GEOG0105 Research Project and Dissertation

Hydrological trends in a warming climate:

A daily trend analysis of 207 Norwegian catchments

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ABSTRACT

Global warming is impacting regional hydrology, especially in mountainous areas where seasonal snowpack plays an important role. Hydrological trends in 207 Norwegian catchments, representing a variety of hydrological regimes, hydro-climates and altitudes, were divided into six subsets according to major watersheds, to limit the interference of latitudinal dependent trends. The catchments were analysed to assess whether, in a warming climate, rainfall and snowmelt trends are the main drivers of streamflow change in Norway.

Both an annual and daily trend analysis approach were used to estimate trends in streamflow, snowmelt, rainfall and temperature, employing the Mann-Kendall test and Sen's Slope Estimator in to assess trend significance and magnitude respectively. Consistency in recent (1983-2012) and long-term (1963-2012) trends, altitude dependence and relationship between streamflow, snowmelt and rainfall trends were examined.

Rainfall and snowmelt are the main drivers of streamflow trends in Norway and are linked to periods of warming, causing phase shifts in precipitation and earlier and slower snowmelt. In high-elevation catchments snowmelt season is ending later due to increased snow accumulation in winter. Catchments east of the Scandinavian Mountains experienced increased summer rainfall. The daily changes are found to be of greater importance when assessing changes to the hydrological regime. Daily trend patterns are generally consistent for both periods. Pronounced warming and increased rainfall in winter during the long-term period is possibly linked to the North Atlantic Oscillation.

Catchment hydrology has been affected differently in Norway, depending on hydrological regime, hydro-climate and altitude. The detected changes have been associated with increased annual and seasonal temperatures. Future research should focus on the attribution of detected trends to climate change and assess the influence of decadal scale climate variability.

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Abbreviations

α	significance level
D&A of impacts	detection and attribution of climate change impacts
EDW	elevation dependent warming
Fig.	figure
GHG	greenhouse gas
HRN	hydrological reference network
IPCC	Intergovernmental Panel on Climate Change
LPSA	low pressure system activity
MA	moving average
m.a.s.l.	meters above sea level
MK	Mann-Kendall
NAO	North Atlantic Oscillation
NVE	Norwegian Water Resources and Energy Directorate
OLS	ordinary least squares regression
RS	combined rainfall and snowmelt trend
SWE	snow water equivalent
Tab.	table
WWDD	wet gets wetter and dry gets drier
5dMA	5-day moving average
10dMA	10-day moving average
30dMA	30-day moving average

1. CLIMATE CHANGE IMPACTS ON HYDROLOGY

1.1. Introduction

This dissertation investigates the evidence for hydrological changes in Norway by analysing 50 years of daily streamflow, rainfall, snowmelt and temperature records from 207 catchments. The observed changes are discussed in the context of rising temperatures and climate change. This chapter therefore begins with a brief review of the effects of a warming climate on hydrology (Section 1.2), before reviewing the current literature on detection and attribution of climate change impacts and approaches for analysing hydrological trends (Section 1.3), followed by a short introduction to the hydro-climatology of Norway (Section 1.4). Finally, the research aims and objectives are presented (Section 1.5).

1.2. The hydrological cycle in a warming climate

Evidence from diverse independent climate indicators have confirmed unequivocally that the world has warmed since the 19th century (Fig. 1.1a; Hartmann et al., 2013). Rising global temperatures have mainly been attributed to anthropogenic radiative forcing caused by greenhouse gas (GHG) emissions, rather than the natural variability of the climate system (Fig. 1.1b; IPCC, 2014). In addition to being the dominant cause of observed warming of the lower atmosphere, anthropogenic influence has been detected in warming of the ocean, in reductions of snow and ice, in global mean sea level rise, and in changes in the global water cycle (IPCC, 2014). This change in the climate system subsequently leads to changes in other systems, physical, biological and human.

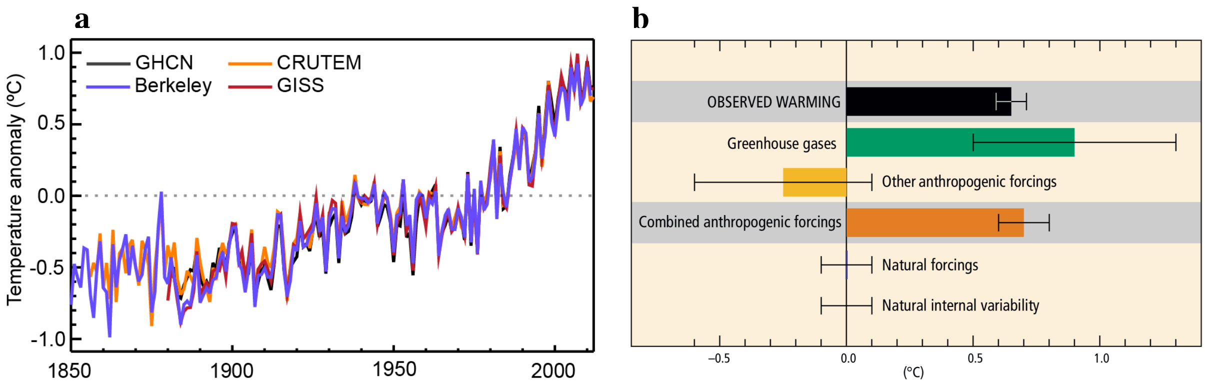

Figure 1.1: (a) Global annual average land-surface air temperature (LSAT) anomalies relative to a 1961–1990 climatology from four different data sets (Source: Hartmann et al., 2013). (b) Contributions to observed surface temperature change over the period 1951-2010 (Source: IPCC, 2014).

There is increased evidence that global warming is changing the hydrological cycle. Changes are particularly evident in precipitation and potential evaporation, as the water holding capacity of the air increases with 7% per 1°C warming (Trenberth, 2011; Jiménez Cisneros et al., 2014). A warming climate has also been linked to an amplification of precipitation extremes (Allan and Soden, 2008; Fischer and Knutti, 2016). Streamflow and runoff show few consistent global changes, but generally increase at high latitudes (Hartmann et al., 2013). Precipitation is the main driver of changes and interannual to decadal variability in discharge (Dai et al., 2009). The most notable projected effects of climate change on the water cycle are less precipitation falling as snow and a reduction in snow cover extent and duration, wet regions becoming wetter and dry regions becoming drier (WWDD), increasing global mean precipitation and evaporation with substantial regional variations (Jiménez Cisneros et al., 2014). However, the WWDD paradigm of amplification of existing hydrological conditions appears not to be robust over land (Greve et al., 2014).

Growing evidence suggests that the rate of warming is amplified with altitude. This phenomenon is still poorly observed, but if true means that mountainous regions will be affected to a greater extent by global warming, especially high altitude ice and snowpack, and associated runoff (Pepin et al., 2015). Whether due to elevation dependent warming or global warming in general, the cryosphere in mountain regions is rapidly declining, which is likely to accelerate over the next decades (Huss et al., 2017). Many regions of the world relying on glaciers and seasonal snowpacks for water supply are likely to be negatively impacted by hydrological changes related to rising temperatures (Barnett et al., 2005). Losing these snowpacks could lead to a cascade of impacts on humans and ecosystems (Huss et al., 2017).

1.3. Detection and attribution of climate change impacts

The motivation for assessment of observed changes and impacts is the possibility of changes in observational records being an indication of and critical to understanding future expected changes (Burn et al., 2012; Cramer et al., 2014). However, some key limitations to such extrapolation exist. Firstly, the nonlinear nature of many systems means that past impacts cannot be linearly scaled to future impacts. Secondly, future impacts may occur where changes are yet to be detected (Cramer et al., 2014). As long as these limitations are kept in mind, the usefulness of examining observed climate change impacts is not diminished. Furthermore, observed trends can be compared to changes predicted by earth system models and global and regional climate models, and serve as an indication of the accuracy and precision of these predictions.

Developing a common methodology for the detection and attribution of climate change, has long been the focus of the research community (e.g. Hegerl, 2010; Stone et al., 2013; Cramer et al., 2014). Assessment of changes within the climate system initially received the most attention of the research community. The detection and attribution of climate change impacts (D&A of impacts), i.e. assessment of the causal relationship between drivers and a responding system, have been the focus of research since changes in the climate system became apparent (Cramer et al., 2014). *Detection* of climate change is demonstrating that a system, be it climate, natural or human system, has statistically changed beyond some reference behaviour (Hegerl, 2010), as opposed to *attribution*, which establishes the causal relationship and magnitude of an observed change (Stone et al., 2013). Attribution requires the contributions of all external drivers to be evaluated, as the confidence with which the system change can be attributed to climate change should be indicated (Cramer et al., 2014). *Confounding factors*, i.e. drivers other than climate change which affect the system, pose a substantial challenge to the assessment of climate change impacts on natural and human systems. In addition, high-quality, long-term observations relating to natural and human systems, and the factors affecting them are needed for a comprehensive attribution analysis (Cramer et al., 2014). This often results in studies making qualitative rather than quantitative assessments of climate change impacts.

1.3.1. Detection of climate change signals in hydrological records

A major unresolved scientific problem in hydrology is whether the regional hydrological cycle is accelerating or decelerating under climate and environmental change, and whether these changes are irreversible, i.e. if there are tipping points involved (Blöschl et al., 2019). Explaining patterns of regional hydrological change is a central challenge in hydrology. Obtaining reliable information on such patterns beyond the basin or national scale will aid in the identification and attribution of flow regime changes caused by climate change (Stahl et al., 2010). Moreover, it is uncertain to what extent cold region runoff and groundwater will undergo changes in a warmer climate (Blöschl et al., 2019).

There are several methods and approaches that can be employed for detection of changes to hydrological systems. Many approaches compare observed or modelled data, or indicators calculated from these, to a stationary or non-stationary baseline (Cramer et al., 2014), e.g. Indicators of Hydrological Alternation (see Laizé et al., 2010). Projections of streamflow or other hydrological variables are frequently used for such comparisons, despite the large uncertainties associated with predictions of global hydrological models (Todd et al., 2011; Shen et al., 2018).

The hydrological cycle of a region or river basin can be viewed as a natural system, which can be affected by external drivers from the human, climate or other natural systems (Stone et al., 2013), meaning forcings other than climate change affects the system. The processes influencing the system can be nonlinear, involve threshold effects, be non-local in both space and time or involve lagged responses (Cramer et al., 2014). Trend detection and attribution in hydrological time series is complicated by the high natural variability, low signal-to-noise ratio, and display of complex behaviour of the time series, in addition to different drivers acting simultaneously in a catchment and some driver-effect mechanisms not being well understood (Blöschl et al., 2007; Delgado et al., 2010; Merz et al., 2012). Hence, streamflow trends must be interpreted with caution. Furthermore, other drivers, e.g. land-use changes, irrigation or urbanisation, may act as confounding factors. Much of the published literature on observed hydrological trends does not clearly differentiate between changes due to climate, land use, and water use (Burn et al., 2012).

Similar to the efforts of creating a unified framework for D&A of impacts, Burn et al. (2012) called for the development of a common analysis protocol for hydrological trend analysis and assessment of the impacts of climate change on hydrology. Studies have found considerable spatial variability in detected streamflow changes, and trend analysis results are sensitive to the selection of data, trend detection method and the time period chosen (Stahl et al., 2010). A unified approach will facilitate the synthesis of results from different regions, enabling the presentation of the various responses of watersheds to climate change in a manner that overcomes the discrepancies of collected studies. Consequently, consistent use of hydrological variables in trend studies is key to facilitate comparisons across regions, while also accommodating regional differences and the suitability of certain variables for particular regions (Burn et al., 2012). Many studies only qualitatively attribute hydrological trends to climate change attribution, or merely interpret the detected trends. There are two methods for quantitatively relating detected hydrological changes to an assumed driver, termed data-based and simulation-based attribution (Merz et al., 2012). The *data-based* approach compares the time series of the potential cause and effect variables, e.g. by evaluating the correlation between them, whereas the *simulation-based* approach utilises models to identify the causal link between an observed trend and the assumed driver (e.g. Hidalgo et al., 2009; Duethmann et al., 2015; Kormann et al., 2016).

In recent decades, many countries have established national hydrological reference networks (HRNs), e.g. the UK (Bradford and Marsh, 2003), USA (Slack and Landwehr, 1992), and Norway (Fleig et al., 2013), which are collections of streamflow records from gauging stations intended to aid in observation of watershed responses to variation and changes in climate (Whitfield et al., 2012). These networks serve as a basis for investigating climate and catchment processes which determine regional hydrological changes (Déry et al., 2009) and are especially suited for the identification of streamflow changes driven by climate change (Burn et al., 2012). Such large-scale streamflow datasets are essential for hydrological change detection, attribution of changes to the associated causes, and identification of the involved processes, which can help predict changes at unmonitored sites (Hannah et al., 2011). International HRNs, which include catchments from a variety of climatic, topographic and ecological regions, can facilitate inter-comparative hydrological trend studies (Burn et al., 2012).

Hydrological trend analysis aims to detect changes in hydrological time series and assess the magnitude of the change. Sometimes the goal is detecting a statistically significant trend, but often the magnitude of the trend is assessed as well. The Mann-Kendall (MK) test is commonly used to detect of significant trends in combination with the Sen's Slope Estimator (SS) to calculate the magnitude of the detected trend (Tab. 1.1). The combination of the MK test and the SS estimator is by far the most common trend analysis method, followed by linear regression, followed by linear regression, and is often applied to monthly, seasonally or annually averaged data (Tab. 1.1 Burn et al., 2012). The hydrological time series is often prewhitened to remove autocorrelation (Tab. 1.1), as autocorrelation interferes with the MK test (Yue et al., 2002). Indiscriminate use of prewhitening has more recently been criticised, as it can impact the detection of trends and cycles in time series, as well as diminish their magnitude and significance (Razavi and Vogel, 2018). However, Kormann et al. (2014) found little difference between the MK test results of prewhitened and non-prewhitened data.

Streamflow change is a direct indicator of a changing regional water balance (Cramer et al., 2014), hence it is the frequent subject of hydrological trend studies (Tab. 1.1). Studies of streamflow trends often focus on regional differences (e.g. Stahl et al., 2010), single catchments (e.g. Folton et al., 2019), or recent vs. long-term trends (e.g. Hisdal et al., 2001). Most common are analyses of trends in mean annual, seasonal or monthly values (e.g. Stahl et al., 2010; Makarieva et al., 2019), extremes events (e.g. Vormoor et al., 2016), or hydrological indices (e.g. Wilson et al., 2010).

Table 1.1: Studies on hydrological trends, mainly streamflow, from different regions of the world.

Study	Focus	Data	Period	PW	Method	Findings
Dai et al., 2009	Streamflow changes in the *world*'s major rivers	Monthly Q	57 years	No	t test	Most of the world's large rivers show large annual variability. A third of the largest 200 rivers show significant trends with climatic forcing being the dominant factor. Some interannual variation is correlated with the ENSO.
Durocher et al., 2019	Trends in discharge to the *Arctic Ocean*	Annual Q based on monthly Q	41 years	No	MK, SS	Increased freshwater discharge to the Arctic Ocean between 1975-2015.
Hisdal et al., 2001	Changes in drought magnitude and frequency in *Europe*	Drought parameters based on daily Q	28-84 years	No	MK, Resampling	No clear indication that recent droughts have become more frequent or severe. Chosen time periods, records and parameters have a strong influence on result. Strong agreement between MK and resampling tests.
Stahl et al., 2010	Streamflow trends in *Europe*	Monthly, annual Q, low flow indices	42-72 years	No	MK, SS	Importance of distinguishing between annual and seasonal trends. Regional coherence of increasing winter streamflows. Northward shift of negative trends in transitions from spring to summer.
Folton et al., 2019	Hydrological trends in a *Mediterranean* catchment	Various hydrological indices based on daily Q, P	50 years	No	MK, SS	Flows respond not only to the magnitude of changes in rainfall and temperature but also to the timing of these changes. Catchments that usually support low flows are the most sensitive to climatic disturbances.
Kormann et al., 2014	Elevation dependent trends in the *Austrian Alps*	Daily T, P, NSH, TSH, Q	30-60 years	Yes	MK, SS	Importance of identification of sub-seasonal trend timing. Trend timing more stable measure than magnitude. Majority of trends occur in spring/ early summer (NSH, SH, Q, T). Runoff trend timing strongly dependent on altitude. Climate change effects on temperature not dependent on altitude.
Wilson et al., 2015	Streamflow changes in the *Nordic countries*	Annual, seasonal, flood and drought indices based on daily Q	39-85 years	Yes	MK	Overall trend of increased annual, spring and winter streamflow. Streamflow trends result from temperature and precipitation changes. Trends mostly agree with available streamflow projections.
Skaugen et al., 2012	Trends in snow water equivalent in *Norway*	Annual SWE	19-79 years	No	MK	Positive trends above 850 m. a.s.l. due to increased precipitation. Seasonal trends of precipitation and temperature must form basis for assessing snow trends. The NAO index correlates positively with SWE at high elevations and negatively at low elevations for periods of high NAO index.
Vormoor et al., 2016	Changes in flood magnitude and timing in *Norway*	POT event series based on daily Q, R, SM	30-50 years	No	MK, PR, SS	Negative trends in flood magnitude found more often than positive. Regional differences reflect flood generating processes. General shift from snowmelt to rainfall dominated flood generation.
Makarieva et al., 2019	Warming impacting hydrometeorological regime of Arctic rivers in *Russia*	Monthly and annual Q, spring freshet start dates	50-80 years	Yes	MK, SS	Increasing temperatures (2°C over 50 years) has shifted the rain-snow ratio. Positive monthly streamflow trends in autumn-winter and spring flood periods over 80 years. Most increases in streamflow are not monotonic, but occur in break points.

Study	Focus	Data	Period	PW	Method	Findings
Déry et al., 2009	Detection of runoff timing changes in *western Canada*	5-day means of Q, P, T, SWE	35 years	No	SS	Proposed method able to detect streamflow timing changes throughout the year.
Kim and Jain, 2010	High-resolution streamflow trends in *western United States*	3-day moving average Q, centre of volume	60 years	No	Median quantile regression	Daily resolved trend analysis enables better understanding of changes in timing and magnitude of streamflow. Trends show large variability between catchments.
Vilanova, 2014	Mean annual streamflow trends in *Brazil*	Mean annual Q, total annual R	18-29 years	Yes	MK, regional MK, SS	No significant streamflow trends in 29 year period. Significant trends in 18 year period was attributed to human activity, as no significant trend in rainfall was found.
Mahmood and Jia, 2019	Hydro-climatic trends in *Lake Chad basin*	Daily Q, P, T, and other meteorological variables	42 years	No	MK, SS, change point analysis	Significant increasing trends in temperature. Very weak to strong decreasing precipitation trends. Found a 40% reduction in inflow to Lake Chad over the whole period, of which 66% and 34% are attributed to human activities and climate change respectively.

Abbreviations: Temperature (T), precipitation (P), new snow height (NSH), total snow height (TSH), runoff/streamflow (Q), peak over threshold (POT), prewhitening (PW), moving average smoothing (MA), rainfall (R), snowmelt (SM), snow water equivalent (SWE), Mann-Kendall test (MK), Sen's Slope estimator (SS), Poisson regression (PR)

Trend detection of monthly, seasonally or yearly average data, while commonly used (Tab. 1.1.; Burn et al., 2012), is not a robust method for detecting climate change signals (Kormann et al., 2014). High resolution trend analysis has the potential to better pinpoint the timing and causes of streamflow changes and has been developed in the last decade (see Déry et al., 2009; Kim and Jain, 2010; Kormann et al., 2014; Kormann et al., 2015; Rottler et al., 2018). Daily resolution trend analysis gives a more detailed picture of how the hydrological regime has changed in a catchment. Furthermore, Kormann et al. (2014) found trend timing to be a more stable measure than trend magnitude to characterise sub-seasonal trends. High resolution trend analysis not only gives a more detailed picture of how and in which periods of the year the hydrological regime is changing, but could also facilitate the attribution of the changes to specific drivers.

1.4. Hydro-climatology of Norway

Norway has a large geographical range stretching from the 57°N to 71°N latitude and 4°E to 31°E longitude. The dominant topographical feature in Norway are the Scandinavian Mountains, which run along the Finnish-Norwegian and Swedish-Norwegian borders and turn westward at 63°N dividing the humid coastal climate of Vestlandet from the more continental climate of Østlandet (Fig. 1.2).

A wide geographical range and complex topography combined with the heat and moisture brought by the North Atlantic Current and westerlies results in diverse hydro-climatic conditions (Mohr and Tveito, 2008; Hanssen-Bauer et al., 2015). The peninsular shape of Fennoscandia limits the size of its rivers and watersheds, but due to the high precipitation, e.g. on the western slopes of the mountains which often have mean annual precipitation rates of 3500-4000 mm (Hanssen-Bauer et al., 2015), high runoff rates are still common (Fig. 1.3a; Hyvärinen and Kajander, 2005). The typical flow regime in Norway consists of a winter low flow period, a clearly defined snowmelt flood in spring, a summer low flow period, and the occasional autumn flood due to precipitation as rain (Gottschalk et al., 1979). The seasonal distribution of runoff is dependent on precipitation, snow accumulation and the timing of snowmelt (Hanssen-Bauer et al., 2015) and can be classified into four prevailing hydrological regimes (Tab. 1.2).

Figure 1.2: *Topography and runoff regions of Norway* (Data source: Kartverket, 2010)

Table 1.2: Hydrological regimes in Norway based on low and high water periods (Gottschalk et al., 1979).

Hydrological regime	Description
Mountain	High water in spring dominated by snowmelt. Dominant low water flow in winter caused by snow accumulation.
Inland	High water both in spring and autumn caused by snowmelt and high precipitation respectively. Dominant low water flow in winter caused by snow accumulation.
Transition	High water both in spring and autumn caused by snowmelt and precipitation respectively. Low flow both in winter and summer caused by snow accumulation or high evaporation and/or low precipitation respectively.
Atlantic	High water in autumn or early winter dominated by rainfall. Low flow in summer caused by high evaporation and/or low precipitation.

Norway is divided into six runoff regions based mainly on the borders of major watersheds and loosely on administrative regions (Tab. 1.3; Fig. 1.2). In all of Norway the annual evapotranspiration is less than annual precipitation. Annual evaporation is about 200-500 mm yr^{-1}, with the highest values in southern Norway and decreasing with increasing altitude and latitude (Hyvärinen and Kajander, 2005; Hanssen-Bauer et al., 2015). With little variation in evapotranspiration, the dominant influence on runoff rates is precipitation, which can be seen in the similarity in runoff and precipitation patterns (Fig. 1.3a; Fig. 1.4). Total runoff can show great interannual variability (Hanssen-Bauer et al., 2015).

Table 1.3: Overview of the runoff regions of Norway and their characteristics, as used by Hanssen-Bauer et al. (2015) based on Pettersson (2012).

Runoff region	Hydrological regime	Hydroclimate	Area (km^2)
Finnmark*	Mountain	Arctic climate, cold and dry with low runoff rates. Coastal areas are slightly warmer and more humid.	98,550
Nordland	Mountain Transition Atlantic	High precipitation and runoff rates. Similar to Vestlandet, but with lower annual temperatures due to more northernly location.	39,740
Trøndelag	Atlantic Inland Transition Mountain	Medium precipitation and runoff rates. Mild winters, especially in coastal areas.	33,372
Vestlandet	Mountain Transition Atlantic	Highest and most intense precipitation due to orographic precipitation and resulting in high runoff rates. Small watersheds due to the steep topography.	60,699
Østlandet	Atlantic Inland Transition Mountain	Has a more continental climate with dry, cold winters and the highest summer temperatures. The bulk of the precipitation falls in summer and autumn. Is usually located in the rain shadow of the mountains, but can experience high precipitation when moisture is brought by southeasterly winds.	83,533
Sørlandet	Inland Transition Atlantic	High to medium precipitation and runoff rates. Cold winters and warm summers.	23,514

* Shortened from the full name “Troms og Finnmark”. This runoff region will be referred to as “Finnmark” throughout this study.

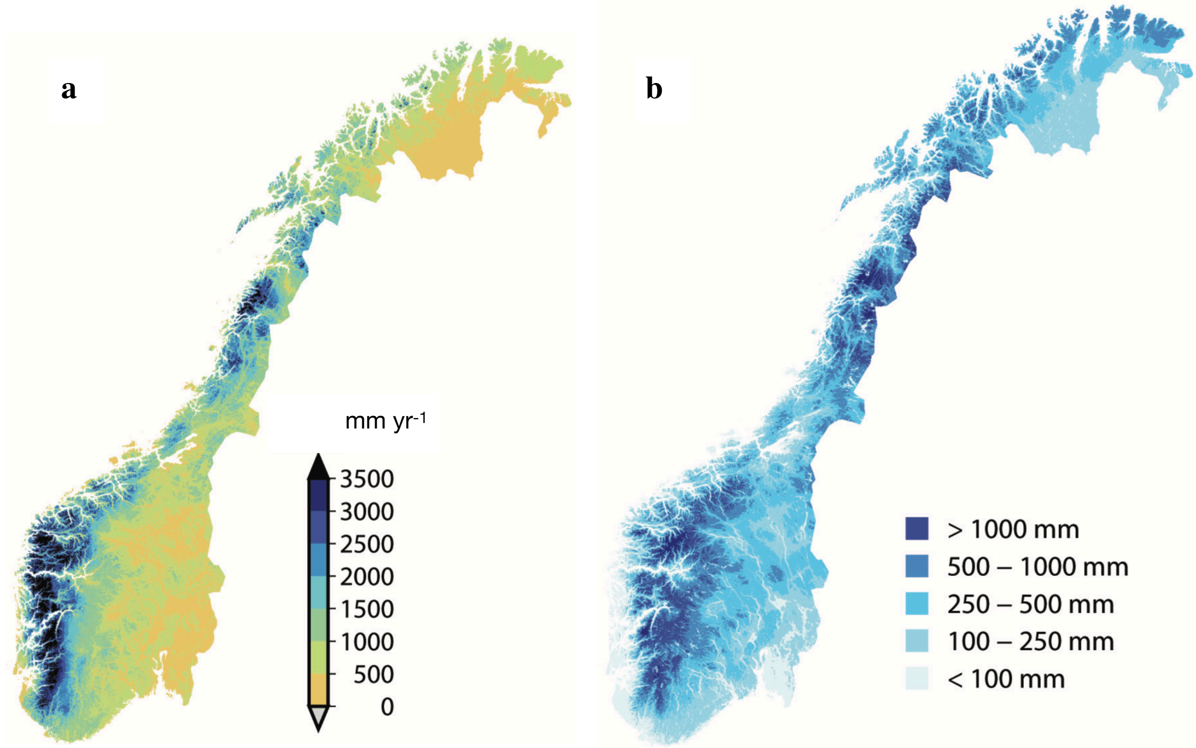

Figure 1.3: *Mean annual runoff (a) and mean annual maximum snow water equivalents (SWE) (b) in Norway for the reference period 1971-2000* (Source: Hanssen-Bauer et al., 2015).

