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Abstract

This chapter examines past and present studies of variability and changes in atmospheric variables within the North Sea region over the instrumental period; roughly the past 200 years. The variables addressed are large-scale circulation, pressure and wind, surface air temperature, precipitation and radiative properties (clouds, solar radiation, and sunshine duration). Temperature has increased everywhere in the North Sea region, especially in spring and in the north. Precipitation has increased in the north and decreased in the south. There has been a north-eastward shift in storm tracks, which agrees with climate model projections. Due to large internal variability, it is not clear which aspects of the observed changes are due to anthropogenic activities and which are internally forced, and long-term trends are difficult to deduce. The number of deep cyclones seems to have increased (but not the total number of cyclones). The persistence of circulation types seems to have increased over the past century, with ‘more extreme’ extreme events. Changes in extreme weather events, however, are difficult to assess due to changes in instrumentation, station relocations, and problems with digitisation. Without thorough quality control digitised datasets may be useless or even counterproductive. Reanalyses are useful as long as biases introduced by inhomogeneities are properly addressed. It is unclear to what extent circulation over the North Sea region is controlled by distant factors, especially changes in Arctic sea ice.

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2.1 Introduction

Situated in northern central Europe, the North Sea exhibits large climate variability with inflow of a wide range of air masses from arctic to subtropical. For this reason, it is difficult to differentiate between natural and externally forced variability, despite large amounts of historical data. This chapter examines past and present studies of variability and changes in atmospheric variables over the instrumental period; roughly the last 200 years. Research areas lacking consensus in the scientific community are highlighted to stimulate further research.

The main driver of atmospheric variability in the North Sea region is the North Atlantic Oscillation (NAO). Despite its apparent long-term irregularity, the NAO exhibits extended periods of positive or negative index values. No consensus exists with respect to the size of the fraction of interannual NAO variance that cannot be explained by random forcing and is therefore probably influenced by external forcing. Slowly varying natural factors with an effect on European climate, such as the Atlantic Multidecadal Oscillation (AMO), may superimpose long-term trends on atmospheric variability and so be difficult to distinguish from the anthropogenic climate change signal.

The source of atmospheric and surface data influences the results obtained, even in the comparatively data-rich North Sea region. Based on reanalysis data, several studies find positive trends in storm activity over the North Sea region and a northeast shift in storm tracks over the past few decades. However, studies based on direct or indirect historical records of long-term variations in pressure, wind or wind-related proxies, mostly do not identify robust long-term trends. This counter-intuitive result is explained by uncertainties in the long-term historical wind and atmospheric pressure observations, and additional uncertainties arising from the lack of quality control when digitising old data as well as potential biases in the reanalyses due to the fact that the underlying amount of available data is not constant in time. Nevertheless, the northeast shift in storm tracks appears to be a new phenomenon. In contrast, the increase in wind speed and storminess in the latter half of the 20th century does not seem to be unprecedented within the context of historical observations. There are indications of an increase in the number of deep cyclones (but not in the total number of cyclones). There are also indications that the persistence of circulation types has increased over the past century.

Temperatures have increased both over land and over the North Sea. There is a distinct signal in the number of frost days and the number of summer days. While there is a clear winter and spring warming signal over the Baltic Sea region, this is not as clear for the North Sea region. As expected, the

variability in marine temperatures on seasonal timescales is less than for the land temperatures.

Precipitation over land and, but to a lesser extent, over sea is positively correlated with the NAO, and on longer time scales with the AMO. There are indications of an increase in precipitation in the north of the region and a decrease in the south, in agreement with the north-eastward shift in the storm tracks. There are also indications that extreme precipitation events have become more extreme and that return periods have decreased.

From the few datasets available on radiative properties, it may be concluded that there are non-negligible trends together with potential uncertainties and land-sea inhomogeneities which make it difficult to assess these quantities in detail.

Climate change in the North Sea region cannot be investigated in isolation. In particular, what the relation is between changes in the Arctic cryosphere and trends in storminess, number of cyclones, persistence of circulation anomalies and extreme events further south, is an open research question. As analyses of the latter often rely on small datasets covering relatively short time scales, it is difficult to draw statistically significant conclusions. It is therefore essential to make available the large amount of data from past decades that have not yet been digitised. However, it is essential to thoroughly quality-check the data.

2.2 Large-Scale Circulation

2.2.1 Circulation Over the North Sea Region in a Climatological Perspective

From a climatological perspective, the North Sea region is characterised by strong ocean-atmosphere interactions, especially during winter, compared to other regions at similar latitudes (Furevik and Nilsen 2005). These interactions involve transfer of momentum, moisture and various trace gases, mainly carbon dioxide (Takahashi et al. 2002). In addition to the recent warming trend (Delworth and Knutson 2000; Johannessen et al. 2004), the North Sea and nearby regions witnessed climate change during the early 20th century, which was large in comparison to similar latitudes elsewhere (von Storch and Reichardt 1997; Gönner 2003).

Atmospheric circulation in the European/North Atlantic region plays an important role in the regional climate of the North Sea and surrounding land areas (Hurrell 1995; Slonosky et al. 2000, 2001). It is mainly described by the NAO (e.g. Hurrell et al. 2003) which is an expression of the zonality of the atmospheric flow. The North Sea region is controlled by two large-scale quasi-stationary atmospheric patterns, the Icelandic Low (IL) and the Azores High (AH) plus a thermally driven

system over Eurasia with high pressure in winter and low pressure in summer. The dominant flow is therefore westerly, although any other wind direction is also frequently observed, and one of the main factors controlling air-sea interactions in the North Sea region is wind stress. Large-scale processes also constitute one of the main driving mechanisms responsible for the connection between local processes and global change. It is therefore important to pay attention to the recent changes in large-scale flow directly affecting the North Sea region.

The remainder of Sect. 2.2 reviews the status of the large-scale atmospheric variability affecting the North Sea region by focusing on the major teleconnection patterns and their effect on the jet stream. The NAO can be seen as the European expression of a larger-scale phenomenon, the Arctic Oscillation (AO). The relationship between the NAO and AO is briefly discussed in the following section, while a general description of the NAO and its properties is given in Annex 1. The NAO varies on a wide range of time scales from days to decades, reflecting interactions with surface conditions including sea-surface temperature (SST) and sea ice. These changes translate into changes in pressure and winds (Sect. 2.3), temperature (Sect. 2.4) and precipitation (Sect. 2.5) and also affect other variables, like sunshine (Sect. 2.6).

2.2.2 NAO and AO

The strength of the westerlies and the eddy-driven jet stream over the North Atlantic and western Europe are controlled by various factors including the pressure difference between the IL and AH as the main centres of action of the NAO (Wanner et al. 2001; Hurrell et al. 2003; Budikova 2009). The NAO and its changes can be understood as signals in the surface pressure field of jet stream variation (Hurrell and Deser 2009) and, as such, are often referred to as the regional expression of the AO, which describes sea-level pressure variations between the Arctic and northern hemisphere lower latitudes (Budikova 2012) or, in other words, variability in the strength of the polar vortex. The AO, also termed the Northern Annular Mode (NAM), was first identified by Lorenz (1951) and named by Thompson and Wallace (1998). Its positive phase is characterised by low surface pressure in the Arctic and a generally zonal (west to east) jet stream, thus keeping cold air in the Arctic. When the AO index is negative, there is high pressure in the Arctic and a stronger meridional (north to south or vice versa) component of the jet stream so that cold air can extend to lower latitudes. With respect to the North Sea region, the state of the AO controls the westerly flow and the storm track.

The AO is strongly correlated with the NAO, and the latter can be viewed as the signature of the former over the

North Atlantic region. Therefore, the discussion below and in Annex 1 focuses on the NAO. Note, however, that there is extensive literature on the relationship between the NAO and AO, including studies on the robustness and consistency of one versus the other (Ambaum et al. 2001). Even though the AO and NAO are strongly correlated, their relationship is not linear (Kravtsov et al. 2006; de Viron et al. 2013). A fully three-dimensional picture discussing the connection between the AO and stratospheric circulation anomalies is provided by, for example, Ripesi et al. (2012).

2.2.3 Temporal and Spatial Changes in the NAO

Given the importance of the NAO in the North Atlantic and European climate, substantial efforts have been made to understand its variability in order to gain insight into its potential predictability. The NAO index varies on a wide range of time scales ranging from days to decades. As Fig. A1.1 in Annex 1 shows, the long-term behaviour of the NAO is irregular, and there is large interannual and inter-decadal variability, reflecting the interaction with the surface, including SST and sea ice. Focusing on the 20th century, a period with predominantly positive NAO index values prevailed in the 1920s, followed by mainly negative values in the 1960s. Since then, a positive trend has been observed, which means more zonal circulation with mild and wet winters and increased storminess in central and northern Europe, including the North Sea region (e.g. Hurrell et al. 2003). This was especially the case in the early 1990s, raising claims that this behaviour was ‘due to anthropogenic climate change’. After the mid-1990s, however, there was a tendency towards more negative NAO index values, in other words a more meridional circulation and according to Jones et al. (1997), Slonosky et al. (2000, 2001) and Moberg et al. (2006), the strongly positive NAO phase in the early 1990s should be seen as an element of multi-decadal variation comparable to that at the start of the 20th century rather than as a trend towards more positive NAO values. It should also be noted that the winter of 2010/2011 had one of the most negative NAO indices in the record (Jung et al. 2011; Pinto and Raible 2012).

Intraseasonal variability in the NAO can be reasonably well described by a Markov process or first-order autoregressive (AR1) model (Feldstein 2000), although the observed skewness of the NAO index (Woollings et al. 2010; Hannachi et al. 2012) or the particularly enhanced persistence of the negative phase (Barnes and Hartmann 2010) are not captured very well. Nonlinear Rossby wave breaking mechanisms have been proposed to explain the intraseasonal variability in the NAO (e.g. Benedict et al. 2004; Franzke et al. 2004; Woollings et al. 2008).

Interannual variability in the winter mean NAO index is also found to be linked to the intraseasonal transitions between the positive and negative phases of the NAO pattern (Luo et al. 2012).

Several studies have attempted to quantify the fraction of interannual NAO variance that can be explained by ‘climate noise’, namely by random sampling of the intraseasonal variations (Feldstein 2000; Keeley et al. 2009; Franzke and Woollings 2011). As the NAO exhibits no preferred periods on the interannual and longer timescale (Hurrell and Deser 2009), the results of these studies differ widely, meaning as yet no consensus on the fraction of interannual NAO variability that cannot be explained by such noise and which is thus likely to be forced externally (Stephenson et al. 2000; Feldstein 2002; Rennert and Wallace 2009). Several external forcing mechanisms have been proposed, including bottom boundary conditions of local SST (Rodwell et al. 1999; Marshall et al. 2001) and sea ice (Strong and Magnusdottir 2011), volcanoes (Fischer et al. 2007), solar activity (Shindell et al. 2001; Spanghel et al. 2010; Ineson et al. 2011), and even stratospheric influence (Blessing et al. 2005; Scaife et al. 2005), including the quasi-biennial oscillation (Marshall and Scaife 2009) and stratospheric water vapour trends (Joshi et al. 2006). Remote SST forcing of the NAO originating from as far as the Indian Ocean was proposed by Hoerling et al. (2001) and Kucharski et al. (2006), while Cassou (2008) proposed an influence of the Madden-Julian Oscillation, but no consensus has been reached.

The positive trend in the NAO from the 1960s to the 1990s has been a particular focus of interest, and it has been a concern that many climate models have been unable to simulate the observed trends (Gillett 2005). Even though some models simulate quite large natural variability (Selten et al. 2004; Semenov et al. 2008), they still fall short of the strongest observed 30-year trend (Scaife et al. 2009). While Gillett et al. (2003) suggested that anthropogenic forcing could have contributed to the positive NAO trend there is no agreement on this, and concerns over the ability of models to represent NAO variability mean that attribution attempts should be treated with caution. The downturn in the NAO since the mid-1990s has brought its relation to climate change under further doubt (Cohen and Barlow 2005).

The NAO pattern is not entirely stationary, neither geographically nor with respect to season (Fig. 2.1). Although the amplitude is greatest and the explained variance highest in winter, the NAO is present all year round with varying strength and position. In particular, there is a westward shift of the southern centre of action in spring and an eastward displacement of both centres in autumn. Portis et al. (2001) developed a seasonally and geographically varying ‘mobile’ NAO, which is obtained from sea-level pressure data by taking into account the migration of the AH nodal points. Other techniques also exist, such as optimally interpolated

patterns, trend empirical orthogonal functions (EOFs; Hannachi 2007a, 2008) and cluster analysis (Cheng and Wallace 1993; Hannachi 2007b, 2010).

Regression of the NAO on near-surface winds (Fig. 2.2) illustrates the well-known fact that the positive NAO phase is accompanied by stronger than average westerlies in the mid-latitudes right across the North Sea region into Scandinavia (Hurrell and Deser 2009), thus contributing to enhanced precipitation in the northern mid-latitudes.

