Detecting and attributing recent warming in Europe

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Thesis submitted for the degree of Philosophiae Doctor (PhD)

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University of Oslo
Oslo, Norway
2017
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Series of dissertations submitted to the
Faculty of Mathematics and Natural Sciences, University of Oslo
No. 1915

ISSN 1501-7710

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Cover: Hanne Baadsgaard Utigard.
Print production: Reprosentralen, University of Oslo.
Abstract

Regional warming trends for Europe during the last decades have been larger than the global mean trend, especially on the seasonal scale. An improved understanding of hydroclimatic drivers of the warming is key to improve estimates of the energy balance and water balance terms. In addition to increased radiative forcing from greenhouse gases, the atmospheric circulation is the key driver of seasonal temperature trends in Europe. Of secondary importance are positive feedback mechanisms between the land surface and the atmosphere that enhance warming. For instance, partitioning of net radiation into latent and sensible heat is controlled by properties of the land surface, such as snow albedo and soil moisture content. In the first part of the thesis, monthly temperature trends and their physical drivers were explored by separating the change signal into circulation changes and other factors, so-called within-type changes, which may arise from local feedback mechanisms. This separation allowed identifying months and regions where temperature changes can mainly be attributed to circulation changes, and where within-type changes may also play a role. Using a novel probabilistic approach for statistical attribution, we showed that circulation changes could not account for all the observed warming in Europe over the period 1981–2010. Significant warming, such as the warming covering large parts of Europe in April, June–August, must also be caused by other factors. The second part of the thesis assessed two possible causes for this warming through a detailed study for Norway, where within-type changes were found to play a role in explaining the warming in April and July–September. This was done, first, by investigating the relationship between snow cover changes and monthly temperature trends for all of Norway (snow albedo feedback), and, second, by performing a diagnostic land surface model experiment testing the hypothesis that dry conditions may enhance the warming (soil moisture–temperature feedback). A significant decrease in snow cover was detected for most of Norway, accompanied by strong temperature trends, particularly in spring and at low elevations. Targeted model experiments with a coupled WRF–Noah-MP model for South Norway confirmed that the snow albedo feedback enhanced warming in spring, and showed that the soil moisture–temperature feedback could enhance warming during parts of the summer, provided that the soil moisture content was sufficiently low. This work has contributed to an improved understanding of the causes of recent temperature changes in Europe and Norway in particular, by assessing the relative role of circulation changes and within-type changes on regional-scale warming, through combining a statistical attribution approach and a physically-based modelling approach in a unique way.
Summary for the layperson

Europe has warmed faster than the rest of the globe during the last decades, especially when considering trends for each season separately. In addition to warming caused by greenhouse gas emissions, the atmospheric circulation is the key driver of seasonal warming in Europe. For example, westerly airflow tends to bring warmer weather in winter. Of secondary importance are climate feedbacks that may enhance warming locally. For example, snow reflects more sunlight than grass or bare ground. A retreating snow cover reveals a darker ground that allows more sunlight to be absorbed, and leaves more energy available for further melting and warming. This is the snow albedo feedback (albedo is a measure of reflectance). A different feedback, the soil moisture–temperature feedback, mainly acts during warm spells when the soil moisture may decrease to such low levels that evaporation is reduced. Since evaporation requires energy, less evaporation implies less cooling, and thus enhanced warming.

In the first part of the thesis, monthly temperature trends and their physical drivers were explored by separating the temperature trend signal into circulation changes (that is, variations in the occurrence of, for instance, westerly weather) and other factors (within-type changes). “Other factors” include local feedbacks enhancing the warming, but may also include warming due to greenhouse gas emissions. Note that this separation does not divide causes of warming into natural variability and anthropogenic forcing. To determine whether the warming could be attributed to circulation changes, we developed a novel statistical approach and applied it to temperature trends for Europe (Paper II). Circulation changes could not account for all the observed warming in Europe over the period 1981–2010. Therefore, other factors than circulation changes contributed; for instance, to the warming covering large parts of Europe in April, June–August. The second part of the thesis assessed the snow albedo feedback and the soil moisture–temperature feedback as possible causes of warming for Norway. Analyses over the past five decades showed strong decreases in snow mass at low elevations, accompanied by warming (Paper III). From Papers I–III, we chose South Norway as a focus region, in April and July–September, for a detailed model study because this region displayed potential for positive feedbacks (within-type changes). Last, we used a regional climate model to test the contributions of the snow albedo feedback and the soil moisture–temperature feedback (Paper IV). The snow albedo feedback contributed to enhanced warming in spring. Further, the positive soil moisture–temperature feedback could enhance warming in summer, provided that the ground was sufficiently dry. It is important to study these feedbacks to correctly predict temperatures, and other variables, in climate models.
Acknowledgments

Completing a PhD is like climbing a mountain; you have to hike a long way to get a good view. In the beginning, you look in one direction and you don’t see the rest of the landscape, much like the start of a PhD project. You see from the map (the literature) that a whole landscape exists, but you don’t appreciate the full view (the body of knowledge) until you are at the peak. And then, you still have a long way down. During the journey, you have to stick to your chosen path (follow the theme of the thesis). I’d like to thank my main supervisor, Lena Tallaksen, for giving me the map and leading me towards the right path up and down the mountain. She could not join me all the way, but she made sure I had the right travelling companions who could point me in the right direction. Thanks for all your support, for being available for discussions, and teaching me how to write science papers. A big thank you to my co-supervisors, Frode Stordal and Chong-Yu Xu, and co-authors for your help. All of you have included me into your scientific life, and you have always introduced me to your scientific colleagues and their PhDs. Keep doing that, your network has been very helpful for me. For example, Paper II would not have developed if it wasn’t for the conference hosted by IAHS FRIEND-WATER in October 2014. Thanks to IIASA in Laxenburg for hosting the annual Young Scientist’s Summer Programme, which I attended in 2014. Last, but not least, many thanks to Helene for guiding me through the recesses of WRF and Noah!

Team Nilsen et al. has consisted of more people than my supervisors and co-authors. I’d like to thank all my fellow colleagues, PhD-students, and my own students in hydrology. The IT section at has always been supportive, especially when I needed it the most (when WRF segfaulted, and when I haphazardly deleted my home directory). A special thanks to Hege Hisdal and Stein Beldring for everything you have taught me, and for giving me so much flexibility in my new job. Thanks to Trine and Helga for the moral support towards the end of my PhD; to the University Library’s Academic Writing Centre for guiding my writing; to Børneblæs brass and barbecue club (Nedre Blindern ryttmeegg- og kremfløye-kjøkken) for the social support; to Climabyte; to the Lund Prestrud family for having me at Lyngør two weeks every summer; to the bug in the webplayer of my favourite radio channel that allowed me to listen to the commercial-free night broadcasts during daytime; to Turid for reading draft upon draft, and Sarah for consulting services on commas and parentheses.

Last, but not least, I would like to thank mom, dad, Turid, and Audun for your everlasting support, for your patience, and for being there for me. I couldn’t have done this without you. Thank you very much!
5 Findings

5.1 Paper I: Recent trends in monthly temperature and precipitation patterns in Europe

5.2 Paper II: A probabilistic approach for attributing temperature changes to synoptic type frequency

5.3 Paper III: Five decades of warming: impacts on snow cover in Norway

5.4 Paper IV: Diagnosing land–atmosphere coupling in a seasonally snow-covered region (South Norway)

6 General discussion

6.1 Trend detection

6.1.1 Comparing detected trends across time periods and datasets

6.1.2 Sensitivity to time period and time scale

6.2 Statistical attribution of recent temperature trends

6.2.1 Differences in circulation changes

6.2.2 Differences in within-type changes

6.2.3 Using only one classification of synoptic types

6.3 Exploring the physical drivers of temperature trends

6.3.1 Linkage between statistical analyses and model studies

6.3.2 Potential to improve land surface models

6.3.3 Future changes

7 Conclusions

8 References

9 Papers
List of papers

Paper I

Paper II

Paper III

Paper IV

For Papers I, II and IV, I was responsible for the programming, modelling, analyses, and writing. P. James provided the SVG data and revised the paragraph about SVG in Paper II. For Paper III, J. Rizzi performed the analyses and drafted the initial version of the paper, whereas I drafted the introduction and discussion sections of the manuscript, and reviewed the manuscript, in addition to performing initial analyses of temperature trends in SeNorge that the current analyses are based on.
List of symbols and abbreviations

$\alpha$, albedo

CTR, control run in the model diagnostic study

ERA-Interim, global, gridded climate reanalysis dataset

GWB, Gridded Water Balance model

$LE$, latent heat [W m$^{-2}$]

LSM, land surface model

$m_{WFDEI}$, WFDEI temperature trend

$m_{SC}$, median of the distribution of circulation-induced trends

NAO, North Atlantic Oscillation

Noah-MP, land surface model

NVE, the Norwegian Water Resources and Energy Directorate

$P$, precipitation

$\rho(\text{SM}, LE)$, terrestrial leg of land–atmosphere coupling

$\rho(LE, T)$, atmospheric leg of land–atmosphere coupling

$r_{circ}$, trend ratio

$R_n$, net radiation

$SCE$, snow cover extent

SeNorge, a high-resolution, gridded reference dataset for Norway

$SH$, sensible heat [W m$^{-2}$]

$SM$, soil moisture [%]

$SM_{crit}$, a soil moisture content below which $LE$ is limited [%]

SP, snow-poor run in the model diagnostic study
SR, snow-rich run in the model diagnostic study
ST, synoptic type
SVG, SynopVis Grosswetterlagen, a classification of synoptic types
SWE, snow water equivalent
$T$, air temperature [$^\circ C$]
$t_{obs}$, observed trend
$t_{circ}$, circulation-induced trend
WFDEI, Watch Forcing Data ERA-Interim, a gridded, global hydroclimatic dataset
WRF, Weather Research and Forecasting model
WRF–Noah-MP, coupled atmosphere and land surface model
Chapter 1

Introduction

1.1 Motivation

The last decades have experienced strong warming globally, with each decade after 1970 being warmer than the previous (Hartmann et al. 2013). This warming has implications for the water balance, for instance: changed precipitation patterns, altered evapotranspiration, a shorter snow season and less snow cover (Hartmann et al. 2013; Vaughan et al. 2013). The warming is partly caused by increased radiative forcing from anthropogenic greenhouse gases emitted since the industrial revolution. Besides the direct influence of increased radiative forcing, the atmospheric circulation is a main driver of hydroclimatological change on the regional scale, through advection of energy and water fluxes. In addition, feedbacks between the land and the atmosphere may lead to warming that is independent of atmospheric circulation changes.

Variations in the atmospheric circulation have been highlighted as the main driver of temperature variability in Europe, especially in winter (e.g. Corti et al. 1999; Vautard and Yiou 2009; Hoy et al. 2013b; Fleig et al. 2015; Saffioti et al. 2015). In these studies, the large-scale circulation is characterised by synoptic types (STs) which are used to separate climatic trends into those caused by circulation changes, and those caused by other factors (Beck et al. 2007; Cahynová and Huth 2009; Jones and Lister 2009; Küttel et al. 2011). Changes in the atmospheric circulation cannot explain all the observed warming, however (Barry and Perry 1973; Beck et al. 2007; Cahynová and Huth 2009). The part of the temperature trend that cannot be attributed to circulation changes (or more specifically, changes in the frequency of synoptic types) is called within-type changes (or more specifically, changes within the synoptic types). Within-type changes imply that the air masses change (warm) over time (Beck et al. 2007). Within-type changes may arise from different sources, including radiative for-
cing from greenhouse gases and positive feedback mechanisms between the land surface and the atmosphere that enhance the warming regionally (Küttel et al., 2011).

Another driver of warming is land-atmosphere interactions. The land surface controls the surface energy balance, for instance through the partitioning of net radiation into latent and sensible heat. This partitioning depends on the reflectivity of the land surface (albedo), which is high for snow and low for soil and vegetation. Strong declines in the snow cover are detected in seasonally snow-covered regions in the Northern Hemisphere, especially during the melt season (Derksen and Brown, 2012; Thackeray and Fletcher, 2016). The snow albedo feedback (Figure 2.2a) is initiated by warming that reduces the albedo and increases the amount of net radiation available on the ground, which enhance temperatures (e.g. Dickinson, 1983; Groisman et al., 1994b; Chapin et al., 2005; Hall and Qu, 2006).

