1990-2010

Citation databases:

- Science Citation Index Expanded (SCI-EXPANDED) –1899-present
- Conference Proceedings Citation Index- Science (CPCI-S) –1990-present

Chemical Databases: Uncheck all boxes

Refine:

- Exclude document types: review;
- Language: English

Retrieve 3,544 articles. Import the first 500 sorted by relevance.

ProQuest Research Library

Search: (forest* or woodland or trees or plant* forest* or production forest* or land clearing or land change) and (precipitation or rain or storm or cloud*) in Abstract AND forest or woodland or trees or precipitation or rain or cloud in Title
Date range: 1990-2010 (no day or month specified)

Source type: all

Document type: uncheck Review

Language: English

Retrieve 9,138 articles. Import the first 500 sorted by relevance.

**ScienceDirect**

Search: (forest* or woodland or trees or plant* forest* or production forest* or land clearing or land change) and (precipitation or rain or storm or cloud*) in Abstract, **Title, Keywords** AND forest or woodland or trees or precipitation or rain or cloud in **Title**

Include Journals and books and all sources and subjects.

Date range: 1990-2010

Retrieve 9 articles.
Appendix B

List of symbols

The definition of symbols used in the thesis and the unit of these parameters.

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Definition [unit]</th>
</tr>
</thead>
<tbody>
<tr>
<td>$a'$</td>
<td>Clapp-Hornberger retention curve parameter in CLASS model [-]</td>
</tr>
<tr>
<td>$\alpha$</td>
<td>Surface albedo [-]</td>
</tr>
<tr>
<td>$\alpha_0$</td>
<td>Bare soil surface albedo [-]</td>
</tr>
<tr>
<td>$\alpha_b$</td>
<td>Vegetation surface albedo [-]</td>
</tr>
<tr>
<td>$\alpha'$</td>
<td>Factor accounts for the aerodynamic component in the Priestley-Taylor equation [-]</td>
</tr>
<tr>
<td>$b'$</td>
<td>Clapp-Hornberger retention curve parameter in CLASS model [-]</td>
</tr>
<tr>
<td>$b$</td>
<td>Retention curve parameter [-]</td>
</tr>
<tr>
<td>$\beta$</td>
<td>Ratio of entrainment flux to surface flux [-]</td>
</tr>
<tr>
<td>$c'$</td>
<td>Clapp-Hornberger retention curve parameter in CLASS model [-]</td>
</tr>
<tr>
<td>$C_{1,sat}$</td>
<td>Coefficient force term moisture in CLASS model [-]</td>
</tr>
<tr>
<td>$C_{2,ref}$</td>
<td>Coefficient restore term moisture in CLASS model [-]</td>
</tr>
<tr>
<td>$CG_{sat}$</td>
<td>Saturated soil conductivity for heat [K m^{-2} J^{-1}]</td>
</tr>
<tr>
<td>$C_{veg}$</td>
<td>Vegetation fraction [-]</td>
</tr>
<tr>
<td>$\Delta$</td>
<td>Slope of the saturation vapour pressure temperature curve [kPa K^{-1}]</td>
</tr>
<tr>
<td>$e_a$</td>
<td>Actual air vapour pressure [kPa]</td>
</tr>
<tr>
<td>$E_l$</td>
<td>Evaporation from canopy interception [mm]</td>
</tr>
<tr>
<td>$ET_p$</td>
<td>Potential evapotranspiration (ET) [mm]</td>
</tr>
<tr>
<td>$e_s$</td>
<td>Saturation vapour pressure [kPa]</td>
</tr>
<tr>
<td>$E_s$</td>
<td>Soil evaporation [mm]</td>
</tr>
<tr>
<td>$E_{sp}$</td>
<td>Maximum soil evaporation [mm]</td>
</tr>
</tbody>
</table>
Appendix B. List of symbols