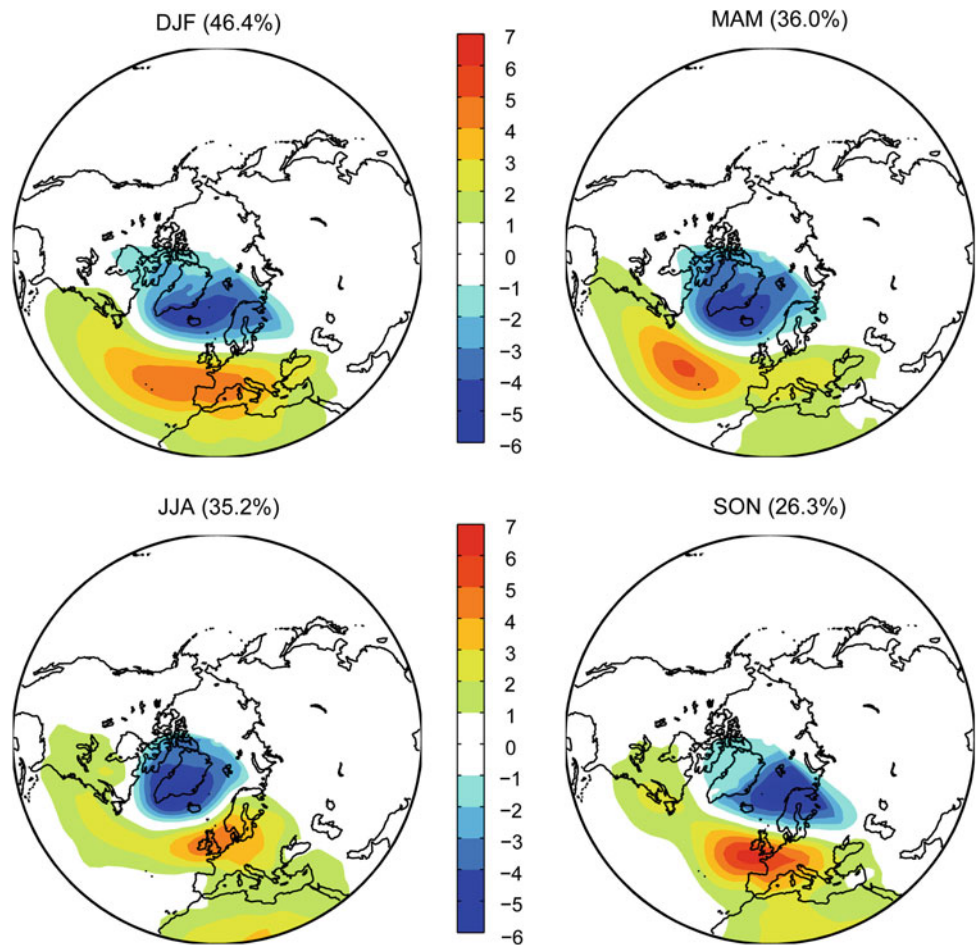
2.2.4 Other Modes of Variability

The NAO is essentially a signal of variations in the Atlantic eddy-driven jet stream, but one pattern is not sufficient to fully describe the jet variability. The eddy-driven jet stream variability and regimes can, in fact, be described by at least two modes of variability—the NAO and the East Atlantic Pattern (EA; Wallace and Gutzler 1981; Woollings et al. 2010; Hannachi et al. 2012). The latter is defined as the second prominent mode of atmospheric low frequency variability over the North Atlantic. It appears throughout the year, but is more prominent in winter. Comprising a north-south dipole of anomalies, the EA pattern¹ resembles the NAO, but with anomaly centres displaced south-eastward and thus is sometimes interpreted as a ‘southward shifted’ NAO (e.g. Barnston and Livezey 1987). The positive phase of the pattern is associated with wetter-than-average conditions over northern Europe and Scandinavia. The EA pattern has particularly strong fluctuations in the low frequency component showing a consistent positive trend over recent decades. However, this is indicative of trends over the Mediterranean region rather than the North Sea region (Woollings and Blackburn 2012). Trouet et al. (2012) introduced a ‘summer NAO’ pattern with a blocking high over the British Isles and a low over south-eastern Europe in its positive phase which is most pronounced on interannual timescales and resembles the EA.

A third pattern that has some impact on the North Sea and Scandinavia is the Scandinavian pattern¹ (Wallace and Gutzler 1981), characterised primarily by a centre over Scandinavia and weaker centres of opposite polarity over western Europe. The positive phase of the Scandinavian pattern is frequently linked to the occurrence of atmospheric blocking over northern Europe (Barriopedro et al. 2006) and is therefore characterised by below-average precipitation in this region, as well as by large interannual and decadal variability over the past 50 years (Crocini-Maspoli et al. 2007; Tyrlis and Hoskins 2008), which may be related to longer term variability of blocking across the Atlantic Ocean

¹www.cpc.ncep.noaa.gov/data/teledoc/telecontents.shtml.

Fig. 2.1 Leading EOF of seasonal mean sea-level pressure (SLP) anomalies over the North Atlantic (20°–0°N, 90°W–40°E for the period 1948–2014. The percentage of explained variance is given above each panel



(Häkkinen et al. 2011). Before the 1980s the positive phase was dominant, this was followed by a negative phase between 1980 and 2000. The pattern amplitude has weakened over the past decade compared to the earlier part of the record.

Finally, on longer timescales, atmospheric conditions in the North Sea region are significantly influenced by the Atlantic Multidecadal Oscillation (AMO—more correctly termed Atlantic Multidecadal Variability, as there is no temporal regularity), which describes basin-wide variations in the temperature of the North Atlantic (Knight et al. 2005a, b) on the order of decades. These are particularly important in summer (Sutton and Hodson 2005) when the atmospheric response resembles a pattern termed the ‘summer NAO’ (Folland et al. 2009; Ionita et al. 2012b), see also Fig. 2.1. Warm periods were observed prior to 1880, between 1930 and 1965 and after 1995, and cool periods between 1900 and 1930 and between 1965 and 1995.

2.2.5 Summary

The NAO is the dominant mode of near-surface pressure variability over the North Atlantic and Europe, including the

North Sea region. Amplitude and explained variance are largest in winter, but the NAO impacts the North Sea region throughout the year. Despite its apparent long-term irregularity, the NAO exhibits extended periods of positive or negative index values. It is therefore important to quantify the fraction of interannual NAO variance that cannot be explained by random forcing and so is likely to be influenced by external forcing. There is no consensus on the size of this fraction, or on the possible external forcing mechanisms.

2.3 Atmospheric Pressure and Wind

A typical characteristic of the climatology of the North Sea region is the large variability in meteorological variables on multiple time scales. The strong increase in wave height and storminess between the 1970s and the 1990s over the North Sea and North Atlantic (Carter and Draper 1988; Hogben 1994) raised public concern about a roughening wind climate and speculations about whether global warming might have an impact on storminess (Schmidt and von Storch 1993). With the availability of many more observations and gridded

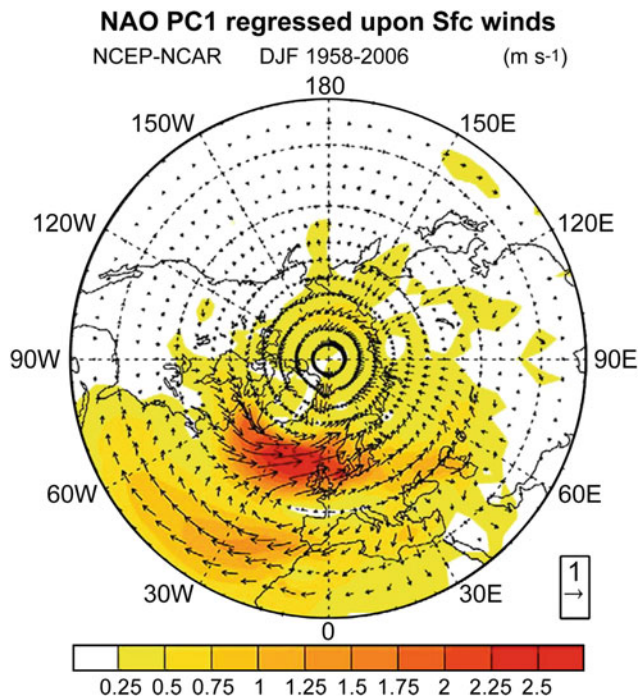


Fig. 2.2 Wind speed and direction associated with a 1 standard deviation change in the NAO index. The index is obtained from an EOF analysis of NCEP/NCAR sea-level pressure data (1958–2006) over the North Atlantic sector. Colour scale: m s^{-1} , unit vectors: 1 m s^{-1} (Hurrell and Deser 2009)

reanalysis data sets, detailed studies have significantly improved understanding of the atmospheric circulation and related winds over the North Atlantic and North Sea.

Owing to the large climate variability, results regarding changes or trends in the wind climate are strongly dependent on the period and region considered (Feser et al. 2015b). Through the strong link to large-scale atmospheric variability over the North Atlantic, conclusions about changes over the North Sea region are best understood in a wider spatial context. The following sections summarise studies on variations and trends in pressure and wind for recent decades and place these in the context of studies about changes over the last roughly 200 years.

2.3.1 Atmospheric Circulation and Wind Since Around 1950

The period since about 1950 is relatively well covered by observational data. The beginning of the satellite period in 1979 led to further substantial improvements in global data coverage, especially over the oceans and in data-sparse regions. Although in situ wind observations allow direct analysis of this variable, in particular over the sea (e.g. International Comprehensive Ocean-Atmosphere Data Set, ICOADS; Woodruff et al. 2011), the information is often

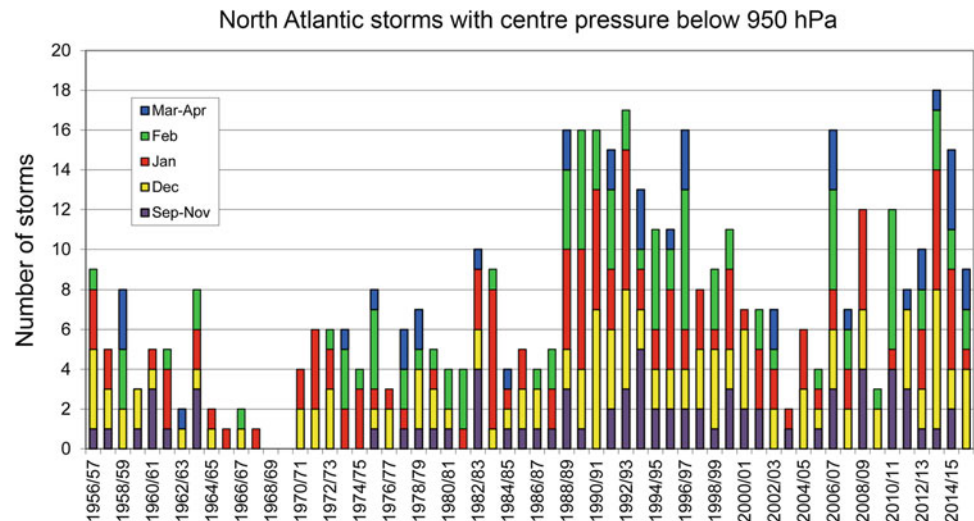
predominantly local and inhomogeneities make the straightforward use of these data difficult, even for recent decades (Annex 1). Examples include an increase in roughness length over time due to growing vegetation or building activities, inhomogeneous wind data over the German Bight from 1952 onwards (Lindenberg et al. 2012) or ‘atmospheric stilling’ in continental surface wind speeds due to widespread changes in land use (Vautard et al. 2010).

Most studies therefore do not use direct wind observations, but instead rely on reanalysis products such as NCEP/NCAR (from 1948 onwards; Kalnay et al. 1996; Kistler et al. 2001), ERA40 (from 1958 onwards; Uppala et al. 2005) or, more recently, ERA-Interim, starting in 1979 (Dee et al. 2011) and the 20th Century Reanalysis 20CR (from 1871 onwards; Compo et al. 2011) and other reanalysis products, see Electronic (E-)Supplement Sect. S2.2. Making use of all available observations, a frozen scheme for the data assimilation of observations into state-of-the-art climate models is used to minimise inhomogeneities caused by changes in the observational record over time. However, studies indicate that these inhomogeneities cannot be fully eliminated (see E-Supplement S2). In addition, systematic differences between the underlying forecast models, such as due to their different spatial resolutions (Trigo 2006; Raible et al. 2008) and differences in detection and tracking algorithms (Xia et al. 2012) may affect cyclone statistics (for example changes in their intensity, number and position). Apart from these differences and inhomogeneities, the number of detected cyclones and their intensities show very high correlations between reanalyses (Weisse et al. 2005; Raible et al. 2008).

Three recent studies cover a continental-scale area. Franke (2009) manually counted the number of strong low pressure systems (central pressure below 950 hPa) over the North Atlantic (north of 30°N) from weather maps of the Maritime Department of the German Weather Service (‘Seewetteramt’), see Fig. 2.3, which shows generally weak activity prior to 1988 and enhanced activity for the following decade, followed by a decrease to 2006.

Since then, the number of deep cyclones has again increased despite the predominantly negative NAO (Fig. A1.1 in Annex 1), and the maximum value with 18 such cyclones was observed in 2013/2014. Despite large decadal variations, there is still a positive trend in the number of deep cyclones over the last six decades, which is consistent with results based on NCEP reanalyses since 1958 over the northern North Atlantic Ocean (Lehmann et al. 2011). Using an analogue-based field reconstruction of daily pressure fields over central to northern Europe (Schenk and Zorita 2012), the increase in deep lows over the region might be unprecedented since 1850 (Schenk 2015). Barredo (2010) investigated adjusted storm losses in the period 1970–2008 on a European scale but did not find any trends despite a roughened wind climate. Based on the CoastDat2 reanalysis

Fig. 2.3 Number of low pressure systems on the North Atlantic with a core pressure of 950 hPa or below, 1956/57–2015/16 (after Franke 2009, updated)



(Geyer 2014) and using E-OBS pressure data (van den Besselaar et al. 2011), von Storch et al. (2014) did not find robust evidence for supporting claims that the intensity of the two strong storms in late 2013 would be beyond historical occurrences and that the recent clustering of storms should be related to anthropogenic influence.

2.3.1.1 North Sea Region

Different studies based on reanalyses confirm the strong increase in wind speeds and wave heights observed over the North Sea region since the 1970s. Covering the period since about 1950, a positive trend is visible in annual storm activity in the NCEP, ERA40 and 20CR datasets, although the most recent decade shows a decrease in wind speed (e.g. Matulla et al. 2007; Donat et al. 2011) with no notable trend in mean wind speed for the period as a whole (1948–2014) over the North Sea (Fig. 2.4).

Siegismund and Schrum (2001) analysed decadal changes in wind forcing over the North Sea based on the NCEP reanalysis for the period 1958–1997 (Fig. 2.4). Over the 40 years, mean annual wind speeds increased by 10 %, mainly due to an increase in autumn and winter (ONDJ) after the 1960s and in late winter (FM) since the mid-1980s, but no trend was found for summer. Increased wind speeds are accompanied by an increase in WSW wind directions in autumn and winter (ONDJ) over the last three decades compared to the first (1958–1967). The enhanced mean winter wind speeds agree with an increase in the mean winter NAO index with a correlation of 0.69 for the year-to-year variations (Fig. 2.4). An update of the graphic for 1948–2014 shows a return to average wind conditions in the last decade with no notable trend remaining. The updated correlation with the NAO is 0.73.