During dry periods, a positive feedback between soil moisture and temperature may arise when warming reduces the soil moisture content, which reduces the latent heat as the soil dries out and leaves more energy available for sensible heat. This can ultimately enhance the initial warming. The soil moisture–temperature feedback is well-studied for transitional zones between dry and humid climates (Koster et al., 2004; Dirmeyer et al., 2013 e.g.), but less so at higher latitudes with a moist climate. In seasonally snow-covered regions, such as in Norway, the snow cover in spring influences the soil moisture storage at the start of summer. A small snow storage or early snowmelt leaves less water available to replenish the soil moisture storage (Wilson et al., 2010; Xu and Dirmeyer, 2013b). Snow cover anomalies may therefore influence soil moisture anomalies at high latitudes.

Identifying regions and time periods where the temperature trend is influenced by circulation changes or within-type changes is a first step towards understanding the detailed mechanisms leading to temperature trends. The importance of within-type changes relative to circulation changes varies depending on the region and time of year (Jones and Lister, 2009; Küttel et al., 2011; Cahynová and Huth, 2016; Fleig et al., 2015). Different feedback mechanisms act at specific times of year and in specific regions. For instance, the snow albedo feedback is limited to snow-covered regions, and is strongest in the melting season, when the incoming sunlight is strongest (Callaghan et al., 2012). Detection of trends on the monthly scale can therefore help explain the driving mechanism. The goal of this thesis is to detect recent trends for Europe on the monthly scale, and to improve our process understanding of key drivers of change. An improved understanding of local processes enhancing warming is key to improve estimates of the energy balance and water balance terms correctly. Because recent seasonal temperature trends in Europe have been accompanied by large changes in
the snow cover and summer drying, we focus on the snow albedo feedback and the soil moisture–temperature feedback.

Last, a word about terminology. The terms “detection” and “attribution” used in this thesis are not to be confused with the scientific field of “detection and attribution” that seeks to explain climate change either by anthropogenic forcing or natural variability (e.g. Stott et al. 2004; Stone et al. 2009; Pall et al. 2011). In this thesis, “detection” or “attribution” refer to the common uses of the words: to detect – calculate – changes in a variable and to attribute – find a cause – of those changes. Note also that we do not separate temperature trends into anthropogenic forcing and natural variability, but rather into circulation changes and within-type changes.

1.2 Objectives

The main objective of this thesis is to detect monthly hydroclimatic trends in Europe and explore their physical drivers through statistical attribution and through a model diagnostic study. In particular, the importance of land–atmosphere interactions in Norway is explored. Four research questions are defined:

1. Which regions and months have experienced significant trends in hydroclimatology in Europe?
2. Can the detected trends for Europe be attributed to circulation changes or within-type changes?
3. How do snow cover changes influence warming in a region with seasonal snow cover?
4. For regions with within-type changes, can the detected trends be attributed to the snow albedo feedback (in spring), and the soil moisture–temperature feedback (in summer)?

1.3 Study design

This work combines traditional statistical methods with model diagnostic studies. Temperature and precipitation trends were detected on the monthly scale for Europe to address research question 1, in Paper I (Nilsen et al., 2014) and Paper II (Nilsen et al., 2017b). Because of the noisy trend patterns for precipitation, we continued
with temperature only. Research question 2 was addressed by statistical attribution methods commonly used in synoptic climatology. The method of hypothetical trends (Leathers and Ellis, 1996) was used in combination with trend ratios (Cahynová and Huth, 2009) in Paper I, whereas Paper II provides a probabilistic expansion of trend ratios. This novel probabilistic approach improves upon the trend ratio method by evaluating the statistical significance of synoptic circulation changes on observed climate trends.

Land surface models (LSMs) are useful tools for detailed process studies of what drives changes in hydroclimatic variables. For that purpose, a smaller target region in Europe was selected for high-resolution model diagnostic studies. South Norway was identified as a region where trends could be attributed to within-type changes in April and July–September. With its complex topography, South Norway spans different climate zones, ranging from a maritime moist and temperate climate in the western part, to a dry inland climate in the eastern part. A detailed dataset of estimated observed hydroclimatic variables is available for Norway (SeNorge), which allows for evaluation of the model results. The influence of a changing snow cover on temperatures was investigated for this dataset (research question 3, Paper III; Rizzi et al., 2017). A modelling study of the snow albedo feedback and the soil moisture–temperature feedback was performed using a regional climate model (WRF) coupled to the Noah-MP land surface model (research question 4, Paper IV Nilsen et al., 2017a). Although this model-diagnostic study for two summers (2006 and 2014) is not able to explain the full within-type trend over the period 1981–2010, it proves that the soil moisture–temperature feedback is one factor explaining the enhanced warming. Together, these four papers form a study of detailed mechanisms of what controls regional temperature trends. This thesis has contributed to the strategic research area LATICE (Land-ATmosphere Interactions in Cold Environments) funded by the Faculty of Mathematics and Natural Sciences at the University of Oslo (http://www.mn.uio.no/geo/english/research/groups/latice/index.html).
Figure 1.1: Flow chart describing the papers, including the research question numbers (in parentheses) and main methodology.
Chapter 2

Scientific background

This chapter presents the scientific background for the work done as a part of the thesis. First, a summary of published seasonal changes in temperature, snow, and soil moisture (relevant for research question 1) is presented before a review of possible processes explaining detected seasonal changes, including attribution studies (relevant for research questions 2–4).

2.1 Detected climate change in Europe and Norway

Due to a large number of detection studies, this summary is limited to i) Europe and Norway, ii) recent changes covering the past decades, iii) seasonal and monthly changes, and iv) temperature and the related variables snow and soil moisture.

2.1.1 Recent seasonal temperature changes

The temperature change for Europe has exceeded the global mean trends (Jones et al. 2012, Hartmann et al. 2013, EEA 2017). Local temperature trends are sensitive to the region, season, period, method, and dataset under study. A wide range of values are therefore reported for Europe, however, clear warming signals emerge across studies. Annual temperature trends from Europe-wide studies range from 0.3 °C/decade for the period 1960–2015 at 2° spatial resolution (EEA 2017) to more than 1 °C/decade for the period 1976–2009 for point observations (Klein Tank et al. 2005), depending on region, time period, and resolution (Table 2.1). On the monthly scale, trend magnitudes are even higher, exceeding 3 °C/decade in northern Europe in winter (DJF) for the period 1976–2009 (Klein Tank et al. 2005). Winter warming in northern Europe and
summer warming in southern and central Europe are consistent across several studies (Klein Tank et al. 2005; EEA 2008, 2017). For Norway, temperature trends for the period 1955–2014 showed warming in all seasons, most pronounced in winter and spring (Førland et al. 2016) (Table 2.1).

2.1.2 Recent seasonal snow changes

The seasonal snow cover influences the climate over large parts of the globe, covering up to 30% of the global land area (Vaughan et al. 2013). The snow conditions over the Northern Hemisphere have changed during the last decades (Dery and Brown 2007). The Northern Hemisphere winter snowpack has increased over the past decades (Dery and Brown 2007; Callaghan et al. 2012) because of increased cold-season precipitation (Hartmann et al. 2013). Still, a dramatic snow cover decline has been observed at the start and end of the snow season, with a stronger decline in spring than in autumn (Fedesco et al. 2009). This asymmetry has been ascribed to the stronger incoming sunlight in the melting season than in autumn (Callaghan et al. 2012). The early spring snow cover in the Arctic has reduced by 18% in May–June, over the period 2008–2012 (Derksen and Brown 2012). Fennoscandia is one of the regions experiencing the strongest reduction in the snow season length (Callaghan et al. 2012). Studies for Norway have detected decreasing snow water equivalent (SWE) for the periods 1931–1960 and 1991–2009 in southern Norway (Skaugen et al. 2012), and decreasing snow depth for the period 1961–2010 (Dyrrdal et al. 2013). Also in Norway, the strongest reductions have been found in spring (Dyrrdal 2010). Snow depth and SWE have increased at high elevations, above approximately 1000 m a.s.l. (Skaugen et al. 2012).

2.1.3 Recent seasonal soil moisture changes

Seasonal soil moisture trends are sparse. Annual soil moisture trends show significant drying in southern Europe and significant wetting in northern Europe (Sheffield and Wood 2008; Hartmann et al. 2013), confirmed by trend studies of precipitation and evapotranspiration (Becker et al. 2013; Greve et al. 2014). Since soil moisture is closely related to precipitation, evapotranspiration, temperature, and streamflow, these variables may be taken as proxies for soil moisture changes. Summer streamflow has decreased in large parts of Europe for the period 1963–2000, particularly for southern and central Europe, and even for parts of northern Europe in August (Stahl et al. 2012). Streamflow trends for Norway decreased in summer (Wilson et al.)
Table 2.1: Comparison of temperature trends in regional trend studies for Europe. Seasonal trends from [Jones et al. (2012)] are averaged over months from their Table 2. The abbreviations in column Reg (for regions) indicate: Gl=Globally, NH=Northern Hemisphere (spatial average), Eur=All of Europe, Alp=The Alps, No=Norway, Fi=Finland.

<table>
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<td><strong>Summer (JJA)</strong></td>
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confirmed by \cite{Stahl_et_al.2012}. Projections show a continued drying trend in summer, both in Europe \cite{Seneviratne_et_al.2006,Eisner_et_al.2017} and Norway \cite{Hanssen-Bauer_et_al.2015}. In particular, the variability of the summer temperatures in Europe is expected to increase, which increases the probability of dry conditions in the future \cite{Schar_et_al.2004,Whan_et_al.2015}.

Figure 2.1: "Updated NAO Index after Jones et al. (1997) for boreal winter (DJFM) 1823/1824 to 2012/2013 re-normalized with respect to the full time period. The black line shows an 11-year running mean." \cite{Feser_et_al.2015}. Reprinted with permission from John Wiley and Sons.
2.2 Drivers of detected changes in the regional climate

2.2.1 Role of atmospheric circulation on the regional climate

The global oceanic and atmospheric circulation is primarily a mechanism to even out differences in the incoming energy between the poles and tropics (Lamb, 1982). In Europe, westerly airflow dominates the atmospheric circulation, advecting oceanic, mild and moist air onto northern Europe and dry air onto southern Europe (Becker et al., 2013). Variations in storm tracks, pressure systems and circulation patterns in the Northern Hemisphere are often described as the North Atlantic Oscillation, NAO (Hurrell et al., 2013) or as other circulation indices (Huth et al., 2008, 2016). Circulation indices are useful because of their strong positive correlation to local temperature and precipitation (Hurrell and Deser, 2009; Hurrell et al., 2013).

The NAO index is the difference between the normalised sea level pressure in the Iceland low and the Azores High (Hurrell et al., 2013). Strong westerly airflow across the North Atlantic and toward Europe is characterised with a positive NAO index (NAO+), whereas a negative NAO index (NAO-) is associated with weaker westerlies. The NAO index increased from NAO- to NAO+ between the early 1970’s and the mid-1990’s (Figure 2.1), which was accompanied by an increased occurrence of westerly airflow and frontal weather during that period (Hurrell et al., 2013). After the year 2000, NAO- prevailed until the winter 2012/2013, but has been positive since then (Hurrell and Staff, 2017). Numerous studies have focused on trends in the atmospheric circulation, or trends in variables that depend on the atmospheric circulation for the decades 1970, 1980 and 1990 (e.g. Klein Tank et al., 2005; EEA, 2008; Cahynová and Huth, 2009; Cahynová and Huth, 2016). Due to the shift in the NAO index, these studies are, however, not representative for the most recent past.

