1 Introduction

Climate is changing and will continue to change. Societies and ecosystems are affected by and often depend on climate and its variability. Already in 1992, the United Nations Framework Convention on Climate Change stated that all parties shall "cooperate in preparing for adaptation to the impacts of climate change" (United Nations 1992). Over the last decades, several countries have developed national adaptation strategies. The EU strategy on adaptation to climate change (European Commission 2013), for instance, acknowledges the need to take adaptation measures at all levels ranging from national to regional and local levels. The Global Framework for Climate Services (GFCS), established in 2009, sets out to develop and communicate climate information to "enable better management of the risks of climate variability and change and adaptation to climate change" (http://www.wmo.int/gfcs/vision). In short, there is an urgent demand for scientifically credible climate change information, in particular at the regional scale (Hewitt et al. 2012). One approach to obtain information about regional climate change is downscaling of global climate projections. In fact, a plethora of different data products have already been made available via internet portals.

Yet the provision of regional climate change information is one of the big challenges in climate science (Schiermeier 2010) and still a subject of essentially basic research (Hewitson et al. 2014). A *Nature* editorial prominently pointed out that "certainty is what current-generation regional studies cannot yet provide" (Nature 2010). Kundzewicz and Stakhiv (2010) argue that climate models have originally been developed to guide mitigation decisions. They could provide a broad picture of global climate change but would not yet be skillful to serve as input for regional adaptation planning. Kerr (2011*b*) brings forward a range of arguments which have been issued against current downscaling practice, and, in a later piece (Kerr 2011*a*), discusses the challenges of providing actionable climate information.

Against this background, the book at hand attempts to provide a reference for a range of approaches and methods often summarised as statistical downscaling. At the same time, the book aims to put the more technical issues of statistical downscaling into the broader context of user needs, regional climate modelling uncertainties and limitations, and good scientific practice. To begin with, we would like to sketch the scientific idea of statistical downscaling and then give some guidance on how to best approach this book.

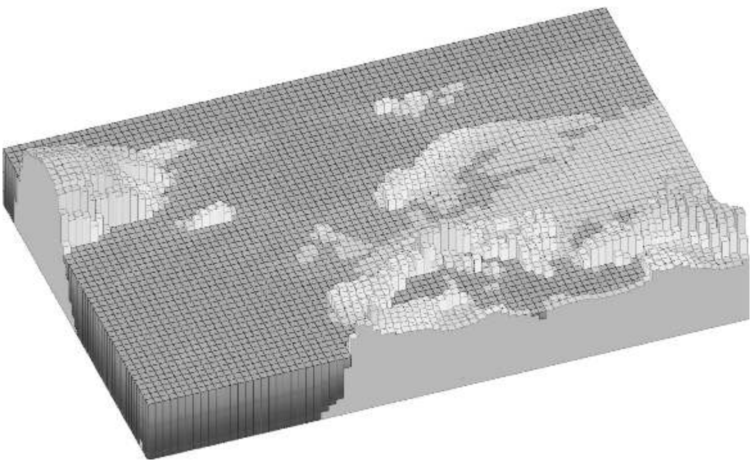

Figure 1.1 Landmask and elevation model of a typical state-of-the-art GCM. The horizontal resolution is approximately $1.13^\circ \times 1.13^\circ$. Adapted from Figure 1.4 (bottom panel AR4), Solomon et al. (2007).

1.1 Statistical Downscaling and Bias Correction in a Nutshell

On the 18th of July 2009, heavy rains fell in the city of Graz, Austria. The soil was still saturated from a wet spell in late June, such that the city's streams burst their banks, and several districts were flooded. Hydrologists, engineers and city planners might all be interested in the risk of such a flooding to happen again: it depends on the precipitation history over the preceding weeks, on the intensity of the rainfall event and on its spatial-temporal distribution. But all these users of climate information are more and more concerned not only with risk in present climate but also with potential changes of risk in a warmer future climate.