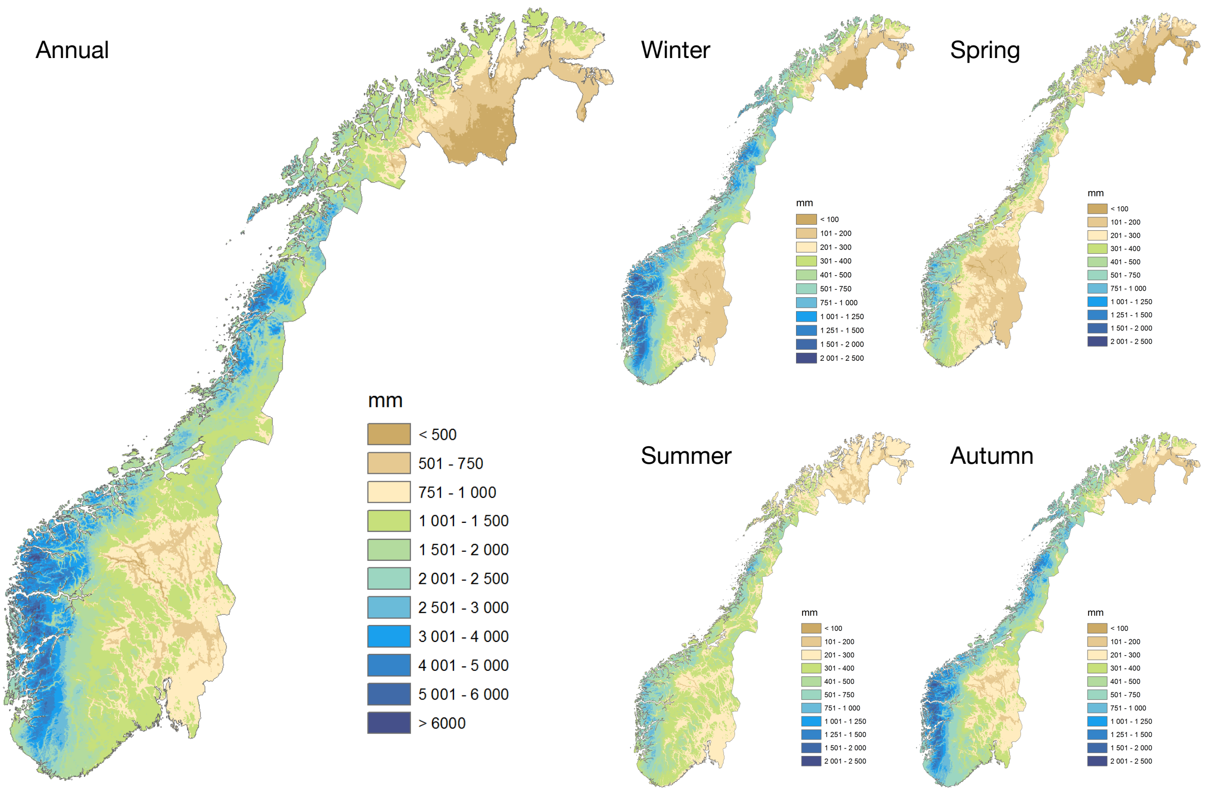

Figure 1.4: *Annual and seasonal precipitation in Norway* (Source: Norsk Klimaservicesenter)

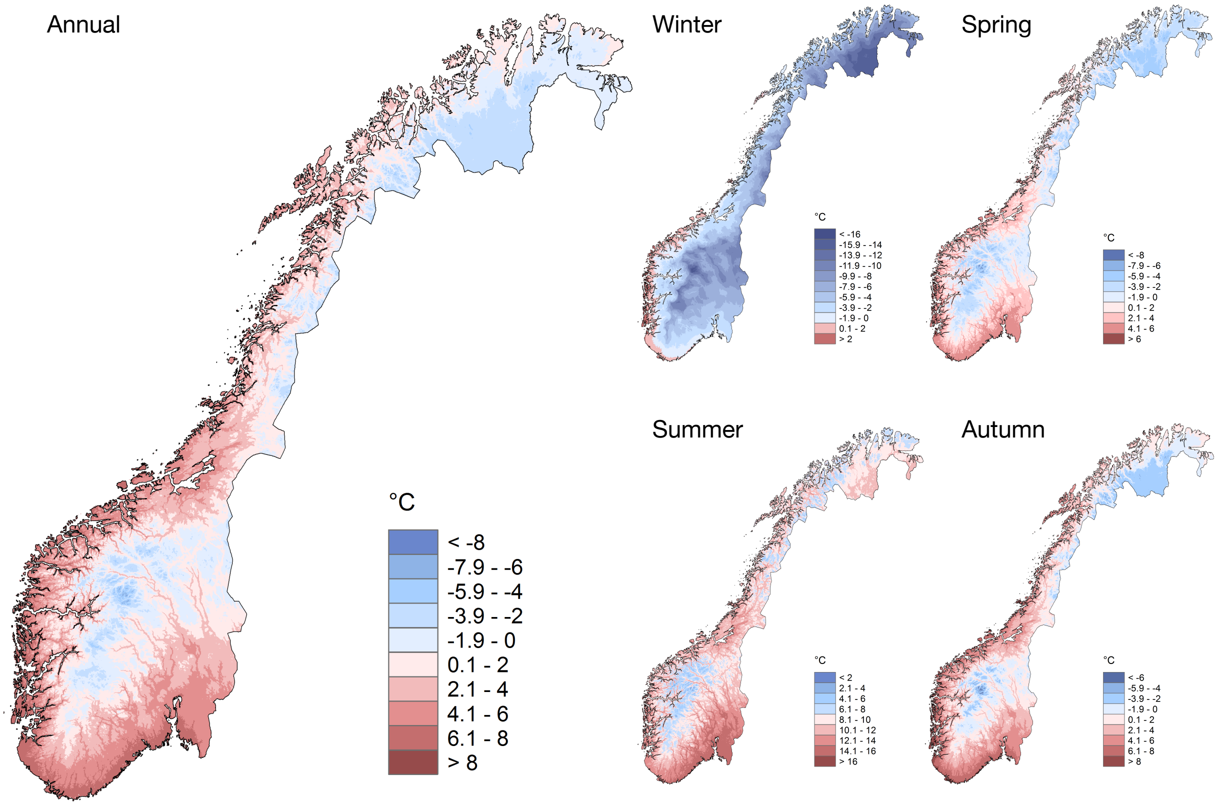

Figure 1.5: *Annual and seasonal mean temperature in Norway* (Source: Norsk Klimaservicesenter)

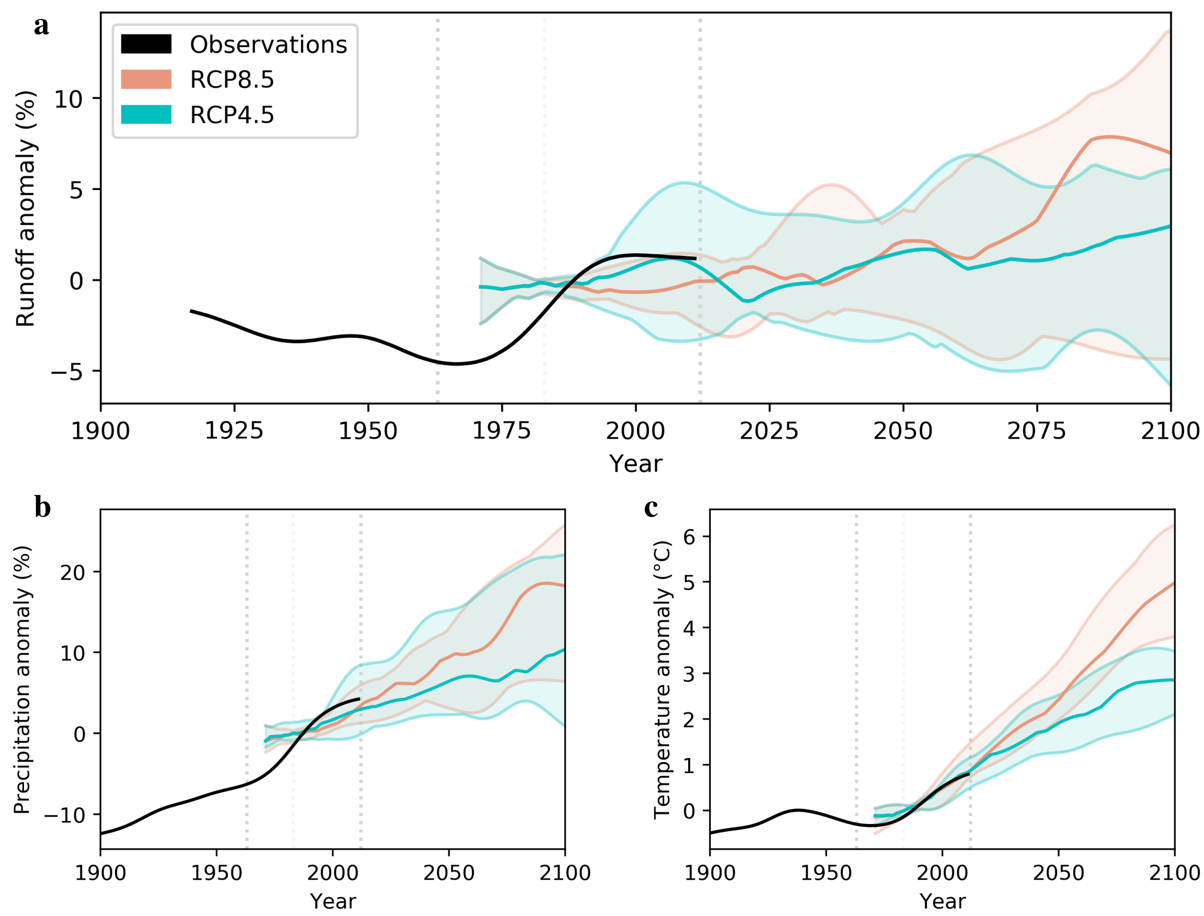

Figure 1.6: *Observed and projected runoff (a), precipitation (b) and temperature (c) in Norway.* Temperature and precipitation given in percent deviation from the 1971-2000 reference period. Projected data from an ensemble of ten EURO-CORDEX GCM/RCM simulations of two representative concentration pathways (RCP), high-emission RCP8.5 and low-emission RCP4.5 (see Wong et al., 2016). Grey vertical lines indicate the period analysed in this study (Data source: Norsk Klimaservicesenter).

1.5. Aims and objectives

Global warming has already started impacting non-climatic systems, both globally and regionally. In Norway, surface temperature, precipitation and runoff has increased since the 1960s (Fig. 1.6). Temperature and precipitation are projected to keep increasing until 2100 (Fig. 1.6b-c), while future runoff will likely not experience a large increase (Fig. 1.6a)[1]. Recent studies on climate change impacts on Norwegian hydrology have focused on changes in the flood regime (e.g. Vormoor et al., 2016), extreme events (Dyrrdal et al., 2012), on specific hydrological variables (e.g. snow: Skaugen et al., 2012; Rizzi et al., 2017), annual to monthly streamflow trends (Stahl et al., 2010; Wilson et al., 2010), or future changes (Hanssen-Bauer et al., 2015). This dissertation will analyse hydrological trends in Norway at a daily resolution, which to the knowledge of the author has not been done before.

[1] The reader is referred to Hanssen-Bauer et al. (2017) for a comprehensive report on future climate change in Norway

The aim of this study is to give a comprehensive analysis of observed, non-anthropogenic hydrological changes in Norway between 1963 and 2012, by analysing daily streamflow, snowmelt, rainfall and temperature data from 207 catchments. Due to the large number of catchments analysed, this dissertation will focus on trend detection.

The main research question is: *Has streamflow in Norway changed in a warming climate through changes in rainfall and snowmelt?* This has been divided into four research objectives:

I. *Are there any differences between recent and long-term trends?* It is well known that trend analyses are heavily affected by the period chosen. To this end, trends in two time periods will be analysed; a recent (1983-2012) and a long-term (1963-2012) period. The focus here lies on whether trends differ in the two studied periods, both in terms of significance, magnitude and timing.

II. *Do detected trends display any dependence on altitude?* Norway is well suited to analyse the relationship between hydrological changes and altitude, as most of the country is mountainous. As Vestlandet and Østlandet have similar elevation and latitudinal ranges, but different hydro-climatological conditions, the altitude dependence of the detected hydrological trends in these two regions will be examined in more detail.

III. *To what extent can changes in streamflow be attributed to changes in snowmelt and rainfall?* Precipitation, snow accumulation and timing of snowmelt are the main factors determining the hydrological regime in Norwegian catchments (Hanssen-Bauer et al., 2015). The author therefore hypothesises that rainfall and snowmelt trends are the dominant factors governing streamflow changes.

IV. *What is the effect of applying different moving average filtering to the original daily data?* The 30-day moving average (30dMA) used by Kormann et al. (2014; 2015) removed much variability from the data, which may have affected the detection of significant trends, the estimation of trend magnitude and the timing of daily trends. Therefore, in addition to 30dMA trends, 5-day MA (5dMA) and 10-day MA (10dMA) trends are calculated and compared.

2. DATA

2.1. Hydro-climatological data

2.1.1. Streamflow records

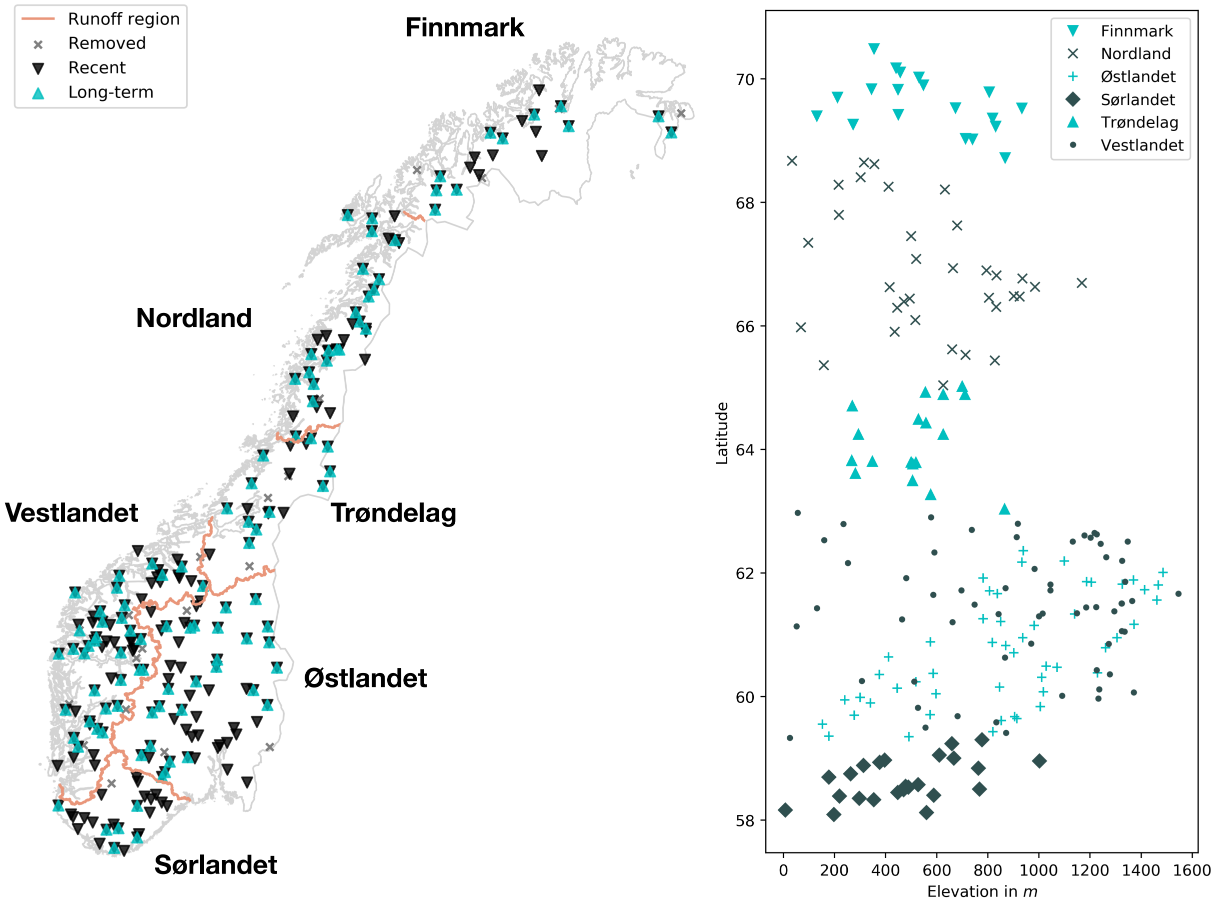

Figure 2.1: *Location of all catchments (left) and altitudinal distribution of catchments used in this study (right)*. Both geographical location and altitude refers to the gauging station of each catchment. Each runoff region covers different latitudinal ranges, with the exception of Vestlandet and Østlandet which overlap. These two regions are separated by the Scandinavian Mountains and also have catchments spanning the same elevation.

Daily streamflow records from 227 Norwegian catchments were available, and these records are the same as used by Vormoor et al. (2016). Data lengths of the available records were varied with longest record starting in 1871 and the most recent data being from 2014 (Tab. A1, Appendix; Fig. 2.3), but a minimum of 30 years were available for all catchments. Of these records, 207 where used for analysis of recent (1983-2012) and 107 for long-term (1963-2012) trends respectively (Fig. 2.1; see Section 2.2 for selection criteria).

The bulk of the catchments (189) are part of the Norwegian Hydrological Reference Network (HRN) compiled by the Norwegian Water Resources and Energy Directorate (NVE) (Fleig et al., 2013). These catchments were selected according to the criteria set out by Whitfield et al. (2012). The catchments from the HRN are pristine or near-natural catchments with less than 10% affected by basin development and absent of significant hydrological alterations, i.e. regulations, diversions and water use (Fleig et al., 2013). The remaining catchments were added by Vormoor et al. for their analysis, and are more affected by land-use, but still unaffected by major hydrological alterations. Due to the pristine or near-natural character of the watersheds any hydrological changes can be attributed to a changing environment rather than direct anthropogenic influences. However, it should be noted that the impacts on streamflow generation as a result of gradual afforestation, which has taken place during the past 100 years due to decreasing grazing in Norway, are currently unquantified (Vormoor et al., 2016).

2.1.2. SeNorge data

In addition to the streamflow records, daily rainfall, snowmelt and temperature data were analysed to identify the source of the change in the hydrological regime. These data were extracted from daily 1 km gridded maps available at www.seNorge.no starting from 1957 to the present. These gridded SeNorge maps are spatially interpolated from daily observations of 24-hour mean temperature from about 150-230[2] stations, and 24-hour accumulated precipitation from about 400-650[1] stations using a bayesian approach. Snowmelt, i.e., the runoff generated by melting snow, is dependent on snow water equivalent (SWE), air temperature and snow wetness, and calculated using a snow model derived from the HBV model (Engeset, 2016).

2.2. Assessing suitability of data for trend analysis

Due to the need to avoid the interference from latitude dependent trends or the Scandinavian Mountains, the data was divided into subsets according to the six runoff regions of Norway (Fig. 2.1; see Section 1.4). This also gave the opportunity to compare how the regions Vestlandet and Østlandet, with similar latitudinal range but different hydro-climatological conditions, in addition to having catchments representing the four prevailing hydrological regimes (see Tab. 1.2).

The record length and quality were visualised (Fig. 2.3), as well as elevation range (Fig. 2.2), to aid in the catchment selection process. The aims of the catchment selection were to have; (1) good spatial and altitudinal coverage in each region, (2) sufficient data quality and record length for trend analysis, and (3) sufficient catchments in each runoff region. A catchment with more than 10% missing days in the chosen period, i.e. 30 years or 50 years, was elimin-

[2] The number of stations varies depending on the year (see Engeset, 2016).

ated from the selection. For some catchments either rainfall or snowmelt data were missing, and these were therefore removed as well. The main reason for catchments being eliminated from analysis, was missing SeNorge records for the recent period, and too much missing data for the long-term (Fig. 2.3).

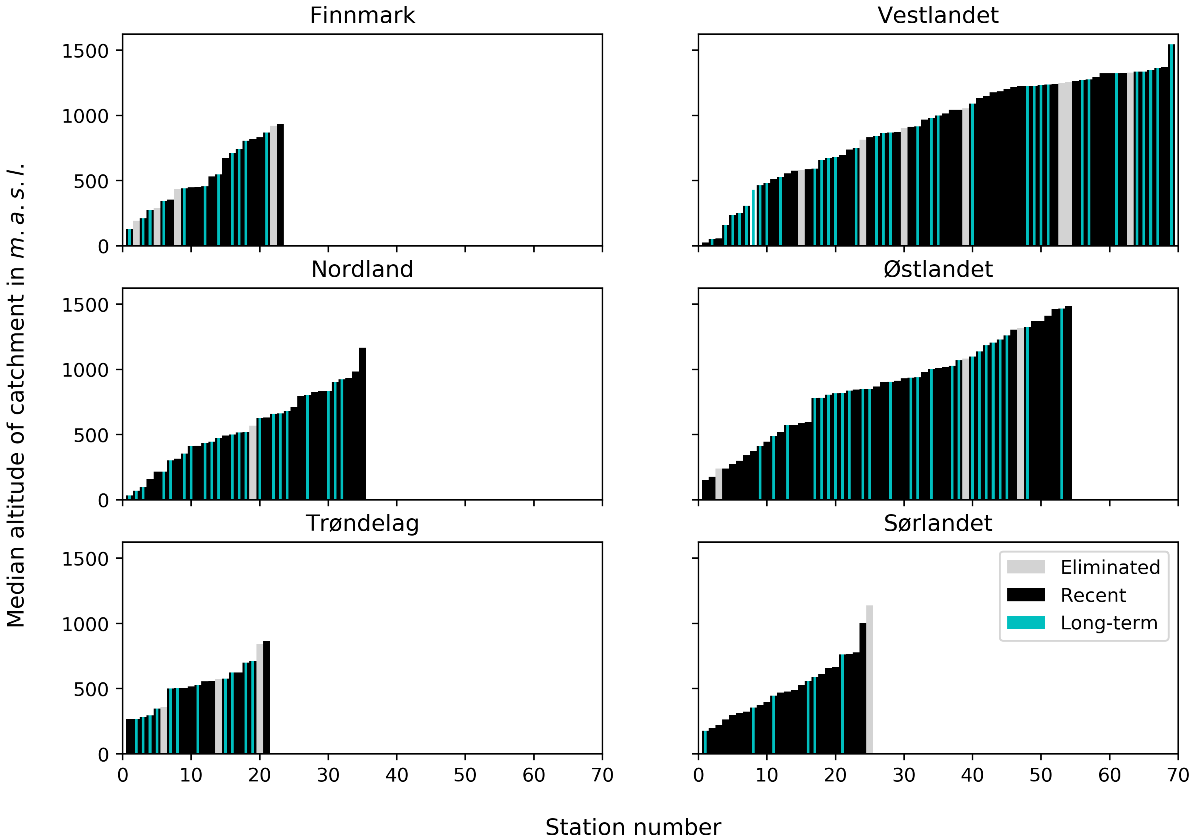

Figure 2.2: *Altitude distribution of selected and eliminated catchments for each of the six runoff regions*. All catchments used in the long-term trend analysis where also used in the analysis of recent trends, with the exception of one catchment in Vestlandet.