Links between the atmospheric circulation and the local climate are studied through the old field of synoptic climatology (Barry and Perry, 1973). Early weather prediction, prior to the advent of numerical weather prediction models, relied on relationships between the local climate and atmospheric circulation, and assumed that local climate trends could only be explained by changes in the atmospheric circulation (Dzerdzeevski, 1963; Yarnal, 1993). However, other factors than circulation changes have also been shown to contribute to local climate trends. For example, temperature changes are to a larger degree explained by changes in the atmospheric circulation in winter than in summer, when other factors contribute to warming (Beck et al., 2007; Cahynová and Huth, 2009; Fleig et al., 2015). Further, the importance of circulation changes varies over multidecadal time scales (Beck et al., 2007; Küttel et al., 2011).
An important topic in synoptic climatology is to separate local climate changes into those caused by changes in the atmospheric circulation, and those caused by other factors (Beck et al., 2007; Jones and Lister, 2009; Cahynová and Huth, 2009; Küttel et al., 2011). To relate the local climate to the atmospheric circulation, the circulation patterns are classified into synoptic types (STs) describing the synoptic situation on a given day (Huth et al., 2016). STs allow separating the cause of temperature trends into two parts: circulation changes and within-type changes. Circulation changes refer to changes in the frequency of STs that induce a climate trend, whereas within-type changes refer to changes within the STs that induce a climate trend (Beck et al., 2007). Within-type changes may include increased radiative forcing from greenhouse gases, and positive feedback mechanisms between the land and atmosphere, but may also stem from uncertainties in the stratification of STs (Küttel et al., 2011). Technically, a positive within-type change means that the STs become warmer over time. For those regions and time periods where the temperature trend is not due to more frequent warm synoptic types, possible interactions between the land surface and the atmosphere may play a role. Note that increased radiative forcing from greenhouse gases may also influence circulation changes.

**Attribution studies** STs are correlated with many local climate variables, which make them useful to study the physical processes controlling local climate change. Circulation changes have explained changes in a range of variables, from daily streamflow (Massei and Fournier, 2012), floods (Prudhomme and Genevier, 2011), and drought (Fleig et al., 2011; Hannaford et al., 2013) to ozone concentrations (Otero et al., 2016).

Beck et al. (2007) analysed monthly temperatures for central Europe for the period 1780–1995 and found that within-type changes were present in all seasons, particularly in summer. Cahynová and Huth (2010) found circulation changes in seasons other than winter to be of minor importance for temperature trends at stations in the Czech Republic (1961–1998). Studies by Beck et al. (2007); Vautard and Yiou (2009); Cahynová and Huth (2010); Hoy et al. (2013b) conclude that within-type changes explained a large part of the observed temperature trends, especially in summer in central Europe (Beck et al., 2007), though they differ slightly in their findings for other seasons. Küttel et al. (2011) analysed gridded temperature and precipitation time series for all of Europe and concluded that within-type changes were also important for temperature changes in winter, especially in eastern Europe and Scandinavia. One of the reasons for the different response in winter and summer is that winter weather is dominated by advective, large-scale processes, whereas summer weather is dominated by radiative, small-scale processes (Huth et al., 2016).
2.2.2 Role of the land surface on the regional climate

Accurate information about properties on the land surface is a vital requirement to estimate the components in the energy and water balances correctly. The land surface controls the local climate by regulating the exchange of heat, moisture and momentum between the land surface and atmosphere (Seneviratne et al. 2010). The detailed mechanisms of this coupling vary from region to region and season to season. For instance, a correct estimation of the net radiation over a given area requires an exact albedo. This is particularly important in seasonally snow-covered regions, where the albedo before snowmelt exceeds 0.5 (and may be as high as 0.9) and drops to lower values after snowmelt (0.3–0.1 in regions with forest, grasslands or bare soil) (Dingman 2002). Through its albedo, the snow cover controls the amount of reflected shortwave radiation (Groisman et al. 1994b; Betts 2000; Dery and Brown 2007). At other times of year, the influence of snow will be exceeded by other land surface properties. When the soil moisture content is so low that evapotranspiration becomes limited, the partitioning into latent and sensible heat fluxes is controlled by the soil moisture storage (Seneviratne et al. 2010).

The soil moisture storage may be affected by the snow cover in seasonally snow-covered regions. During years with less snow in stock less water is available to replenish the soil moisture storage during snowmelt (Wilson et al. 2010; Xu and Dirmeyer 2013b). Further, a longer snow-free period means that the soil moisture storage is allowed to decline for a longer period during the summer half year. The effect of snowmelt anomalies on the water balance weeks and months after snowmelt is an emerging field of study (Wetherald and Manabe 1995; Rowell and Jones 2006; Rowell et al. 2009; Im et al. 2010; Xu and Dirmeyer 2011; 2013a,b; Saini et al. 2016), and may be relevant for Norway because of the detected snow reductions in spring (Dyrrdal 2010) and summer drying (Wilson et al. 2010).

Positive feedback mechanisms Temperature changes may be modulated by feedback mechanisms between the land surface and the atmosphere that enhance or diminish the initial warming (Bony et al. 2006). Many feedback mechanisms are relevant to the climate system, such as the water vapour feedback and cloud feedbacks (Held and Soden 2000; Bony et al. 2006). In this thesis, we focus on feedback mechanisms involving changes in the hydrologic cycle that can alter an initial warming, namely the snow albedo feedback and the soil moisture–temperature feedback.
The snow albedo feedback (Figure 2.2a) is initiated by warming that reduces the albedo and increases the amount of net radiation available on the ground, which leaves more energy available for further melting and warming (e.g. Dickinson, 1983; Groisman et al., 1994b; Chapin et al., 2005; Hall and Qu, 2006).

A feedback between temperature and soil moisture is well established for periods with a strong soil moisture limitation on evapotranspiration (e.g. Koster et al., 2004; Seneviratne et al., 2006; Koster et al., 2010; Fischer et al., 2012; Miralles et al., 2014). In the soil moisture–temperature feedback loop (Figure 2.2b), a spell of warm and dry weather causes an anomalously low soil moisture storage that limits evapotranspiration (evaporative cooling) and thus enhances the initial warming (Seneviratne et al., 2010). This soil moisture–temperature feedback does not arise until the soil moisture has dropped below a critical point where evapotranspiration becomes limited by soil moisture (Figure 2.2c). However, a low soil moisture storage does not unconditionally feed back to temperature, as demonstrated by e.g. Quesada et al. (2012) for Europe, because high temperatures also depend on the atmospheric circulation situation: a blocking high pressure system favour heat waves, whereas cyclonic situations inhibit them.

A useful framework for determining if the conditions are sufficiently dry was first presented by Budyko in 1958, see a review in Greve et al. (2016) and Figure 2.2. The Budyko curve defines a relationship between an evaporative index, \( \frac{LE}{LE + SH} \) (or \( \frac{ET}{ET_{max}} \)), and a dryness index, \( SM \), in which evapotranspiration is limited by the available water (a supply limit) and by the available energy (a demand limit) (e.g. Berghuijs et al., 2014; Greve et al., 2016). Three regimes exist, depending on the soil moisture’s control on evapotranspiration. In humid climates, evapotranspiration is governed by other factors than soil moisture; this regime is called energy-limited. Below a critical soil moisture threshold, that depends on the land cover (Gallego-Elvira et al., 2016), the soil moisture limits evapotranspiration. Here, a soil moisture–temperature feedback may arise (Figure 2.2) that gradually reduces the soil moisture. This stage is a transition regime between dry and humid climates. Although such transitional regimes are more common in certain geographical regions, such as southern Europe (Greve et al., 2014), a humid climate may become transitional for periods without rainfall, and a dry climate may become transitional for wet periods (Zampieri et al., 2009; Dirmeyer et al., 2009; Seneviratne et al., 2010; Halder and Dirmeyer, 2017). If the drying continues until the wilting point (which is also land cover-dependent), the evapotranspiration seizes. This characterises the dry regime. Now, the feedback stops, because there is no more water left to regulate the evapotranspiration (Seneviratne et al., 2010; Gallego-Elvira et al., 2016). The transitional and the dry regimes are
termed soil moisture-limited. Because the processes controlling evapotranspiration are so different between an energy-limited and a soil moisture-limited regime, a correct process understanding depends on a correct characterisation of the regime.

A common way of diagnosing the soil moisture–temperature feedback is to recognise that the feedback pathway consists of two segments: the terrestrial leg and the atmospheric leg (Dirmeyer et al., 2006, 2014). The terrestrial leg is characterised as the correlation between soil moisture, $SM$, and latent heat, $LE$, whereas the atmospheric leg is characterised as the correlation between $LE$ and air temperature, $T$ (details are given in Section 4.3 and Paper IV). If both legs are present, it suggests that indeed, the land surface is contributing to the detected atmospheric change (Orlowsky and Seneviratne, 2010). It is worth noting that a correlation does not imply causality. However, it is a commonly used metric (Dirmeyer et al., 2006, 2014; Halder and Dirmeyer, 2017) that points in the direction of a feedback.

**Attribution studies** In this paragraph, we concentrate on drivers of spring warming, as detected for Norway by Førland et al. (2016), and summer warming, as detected for southern and central Europe by Klein Tank et al. (2005) and EEA (2008). The snow albedo feedback is widely studied as a driver of spring warming (see e.g. Chapin et al., 2005; Hall and Qu, 2006; Peng et al., 2013). Spring snow cover reductions have accelerated over the past decades, largely driven by increasing temperatures (Brown and Robinson, 2011; Derksen and Brown, 2012), enhanced by the snow albedo feedback (Groisman et al., 1994a; Dery and Brown, 2007).

Much research has dealt with the relationship between summer warming and dry soil moisture conditions in southern and central Europe, which are considered hot-spots of land–atmosphere coupling (Seneviratne et al., 2006). The soil moisture–temperature feedback has been documented to drive summer warming in southern and central Europe (see e.g. Seneviratne et al., 2006; Zampieri et al., 2009; Orlowsky and Seneviratne, 2010; Miralles et al., 2014), especially during heat waves (Schar et al., 2004; Zaitchik et al., 2006; Vautard et al., 2013; Miralles et al., 2014; Whan et al., 2015). Summer warming has also been explained with drier than normal conditions in preceding seasons, such as a soil moisture anomaly in spring as documented by Whan et al. (2015) or earlier snowmelt due to higher temperatures, as suggested by Wilson et al. (2010). The soil moisture–temperature feedback has received little attention at northern latitudes, where the climate is usually moist.
Figure 2.2: a) The snow albedo feedback. An initial warming leads to reduced snow cover, $S$, and reduced albedo, $\alpha$, that allows more shortwave radiation to be absorbed in the ground. The result is an increase in the net radiation, $R_n$, that leaves more energy available for latent heat, $LE$ and sensible heat, $SH$, ultimately increasing the air temperature, $T$. b) Soil moisture–temperature coupling. In energy-limited regimes, the soil moisture, $SM$, does not control the latent heat, rather, when the latent heat increases, it depletes the soil moisture storage. An initial warming that reduces the soil moisture below a critical value, $SM_{crit}$, shifts the regime into a soil moisture-limited regime. Here, the soil moisture limits latent heat (positive correlation; the terrestrial leg), and the warming may be enhanced by the reduced evaporative cooling (negative correlation; the atmospheric leg). c) Schematic illustration of the Budyko framework. Above $SM_{crit}$, latent heat is not limited by the soil moisture. Below $SM_{crit}$, latent heat becomes limited by soil moisture, protentially triggering a soil moisture–temperature feedback.
Chapter 3

Data

Figure 1.1 gives an overview of the data and methods used in Papers I–IV. Trends were detected for the gridded climate dataset WFDEI in Paper I (air temperature and precipitation) and Paper II (air temperature only), and for a high-resolution gridded (1×1 km) reference dataset for Norway, SeNorge, in Paper III. Point observations as well as SeNorge data were used in the model setup and validation in Paper IV, whereas ERA-Interim reanalysis data were used to force the model. A classification of synoptic types, SynopVis Grosswetterlagen (SVG), was used for statistical attribution in Papers I and II.

3.1 Gridded hydroclimatic datasets

3.1.1 The WFDEI dataset

Temperature and precipitation from the bias-corrected reanalysis dataset, i.e., the Watch Forcing Data ERA-Interim (WFDEI; Weedon et al., 2014) were used for trend detection in Paper I and II. These gridded (0.5°×0.5°), daily historical data are available at 3-hourly resolution, globally (except over oceans) (ftp://rfdata:forceDATA@ftp.iiasa.ac.at). We used daily averages for Europe for the recent past (1979–2009 in Paper I and 1981–2010 in Paper II).

