



1 **Implications on atmospheric dynamics and the effect on black**
2 **carbon transport into the Eurasian Arctic based on the choice of**
3 **land surface model schemes and reanalysis data in model**
4 **simulations with WRF.**

5 Carolina Cavazos Guerra¹, Axel Lauer², Andreas B. Herber³, Tim M. Butler¹, Annette Rinke⁴,
6 Klaus Dethloff⁴

7
8 ¹. Institute for Advanced Sustainability Studies (IASS), Potsdam, Germany.

9 ². Deutsches Zentrum für Luft- und Raumfahrt (DLR), Institut für Physik der Atmosphäre, Oberpfaffenhofen,
10 Germany.

11 ³. Alfred Wegener Institute, Helmholtz Centre for Polar and Marine Research, Research Unit Bremerhaven, Germany.

12 ⁴. Alfred Wegener Institute, Helmholtz Centre for Polar and Marine Research, Research Unit Potsdam, Germany.

13 *Correspondence to:* Carolina Cavazos-Guerra (carolina.cavazosguerra@iass-potsdam.de)

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16 **Abstract**

17 A realistic simulation of physical and dynamical processes in the Arctic atmosphere and its feedbacks with the surface
18 conditions is still a challenge for state-of-the-art Arctic climate models. This is of critical importance because studies of,
19 for example, transport of pollutants from middle latitudes into the Arctic rely on the skill of the model in correctly
20 representing atmospheric circulation including the key mechanisms and pathways of pollutant transport. In this work the
21 performance of the Weather Research and Forecast model (WRF) with two land surface model schemes (Noah and
22 NoahMP) and two reanalysis data sets for creation of lateral boundary conditions (ERA-interim and ASR) is evaluated
23 focusing on meteorological surface properties and atmospheric dynamics. This includes the position and displacement of
24 the polar dome and other features characterizing atmospheric circulation associated to sea ice maxima/minima extent
25 within the Eurasian Arctic. The model simulations analyzed are carried out at 15-km horizontal resolution over a period
26 of five years (2008 to 2012). The WRF model simulations are evaluated against surface meteorological data from
27 automated weather stations and vertical profiles from radiosondes. Results show that the model is able to reproduce the
28 main features of the atmospheric dynamics and vertical structure of the Arctic atmosphere reasonably well. The influence
29 of the choice of the reanalyses used as initial and lateral boundary condition and of the LSM on the model results is
30 complex and no combination is found to be clearly superior in all variables analyzed. The model results show that a more
31 sophisticated formulation of land surface processes does not necessarily lead to significant improvements in the model
32 results. This suggests that other factors such as the decline of the Arctic sea ice, stratosphere-troposphere interactions,
33 atmosphere-ocean interaction, and boundary layer processes are also highly important and can have a significant
34 influence on the model results.

35



36 The “best” configuration for simulating Arctic meteorology and processes most relevant for pollutant transport (ASR +
37 NoahMP) is then used in a simulation with WRF including aerosols and chemistry (WRF-Chem) to simulate black
38 carbon (BC) concentrations in and around the Arctic and to assess the role of the modeled atmospheric circulation in the
39 simulated BC concentrations inside the Arctic domain. Results from simulations with chemistry are evaluated against
40 aerosol optical depth from several Aeronet stations and BC concentrations and particle number concentrations from
41 several stations from the EBAS database. The results with WRF-Chem show a strong dependency of the simulated BC
42 concentration on the modeled meteorology and the transport of the pollutants around our domain. The results also show
43 that biases in the modeled BC concentrations can also be related to the emission data. Significant improvements of the
44 models and of our understanding of the impact of anthropogenic BC emissions on the Arctic strongly depends on the
45 availability of suitable, long-term observational data of concentrations of BC and particulate matter, vertical profiles of
46 temperature and humidity and wind.
47

48 **1 Introduction**

49

50 The Arctic atmosphere is often characterized by a stable boundary layer and strong near-surface temperature inversions,
51 which limit turbulent mixing and vertical transport due to the combined effects of efficient cooling by upward long wave
52 radiation and reflection of short wave at the surface as a result of the high sea ice concentration and relatively flat and
53 homogeneous surface in the inner part of the Arctic Ocean. This negative radiation budget at the surface is amplified by
54 the typically low atmospheric moisture and cloud free conditions (Anderson and Neff, 2008). This effect is most
55 pronounced during winter when there is little or no sunlight and the surface is frequently covered by snow or ice
56 (Bradley et al., 1992). Consequently, a ‘dome’ forms that is characterized by low and constant potential temperatures and
57 isolates the Arctic lower troposphere from the rest of the atmosphere by acting as a barrier (the so-called Arctic front).
58 The Arctic front separates the cold Arctic air from warmer air in the south and can reach as far south as 40°N during the
59 coldest periods of the year (Stohl, 2006). The pronounced seasonal cycle of these atmospheric features strongly
60 determines transport mechanisms and pathways of pollutants into the Arctic (Schnell, 1984; Sharma et al., 2006).