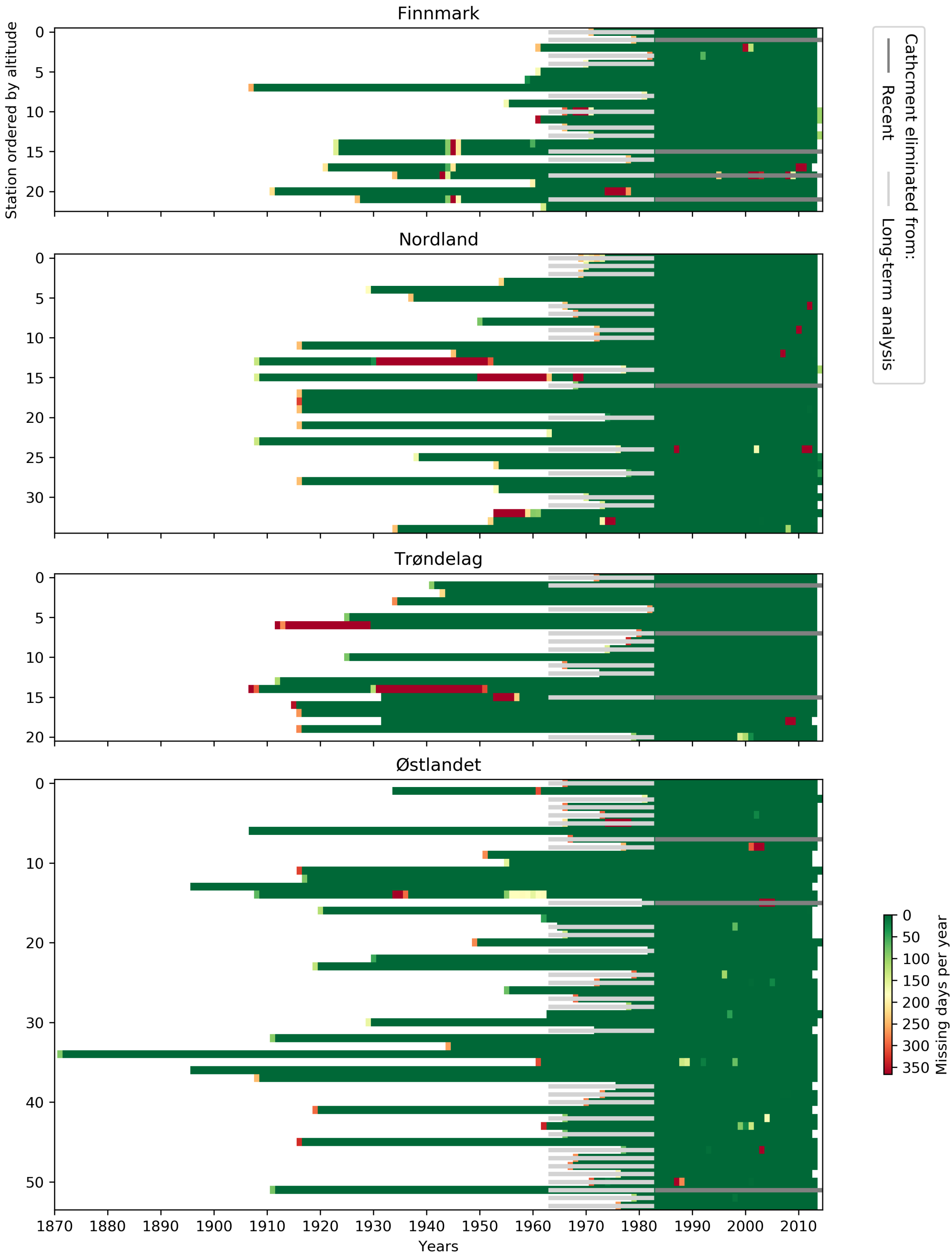

Figure 2.3: *Overview of streamflow records for catchments in each runoff region ordered by altitude*. Data quality is indicated by number of days per year with missing data. Catchments marked in dark/light grey were eliminated from the short-/long-term trend analysis.

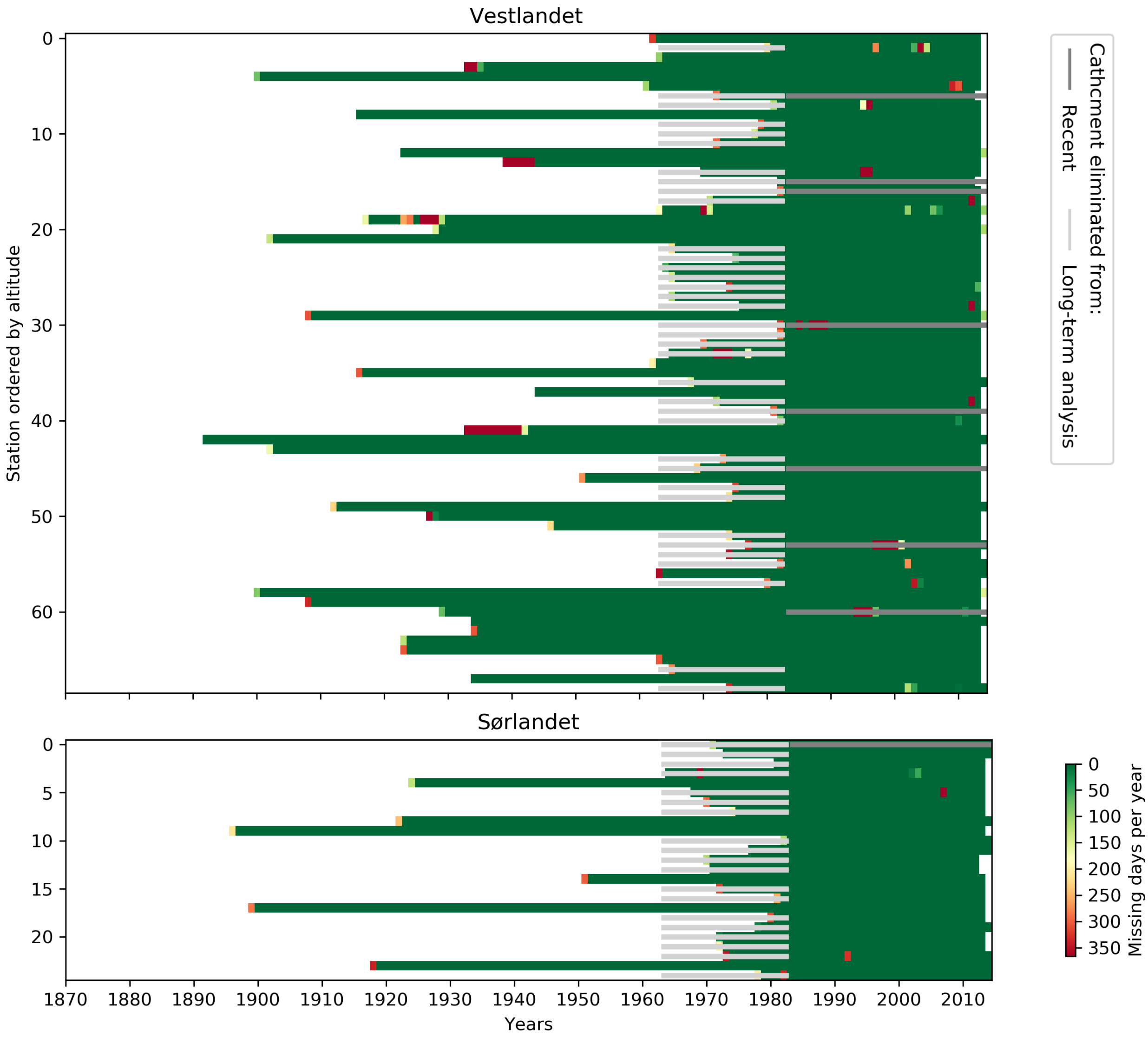

Figure 2.3 (continued)

3. METHODS[3]

This chapter first introduces the trend analysis procedure employed for analysing both annual and daily resolved trends (Section 3.1). Then, the calculation of annual trends is described in Section 3.2, and, finally, the daily resolved trend analysis approach is explained in Section 3.3. To enable direct comparison with the findings of Kormann et al. (2014) the methods were changed as little as possible, a notable exception being the MA filtering (see Section 3.3).

3.1. Statistical trend analysis methods

The Mann-Kendall (MK) test is widely applied in hydrology for the detection of significant trends (Burn et al., 2012). It is a non-parametric trend test that applies to monotonic trends only (Chandler and Scott, 2011) and assumes that there is no autocorrelation present in the time series (Helsel and Hirsch, 2002). The main advantage of the MK test is the non-parametric nature of the test, meaning it does not require assumptions about the underlying statistical distribution of the time series, which makes it robust to the skewed distributions that are characteristic of environmental data (Skaugen et al., 2012).

The MK test statistic is given by

$$S = \sum_{i=1}^{n-1} \sum_{j=i-1}^{n} Sgn(X_j - X_i) \tag{1}$$

where X_j and X_i are sequential values in and n the length of the time series to which the test is being applied, and

$$Sgn(x) = \begin{cases} +1 & & x > 0 \\ 0 & if & x = 0 \\ -1 & & x < 0 \end{cases} \tag{2}$$

The *Sen's Slope* (SS) estimator (Sen, 1968) was used to estimate the magnitude of a detected significant trend and is defined as the median of the slopes for each $X_j - X_i$ pair.

A prewhitening procedure was performed to correct for any detected autocorrelation in the time series to which the MK test was applied to. It should however be noted that there is a chance of a portion of the trend in the time series being removed with prewhitening (Yue et al., 2002). Especially extreme hydrological and climatological conditions tend to be underes-

[3] All analysis was run with Python 3.7 (Python Software Foundation, 2018) and python package *trend* (Hodson, 2018; available at https://github.com/USGS-python/trend) was used when computing the MK test and SS estimator. The code can be viewed at https://github.com/skalevag/HydroTrendsNorway

timated when a time series is prewhitened. As extreme events are not the focus of this study, implementing a prewhitening procedure as part of the trend analysis was deemed necessary to ensure that significant trends were not detected where none exist. Moreover, as the original data has been modified by a MA filter, meaning the magnitude of any extreme events have been altered.

To account any autocorrelation present in the data the following prewhitening procedure was used: First, the time series was tested for the presence of a significant ($\alpha = 0.05$) lag-1auto-correlation using the Ljung-Box test (Ljung and Box, 1978). If significant autocorrelation was detected, the time series was then modified using a prewhitening procedure based on Wang and Swail (2001)

$$W_t = \frac{Y_t - cY_{t-1}}{1 - c} \tag{3}$$

where Y_t is the original time series, c the autocorrelation coefficient, and W_t the modified time series.

Hydro-climatological data from sites in the same geographical area often cross-correlated (Wilks, 2006; Renard et al., 2008). This cross-correlation, i.e. correlation between sites, becomes an issue when multiple statistical tests are being simultaneously applied (Wilks, 2006). Calculating the field significance can determine whether the trends detected are significant at the field level, by determining the percentage of trends that are expected to be detected purely by chance (Burn et al., 2012).

The field significance was calculated using the bootstrap approach after Burn and Hag Elnur (2002), which is considered a robust tool for determining the field significance (Renard et al., 2008). This method removes any temporal structure in the data by randomly selecting years to be included in the resampled data set. However, as the values from all catchments in a given year is included, the cross-correlation in the original data is preserved. The data is resampled NS times[4] and the MK test applied to each resampled time series. Next, the percentage of significant trends in each resampling is calculated to create a distribution. The local significance level α_{local}, i.e. the significance level for the MK test, remains unchanged when applied to the resampled data, whereas the field, or global, significance level α_{field} determines the critical value p_{crit}, which is defined as the $1 - \alpha_{field}$ quantile of the distribution. If the percentage of

[4] For reasons of time constraint *NS* was limited to 400

significant trends found exceeds p_{crit} then the results will be considered significant at the α_{field} level, likely not caused by randomness, and not significantly impacted by cross-correlation.

3.2. Annual trends

An analysis of annual trends was included to enable better comparisons with other studies, as annual trends, especially annual streamflow, is a commonly analysed variable in hydrological trend studies as it represents the overall water availability (Burn et al., 2012). Moreover, changes in the annual water balance will affect daily trends. The annual trend analysis was performed on total annual runoff, snowmelt and rainfall, in addition to evapotranspiration ET, which was estimated from the water balance equation:

$$P - ET - Q \pm \Delta S = 0 \tag{4}$$

The change in interannual storage ΔS was assumed to be zero. However, in glaciated catchments, this assumption is incorrect. Therefore, ET was not calculated in catchments where more than 10 % is covered by glaciers. The total precipitation P was calculated from the sum of annual rainfall and snowmelt, and annual runoff Q from the daily streamflow records. The annual values were summed for each Norwegian hydrological year, i.e. from September 1st to August 31st, meaning that the analysis period was slightly adjusted to 1982-2012 and 1962-2012, to ensure full data coverage for 30 and 50 years. If more than 10% of a hydrological year were missing from the daily streamflow, rainfall or snowmelt time series, then the annual value of that variable was not calculated for that year. Since ET is based on these variables then ET for the corresponding year was also not calculated. Trends in mean annual temperature were also calculated.

3.3. Daily resolution trend analysis

The daily resolution trend analysis approach used in this study is based on the approach developed by Kormann et al. (2014; 2015) for the analysis of elevation dependent trends in the Alps. This approach is similar to the seasonal MK test (Hirsch et al., 1982), which analyses the trends of each season or month separately and was designed for monthly resolved trend analysis. The approach of Kormann et al. takes it further to have highly resolved trends, and also gives the opportunity to create trend hydrographs similar to Déry et al. (2009). The method has also been used to analyse elevation dependent temperature trends in the Swiss Alps (Rottler et al., 2018).

Time series for each day of year (DOY) were compiled from the daily data for the chosen time period, i.e. 30 or 50 years, meaning the time series had a length of either 30 or 50 values. These DOY time series where then analysed according to the trend analysis procedure described in Section 3.1. The results were then aggregated into an array with the trend magnitude of each DOY. Daily trends were calculated for streamflow, rainfall, snowmelt, and temperature (Fig. 3.1).

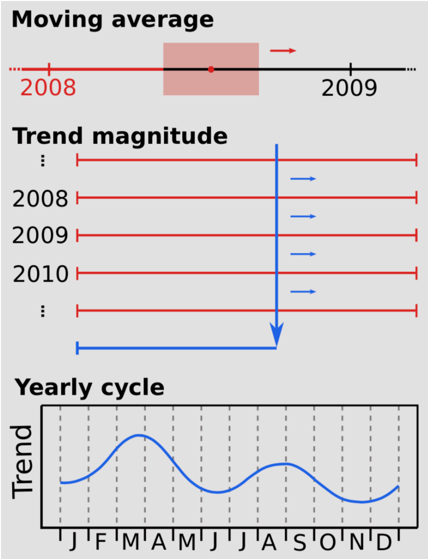

Figure 3.1: *Schematic representation of the daily resolution trend analysis approach*. First, a moving average filter is applied to the daily time series data. Second, the trend significance and magnitude is calculated for each DOY time series (large blue arrow). Third, the individually calculated trend magnitudes are aggregated into the yearly cycle of daily resolved trends (Source: Rottler et al., 2018).

The interannual variability of the DOY time series is high, making the MK test less capable of detecting trends. Consequently, the original daily data was smoothed using a moving average (MA) filter, before extracting the DOY time series. This idea stems from Kim and Jain (2010), who used a 3-day MA filter, and was taken up by Kormann et al. (2014), who used a 30-day MA (30dMA) filter to further lower the variability. A 5-day average was judged by Déry et al. (2009) to be sufficient to obtain similar hydrological responses from watersheds of varying sizes, in addition to minimising the effects of transient storms on precipitation fluctuations (Whitfield and Cannon, 2000). One of the aims of this study was to examine the effect of different MA filters, therefore 5-day, 10-day and 30-day MA filters were used to smooth the data before applying the trend procedure. The resulting trend magnitude array for each DOY has clusters of significant trends. To enable a comparison between different catchments, the timing of these trend clusters were calculated. The *trend timing* metric (Kormann et al., 2014) is designed to detect shifts in the trend clusters detected by the MK test and SS estimator by calculating the central moment of a cluster (Fig. 3.2). The trend timing is given in DOY and defined as

$$trend\ timing = \frac{\sum_{i=p}^{q} DOY_i \cdot trend\ magnitude_i}{\sum_{i=p}^{q} trend\ magnitude_i} \tag{5}$$

where p is the first and q the last DOY in the trend cluster.

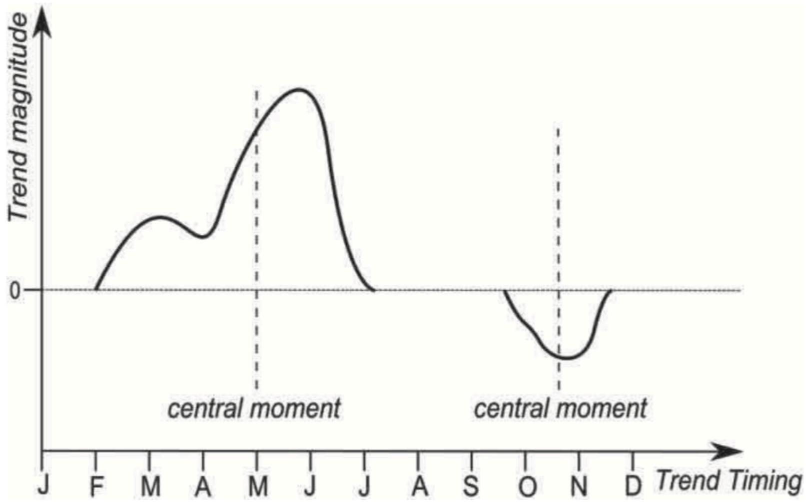

Figure 3.2: *Schematic illustration of trend timing*, i.e. the central moment of a cluster magnitudes of significant daily trends (trend cluster) (Source: Kormann et al., 2014).

4. RESULTS

This chapter presents the detected trends in annual and daily streamflow, snowmelt, rainfall, and temperature. First, the annual trends and their altitude dependence are briefly described (Section 4.1), followed by a more thorough description of the daily 10dMA[5] trends in streamflow, snowmelt and rainfall, both recent and long-term (Section 4.2) and their altitude dependence (Section 4.2.1). Next, results from the runoff regions Vestlandet and Østlandet are presented in more detail (Section 4.2.2) and detailed results from a few selected catchments in these two runoff regions are presented (Section 4.2.3).

4.1. Annual trends

Annual hydrological changes show a variety of significant trends (Fig. 4.1; Fig. 4.2), of which most are field significant in the long-term period, and only a few in the recent (Fig. 4.3). The detected significant *rainfall* trends are mainly all positive in both the short- and long-term period, although for the recent period only the trends in Finnmark are field significant. Recent *snowmelt* trends are all negative, but only field significant for Nordland where a large percentage of the catchments show significant trends. In contrast, annual snowmelt increased in a few high altitude and latitude catchments in 1963-2012, while the spatial pattern of positive trend remained the same. Increases in annual *runoff* are concentrated in the northern parts of Østlandet and Vestlandet. These trends are only field significant in the long-term period. Long-term e*vapotranspiration* trends are mainly positive, while recent trends are largely negative in Nordland and Vestlandet.

The annual trends in runoff, snowmelt, rainfall, and evapotranspiration are not altitude dependent, except snowmelt trends in Nordland, which show significant negative correlation with altitude. All significant mean annual temperature trends are positive (Fig. 4.4), showing about the same rate of warming for both periods, $0.02 - 0.09\ ^{\circ}C\ \ yr^{-1}$ in 1983-2012 and $0.02 - 0.08\ ^{\circ}C\ \ yr^{-1}$ in 1963-2012. Nordland, Sørlandet and Vestlandet show no significant correlation between altitude and annual temperature trends in both periods. Trøndelag and Finnmark show a positive correlation between altitude and temperature trend magnitude, while Østlandet, located east of the mountains, exhibits negative correlation (only significant in long-term period for Østlandet and Finnmark).

[5] Note that only 10dMA daily trends are presented in this chapter. The effect of different MA on the daily trends is discussed in Section 5.4.1

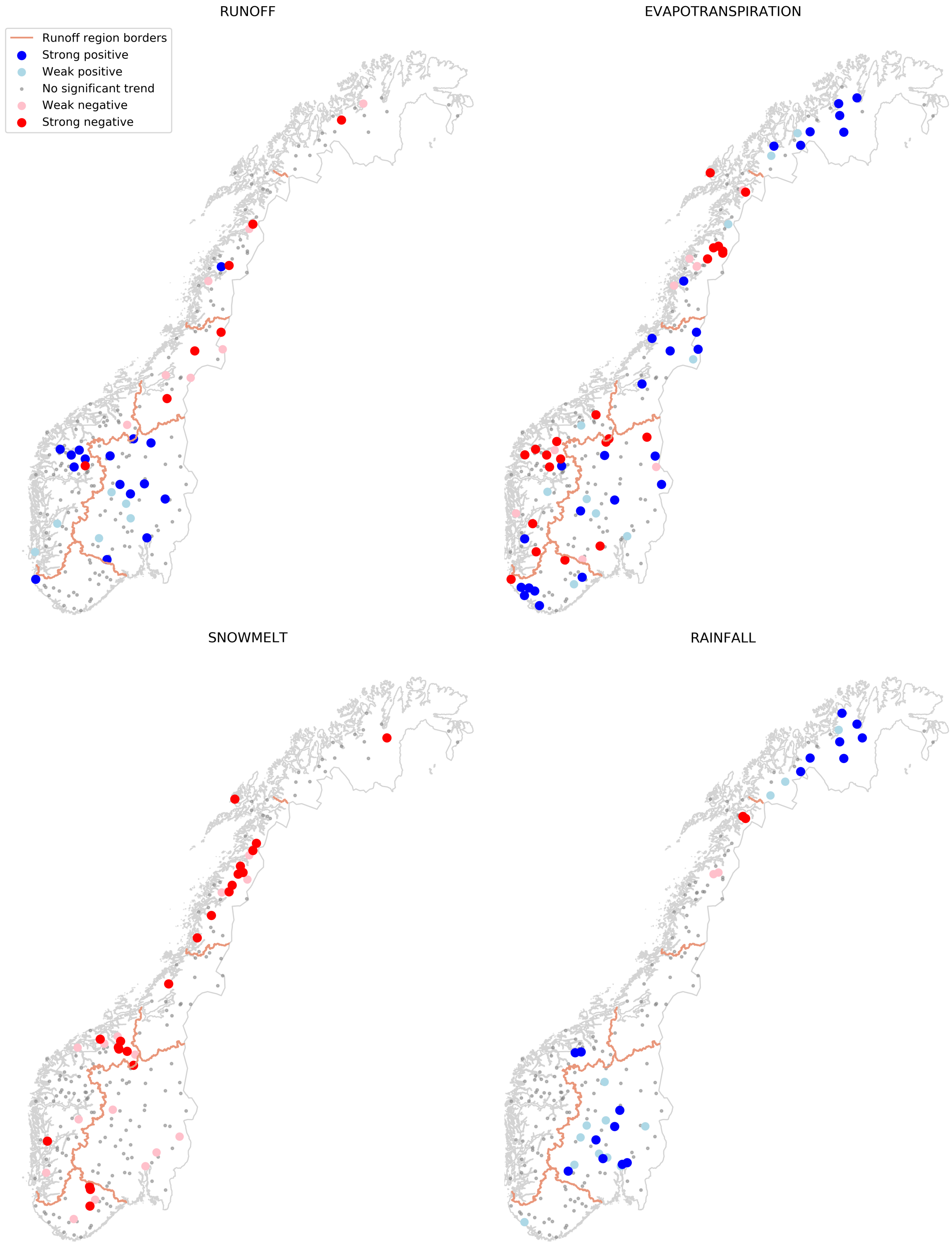

Figure 4.1: *Recent trends (1982-2012) in total annual runoff, evapotranspiration, snowmelt and rainfall*. Catchments with a significant positive or negative trend at two significance levels, $\alpha_{weak} = 0.1$ and $\alpha_{strong} = 0.05$, are indicated by colour. Refer to Section 1.4 for runoff region names and characteristics.

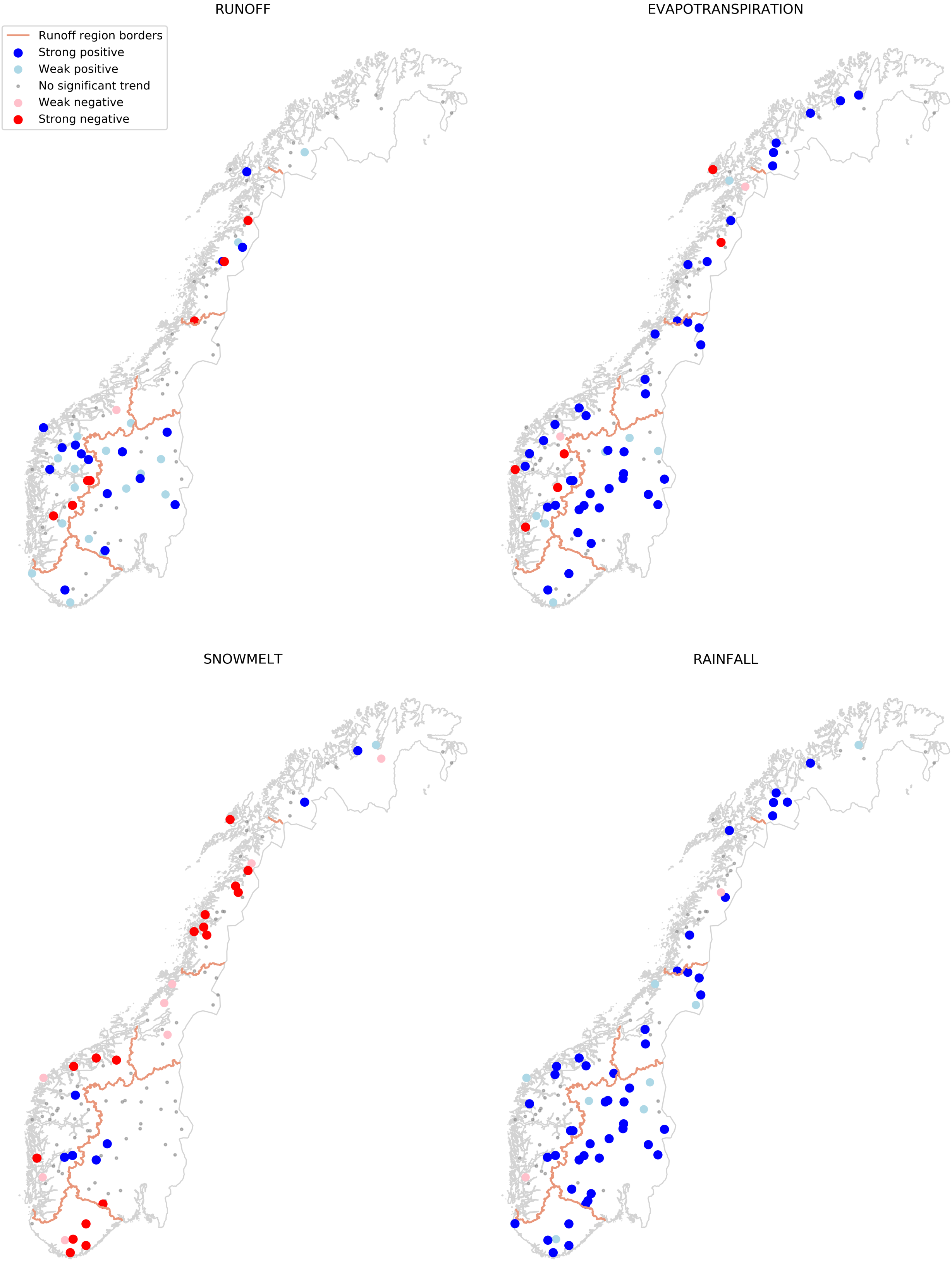

Figure 4.2: *Long-term trends (1962-2012) in total annual runoff, evapotranspiration, snowmelt and rainfall*. Same as Fig. 4.1.