The variables making up WFDEI originate from the ERA-Interim reanalysis (Dee et al., 2011). The temperature has been elevation corrected using the environmental lapse rate and bias-corrected against gridded monthly observations from the Climate Research Unit, University of East Anglia (CRU TS 3.1; Harris et al., 2014). The precipitation has been corrected for undercatch and bias-corrected against Global Precipitation Climatology Centre (GPCC) data. It is worth noting that WFDEI variables
are spatially consistent after bias-correction (Weedon et al., 2014), which is necessary when linking local variables to the large-scale atmospheric situation.

### 3.1.2 The seNorge dataset

We used a high-resolution dataset for Norway, the 1×1 km SeNorge dataset (Engeset et al., 2004; Tveito and Roald, 2005; Saloranta, 2012, senorge.no). SeNorge version 1.1 contains gridded hydroclimatic data at a daily time step from 1957–today, covering mainland Norway (Engeset et al., 2004).

Interpolated temperature and precipitation are produced by the Norwegian Meteorological Institute (Tveito and Roald, 2005). Interpolation is performed after subtracting a reference level of the variable, which is created by regressing physical characteristics such as latitude, elevation, distance from the coast etc. After interpolation, the reference level is added back to obtain the original physical characteristics. Monthly lapse rates are used to adjust temperature and precipitation to the terrain. Precipitation measurements are corrected for undercatch before gridding.

Hydrological variables are modelled with the GWB model, developed by the Norwegian Water Resources and Energy Directorate (NVE) (Beldring et al., 2003). The model core in GWB is the commonly used HBV model (Bergström, 1995; Sælthun, 1996), forced with the interpolated temperature and precipitation. GWB is a conceptual hydrological model calibrated against runoff data for boreal conditions (shallow soils, short roots etc). Snow variables are calculated separately in a snowmap model (Saloranta, 2012).

In SeNorge, each grid cell may contain lakes, glaciers and up to two varying land use classes (“open land”, “forest”, “alpine”, “heather”, or “bedrock”). Evapotranspiration for one grid cell is modelled as a weighted average of evapotranspiration calculated for the four land cover classes, including actual evaporation and interception (over vegetation), lake evaporation (over lakes). There is no evapotranspiration over glaciers and snow (Sælthun, 1996; Beldring et al., 2003). The actual evaporation is limited by soil moisture below a certain threshold (depending on the land cover class), and follows the potential evapotranspiration above the threshold. Potential evapotranspiration is a function of temperature and monthly correction factors. The correction factors are highest in June and lowest in November–March (Engeland et al., 2004).

Despite being a state-of-the-art, high-resolution dataset, SeNorge is not as homogenous as it may seem at first. For instance, because most measuring stations are located at low elevations and close to settlements, the SeNorge data are most un-
certain at high elevations and remote places. The number of stations in the station network has changed over time, meaning that the uncertainty is not homogenous in time. The inconsistent number of stations may influence trend calculations.

3.2 Point observations

Station observations of snow, soil moisture, soil temperature, and air temperature were provided by NVE and were used to validate the WRF–Noah-MP model for South Norway in Paper IV. As summarised in Table 3.1, snow water equivalent, SWE and snow depth, SD, were retrieved from 12 and 11 stations, respectively. Soil moisture and soil temperature were retrieved from four stations. These 23 snow stations and four soil moisture/soil temperature stations display a range of soil and vegetation types, from lowland sites with meadows (e.g., Ås), inland sites with forests (e.g., Brunkollen) and mountain sites with birch forest (e.g., Groset) (Colleuille, 2000). A map of the stations is shown in Figure 3.1.

SWE is measured using a snow pillow, i.e., a flexible, antifreeze-filled container measuring the weight of the overlying snow (Ree et al., 2011). Snow depth is measured by timing a reflected ultrasound pulse. Soil moisture is measured in six segments in the vertical, at -10 cm, -20 cm, -30 cm, -40 cm, -60 cm and -100 cm. Soil moisture is measured with Time Domain Reflectometry (TDR). Soil temperature is measured in nine segments in the vertical, 15 cm apart. Air temperature at 2 m was also provided for the stations measuring soil temperature and soil moisture (Abrahamsvoll, Kise, Groset and Ås), but due to a bias in the temperature at Groset and Ås, we used data from the Norwegian Meteorological Institute at Møsstrond (close to Groset) and Ås (in Ås).

3.3 A catalogue of synoptic types: SynopVis Grosswetterlagen (SVG)

Classifications of circulation indices aim to characterise the atmospheric circulation, based on the location of pressure patterns and storm tracks, as one categorical variable (e.g., Huth et al., 2008; Cahynová and Huth, 2016; Fleig et al., 2015; Huth et al., 2016). Such classifications are often used to separate trends into circulation changes and within-type changes (Beck et al., 2007; Cahynová and Huth, 2009, 2010; Küttel et al., 2011). In Papers I and II, we used a daily catalogue of synoptic types (STs) for
Table 3.1: Point measurements used for validation in Paper IV. All data were available from the Norwegian Water Resources and Energy Directorate, except air temperature from Møsstrond, available from the Norwegian Meteorological Institute.

<table>
<thead>
<tr>
<th>Var.</th>
<th>Station name</th>
<th>Municipality</th>
<th>[masl]</th>
<th>East [°]</th>
<th>North [°]</th>
<th>Station no.</th>
</tr>
</thead>
<tbody>
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<td>Hol</td>
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<td>8</td>
<td>60.68</td>
<td>012.142.0</td>
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<tr>
<td></td>
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<td>Bærum</td>
<td>370</td>
<td>10.55</td>
<td>59.97</td>
<td>008.005.0</td>
</tr>
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<td>Follo</td>
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<td>9.99</td>
<td>62.06</td>
<td>002.373.0</td>
</tr>
<tr>
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<td>8.31</td>
<td>59.83</td>
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</tr>
<tr>
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<td>8.34</td>
<td>59.17</td>
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<td>59.85</td>
<td>MET station</td>
</tr>
</tbody>
</table>
Figure 3.1: Location of snow stations (measuring SWE) in blue and soil moisture stations in brown. Note that Groset is both a snow station and a soil moisture station.
Europe, the SynopVis Grosswetterlagen (SVG; provided by P. James, 2013; personal communication), based on James (2007). SVG is derived from mean sea level pressure, precipitable water, 500−1000 hPa thickness, and 500 hPa geopotential height, all taken from the Twentieth Century Global Reanalysis (Compo et al., 2011) and supplemented by NCEP/NCAR reanalyses (Kalnay et al., 1996) for the period after 2010 (P. James, 2016; personal communication), see details in Paper II.

**Representativity of SVG for Europe** The separation of trends into circulation changes and within-type changes was conducted under the assumption that SVG is able to discriminate between STs. In other words, an ST classification must be able to separate warm from cold STs. If they are mixed, a possible warming signal related to increasing warm STs would not be visible. This implies that results showing within-type changes may not explain a change in the properties of STs, but rather deficiencies related to deriving the ST classification, see Beck et al. (2007) for a discussion of possible factors giving rise to within-type changes. See also Sections 2.2 and 4.2.

Figure 3.2 shows the discriminative capability of SVG. The Shannon Entropy used to measure SVG’s ability to separate warm STs from cold STs (described in James, 2007) is calculated using five bins of temperature data (“very cold”, “cold”, etc. for each grid cell) and assessing whether the temperatures for each ST are evenly distributed across the bins or are biased towards cold or warm bins. The discriminative capability drops with distance away from central Europe, which is the region that the original Hess-Brezowsky Grosswetterlagen was defined for (P. James, 2014; personal communication). Consequently, the SVG classification is most robust for central parts of Europe. In South Norway, the discriminative capability is lower than for, for instance, Germany but we have assumed it to be sufficiently high to separate out circulation changes from within-type changes. In northern Norway, however, SVG is less reliable. Note also that the discriminative capability of precipitation is lower than that for temperature.

**Frequencies and patterns of SVG synoptic types** The frequency distribution of SVG synoptic types for 1948–2017 is shown in Figure 3.3a. Although this time period exceeds the one used in this thesis, it gives an impression of the relative frequencies of the different STs for each month. These frequencies vary in time, as shown in Figure 3.3b–d) for the period 1979–2010, covering the full period used in both Paper I (1979–2009) and Paper II (1981–2010). Figure 3.3 shows smoothed values for five flow directions: ”westerly” (W and SW; STs 1–6), ”northerly” (N and NW; STs 7–8, 12–17), ”central” (C; STs 9–11), ”easterly” (E and NE; STs 18–23), and ”southerly” (S and SE;
STs 24–29). The ST numbering follows the rows in Figure 3.4 (and in James, 2007), e.g., the synoptic type in the first row (Wa) is number 1. Trends in the frequency of STs are calculated using linear regression, and significance is calculated using the t-test.

When considering the whole year (Figure 3.3b), westerly STs increased significantly, whereas no other flow directions showed significant trends. Identical STs may exhibit different synoptic features in summer and winter (James, 2007). Therefore, summer trends (Figure 3.3c) and winter trends (Figure 3.3d) are shown separately. For summer, westerly types increased at the expense of northerly types. For winter, no significant trends could be detected over the period. Note that Figure 3.3 does not state whether an ST is warm or cold, which requires additional information on local temperature.

Climatological composite means for the 29 STs are shown in Figure 3.4.
Figure 3.2: Quantile entropy reduction above noise for maximum temperature (top) and precipitation (bottom). Because some STs are colder than average and other STs are warmer than average, the distribution of quantiles for each grid cell, month and ST will be biased. The larger the bias, the higher the entropy reduction (a measure of anomalies from an unbiased ST). Even randomly generated ST classifications have an entropy reduction by chance, termed "noise". The entropy reduction above noise means that the noise is subtracted from the entropy reduction. Figures provided by P. James, 20.11.2014. Reprinted with permission from the author.
Figure 3.3: Frequencies of synoptic types sorted by flow direction as a) percentages for the period 1948–2017 (data provided by P. James, 11.10.2017), and as a 3-year moving average for 1979–2010 comprising b) all days in the year c) summer days (15 April–14 October) and d) winter days (15 October–14 April), as a percentage of the year. Significant trends at the 0.05 level are shown with a linear trendline.
Figure 3.4: Climatological composite means for the 29 STs. Left: mean sea level pressure (contours, in hPa) and precipitable water (colours, in mm). Right: geopotential height at 500 hPa (in decametres, dam) and 500–1000 hPa thickness (in dam). Figures provided by P. James, 12.10.2017. Reprinted with permission from the author.
Figure 3.5: Continued from previous page.
Figure 3.7: Continued from previous page.
Figure 3.8: Continued from previous page.
Figure 3.9: Continued from previous page.
Figure 3.10: Continued from previous page.
Figure 3.11: Continued from previous page.
Figure 3.12: Continued from previous page.
Chapter 4

Methods

This section summarises trend detection methods, statistical attribution methods, which include the method of hypothetical trends, and a novel probabilistic approach expanding on this method, and land surface modelling performed in the model diagnostic study.

4.1 The Theil-Sen slope and Mann–Kendall trend test

Air temperature trends were calculated using the Sen slope (also called the Theil-Sen estimator or Kendall–Theil Robust Line; [Theil, 1950; Dery et al., 2005]). The trend magnitude $m$ is the median of the linear slopes between all pairs of values:

$$m = \text{median} \left(\frac{y_i - y_j}{t_i - t_j}\right) = \text{median}(m_k)$$  \hspace{1cm} (4.1)

where $y_i - y_j$ is the temperature difference between all possible values in the time series, $t_i - t_j$ is the corresponding time difference between each pair of values in the time series, and $m_k$ is a vector consisting of all possible linear slopes. A more thorough description is given in Paper II, Section 3.1. The Sen slope was selected, rather than the linear regression trend, because it is less influenced by outliers (Stahl et al., 2012). However, it is still sensitive to the time period chosen (Hisdal et al., 2001).

The Mann–Kendall trend test (Sneyers, 1990) was used in conjunction with the Sen slope to assess whether the estimated observed trend is significantly different from zero. When testing the hypothesis that no monotonic trend is present in WFDEI temperatures, a standard significance level of 0.05 was assumed (two-sided test). The
Mann–Kendall test statistic, $Z_{MK}$, is defined as (Dery et al., 2005):

$$Z_{MK} = \begin{cases} 
\frac{S-1}{\sigma_S}, & S > 0 \\
0, & S = 0 \\
\frac{S+1}{\sigma_S}, & S < 0 
\end{cases}$$ (4.2)

where $S$ is the test statistic and $\sigma_S$ is the variance of $S$. To calculate $S$, the signs of all possible trendlines $m_k$ are summed:

$$S = \frac{n(n-1)/2}{\sum_{k=1}^{n} \text{sgn}(m_k)}$$ (4.3)

where $\text{sgn}(m_k)$ is 1, 0 or -1, depending on the sign of $m_k$.