Weisse et al. (2005) compared an NCEP-driven regional climate simulation (50 km resolution) with wind speeds

from marine stations and found relatively good agreement. For the period 1958–2001, they found increasing storminess over most marine areas north of 45°N with a small, but significant positive trend over the North Sea and Norwegian Sea. The relative increase in storm frequency is largest over the southern North Sea across Denmark towards the Baltic Sea (1–2 % per year). The number of storms was lowest during the 1970s (with some notable exceptions, in particular the ‘Capella’ storm in January 1976) and peaked around 1990–1995. However, since then, a decrease in storm frequency has been observed which is confirmed by other studies (e.g. Matulla et al. 2007). Based on a high-resolution model hindcast forced by NCEP reanalyses for the storm season (November to March) 1958–2002, simulated storm-related sea-level variations confirm a significant positive trend for the Frisian and Danish coast (Weisse and Plüß 2006) while insignificant changes in mean and 90th percentile water levels are found for the UK, the Dutch coast and the German Bight. However, Weisse and Plüß also noted that positive trends in observations are higher than those in the NCEP-driven hindcast (see Chap. 3).

In contrast to the strong increase in wind speed in the NCEP reanalysis, Smits et al. (2005) found no increase in geostrophic wind speeds and even a decrease in homogenised wind observations for inland stations in the Netherlands for the period 1962–2002, while coastal stations show an increase consistent with NCEP. Smits et al. (2005) claimed that this is due to inconsistencies in the NCEP data, but it might also be that the ‘atmospheric stilling’ postulated by Vautard et al. (2010; see E-Supplement Sect. S2.1) can explain these differences.

2.3.1.2 Northern North Atlantic Region

As variations in atmospheric circulation and the wind climate over the North Sea show a high co-variability with

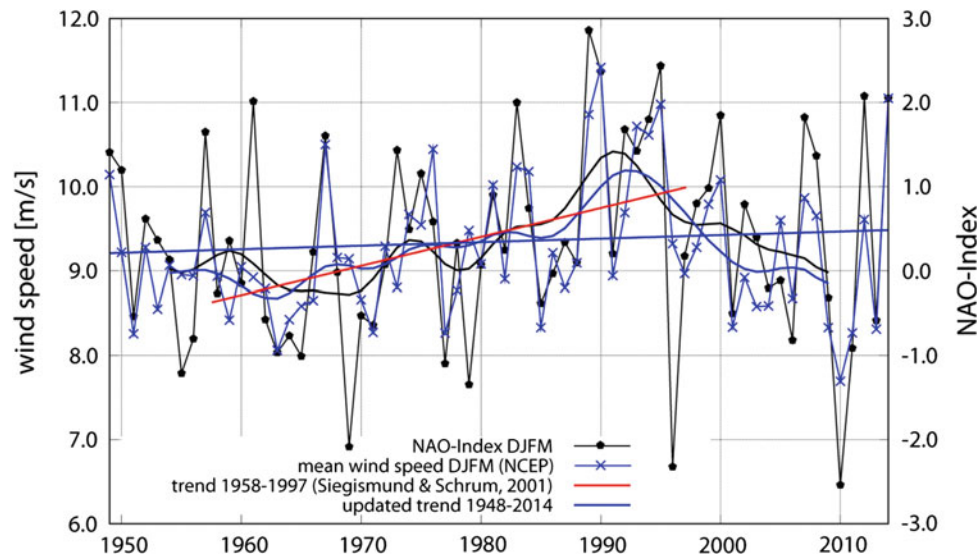


Fig. 2.4 Time series of mean seasonal wind speed derived from NCEP/NCAR reanalysis over the North Sea (blue) and the NAO index (black) for winter (DJFM) 1948–2014 recalculated and updated as in Siegmund and Schrum (2001), by F. Schenk. The positive trend (red

linear fit) from this study has ended due to more average wind conditions in the last decade (blue linear fit). Smoothed lines are shown to highlight decadal-scale variations (11-year Hamming window)

large-scale variations and cyclonic activity over the North Atlantic, the majority of studies focus on the whole Euro-Atlantic region rather than on the North Sea alone. These studies show that the increase in wind speed or storminess is related to a general intensification of storm tracks over the North Atlantic north of about 55°N.

Based on NCEP reanalyses, Chang and Fu (2002) found a significant increase of about 30 % in decadal mean winter (DJF) storm track intensity for the period 1948–1998, with values about 30 % higher in the 1990s than during the late 1960s and early 1970s. The strengthening of storm track intensity is most pronounced over the North Atlantic albeit the trend compared to the few available conventional measurements seems to be overestimated in the NCEP reanalysis (Chang and Fu 2002; Harnik and Chang 2003; see E-Supplement Sect. S2.2). Geng and Sugi (2001) also found a significant increase in the number of North Atlantic cyclones and a significant intensifying trend for the cyclone central pressure gradient.

Similar results were found by Raible et al. (2008) based on ERA40 and NCEP for the period 1958–2001. The authors highlighted seasonal differences and reported a slight increase in number and a significant increase in intensity in winter (DJF) for the northern North Atlantic including the North Sea region (55°N–70°N, 45°W–15°E), a negative tendency in summer (JJA) and a non-significant increase in autumn (SON). The intensification in winter is stronger in NCEP than in ERA40 and also includes spring (MAM) in agreement with a similar increase in the number of deep

lows (<980 hPa) in both seasons (Lehmann et al. 2011). The number of deep lows shows a minimum in the 1970s, followed by a strong increase.

The general enhancement in winter storm track intensity is accompanied by a northward shift in the storm track of 2–5° (depending on the data set) for NCEP (1948–1997; Chang and Fu 2002), ERA15 (1979–1997; Sickmoeller et al. 2000) and ERA40 (1958–2001; Wang et al. 2006), in agreement with a shift and intensification of deep lows (<980 hPa) towards the NE over the North Atlantic in the period 1948–2008 (Lehmann et al. 2011).

2.3.1.3 Southern North Atlantic Region

The general increase in the number of deep cyclones and storminess over the northern North Atlantic and North Sea is accompanied by partly opposing tendencies for the mid-latitudes south of 55–60°N (Gulev et al. 2001), suggesting a general northward shift in the cyclone tracks, consistent with findings of McCabe et al. (2001).

The north-south contrast in the sign of trends was also confirmed by Trigo (2006) who applied an objective detection and tracking algorithm to NCEP and ERA40 for winter (DJFM) 1958–2000 to produce a storm-track database for different stages in the cyclone lifecycle over the North Atlantic and Europe (20°–70°N; 85°W–70°E). On a seasonal basis, the trend is generally positive at higher latitudes (mostly due to an increased frequency of moderate and intense storms) and negative in the subtropical belt. Wang et al. (2006) and Raible et al. (2008) also drew similar

conclusions. All these results consistently show a northward shift in mean storm track position since about 1950 (Feser et al. 2015a).

2.3.2 Regional Variations in Pressure and Wind Since Around 1800

Observed changes in cyclone characteristics and winds over the last 40–60 years pose the question as to whether these changes merely reflect (multi-)decadal variations or whether they reflect long-term change. This section summarises current knowledge about the historical evolution of pressure and wind in the last 200 years over the Euro-Atlantic region.

Information about long-term variations in pressure and wind rely on multiple direct and indirect observations. They provide qualitative to semi-quantitative historical descriptions (the latter quite far back in time), such as storm surge-related damage on the Dutch coast since the 15th century (de Kraker 1999) or daily weather diaries like those at the observatory of Armagh (Ireland) since 1798 (Hickey 2003). Direct measurements such as surge levels at Liverpool since 1768 (Woodworth and Blackman 2002) and Cuxhaven since 1843 (Dangendorf et al. 2014) or wind records in the Dublin region since 1715 (Sweeney 2000) also provide important information on variations in the storm climate. The difficulty with these observations is that they often represent local conditions and so often exhibit inhomogeneities to an unknown extent.

As recommended by WASA Group (1998), most studies use pressure observations (which are more homogeneous over time than wind measurements) to derive wind and storm indices (e.g. WASA Group 1998; Klein Tank et al. 2002). The usefulness of typically-used pressure-based indices has recently been re-assessed and confirmed. Even single-station pressure indices such as strong pressure changes over 6- or 24-h periods or the annual number of deep lows provide useful information about long-term variations in the wind and storm climate (Krueger and von Storch 2011). The information content to describe long-term variations in the statistics of pressure and wind is even higher for indices of geostrophic wind speeds calculated from triangles of daily pressure observations (Krueger and von Storch 2012). The correlation of geostrophic wind speeds calculated from station triplets with real model wind speeds is especially high over open terrain and sea areas (i.e. regions that often lack conventional observations). Although pressure-based indices provide only an indirect link to real wind speeds or storminess, they can be considered a valid approach for assessing long-term statistics of pressure and wind (Krueger and von Storch 2011, 2012).

As historical information on pressure and wind mostly relates to regional or local scale rather than gridded fields,

the studies presented in the following sections are discussed by region. Figure 2.5 provides an overview of potential long-term trends in the wind and storm climate.

2.3.2.1 North Atlantic and Iceland

The region north of around 55–60°N is of special interest regarding changes in the intensity or position of the main storm tracks. Historical information here is limited to the Shetland, Orkney and Faroe Islands as well as Iceland. The longest pressure-based wind index, the annual mean of absolute pressure changes over 24 h, suggests a significant positive trend over Iceland in the period 1823–2006 (Hanna et al. 2008), but no robust trend exists over the Norwegian Sea (since 1833) or the North Sea (since 1874). The latter is consistent with the result of Schmith et al. (1998) who found no significant trend for absolute daily pressure tendencies for stations around the NE Atlantic for winter 1871–1997. Analysis of high annual geostrophic wind speed percentiles over the NE Atlantic also indicates no significant change since the late 19th century (WASA Group 1998; Alexandersson et al. 2000; Matulla et al. 2007; Wang et al. 2009a, 2011). For the shorter period 1923–2008, positive trends exist over the northern NE Atlantic for spring (Wang et al. 2009a) which is in agreement with the intensification and northeast shift in cyclone activity in the last 60 years.

A positive trend also exists for the annual frequency of zonal weather types in the winter half-year 1881–1992 (Schiesser et al. 1997). As this weather type (Großwetterlagen) classification (Baur 1937; Hess and Brezowsky 1952, 1977; Hoy et al. 2012) relies on historical weather maps over the North Atlantic and Europe, the trend should be viewed with caution due to an improvement over time in detecting smaller lows (E-Supplement S2).

2.3.2.2 British Isles

There are many historical wind and wind-related documents and records for Great Britain and Ireland. Although robust trend estimates have not been undertaken, available information suggests large multi-decadal variations but no overall long-term trends (e.g. Sweeney 2000 for the number of storms per decade from historical reports of the Dublin region 1715–1999). For the shorter period 1903–1999, however, adjusted wind observations do show a decrease in the decadal number of storms exceeding 50 knots (25.7 m s^{-1}). The record of storms from a daily weather diary of Armagh (Ireland) 1798–1999 (Hickey 2003) also indicates similarly large variations to those of recent decades, although observer bias reduces reliability over time. Anemometer readings at the station show no obvious trend in the number of gale days per year for the period 1883–1999.

For the Irish Sea, tide gauge records at Liverpool provide an indirect estimate of long-term variations in storms. These show a negative tendency for the annual maximum surge at

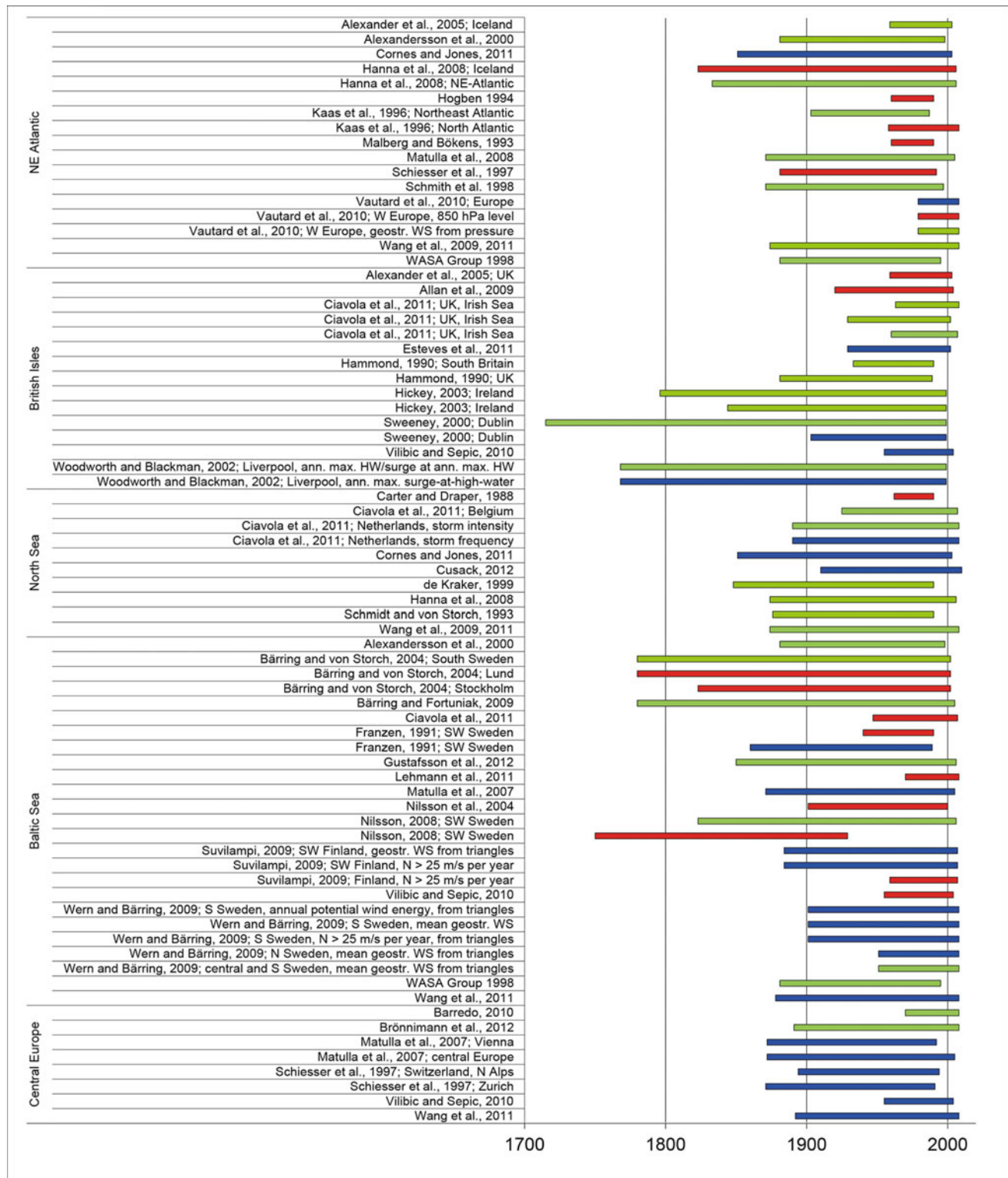


Fig. 2.5 Long-term trends for storminess over the Euro-Atlantic region based on different historical and observational sources (Feser et al. 2015a). Red, blue and green colours indicate positive, negative or no trend, respectively

high water for the period 1768–1999 (Woodworth and Blackman 2002) but no trend in the annual maximum high water level or the surges at annual maximum high water for this period. For shorter periods, no trends are found for maximum monthly wind speed observations for the Irish Sea 1929–2002 (Ciavola et al. 2011) while Esteves et al. (2011) found a weak but significant negative trend for monthly mean wind speeds at the Bidston observatory on the northern Irish Sea coast over the same period.