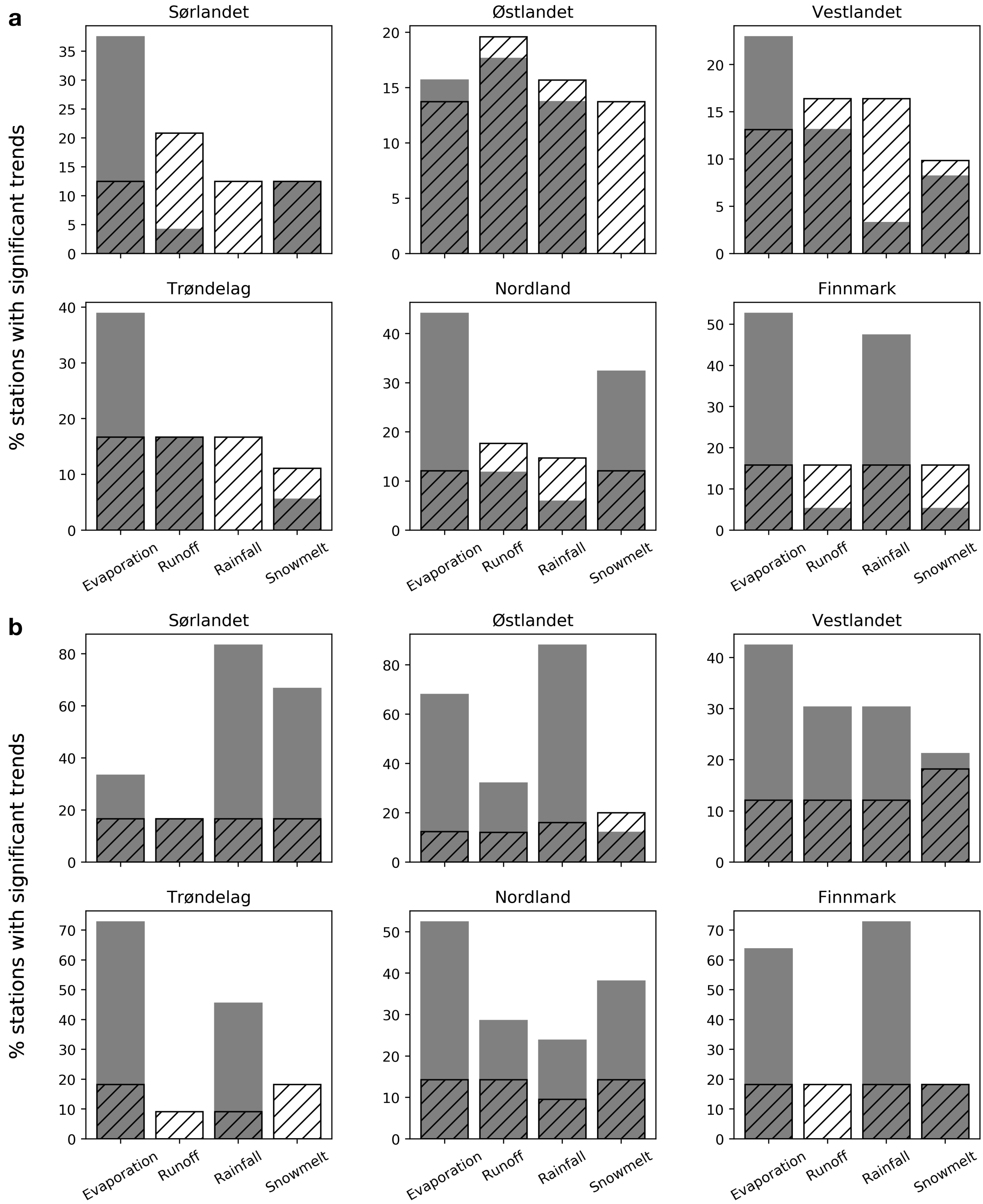

Figure 4.3: *Field significance of annual trends for the (a) recent (1983-2012) and (b) long-term (1963-2012)*. Where the percentage of significant trends (grey) at the local level $\alpha_{local} = 0.05$ exceeds p_{crit} (hatched), the detected trends are considered significant at the field significance level $\alpha_{local} = 0.1$.

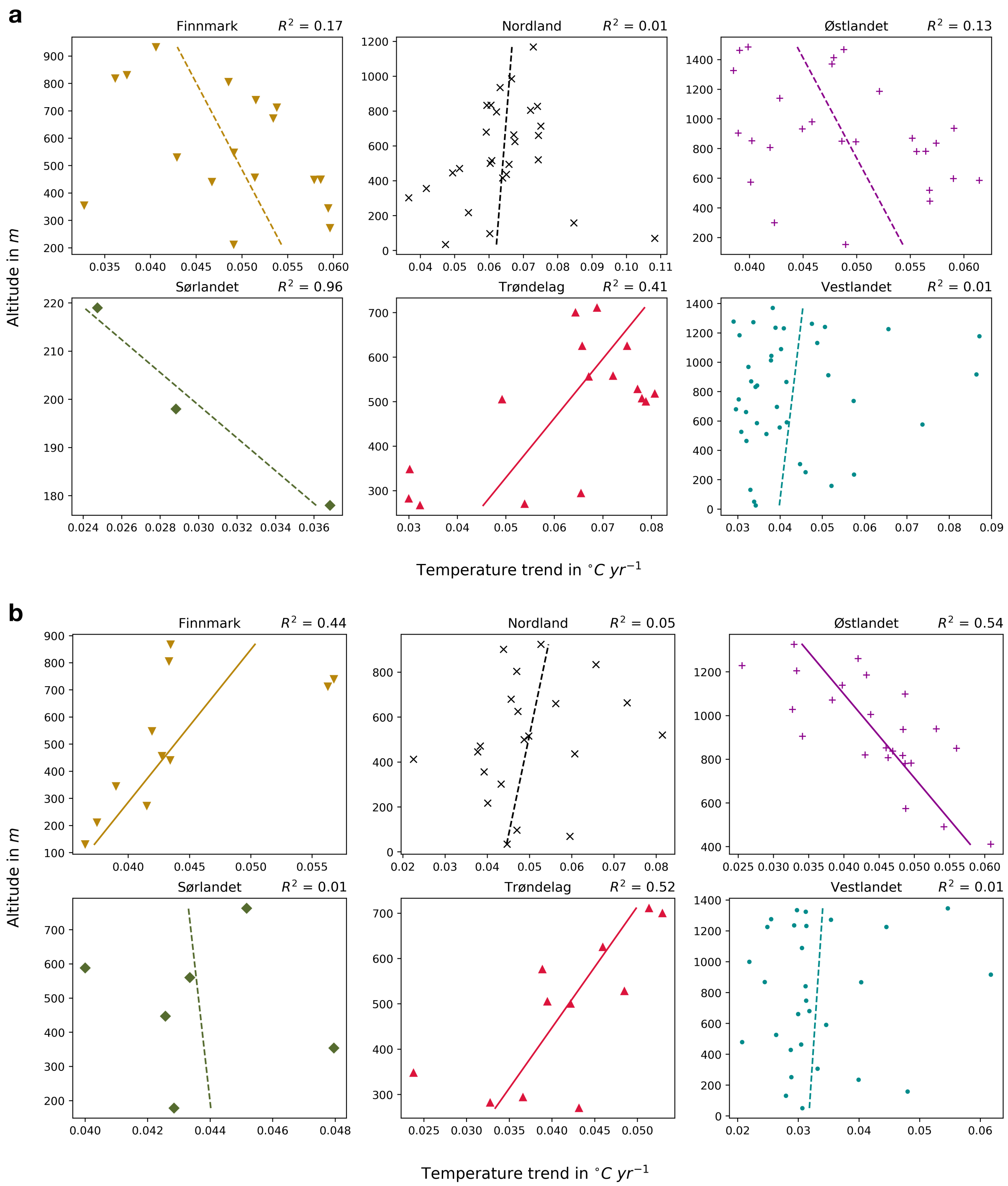

Figure 4.4: Correlation between annual temperature trends and altitude for catchments with significant trends ($\alpha = 0.05$) calculated for the recent (a) and long-term (b) time periods. Lines calculated by linear ordinary least squares regression (OLS), dashed line represents non-significant correlation ($\alpha = 0.05$).

4.2. Daily trends

Streamflow changes overall have a small rate of change relative to the daily streamflow, and large parts of the year show no significant changes. A recent streamflow change that is consistent and field significant across all catchments, is an increase in March-April streamflow (Fig. 4.5a), which is concurrent with positive snowmelt trends (Fig. 4.6a), and significant warming (Fig. 4.8a). In Finnmark, Nordland, Trøndelag and Vestlandet this positive streamflow trend coincides with positive rainfall trends (Fig. 4.7a). Following the positive spring streamflow trend is a period of streamflow reduction, which occurs in late spring in Sørlandet and Østlandet, during summer in Finnmark, Nordland, and Trøndelag, and is not clearly present in Vestlandet. The consistent pattern of positive to negative daily trends during spring in both recent and long-term streamflow and snowmelt trends is indicative of a shift in snowmelt to earlier in the year.

Some catchments in Vestlandet show a strong increase during summer, which is not explained by changes in snowmelt or rainfall, but occurs in a period of significant warming. These changes are likely caused by increased glacial melt (see Section 4.2.3). Østlandet and Sørlandet experience similar positive streamflow changes during summer, but these are likely caused by increases in rainfall (Fig. 4.7a). Vestlandet, Trøndelag and Nordland show similar patterns of sporadic positive recent rainfall trends January-June, followed by intermittent negative trends. These positive trends for the first 6 months of the year is also evident in the long-term trends (Fig. 4.7b).

The long-term hydrological trends (Fig. 4.5-8b) largely reflect the patterns of the recent trends, except for having a clearly lower trend magnitude and larger parts of the year with field significant trends in each region. However, Vestlandet shows clear differences between recent and long-term trends. During summer there are strong positive snowmelt trends above 748 m (Fig. 4.6b) and large increases in winter to late spring streamflow at lower altitude (Fig 4.5b). The latter appears to be driven by positive rainfall trends in the same period (Fig. 4.7b) and is likely caused by a shift from snow to rain in precipitation driven by increasing temperature (Fig. 4.8b).

The temporal pattern of temperature trends was similar for all regions with significant warming of about $0.10 - 0.15\ °C$ per year in three distinct periods: April, July-September and late November (Fig. 4.8). There are some negative temperature trends, but these are not field significant and therefore do not represent a consistent change across the runoff region.

In summary, there are many similarities between the runoff regions, but also great variability, especially in streamflow trends. Notably, a catchment with positive annual trends may contain both positive and negative daily trends. Even in catchments with a non-significant annual trend, significant changes have occurred at the daily scale. Every catchment has experienced changes in all variables which are significant both at the local and field significance level. As shown by the daily snowmelt trends in Østlandet (Fig. 4.6), even where no significant annual trends are detected, there can be consistent significant daily trends.

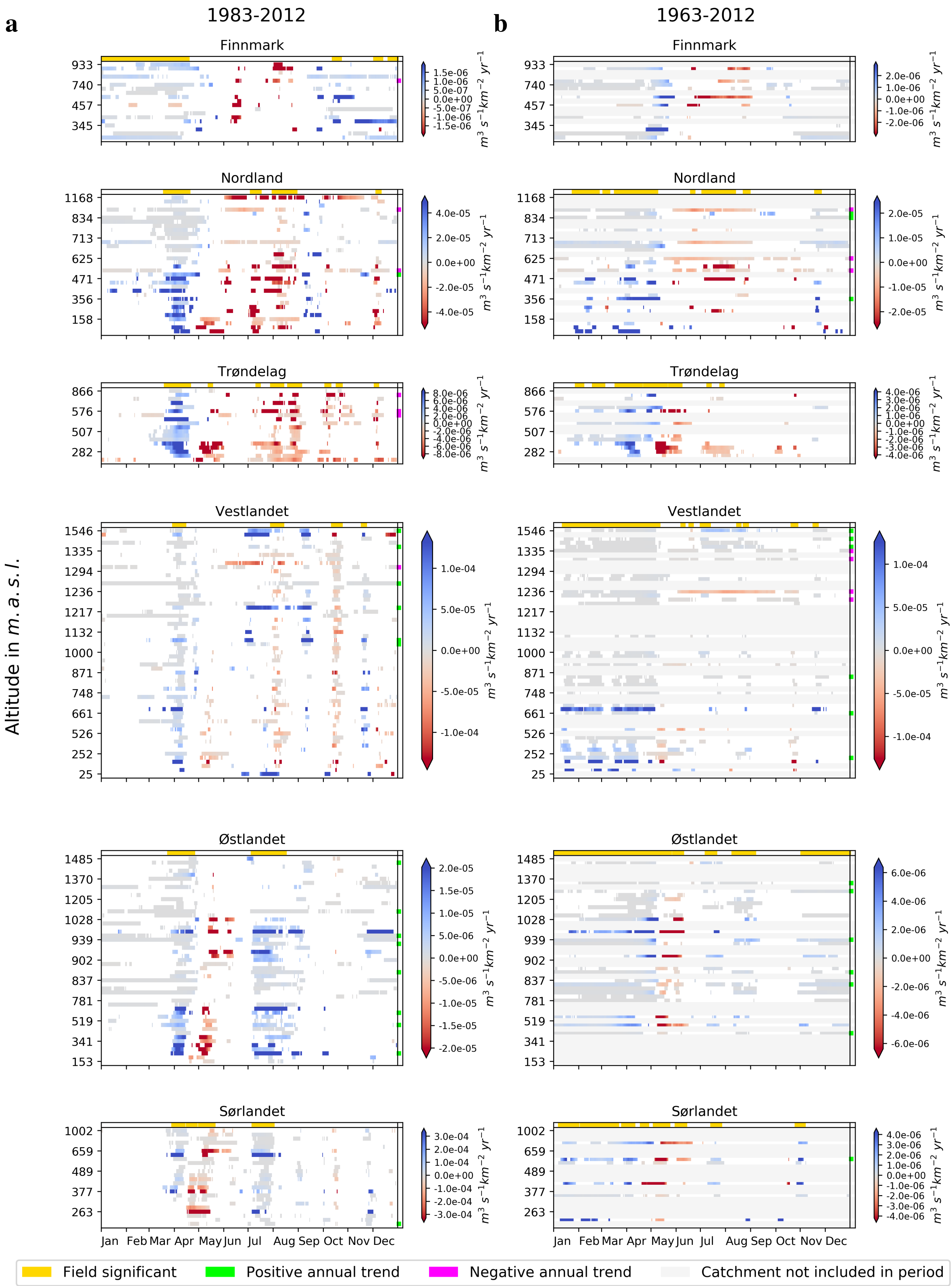

Figure 4.5: *10dMA streamflow trends for (a) recent (1983-2012) and (b) long-term (1963-2012) periods*. Trend magnitude of each DOY time series indicated by colour for each catchment, but only where a significant trend was detected ($\alpha_{local} = 0.1$). Upper bar indicates where a daily trend is field significant ($\alpha_{field} = 0.1$), while right-hand bar indicates a significant positive or negative annual trend ($\alpha = 0.05$). Catchments are ordered according to altitude, which is indicated for every fifth catchment.

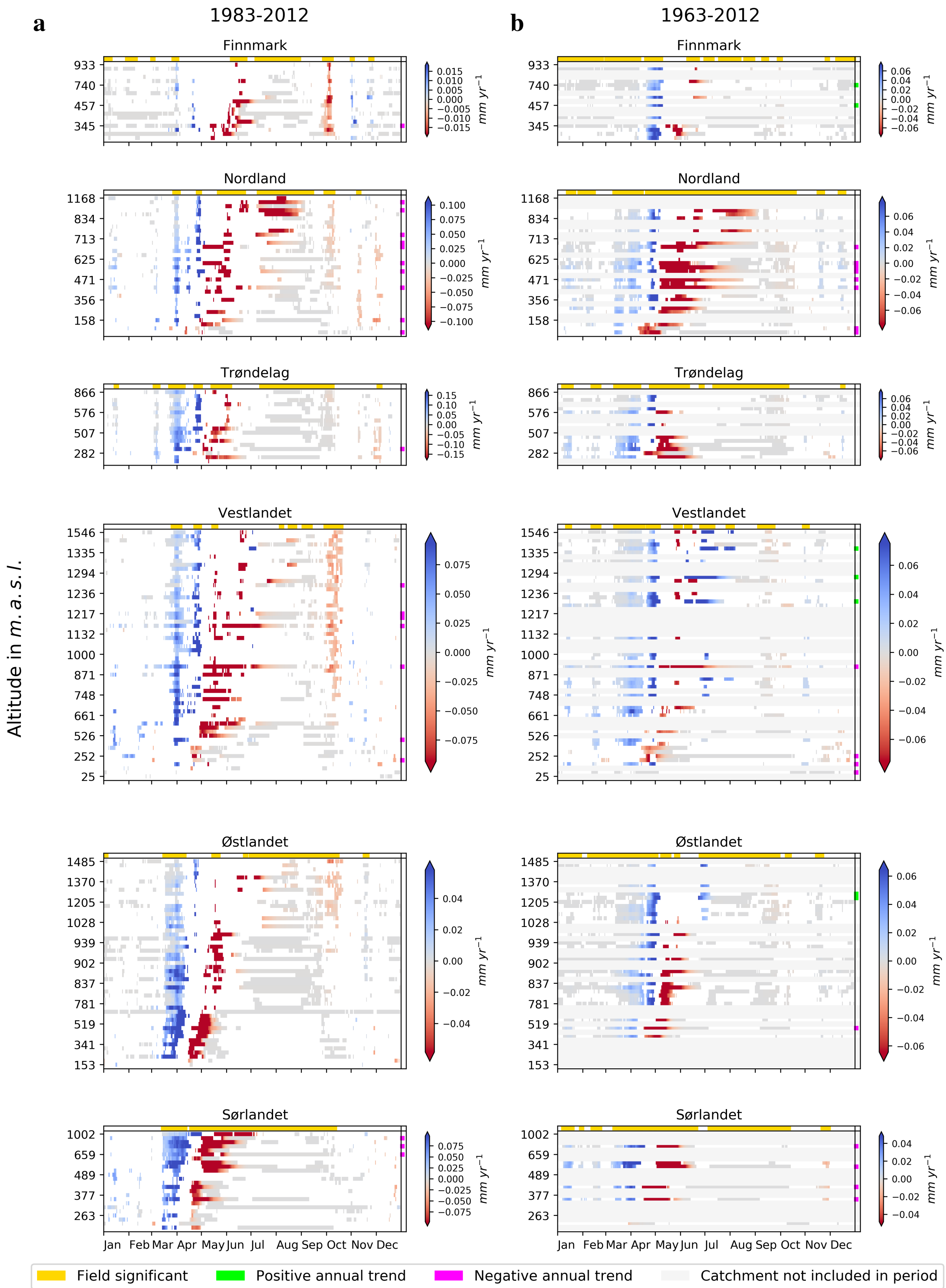

Figure 4.6: *Daily snowmelt trends for (a) recent (1983-2012) and (b) long-term (1963-2012) periods*. Same as Fig. 4.5.

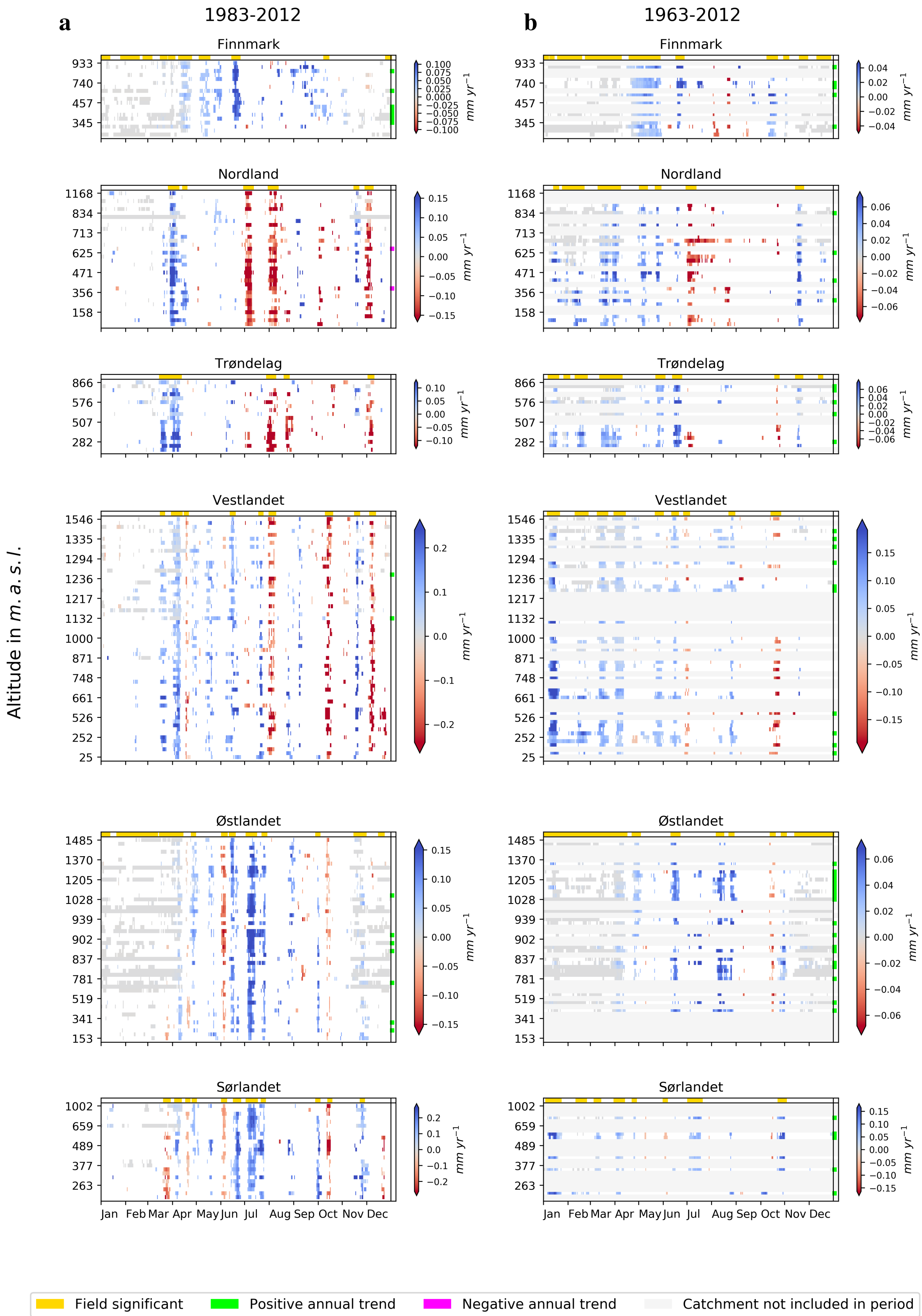

Figure 4.7: *Daily rainfall trends for (a) recent (1983-2012) and (b) long-term (1963-2012) periods*. Same as Fig. 4.5.

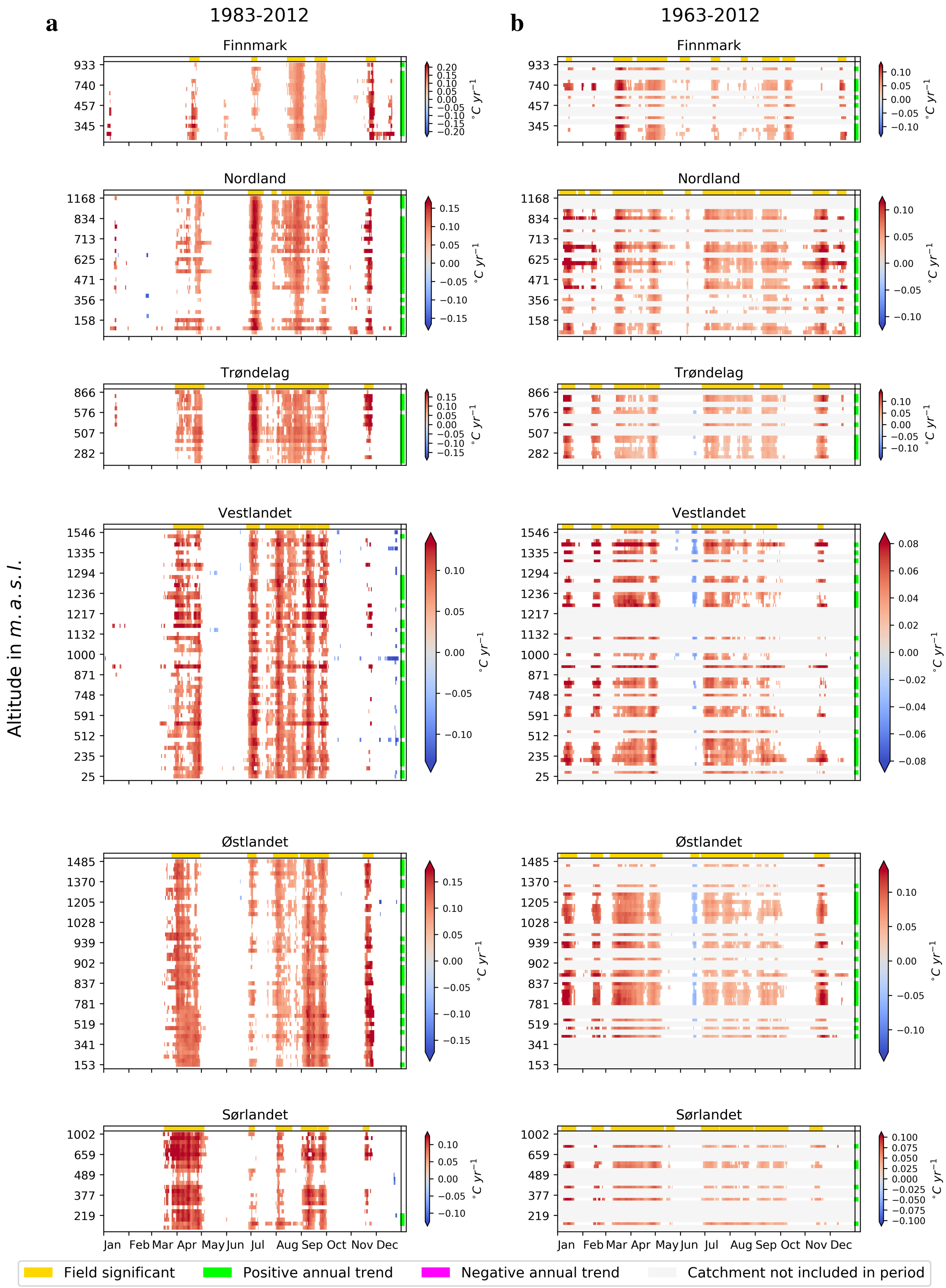

Figure 4.8: *Daily temperature trends for (a) recent (1983-2012) and (b) long-term (1963-2012) periods*. Same as Fig. 4.5.