Much of our knowledge about future climate change stems from projections with global general circulation models (GCMs). For instance, the ensemble simulations carried out within the coupled model intercomparison project (CMIP, Meehl et al. 2007*a*, Taylor et al. 2012) have been the backbone of many prominent messages published in the last Intergovernmental Panel on Climate Change (IPCC) assessment reports (e.g. Meehl et al. 2007*b*, Collins et al. 2013). But even state-of-the-art GCMs still have a rather coarse resolution (Figure 1.1). As a consequence, regional-scale topography and meteorological processes, in particular those responsible for many types of extreme events, are not represented by these models.

The idea of downscaling is to bridge the gap between the large spatial scales represented by GCMs to the smaller scales required for assessing regional climate change and its impacts. Two major types of downscaling exist: in dynamical downscaling, a high-resolution regional climate model (RCM) is nested into the GCM over the domain of interest (Rummukainen 2010). In statistical downscaling, empirical links between the large-scale and local-scale climate are identified and applied to climate model output.

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Douglas Maraun , Martin Widmann
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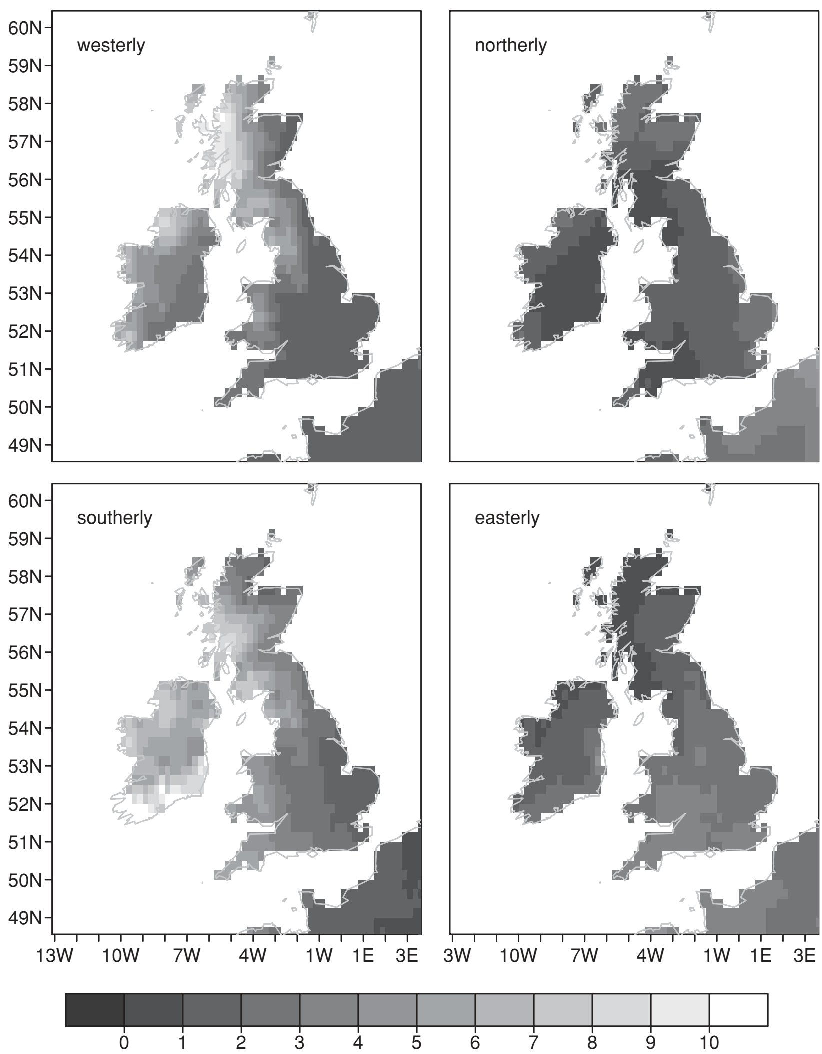

Figure 1.2 Precipitation composites [mm/day] on the British Isles for selected Lamb weather types (Lamb 1972), which describe the main atmospheric circulation patterns over the British Isles. Based on E-OBS daily data (Haylock et al. 2008) and the Lamb weather types from the Climatic Research Unit (Jones et al. 2013) for the period 1950 to 2016.