### 4.2 Statistical attribution

#### 4.2.1 Terminology

Before introducing the statistical attribution, a note on terminology is required. Paper I used terminology from Cahynová and Huth (2010) but this was not precise enough for our use of the SVG classification of synoptic types (see referee comments to Fleig et al. (2015)). Therefore, Paper II used terminology from Fleig et al. (2015), who also used the SVG classification. Table 4.1 lists the terminology used in Papers I and II, and in this thesis. For instance, the term “circulation type (CT)” was changed into “synoptic type (ST)” in Paper II to reflect that SVG is classified based on not only sea level pressure and geopotential height, but also precipitable water and the 500–1000 hPa thickness. Further, the term “hypothetical trend” used in Paper I is synonymous with “synoptic-circulation-induced trend” used in Paper II. For simplicity, we use the terms “circulation-induced trend” throughout this thesis, also for the synoptic-circulation-induced trends that result from the probabilistic approach in Paper II. Similarly, atmospheric circulation is in Paper II referred to as “synoptic circulation” to emphasise the methodology behind SVG, but for simplicity, we use “atmospheric circulation” throughout this thesis.
Table 4.1: Terminology in Papers I and II. Text in bold indicate the wording used in this thesis, when two terms are used synonymously.

<table>
<thead>
<tr>
<th>Term used</th>
<th>Definition</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Paper I Circulation type (CT)</td>
<td>Classified based on pressure variables</td>
<td></td>
</tr>
<tr>
<td>Paper II Synoptic type (ST)</td>
<td>Classified from more variables than pressure</td>
<td></td>
</tr>
<tr>
<td>Paper I Observed trend, $t_{obs}$</td>
<td>Climatic trend based on the WFDEI dataset</td>
<td></td>
</tr>
<tr>
<td>Paper II WFDEI trend, $m_{WFDEI}$</td>
<td>The same as $t_{obs}$</td>
<td></td>
</tr>
<tr>
<td>Thesis Estimated observed trend</td>
<td>General term, reflecting that reanalyses are not observations</td>
<td></td>
</tr>
<tr>
<td>Paper I and II Circulation changes</td>
<td>Changes in the frequency of STs that induce a trend</td>
<td>(Beck et al., 2007)</td>
</tr>
<tr>
<td>Paper I and II Within-type changes</td>
<td>Changes within the STs that induce a trend</td>
<td>(Beck et al., 2007)</td>
</tr>
<tr>
<td>Paper I Method of hypothetical trends</td>
<td>Method to calculate the influence of circulation changes on a trend</td>
<td>(Leathers and Ellis, 1996) ; Cahynová and Huth, 2009</td>
</tr>
<tr>
<td>Paper II Probabilistic approach</td>
<td>Expanded hypothetical method</td>
<td>(Nilsen et al., 2017b)</td>
</tr>
<tr>
<td>Paper I Hypothetical trend, $t_{circ}$</td>
<td>Trend resulting from changed frequency of STs</td>
<td>(Huth, 2001)</td>
</tr>
<tr>
<td>Thesis Circulation-induced trend, $t_{circ}$</td>
<td>The same as hypothetical trend</td>
<td>(Fleig et al., 2015) ; Cahynová and Huth, 2016</td>
</tr>
<tr>
<td>Paper II Synoptic-circulation-induced trend</td>
<td>Hypothetical trend resulting from the probabilistic approach</td>
<td>(Nilsen et al., 2017b)</td>
</tr>
<tr>
<td>Paper II Circulation-induced component of the trend, $m_{SC}$</td>
<td>Median of the distribution of synoptic-circulation-induced trends</td>
<td>(Nilsen et al., 2017b)</td>
</tr>
<tr>
<td>Paper I and II Within-type trend</td>
<td>Component of the trend that is independen of circulation changes</td>
<td></td>
</tr>
</tbody>
</table>
4.2.2 Method of hypothetical trends

The method of hypothetical trends (Leathers and Ellis, 1996; Huth, 2001; Cahynová and Huth, 2009) (also called method of circulation-induced trends, Cahynová and Huth 2016) provides a way to separate circulation changes from within-type changes. A new, hypothetical time series is constructed by replacing all observations in a month and ST with the long-term mean of all days having the same month and ST. The trend of the hypothetical time series is called the hypothetical trend. To assess the influence of circulation changes on the estimated observed trend, the hypothetical trend, $t_{circ}$, is divided by the estimated observed trend, $t_{obs}$, to obtain the trend ratio, $r_{circ}$. If the trend ratio is close to 1, the hypothetical trend can explain the observed trend well, and the observed trend can be attributed to circulation changes.

For trend ratios close to 1 (in Paper I, 0.5–1.5), the estimated observed trend cannot be explained by circulation changes alone, and the estimated observed trend is partly attributable to within-type changes. Small trend ratios (below 0.5) indicate within-type changes. In that case, the temperature trend could not be explained by circulation changes alone. Trend ratios exceeding 1.5 were not considered because they often stem from dividing by small estimated observed trends (Fleig et al., 2015).

4.2.3 Probabilistic approach

The method of hypothetical trends and trend ratios (Section 4.2.2) does not contain any way of assigning a statistical significance to the circulation changes, nor any way of avoiding high trend ratios when dividing by a small estimated observed trend. Paper II contains an expansion of the method of hypothetical trends, denoted the probabilistic approach for attributing temperature changes to synoptic type frequency, or probabilistic approach, for short.

The probabilistic approach developed in Paper II provides a statistical test of whether the estimated observed trend can be explained by changes in circulation alone. First, we derived the empirical distribution of circulation-induced trends by repeated historical resampling. Building on the method of hypothetical trends by Cahynová and Huth (2009), we generated a large set of circulation-induced trends by resampling 10,000 times. Second, for the resulting distribution of circulation-induced trends, we applied a Monte Carlo test to test the likelihood that the estimated observed trend could be randomly generated from the following empirical distribution.
\( H_0: \) The estimated observed trend can be explained by changes in atmospheric circulation alone.

\( H_a: \) The estimated observed trend cannot be explained by changes in atmospheric circulation alone.

The closer the estimated observed trend is to the median of the circulation-induced trends, the more likely it is that the estimated observed trend is caused by circulation changes alone. If the estimated observed trend is significantly different from the median, it is not likely caused by circulation changes. Two requirements must be fulfilled before the temperature trend can be attributed to within-type changes: i) the estimated observed temperature trend must be statistically significant, and ii) the alternative hypothesis must be valid, see Figure 4.1 panel (b). The probabilistic approach does not quantify the magnitude of within-type trends, but tests whether within-type changes contribute to the estimated observed temperature trend or not.

\subsection*{4.3 Assessment of the physical drivers of temperature trends}

The climatological cause of trends was assessed in a detailed study of concurrent changes in snow indices and temperature trends for Norway (focusing on the snow albedo feedback; Paper III) and through a model diagnostic study (Paper IV). This last paper explored both the snow albedo feedback and the soil moisture–temperature feedback, their sensitivity to a longer snow-free season and anomalously dry weather situations.

Many different methods of diagnosing land–atmosphere coupling exist, ranging from correlations between the land state and atmospheric state (e.g. Dirmeyer et al. 2009; Halder and Dirmeyer 2017), to running long-term climatology models (Miralles et al. 2014; Walton et al. 2017), and to running complex ensemble models comparing one fully coupled soil moisture (or snow) ensemble with a prescribed soil moisture (or snow) ensemble (Koster et al. 2004; Seneviratne et al. 2006, 2010). Many of these methods are costly in terms of computing power; running an ensemble, or running a 30-year period for climatology would increase the computation time in Paper IV by an order of magnitude. The snow albedo feedback was assessed through correlations in Paper III (see Figure 2.2h) and through comparing temperatures and radiation fluxes across different model runs in Paper IV.
Figure 4.1: Figure 1 from Paper II, explaining the probabilistic approach. From Nilsen et al. (2017b).
For the soil moisture–temperature feedback, we chose the two-legged metric, consisting of correlations between the soil moisture and latent heat, $\rho(SM, LE)$ and between latent heat and air temperature, $\rho(LE, T)$. As depicted in Figure 2.2b, $\rho(SM, LE)$ reveals whether the soil moisture controls the latent heat flux, or the other way around. Coupling leading to a positive feedback occurs only if the soil moisture is low enough to control the latent heat, i.e., a positive $\rho(SM, LE)$ (see ”Terrestrial leg” in Figure 2.2b). If the temperature rises as a result of reduced evaporative cooling, a negative correlation between latent heat and air temperature, $\rho(LE, T)$ indicates a soil moisture limitation on evapotranspiration (Seneviratne et al., 2006; Dirmeyer et al., 2009; Lorenz et al., 2012) (see ”Atmospheric leg” in Figure 2.2b). In that case, the reduced latent heat may accelerate warming. For land–atmosphere coupling to be present, both the terrestrial and atmospheric legs must be in place (Dirmeyer, 2011; Dirmeyer et al., 2014; Halder and Dirmeyer, 2017).

### 4.3.1 Model diagnostic study

The regional climate model Weather Research and Forecasting (WRF) (Skamarock and Klemp, 2008), coupled to the land surface model Noah-MP (Niu et al., 2011) was used to explore the physical drivers of temperature trends (Paper IV). Details of the model setup are given in Table 4.2.

Land surface models are employed to handle the land surface component of regional climate models. Noah-MP (Niu et al., 2011), a multi-parameterisation development of the Noah land surface model, was created to allow for a large physical ensemble of model simulations within the same model. Noah-MP and its predecessors Noah-LSM/Noah-UA have been applied for different applications in central and northern Europe (Schwitalla et al., 2011; Greve et al., 2013; Gayler et al., 2014; Mayer et al., 2015; Rydsaa et al., 2016; Aas et al., 2017; Erlandsen et al., 2017).

In Paper IV, we hypothesised that the warming detected for April was enhanced by the snow albedo feedback, and that the warming detected for July–September was enhanced by the soil moisture–temperature feedback (provided sufficiently dry conditions). The warm summer of 2014 was a plausible candidate for dry conditions and was chosen for control simulations with a coupled WRF–Noah-MP model. To simulate drier summer conditions, we i) increased the length of the snow-free season, and ii) introduced boundary forcing from a warm and dry summer (2006). In total, six model runs were performed (see Figure 4.2e): The first three runs differed only in their initial ground conditions (snow-poor, control and snow-rich) and used boundary forcing from the warm summer of 2014. The next three runs kept the same initial
Table 4.2: WRF–Noah-MP model configuration for Paper IV. From (Nilsen et al., 2017a) (Table 2 in Paper IV).