4.2.1. Altitude dependence

The streamflow trends are very variable with some catchments showing large significant changes and other barely any (Fig. 4.9). Due to this large variability there is no clear correlation between trend magnitude and altitude for streamflow.

In comparison, daily snowmelt trends show clear dependence on altitude in certain regions (Fig. 4.10). Snowmelt trends in Sørlandet show a clear dependence on altitude, with strong negative (5th percentile) trends being negatively, and strong positive (95th percentile) being positively correlated with altitude. Moreover, this pattern is to some extent present in Nordland, Vestlandet (below 1000 m.a.s.l.), and Østlandet (below 500 m.a.s.l.). This pattern is indicative of the previously mentioned shift to earlier snowmelt. However, the negative trends are larger than the positive, meaning there is an overall negative trend in snowmelt, evident by the negative mean daily trend. The relationship between snowmelt trend magnitude and altitude appears to be linear over smaller altitude ranges, e.g. Sørlandet, Nordland, but non-linear for Vestlandet and Østlandet.

Only rainfall trends in Vestlandet display a clear altitude dependence (Fig. 4.11). The relationship appears to be non-linear, with the largest changes occurring at mid-altitude (500-1000 m.a.s.l.). However, the mean trend is close to zero, meaning the changes are largely interannual shifts.

Temperature trends are mainly all positive, with the exception of a few catchments (Fig. 4.12). The mean and weaker (5th percentile) temperature trends appear to be moistly uniform across most altitudes and regions, while the strongest (95th percentile) appear to be positively correlated with altitude in Nordland, Trøndelag and possibly Sørlandet, and negatively in Østlandet.

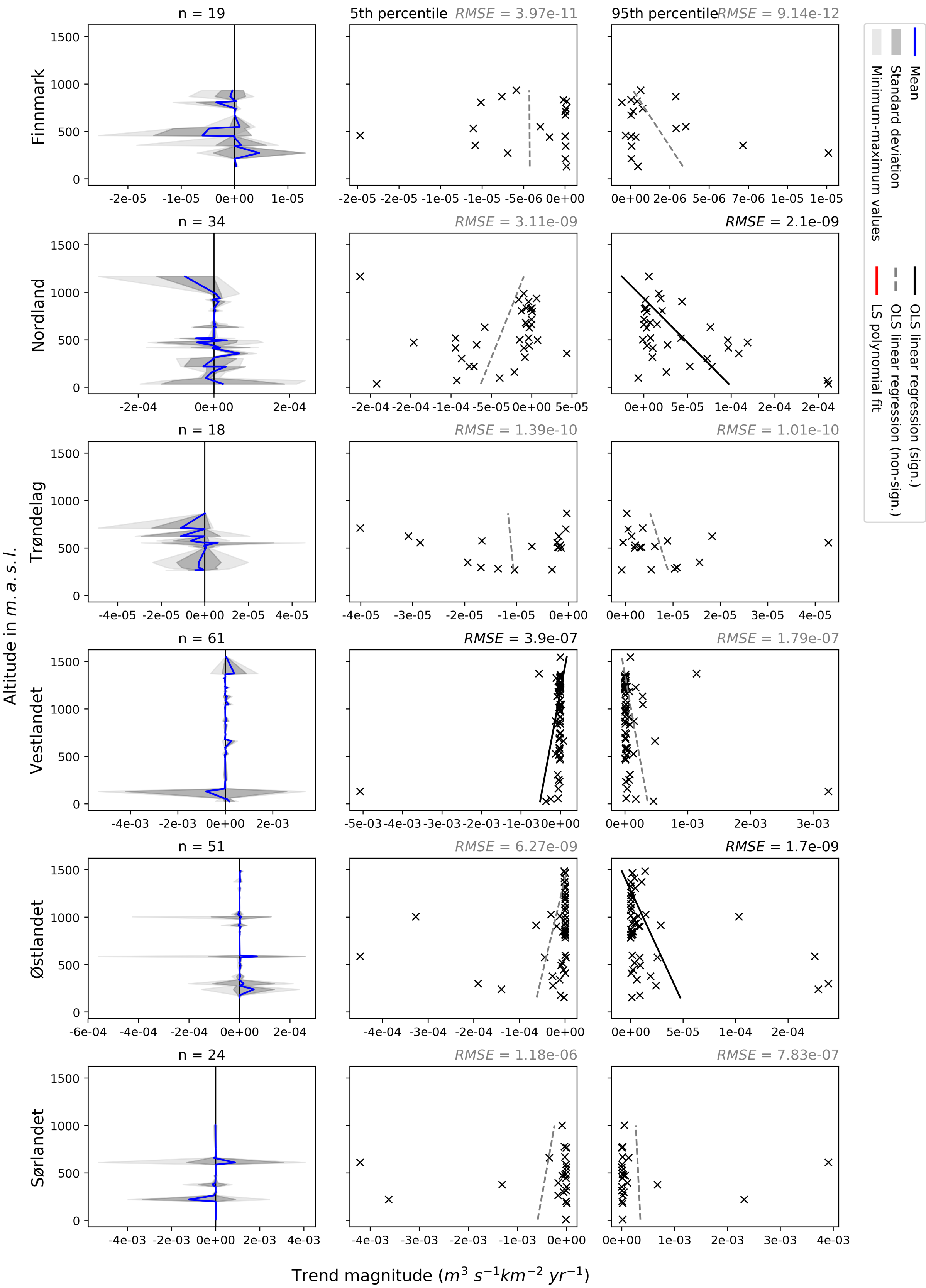

Figure 4.9: *Altitude dependence of significant 10dMA daily streamflow trends* in the recent (1983-2012) period. The first column shows the mean and standard deviation of the daily trends for each catchment plotted against the altitude of the catchment, the second correlation between altitude and the 5th percentile, and the third the correlation between the 95th percentile of significant daily trends and altitude. *n* refers to the number of catchments in each region.

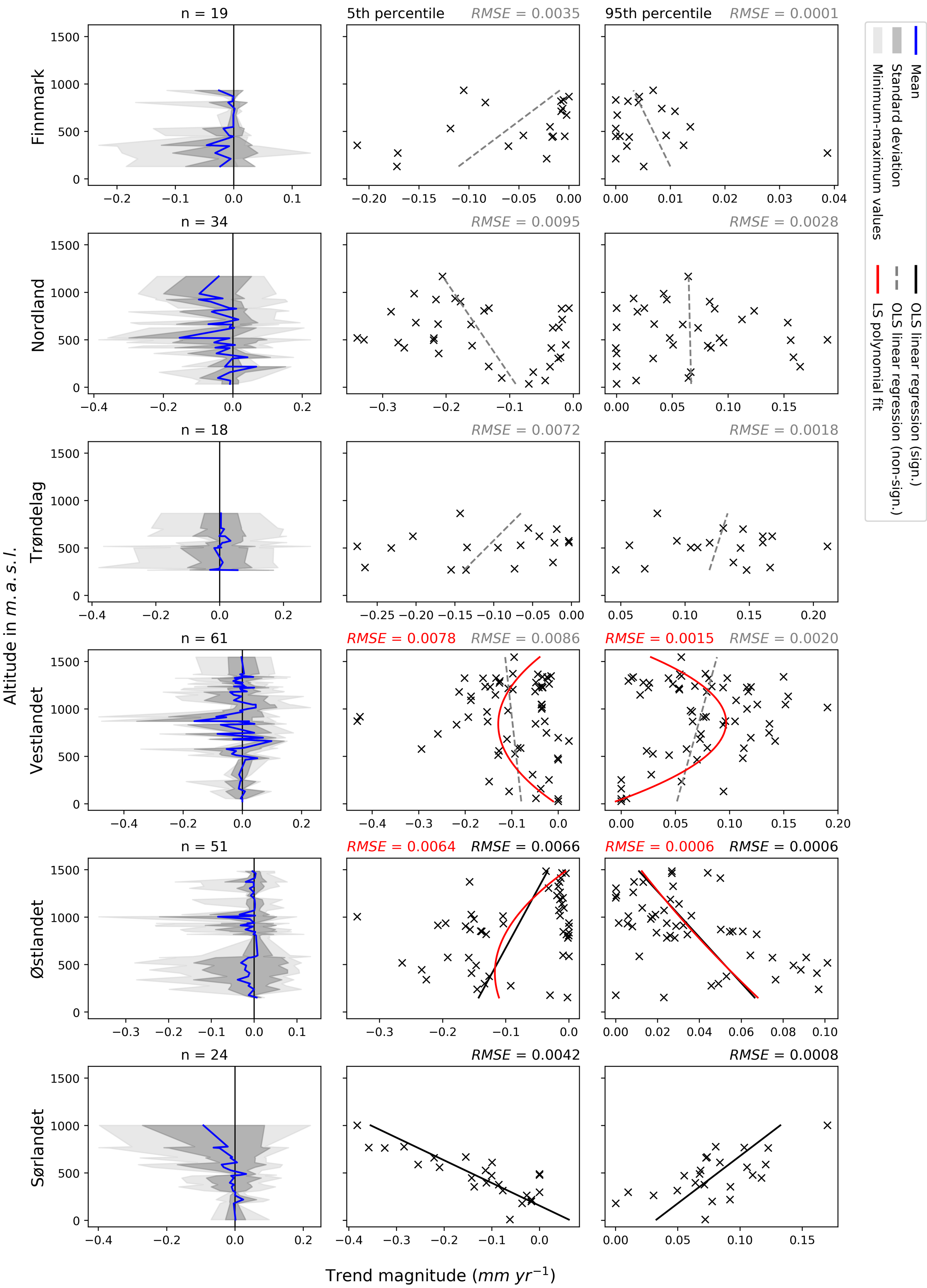

Figure 4.10: *Altitude dependence of significant 10dMA daily snowmelt trends* in the recent (1983-2012) period. See Fig. 4.9.

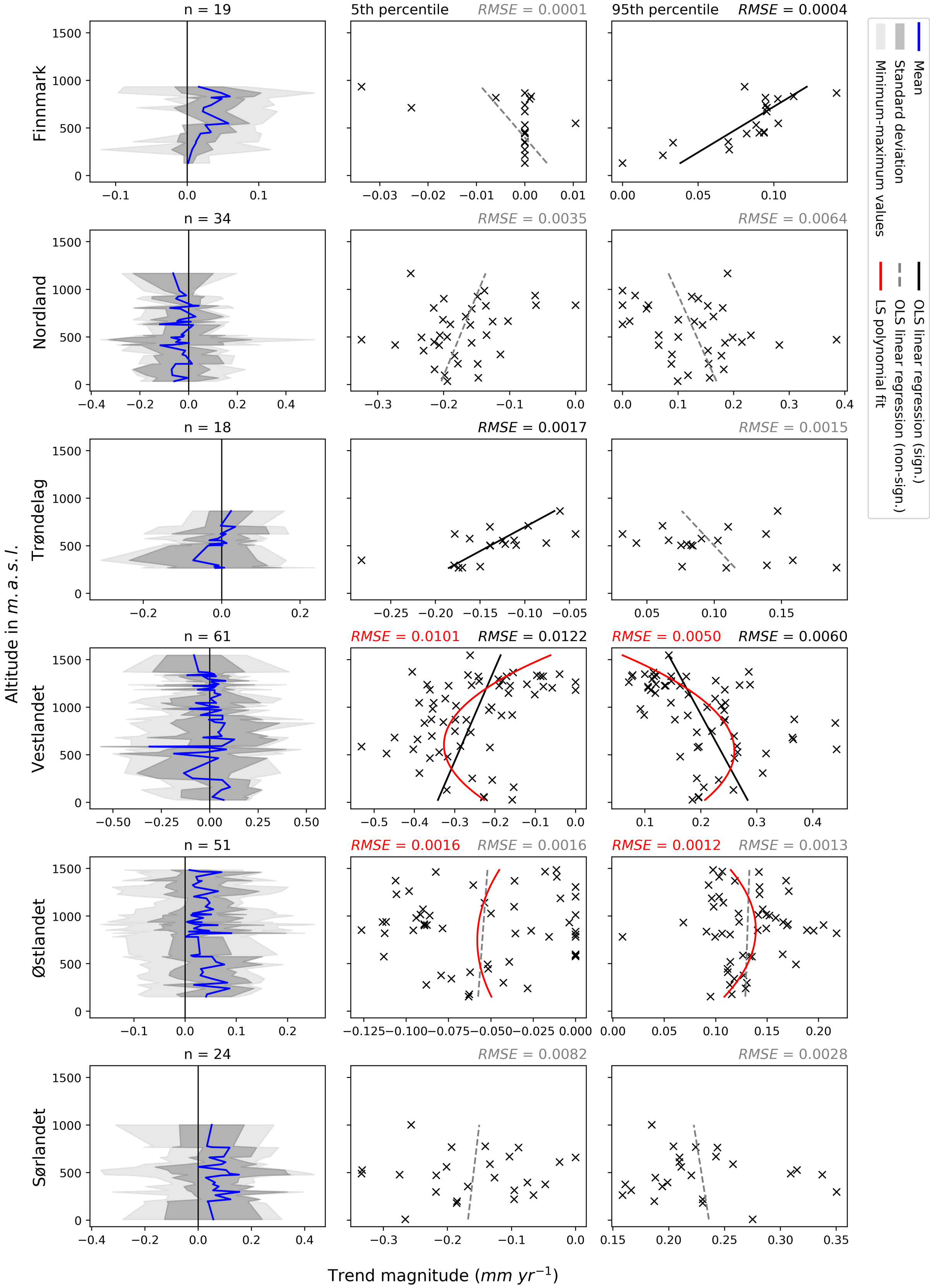

Figure 4.11: *Altitude dependence of significant 10dMA daily rainfall trends* in the recent (1983-2012) period. Same as Fig. 4.9.

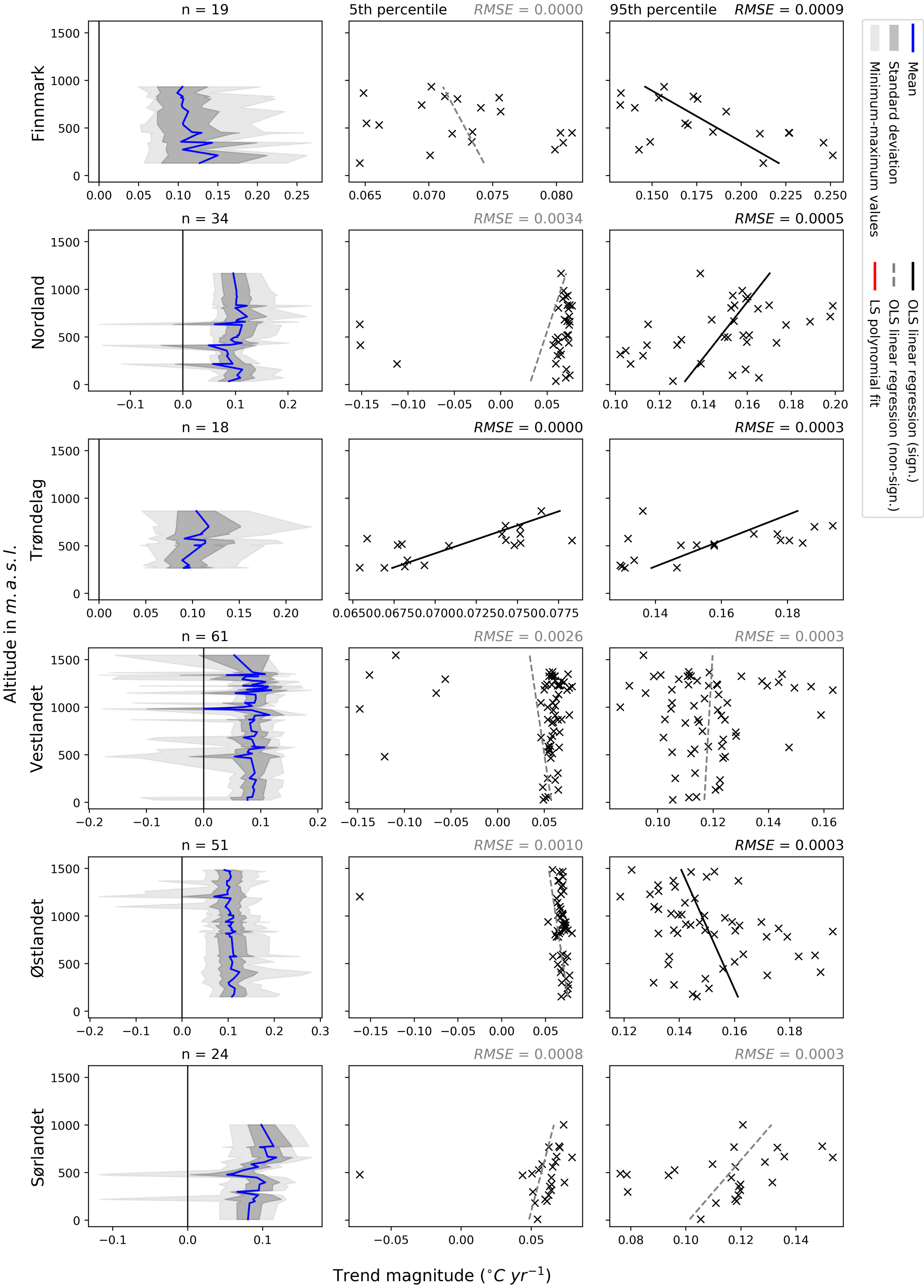

Figure 4.12: *Altitude dependence of significant 10dMA daily temperature trends* in the recent (1983-2012) period. Same as Fig. 4.9.

4.2.2. *Selected runoff regions: Vestlandet and Østlandet*

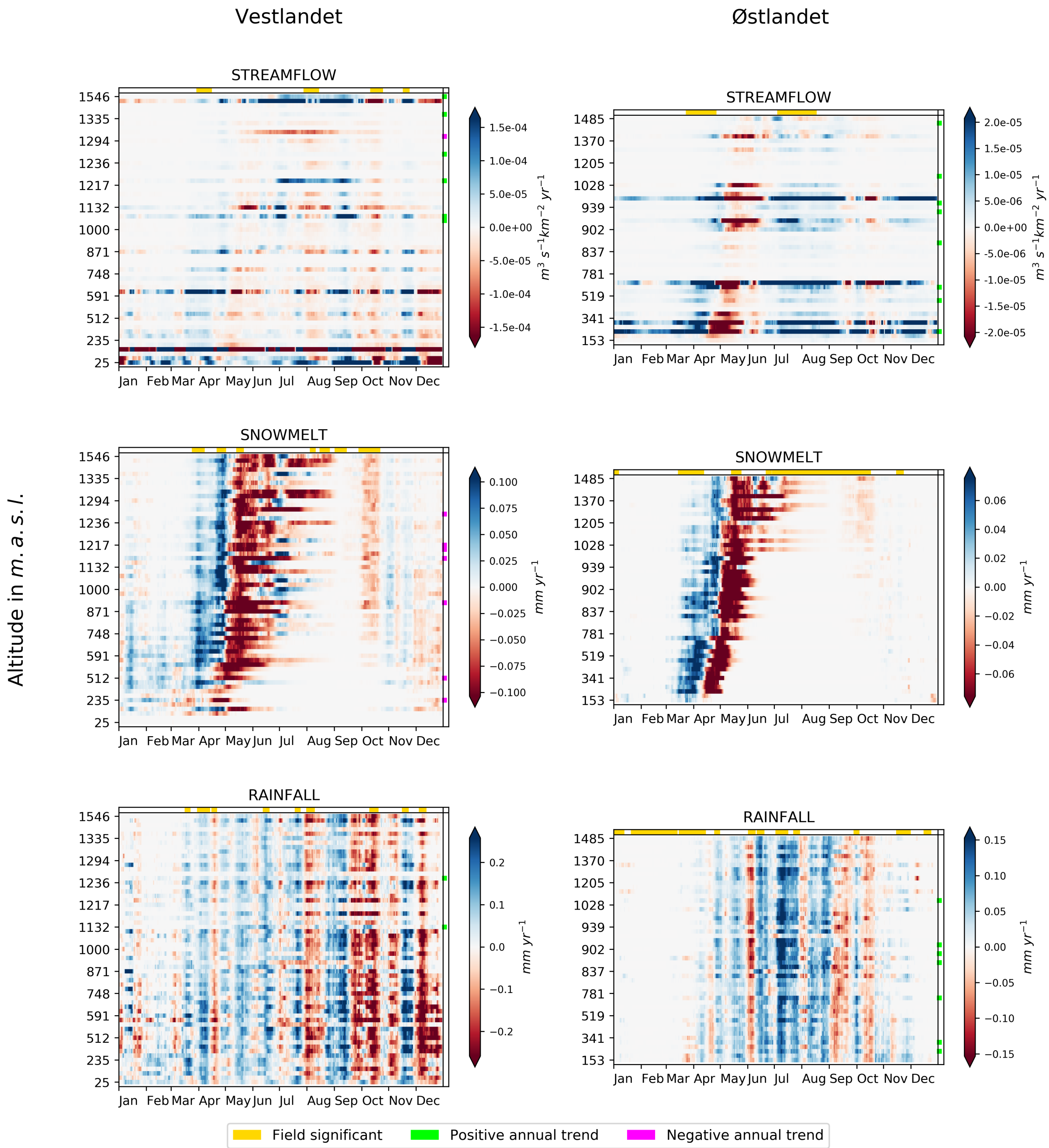

Figure 4.13: *10dMA daily trend magnitudes* (significant and non-significant) in streamflow, snowmelt and rainfall for Vestlandet and Østlandet. Upper bar indicates (yellow) where a daily trend is field significant ($\alpha_{field} = 0.1$), while right-hand bar indicates a significant positive (green) or negative (pink) annual trend ($\alpha = 0.05$).

Examining the trend magnitudes of both significant and non-significant daily trends (Fig. 4.13), a more coherent pattern becomes apparent. The effect of snowmelt change on streamflow can be seen clearly in Østlandet, but is less evident in Vestlandet where streamflow changes have been more affected by rainfall trends (Fig. 4.13).

The distinctive pattern of a shift to earlier snowmelt can be seen more clearly in Fig. 4.13, where both significant and non-significant trends are included (compared with Fig. 4.6). In Østlandet below 1000 m the shift occurs over a period of roughly two months, while in Østlandet above 1000 m and in Vestlandet the period of change is longer, in addition to a field significant negative trend in October. The negative snowmelt trends are larger than the positive trends in both catchments, which could point to a reduced snowpack. The largest changes in rainfall occurred during the summer months in Østlandet (Fig. 4.14) and are mainly positive. In contrast, rainfall changes in Vestlandet are spread evenly over the entire year, with the largest changes in August-December (Fig. 4.15).

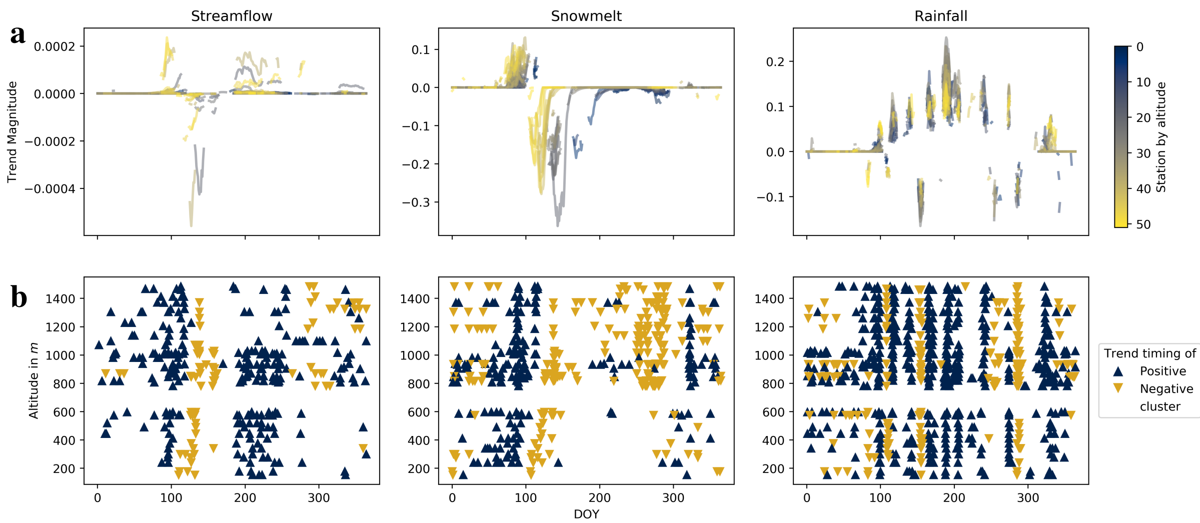

Figure 4.14: *Magnitude and timing of significant 10dMA trends in Østlandet for recent period (1983-2013).* (a) Clusters of significant trends for each DOY. The trend magnitude line for each catchment is colour coded according to station altitude. (b) Simplification of the daily MA plots (e.g. Fig. 4.5), where the timing of each trend cluster is indicated.