Figure 1.2 illustrates such an empirical relationship for the British Isles. The panels show the average precipitation which falls under four different situations of the large-scale atmospheric circulation: in case of a westerly flow, the highest precipitation is expected along the west coast of Ireland and Great Britain. Highest intensities occur in particular in the western Scottish Highlands – the South East of England is typically dry under such conditions. Northerly airflow instead brings cooler air, which typically carries less moisture. Precipitation intensities are thus lower – with relatively high values in the exposed regions of Northern Scotland, the North East of Ireland and East

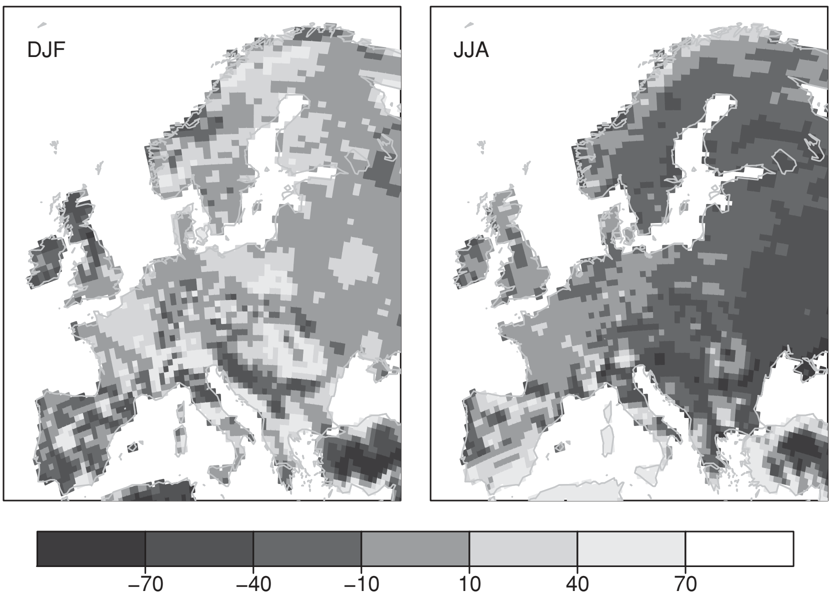

Figure 1.3 Relative precipitation bias [%] of the RCM RACMO2 (van Meijgaard et al. 2008), driven by the GCM EC-EARTH (Hazeleger et al. 2010) compared to E-OBS data (Haylock et al. 2008), for the period 1971–2000. Left: winter; right: summer.

Anglia. If the flow arrives from the south, precipitation is highest in the South of Ireland but also along the west coast of Great Britain. Finally, easterly flow brings higher intensities to the East coast and to the hills of South-West England and Wales, the first orographic barrier in the south of Great Britain. Other such situations – called weather types in meteorology – would also be associated with typical precipitation patterns. Thus, knowing the large-scale flow, one can predict the regional distribution of precipitation, including the effects of regional orography. Applying this empirical relationship to the large-scale circulation simulated by a GCM would thus downscale the GCM and generate regional-scale precipitation fields. In statistical downscaling jargon, this would be a weather-type-based perfect prognosis model. Under the assumption that the empirical link between large-scale circulation and local precipitation remains valid in a future climate, one could apply the model to generate regional precipitation projections.

But even if a GCM would resolve the climate processes relevant for a particular user, the simulated climate would typically still substantially deviate from real-world climate. In fact, even after dynamical downscaling the simulated regional climate is in general biased compared to observations. Figure 1.3 shows the relative error between simulated and observed mean winter (left) and summer (right) precipitation climate. The simulation has been conducted with the RCM RACMO2, driven by the GCM EC-EARTH, two well-performing climate models. In some parts of Europe, the relative error is below 10%, but in many regions it exceeds plus or minus 70%. Impact modellers often cannot use such simulations directly; they demand some form of statistical post-processing to adjust the model output towards observations. Again, one could establish an empirical link: here the ratio between the simulated and the observed mean precipitation. Applying this scaling factor to the simulation, one would "remove" the model bias. This bias correction procedure is a simple form of model output statistics. Under the assumption that the correction function is applicable in a future climate, one could post-process future precipitation projections.