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>$f_c$</td>
<td>Fraction of rainfall intercepted [-]</td>
</tr>
<tr>
<td>$F_q$</td>
<td>Lateral flux convergence [mm hr$^{-1}$]</td>
</tr>
<tr>
<td>$G$</td>
<td>Ground heat flux [J hr$^{-1}$]</td>
</tr>
<tr>
<td>$g_{D}$</td>
<td>Vapor pressure deficit correction factor for surface resistance [-]</td>
</tr>
<tr>
<td>$g_{s\text{max}}$</td>
<td>Maximum stomatal conductance [m s$^{-1}$]</td>
</tr>
<tr>
<td>$\gamma$</td>
<td>Psychrometric constant [kPa K$^{-1}$]</td>
</tr>
<tr>
<td>$\gamma_q$</td>
<td>Humidity lapse rate [g kg$^{-1}$ m$^{-1}$]</td>
</tr>
<tr>
<td>$\gamma_\theta$</td>
<td>Potential temperature lapse rate [K m$^{-1}$]</td>
</tr>
<tr>
<td>$h$</td>
<td>Convective boundary layer height [m]</td>
</tr>
<tr>
<td>$h_{\text{veg}}$</td>
<td>Vegetation height [m]</td>
</tr>
<tr>
<td>$I_p$</td>
<td>Canopy interception [mm]</td>
</tr>
<tr>
<td>$k$</td>
<td>von Karman’s constant [0.41]</td>
</tr>
<tr>
<td>$K_s$</td>
<td>Saturated hydraulic conductivity [kg m$^{-2}$ hr$^{-1}$]</td>
</tr>
<tr>
<td>LAI</td>
<td>Leaf area index [-]</td>
</tr>
<tr>
<td>$L_e$</td>
<td>Specific latent heat of evaporation [2.45e6 J Kg$^{-1}$]</td>
</tr>
<tr>
<td>LE</td>
<td>Latent heat flux [W m$^{-2}$]</td>
</tr>
<tr>
<td>$L_{\text{in}}$</td>
<td>Downward longwave radiation [W m$^{-2}$]</td>
</tr>
<tr>
<td>$L_{\text{out}}$</td>
<td>Upward longwave radiation [W m$^{-2}$]</td>
</tr>
<tr>
<td>$P_{\text{hPa}}$</td>
<td>Surface pressure [hPa]</td>
</tr>
<tr>
<td>$P$</td>
<td>Precipitation [mm]</td>
</tr>
<tr>
<td>$P_{\text{eff}}$</td>
<td>Effective precipitation [mm]</td>
</tr>
<tr>
<td>$\phi$</td>
<td>Soil porosity [-]</td>
</tr>
<tr>
<td>$q$</td>
<td>Specific humidity in the convective boundary layer [g kg$^{-1}$]</td>
</tr>
<tr>
<td>$\Delta q$</td>
<td>Humidity jump at the entrainment layer [g kg$^{-1}$]</td>
</tr>
<tr>
<td>$\Delta q_\text{i}$</td>
<td>Convective moisture [g kg$^{-1}$]</td>
</tr>
<tr>
<td>$q_{\text{sat}}$</td>
<td>Saturation humidity [g kg$^{-1}$]</td>
</tr>
<tr>
<td>$\delta q_{\text{sat}}$</td>
<td>Derivative of saturation humidity with respect to temperature [g kg$^{-1}$ K$^{-1}$]</td>
</tr>
<tr>
<td>$r_a$</td>
<td>Aerodynamic resistance [s m$^{-1}$]</td>
</tr>
<tr>
<td>RH</td>
<td>Relative humidity [-]</td>
</tr>
<tr>
<td>$\rho_a$</td>
<td>Air density [kg m$^{-3}$]</td>
</tr>
<tr>
<td>$\rho_s$</td>
<td>Soil density [kg m$^{-3}$]</td>
</tr>
</tbody>
</table>
Appendix B. List of symbols

\( R \)  
Dry air gas constant \([287 \text{ J kg}^{-1}\text{K}^{-1}]\)

\( R_{sn} \)  
Net radiation which is the difference between incoming and outgoing radiation \([\text{W m}^{-2}]\)

\( r_s \)  
Canopy surface resistance \([\text{s m}^{-1}]\)

\( R_{sin} \)  
Incoming radiation \([\text{W m}^{-2}]\)

\( r_{s,min} \)  
Minimum leaf surface resistance \([\text{s m}^{-1}]\)

\( r_{s,soil,min} \)  
Minimum soil surface resistance for evaporation \([\text{s m}^{-1}]\)

\( S_0 \)  
Constant solar irradiance at top of atmosphere \([\text{W m}^{-2}]\)

\( SH \)  
Sensible heat flux \([\text{W m}^{-2}]\)

\( S_{in} \)  
Incoming solar (shortwave) radiation \([\text{W m}^{-2}]\)

\( S_{out} \)  
Outgoing solar (shortwave) radiation \([\text{W m}^{-2}]\)

\( T_a \)  
Specific air temperature in the convective boundary layer \(\[^{\circ}\text{C}\]\)

\( Tr \)  
Transpiration \([\text{mm}]\)

\( Tr_p \)  
Maximum transpiration \([\text{mm}]\)

\( T_s \)  
Surface layer specific air temperature \(\[^{\circ}\text{C}\]\)

\( T_{soil} \)  
Soil temperature \(\[^{\circ}\text{C}\]\)