61

62 Sea ice is part of a complex system that acts as an interface between the atmosphere and the ocean: the ice inhibits
63 vertical heat transfer between the atmosphere and the ocean, and contributes to the ice-albedo feedback mechanism. The
64 Arctic summer sea ice extent which has been observed over the last 30 years by satellites (Meier et al, 2006) has an
65 effect on the atmospheric temperature gradient in the lower atmosphere (Serreze et al., 2000). For example, a larger sea
66 ice-free area results in an increase in absorbed heat in the upper ocean, leading in in turn to the increase in the near-
67 surface temperature maximum (Jackson et al. 2010). This pattern affects atmospheric circulation by modifying weather
68 patterns in the Arctic and beyond, an effect also referred as the Arctic amplification (e.g. Screen and Simmonds, 2010;
69 Cohen, et al., 2014). These changes in atmospheric circulation connected to the decline of Arctic sea ice and changes in
70 continental snow-cover may also disturb temperature and precipitation patterns and increase the likelihood of extreme
71 weather events in mid-latitudes (Overland and Wang, 2010; Jaiser, et al., 2012; Handorf et al., 2015). The Arctic is
72 usually dominated by low pressure in winter forming a “Polar Vortex” of counter-clockwise circulating winds around the
73 Arctic. In a warming Arctic, high pressure replaces the low pressure, weakening the Polar Vortex and reversing the
74 circulation in which cold air flows southwards and warm air moves poleward (Honda, et al 2009; Petoukhov and
75 Semenov, 2010). This happened in winter 2009/2010 leading to unusually cold and snowy winter conditions in China,
76 Eastern Asia, the eastern United States and Europe, whereas West Greenland and the Bering Strait experienced
77 anomalous warm temperatures (Seager et al., 2010).

78

79 The pronounced seasonal cycle of the extent of the Arctic front plays a key role in the transport of black carbon (BC) and
80 other pollutants from source regions in mid-latitudes including Asia, Russia, North America and North Europe into the
81 Arctic. The maximum advection of pollution into the Arctic occurs usually in winter and early spring (*i.e.*, the ‘Arctic
82 haze’ season) and the minimum in summer when the Arctic aerosol is diminished by clouds and precipitation. In winter
83 (particularly in January), northern Eurasia is one of the major source regions for the Arctic pollution. Polluted air masses
84 from densely populated areas over East-Asia and North America are typically too warm and moist to directly penetrate
85 the polar dome, but they can descend into the Arctic middle or upper troposphere creating the Arctic haze (Stohl, 2006).

86



87 An important aim of many Arctic modeling studies is to improve our understanding of the causes, governing
88 mechanisms, and effects of Arctic amplification in order to better characterize the relevant physical processes in the
89 Arctic ocean–atmosphere–cryosphere system (in particular atmospheric boundary-layer processes). However, the
90 complexity of these processes and their interactions are still a challenge for modeling the Arctic (Vihma et al., 2014)
91 leading to a large spread in model results and a large sensitivity to a wide set of parameters regarding model
92 configuration and initialization.

93

94 Model biases in global climate models (GCMs) used for present-day and future climate simulations are partly related to
95 the horizontal and vertical resolution (e.g. Duffy et al., 2003; Wehner et al., 2010). Coarse horizontal resolution, for
96 example, can lead to the overestimation in sensible heat fluxes (Schmittner et al., 2002). In recent years Regional
97 Climate Models (RCMs) have gained popularity as they are computationally less expensive and are capable of
98 capturing mesoscale coupled processes and regional climatic evolution due to the ability to run at very high spatial and
99 temporal resolutions (Dethloff et al., 1996).

100

101 Model parameterizations play an important role in representing the vertical stratification and atmosphere–surface
102 energy exchange in model simulations (Dethloff et al., 2001). For example, some differences in regional model
103 simulations can be attributed to the boundary layer and surface parameterizations used, which result in surface flux
104 differences, and to the lateral moisture forcing, both of which affect moisture availability in the atmosphere (Rinke et
105 al., 2000). Therefore, it is critical to choose schemes suitable for the relevant Arctic processes as the application of non-
106 suitable schemes may introduce biases into the model outputs (Misenis and Zhang, 2010). This is particularly relevant
107 for Arctic studies as model parameterizations are typically developed and tested for mid-latitude conditions which may
108 not be suitable for use in the Arctic. Optimizing the model configuration for Arctic simulations is also needed in order
109 to be able to reduce the uncertainties in high-resolution projections of future Arctic climate and to be able to better
110 assess the impacts of climate change over the Arctic and beyond. For example, Land Surface Models (LSMs)
111 implemented in climate models (designed to calculate processes at the surface-atmosphere interface) play a critical role
112 in the representation of the heat budget.