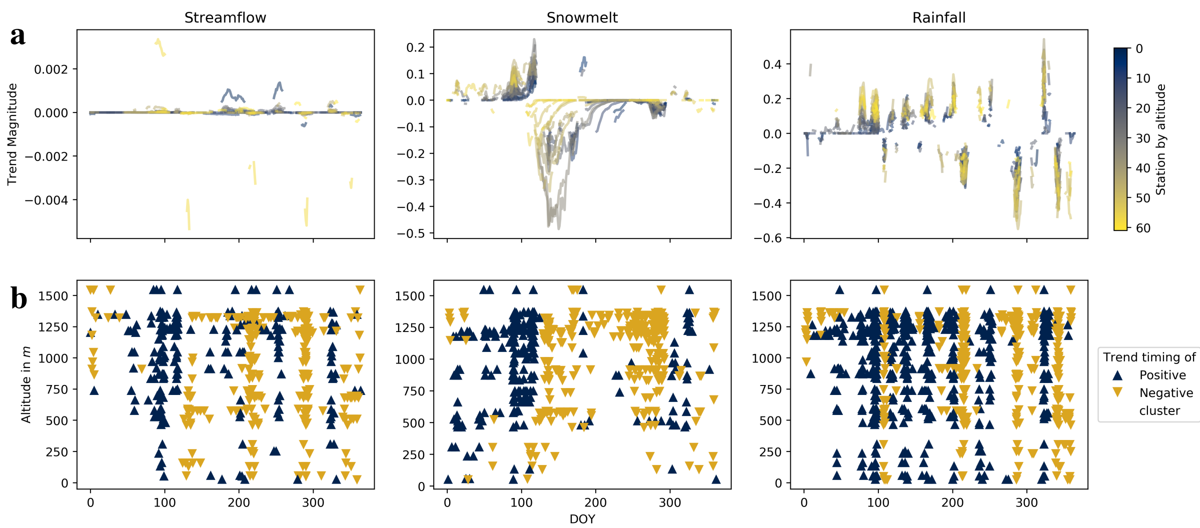

Figure 4.15: *Magnitude and timing of significant 10dMA trends in Vestlandet for recent period (1983-2013).* Same as Fig. 4.14.

4.2.3. Selected catchments

As previously mentioned, the streamflow changes in both periods were small, leading to little to no noticeable change in most catchments. The catchments presented in this section were chosen due changes in their hydrographs being more distinguishable, with an attempt to represent various elevations and hydrological regimes (see Tab. 1.2) in the regions Østlandet and Vestlandet. All of the selected catchments are of relatively small size. The sum of rainfall and snowmelt trends appear to follow the pattern of streamflow change well (Fig. 4.16-19), although often leading or lagging, and/or with reduced magnitude, which can be attributed to the different characteristics and response time of each catchment.

Sæternbekken (Fig. 4.16) is representative of the inland hydrological regime. The early spring snowmelt increased, while peak flow is diminished (Fig. 4.16a), caused by a shift in snowmelt. From May-February only rainfall trends are present, and appear to fit well with streamflow change pattern, especially in the autumn months.

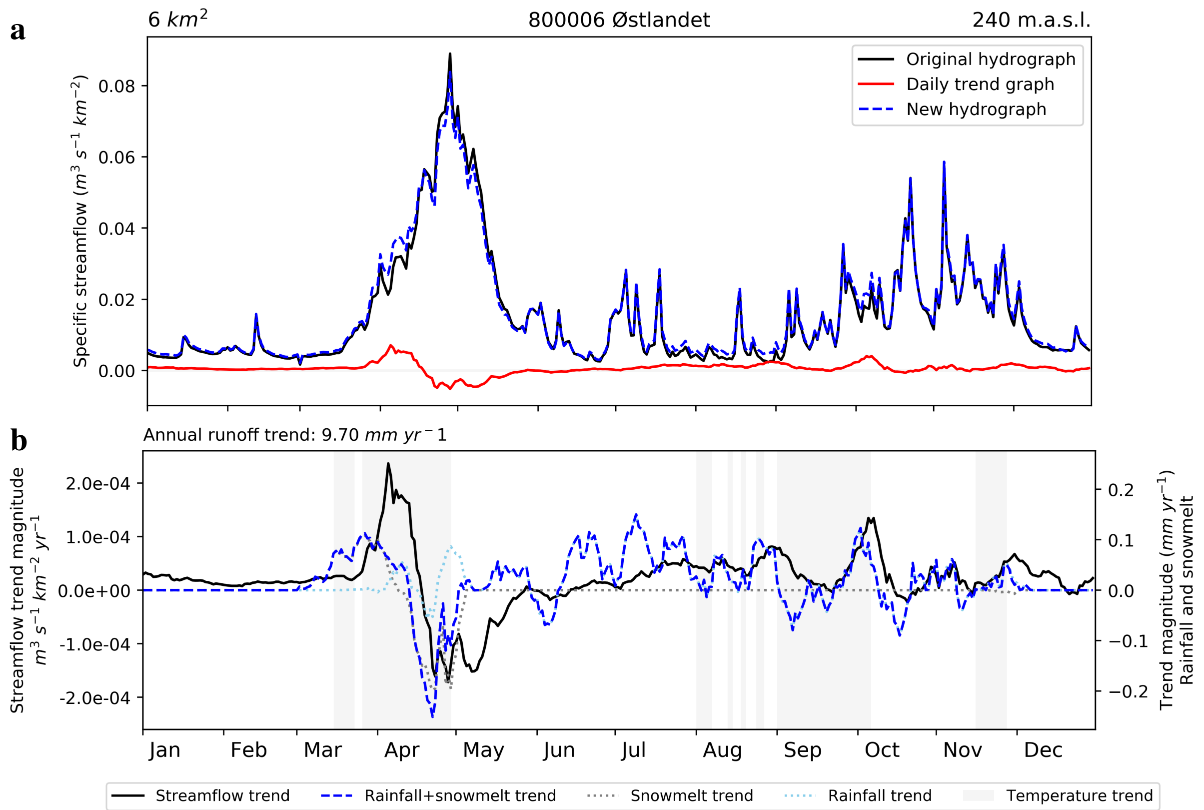

Figure 4.16: *Hydrological trends in 800006 Sæternbekken (1983-2012), a spring snowmelt dominated, low altitude catchment in Østlandet, with summer and autumn rainfall.* (a) Original hydrograph calculated from 1973-1982 streamflow observations, trend hydrograph with the trend magnitude per 30 years for each DOY, and the new hydrograph calculated by adding the trend hydrograph to the original hydrograph (inspired by Déry et al., 2009). (b) Magnitude of 10dMA trends (significant and non-significant) in streamflow, snowmelt and rainfall for 1983-2012 period, periods with significant temperature trends (all positive) shaded. Significant annual runoff or temperature trends included if present.

Grosettjern (Fig. 4.17) is an example of a catchment with a mountain hydrological regime, with snowmelt dominated high flow in spring and winter low flow (Fig. 4.17a). Significant warming in spring is followed by a positive streamflow trend, then a larger negative trend caused by snowmelt trends (Fig. 4.17b). This decrease in snowmelt in May caused a reduction in peak flow, and a slightly earlier onset of the freshet. The increase in summer streamflow is caused by a positive rainfall trend.

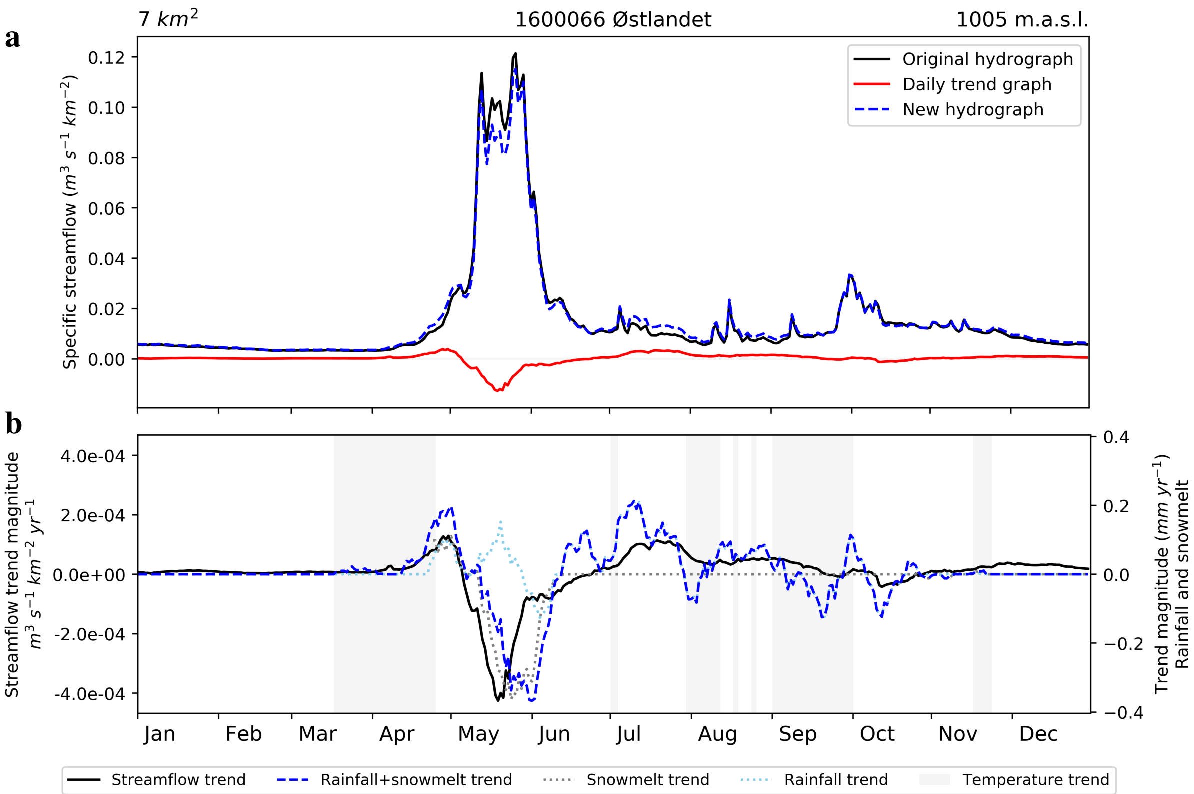

Figure 4.17: *Hydrological trends in 1600066 Grosettjern (1983-2012), a high-altitude, snowmelt dominated catchment in Østlandet*. See Fig. 4.16.

In Fønnerdalsvatn (Fig. 4.18) there is a similar positive snowmelt trend during a significant warming in spring, but which has less impact on streamflow. Streamflow changes in autumn appear to be related to rainfall trends. However, in June-August the combined rainfall and snowmelt (RS) trends are distinctly different from the streamflow trend, which is likely due to 43.5 % of the catchment being glaciated. The discrepancy between streamflow and RS trends is greatest in two periods of significant warming (early July and August). The warming has likely led to increased glacial melt in these periods, causing the discrepancy.

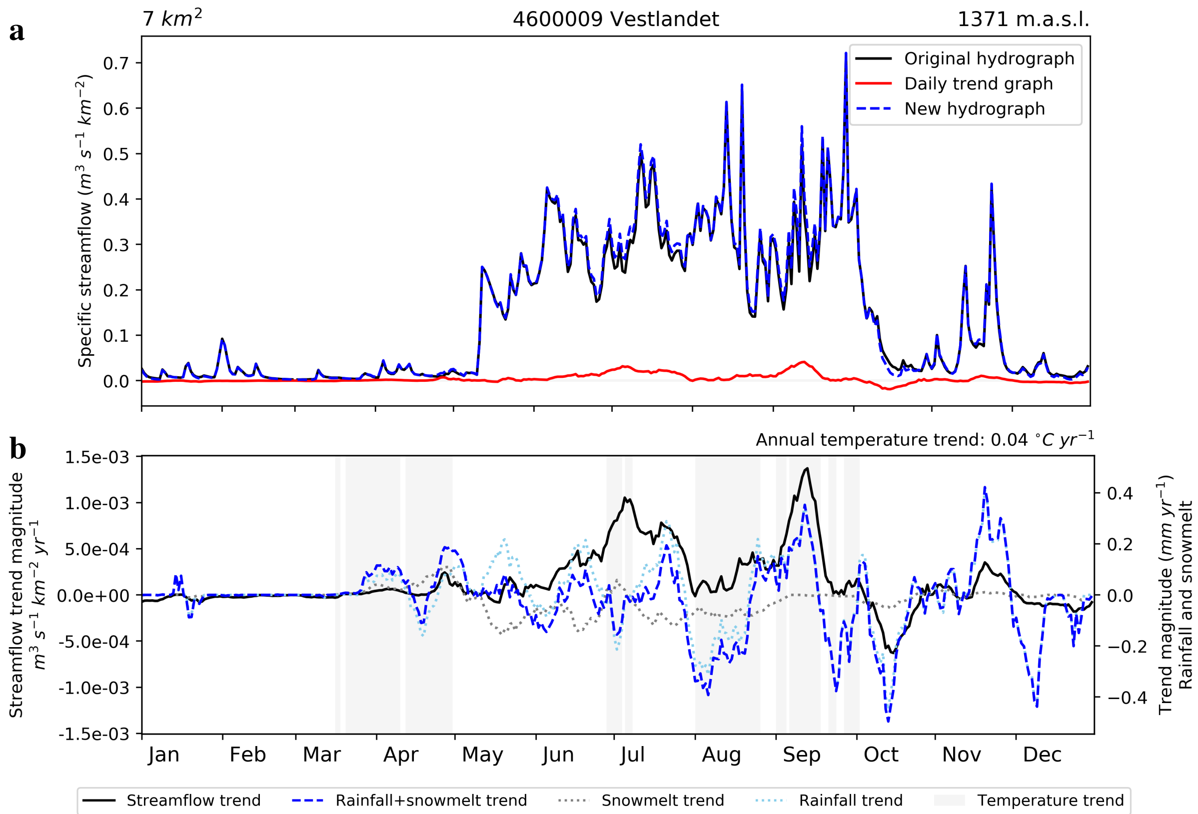

Figure 4.18: *Hydrological trends in 4600009 Fønnerdalsvatn (1983-2012), a high-altitude, glaciated catchment in Vestlandet*. See Fig. 4.16.

Holsenvatn (Fig. 4.19) is an example of a transition hydrological regime. Two peaks in spring streamflow trends are concurrent with positive temperature and snowmelt trends (Fig. 4.19b), and the pattern of streamflow changes for the rest of the year almost exactly match the rainfall trends.

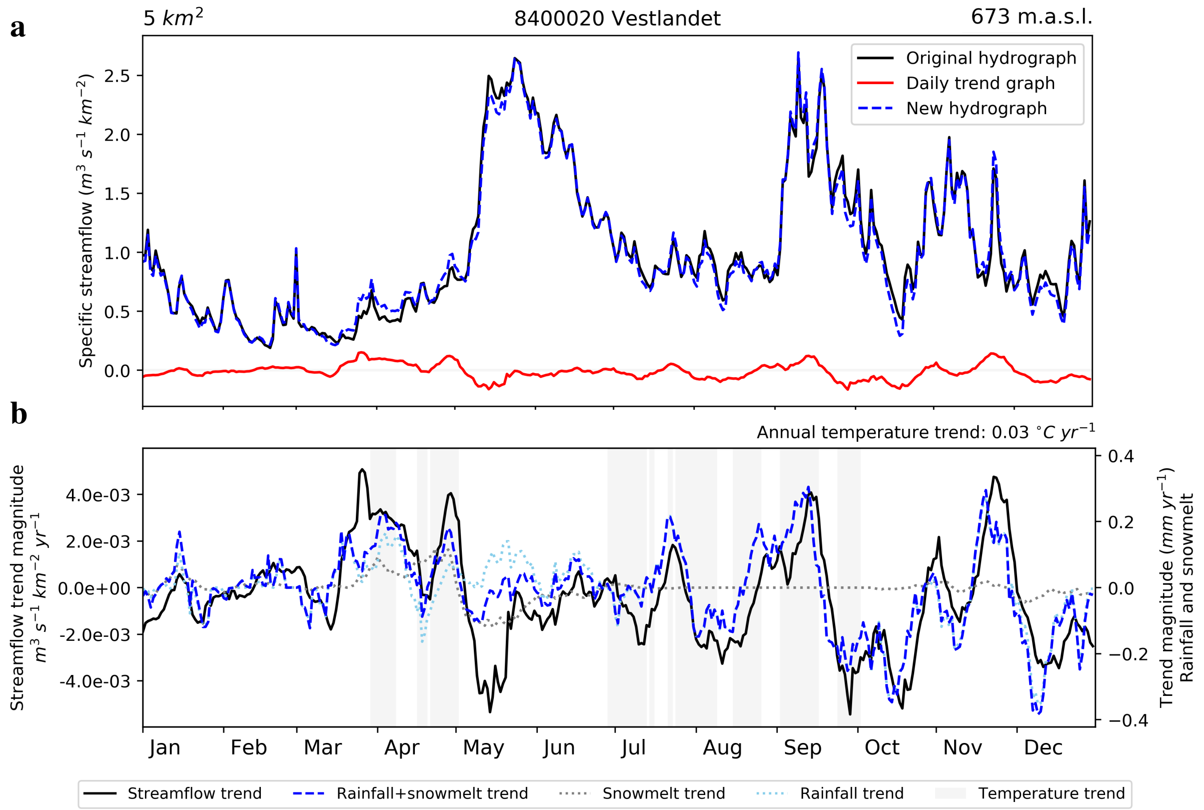

Figure 4.19: *Hydrological trends in 8400020 Holsenvatn (1983-2012), a catchment with spring and autumn high flow in Vestlandet*. See Fig. 4.16.

Tysvær (Fig. 4.20) has an Atlantic hydrological regime, with summer low flow and late autumn-winter high flow caused by rainfall, with little to no contribution from snowmelt. The streamflow changes in this catchment is entirely driven by rainfall changes. From March to August there is a distinct pattern of each peak in the streamflow trend being preceded by a positive trend in rainfall (Fig. 3.20b), although the largest impact on streamflow occurred in autumn (Fig. 3.20a), where the pattern of rainfall and streamflow trends match well.

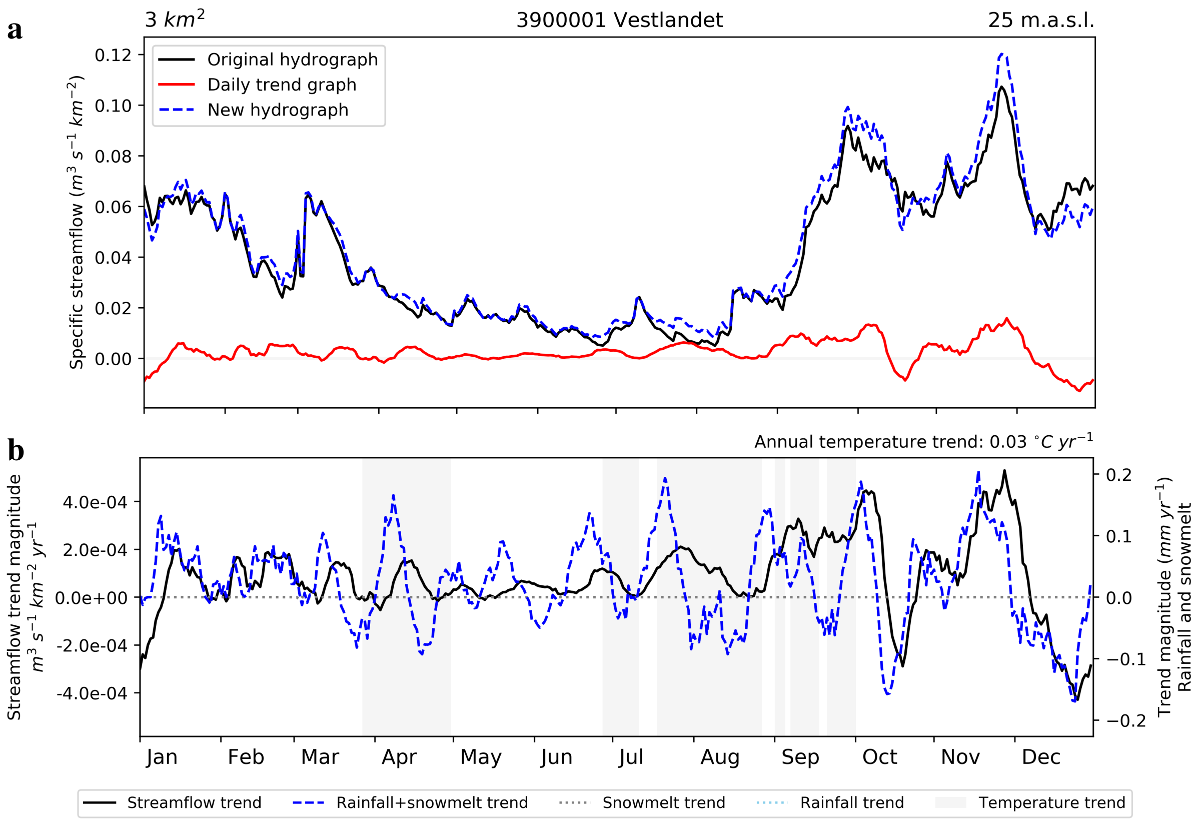

Figure 4.20: *Hydrological trends in 3900001 Tysvær (1983-2012), a rainfall dominated, coastal catchment in Vestlandet*. See Fig. 4.16.

5. DISCUSSION

5.1. Recent and long-term trends

The two periods analysed exhibit a steady increase in mean surface temperature, both globally and in Norway (Fig. 1.1; Fig. 1.6). This study has confirmed significant temperature increases in both annual and daily trends for both periods. However, this warming is not seasonally uniform, with significant warming in April, July-September and late November for both periods, in addition to strong warming in January and February in 1963-2012. Timing and magnitude of temperature trends is largely consistent across regions and altitudes. A recent study which analysed the entire SeNorge temperature dataset with full spatial coverage of Norway, but using monthly averages, found the same significant warming in April and July-September (Rizzi et al., 2017). They further concluded that accelerated warming in spring is due to the snow albedo feedback. Similar patterns of a pronounced warming in spring has been found in the Alps (Kormann et al., 2014; Rottler et al., 2018), but here snow-albedo feedback was not thought to influence the temperature (Rottler et al., 2018). In this study, warming during winter is only evident in 1963-2012. Rizzi et al. (2017) found indications of accelerated warming, as the most recent period analysed, 1981-2010, had stronger temperature trend magnitudes, compared to 1961-1990 and 1971-2000. In this study, the rate of recent change was similarly found to be somewhat larger compared to long-term trends. This accelerated warming could be influencing hydrological changes, as recent streamflow, snowmelt and rainfall trends also show greater rates of change compared with long-term changes.

This study found significant changes in seasonal streamflow distribution, even when few significant changes in annual runoff were found (see Fig. 4.3). These findings are in agreement with Wilson et al. (2010), who found changes in seasonal distribution of discharge since the 1940s have generally been larger than changes in annual flow in the Nordic countries. The temporal pattern of streamflow and snowmelt changes are mostly consistent in both periods, with some differences in timing and duration of trends. The positive long-term streamflow trends in January-April is in agreement with the findings of Stahl et al. (2010), who found positive monthly streamflow trends in January-April for the period 1962-2004 in Norway. Moreover, their findings of a negative trend in May-June streamflow, concur with the findings of this study. Stahl et al. (2010) assumed these positive streamflow changes in spring to be caused by an earlier onset snowmelt. The large agreement between the patterns of snowmelt and streamflow trends found in this study confirms this assumption. Stahl et al. (2010) further found mixed positive and negative streamflow trends in summer. In this study, where the trends are grouped by runoff region, is it clear that these mixed trends result from the various hydrological regimes and different hydro-climatic conditions of each region. Both long-term and recent summer streamflow trends are negative in Finnmark, Nordland, and Trøndelag, positive in Sørlandet and Østlandet, and mixed in Vestlandet. The positive summer streamflow trends are caused by increased rainfall and the negative could be caused by increased evapotranspiration.

Snowmelt is occurring earlier, which has also been observed in many European rivers dominated by snowmelt-induced peak flow (Madsen et al., 2014) and other mountainous regions of the world, e.g. Colorado Mountains (Clow, 2009). The detected warming in spring appears to be the cause of this shift to an earlier onset of snowmelt, as evidenced by positive spring snowmelt trends in all regions concurrent with significant warming. However, the resulting effect on streamflow appears to be a flattening of the snowmelt-induced peak flow (see Section 4.2.3). Earlier and slower spring snowmelt is consistent with trends across the northern hemisphere from 1980 to 2017 (Wu et al., 2018). This change is due to (1) an earlier onset of snowmelt and (2) the negative snowmelt trends having greater magnitude than the positive. These greater negative trends indicate an overall reduction in snowpack, supported by a reduction in snow cover extent (Rizzi et al., 2017). Conversely, trends in SWE for the analysed period is mixed, with prevailing positive trends in 1961-1990 and negative in 1991-2009 (Skaugen et al., 2012). Moreover, with the exception of Nordland, few significant changes in total annual snowmelt were detected, which gives an indication of annual snowpack. Furthermore, net negative snowmelt trends are mostly present in the mid-altitude catchments in Vestlandet and Østlandet (see Fig. 4.14-15a). Fewer and less intense snowmelt generated floods in the analysed period (Vormoor et al., 2016), suggests that slowing snowmelt rates has affected the flood regime.