For southern England, Hammond (1990) used different stations to calculate an annual windiness index taking the annual average of monthly mean wind speeds for the Boscombe Down area. For the period 1881–1989, the annual windiness index does not show any long-term trend. For the end of the 17th and the first half of the 18th century (Late Maunder Minimum), wind indices were derived from ship logbooks for the Øresund region (Frydendahl et al. 1992) and the English Channel (Wheeler et al. 2009). Ship logbooks offer a unique source of information about past wind climates, as discussed for example by Küttel et al. (2009). Wind indices based on these logbooks suggest a generally stationary wind climate with large decadal variations. For the period 1920–2004, the 1930s show another period of more severe storms for the British Isles based on extreme three-hourly pressure changes (Allan et al. 2009).

2.3.2.3 North Sea Region

Limited historical information about dike repair costs for northern Flanders indicates no obvious visible long-term trend for the period 1488–1609 (de Kraker 1999). However, storm-related damage does appear to reflect similarly large multi-decadal variations as for storm observations over recent decades. An update of the historical index using water level observations at Flushing at the mouth of the Western Scheldt estuary for 1848–1990 shows a notable increase in spring tides in the 1990s which at least partly reflects the 30 cm rise in sea level over this period.

Other datasets do not indicate long-term trends in storm intensity. Surge information for the Netherlands shows a decrease in storm frequency over the period 1890–2008 (Ciavola et al. 2011). Also, Cusack (2012) found a weak negative tendency for the decadal running mean of the annual number of damaging storms and a related storm loss index calculated from homogenised wind observations of the Netherlands for 1910–2010, but did find large decadal variations for stormy conditions in the 1920s and 1990s. Storm intensity estimates derived from wind, wave and surge observations from Belgium for the period 1925–2007 show no trend (Hossen and Akhter 2015).

An analysis of geostrophic wind speeds for the German Bight shows no robust trends for the period 1876–1990 (Schmidt and von Storch 1993). But when the record is extended to include the period up to 2012 (Fig. 2.6) a

tendency for decreasing wind speed in the upper percentiles becomes visible, corroborating direct wind, surge and wave observations from Belgium and the Netherlands, and findings by Rosenhagen et al. (2011). Analogue-based storminess shows a good correlation with the German Bight index and indicates no long-term trend since 1850 (Schenk 2015). Wang et al. (2011) found significant negative trends over the North Sea and surrounding land areas for the 99th percentile of geostrophic wind in summer, but no robust trends in other seasons. Direct wind observations from Skagen in northern Denmark also suggest decreasing overall storminess for the period 1860–2012 with extremely high storminess prior to 1875 (Clemmensen et al. 2014).

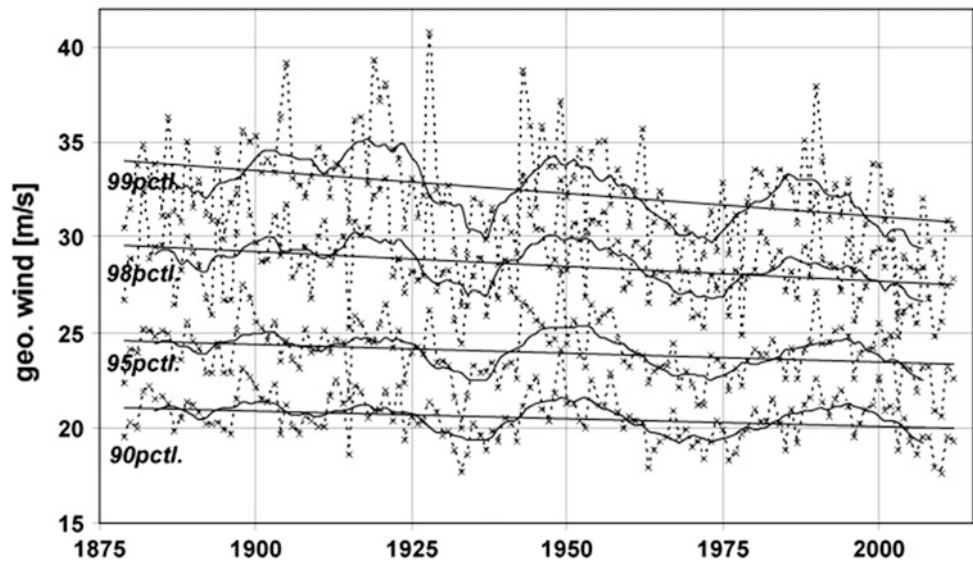
2.3.2.4 Northern Alps and Central Europe

Although not directly linked to the North Sea wind climate, observations from further south are also useful to help understand variations in large-scale atmospheric circulation and give indications about a northward displacement in storm tracks. For Vienna, the number of gale days (above 8 Bft, 17.2 m s^{-1}) show a clear decrease for the period 1872–1992 (Matulla et al. 2007), however based on non-homogenised observations. This is corroborated by significant negative trends in the number of days exceeding 7, 8 and 9 Bft (>13.9 , 17.2 and 20.8 m s^{-1} , respectively) in northern Switzerland for the period 1894–1994 (Schiesser et al. 1997). The duration of strong winds (>7 Bft) also shows a negative trend for Zürich for 1871–1991 except in winter (Brönnimann et al. 2012). Negative trends are also found for central Europe (Matulla et al. 2007; Wang et al. 2011). In contrast, Stucki et al. (2014) found no clear trends, but large interdecadal variability, over Switzerland.

2.3.3 Trends in the 20th Century Reanalysis Since 1871

As shown in Fig. 2.5, the majority of studies using observational storm proxies find no robust trends, some even a negative tendency, for the wind and storm climate in historical pressure and wind observations. In contrast, 20CR (Compo et al. 2011) suggests significant upward trends for storminess over data-sparse regions like the NE Atlantic (Donat et al. 2011) while Bett et al. (2013) did not find a clear trend over Europe. Closer inspection reveals that the agreement of 20CR and wind observations over land like Zürich is reasonable (Brönnimann et al. 2012), but there are discrepancies over sea (Krueger et al. 2013; Schenk 2015). This is because 20CR, like other reanalyses, assimilates all available pressure observations at a given time step which leads to a strong increase in assimilated land pressure observations (and to a lesser extent also sea pressure observations) over time. Following Krueger et al. (2013),

Fig. 2.6 High annual percentiles of geostrophic wind speeds over the German Bight after Schmidt and von Storch (1993) updated and reproduced for 1879–2012. Running 11-year means and linear trends are displayed to highlight long-term variations (data by G. Rosenhagen, figure by F. Schenk)



inconsistencies between 20CR and pressure-based storm indices over data-sparse regions increase back in time as the number of assimilated stations by 20CR, mainly over sea areas, decreases. Spurious pressure trends in data-sparse regions, identified in the NCEP/NCAR reanalysis (Hines et al. 2000) might also affect 20CR (E-Supplement Sect. S2.3).

2.3.4 Summary

Different studies mainly based on reanalysis data show positive trends in storm activity over the NE Atlantic and North Sea together with a northeast shift in the position of storm tracks over the last 40–60 years. This is also reflected in a roughened wind and wave climate, although a return to average conditions beginning at the end of the 20th century has clearly reduced trends from earlier publications. As summarised in Fig. 2.5 direct or indirect historical records of long-term variations in pressure, wind or wind-related proxies mostly show no robust long-term trends for the last 100 years or more. Large decadal variations seem to dominate for centuries.

While the increase in wind speeds and storminess in the latter half of the 20th century does not seem unprecedented in the context of historical observations, the northeast shift in storm tracks in this period may be a new phenomenon. The long-term decrease north of the Alps mainly results from a less stormy period during the 1990s compared to the North Sea and North Atlantic while the period at the end of the 19th century is comparably windy. The less stormy 1990s further south are consistent with the northeast shift in storm

tracks and the decrease in winter cyclone activity in the mid-latitudes. This northeast shift together with the trend pattern of decreasing cyclone activity for southern mid-latitudes and increasing trends north of 55–60°N after around 1950 seems consistent with scenario simulations to 2100 under increasing greenhouse gas concentrations (e.g. Ulbrich et al. 2009; Feser et al. 2015a; see Chap. 5). This corroborates the findings by Wang et al. (2009b) that combined anthropogenic and natural forcing had a detectable influence on this pattern of atmospheric circulation, storminess and ocean wave heights during boreal winter 1955–2004 while an analysis for the first half of the 20th century is less likely to be dominated by external forcing.

Uncertainties remain not only for long historical wind and pressure observations (e.g. Lindenberg et al. 2012; Wang et al. 2014), but also for 20CR that to a large extent relies on these observations (Brönnimann et al. 2013; Krueger et al. 2013; Dangendorf et al. 2014; Schenk 2015). These are discussed in detail in E-Supplement S2. As a better understanding of long-term variations versus trends, and their link to atmospheric circulation is crucial for any regional climate change analysis, data rescue initiatives and digitisation initiatives such as data.rescue@home (www.data-rescue-at-home.org) or oldWeather (www.oldweather.org) are essential for further improvements towards the homogenisation of observations and reanalyses prior to about 1950. Such data can then be used in reanalysis projects such as ACRE (Atmospheric Circulation Reconstructions over the Earth; www.met-acre.org). However, in the light of problems apparently introduced into the WASA dataset during the digitisation step (see E-Supplement Sect. S2.3), it is also essential to thoroughly quality-check this type of data.

2.4 Surface Air Temperature

Despite the large variability in temperature, the warming trend of recent decades is strong enough to be discernible in local temperature observations, and it is larger than the warming trend simulated by state-of-the-art climate models. The principal drivers for this ‘excess warming’ appear to be changes in atmospheric circulation, mainly in winter and spring, and feedbacks involving soil moisture and cloud cover, mainly in summer and autumn (Van Oldenborgh et al. 2009).

The data sources for near-surface air temperature are different over land and sea. Terrestrial measurements are made at fixed locations, with typically standardised installations (WMO 2010) and at a reference height of 2 m (e.g. Klein Tank et al. 2002). In contrast, marine air temperature observations are typically made aboard moving ships (ICOADS; Woodruff et al. 2011), adjusted to a common reference height of 10 m (necessary because the typical observation height has increased by about 20 m over the period of record; Kent et al. 2013). Only a few fixed station measurements exist, such as on oil platforms. Observations of marine air temperature from ships are affected by daytime heating biases, and to avoid these problems datasets (for example from the Hadley Centre) are constructed using night-time observations only. Alternatively, both day and night observations, with adjustments for daytime heating following Berry et al. (2004), can be used. The North Sea region is relatively well sampled, but observations are sparse in the 19th Century and, more recently, during the Second World War.

2.4.1 Terrestrial Surface Air Temperature

The first decade of the 21st century was characterised by some extreme seasons. The hot summer of 2003 was probably unprecedented for at least 500 years in western Europe (Luterbacher et al. 2004), but was even surpassed in extremity by the East European summer of 2010 (Barriopedro et al. 2011). Summer and autumn 2006 and winter 2006/2007 were also exceptionally warm (Luterbacher et al. 2007; Cattiaux et al. 2009). On the other hand, winter 2010/2011 had a very negative NAO index (see Sect. 2.2), but was much warmer than comparable winters with a similarly negative NAO index (Cattiaux et al. 2010). Figure 2.7 shows time series of annually averaged land air temperature for the North Sea region, defined here as the area between 48°N and 62°N and 6°W and 10°E, for various data sets. The graphic shows 2014 to be unprecedentedly warm, even though none of the four seasons was the warmest on record (winter ranks 2nd after 2006/2007, spring 3rd after 2007 and 2011, summer 14th and autumn 2nd after 2006), and the previous maximum from 2011 was exceeded

by almost 0.5°C. The datasets used are the CRUTEM4v (from UEA/CRU; Jones et al. 2012), GHCN-M version 3 (NOAA/NCDC, Peterson and Vose 1997; Jones and Moberg 2003), GISTEMP (NASA/GISS; Hansen et al. 2010) and BerkeleyEarth (<http://berkeleyearth.org/>), which have been subject to a homogeneity adjustment, supplemented by the E-OBS daily dataset version 10.0 (Haylock et al. 2008). To compare the different datasets, the grids of the global datasets are regridded to match that of the E-OBS grid (0.5° × 0.5°; van der Schrier et al. 2013).

The similarity between these estimates of temperature over the North Sea region is evident, with only minor differences in trend values (Table 2.1). Over the period 1980–2010, the trend in annual averaged daily mean temperature is approximately 0.38 °C decade⁻¹. Trend values are based on a linear least-square approximation to the data. Table 2.1 also gives temperature change for the whole of Europe (30°–75°N, 12°W–45°E plus Iceland, based on E-OBS), the northern hemisphere land and the global land area, both based on CRUTEM4. For 1980–2010, the warming trend in the North Sea region is smaller than that of Europe as a whole, but larger than the average over the northern hemisphere and global land areas.