113

114 Data used as initial and lateral boundary conditions also play an important role in the results of a RCM as the quality of
115 the data used have direct implications on the quality of the model outputs (Denis et al., 2003; Diaconescu et al., 2007).
116 However, the quality of a reanalysis may also vary regionally in some variables, especially in areas where observations
117 are sparse. Despite these limitations, reanalysis data provide gridded, self-consistent datasets suitable for model
118 evaluation and to perform a comprehensive examination of climate variability (Trenberth et al., 2008).

119

120 In addition to these challenges for numerical modeling of the Arctic atmosphere, deficiencies in the treatment of aerosols
121 and clouds may reduce the accuracy in climate model projections (Overland and Wang, 2013). The wide spread of model
122 results for the distribution of aerosols over the Arctic can be attributed to, for instance, differences in emissions,
123 chemistry and transport schemes (Shindell et al., 2008). Even though simulations of BC concentrations in the Arctic from
124 recent studies are in better agreement with observations, the amplitude of the seasonal cycle of BC is still underestimated
125 by most models (Eckardt et al., 2015). Because of the strong dependency of the radiative forcing by aerosols on the
126 surface albedo (Haywood and Shine, 1997), uncertainties in the surface albedo in the models translate directly into
127 uncertainties in estimates of the absorption of solar radiation by BC over high albedo surfaces (Myhre et al., 2009).



128

129 In this work we assess the performance of the Weather and Research Forecasting (WRF) model in representing
130 surface/atmosphere dynamics depending on the selection of the Land Surface Model (LSM) parameterization and
131 reanalysis datasets for initialization and provision of lateral boundary conditions (ERA-Interim and Arctic System
132 Reanalysis). The meteorological study is conducted over a five year period (2008-2012) for two seasons: February,
133 March and April (hereafter FMA) and July, August and September (hereafter JAS) coinciding with the maximum and
134 minimum sea ice extent in the region, respectively, within the Eurasian Arctic region. In particular, we assess the
135 magnitude of model biases on the surface meteorology including air temperature, wind speed, geopotential, the
136 evolution of the planetary boundary layer (PBL) and the position of the Arctic dome over the Eurasian Arctic. We then
137 investigate the implications of these model biases on the atmospheric circulation influencing concentrations of black
138 carbon in and advected towards the Eurasian Arctic by comparing results for aerosol optical depth (AOD) and
139 concentrations of black carbon from corresponding simulations with WRF-Chem with observations. The selected case
140 studies using the WRF-Chem are conducted for periods characterized by the minimum (warm period) and maximum
141 (cold period) observed tropospheric concentrations of BC over the Arctic north of 66°N (Stohl, 2006).

142

143 The paper is organized as follows: In section 2, the model configuration, experimental design and observational data
144 used for model validation are described. In section 3 we present the analysis of biases between model outputs and
145 observational data for the set of simulations with the WRF model focusing on meteorology, followed by results from
146 selected case studies using a model configuration including air chemistry and aerosols (WRF-Chem). The discussion
147 and conclusions are presented in section 4.

148

149 **2 Methodology**

150 **2.1. Model description and simulations**

151

152 **2.1.1 The weather research and forecasting model (WRF)**

153

154 For the sensitivity analyses we use the Advanced Research version 3.7.1 of the weather research and forecasting model
155 (WRF). WRF is a mesoscale forecast model and assimilation system developed by the National Center for
156 Atmospheric Research (NCAR) together with several partners. More details about the development of WRF can be
157 found in Michalakes et al. (2005). The WRF model is designed for a wide range of applications from idealized research
158 to operational forecasting, with an emphasis on horizontal grid sizes in the range of 1–10 km. Several physics options
159 in the form of different packages are available for relevant physical processes including microphysics of clouds and
160 precipitation, cumulus convection, planetary boundary layer and surface layer physics, turbulence and diffusion and
161 radiation (longwave and shortwave). All of these schemes consist of solver-dependent routines approximating or
162 parameterizing physical processes that are too complex or too computationally costly to be explicitly represented
163 (Skamarock, 2005).