The terms "statistical downscaling" and "bias correction" are used differently in different communities and countries. Many US researchers use the terms essentially interchangeably. In other countries, climatologists often reserve the term "statistical downscaling" or even "empirical statistical downscaling" for the first approach, which we call perfect prognosis. The term "bias correction" is used by dynamical downscalers and hydrologists exclusively for the second approach, but some users of empirical statistical downscaling would claim that also their approaches are bias correcting. Recently, some authors began to argue that bias correction, as it does only post-processes model output, should better be called bias adjustment. And being slightly meticulous, one could even argue that many statistical methods from either approach do not generate time series representing local climate – that is, they are not really downscaling. We therefore decided to follow a semi-pragmatic approach. In general, and in particular when we compare the two approaches, we use the terminology originally proposed by Klein and Glahn 1974: we call the first approach perfect prognosis (PP) and the second model output statistics (MOS). The key advantage is that these terms are precisely defined and at the same time get more and more used across disciplines. But since most MOS approaches in this book are mere bias corrections, we often use this simpler term. The term "statistical downscaling" is used rather colloquially to subsume both approaches.

1.2 How to Read This Book

We hope the book will prove useful for different audiences, each with its specific backgrounds and needs. One could approach the book simply by reading it in the given order or use selected chapters as reference. In particular, these would be the technical chapters on statistical methods (Chapter 6) and dynamical modelling (Chapter 8), the different downscaling approaches (Chapters 11–14), and finally the evaluation and performance (Chapters 15 and 16).

Readers who are new to the field or who are mainly interested in using and interpreting downscaling results may instead start reading the book from Chapter 18. This chapter provides a condensed summary of how to best apply statistical downscaling in practice: what are important issues to be considered? Which methods are useful in which context? How could one deal with uncertainties? In each section, the reader is then directed towards more in-depth discussions in the preceding parts of the book. We will sketch these in the following.

Part I provides both a broader context and the necessary technical background. In Chapter 2 we introduce climate and weather phenomena governing regional climates. After a historical overview (Chapter 3), we discuss the main assumptions, requirements and concepts of downscaling (Chapter 4). User needs are reviewed and discussed in

Chapter 5. Chapters 6 to 9 provide background in statistical modelling, a summary of observational data and dynamical climate modelling and a discussion of their limitations and uncertainties.

Part II is the core of the book and introduces the overall structure of downscaling methods (Chapter 10), as well as the major approaches PP, MOS, weather generators and combinations of these approaches (Chapters 11–14). Each of these chapters provides an overview of widely used methods as well as their structural limitations and the assumptions underlying their use.

Part III discusses the performance of statistical downscaling and regional modelling and its use in practice. A general framework for the evaluation of regional climate projections is presented in Chapter 15 and a synthesis of actual performance in Chapter 16. The ongoing debate of the applicability of downscaling is critically reviewed in Chapter 17. Finally, Chapter 18 provides a synthesis of the book and guidelines for the use of downscaling in practical applications.

Part I

Background and Fundamentals

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2 Regional Climate

Before discussing regional climate modelling in more detail, it is sensible to briefly sketch regional climate itself, and the factors controlling regional climate and climate change. We use the term rather loosely, spanning a range of scales below the continental and synoptic scale. Some may prefer to distinguish between regional and local climate – we decided to use only one term – which scales we refer to should become clear from the context. In fact, we will discuss that it is essential to define the relevant scales for any user case individually. In some situations, a region might be a whole country, in other situations a district or even just a specific valley.

Regional climate is determined by a vast number of climatic processes spanning global to local scales. To successfully model regional climate change, it is essential to successfully model the processes relevant for the specific application. We will therefore sketch these processes in the following, considering the Alps as a showcase.