<table>
<thead>
<tr>
<th>Reference</th>
<th>Atmospheric model</th>
<th>WRF v 3.7.1</th>
<th>Skamarock et al., 2008</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Land surface</td>
<td>Noah-MP with default options</td>
<td>Niu et al., 2011</td>
</tr>
<tr>
<td></td>
<td>Microphysics</td>
<td>WRF Single-Moment 3-class (WSM3)</td>
<td>Hong et al., 2004</td>
</tr>
<tr>
<td></td>
<td>Boundary layer physics</td>
<td>Mellor-Yamada-Janjić (MYJ)</td>
<td>Janjić, 1994</td>
</tr>
<tr>
<td></td>
<td>Surface layer</td>
<td>Eta surface layer scheme</td>
<td>Janjić, 2002</td>
</tr>
<tr>
<td></td>
<td>Radiation (LW &amp; SW)</td>
<td>RRTMG</td>
<td>Iacono et al., 2008</td>
</tr>
<tr>
<td></td>
<td>Cumulus</td>
<td>Kain-Fritsch in d01, resolved in d02</td>
<td>Kain, 2004</td>
</tr>
<tr>
<td></td>
<td>Simulation period</td>
<td>Experiments: 12 May–30 September</td>
<td>Dee et al., 2011</td>
</tr>
<tr>
<td></td>
<td>Spin-up period</td>
<td>10.5–12.5 months (depending on run)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Land cover</td>
<td>20-class MODIS 15 arc seconds</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Timestep</td>
<td>45 s</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Nesting</td>
<td>Two-way nesting</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Resolution in d01</td>
<td>15 km (100×100 cells)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Resolution in d02</td>
<td>3 km (155×185 cells)</td>
<td></td>
</tr>
</tbody>
</table>

conditions as before and used boundary forcing from the warm and dry summer of 2006.
Figure 4.2: Figure 2 from Paper IV. a) Model domains, b–d) initial SWE at the start of the model run at b) 12 May 2014 (CTR), c) 11 July 2014 (snow-poor, two months earlier melting), and d) 12 May 1995 (snow-poor). e) Schematic of the different model runs. Note that in addition to SWE, we swapped snow depth, snow albedo, soil moisture, skin temperature, initial temperature, soil category fraction at the bottom and top, and an indicator function for snow. From (Nilsen et al., 2017a).
Chapter 5

Findings

This section summarises the main results of the thesis, based on the four papers. A schematic of the papers is shown in Figure 1.1 and the results are further discussed in Section 6 (General discussion). Figure numbers refer to this thesis, unless stated otherwise.

5.1 Paper I: Recent trends in monthly temperature and precipitation patterns in Europe

Climate change is characterised by large spatial and temporal variations. In this study, we investigated the spatial pattern of precipitation and temperature trends in Europe for each calendar month. For this purpose, we used gridded precipitation and temperature data from WFDEI (Weedon et al., 2014). The period 1979–2009 was chosen because it is a period commonly used for climate trends, and because of the availability of WFDEI data at time of the study. Further, a catalogue of synoptic types, SVG by P. James, was used in the method of hypothetical trends (Section 4.2.2) to attribute the WFDEI trends either to changing frequency of synoptic types or within-type changes. Note also that some of the terminology changed from Paper I to Paper II, as summarised in Table 4.1 and Section 4.2.1.

Main results  Results were shown as “observed” WFDEI trends, $t_{\text{obs}}$, hypothetical trends, $t_{\text{circ}}$, and trend ratios, $r_{\text{circ}}$ ($r_{\text{circ}} = t_{\text{circ}}/t_{\text{obs}}$) for the annual scale and for February in the paper. A complete figure for all months is shown in Figure 5.1, for temperature. This figure shows winter/spring and summer/autumn on two different pages.
Figure 5.1: Complete Figure 2 (Paper I) of temperature, for winter and spring months (1979–2009). Left: WFDEI trend, $t_{obs}$, middle: circulation-induced trend, $t_{circ}$, right: trend ratio $r_{circ}$. Modified from [Nilsen et al., 2014]. Continued on the next page.
Figure 5.1: (contd.) Complete Figure 2 (Paper I) of temperature, for summer and autumn months (1979–2009). Left: WFDEI trend, $t_{\text{obs}}$, middle: circulation-induced trend, $t_{\text{circ}}$, right: trend ratio $r_{\text{circ}}$. Modified from (Nilsen et al., 2014).
A dominant warming signal was observed in winter, spring and summer; during autumn, weak cooling trends were observed in southern Europe (Figure 5.1, upper row). Strong warming, up to 2.5 °C/decade, was seen in North Scandinavia in December and January (0.8 °C/decade on the annual scale). The summer warming occurred concurrently with negative streamflow trends for central Europe (Stahl et al., 2010).

Circulation-induced trends generally resembled observed trends, but displayed a smoother spatial pattern and weaker trend magnitudes (Figure 5.1, middle row). The trend ratio highlighted regions in almost all months in which WFDEI trends could be explained by the circulation changes. The most wide-spread signal of circulation changes was found in February followed by December, September, May and April (Figure 5.1 marked as “Circ” in the bottom row). This strong influence of circulation changes in winter is consistent with an increased NAO index until the 1990s. Trend ratios less than 0.5 suggest that the trend was influenced by other factors than circulation changes (Figure 5.1 marked as ”NoCirc” in the bottom row). This was found for most other months than the ones listed above. Only a few regions are marked “NoTrend”, that is, where $t_{obs}$ was so small that the resulting trend ratio exceeded 1.5.

Similar figures for precipitation (not shown) were more patchy than temperature trends. This is partly because of the more local nature of precipitation than temperature, especially in summer when convective storms dominate. Circulation-induced precipitation trends could also be influenced by the performance of the SVG, see Figure 3.2, bottom. Based on this figure, we decided not to use precipitation in the further studies.

Strong warming was detected for Scandinavia (exceeding 2 °C over the 31-year period, or 0.8 °C/decade, on the annual scale, and higher for December and January). This strong warming motivated a review of previous regional temperature trends in Europe, shown in Table 2.1. All trends are sensitive to the time period chosen, thus, other periods than the one chosen in Paper I (1979–2009) would result in different results. Comparing Figure 5.1 to Figure 5.2 shows the effect of changing the analysis period from 1979–2009 to 1981–2010. The winter 2009/2010 was very cold in Europe due to weakened westerly air flow, in fact, the NAO index was record low (Osborn, 2011). A broader picture of trends could be explored by analysing running 30-year trends, similar to Hannaford et al. (2013).

Unrealistically high trend ratios may result when the observed trend (in the denominator) is close to zero. In Paper I, we solved this by adding cut-off values of 0.5 and 1.5, inspired by Cahynová and Huth (2009) and Fleig et al. (2015). Further, the trend ratio fails to capture changes when both circulation changes and within-type changes play a role but in opposite directions. These two arguments motivated us to
extend the method of hypothetical trends in Paper II.

5.2 Paper II: A probabilistic approach for attributing temperature changes to synoptic type frequency

The probabilistic approach developed in Paper II provides an assessment of the likelihood that the estimated observed trend is attributable to circulation changes alone and allows assigning a statistical significance to the circulation changes (described in Section 4.2). The objective of Paper II was to separate the cause of temperature change into within-type changes and circulation changes (similar to Paper I), however, by applying the probabilistic approach to temperature changes in Europe (WFDEI). The period 1981–2010 was chosen so the results could be comparable with other studies using the World Meteorological Organisation’s reference period.

Main results

Annually, WFDEI temperature trends calculated using Sen slopes were positive for nearly all of Europe. On the monthly scale, significant temperature trends were found in large parts of Europe in April–August, and in November (Figure 5.2 upper row). The circulation-induced component of the temperature trend (median of the distribution; middle row) could explain warming trends for western Europe in May (and in the Balkans in February and Belarus in August). Here, both the WFDEI trends and circulation-induced trends were significant. Regions where the within-type changes played a role are presented in the bottom row of Figure 5.2. In the bottom row, dots are the same as in the upper row, denoting where the WFDEI trend is significant. Dots on a white background in the bottom row mean that the temperature trend can be explained by circulation changes; dots on colour mean that within-type changes are responsible for at least a part of the temperature trend. Figure 4.1 shows the difference between dots on white (panel (a)) and dots on colour (panel (b)). Overall, within-type changes could partly explain the warming in April, June–August for large parts of Europe, and in November north of the Black Sea. For South Norway, April and July–September temperature trends could partly be explained by within-type changes. Regions where temperature changes could partly be attributed to within-type changes may potentially suggest the presence of feedbacks between the land surface and atmosphere. A closer look reveals within-type changes along the coast and inland in June and along the southern coast in September.

When comparing Figure 5.2 to Figure 5.1, similarities in the WFDEI temperature trends are evident, for instance in the strong winter warming, and the cooling in
Figure 5.2: Figure 3 from Paper II. Temperature trends for winter and spring (1981–2010) for a) estimated observed WFDEI trends. b) circulation-induced component of the WFDEI trend. c) Within-type changes (in colour). Regions shaded by dots mark significant trends. From (Nilsen et al., 2017b).
Figure 5.2: Figure 3 from Paper II. Temperature trends for summer and autumn (1981–2010) for a) estimated observed WFDEI trends. b) circulation-induced component of the WFDEI trend. c) Within-type changes (in colour). Regions shaded by dots mark significant trends. From (Nilsen et al., 2017b).
May, September and December. In January and February, Figure [5.2] displays strong cooling (in eastern Europe in January and in Scandinavia in February), whereas Figure [5.1] displays warming or no trend. Circulation-induced trends (middle row) differed in January. This example shows that trends are sensitive to a slight change in time period (1979–2009 vs 1981–2010), but agreed well for months other than January and February. Trends explained by circulation changes (compare the bottom row of Figure [5.1] with dots in the middle row of Figure [5.2]) aligned reasonably well, for instance for February (eastern Europe), May (France, Spain), July (eastern Europe) and September (Norway). The two bottom rows of Figure [5.2] and Figure [5.1] deviates to a large extent, however. This deviation reflects the different methods used in Paper I and Paper II (discussed further in Section 6.2). The circulation-induced warming May and to a smaller extent February is captured in both Figures, but Figure [5.1] showed circulation-induced trends in other months and regions as well. Last but not least, regions of within-type change differed greatly.

Regions where temperature changes could partly be attributed to within-type changes may potentially suggest the presence of feedbacks between the land surface and atmosphere. In Paper II, that was the case for South Norway in spring and summer.

5.3 Paper III: Five decades of warming: impacts on snow cover in Norway

Strong decreases in the snow season duration in spring have been linked to accelerated spring warming (Thackeray and Fletcher, 2016). Paper III (led by J. Rizzi) was motivated by the strong winter trends detected in Paper I. Further, in Paper II, April was highlighted as a month with significant warming that could not be attributed to changes in the synoptic-type frequency. The aim of Paper III was to detect changes in snow indices and climate variables in Norway and to assess the snow albedo feedback as a possible driver of these changes. In this study, we performed a detailed analysis of trends and changes in snow climatology using the dataset SeNorge ($1 \times 1$ km, covering mainland Norway) for different 30-year periods within 1961–2010. Trend analyses of monthly temperature were done for 1981–2010, the same period as in Paper II. Relationships between snow water equivalent (SWE), snow cover extent (SCE), temperature (T), and precipitation (P) were presented for Norway as a whole, and for four macro regions (North, South-East, South-West and Mountains) and 19 sub-regions.
Main results The snow cover extent declined in all four macro regions. The strongest reduction occurred in spring, particularly in April for low-lying regions in South Norway and in May, further inland and further north (see Figure 3 in Paper III). Between 1981–2010 and 1961–1990, SCE decreased by nearly 25 000 km$^2$ in April and more than 25 000 km$^2$ in May (8% of the total land area of mainland Norway). November and December experienced the second largest loss of SCE (about 15 000 km$^2$; Figure 5a in Paper III). The regions South-East and South-West experienced the most dramatic reduction in SCE, relative to their areas. The region Mountain experienced a decline in SCE, but an increase in SWE due to increased precipitation rates (Figure 5a and b, in spring).

Snow changes were accompanied by a decrease in the length of the snow season, particularly in spring. Correlation analyses showed that SCE was negatively correlated with temperature (significant in all months except summer), supporting our hypothesis that the snow albedo feedback enhances snowmelt in spring. A stronger difference in spring than in autumn was expected because both snow cover and sunlight is abundant in spring. Further, strong temperature trends were found in April across all of Norway, accompanied by strong SCE reductions in the same month. These findings support the hypothesis of an active snow albedo feedback, particularly in spring.

Further analyses of SeNorge temperature trends showed widespread warming in July, August and September, in addition to April (for the period 1981–2010; Figure 7 in Paper III), confirming trends detected in WFDEI from Paper II. The strongest warming was observed in North Norway in December, close to 3 °C/decade, however, only a part of these trends were significant. In winter, the interannual variability of trends was high (shown as a large inter-quartile range), which explained why the highest trends in North Norway were not significant in all grid cells. Boxplots of temperature trends (Sen slopes) for three 30-year periods revealed that trends were sensitive to the time period chosen (Figure 8 in Paper III). For April, the warming accelerated between 1961–1990, 1971–2000 and 1981–2010. August and September also displayed accelerated warming.