In Nordland, Trøndelag, and Vestlandet the earlier spring snowmelt is concurrent with positive rainfall trends, which suggests that both rainfall and temperature are driving the positive snowmelt trend in these regions, as rainfall is known to accelerate snowmelt (Singh et al., 1997). In addition, these regions all have strong positive rainfall trends in winter and spring, likely due to an increased fraction of liquid precipitation (Hynčica and Huth, 2019), further supported by significant warming coinciding with these positive rainfall trends (Fig. 4.7-8). There is no increase in rainfall during winter in Finnmark and Østlandet, although these regions also experienced significant warming. As winter temperatures are very low in these regions (see Fig. 1.5), the warming was not sufficient to cause a phase change in precipitation. Increasing trends in summer rainfall in Østlandet and Sørlandet, are not linked to precipitation phase shifts, but an overall increase in precipitation (Kovats et al., 2014), and drives and increase in summer streamflow in these regions.

5.2. Altitude dependence of hydrological trends

Elevation dependent warming (EDW) is thought to cause larger temperature increases at higher altitudes (Pepin et al., 2015). Neither annual nor daily trends show in this study a consistent pattern that agrees with EDW. Although Trøndelag shows greater warming with altitude, the two regions with the largest elevation range show either no consistent elevation dependent warming (Vestlandet), or greater warming at lower elevations (Østlandet). This latter finding agrees with temperature trends in the Swiss Alps, where trends were stronger at lower elevations (Rottler et al., 2018). The findings of this study are not consistent enough to agree with EDW, but since temperature data was extracted only for the catchments used in this study, an analysis of the entire data set could be more conclusive.

Changes in winter and spring rainfall are altitude dependent in Vestlandet, Trøndelag and Nordland. This is especially noticeable in catchments below 600-750 m, which show strong positive streamflow trends in January-April caused by increased rainfall. The phase of winter precipitation is known to be elevation dependent in maritime mountainous regions like these, making them especially sensitive to climate change (Jefferson, 2011).

The aforementioned net reduction in snowmelt trends is less apparent in recent significant daily trends above 1000 m in Vestlandet and Østlandet (Fig. 4.6). These catchments further show strong positive long-term trends during summer. When non-significant daily trends are included (Fig. 4.13; Fig. 5.1), this pattern of positive-negative-positive snowmelt change becomes more apparent, i.e. the period when snowmelt occurs is growing longer. This change occurs above 750 m in Vestlandet and 1000 m in Østlandet, and is present in both recent and long-term periods. The extension of the snowmelt season is caused by (1) increasing temperature causing an earlier start, and (2) increased snow volume causing the season to last longer. Areas above 850-1000 m experienced a later end to the snow season due to higher snow accumulation caused by higher precipitation and sufficiently low temperatures, and an increase in SWE (Skaugen et al., 2012; Rizzi et al., 2017).

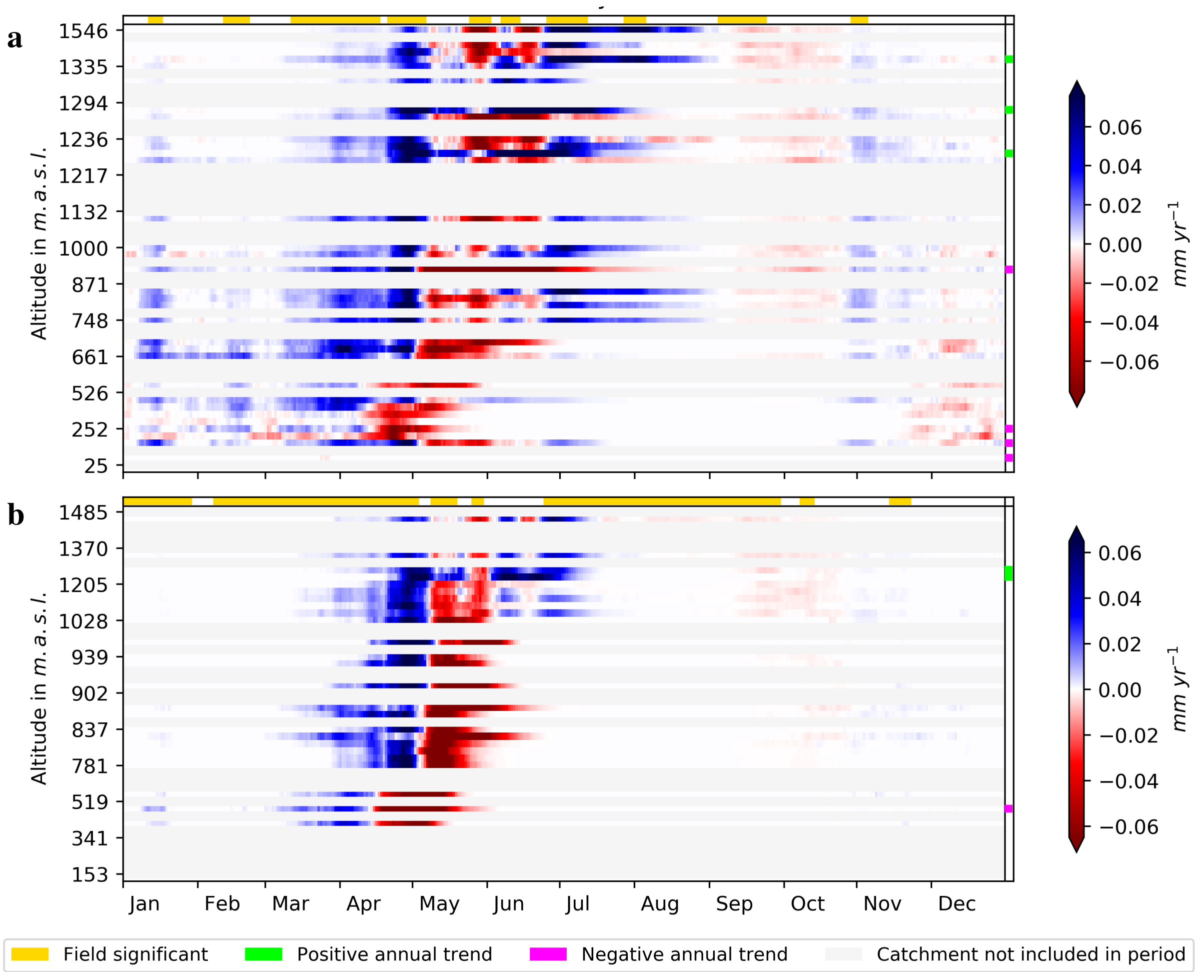

Figure 5.1: Long-term 10dMA snowmelt trend magnitudes of both significant and non-significant trends in (a) Vestlandet and (b) Østlandet.

5.3. Attribution of streamflow changes

The majority of the detected streamflow can be attributed to rainfall and snowmelt trends. Both the larger daily trend plots and the more detailed evaluation suggest an overall agreement between snowmelt and rainfall as drivers of streamflow change. Rainfall and snowmelt are both closely related to precipitation, which is a major driver of streamflow trends globally (Dai et al., 2009). However, glacial melt has also been found to influence some catchments, and the majority of the catchments showed positive long-term annual evapotranspiration trends (Fig. 4.2). These evapotranspiration trends should however be interpreted with caution, as the annual evapotranspiration was not measured, but calculated from the hydrological records. The correlation between streamflow and combined snowmelt and rainfall trends suggests that larger streamflow changes are driven by snowmelt and rainfall changes, but the largest RS changes show no response in streamflow (Fig. 5.2). This could be due to the lag between the timing of the streamflow and RS trends (see Section 4.2.3) caused by the different response time of each catchment. Another explanation is that there are other hydrological variables influencing streamflow, e.g. glaciers, evapotranspiration or groundwater.

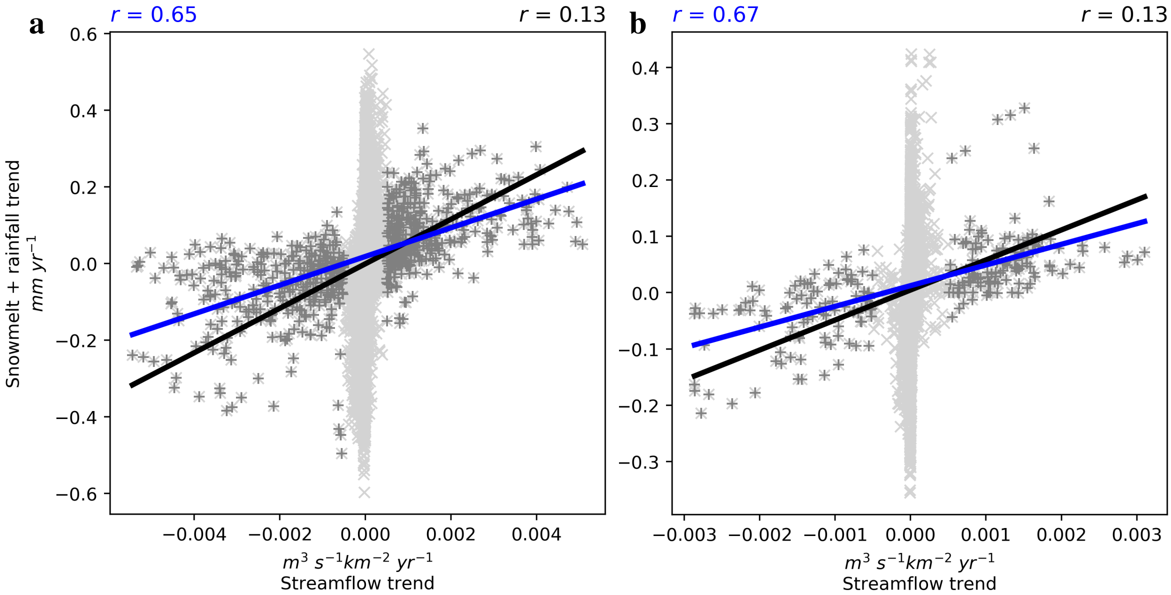

Figure 5.2: Correlation between 10dMA streamflow and RS trends in (a) 1983-2012 and (b) 1963-2012 from all six runoff regions. Lines are fitted by OLS to all trends (black line) and discounting streamflow trends between ± 0.0005 (blue line).

Even in natural catchments, streamflow is the integrated response of many driving processes (Madsen et al., 2014) which can act as confounding factors. Although the catchments in this study were selected as to keep these factors to a minimum, they cannot be completely ruled out. Apart from evapotranspiration, and other hydro-climatic variables directly influencing the water cycle in each catchment, vegetation also responds to a warming climate. A change in vegetation cover will impact the characteristics of the catchment, which in turn could lead to alterations in streamflow, which could be misconstrued as a trend. Furthermore, vegetation responses to climate change are also elevation dependent, e.g. changes in tree and forest lines, which in Norway have been rising since the 1920s (Bryn and Potthoff, 2018). Additionally, vegetation responses to past disturbances, e.g. forest fires, also causes gradual changes in vegetation impacting streamflow (Jones, 2011). Other than by selecting pristine or near-natural catchments, the influence of vegetation responses to climate change was not explicitly accounted for in this study. Neither were other hydrological variables that influence streamflow analysed, e.g. evapotranspiration. Despite establishing a strong relationship between streamflow and RS trends, which in turn have been connected to rising temperatures, these changes have not been directly attributed to climate change. Furthermore, the influence of catchment size and other hydrological variable trends on streamflow trends should be quantitatively assessed and is suggested as a topic of future research.

5.4. Suitability of approach for trend detection and attribution

There is no question that daily trend analysis reveals more about how the hydrological regime is changing than annual trends. By aggregating the daily data to annual values, all information on seasonal and sub-seasonal changes are lost. In this study the annual analysis only detected a few significant changes, while the daily analysis detected significant changes in all catchments. This disparity between the annual and daily trends is caused by intra-annual shifts, while the annual amounts remain unchanged (Déry et al., 2009; Kormann et al., 2015). Such shifts will not be detected by annual analysis, and further emphasises the importance of differentiating between annual and seasonal trends already highlighted by other studies (e.g. Stahl et al., 2010). Using a daily resolved trend analysis it is possible to precisely determine the time of year when hydrological variables are changing, thus facilitating the attribution of changes to the underlying causes, while also being of more use in water management. Despite the clear advantages of analysing daily trends, annual trends could be a more useful tool for analysing the changes in water balance on the basin or regional scale. Moreover, changes in the annual values will affect the daily trends, as illustrated by the larger negative than positive trends in snowmelt. However, this can also be accomplished by the annual trend integral (Kormann et al., 2015), or in the case of trends in mean annual temperature, the mean of daily trends. A complication of high-resolution trend analysis identified by Déry et al. (2009) is whether the detected streamflow trends are primarily driven by the timing of the intensity of an event. However, by incorporating variables affecting streamflow, and analysing the trends of these variables in conjunction with streamflow trends, as done in this study, it is possible to identify the mechanisms driving streamflow changes.

5.4.1. Effect of moving average smoothing

The main effect of the MA filter is to reduce the magnitude of streamflow peaks, but it does not substantially increase the values in low flow periods (Fig. 5.3). A 30dMA removes a large portion of the variability from the record and makes the changes appear to occur more gradually over a longer time period. This gives the impression that the onset of the spring snowmelt occurs about two weeks earlier that in the original record (Fig. 5.3).

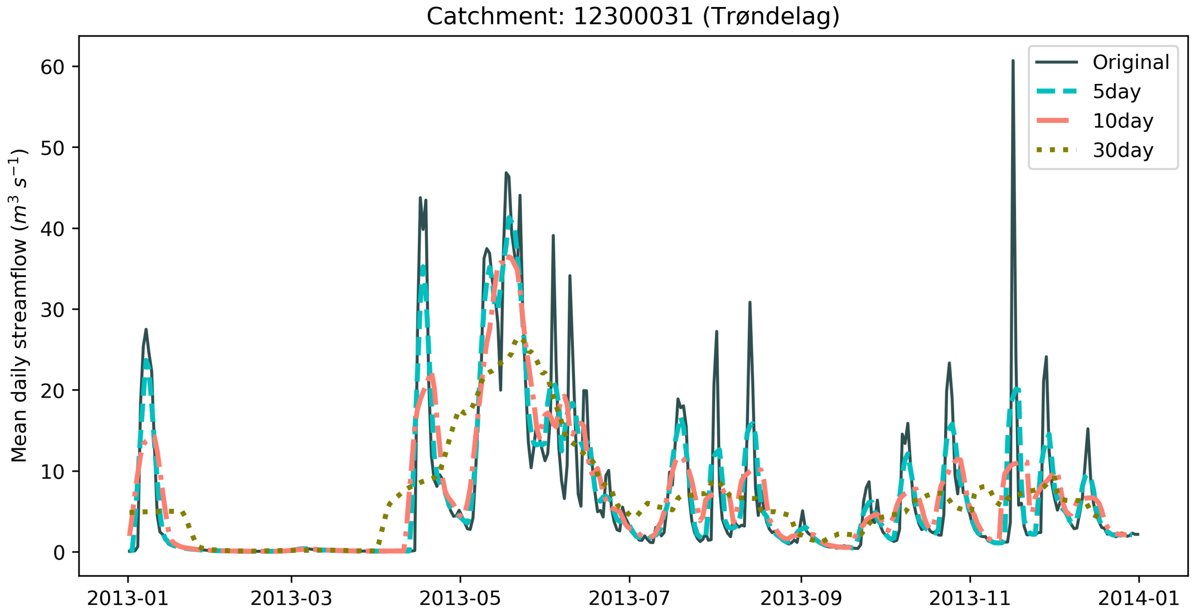

Figure 5.3: *Example of the effect of applying MA filters to daily data*. The original time series, as well as the time series after moving average smoothing with a window of 5, 10, and 30 days centred on the reference day.

This temporal stretching effect is apparent in the daily trends (Fig. 5.4). The 30dMA makes the changes appear more coherent and spread over a larger period than is the case. On one hand, this improves the readability of the plot and makes it easier to identify the broader changes. On the other hand, some temporally focused changes are not identified in the 30dMA trends, e.g. the increase in late November streamflow driven by increased rainfall (Fig. 5.4a; Fig. 5.4c). The 5dMA and 10dMA trends are able to pick up these shorter shifts in streamflow. However, 5dMA or 10dMA trends exhibit more and shorter trend clusters, which makes the detection of a pattern more difficult. The rainfall trends appear to be more susceptible to the effects of the MA filter than the other variables. This is especially noticeable by the high trend magnitudes in January-March (Fig. 5.4c).

In summary, using a 10dMA appears to be a suitable compromise for detecting sub-monthly trends. However, the choice of MA filter depends on the application of the daily resolution trend plots. If a more general period of change is desired, a 30-day moving average is suitable. Additionally, the effect of different MA filters on results of the daily trend analysis should be explored further, especially how different variables are affected, and comparing daily trend calculated from the original daily data with smoothed data.

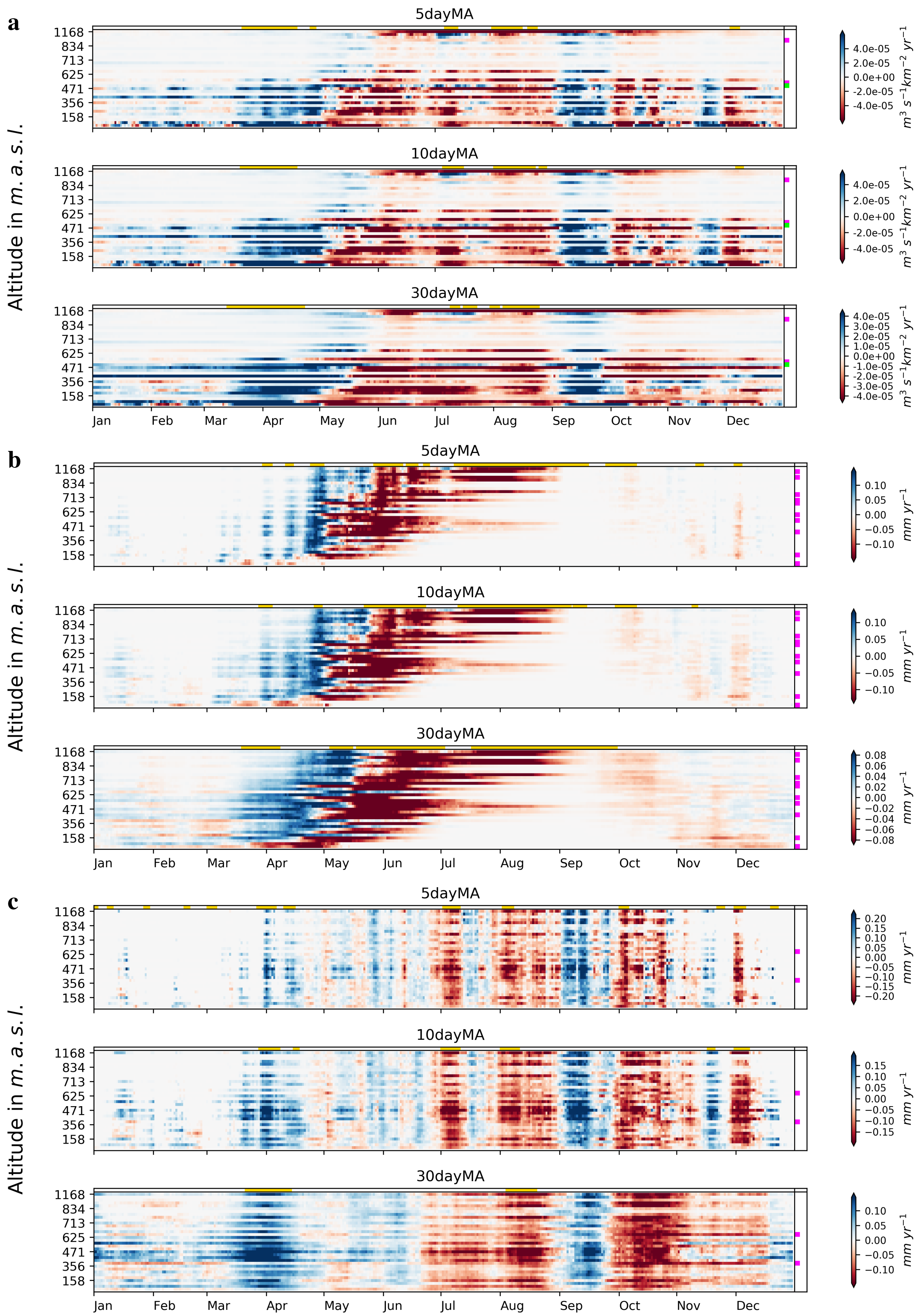

Figure 5.4: *Comparison of the effect different MA filters on daily trends*. 5dMA, 10dMA, and 30dMA daily trends for Nordland without filtering by local significance level.

5.4.2. Data requirements and uncertainty

A major advantage of using the interpolated SeNorge data, was that snowmelt and rainfall data were available for every catchment, and enabled a direct comparison of the streamflow and RS trends for each catchment. Nevertheless, some of the disagreement found between the streamflow and RS trends could result from the uncertainty introduced from the SeNorge data, since it is modelled and interpolated from observational measurements. However, as meteorological data is generally measured at stations, some sort of interpolation method must be applied to obtain full spatial coverage of a catchment. There are two aspects of the SeNorge data which influence the trend analysis. Firstly, the data might be biased at higher elevations as the spatial coverage of temperature and precipitation stations in Norway is denser at low elevations (Rizzi et al., 2017). Secondly, the number of stations from which the data is interpolated vary from year to year, with generally better coverage in recent years (Engeset, 2016).

The method requires long daily records of streamflow and other hydro-climatological variables. Such records are not available in many regions of the world, which highlights the importance of HRN. The time and computational requirements of this method is moderate. The majority of the time was spent organising the records, assessing the quality of the data, and writing the scripts, while the actual running of the trend analysis for all 207 catchments, four variables and two time periods, was accomplished in a day. Calculating the field significance was more time consuming, since it involved resampling.

5.4.3. Non-linearity and significance testing

As previously mentioned, the consistently higher trend magnitudes for 1983-2012 compared to 1963-2012, could be caused by non-linear or abrupt change. The MK test should be able to detect non-linear changes, as long as they are monotonic, but has been shown not to detect abrupt changes (Zhao et al., 2016). However, the SS estimator can only estimate the magnitude of linear trends (Helsel and Hirsch, 2002). Both non-linear and abrupt changes could be present in the DOY time series of this study, as for example abrupt change has been found in arctic catchments (Makarieva et al., 2019). Burn et al. (2012) emphasised the need to examine hydrological time series for non-linear and abrupt change behaviour. However, as over 450 000 DOY time series was analysed with the MK test and SS estimator in this study, it was not feasible to examine each of them for abrupt change or non-linearity. If present, these phenomena would lead to over- or underestimation of certain trends, and could result in significant trends not being detected.

The applicability of trend testing in hydrology has been called into question (Serinaldi et al., 2018), and some studies have therefore opted to only estimate trend magnitudes (e.g. Stahl et al., 2010; Kormann et al., 2016). In this study, some changes in hydrological variables became more apparent when all daily trends where presented, and not “filtered” by significance. Especially when attributing streamflow changes to snowmelt and rainfall the inclusion of all trends where necessary (Sections 4.2.3; 5.4). However, for other applications, “filtering” by significance could be appropriate, e.g. when looking for consistent (and field significant) changes for larger regions.

5.4.4. Influence of decadal scale variability

Owing to the natural variability of the climate system at the decadal to interannual scale, trends calculated from short records are very sensitive to the beginning and end of the chosen period (Burn et al., 2012; IPCC, 2014). The periods chosen for this study are considered to be sufficiently long to reduce the influence of interannual variability, but not decadal variability. Finding consistent changes in both periods increases the likelihood of detected changes being caused by long-term persistence, and indeed many of the hydrological trends found in this study is consistent in both periods. However, decadal scale climate variability could still be influencing temperature and hydrological trends.

Precipitation in Norway is to a large extent governed by atmospheric circulation patterns, and mainly influenced by low pressure system activity (LPSA) in the North Atlantic (Hanssen-Bauer et al., 2015). The North Atlantic Oscillation (NAO) is the leading mode of atmospheric circulation variability over the North Atlantic region (Pinto and Raible, 2012) and a good indicator of LPSA in the Nordic Sea (Hanssen-Bauer et al., 2015), particularly in winter, when atmospheric circulation is more variable (Hurrell and Deser, 2010). A high NAO-index indicates high LPSA bringing warm, moist air in over Norway, resulting in warm, wet winters, while a low NAO-index indicates low LPSA, causing dry, cold winters (Hanssen-Bauer et al., 2015). The NAO also strongly influences winter temperatures in northern Europe (Hurrell and Deser, 2010). During winters of the 1960s, the NAO was in a pronounced negative phase, followed by prevailingly positive phases (Fig. 5.5). This transition from negative to positive winter NAO index occurred during the long-term time period analysed in this study, and could explain why significant winter warming and increased winter rainfall was only detected in this period. Furthermore, a high NAO index is linked to increased SWE at high elevations and reduced SWE at low elevations (Skaugen et al., 2012). Moreover, the NAO has been shown to influence streamflow trends in Europe (Hannaford and Marsh, 2006), and annual and winter streamflow trends in Norway (Hannaford et al., 2013).