164

165 The model physics packages used in this study for the meteorological WRF model set up (also defined in Table 1) are:
166 (1) the atmospheric surface layer scheme which calculates momentum and heat fluxes at the surface uses the MM5
167 similarity scheme based on Monin-Obukhov with Carlsol-Boland viscous sub-layer and standard similarity functions



168 from look-up tables (Beljaars, 1994), (2) convective processes following the Grell 3D parameterization, a multi-closure,
169 multi-parameter ensemble method which is an improved version of the Grell-Devenyi ensemble scheme for horizontal
170 grid sizes larger than 10 km (Grell and Devenyi, 2002), (3) the PBL (Planetary Boundary Layer) parameterization
171 implemented is the non-local Yonsei University (YSU) scheme (Hong, 2010) which calculates atmospheric tendencies
172 of temperature, moisture with clouds, and horizontal momentum (4) Lin et al. (1983) cloud microphysics including
173 cloud ice, snow and graupel processes, (5) radiative processes are parameterized using the Rapid Radiative Transfer
174 Model for Long Wave (LW) radiation (Mlawer et al., 1997) and the Dudhia (1989) scheme for short wave (SW)
175 radiation, and finally (6) the land-surface parameterization (hereafter LSM) which characterizes heat, moisture and
176 radiation fluxes at the surface and resulting feedbacks with the atmosphere. Sensitivity analyses were conducted using
177 two available LSM schemes, the Unified Noah Land Surface Model (Chen and Dudhia, 2001) and the Noah-MP Land
178 Surface Model (Niu et al., 2011; Yang et al., 2011). A more detailed description of both LSM schemes is given in the
179 next section.

180

181 **Land Surface Models (LSMs)**

182

183 LSMs simulate surface/atmosphere dynamics and the land surface variability by including the relevant land
184 surface/hydrology processes. The variables resolved in the LSM parameterization interact directly with other physical
185 parameterizations in WRF such as cloud microphysics and cumulus schemes, radiation schemes and certain land and
186 ocean variables that impact the vertical transport (surface layer) and the PBL. LSMs also calculate sea ice processes
187 affecting ice temperature, skin temperature, snowpack water content, snow depth, and surface energy (note that LSMs
188 in WRF currently do not include sea ice generation, melting, and change in thickness).

189

190 Some modeling studies (e.g. Misemis and Zhang, 2010) have suggested that there is a higher sensitivity of WRF to the
191 LSM schemes than to PBL schemes for simulating standard meteorological variables including surface temperature,
192 relative humidity and wind vectors, due to differences in surface fluxes. This fact may affect performance of the
193 atmospheric model over the Arctic. Thus, we performed sensitivity studies with both LSM schemes to investigate their
194 strengths and limitations for Arctic applications and the implications in reproducing important meteorological
195 variables. The experimental design of the simulations performed in this work is well detailed in section 2.1.4 and also
196 shown in Table 1b.

197

198 The two tested LSM schemes used in this work are the Noah Land Surface Model (Noah LSM) developed by the
199 Oregon State University and a more detailed and updated version, the Noah Multi-Physics (NoahMP). The Noah
200 scheme calculates the soil temperature, as well as the soil and canopy moisture in four layers (10, 30, 60, and 100 cm
201 thick) and includes fractional snow cover and frozen soil physics (Chen and Dudhia, 2001; Ek and Mitchell, 2003). The
202 surface skin temperature is calculated with a single linearized surface energy balance equation representing the
203 combined ground-vegetation surface. Soil temperature is calculated solving the thermal diffusion equation, soil
204 moisture is predicted using the Richards equation (Richards, 1931). Known limitations of the Noah LSM model include
205 the tendency of the Noah scheme to produce too little snow cover during spring (e.g., Barlage et al., 2010), significant
206 biases in surface temperature due to structural limitations when heterogeneities exist at the surface (Miguez-Macho et
207 al., 2008; Fan et al., 2007) and larger seasonal variations in snow water equivalent compared to other LSM schemes
208 (Chen et al., 2014). As a side note, one of the reasons for using the latest version of WRF 3.7.1 is due to the fact that in



209 previous versions, the Noah LSM was overestimating surface temperature over sea ice in case of snow melt over ice,
210 particularly in late spring and summer as found by Marelle et al (2016). This can interfere with the air mass exchange
211 with mid-latitudes, and increased vertical mixing from lower stability.

212

213 The other LSM model used in this study is the NoahMP scheme (Yang et al., 2011). NoahMP contains a multi-layer
214 snow pack with liquid water storage and melt/refreeze capability and a snow-interception model describing
215 loading/unloading, melting/refreezing, and sublimation of the canopy-intercepted snow (Niu, et al., 2011). It considers
216 equations for the snow albedo, snow temperature, density, total water content and content of liquid water which help to
217 improve simulations of the diurnal variations of the snow skin temperature, which is critical for computing available
218 energy for melting (Chen et al., 2014; Niu, et al., 2011). Studies comparing Noah and NoahMP (e.g. Miguez-Macho
219 and Fan, 2012) have shown that NoahMP tends to show larger daytime biases in the diurnal cycle of the 2-m
220 temperature and dew point during wet months compared to dry months. The same studies also indicated that monthly
221 absorbed SW radiation and sensible heat shows that less solar radiation is absorbed in Noah then in NoahMP resulting
222 in a colder surface and a lower (or negative) sensible heat flux.