In all datasets the period from the early 1990s onwards is warmest. Figure 2.8 highlights annual temperatures of the past few decades, averaged over the North Sea region and relative to the 1961–1990 climatology, based on the E-OBS dataset. The grey bars in Fig. 2.8 indicate the estimated uncertainties which take into account both errors introduced by spatial interpolation over areas without observations, by inhomogeneities in the temperature data that result from station relocations or instrument changes etc., and by urbanisation, as documented by van der Schrier et al. (2013) and Chrysanthou et al. (2014). The uncertainties indicate that although it is not possible to be 100 % certain about the ranking of individual years, the positive overall trend since the 1980s is very pronounced and 2014 stands out, even taking the uncertainties into account.

Ionita et al. (2012b) examined the connection between diurnal temperature range (DTR) and atmospheric circulation. They found that modes of interannual winter DTR variability are strongly related to the NAO and, to a lesser extent, the AMO, whereas in summer DTR variability is mainly influenced by a blocking pattern over Europe.

2.4.2 Number of Frost Days and Summer Days

According to Della-Marta et al. (2007), the length of western European heat waves has doubled since 1880 and Europe’s climate has seen more warm extremes. This is illustrated in Fig. 2.9 which shows the difference in the annual number of frost days (minimum temperature <0 °C) and summer days

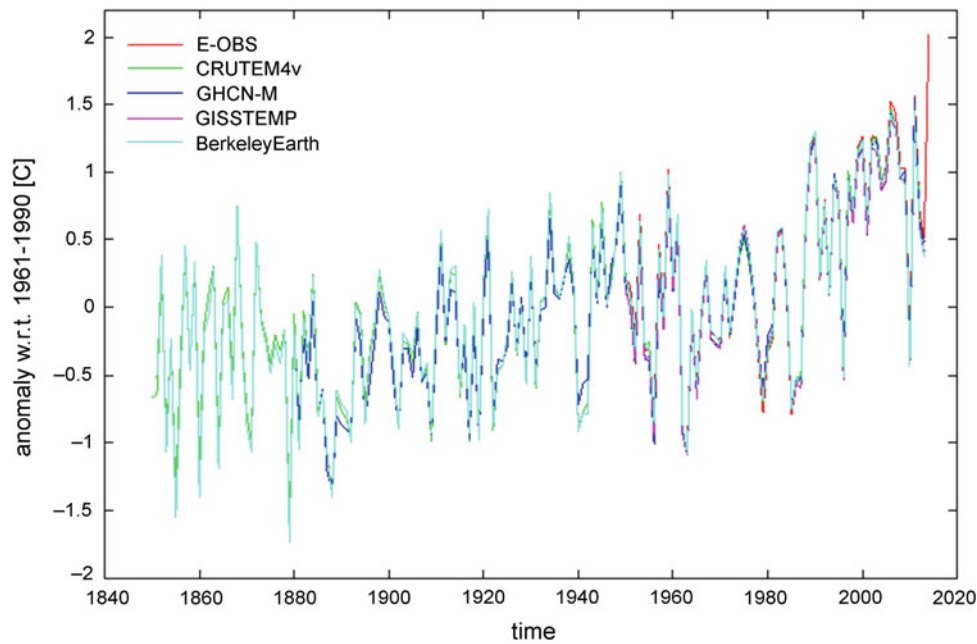


Fig. 2.7 Land-based annual mean air temperatures averaged over the North Sea region (48°–62°N, 6°W–10°E) with respect to the 1961–1990 climatology as calculated by E-OBS (red), CRUTEM4v (green), GHCN-M (blue), GISTEMP (purple) and the Berkeley Earth dataset (light blue)

Table 2.1 Linear temperature trends ($^{\circ}\text{C decade}^{-1}$) over 1950–2010 and 1980–2010 for the North Sea region for CRUTEM4v, GHCN-D, GISTEMP, BerkeleyEarth and E-OBS

	CRUTEM4v	GHCN-D	GISTEMP	Berkeley Earth	E-OBS	Europe	NH land	Global land
1950–2010	0.210	0.174	0.228	0.157	0.204	0.179	0.199	0.172
1980–2010	0.383	0.389	0.389	0.353	0.408	0.414	0.337	0.267

The last three columns give trends for Europe (based on E-OBS), the northern hemisphere land and global land temperatures (based on CRUTEM4), respectively. Numbers in bold indicate that the trend is statistically significant at the 5 % level based on a t-test accounting for the reduced degrees of freedom due to autocorrelation (von Storch and Zwiers 1999)

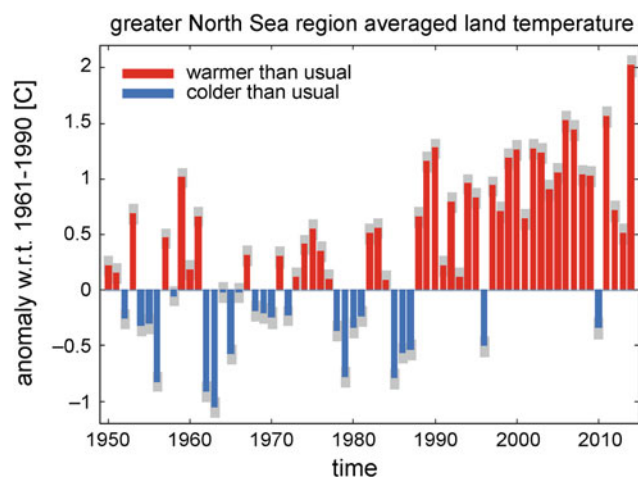


Fig. 2.8 Annual averages for land-based air temperature over the North Sea region with respect to the 1961–1990 climatology as calculated by the E-OBS dataset. The uncertainty estimate for the E-OBS data is included as grey boxes

(maximum temperature $\geq 25^{\circ}\text{C}$) between 1981–2010 and 1951–1980 (based on E-OBS data). The change in these indices is not spatially consistent (in contrast to the increase in annual averaged temperature—not shown). All differences are statistically significant at the 5 % level using a one-sided Student t-test. The figure shows that the number of frost days has declined almost everywhere, with the strongest decreases found in the northern and eastern parts of the domain. The number of summer days has also increased almost everywhere, with the smallest increases in Scotland, northern England and Scandinavia and the largest in northern France.

2.4.3 Night Marine Air Temperature

As for land-based temperatures, the night marine air temperature (NMAT) also increased over the period 1856–2010 (Fig. 2.10). Two datasets were used, an uninterpolated 5° monthly mean dataset for 1880–2010 (HadNMAT2; Kent et al. 2013) and an interpolated (using a large-scale

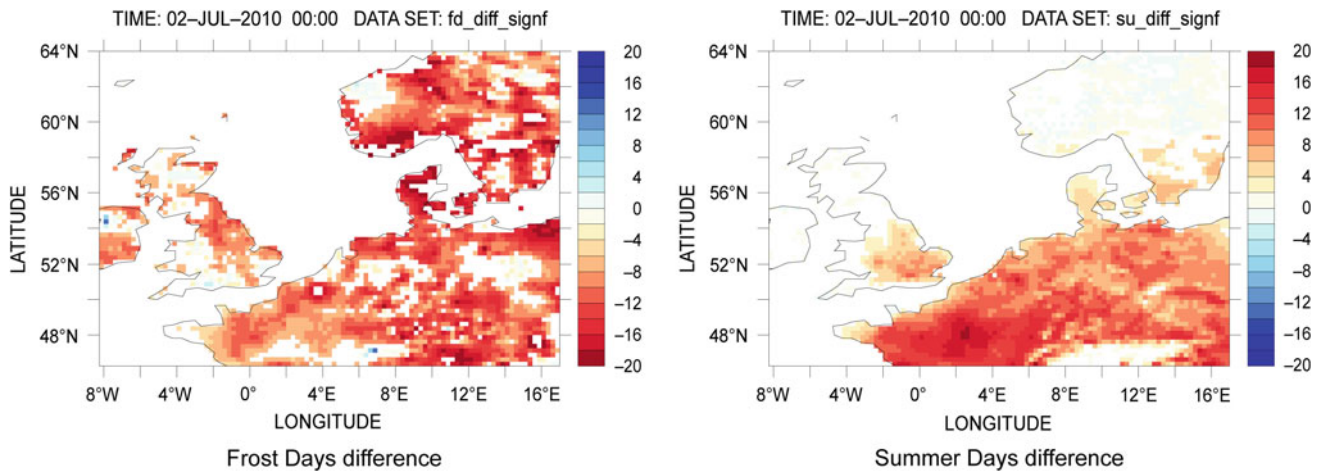


Fig. 2.9 Difference between the 1981–2010 and 1951–1980 climatological values of the annual number of frost days (*left*, daily minimum temperature $<0^{\circ}\text{C}$) and summer days (*right*, daily maximum

temperature $\geq 25^{\circ}\text{C}$). Grid squares with missing data or where the difference did not pass the 95 % significance level using a Student t-test, are white. Calculations based on E-OBS data

reconstruction technique; Rayner et al. 2003) 5° monthly mean dataset for 1856–2001. Differences between these datasets are larger than for land temperatures, especially around 1900 and during and just after the Second World War. In the latter period, sampling is sparse and non-standard observing practices necessitated adjustments to the observations (Kent et al. 2013). After about 1950, agreement improves. Linear trends in air temperature, adjusted for day-time heating biases (Berry and Kent 2009) show similar values.

Figure 2.10 also indicates that for marine air temperature the values in the most recent decade are likely to be the warmest on record, although uncertainty is large in the early part of the record due to sparse sampling.

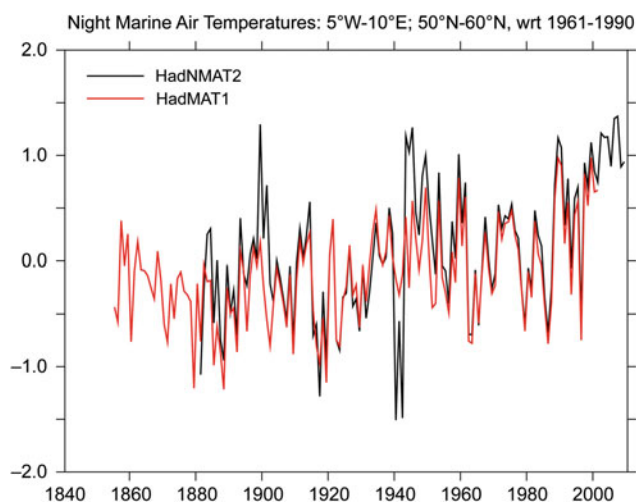


Fig. 2.10 Annual average night marine air temperature anomalies ($^{\circ}\text{C}$) for the region 50° – 60°N , 5°W – 10°E from 5° monthly mean datasets: HadNMAT2 (*black*) and HadMAT1 (*red*)

Seasonal time series of the marine air temperature data sets show broadly similar variability to land-based temperatures (Fig. 2.11), but with a smaller amplitude. Very recently, the differences again increase, but this seems to be due to sparse observations and changes in the marine observing system (Kent et al. 2007, 2013).

2.4.4 Comparison of Land and Marine Air Temperatures

A comparison of land and marine temperatures (Fig. 2.12) shows general agreement. The lower plot in each panel depicts three estimates of the land-marine air temperature difference over the North Sea region based on E-OBS data for the land component and three different marine air temperature datasets: the NOCv2.0 dataset (Berry and Kent 2009, 2011), the HadNMAT2 dataset and the HadMAT1 dataset. Due to the much larger heat capacity of water, the difference series between the land and marine air temperature shows a residual positive trend over the last few decades of the record.

2.4.5 Summary

There is generally good agreement between the different temperature data sets over the oceans and over land. While temperatures have clearly increased over land, the NMAT shows there has also been an increase over the North Sea, even though the variability on seasonal timescales is smaller than for the land temperatures. Furthermore, due to the large heat capacity of water it takes much longer to warm the ocean than the land. In addition, heat is transported away

Fig. 2.11 Seasonal mean night marine air temperature (°C) from NOCv2.0 (1970–2010), HadNMAT2 (1950–2010) and HadMAT1 (1950–2001). HadNMAT2 and HadMAT1 were averaged to the same 1° grid as NOCv2.0 and masked to the NOCv2.0 land mask

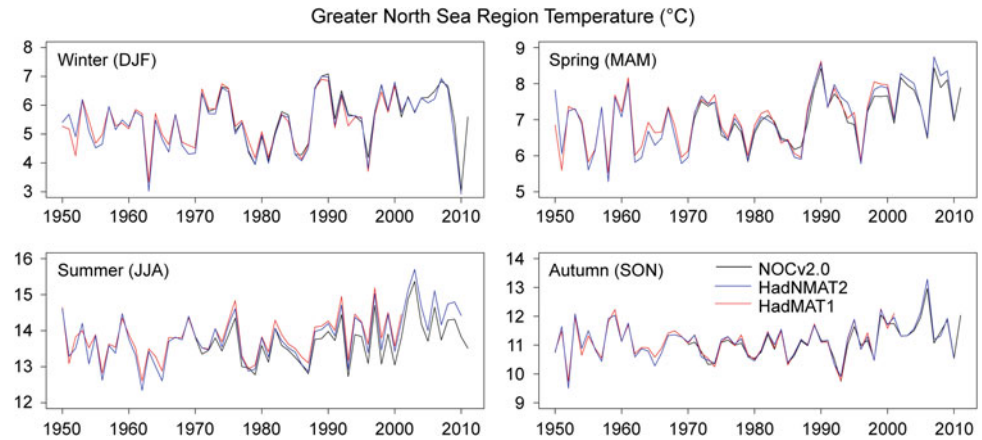
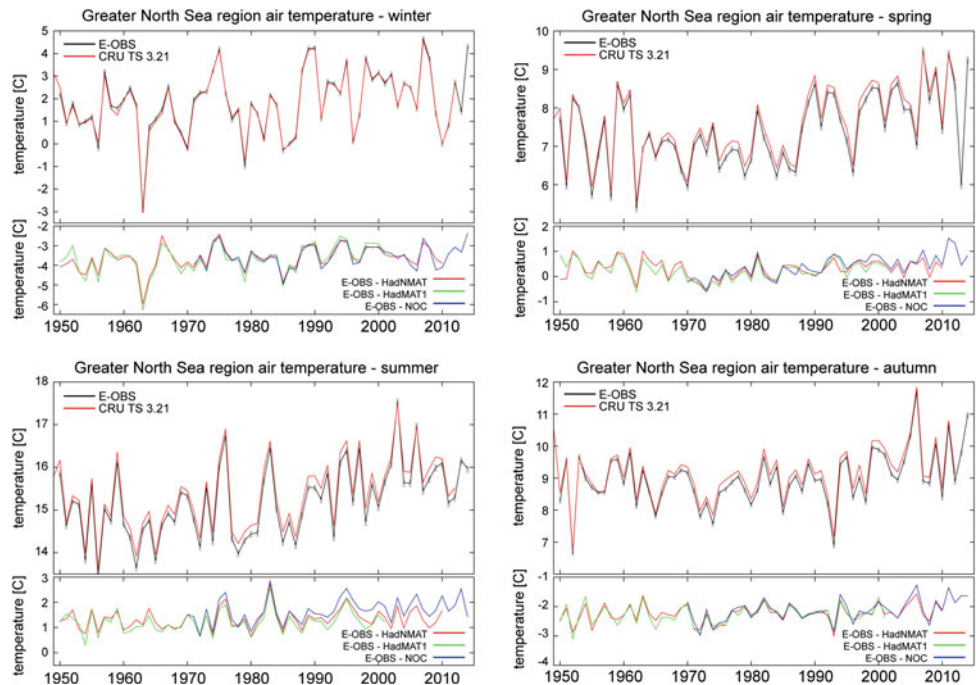


Fig. 2.12 Land-based air temperature over the North Sea region for winter (*upper left*), spring (*upper right*), summer (*lower left*) and autumn (*lower right*) based on the CRU TS 3.10 (Harris et al. 2014; *red*) and E-OBS (*black*) datasets. The uncertainty estimate for the E-OBS data is included as *grey boxes*. The *lower plots* show the difference between land-based air temperature and marine air temperature for this region, for three marine air temperature datasets



from the surface into deeper waters, where it cannot be directly measured. Thus, it would be expected that the land warms faster than the sea as long as the radiative forcing is positive. This is exactly the situation being observed, and according to the datasets, the imbalance is up to several tenths of a degree.