5.4 Paper IV: Diagnosing land–atmosphere coupling in a seasonally snow-covered region (South Norway)

The processes driving within-type changes detected in Paper II, supported by strong snow and temperature changes in Paper III, were further explored in the model dia-
gnostic study in Paper IV. South Norway was chosen as the study area for modelling based on a range of factors:

i) the presence of within-type changes in April and July–September,

ii) availability of a high-resolution dataset for Norway (SeNorge),

iii) because this region allowed testing both the snow albedo feedback and the soil moisture–temperature feedback in one study region,

iv) because previous studies have documented drying summer conditions in South Norway [Wilson et al., 2010; Stahl et al., 2012] and

v) the discriminative capability of the SVG classification was considered sufficiently high to give trust in the main results for South Norway.

Although North Norway displays strong within-type changes in many months (particularly in winter), this region was not considered for the model-diagnostic study, partly because of the low discriminative capability far away from central Europe (as discussed in Section 3.3). Because we are interested in the snow albedo feedback and soil moisture–temperature feedback, the location of within-type changes in April and summer were given more weight than the strong within-type signal in North Norway in winter (when the ground is expected to be fully snow-covered throughout the period and no snow albedo feedback is expected).

To our knowledge, the soil moisture–temperature feedback has not previously been studied for Norway, where evapotranspiration is generally energy limited rather than soil moisture limited. The experiments for the coupled model WRF–Noah-MP were therefore designed to simulate anomalously dry conditions to trigger soil moisture-limited conditions. The warm summer of 2014 and the warm and dry summer of 2006 were chosen for the purpose.

**Main results** Changes in the snow cover had direct effects on temperature, soil moisture, and evapotranspiration during the melting period, supporting that the snow albedo feedback enhanced the spring warming in the region influenced by snow changes and, for a large part of South Norway, even after snowmelt. For example, air temperatures over snow-free ground were about 5 °C higher than air temperatures over snow-covered ground (at station Kyrkjestølen).

The soil moisture–temperature feedback was diagnosed by correlating soil moisture and latent heat, $\rho(\text{SM, LE})$; i.e., the terrestrial leg of land–atmosphere coupling,
and by correlating latent heat and air temperature, $\rho(LE, T)$; i.e., the atmospheric leg (Dirmeyer, 2011). The positive soil moisture–temperature feedback acted in the Oslofjord region (along the southeastern coast of South Norway) during 2014 and 2006. Both the atmospheric leg, Figure 5.3a–f, and the terrestrial leg were present during 15 July–14 August, to a larger degree in 2006 than in 2014. Thus, a positive feedback on temperature was initiated by the warmer than average conditions in 2014 (and for 2006: warmer and drier than average conditions), and further enhanced by a soil moisture limitation on evapotranspiration. A longer snow-free season did not influence this soil moisture–temperature feedback, likely because the changes in the initial ground conditions were not large enough to induce a sufficiently low soil moisture content. Changing the atmospheric boundary forcing to the warm and dry 2006 summer had a larger effect on the strength and extent of the coupling leading to a positive feedback, than changing the initial conditions. Thus, the enhanced warming was caused by a lack of rainfall rather than a longer snow-free season.

To strengthen our confidence in the model results, correlations were calculated for the estimated observed dataset of interpolated temperatures, and modelled evapotranspiration and soil moisture deficits (SeNorge). SeNorge confirmed WRF–Noah-MP results, displaying enhanced warming in the Oslofjord region (Figure 5.3g–h). SeNorge also confirmed the more widespread pattern of the terrestrial leg than the atmospheric leg. However, differences in the extent of positive coupling between WRF–Noah-MP and SeNorge were seen. In particular, coupling was detected for the (north-)western coast in 2006 and the mountains in WRF–Noah-MP, but not in SeNorge. We proposed some improvements to the evapotranspiration estimations in Noah-MP and in SeNorge that could alleviate discrepancies between the two datasets in further studies, namely i) improper representation of soil depth/root depth and ii) improper parameterisations of for instance evapotranspiration.
Figure 5.3: Figure 7 from Paper IV. Atmospheric leg of land–atmosphere coupling (negative correlation, $\rho(LE, T)$ indicates coupling) for the period 15 July–14 August 2014 (left) and 2006 (right). Rows 1–3 show CTR, SP and SR, respectively, from the WRF-Noah-MP runs. SeNorge is shown in the last row. A red colour indicates coupling in all plots. From (Nilsen et al. 2017a).
Chapter 6

General discussion

Understanding circulation changes and land–atmosphere interactions are crucial for our understanding of what drives hydroclimatic change. Although the broader picture of climate trends on a global and annual scale is well understood, research efforts are now moving towards understanding the detailed mechanisms leading to temperature change at the regional and seasonal scale. This section discusses the main outcomes of this work, including uncertainties. Suggestions for further studies are integrated in the text.

6.1 Trend detection

6.1.1 Comparing detected trends across time periods and datasets

The trend results depend on choices and assumptions made, specifically, on the chosen dataset, method (e.g. [Cahynová and Huth] 2009), time period ([Hannaford et al.] 2013), and spatial and temporal aggregation. Robust results emerged, however, across different datasets and slightly different periods in Papers I, II and III. SeNorge confirmed the strong winter trends in northern Norway, and significant April, and July–September warming in South Norway (Figure 7 in Paper III). This similarity was expected because temperature datasets are well-constrained by observations and well-tested. SeNorge temperatures are interpolated from stations run by the Norwegian Meteorological Institute, whereas WFDEI is based on a reanalysis dataset (ERA-Interim), bias-corrected against CRU stations. Thus, some of the same stations have been used both in SeNorge and WFDEI, giving rise to the same result. However, the station network in SeNorge is finer than the one used by CRU, and has a much finer spatial resolution, which results in a more reliable dataset. Obtaining the same
results with different datasets strengthens our confidence in the results. Similarly, large patterns of trends in the same direction also strengthens our confidence in the results.

### 6.1.2 Sensitivity to time period and time scale

Changing the period of WFDEI trends from the period 1979–2009 to 1981–2010 yielded similar results, except for winter trends. Results in Figure 5.1 and Figure 5.2 mainly differed in January and February, when the interannual variability was high. This example shows that a small change at the beginning or end of the period (shifted two years and from 31 years to 30 years) was sufficient to influence the results, at least for certain months. The latest period included the cold winter 2009/2010 (Hurrell and Deser, 2009), which likely caused the cooling trends seen for the 1981–2010 period (in eastern Europe in January and in Scandinavia in February).

In Paper III (Figure 8), the periods were shifted by 10 years, showing a $>0.5$ °C higher trend for the period 1981–2010 than 1961–1990, for December and August. NAO was highlighted as the main cause of winter warming in Paper II, and the importance of NAO has been pointed out in many studies of circulation-induced trends in Europe for the period starting before $\approx 1980$ and ending after $\approx 2000$ (Cahynová and Huth 2010, (1961–1998), Cahynová and Huth 2016, (1961–2000), Fleig et al. 2015, (1963–2001), Kučerová et al. 2017, (1957–2002)). Given that NAO explains a large degree of the winter weather in Europe, the temperature trend is strongly related to the chosen time period (see Figure 2.1).

The choice of temporal aggregation also influences the results. Monthly trends, calculated in Papers I, II, and III may mask important sub-monthly signals. One example is the increased evapotranspiration related to leafing in spring, which may occur in the course of a few days or weeks. Still, a monthly time scale is finer than in most published trend studies and reveals strong trends that point towards specific physical processes influencing the warming.

A general recommendation for future studies is to use more than one dataset (STs and climate data), using more than one trend period, and possibly different trend methods and time resolutions. Using gridded data allows for spatially coherent patterns that increases our confidence in the results. Those trends that emerge across datasets, trend periods etc. would be considered (very) robust. This is normally outside of the scope of many studies, but it is possible given the increasing computing power and collaborative efforts in the scientific community today.
6.2 Statistical attribution of recent temperature trends

The role of atmospheric circulation on the local climate has been a topic of research for decades, reviewed by e.g. Huth et al. (2008, 2016). In particular, attribution of climatic trends into either circulation changes or within-type changes has received much attention (e.g. Beck et al., 2007; Cahynová, 2010; Küttel et al., 2011; Hertig et al., 2015; Cahynová and Huth, 2016). We developed a probabilistic approach to attribute the relative importance of circulation changes on climate trends (Paper II) to guide us towards regions of possible feedback mechanisms (regions of within-type changes). The following sub-section contains comments on differences between regions detected as circulation changes in Papers I and II. Sub-section 6.2.2 contains comments on differences between regions detected as within-type changes in Papers I and II.

6.2.1 Differences in circulation changes

Circulation changes marked in the bottom rows of Figure 5.1 and Figure 5.2 agreed reasonably well despite the different methods used in Papers I and II. Paper I used the simple trend ratio to compare the estimated observed trend to a circulation-induced trend that would have resulted if changes in the frequency of synoptic types governed the estimated observed trend. Paper II used a probabilistic approach to derive a distribution of circulation-induced trends by resampling. Figure 5.2 marks regions where the median of the distribution was significantly different from zero (dots in the middle row of this figure), that is, for western Europe in May and to a smaller extent central Europe in July and in the Balkans in February (where the estimated observed trend covered limited regions). Thus, the method for determining whether trends are explained by circulation changes is quite similar in the trend ratio method and the probabilistic approach. It is worth noting that circulation changes did not explain the strong warming in Finnmark, northern Norway, in December and January, in either of the methods.

6.2.2 Differences in within-type changes

For within-type changes, however, the two methods differed substantially. In the probabilistic approach, within-type changes are determined by comparing the estimated observed trend to the distribution of circulation-induced trends. The estimated observed trend is attributed to local effects if it is sufficiently far into the tail of the distribution. This is in contrast to the trend ratio, where the trend is either explained by
circulation changes or by within-type changes. Thus, the probabilistic approach has a stricter criterion to assign the within-type change than the trend ratio, and leaves many grid cells blank. There are two reasons for this. The first reason is the requirement for a significant estimated observed trend – in Paper I, we did not calculate significance, but discarded trend ratios exceeding 1.5. A less strict significance level on estimated observed trends would render larger regions of significant estimated observed trends. For instance, single dots in Figure 5.2, such as for South Norway in June, would be accompanied by dots covering a larger region with a less strict significance level. The other reason is that the probabilistic approach uses a standard significance level for the difference between $m_{\text{circ}}$ and $t_{\text{WFDEI}}$ of 0.05. Larger regions of within-type change would appear in case of a less strict significance level on the test that the likelihood that the estimated observed temperature trend belongs to the distribution of circulation-induced trends ($H_0$ and $H_1$ in Paper II).

Both Cahynová and Huth (2009) and Fleig et al. (2015) define within-type changes as changing internal properties of STs (e.g. that STs warm over time). Beck et al. (2007), point out that these changes may be caused by uncertainties in the climate data and ST classification, as well as increased radiative forcing from greenhouse gases. If, for instance, a warm ST is consistently mis-classified as a cold ST towards the end of the period, the resulting hypothetical time series would not capture an increased frequency of warm STs, although the estimated observed trend would increase. Further, the trend ratio treats circulation changes and changing properties of STs as mutually exclusive and thus provides a too large region of within-type changes. In reality, there is a gradual transition from circulation changes to within-type changes. The probabilistic approach takes this gradual transition into account, and focuses on regions that are evaluated as within-type changes with high confidence. For the purpose of identifying possible feedback mechanisms, it is important that within-type changes do actually reflect changing properties of STs as a result of a physical change, therefore, trend ratios are not suited as guides towards regions experiencing potential land–atmosphere feedbacks.

Although our probabilistic approach provides more focused regions, and a way of identifying within-type trends that are significantly different from circulation-induced trends, it does not separate the changing properties of STs into influences from land–atmosphere feedbacks, radiative forcing, or uncertainties. Further research is needed to separate out uncertainties from part of the within-type changes that is caused by physical processes. An approach to separate out these causes may be to calculate within-type changes in maximum and minimum temperature and other variables. Given that the soil moisture–temperature feedback is associated with heat waves,
maximum temperatures would provide a better way of distinguishing it from other processes. Further studies could also improve upon our use of fixed months that is not able to detect whether a circulation change in January influences temperatures in February. Detecting changes between months would require using a moving average of e.g. 30 days, or different day lengths. Moreover, Paper II showed many regions and months of strong within-type changes in Europe that would be worth exploring further. The region north of the Black Sea in November, for instance, displayed interesting signals described in Paper II. These signals could be pursued further.