2.1 Large- and Planetary-Scale Processes

The global temperature field is to a first order determined by the influence of latitude and land–sea distribution on the radiative balance, and the different thermal capacity of land and ocean. The Alps are located in the mid-latitudes, in a temperate climate (Figure 2.1, left). The mid-latitudes are a region with a strong meridional temperature gradient and high baroclinic instability, which controls the position of the jet stream and fuels the North Atlantic storm track (Hoskins and Valdes 1990, Lynch and Cassado 2006). Jet streaks, regions of acceleration and divergence in the jet stream, control the genesis of cyclones. The upper-level winds also steer the path of cyclones. Conversely, the passage of cyclones along the storm track drives the jet stream, intricately linking the two phenomena (Woollings 2010). The Alps are located just south of the climatological mean position of the polar front and polar jet in both summer and winter. Cyclones and anticyclones advect different air masses to central Europe. For instance, arctic continental air brings cold and dry air, tropical maritime air is warm and moist, tropical continental air is hot and dry. The jet stream itself follows planetary-scale Rossby waves meandering around the globe. If the meanders have a large amplitude, cold air is transported far south and warm air far north (see Figure 2.2). Such situations are often persistent in time, associated with blocking events: high-pressure systems that block the passage of cyclones and cause heat waves and drought in summer and cold spells in winter.

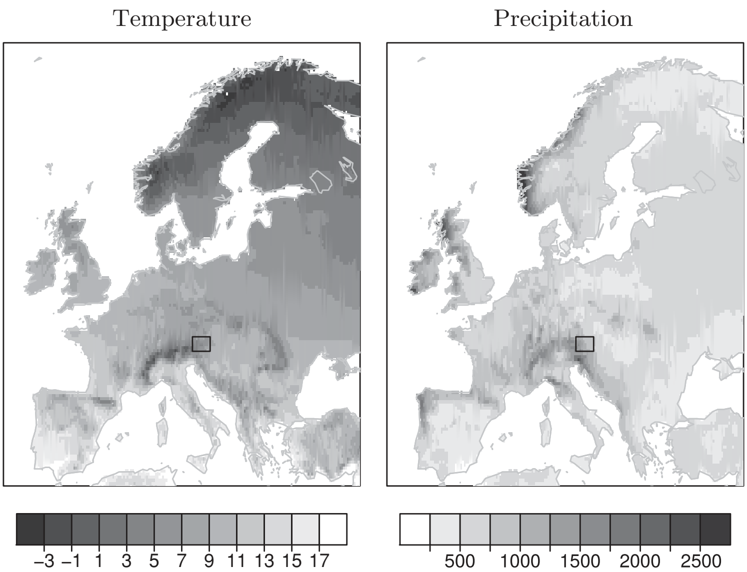

Figure 2.1 Annual mean temperature [°C] and annual mean precipitation [mm] in Europe for the reference period 1971–2000, calculated from the EOBS data set, version 14.0 (Haylock et al. 2008). See Chapter 7 for a discussion of gridded data sets.

If the meanders are weak, the flow is mainly zonal, corresponding to a positive phase of the North Atlantic Oscillation (NAO) – the pressure gradient between the Icelandic low and the Azores high. The mean position of Rossby wave meanders – and hence the position of the stationary mid-latitude high- and low-pressure systems – is controlled by the land–sea contrast and major mountain ranges such as the Rocky Mountains or the Greenland ice shield but also the sharp sea surface temperature gradient in the north Atlantic and the north-eastward orientation of the eastern North American coastline (Hoskins and Karoly 1981, Held et al. 2002, Minobe et al. 2008, Brayshaw et al. 2015).

Mid-latitude winter climate is influenced by processes in the stratosphere (Limpasuvan et al. 2004). During the polar night, a strong circumpolar vortex of westerly winds rotates around the pole. In the Northern Hemisphere, tropospheric Rossby waves penetrating into the stratosphere may slow down these winds and ultimately break down the vortex. As a result, stratospheric temperatures may suddenly rise by several tens of degrees. The breakdown of the polar vortex weakens the polar jet (Baldwin and Dunkerton 2001) and favours cold air outbreaks into the mid-latitudes (Thompson et al. 2002). El Niño/Southern Oscillation influences European winter climate, arguably via a stratospheric teleconnection (Ineson and Scaife 2009).