2.5 Precipitation

2.5.1 Precipitation Over Land in the North Sea Region

In a warmer climate, the atmospheric water vapour content is likely to rise due to the increase in saturation water vapour

pressure with air temperature, as described by the Clausius-Clapeyron relation, and to result in an intensification of rainfall (Held and Soden 2006; O’Gorman and Schneider 2009). Evidence of higher amounts and more extreme precipitation has already been reported (e.g. Groisman et al. 2005; Moberg et al. 2006; Donat et al. 2013; Hartmann et al. 2013). Even though floods are a recurring event in Europe, attempts have been made to link increased flood risk to changes in the frequency of atmospheric blocking events (Lavers et al. 2012) or to anthropogenic climate change (Pall et al. 2011).

In a global study, Donat et al. (2013) showed a weak increase in the number of days exceeding 10 mm of precipitation (R10 mm) over the northern parts of Europe (but statistically significant only over eastern Europe at the 96 %

level). Over the Iberian Peninsula, a non-significant decrease in this metric is observed. A non-significant increase in the contribution of extreme precipitation events to the total precipitation amount (R95pTOT) is also observed over Great Britain and Scandinavia.

The European Climate Assessment and Dataset (ECA&D, Klein Tank et al. 2002) is a collection of daily station observations of 12 elements (of which five are gridded) and contains (as of March 2014) data from nearly 8000 stations across Europe and the Mediterranean. The station time series are updated on a regular basis using data provided by the national meteorological and hydrological services (NMHSs), universities or, before updates from these institutions are available, synoptic messages from the Global Telecommunication System (GTS). ECA&D receives time series from 61 data providers for 62 countries (as of March 2014).

Figure 2.13 compares precipitation for three stations with data since the beginning of the 20th century; Cambridge (UK), Stromsfoss Sluse (Norway) and De Bilt (Netherlands). The time series show strong interannual and decadal variability. A general upward trend is visible in the Dutch and Norwegian time series with trends of $14.52 \text{ mm decade}^{-1}$ in the annual data over the 1901–2014 period for De Bilt and $28.81 \text{ mm decade}^{-1}$ over the 1901–1950 period for Stromsfoss Sluse. Both trends are (just) statistically significant at the 5 % level following a t-test accounting for autocorrelation in the time series (von Storch and Zwiers 1999). A long-term trend in the UK time series is less pronounced. The Norwegian time series exhibits an enhanced trend since the mid-1990s, especially in summer. A weak drying trend since the 1990s, although not unprecedented, is

visible in the UK series. It is also clear that, under certain circumstances, the entire area is influenced by high pressure for extended periods (e.g. in 1921) such that the whole North Sea area remains very dry.

Trends in annual land precipitation are positive almost everywhere over the North Sea region for the period 1951–2012 (Fig. 2.14, top panel). The greatest increase in precipitation is observed in winter (Fig. 2.14, lower left), especially along the west coast of Norway, over southern Sweden, parts of Scotland and the Netherlands and Belgium. Further inland, trends are much smaller and statistically non-significant almost everywhere. In summer, there is no evidence of increasing precipitation trends along the coast of western Norway, while the contrast between trends in coastal regions and more inland regions of the European mainland increases considerably as the latter show negative trends in summer (Fig. 2.14, lower right).

Winter and spring in northern Europe (defined as the land area north of 48°N) show an overall decreasing trend in return periods of extreme precipitation (van den Besselaar et al. 2013), which is indicative of increasing precipitation extremes. The trend is most pronounced in the 5-day precipitation amount in northern Europe during spring. The 5-day amount which is statistically a 20-year event over the 1951–1970 period becomes an approximately 8-year event in the 1991–2010 period.

For annual 5-day and 10-day precipitation amounts in the UK, Fowler and Kilsby (2003) found significant decadal-level changes in many regions. For the 10-day precipitation amount, the 50-year event during 1961–1990 became an 8-, 11- and 25-year event in eastern, southern and

Fig. 2.13 Annual, winter and summer precipitation series for three stations from the ECA&D dataset; De Bilt (Netherlands, top left), Stromsfoss Sluse (Norway, top right) and Cambridge (UK, lower left). A low-pass filter is applied for the black curves

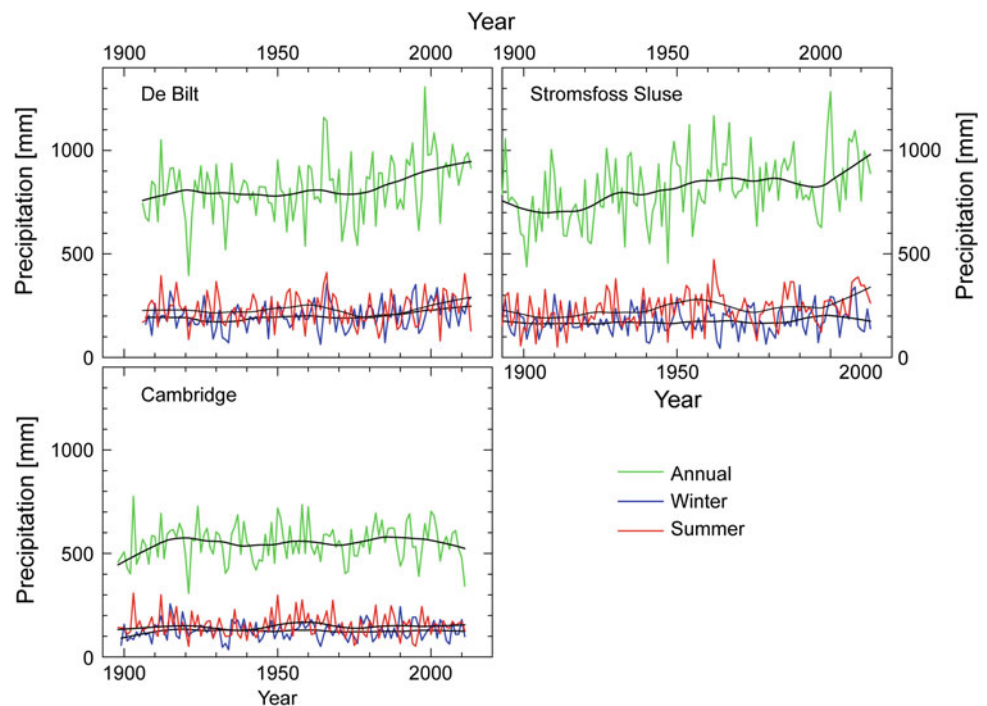
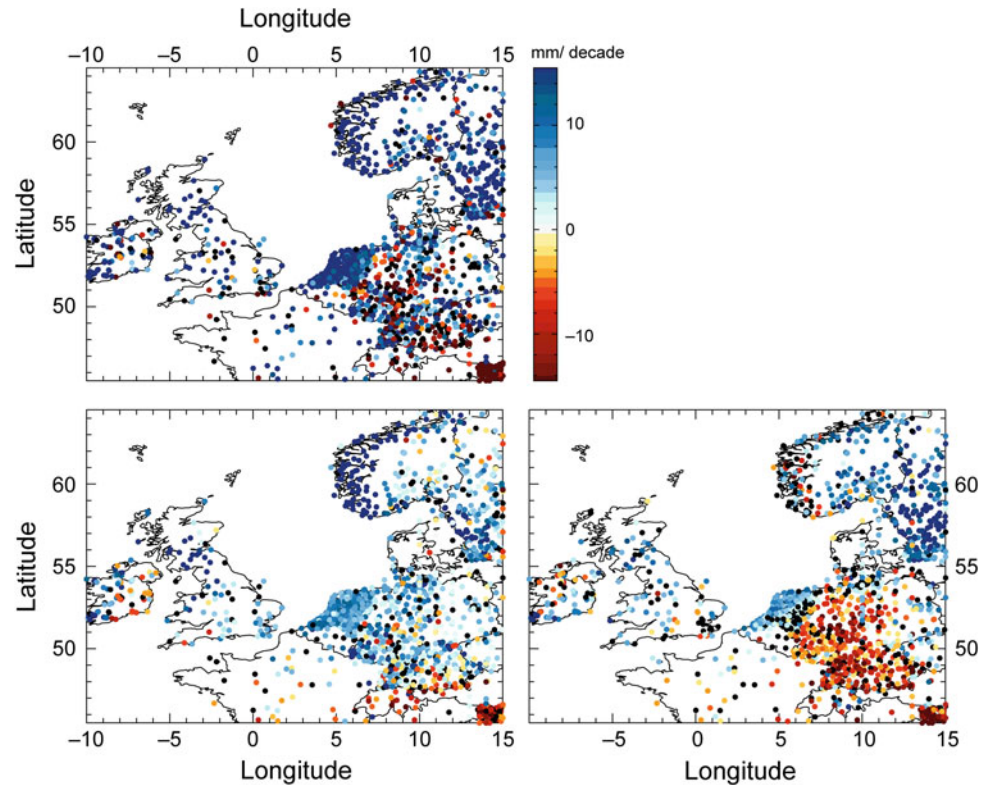


Fig. 2.14 Linear least-squares fit trends in annual (*top*), winter (*lower left*) and summer (*lower right*) precipitation over the period 1951–2012 in mm decade⁻¹. *Blue circles* denote a trend towards wetter conditions, while *orange* and *red circles* denote a trend towards drier conditions, both significant at the 5 % level ($p \leq 0.05$). *Black circles* fail to be significant at the 25 % level ($p \leq 0.25$) and are added to the figure to illustrate areas without any significant trend. *Source* ECA&D



northern Scotland, respectively, during the 1990s. In northern England the average return period has also halved.

Cortesi et al. (2012) analysed the precipitation concentration index, which is a measure of the amount of precipitation on a day with precipitation. The north-western coast of Europe shows relatively low values for this index (i.e. evenly distributed precipitation) compared to more Mediterranean climate types. No clear spatial pattern was detected in the trends in the index.

Groisman et al. (2005) studied total precipitation and frequency of intense precipitation in several regions of the world, including Fennoscandia. They found a significant increase in the annual totals and in the frequency of very heavy annual and summer precipitation events, where ‘very heavy’ precipitation events are defined by counting the upper 0.3 % of daily rainfall events (relating to a daily event that occurs once every 3–5 years).

There is temporal variability in trends when studying precipitation indices of extremes. For example, the trend in precipitation fraction due to very wet days, related to the 95th percentile in daily sums (R95pTOT), shows a different picture when the trends are determined over the period 1951–1978 compared to 1979–2012 (Fig. 2.15). Along the coasts of south-eastern England and the Netherlands, there is no trend apparent for the period 1951–1978, while the period 1979–2012 has an increasing trend for several stations in these areas.

Care should be taken if the precipitation fraction exceeding the 95th percentile (R95pTOT) is determined over a climatological period of several decades, since extremes may have increased disproportionately and thus the shape of the distribution may have changed. For example, an index S95pTOT, using the Weibull shape parameter instead of an explicit estimate of the 95th percentile, can be used (Leander et al. 2014). Northern Europe shows a (significant) increase in R95pTOT, but this is far less pronounced for S95pTOT. Since R95pTOT cannot distinguish between a shift in the median of the probability distribution for precipitation and a change in only the tail of the distribution, trends are generally ‘more negative’ for S95pTOT, especially over southern Scandinavia, the Netherlands, Germany and the UK.

Zolina et al. (2009) introduced a new index for R95pTOT, making use of a gamma distribution for wet day precipitation amounts and the associated theoretical distribution of the fractional contribution of the wettest days to the seasonal or annual total. The trend results for their new index are similar to R95pTOT.