223

224 **2.1.2 Initial and lateral boundary conditions**

225

226 High quality data are needed in order to minimize the uncertainties introduced by the initialization of the model and the
227 provision of the lateral boundary conditions required during run-time. Reanalysis datasets used for initialization and to
228 create lateral boundary conditions are created from a combination of both observed (assimilated) variables (e.g.,
229 temperature,) obtained from different sources (e.g. ground-based observations and radiosondes), and derived fields
230 (e.g., precipitation and cloudiness). These variables are assimilated into a global, gridded, and temporally homogeneous
231 dataset making the best possible use of a large number of observations (Dee et al., 2014).

232

233 In this work two reanalysis datasets are used for the model initialization and to recreate the lateral boundary conditions:
234 the ERA-Interim reanalysis (Dee et al., 2011) produced by the European Centre for Medium-Range Weather Forecasts
235 (ECMWF) and the Arctic System Reanalysis (ASR) (Bromwich et al., 2010). ERA-Interim is a global reanalysis of
236 recorded climate observations over the last 30 years with an improved atmospheric model and assimilation system and
237 replaces its predecessor ERA-40 using four-dimensional variational data assimilation (4D-Var). ERA-interim is
238 available as a gridded data set at approximately 0.7° spatial resolution and 37 atmospheric pressure levels (Dee et al.,
239 2011). Some studies have shown that ERA-Interim agree well with observations of sea level pressure, near-surface air
240 temperature, surface shortwave and longwave radiative fluxes, precipitation, and wind speed when compared against
241 other reanalysis datasets (refer to Lindsay et al., 2014). However, global reanalyses like ERA-interim have many
242 known problems at high latitudes for different reasons including the scarcity of observational data at these remote
243 areas.

244

245 On the other hand, the ASR, developed by the Polar Meteorology Group (PMG) of the Byrd Polar Research Center
246 (BPRC) at the Ohio State University in collaboration with several other institutions is a reanalysis recently developed
247 which implements parameterizations optimized for the Arctic. It consists of a high-resolution version of the Polar
248 Weather Forecast Model (PWRF) and the WRF-VAR and High Resolution Land Data Assimilation (HRLDAS)
249 (Bromwich et al., 2016). PWRF is available through the NCAR Research Data Archive (NCAR, 2015). In this study,



250 we use the ASRv1 30-km consisting of a gridded dataset at a spatial resolution of 30 km and with 29 pressure levels.
251 The data set consists of 27 surface and 10 upper air (measurements in the part of the atmosphere above Earth's surface)
252 variables and 3 soil variables. Differences in the reanalyses used to create the forcing data for our WRF simulations
253 might have a significant impact on the modeled meteorological fields. A comparative analysis of both reanalyses is
254 given in the following section whereas the results from the WRF model simulations to assess the differences resulting
255 from the initial and boundary conditions are presented in section 3.

256

257 **Comparison between ERA-interim and ASR reanalysis**

258

259 The capability of both reanalysis to represent the Arctic atmosphere has been recently compared and documented (e.g.
260 Wesslén et al., 2014; Bromwich et al; 2016). For example, Wesslén et al. (2014) found that while ERA-Interim has a
261 systematic warm bias in the lowest troposphere, ASR has on average a cold bias of about the same magnitude. These
262 authors also found that improvements in the modeled cloud properties, radiation budget, and surface temperature in
263 ASR were not significant in spite of its more sophisticated parameterizations of cloud microphysics. In a more
264 comprehensive assessment of the differences between both reanalysis, Bromwich, et al (2016) found that in higher
265 latitudes throughout Europe, Siberia and North America, the ASR biases are small and cool, whereas ERA-Interim
266 shows warm biases across much of the domain, particularly across Siberia, Alaska and western North America. The
267 authors argue that this trend is probably related to the complex topography in these regions. This is because accurate
268 simulation of near-surface winds over complex terrain is particularly challenging due to highly variable small scales
269 resulting in winds that are not always well represented on a coarse model grid.