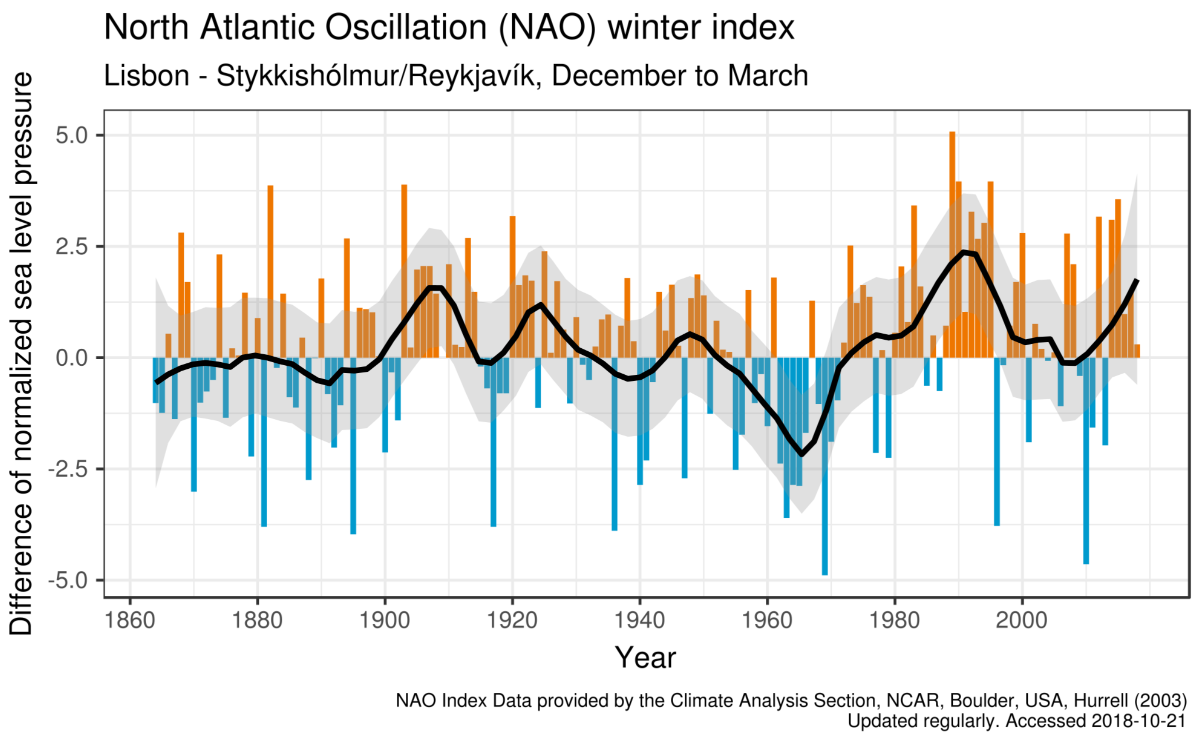

Figure 5.5: NAO winter index (Source: Delorme, 2017)

The influence of the NAO and other decadal scale climate variability was not taken into account in this study. Therefore it uncertain whether the hydrological changes detected by the daily resolved trend analysis has been primarily influenced by climate variability or climate change. Assessing how decadal scale climate variability affects daily hydrological trends should be a topic of future study. Furthermore, as sea surface temperatures (SST) affects the climate of coastal regions in Norway (Hanssen-Bauer et al., 2015), the influence of ocean circulation changes affecting SST, e.g. the Atlantic Multidecadal Oscillation, should also be accounted for. Understanding the impact of modes of climate variability on hydrological trends is important because these may change as a response to climate change (Christensen et al., 2013).

6. CONCLUSIONS

This study analysed hydrological trends in 207 Norwegian catchments between 1963 and 2012. A daily resolution trend analysis approach by Kormann et al. (2014; 2015), complemented by a commonly used annual trend analysis approach, of daily streamflow, rainfall, snowmelt and temperature has demonstrated how a warming climate has affected the hydrological regime of catchments in Norway at the sub-seasonal level.

The detected hydrological trends are for the most part consistent in trend timing for both the recent (1983-2012) and long-term period (1963-2012), although with a consistently smaller magnitude for the long-term trends. Detected streamflow changes are small, and there are little to no noticeable changes in most catchments. Patterns consistent in both periods could indicate how catchments will respond to continued future warming. Consistent trends in both periods include; an earlier start to the snowmelt season, slowdown of snowmelt, and increased summer rainfall east of the Scandinavian Mountains. Notable inconsistencies between periods are; streamflow trends in Vestlandet, summer snowmelt trends in high-elevation catchments in Vestlandet and Østlandet, and warming during winter.

The magnitude of streamflow trends display no clear dependence on altitude. Rainfall trends overall were generally consistent across altitudes, except positive rainfall trends found in low-elevation catchments in Vestlandet, Trøndelag and Nordland, likely caused by a phase shift in precipitation from snowfall to rainfall. High-elevation catchments show an earlier start and later end to the snowmelt season, due to warming in spring and increased snowpack. Snowmelt trend magnitudes were found to be altitude dependent in only two regions, Sørlandet and Østlandet. Temperature showed no consistent dependence on altitude, except a positive correlation with altitude in Trøndelag and negative correlation with altitude in Østlandet.

Streamflow changes in Norway between 1963 and 2012 can largely be attributed to changes in timing and magnitude of rainfall and snowmelt. Especially larger magnitude daily streamflow changes are well correlated with combined rainfall and snowmelt trends, while large changes in combined rainfall and snowmelt trends show no response in streamflow. This discrepancy could caused by the catchment response time or influence of other hydroclimatological variables. In higher elevation catchments, glacial melt is an additional factor, and increased evapotranspiration may influence summer streamflows. Positive temperature trends appear to drive some changes, notably timing of snowmelt trends and phase shifts in precipitation.

Different moving average filters were applied to the original data before calculating trends. In this study, the 10-day moving average was found to be the most suitable. A 30-day moving average gives a misleading impression of coherency in timing of trends.

This study focused on detection of sub-seasonal hydrological trends, impacts on the hydrological regime resulting from these trends, and identifying the main drivers of streamflow changes. As the detected trends were not directly attributed to climate change, quantitatively assessing this relationship is suggested for future research.

AUTO-CRITIQUE

The potential impacts of climate change on hydrology is a topic I find interesting. I stumbled upon this dissertation topic while looking though a list of available topics from my old university, Universität Potsdam, proposed by Dr. Klaus Vormoor. What most drew me to the topic was that the connection to Norway, my home country. The opportunity to use the relatively new method of daily resolved trend analysis was a further motivation. I also wanted the experience in managing and analysing a large dataset, and apply some of the skill I had learned in our Scientific Computing module.

One of the strength of the dissertation was the large number of catchments and spatial coverage of the dataset. The approach yields a lot of information on detected hydrological trends and most of the time was dedicated to the trend analysis and interpretation of the four variables. In hindsight, I should have thought more about attribution earlier in the process and dedicated more time to it. Especially, I think the connection to global warming and climate change should have been explored in more detail. While the altitude dependence of hydrological trends were discussed, it could have been assessed more quantitatively.

A lot of hard work was put into making the figures, and I find them to be very illustrative and able to convey a lot of information in a relatively compact way.

While interpreting the results I kept thinking of additional things I could include, e.g. a comparison between the daily trend integrals and annual trend magnitude, or the influence of decadal climate variability. At the same time I started the analysis in good time, and there is simply a limit to what can be explored and accomplished in four months.

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APPENDIX

Table A1: List of the 227 hydrological records compiled by Vormoor et al. (2016). The record length refers to the years where streamflow records were available.

Runoff region	Catchment number	Name	Record length [year]	Area [km²]	Altitude [m.a.s.l.]	Longitude	Latitude
FINNMARK	20600003	Manndalen bru	1971 - 2013	200.5	933	20.53	69.52
	20500006	Didnojokka	1979 - 2013	110.9	922	20.78	69.18
	19600012	Lundberg	1961 - 2013	246.5	868	18.55	68.72
	20500008	Helligskogen	1982 - 2013	371.1	831	20.66	69.23
	20500003	Skibotn bru	1970 - 2013	727.2	819	20.27	69.37
	20900004	Lillefossen	1961 - 2013	331.8	806	21.92	69.79
	19600011	Lille Rostavatn	1959 - 2013	638.4	740	19.58	69.02
	19600035	Malangsfoss	1907 - 2013	3110.7	713	18.66	69.03
	20800003	Svartfossberget	1981 - 2013	1932.3	673	21.38	69.53
	20800002	Oksfjordvatn	1955 - 2013	265.6	548	21.38	69.90
	21200049	Halsnes	1966 - 2014	144.9	531	22.94	70.03
	21300002	Leribotnvatn	1961 - 2014	135.5	457	23.55	70.11
	21200010	Masi	1966 - 2013	5620.8	450	23.64	69.42
	21200011	Kista	1971 - 2014	6181.8	449	23.52	69.83
	22300001	Stabburselv	1923 - 2013	1067.3	441	24.88	70.18
	22300002	Lombola	1923 - 2013	877.1	435	24.76	70.14
	21300004	Kvalsund	1978 - 2013	124.8	355	23.95	70.49
	22400001	Skoganvarre	1921 - 2012	942.3	345	25.09	69.84
	19400001	Lysetvatn	1934 - 2013	129.7	292	17.82	69.39
	19600007	Ytre Fiskeløsvatn	1960 - 2013	54.4	273	18.88	69.26
	24400002	Neiden	1911 - 2013	2947.2	212	29.32	69.70
	24700003	Karpelva	1927 - 2013	128.9	192	30.38	69.66
	24600009	Sametielv	1962 - 2013	255.7	131	29.72	69.40
NORDLAND	15900003	Engabrevatn	1969 - 2013	53.3	1168	13.77	66.69
	15600024	Bogvatn	1970 - 2013	36.2	985	14.49	66.63
	16300007	Kjemaavatn	1969 - 2013	36.5	935	15.41	66.77
	15600013	Bjørnfoss	1954 - 2013	306.1	924	14.32	66.47
	15600008	Svartisdal	1929 - 2013	122.0	902	14.22	66.48
	16300005	Junkerdalselv	1937 - 2013	422.0	834	15.41	66.81
	15600017	Virvatn	1966 - 2013	79.1	833	15.36	66.31
	15100015	Nervoll	1968 - 2013	653.4	827	13.99	65.44

	15600010	Berget	1950 - 2013	210.6	804	13.88	66.46
	16100007	Tollaaga	1972 - 2013	225.1	795	14.85	66.90
	15100021	Joibakken	1972 - 2013	2623.1	713	13.45	65.53
	16700003	Kobbvatn	1916 - 2013	387.1	680	15.99	67.62
	16300006	Jordbrufjell	1945 - 2013	69.9	664	15.15	66.94
	15100028	Laksfors	1908 - 2013	3653.4	660	13.29	65.62
	17200008	Rauvatn	1977 - 2014	21.2	632	16.88	68.21
	14400001	Åbjørvatn	1908 - 2013	389.0	625	12.67	65.04
	15100013	Øvre Glugvatn	1968 - 2013	60.7	570	13.55	65.67
	16200003	Skarsvatn	1916 - 2013	145.4	520	14.98	67.08
	15300001	Storvatn	1916 - 2013	48.0	516	13.12	66.09
	16600001	Lakshola	1916 - 2013	230.8	500	15.76	67.45
	15600027	Leiraaga	1974 - 2013	44.1	494	13.86	66.44
	15700003	Vassvatn	1916 - 2013	16.3	471	13.18	66.39
	15600015	Forsbakk	1963 - 2013	56.0	445	13.81	66.29
	15200004	Fustvatn	1908 - 2013	525.7	436	13.31	65.91
	15900005	Strømdalen	1976 - 2013	22.4	416	13.42	66.63
	17200005	Melkedal (Litlevatn)	1938 - 2014	92.2	412	16.72	68.25
	17800001	Langvatn	1953 - 2014	18.4	356	15.72	68.62
	18900003	Tennevikvatn	1978 - 2014	85.4	315	16.72	68.64
	17700004	Sneisvatn	1916 - 2014	29.3	302	15.71	68.41
	16800003	Laksaa bru	1953 - 2013	26.7	217	15.30	67.80
	17200007	Leirpoldvatn	1970 - 2014	18.8	216	16.43	68.28
	14800002	Mevatnet	1973 - 2013	108.5	158	12.54	65.36
	16600013	Vallvatn	1953 - 2014	53.0	97	15.54	67.34
	15000001	Sørra	1952 - 2013	6.6	69	12.57	65.98
	18500001	Gaaslandsvatn	1934 - 2013	7.7	34	14.63	68.67
TRØNDELAG	12100022	Syrstad	1972 - 2013	2278.1	866	9.73	63.03
	12200011	Eggefoss	1941 - 2013	654.2	844	11.18	62.89
	30700007	Landbru	1943 - 2013	61.4	711	13.92	64.89
	13900015	Bjørnstad	1934 - 2013	1037.6	700	13.26	65.02
	14200001	Første Aunvatn	1982 - 2013	87.3	625	12.48	64.89
	30800001	Lenglingen	1925 - 2013	450.0	625	13.76	64.24
	12300031	Kjeldstad i Garbergelva	1912 - 2013	145.0	576	11.13	63.27
	13900026	Embreth?er	1980 - 2013	494.8	574	12.44	64.39
	13900035	Trangen	1978 - 2013	852.4	558	12.48	64.43

	13900020	Moen	1974	-	2013	64.0	556	13.08	64.93
	30700005	Murusjø	1925	-	2013	345.6	528	14.02	64.49
	12700011	Veravatn	1966	-	2013	176.1	518	12.33	63.79
	12700013	Dillfoss	1973	-	2013	479.6	507	11.77	63.76
	12400002	Høggaas bru	1912	-	2013	494.7	505	11.36	63.49
	12700006	Grunnfoss	1907	-	2013	880.4	500	11.81	63.79
	12800005	St?foss	1932	-	2013	476.6	358	11.72	64.02
	13300007	Krinsvatn (Krings-vatnet)	1915	-	2013	206.6	348	10.23	63.80
	13800001	Oyungen	1916	-	2013	239.1	294	11.08	64.24
	12500002	Fossing	1932	-	2012	163.4	282	11.06	63.61
	14000002	Salsvatn	1916	-	2013	431.9	270	11.45	64.71
	13100001	Oppgrande bru	1979	-	2013	108.8	267	10.98	63.82
VESTLANDET	7600005	Nigardsbrevatn(Nigardsjøen)	1962	-	2013	65.3	1546	7.24	61.67
	4600009	Fønnerdalsvatn	1980	-	2013	7.0	1371	6.28	60.07
	7500022	Gilja	1963	-	2013	203.4	1364	7.62	61.55
	10900009	Driva v/Risefoss	1933	-	2013	745.4	1347	9.59	62.51
	8800004	Lovatn	1900	-	2013	234.9	1337	6.89	61.86
	7300004	Saelthun	1961	-	2013	790.0	1335	7.70	61.05
	7400016	Langedalen	1972	-	2012	23.8	1328	7.71	61.40
	9900017	Kjeldstad i Garbergelva	1981	-	2013	49.9	1325	7.50	62.20
	7300001	Lo bru	1916	-	2013	562.4	1324	7.82	61.06
	7600011	Vigdøla	1979	-	2013	45.5	1324	7.33	61.51
	7600010	Myklemyr	1978	-	2013	575.8	1323	7.27	61.51
	7500028	Feigumfoss	1972	-	2013	48.0	1294	7.45	61.38
	5000001	Hølen	1923	-	2014	232.7	1277	6.74	60.36
	7200005	Brekke bru	1939	-	2013	268.2	1273	7.11	60.85
	10900021	Driva v/Svon	1970	-	2013	136.0	1263	9.55	62.26
	8800016	Hjelled?a	1982	-	2012	228.9	1255	7.12	61.92
	5000013	Bjoreio	1982	-	2013	262.6	1249	7.40	60.39
	10300040	Rauma v/Horgheim	1971	-	2013	1099.4	1241	7.78	62.47
	4600004	Bondhus	1963	-	2014	60.5	1236	6.27	60.12
	4800005	Reinsnosvatn	1917	-	2013	120.5	1232	6.73	59.97
	5000003	Eidfjordvatn	1928	-	2014	1166.9	1226	7.11	60.43
	10400002	Eikesdalsvatn	1902	-	2013	1093.4	1226	8.12	62.63
	7800008	Bøyumselv	1965	-	2013	40.4	1224	6.74	61.45

10900042	Driva v/Elverhoy bru	1975	-	2013	2437.4	1217	8.69	62.65
10900020	Driva v/GrensehOlen	1964	-	2013	1632.6	1202	9.16	62.57
8300007	Grønengstølsvatn	1965	-	2013	65.7	1184	6.47	61.44
10900029	Dalavatn	1974	-	2013	85.4	1178	8.73	62.61
7500023	Krokenelv	1965	-	2013	45.9	1148	7.40	61.35
10400022	Midtre Mardalsvatn	1976	-	2013	13.7	1132	8.07	62.51
4800001	Sandvenvatn	1908	-	2014	470.2	1090	6.55	60.01
7400018	Fornabu	1982	-	2013	53.1	1054	7.55	61.23
8600012	Skjerdalselv	1982	-	2013	23.8	1045	5.94	61.82
8700010	Gloppenelv v/ Bergheim	1970	-	2013	217.1	1045	6.52	61.72
8300006	Byttevatn	1965	-	2013	104.5	1014	6.35	61.34
7700003	Sogndalsvatn	1962	-	2013	110.9	1000	7.01	61.30
9800004	Oye ndf.	1916	-	2013	138.8	982	6.93	62.07
7100001	Skjerping	1968	-	2014	267.8	969	6.79	60.86
11100005	Toaaa v/Talgoyfoss	1944	-	2013	150.5	917	8.78	62.80
10300020	Isa v/MorstOl bru	1972	-	2013	44.4	912	7.95	62.58
4100008	Hellaugvatn	1981	-	2013	27.5	904	6.19	59.72
3500009	Osali (Botnavatnet)	1982	-	2013	22.5	871	6.62	59.41
8700002	Gloppenelva v/ Eidsfoss	1933	-	2013	614.3	869	6.24	61.76
6200005	Bulken (Vangsvatnet)	1892	-	2014	1092.0	867	6.29	60.63
8300002	Viksvatn (Hestadf- jorden)	1902	-	2013	508.1	842	5.89	61.33
3600013	Grimsvatn	1973	-	2013	34.5	833	6.54	59.58
11200008	Rinna	1969	-	2013	86.2	814	9.40	62.98
8400015	Jølstervatn ndf.	1951	-	2013	384.5	748	6.11	61.49
10400023	Vistdal	1975	-	2013	66.5	737	7.96	62.70
8600010	Åvatn (Ommedals- vatnet)	1974	-	2013	162.1	696	5.92	61.72
4100001	Stordalsvatn	1912	-	2014	130.6	681	6.01	59.68
8000004	Ullebøelv	1927	-	2013	8.3	661	5.79	61.20
9700001	Fetvatn (Fitjavatnet)	1946	-	2013	89.0	591	6.59	62.33
8500004	Straumstad (Sol- hiemsvatnet)	1974	-	2013	109.7	586	5.44	61.65
5500005	Dyrdalsvatn	1977	-	2014	3.3	581	5.52	60.35
11100009	Søya v/Melhus	1974	-	2013	137.6	577	8.60	62.90
3800001	Holmen	1982	-	2014	116.9	556	5.91	59.50

	4200002	Fjellhaugvatn	1963	-	2014	31.9	526	5.83	59.82
	5500007	Eikelandsosen	1980	-	2013	42.7	512	5.76	60.25
	8900001	Horindalsvatn	1900	-	2014	381.7	480	6.12	61.92
	8200004	Nautsundvatn	1908	-	2013	219.0	464	5.39	61.25
	8500002	Blaamannsvatn	1929	-	2013	225.6	429	5.52	61.57
	5500004	Røykenes	1934	-	2014	50.1	307	5.44	60.25
	9100002	Dalsbøvatn	1934	-	2013	25.7	252	5.17	62.16
	10500001	Osenelv v/Oren	1923	-	2013	137.6	235	7.73	62.79
	10100001	Engsetvatn	1923	-	2013	39.9	159	6.62	62.53
	8400020	Holsenvatn	1963	-	2013	5.1	131	6.06	61.43
	10700003	Farstadelva v/Farstad	1965	-	2013	24.2	56	7.16	62.97
	8100001	Hersvikvatn (Hagev-atnet)	1934	-	2013	7.1	51	4.94	61.14
	3900001	Tysvaer	1974	-	2014	3.3	25	5.44	59.33
ØSTLANDET	200291	Tora	1966	-	2013	262.3	1485	7.87	62.01
	200268	Akslen	1934	-	2013	790.9	1467	8.45	61.80
	200013	Nedre Sjodalsvatn	1981	-	2014	480.0	1462	8.93	61.56
	200290	Brustuen	1966	-	2013	253.9	1413	8.30	61.73
	1200013	Rysna	1973	-	2013	50.8	1372	8.72	61.17
	200284	Saelatunga	1966	-	2013	454.8	1370	9.06	61.88
	200025	Lalm	1907	-	2013	3982.4	1326	9.27	61.82
	200303	Dombas	1967	-	2013	497.2	1318	9.10	62.09
	1200197	Grunke	1977	-	2013	184.5	1305	8.70	60.95
	1200137	Gjerdeslaatta	1951	-	2012	774.6	1261	8.73	60.79
	1500079	Orsjoren	1955	-	2012	1177.3	1229	8.26	60.38
	200032	AtnasjO	1916	-	2014	463.2	1205	10.22	61.85
	200614	Rosten	1917	-	2013	1833.9	1186	9.41	61.86
	200145	BaattstO	1896	-	2012	11212.8	1139	10.28	61.33
	200129	DOlplass	1908	-	2013	2014.3	1099	10.45	62.19
	1600132	Gjuvaa	1981	-	2013	33.1	1084	8.80	59.76
	1200097	Bergheim	1920	-	2012	4248.5	1071	9.23	60.47
	1500049	Jalledalsvatn	1962	-	2012	59.1	1028	8.48	60.49
	1200099	Skaalfoss	1965	-	2012	5126.7	1017	9.83	60.08
	1500053	Borgaai	1966	-	2012	94.1	1011	9.01	60.31
	1600066	Grosettjern	1949	-	2014	6.6	1005	8.32	59.84
	1200207	Vinde-elv	1982	-	2013	269.3	981	9.08	61.15
	200011	NaersjO	1930	-	2013	119.1	939	11.48	62.36

	1200070	Etna	1919 - 2013	570.2	936	9.63	60.95
	200607	VaalaasjO	1979 - 2012	126.4	933	9.42	62.17
	1600122	Grovaai	1972 - 2013	42.7	914	8.31	59.64
	1600075	Tannsvatn (Lognvik-vatnet)	1955 - 2013	118.3	905	8.07	59.67
	1200171	HOlervatn	1968 - 2013	79.8	902	9.46	60.71
	1200200	KolbjOrnshus	1978 - 2013	2064.5	870	10.07	60.82
	1600010	Omnesfoss	1963 - 2014	807.7	852	8.99	59.61
	200028	Aulestad	1929 - 2013	866.3	850	10.27	61.22
	1200178	Eggedal	1972 - 2012	309.8	846	9.43	60.15
	31100460	Engeren	1911 - 2013	394.8	837	12.02	61.66
	1600051	Hagadrag	1944 - 2013	727.0	820	8.87	59.43
	200604	Elverum	1871 - 2013	15451.8	817	11.56	60.87
	200267	Mistra bru	1961 - 2013	549.9	807	11.24	61.71
	31100004	Femundsenden (Femunden)	1896 - 2013	1794.0	782	11.94	61.92
	31100006	Nybergsund	1908 - 2013	4424.9	781	12.32	61.26
	1200113	Kraakefjord ndf.	1976 - 2012	702.2	597	9.63	60.04
	1200188	Langtjernbekk	1973 - 2013	4.8	586	9.73	60.37
	200323	Fura	1970 - 2013	42.5	575	11.33	60.89
	1500021	Jondalselv	1919 - 2012	125.9	574	9.55	59.71
	200280	Kringlerdal	1966 - 2013	265.4	519	10.99	60.24
	1600194	Kilen	1962 - 2013	118.5	491	8.79	59.35
	200279	Kraakfoss	1966 - 2012	432.4	445	11.08	60.13
	200142	Knappom	1916 - 2013	1646.0	411	12.05	60.64
	200616	Kuggerud	1977 - 2013	48.4	376	11.73	60.36
	800002	Bjørnegaardssvingen	1968 - 2013	190.2	341	10.51	59.89
	600010	Gryta	1967 - 2013	7.0	300	10.80	59.99
	1200193	Fiskum	1976 - 2012	51.6	277	9.79	59.70
	800006	Sæternbekken	1971 - 2013	6.2	240	10.57	59.94
	31300010	Magnor	1911 - 2013	360.0	240	12.19	59.95
	200633	Stortorp	1979 - 2013	87.1	178	11.52	59.36
	300022	Høgfoss	1976 - 2013	299.3	153	10.86	59.55
SØRLAN	2500024	Gjuvvatn	1971 - 2014	96.8	1139	7.21	59.16
	2600026	Jogla	1973 - 2014	31.1	1002	6.94	58.96
	1900104	Sognedalsåi	1981 - 2013	65.5	777	7.99	59.30
	2200020	Haaverstad	1964 - 2013	1055.0	767	7.43	58.50

N D E T	2000002	Austenaa	1924	-	2013	276.4	763	8.10	58.84
	1900073	Kilåi bru	1968	-	2013	64.4	667	8.30	59.00
	1900079	Gravaa	1970	-	2013	6.3	659	8.04	59.23
	1900096	Storgama ovf.	1974	-	2013	0.5	610	8.65	59.05
	2400009	Tingvatn (Lygne)	1922	-	2014	272.2	588	7.22	58.40
	2200004	Kjølemo	1896	-	2013	1757.7	560	7.53	58.12
	2700025	Gjedlakleiv	1982	-	2014	635.2	527	6.07	58.57
	2700024	Helleland	1977	-	2014	184.7	489	6.15	58.53
	2600020	Årdal	1970	-	2012	77.3	478	6.50	58.54
	2600021	Sandvatn	1971	-	2012	27.5	470	6.77	58.49
	2200016	Myglevatn ndf.	1951	-	2013	182.2	447	7.58	58.45
	1900082	Rauaana	1972	-	2013	8.9	396	8.51	58.97
	1800011	Tjellingtjernbekk	1981	-	2013	2.0	377	8.86	58.93
	2000003	Flaksvatn	1899	-	2013	1780.7	354	8.20	58.33
	1800010	Gjerstad	1980	-	2013	236.2	313	9.03	58.88
	2600029	Refsvatn	1978	-	2014	53.0	297	6.34	58.35
	1900080	Stigvassaai	1972	-	2013	14.5	263	8.52	58.75
	2000011	Tveitdalen	1972	-	2013	0.4	219	8.24	58.39
	2200022	SOgne	1973	-	2014	203.6	198	7.84	58.09
	2800007	Haugland	1918	-	2014	139.4	178	5.65	58.69
	2400008	Møska (Skolandsvat-net)	1978	-	2014	121.0	8	7.07	58.16