Another way of analysing changes in precipitation is by counting the number of wet days. An example of this is the index CWD (maximum number of consecutive wet days, here defined as the number of days with precipitation ≥ 1 mm). Trends in the station records for the period 1951–2012 are shown in Fig. 2.16, which indicates that most of the stations in the North Sea region show a slight increasing

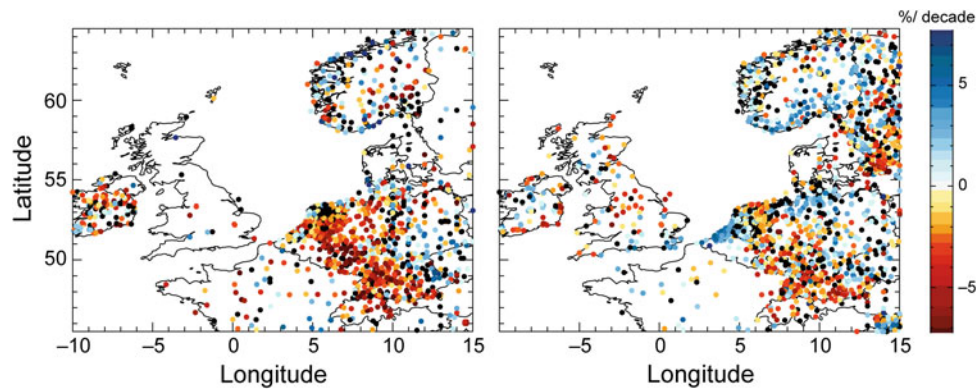


Fig. 2.15 Linear trends in the precipitation fraction due to very wet days (R95pTOT) in winter over the periods 1951–1978 (*left*) and 1979–2012 (*right*). *Blue circles* denote a trend towards wetter conditions, while *orange* and *red circles* denote a trend towards drier conditions,

both significant at the 5 % level ($p \leq 0.05$). *Black circles* fail to be significant at the 25 % level ($p \leq 0.25$) and are added to the figure to illustrate areas without any significant trend. *Source* ECA&D

trend in the annual number of consecutive wet days. A similar map for trends in the maximum number of consecutive dry days (CDD) does not indicate a coherent change in the region. The majority of stations show trend values that do not meet even the 25 % significance level.

An example of interaction between North Sea waters and coastal climate was documented by Lenderink et al. (2009) for a month with extreme precipitation in the coastal region of the Netherlands (August 2006), where precipitation amounts were four times higher than the climatological average. Preceded by an extremely warm July (see Sect. 2.4) with very high sea surface temperatures in the North Sea at the end of July, favourable atmospheric flow conditions transported large amounts of moisture onto land, producing excessive rainfall in an area less than 50 km from the

coastline. This phenomenon seems to be a robust finding since the positive trend in the difference between coastal and inland precipitation observed in the Netherlands is not sensitive to the period analysed.

2.5.2 Precipitation Over the North Sea

Only limited information is available for precipitation over oceans in general and the North Sea in particular. Almost no in situ measurements exist, which means it is necessary to rely on satellite observations using passive microwave detectors. HOAPS (Hamburg Ocean-Atmosphere Parameters and Fluxes) is one such dataset (Andersson et al. 2010, 2011). This covers the period 1988–2008 and is the only generally available satellite-based dataset for which fields of precipitation and evaporation over the oceans are consistently derived (Andersson et al. 2011). Over land, the dataset is gauge-based. Figure 2.17 shows the geographical distribution of annual average precipitation for the North Sea region (Fennig et al. 2012). Over most of the North Sea, the mean annual precipitation is between 600 and 800 mm, although values below 600 mm are also found off the east coast of England. Most coastal regions receive more than 800 mm, and in some mountainous regions (Scotland, Norway) more than 2000 mm are observed. While land stations in the south of the region have most rain in winter with a weak secondary summer maximum, further north and generally over the sea, there is only a maximum in winter, and May and June are the driest months.

Table 2.2 shows annual precipitation totals over the central North Sea region (54°–58°N, 1.5°–5.5°E) from the few available datasets.

There are considerable differences between the various estimates of precipitation, even in reasonably data-rich regions like the North Sea. There are also large differences

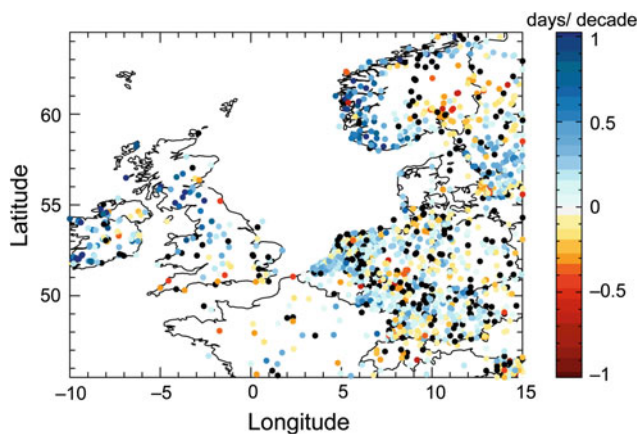


Fig. 2.16 Linear trend in annual maximum number of consecutive wet days (CWD) over the period 1951–2012. *Blue circles* denote a trend towards wetter conditions, while *orange* and *red circles* denote a trend towards drier conditions, both significant at the 5 % level ($p \leq 0.05$). *Black circles* fail to be significant at the 25 % level ($p \leq 0.25$) and are added to the figure to illustrate areas without any significant trend. *Source* ECA&D

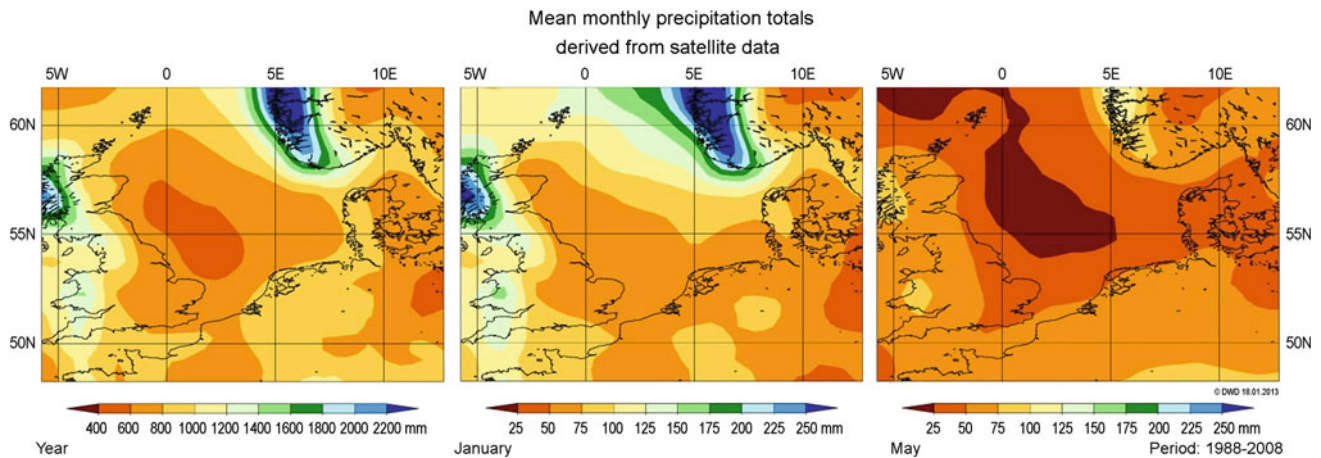


Fig. 2.17 Precipitation over the North Sea area. Ocean: HOAPS dataset (Fennig et al. 2012), land: gauge-based. Annual sum (*left*), monthly sum for January (*centre*), and monthly sum for May (*right*) for the period 1988–2008. All data in mm

Table 2.2 Estimates of annual average precipitation over the North Sea region from different reanalyses and satellite-based datasets

Dataset	1979–2001	1988–2008	Source
HOAPS	–	643	Andersson et al. (2010)
ERA-Interim	812	800	Simmons et al. (2010) and Berrisford et al. (2009)
ERA40	691		Uppala et al. (2005)
Coastdat2 (cDII.00)	853	861	Geyer (2014)
NCEP-CFSR	966	1000	Saha et al. (2010)
MERRA	754	772	Rienecker et al. (2011)

All units in mm

between periods, highlighting the problems in deriving trends in precipitation (Bengtsson et al. 2004). Precipitation in reanalyses depends on the moisture flux divergence, a rather weakly constrained quantity, which in turn does not depend on direct observations, but on the assimilation of satellite radiances (see e.g. Lorenz and Kunstmann 2012).

2.5.3 Summary

An assessment of temporal variability shows that precipitation over land and, but somewhat weaker, over sea is positively correlated with the NAO. Winters with strong positive NAO anomalies show distinct peaks in precipitation and amounts up to twice the average over the North Sea region (Andersson et al. 2010). On longer time scales, drought conditions over central Europe are also connected to the AMO (Atlantic Multidecadal Oscillation; Ionita et al. 2012a). Generally speaking, precipitation is more variable than temperature, and agreement between datasets is less. Nevertheless, there are indications of an increase in precipitation to the north of the North Sea region and a decrease to the south, in agreement with the projected north-eastward shift in the storm tracks. In many regions, there are also

indications that extreme precipitation events have become more extreme and that return periods have decreased.

2.6 Radiative Properties

Meteorological observations aboard ships usually do not include measurements of sunshine duration and radiation. As a result there are few data available, and these are mainly from isolated field campaigns on research vessels. In contrast, cloud parameters are often observed routinely, although the quality of observations varies widely. The following discussion of clouds, solar radiation and sunshine duration therefore relies mainly on studies concerning a wider area, but these data should also be valid for the North Sea region.

2.6.1 Clouds

Clouds have a significant impact on the Earth's radiation budget. They affect incoming solar shortwave (SW) radiation (by reflecting this back to space) as well as outgoing thermal longwave (LW) radiation (by reducing its emission

to space). The difference between the actual radiative flux and that under clear sky conditions is referred to as cloud radiative forcing (CRF). The largest contribution to LW CRF is made by high clouds, whereas the largest contribution to SW CRF is from optically thick clouds due to their higher albedo compared to the clear sky surface albedo. Thus, variations in cloudiness are of great interest in relation to rising global temperatures. The Extended Edited Cloud Report Archive (EECRA; Warren et al. 1986, 1988, 2006) consists of quality controlled climatologies of total cloud cover and cloud type amounts over land and ocean, respectively, based on surface synoptic cloud observations.

Few analyses of changes in cloud cover exist, and even less for the North Sea region. A decrease has been observed in global high cloud cover over almost all land regions since 1971 and most ocean regions since 1952. Norris (2008) analysed global mean time series from gridded surface observations of low-, mid- and upper-level clouds as well as total cloud cover and satellite cloud observations over land and ocean based on EECRA, other surface synoptic cloud reports from land since 1971, the ship-based ICOADS which includes observations since 1952, and satellite observations available from July 1983 in the International Satellite Cloud Climatology Project (ISCCP). Norris (2008) found inconsistencies for the overlapping period of in situ and satellite data except for high clouds.

Over Europe, variability in total winter cloud cover is strongly connected to the NAO. Because the NAO was undergoing a positive trend during a study by Warren et al. (2006), there is a strong positive trend in total cloud cover over Norway at this time (Fig. 2.18). No clear trend is visible further south in the North Sea region (Thompson et al. 2000; Hense and Glowienka-Hense 2008).

Warren et al. (2006) also analysed cloud types observed at European land stations in relation to the NAO/AO signal

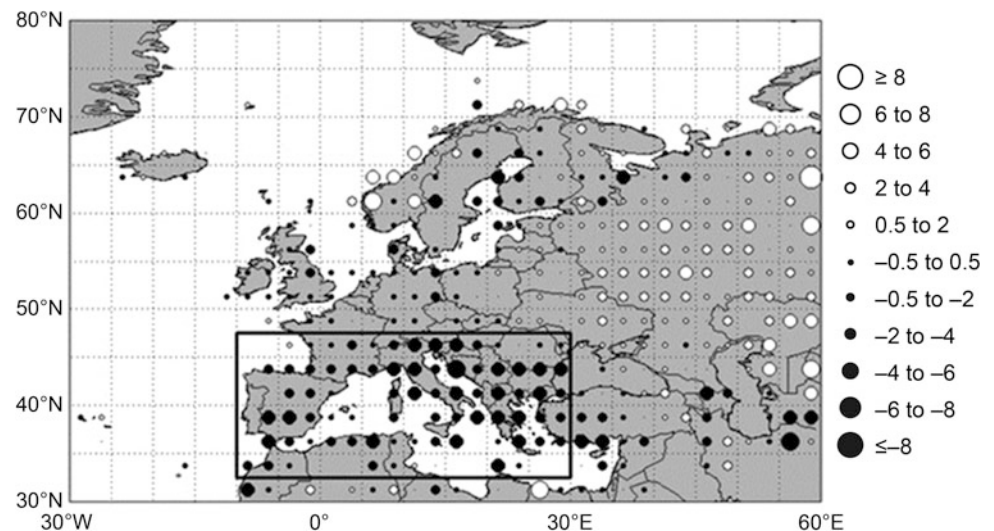
in winter for the period 1971–1996. Not surprisingly, the strongest correlation was for nimbostratus (Fig. 2.19). Across the western parts of Europe correlations are negative, but high positive correlations exist in the northern part of the North Sea region from northern Scotland to Norway.

2.6.2 Solar Radiation

Time series of measured solar radiation data at various sites around the globe show decreasing irradiances on the order of $6\text{--}9\text{ W m}^{-2}$ (corresponding to a decline of 4–6 % over 30 years) after the mid-1950s ('global dimming'; Gilgen et al. 1998; Stanhill and Cohen 2001; Liepert and Tegen 2002) and mainly over land, and subsequent increases since the mid-1980s ('global brightening'; Wild et al. 2005; Norris and Wild 2007) which cannot be explained by variations in solar irradiance or cloudiness alone (Wild 2009), but are largely due to marked changes in the amount of anthropogenic aerosol particles after the Second World War (Stanhill and Cohen 2001; Liepert and Tegen 2002; Streets et al. 2006; Norris and Wild 2007). Since the 1980s, air-quality regulations have led to a decline in air pollution, as can be seen from time series of optical depth (Mishchenko et al. 2007; Ruckstuhl et al. 2008). More recent studies (e.g. Granier et al. 2011; Lee et al. 2013; Myhre et al. 2013; Shindell et al. 2013) corroborate these findings.