270

271 In our study, we evaluate the differences in four important variables (2m temperature and 10m wind speed, sea ice
272 fraction and sea level pressure) between both reanalyses over our studied domain. Fig. 1 shows the differences between
273 both reanalyses for the five-year average for the summer (JAS) and winter (FMA). Biases in the near-surface variables
274 are generally smaller and temperatures are generally colder in the ASR than in ERA-interim. This agrees with the
275 typical biases described in Bromwich et al. (2016) who also found a good agreement in the near-surface moisture and
276 wind fields when compared against observational data. ERA-Interim tends to show larger values than ASR of 10m
277 wind speed over the high Arctic Ocean characterized by sea-ice covering, whereas 10m wind speed tends to be
278 underestimated over continental area, specifically over the east coast of Greenland (Fig. 1). This is consistent with the
279 findings from Moore et al (2013). In their study they found that ERA-Interim has surface wind speeds over the
280 Scoresby Sund region of east Greenland that are too low and not entirely consistent with observations of
281 topographically forced drainage flow in this region. However, when compared against observations in our study, both
282 reanalyses tend to slightly underestimate wind speeds (see section 3.2). Previous model studies have demonstrated that
283 ERA-Interim leads to overly strong turbulent mixing (i.e. Jung et al., 2010), which leads to weak inversions, warmer
284 temperatures near the surface and stronger horizontal wind. As for sea-ice (fourth row in Fig. 1), both datasets are very
285 consistent for both FMA and JAS seasons. However, in the ERA-Interim reanalysis, several locations near the Russian
286 coastline and in some water bodies in North Europe have a larger sea-ice fractions (~0.3) compared to the ASR,
287 especially during the FMA period. This can be explained by the fact that ASR uses MODIS sea ice data. Also the sea
288 ice thickness and snow cover on sea ice vary in ASR whereas these values are fixed in ERA-Interim. Not included in
289 our analyses are the differences at all atmospheric levels; Bromwich et al (2016) found that temperatures are similar
290 with annual mean biases within $\pm 0.2^{\circ}\text{C}$ at nearly all levels from both reanalyses. However, Jakobson et al. (2012)



291 found that during spring and summer over the Arctic Ocean most reanalyses (including ERA-Interim) have large errors
292 in the vertical profiles of air temperature as well as specific and relative humidity.

293

294

295

296 **2.1.3 The WRF-Chem model**

297

298 We also conduct selected case studies with the Advanced Research WRF (ARW) including chemistry and aerosols
299 (WRF-Chem) version 3.7.1. WRF-Chem is capable of calculating chemical processes online including the feedback of
300 radiatively active trace gases and aerosols on atmospheric dynamics. Additional processes simulated by WRF-Chem
301 include emissions, transport of chemical species and aerosols, chemical transformation of species, aerosol physics (e.g.,
302 nucleation, condensation and coagulation), interaction of particles with radiation (photolysis and heating rates),
303 aerosol-cloud interactions (cloud condensation nuclei), and wet and dry deposition to the surface (Grell et al., 2005;
304 Peckham et al., 2011). In earlier versions of the model, the Noah LSM was overestimating surface temperatures over
305 sea ice in case of snow melt over ice leading to possibly increased exchange with the mid-latitudes and increased
306 vertical mixing from lower stability. This may be one of the reasons why concentrations of BC have been simulated
307 poorly in versions of WRF-Chem prior to v3.7 such as the results published in Eckhardt et al (2015).

308

309 For the case studies with the WRF-Chem model, we use the same physical parameterizations summarized in Table 1
310 plus the input data and chemistry/aerosols configuration (Table 2) applied in previous modeling studies over the Arctic
311 (e.g., Grell, et al., 2011; Marelle et al., 2016). For the simulation of aerosol processes, we use the MADE/SORGAM
312 scheme (Modal Aerosol Dynamics model for Europe, Ackermann et al., 1998 / Secondary Organic Aerosol Model,
313 Schell et al., 2001). The MADE/SORGAM scheme simulates the aerosol size distribution using a modal approach with
314 three overlapping log-normal modes. Within each mode, all particles are assumed to have the same chemical
315 composition (internal mixture). This is important factor because the interaction of aerosols with chemistry or
316 atmospheric processes depends on their size and the chemical composition (Baklanov et al., 2014). The aerosol
317 processes are coupled to the clouds and radiation, including photolysis, and gas-phase chemistry. Aerosol and
318 chemistry variables are initialized from the global chemistry model MOZART. More details on WRF-Chem can be
319 found in Peckham et al. (2011), Grell et al. (2005) and Fast et al. (2006).