Severe weather often happens along the fronts between different air masses. Winter storms typically pass north of the Alps, supported by the orientation of the main ridge of the Alps, which spans an arc from south-west to north-east. As a result, precipitation is typically higher along the north-western flank (Figure 2.1, right). The north-west of the Alps is still under maritime influence from the Atlantic, the south from the

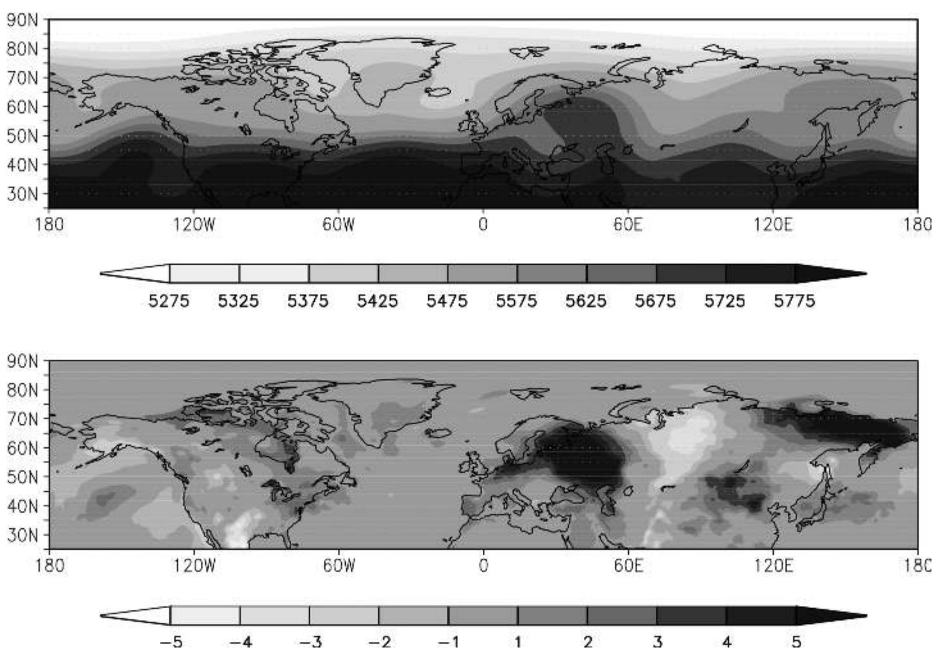

Figure 2.2 Extreme events of July 2010. Monthly mean 500hPa geopotential height [gpm] (top) and maximum temperature anomaly [K] (bottom). The polar jetstream, in the position of the strong geopotential height gradient, meanders strongly as a Rossby wave with a particularly high amplitude over eastern Europe. Hot air is carried northwards in eastern Europe and eastern Siberia, causing, e.g., the severe Russian heat wave. Likewise cold air is transported southwards, causing a cold wave over central Siberia. Based on ERA-Interim (Dee et al. 2011*b*), adapted from a figure created with KNMI Climate Explorer (https://climexp.knmi.nl).

Mediterranean. The Alps themselves contribute to lee cyclogenesis in the gulf of Genoa (Barry 2008). Some of these Genoa lows, called Vb cyclones, travel across Italy and the Adriatic towards eastern Europe. They are responsible for the precipitation maxima in the Julian and Dinaric Alps, north-east of the Adriatic.

In other regions of the world, similar large-scale processes influence the particular climate. Tropical and sub-tropical climates are controlled by the inter-tropical convergence zone (ITCZ), the Hadley and Walker circulation and the Monsoons (Goosse 2015). Along the ITCZ, moist air converges and deep, organised convection occurs. Along the tropopause, the air travels polewards, cools, is deflected by the Coriolis force and sinks over the subtropics, where the Earth's major deserts are located. Temperature contrasts between the oceans and land induce the continental-scale Monsoonal circulation. During summer, the winds blow onshore, carry moisture and cause heavy rains. The orographic forcing of the Himalaya is responsible for the highest rainfall totals worldwide. During winter, the dry winds blow offshore. Specific to the climate within about 5° of the equator is the vanishing Coriolis force: here, no tropical cyclones can form. Oceans are a major source of moisture, and their high heat capacity dampens seasonal temperature variations and thus shapes maritime climates.