In addition to providing a tool for identifying regions of potential feedbacks, the probabilistic approach may prove useful for statistical downscaling applications. Downscaling approaches that link the local climate to the large-scale atmospheric circulation assume that trends are caused by circulation changes and that they remain stable over time, but this assumption is violated when within-type changes are present (Beck et al., 2007; Vautard and Yiou, 2009; Cahynová and Huth, 2010; Cahynová, 2010). There is thus a need to confirm that circulation changes explain changes in the local climate before applying statistical downscaling methods, a step that is bypassed in many studies. The trend ratio provides a simple means to attributing changes in the local climate to circulation changes, and our method allows the user to define a level of rigidity by assigning a statistical significance to the circulation changes. Our probabilistic approach could be used to highlight regions where the application of the downscaling approach would (not) be considered appropriate.

6.2.3 Using only one classification of synoptic types

Many ST classifications exist, and using more than one classification is recommended by e.g. (Cahynová and Huth, 2016). For Paper II, we think that using only the SVG classification is sufficient to serve as an example of our probabilistic approach and to suggest regions where temperature trends may be explained by circulation changes or within-type changes. However, when using the results from Paper II to select a region for diagnostic modelling studies, results based on several classifications would be more robust.

Most ST classifications are valid for specific regions. For example, Hess Brezowsky Grosswetterlagen (GWL) are developed for central Europe and covers a smaller spatial domain than SVG (P. James, 2014; personal communication); the Lamb catalogue covers the British Isles and Vangenheim-Girs is most useful for eastern Europe (Hoy et al., 2013a). Re-running the analyses from Paper II with any of these classifications would likely not improve the quality of the results. Ideally, the analyses should be
re-run with a classification developed for northern Europe/Scandinavia. Only a few such ST classifications are available that also fulfill the requirement of being updated to 2010. Linderson (2001) applied the Lamb classification to southern Scandinavia, and Tveito (2007) used the Lamb approach to southern Norway. A classification has also been developed for Arctic Norway (Spitsbergen), by Niedzwiedz (2013), used by Isaksen et al. (2016). Neither of these classifications were considered in this thesis but it opens up possibilities for new research.

A further option would be to use the cost733class software (Philipp et al., 2010) to develop an updated classification for a given region. This is recommended because low discriminative capability of the classification is one source of within-type change. A poor classification may therefore obscure true signals. A further point to consider if developing a new classification is that the complex topography of Norway will be better resolved by fine-resolution input (for instance by the new ERA5 reanalysis or downscaled products of atmospheric variables.

### 6.3 Exploring the physical drivers of temperature trends

As a selected case of a region with potential within-type changes, we initially looked into the winter warming in an Arctic station, Karasjok in northern Norway. This station has frequent inversions in winter. The warming at this station may therefore be related to less frequent occurrences of inversions. Although more frequent westerly airflow may cause less frequent inversions, these temperature trends were much stronger than the circulation-induced trends, and could therefore not be explained by circulation changes alone. Other influences on temperature trends in northern Norway include changes in the sea ice extent, sea surface temperature (Hanssen-Bauer and Førland, 1998) and degrading permafrost, which will not be discussed in this thesis. To which degree these processes explain today’s changes remains to be tested.

The snow albedo feedback played a role in regions of snowmelt in spring (Papers III and IV). Because of the importance of snow cover on the spring climate at high latitudes, it was expected that the snow albedo feedback would explain the detected spring warming in Norway. Paper IV shows the presence of a positive soil moisture–temperature feedback in parts of South Norway during dry periods in summer, using a coupled land surface–atmosphere model framework. This is likely the first time the soil moisture–temperature feedback has been documented for Norway.
Linkage between statistical analyses and model studies

In this thesis, statistical attribution is employed to separate the temperature trend into a component attributed to circulation changes and one component attributed to within-type changes, and were used as a guide towards potentially interesting regions and months. The model-diagnostic studies is employed to delve deeper into causes of within-type changes, but the model study does not give an exhaustive account of all causes of within-type changes. Further, the model study of two summers (2006 and 2014) is not able to explain the full trend over the period 1981–2010.

Although the snow albedo and soil moisture–temperature feedbacks are not the only local factors playing a role in explaining within-type changes, we have showed that these feedbacks are relevant causes of temperature anomalies. In particular, the study proves that the soil moisture–temperature feedback could explain enhanced warming, provided sufficiently dry conditions. Combined with observed drying trends (Wilson et al., 2010), this result is a step towards explaining the within-type trends over the full period 1981–2010, despite short model runs. The model-diagnostic approach is the closest we can get to assessing causality without performing controlled physical experiments (which is nearly impossible given the spatial and temporal scales in question).

Further steps are needed to extend the model analyses to the full 30-year period, which would require additional computing power and storage. Moreover, future studies should quantify the influence of improper circulation classification on within-type changes, and should test other feedback mechanisms. Within-type changes may arise from a range of feedback mechanisms, which to a varying degree play a role in different regions and months. Temperature trends at high latitudes may be influenced by a range of local factors, whose influence could be diagnosed for Norway:

i) snow albedo feedback (Dery and Brown, 2007)

ii) changes in wind leading to changes to inversions,

iii) water vapour feedback (Ingram, 2013),

iv) reduced Arctic sea ice cover (Hanssen-Bauer and Forland, 1998; Cohen et al., 2014; Simmons and Poli, 2014),

v) changing sea surface temperatures (Deser et al., 2007),

vi) Planck feedback (Pithan and Mauritsen, 2014),
vii) soil moisture–temperature feedback, which has not received much attention at high latitudes,

viii) low discriminative capability of the ST classification, which may also lead to within-type changes.

### 6.3.2 Potential to improve land surface models

Improving our understanding of the drivers of hydroclimatological change helps the scientific community to improve land surface models (LSMs). Although LSMs are constantly being improved with better model structure and parameterisations, they still lack a number of processes, for instance lateral transport and subgrid scale processes (Niu et al., 2011; Davison et al., 2016). Moreover, LSMs perform poorly in regions of complex terrain and with a seasonal snow cover (Aas et al., 2017), such as in Norway. To improve modelling in these regions, we must determine in which way the LSMs perform poorly. In Paper IV, we mentioned two sources of uncertainty: i) improper representation of soil depth and root depth, and ii) improper parameterisations of e.g. evapotranspiration. Both potentially lead to too much soil moisture. Further studies are planned to explore differences in coupling between SeNorge and WRF–Noah-MP, with the aim to separate the sources i) from ii). Specifically, effects of land cover classes will be eliminated by implementing the same land cover and vegetation maps in WRF–Noah-MP as in SeNorge.

### 6.3.3 Future changes

Both warming and increased variability in summer dryness are expected to continue with climate change. Seneviratne et al. (2006) and Zampieri et al. (2009) describe a northward shift of climate regimes with climate change, such that central and eastern Europe become transitional climate regimes during dry periods. Vautard et al. (2007) established a link between summer dry conditions in Northern Europe and winter droughts in southern Europe. Trends towards both winter and summer drying are projected for southern Europe (Collins et al., 2013, for the period 2081–2100 relative to 1986–2005). From the model diagnostic study in Paper IV and projections of summer dryness in South Norway (Hanssen-Bauer et al., 2015), it is likely that also South Norway in the future will turn into a transitional evapotranspiration regime during dry periods in summer. This highlights the importance of an increased awareness about enhanced warming trends in Norway and other seasonally snow-covered regions.
Different factors govern evapotranspiration in energy-limited regimes versus soil moisture-limited regimes. Evapotranspiration trends increase with warming in both regimes. In moist climates, evapotranspiration increases with incoming short wave radiation, and in both moist and dry climates, evapotranspiration increases soil moisture, but more strongly in soil moisture–limited climates (van Heerwaarden et al., 2010). When projecting trends in evapotranspiration, it is therefore important to know which regime applies. For regions that exhibit energy-limitation during parts of the year and soil moisture-limitation during other parts of the year, a correct characterisation of the regime is a requirement to correctly estimate the evapotranspiration. A shift from one evapotranspiration regime to another must be taken into account when projecting trends in evapotranspiration. Paper IV provides information about evapotranspiration regime shifts in Norway during two warm summers, which may prove valuable for improved evapotranspiration estimation in Norway.
Chapter 7

Conclusions

The main goal of this thesis was to detect recent monthly hydroclimatic trends and assess the climatological cause of those trends through statistical attribution and a model diagnostic study (the research papers are summarised in Figure 1.1 and Chapter 5). The main outcomes of this thesis is a novel probabilistic approach for statistical attribution of trends (Paper II) and an improved understanding of local factors enhancing temperatures (Paper IV). Paper II improves upon limitations inherent in existing methodologies by evaluating the statistical significance of synoptic circulation changes on observed climate trends. Using a classification of synoptic types, we attributed temperature trends either to changes in the frequency of synoptic types (circulation changes), or to changes within the synoptic types (within-type changes). South Norway was identified as a region with a potential for positive land–atmosphere feedbacks (within-type changes), and was therefore selected as the study region for targeted experiments using the regional climate model WRF coupled to the land surface model Noah-MP. Because recent seasonal temperature trends in Europe have been accompanied by large changes in the snow cover and summer drying, we focused on the snow albedo feedback and the soil moisture–temperature feedback.

The four research questions from the introduction may be answered as follows:

1. Which regions and months have experienced significant trends in hydroclimatology in Europe?

Temperature trends for the period 1979–2009 showed a general warming, which was most pronounced in winter, spring and summer (Paper I). Annual temperature trends for the period 1981–2010 showed warming for nearly all of Europe. This warming was distributed unequally across months,
displaying significant temperature trends in large parts of Europe for April–August and November (Paper II). For Norway, significant temperature trends were detected in April and July–September. Strong winter warming was detected in northern Norway in Papers I and II.

Strong trends in snow indices (snow cover extent, snow water equivalent and snow duration) were detected for Norway, particularly in spring and at low elevations (Paper III). Between 1981–2010 and 1961–1990, the snow cover extent in mainland Norway decreased by more than 6% (20 000 km²) in April and May. Temperature trends were detected for the estimated observed dataset for Norway (SeNorge) for April, July–September, as well as in northern Norway in winter, confirming results in 1a).

2. Can the detected trends for Europe be attributed to circulation changes or within-type changes?

Circulation changes could explain the large-scale warming for Europe in February for the period 1979–2009, as well as localised warming in other months (Paper I). Elsewhere, circulation changes could not account for all the estimated observed warming, which implies that other factors, such as positive land–atmosphere feedbacks, contributed to the warming.

By applying the probabilistic approach to Europe for the period 1981–2010, we found that warming in large parts of Europe for April, June–August could partly be attributed to within-type changes. For Norway, within-type changes could partly explain the warming in April and July–September (Paper II).

3. How do snow cover changes influence warming in a region with seasonal snow cover?

Significant negative correlation between temperature and snow cover extent strongly suggested the snow albedo feedback as a main cause for April warming in South Norway (Paper III), confirmed by the model diagnostic experiments (Paper IV).

4. For regions with within-type changes in Norway, can the detected trends be attributed to the snow albedo feedback (in spring) or the soil moisture–temperature feedback (in summer)?
During the anomalously warm summer of 2014 and the warm and dry summer of 2006, the warming could partly be attributed to the positive soil moisture–temperature feedback, at least during parts of the summer (Paper IV). This signal appeared in two independent datasets – the WRF–Noah-MP model results and SeNorge. Further studies are required to determine which processes cause these differences between WRF–Noah-MP and SeNorge, which further contributes to identify needs for improvement in the model parameterisation or input data.

Decreases in the snow cover during spring are projected for Norway, and an increased probability of dry conditions during summer is projected for South Norway. Therefore, the snow albedo feedback and the soil moisture–temperature feedback are expected to continue to enhance warming in the future. The combined methodology of identifying within-type changes with land–atmosphere feedbacks may prove valuable when predicting future temperatures. In particular, a correct characterisation of the evapotranspiration regime (soil moisture-limited versus energy-limited) is key because evapotranspiration estimates are very sensitive to the regime. This thesis shows the added value of bringing together different disciplines, namely synoptic climatology and land surface modelling.
Chapter 8

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Appendix

Conference Presentations (oral)


### Conference Posters