320

321 **Emissions**

322

323 Emission data are an important model input as their quality determines to a great extent the results when modeling air
324 pollutants. In this study, anthropogenic emissions are obtained from the EDGAR HTAP v2.2 inventory (EDGAR:
325 Emission Database for Global Atmospheric Research), created in joint collaboration between the European
326 Commission, and the Task Force on Hemispheric Transport of Air Pollution (TF HTAP). EDGAR HTAP v2.2 reports
327 monthly anthropogenic emissions of greenhouse gases (e.g. CO₂, CH₄ and N₂O), precursor gases (e.g. CO, NO_x and
328 SO₂) and aerosols (PM₁₀) per source category at country level on 0.1° × 0.1° grid maps from the energy, industry,
329 transport and residential sectors and annual emissions from shipping and aviation. Emission data from small
330 agricultural fires are not included in v2.2 and were taken from the previous version EDGAR HTAP v1.0. Information
331 about methodology and emission factors used for the EDGAR emission calculations is well documented in Janssens-



332 Maenhout et al. (2015). Biological emissions are included applying MEGAN (Model of Emissions of Gases and
333 Aerosols from Nature), which calculates biogenic emissions online based on simulated temperature, leaf area index and
334 land use data (Guenther et al., 2006). For biomass burning emissions, we used the Fire Inventory (FINN) version 1 also
335 developed by NCAR. FINN is based on daily satellite observations of fires and land cover, combined with emission
336 factors and estimated fuel loadings (Wiedinmyer et al., 2011).

337

338 **2.1.4 Model domain and simulations**

339

340 The WRF model setup used in this study consist of a single domain covering the Eurasian Arctic region at a horizontal
341 resolution of 15 km over an area of approximately 4000 x 3800 km (centered at 70°N and 40°W), which includes large
342 parts of northern Europe and North-west Russia (Fig. 2). The atmosphere is vertically resolved in 30 full σ -levels with
343 15 model levels below $\sigma = 0.9$. The lowest model level in this configuration is centered at 10 meters, whereas the
344 model top-level pressure is set at 100 hPa. The reason for such high vertical resolution in the lowermost troposphere is
345 based on the premise that vertical mixing and turbulent fluxes which are key processes in the PBL may be affected by
346 the total number of vertical levels and in particular by the number of vertical levels close to the surface. For example,
347 recent studies suggest a strong sensitivity to the modeled 10-m wind speed on the model's vertical resolution (i.e.
348 Chou, 2011; Kleczek et al., 2014).

349

350 Simulations with the atmosphere -only model WRF were performed over 5 cold (February, March and April) and 5
351 warm (July, August and September) seasons from the year 2008 to 2012 coinciding with the maximum and minimum
352 sea ice extent in the region. These periods are also characterized with the minimum (warm period) and maximum (cold
353 period) observed concentrations of BC over the Arctic above 66°N (Stohl et al., 2013). Four different sensitivity
354 analyses were performed and named according to the configuration chosen, as described later and also shown in Table
355 1b. The sensitivity analyses are aiming at quantifying the impact of the different LSMs and reanalyses used for the
356 creation of the initial and lateral boundary conditions on model biases compared with surface measurements as well as
357 observed vertical profiles of meteorological variables from stations in the Eurasian Arctic.

358

359 Finally, we analyze a suite of selected cases using WRF-Chem (summarized in Table 2) to assess how the model
360 simulates BC transport into the domain. This is important to estimate the uncertainty of the climate impact of BC
361 transported into the Arctic when assessing different emission scenarios. These case studies with WRF-Chem include 1-
362 month simulations for April 2008, July 2008, April 2009 and July 2010. The case studies were selected based on the
363 availability of observational data used for comparison with the model results. The year 2008 coincides with the Polar
364 International Year in which several ground and aircraft campaigns took place. Also, some authors have described
365 exceptionally high forest and agricultural fire activity in Russia, Kazakhstan and Southeast Asia in April 2008
366 (Warneke et al., 2009; Fisher et al., 2010). We therefore also include April 2009 as an additional case study for
367 comparison with the previous year. July 2010 was an exceptionally hot summer (the hottest summer since the year
368 1500) impacting in particular Western Russia and resulting in a historical heat wave which triggered 500 wildfires
369 around Moscow (Barriopedro et al., 2011). Such wildfires could be important emission sources of BC transported into
370 the Arctic. Previous studies have demonstrated that fires in Eastern Europe and Russia can lead to substantial increases
371 in the atmospheric loading of pollutants at several surface sites in the European Arctic (Stohl et al., 2007) which we
372 aim to confirm with the simulations carried out in the present study.