\( T_{soil1} \)  
Upper soil layer temperature \(\[^{\circ}\text{C}\]\)

\( T_{soil2} \)  
Deeper soil layer temperature \(\[^{\circ}\text{C}\]\)

\( \theta \)  
Potential temperature in the convective boundary layer \(\text{[K]}\)

\( \theta_a^* \)  
Relaxation temperature for PBL \(\text{[K]}\)

\( \theta_e \)  
Equivalent potential temperature \(\text{[K]}\)

\( \Delta \theta \)  
Potential temperature jump at the entrainment layer \(\text{[K]}\)

\( \Delta \theta_s \)  
Temperature change associated with convective cooling \(\text{[K]}\)

\( \theta_s \)  
Soil wetness [-]

\( \theta_{s,*} \)  
Soil moisture for maximum plant efficiency [-]

\( \theta_{s,h} \)  
Hygroscopic point soil moisture [-]

\( \theta_{s,fc} \)  
Field capacity [-]

\( \theta_{s,w} \)  
Wilting point soil moisture [-]

\( w_1 \)  
volumetric soil water content at upper soil \([\text{m}^3 \text{ m}^{-3}]\)

\( w_2 \)  
volumetric soil water content at deeper soil \([\text{m}^3 \text{ m}^{-3}]\)

\( w_{fc} \)  
Field capacity volumetric water content \([\text{m}^3 \text{ m}^{-3}]\)

\( w_{sat} \)  
Saturation volumetric water content \([\text{m}^3 \text{ m}^{-3}]\)
Appendix B. List of symbols

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>$w_{wilt}$</td>
<td>Wilting point volumetric water content</td>
<td>$\text{m}^3 \text{m}^{-3}$</td>
</tr>
<tr>
<td>$u$</td>
<td>PBL average wind speed</td>
<td>$\text{m s}^{-1}$</td>
</tr>
<tr>
<td>$u_g$</td>
<td>Geostrophic wind in x direction in the CLASS model</td>
<td>$\text{m s}^{-1}$</td>
</tr>
<tr>
<td>$v_0$</td>
<td>Initial wind speed in y direction in the CLASS model</td>
<td>$\text{m s}^{-1}$</td>
</tr>
<tr>
<td>$z_0$</td>
<td>Height of surface layer as proportion of boundary layer height</td>
<td>[-]</td>
</tr>
<tr>
<td>$z_{0m}$</td>
<td>Roughness length for momentum</td>
<td>$\text{m}$</td>
</tr>
<tr>
<td>$z_{0h}$</td>
<td>Roughness length for scalar</td>
<td>$\text{m}$</td>
</tr>
</tbody>
</table>
# Appendix C