For Europe, Norris and Wild (2007) examined changes in SW downward radiation and total cloud cover to distinguish the effects of cloud variability from long-term aerosol influences in the period 1971–2002 (Fig. 2.20). Their 'cloud cover radiative effect' (CCRE), defined as the radiative effects of changes in cloud cover, is derived from daytime synoptic surface observations and ISCCP data, subtracted from the downward radiation obtained from the Global

Fig. 2.18 Linear trends in total cloud cover in percent per decade for $2.5^\circ \times 2.5^\circ$ boxes in Europe and North Africa in winter (DJF) for the period 1971–1996. The size of each dot indicates the magnitude of the trend (Warren et al. 2006)



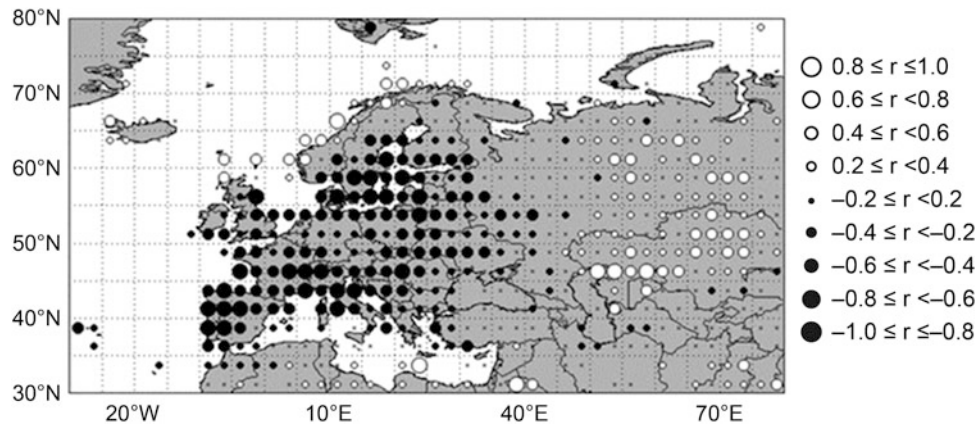


Fig. 2.19 Correlation of nimbostratus anomalies with the Arctic Oscillation index, for $2.5^\circ \times 2.5^\circ$ boxes in Europe and North Africa in winter (DJF) for the period 1971–1996. The size of each *dot* indicates the magnitude of the correlation coefficient (Warren et al. 2006)

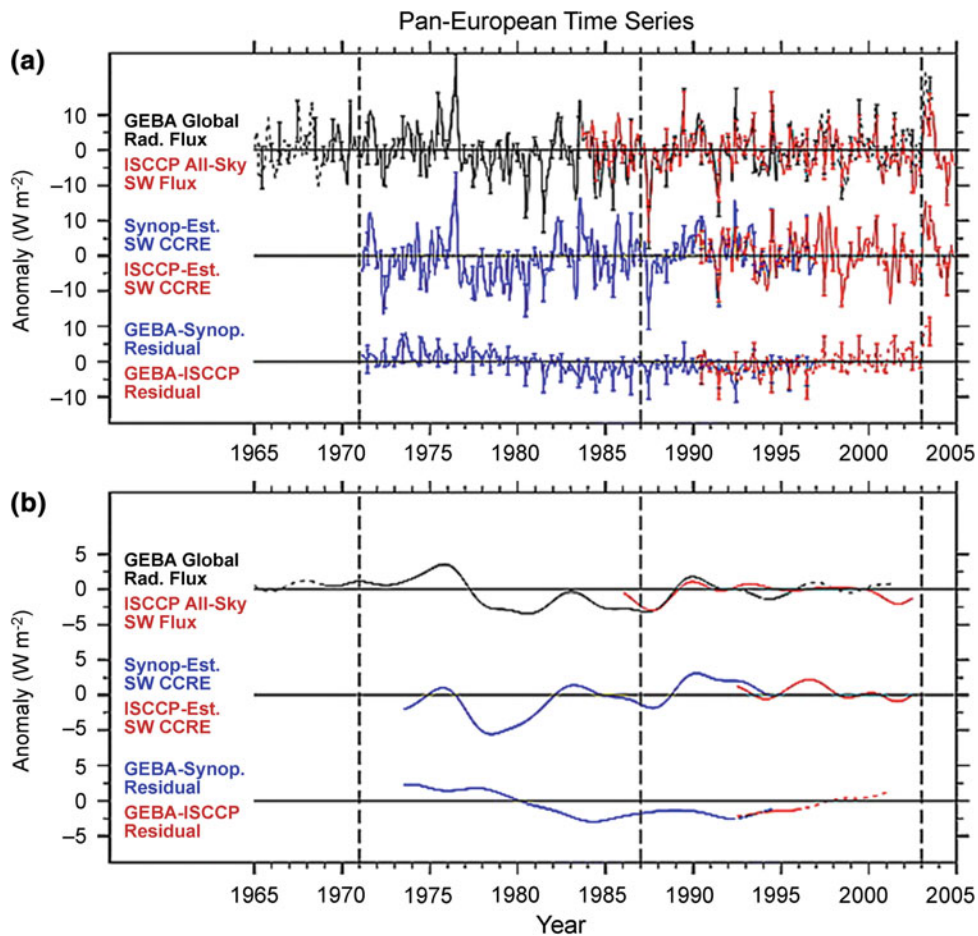


Fig. 2.20 Time series of monthly anomalies averaged over grid boxes covering most of Europe with **a** a 1-2-1 filter and **b** a 61-point 5-year Lanczos low-pass filter for GEBA global radiation flux (upper, black), ISCCP all-sky downward SW radiation flux (upper, red), SW cloud cover radiation effect (CCRE) estimated from synoptic reports of total cloud cover (middle, blue), SW CCRE estimated from ISCCP total cloud cover amount (middle, red), residual anomalies after removing synoptic-estimated SW CCRE from GEBA global radiation (lower,

blue) and residual anomalies after removing ISCCP-estimated SW CCRE from GEBA global radiation (lower, red). Dashed values indicate where less than 75 % of the grid boxes contributed to the GEBA time series. Small vertical bars denote 95 % confidence intervals for June and December anomalies, and vertical dashed lines mark the start and end times for trend calculations (Norris and Wild 2007)

Energy Balance Archive (GEBA). The resulting time series comprises variations in clear-sky solar flux as well as radiative effects of changes in cloud albedo that are not linearly correlated to the cloud cover. They found a high correlation ($r = 0.88$) between global radiation anomalies and the estimated cloud cover radiative effect on monthly and sub-decadal timescales, but the time series of differences show dimming and brightening as well as low frequency trends with minima related to the volcanic eruptions of El Chichón (Mexico) and Pinatubo (Philippines). Decreasing trends for the period 1971–1986 and then increasing trends for 1987–2002 are found for coastal areas of the North Sea region (not shown), but these are mostly not statistically significant (Ruckstuhl et al. 2008, Ruckstuhl and Norris 2009).

Aerosol particles influence the radiation budget and hence air temperature in two ways. The direct aerosol radiative effect refers to clear-sky cases, when solar radiation is directly scattered (mainly by sulphate) or absorbed (mainly

by black carbon), while the indirect aerosol effect enhances cloud albedo via an increase in the number of aerosol particles that act as condensation nuclei creating smaller droplets and prolonging cloud lifetime due to a decrease in droplet size and less precipitation loss.

For the period 1981–2005, Ruckstuhl et al. (2008) estimated the direct and indirect aerosol effects by determining the SW downward radiation for cloud-free and cloudy conditions from eight sites in northern Germany. Excluding the sunny year 2003, the net LW forcing under cloud-free skies is $0.84 \text{ W m}^{-2} \text{ decade}^{-1}$ (range: $0.49\text{--}1.20$), whereas the SW net forcing from changes in cloudiness is $0.56 \text{ W m}^{-2} \text{ decade}^{-1}$ (range: -0.91 to 2.00), resulting in a total cloud forcing of $0.16 \text{ W m}^{-2} \text{ decade}^{-1}$ (range: -0.26 to 0.57). Thus, the direct aerosol effect has a much larger impact on climate forcing than the indirect aerosol and other cloud effects.

Philipona et al. (2009) found an increase in LW downward radiation over Germany, based on observations for 1981–2005, due to rising temperature and humidity and to the increase in greenhouse gas concentration, but found no effect of changes in cloudiness (Fig. 2.21). The total net LW radiation (the difference between incoming and outgoing radiation at the surface) depends on temperature and absolute humidity, which itself is dependent on temperature.

For northern Germany, the LW forcing due to greenhouse gases including water vapour resulted in $0.95 \text{ W m}^{-2} \text{ decade}^{-1}$ (range: $0.26\text{--}1.64$), while the part due to water vapour feedback alone is $0.60 \text{ W m}^{-2} \text{ decade}^{-1}$ (range: $0.16\text{--}1.04$) (Philipona et al. 2009). Thus, the total SW forcing is three times larger than the LW forcing from rising atmospheric levels of anthropogenic greenhouse gases.

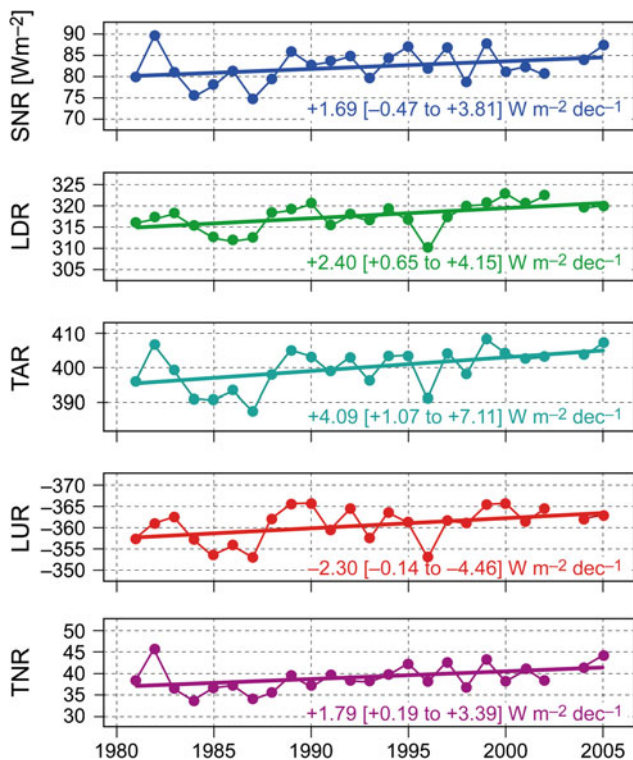


Fig. 2.21 Radiation budget and surface forcing. Annual mean values (W m^{-2}) for the individual components of the surface radiation budget for eight stations in northern Germany: SW net radiation (SNR), LW downward radiation (LDR), total absorbed radiation (TAR), LW upward radiation (LUR) and total net radiation (TNR) from 1981 to 2005 (missing 2003 data). Downward fluxes are positive and upward fluxes negative. Trends in $\text{W m}^{-2} \text{ decade}^{-1}$ with the 95 % confidence interval in brackets. TNR, the balance between downward and upward fluxes at the surface representing the energy available for sensible and latent energy fluxes increases primarily due to the increase of water vapour in the atmosphere (Philipona et al. 2009)

2.6.3 Sunshine Duration

Operational measurements of sunshine duration started at most weather stations in the 1930s or 1940s, mainly using the Campbell-Stokes heliograph. However, over recent decades, new electronic-optical equipment has increasingly been used, causing data quality issues and thus consistency problems within the time series of sunshine duration data (Augter 2013). As the existing sunshine duration database for the open sea is insufficient for climate analyses, the results presented in this chapter are based on coastal or island stations only. Sunshine duration depends on three factors: daylength (which is a function of latitude and season), amount of daytime clouds, and atmospheric opacity. Cloudiness and opacity are influenced by meteorological conditions and the latter also by aerosol concentration, which can be very different over land and sea. For Germany, Schönwiese and Janoschitz (2005) analysed changes in sunshine duration for the periods 1951–2000 and 1971–

2000 and found no obvious trends. It is not clear whether the increase in global radiation is related to an increase in sunshine duration.

2.6.4 Summary

From the few available datasets on radiative properties, it may be concluded that there are non-negligible trends together with potential uncertainties and land-sea inhomogeneities which make it difficult to assess these quantities in detail.

2.7 Summary and Open Questions

It is not obvious how atmospheric circulation has changed in the North Sea region over the last roughly 200 years. Further research is therefore necessary to understand climate change versus climate variability. One open research question is the extent to which circulation over the North Sea region is controlled by distant factors. In particular, whether there is a link between changes in the Arctic cryosphere and atmospheric circulation further south, including over the North Sea region. Overland and Wang (2010) highlighted a connection between the recent decrease in Arctic sea ice and cold winters in several areas of Europe. With the ongoing decline in sea ice in the Arctic, any such effect on circulation patterns would be important for climate in the North Sea region. Rahmstorf et al. (2015) proposed a proxy-based connection between the observed cooling in the North Atlantic south of Greenland and a weakening of the Atlantic Meridional Overturning Circulation (AMOC) partly due to increased melting of the Greenland Ice Sheet and subsequent freshening of the surface waters. Changes in the strength of the AMOC, however, are still debated and Zhang (2008) and more recently, Tett et al. (2014) have stated that its strength has actually increased.

Owing to large internal variability, it is unclear which part of the observed atmospheric changes is due to anthropogenic activities and which is internally forced. Slowly varying natural factors with an effect on European climate, such as the AMO (Petoukhov and Semenov 2010), may superimpose long-term trends and therefore be difficult to distinguish from the anthropogenic climate change signal.

There are signs of an increase in the number of deep cyclones (but not in the total number of cyclones). There are also indications that the persistence of circulation types has increased over the last century or so (Della-Marta et al. 2007). It is an open question whether this is also related to the decline in Arctic sea ice.

Another open question is whether there have been changes in extreme weather events. However, most studies

rely on small datasets covering relatively short time periods, which makes it difficult to draw statistically significant conclusions. As short time series and a lack of homogeneous data make it impossible to obtain reliable trend estimates, it is important to make available and homogenise the large number of data from past decades that have not yet been digitised. However, shown by the case of the erroneous pressure digitisations in the WASA dataset (see E-Supplement Sect. S2.3), it is essential for data to be thoroughly quality-checked. Experience from the WASA data suggests that this step requires human expertise and cannot be fully automated. On the other hand, further reanalyses, which may be considered a ‘best-possible’ time-space interpolator for observed data, can be useful as long as any bias that is potentially introduced through new instruments, station relocations etc. is properly addressed. The same is true for existing reanalyses, as it is unclear how homogeneous reanalyses can be that rely only on surface observations such as 20CR (Compo et al. 2011).

Temperature has increased in the North Sea region, and there is a clear signal in the annual number of frost days or summer days. While there is a clear winter and spring warming signal over the Baltic Sea region (Rutgersson et al. 2014), this is not as clear for the North Sea region. For precipitation, it is difficult to deduce long-term trends; however, there are indications of longer precipitation periods and ‘more extreme’ extreme events.

Other quantities, such as clouds, radiation or sunshine duration, are difficult to judge owing to a general lack of data.

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