## List of Abbreviations

<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>ABARES</td>
<td>Australian Bureau of Agricultural and Resource Economics and Sciences</td>
</tr>
<tr>
<td>ACLUMP</td>
<td>Australian Collaborative Land Use and Management Program</td>
</tr>
<tr>
<td>ACT</td>
<td>Australian Capital Territory</td>
</tr>
<tr>
<td>AEST</td>
<td>Australian Eastern Standard Time</td>
</tr>
<tr>
<td>AGCM</td>
<td>Atmospheric General Circulation Model</td>
</tr>
<tr>
<td>ARW</td>
<td>Advanced Research Weather Research and Forecasting</td>
</tr>
<tr>
<td>ASTER</td>
<td>Advanced Spaceborne Thermal Emission and Reflection Radiometer</td>
</tr>
<tr>
<td>ASRIS</td>
<td>Australian Soil Resource Information System</td>
</tr>
<tr>
<td>AVHRR</td>
<td>Advanced Very High Resolution Radiometer</td>
</tr>
<tr>
<td>BA</td>
<td>Baudena et al. (2008)</td>
</tr>
<tr>
<td>BATS</td>
<td>Biosphere-Atmosphere Transfer Scheme</td>
</tr>
<tr>
<td>BoM</td>
<td>Australian Bureau of Meteorology</td>
</tr>
<tr>
<td>CAM</td>
<td>Community Atmosphere Model</td>
</tr>
<tr>
<td>CBL</td>
<td>Convective Boundary Layer</td>
</tr>
<tr>
<td>CCM</td>
<td>Community Climate Model</td>
</tr>
<tr>
<td>CGCM</td>
<td>Coupled Atmosphere-Ocean General Circulation Model</td>
</tr>
<tr>
<td>CLM</td>
<td>Community Land Model</td>
</tr>
<tr>
<td>CSIRO</td>
<td>Commonwealth Scientific and Industrial Research Organisation</td>
</tr>
<tr>
<td>DA</td>
<td>D'Andrea et al. (2006)</td>
</tr>
<tr>
<td>DE</td>
<td>Differential equation</td>
</tr>
<tr>
<td>DLCD</td>
<td>The National Dynamic Land Cover Dataset</td>
</tr>
<tr>
<td>ENSO</td>
<td>El Niño-Southern Oscillation</td>
</tr>
<tr>
<td>Abbreviation</td>
<td>Full Form</td>
</tr>
<tr>
<td>--------------</td>
<td>-----------</td>
</tr>
<tr>
<td>ET</td>
<td>Evapotranspiration</td>
</tr>
<tr>
<td>EVI</td>
<td>Enhanced Vegetation Index</td>
</tr>
<tr>
<td>FAO</td>
<td>Food and Agriculture Organization of the United Nations</td>
</tr>
<tr>
<td>GAM</td>
<td>generalised additive model</td>
</tr>
<tr>
<td>GCM</td>
<td>General Circulation Model</td>
</tr>
<tr>
<td>IOD</td>
<td>Indian Ocean Dipole</td>
</tr>
<tr>
<td>ITCZ</td>
<td>Inter Tropical Convergence Zone</td>
</tr>
<tr>
<td>IPCC</td>
<td>Intergovernmental Panel on Climate Change</td>
</tr>
<tr>
<td>JAMSTEC</td>
<td>Japan Agency for Marine-Earth Science and Technology</td>
</tr>
<tr>
<td>JISAO</td>
<td>Joint Institute for the Study of the Atmosphere and Ocean, University of Washington</td>
</tr>
<tr>
<td>LAI</td>
<td>Leaf area index</td>
</tr>
<tr>
<td>LCC</td>
<td>Land cover change</td>
</tr>
<tr>
<td>LCL</td>
<td>Lifting condensation level</td>
</tr>
<tr>
<td>LE</td>
<td>Latent heat flux</td>
</tr>
<tr>
<td>LSM</td>
<td>Land Surface Model</td>
</tr>
<tr>
<td>LULCC</td>
<td>Land use and land cover changes</td>
</tr>
<tr>
<td>MDB</td>
<td>Murray Darling Basin</td>
</tr>
<tr>
<td>MJO</td>
<td>Madden-Julian oscillation</td>
</tr>
<tr>
<td>MODIS</td>
<td>Moderate Resolution Imaging Spectroradiometer</td>
</tr>
<tr>
<td>NCAR</td>
<td>National Center for Atmospheric Research</td>
</tr>
<tr>
<td>NCEP</td>
<td>National Centers for Environmental Prediction</td>
</tr>
<tr>
<td>NDVI</td>
<td>Normalized Difference Vegetation Index</td>
</tr>
<tr>
<td>NFI</td>
<td>National Forest Inventor</td>
</tr>
<tr>
<td>NMC</td>
<td>National Meteorological Center</td>
</tr>
<tr>
<td>NOAA</td>
<td>National Oceanic and Atmospheric Administration</td>
</tr>
<tr>
<td>MMM</td>
<td>Mesoscale and Microscale Meteorology</td>
</tr>
<tr>
<td>PBL</td>
<td>Planetary boundary layer</td>
</tr>
<tr>
<td>PDO</td>
<td>Pacific Decadal Oscillation</td>
</tr>
<tr>
<td>RCM</td>
<td>Regional or mesoscale climate model</td>
</tr>
<tr>
<td>RH</td>
<td>Relative humidity</td>
</tr>
<tr>
<td>RMSE</td>
<td>Root mean squared error</td>
</tr>
<tr>
<td>Abbreviation</td>
<td>Full Form</td>
</tr>
<tr>
<td>--------------</td>
<td>-----------------------------------------------</td>
</tr>
<tr>
<td>SAM</td>
<td>Southern Annular Mode</td>
</tr>
<tr>
<td>SH</td>
<td>Sensible heat flux</td>
</tr>
<tr>
<td>SST</td>
<td>Sea surface temperature</td>
</tr>
<tr>
<td>SLATS</td>
<td>Queensland Statewide Land Cover and Trees Study</td>
</tr>
<tr>
<td>SOI</td>
<td>Southern Oscillation Index</td>
</tr>
<tr>
<td>TRMM</td>
<td>Tropical Rainfall Measuring Mission</td>
</tr>
<tr>
<td>USGS</td>
<td>The U.S. Geological Survey</td>
</tr>
<tr>
<td>UTC</td>
<td>Coordinated Universal Time</td>
</tr>
<tr>
<td>WRF</td>
<td>Weather Research and Forecasting</td>
</tr>
</tbody>
</table>
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