373

374 **2.2 Observational data for model evaluation**

375

376 **2.2.1 Meteorological data**

377

378 Surface meteorology in the model is evaluated against observations from 24 weather stations in North Europe and
379 Russia spread throughout the model domain with data available for the studied period and acquired from the Met
380 Office Integrated Data Archive System (MIDAS) (Fig. 2). The southernmost station is “*Leningrad*” at 59.95°N, located
381 in the city of Saint Petersburg, whereas the northernmost station is “*Krenkelja*” at 80.64°N in the Franz Josef Land
382 located between the Arctic Ocean, and the Barents and Kara Seas. The MIDAS dataset comprises daily and hourly
383 weather observations including hourly wind parameters, maximum and minimum air temperature, soil temperature,
384 sunshine duration and radiation fluxes as well as daily, hourly and sub-hourly precipitation measurements, some
385 climatological data and marine observations. The reporting frequency varies among the different stations. At best the
386 data are hourly although some stations report only at 0600, 1200 and 1800 local time (*Met Office*, 2012). In this study
387 we evaluate the root mean square error (RMS) of both reanalysis used for creating the initial and boundary conditions
388 (ERA-Interim and ASR) together with the time series from the different WRF sensitivity experiments for temperature,
389 wind speed, water vapor mixing ratio and snow depth. Bias and RMS are calculated from the comparison between
390 model simulations and surface observations over the five year period 2008-2012. In order to facilitate the analysis, the
391 24 observational stations were divided into two groups: 1) stations located inside the Arctic Circle (here defined as
392 66°N 34’) and 2) those stations located below the Arctic Circle (Table 3).

393

394 We also use observed vertical profiles of temperature, wind speed and relative humidity from the enhanced version of
395 the Integrated Global Radiosonde Archive (IGRA) (<http://www1.ncdc.noaa.gov/pub/data/igra/>) to examine both the
396 variability during the five year period and to assess the model performance and the reanalyses in reproducing the
397 vertical structure of these variables. The IGRA dataset consists of quality-controlled radiosonde and pilot balloon
398 profiles, freely available at more than 1500 globally distributed stations covering varying time periods, many with
399 observations extending from the 1960s to the present (Durre et al., 2006). Upper air data are usually measured by twice
400 daily by radiosonde soundings, taken at 00 and 12Z (Greenwich time). As can be seen in Fig. 2, there are only very few
401 long-term surface measurements in the high Arctic, especially in the North Russian Arctic. Nevertheless, the available
402 stations are helpful to provide a snapshot of the vertical structure of the Arctic atmosphere and to test model
403 performance.

404

405 **2.2.2 Aerosol observations**

406

407 Routine measurements to monitor atmospheric chemistry components in the Eurasian Arctic are very scarce. The
408 limited availability of long-term measurements results in a high uncertainty of the estimated climate impacts of
409 aerosols in the Arctic. In order to help filling these gaps and to improve our understanding of Arctic pollution, intensive
410 field campaigns including surface and aircraft observations have been conducted in recent years. In addition, surface
411 based sun photometer instruments like the AERosol NETwork (AERONET) provide information on aerosol loading
412 and aerosol optical depth as well as other physical properties. AERONET is an optical ground-based aerosol-
413 monitoring network and data archive consisting of identical automatic sun-sky scanning spectral photometers operated



414 by national agencies and universities (Holben et al., 1998). There are, however, only very few long-term observations
415 from the AERONET network available close to the Arctic Circle. Seven AERONET stations were selected in this study
416 (Fig. 2) for an initial evaluation of the model performance. In addition, available observations of Equivalent Black
417 Carbon and measurements of the size integrated aerosol particle number concentration (PNC) measured by a
418 Condensation Particle Counter (CPC) from the monitoring EBAS database (EBAS, 2013) operated by the Norwegian
419 Institute for Air Research (NILU) are used for comparison with our model simulations. All data are available from the
420 EBAS database (<http://ebas.nilu.no/>).

421

422 **3 Results**

423

424 **3.1 Meteorological analyses**

425 **3.1.1 Large scale atmospheric circulation**

426

427 The pattern of the large scale circulation during the five year period is shown for two sensitivity experiments
428 (*noahmp_asr* and *noahmp_ecmwf*) in Fig. 3 for winter (February, March and April) and in Fig. 4 for summer (July,
429 August, September) averaged over five years (2008-2012). The position of the polar dome in our domain is also
430 calculated given its importance as proxy for studying the transport of pollutants and its implications (Stohl, 2006). Here
431 we define the polar dome following Jiao and Flanner (2016) by calculating the maximum latitudinal gradient of the 500
432 hPa geopotential height. The reasoning for this approximation is that where the latitudinal geopotential height gradient
433 is largest, the zonal component of the geostrophic wind is likely to be strongest, and this narrow band of strong
434 geostrophic wind plays an important role for tracer transport in the middle troposphere (we refer to Jiao and Flanner
435 (2016) for further details). Jiao and Flanner (2016) also suggest that the northward expansion of the polar dome
436 correlates with warming in some regions in Alaska and Chukchi Sea. In this work, the calculated isopleth defining the
437 polar dome (gray line in Figs. 3 and 4) extends to middle-latitude regions in the winter months (FMA) when the jet
438 wind speed reaches its maxima as a result of the stronger temperature gradient between low- and high-latitude regions.

439

440 In general, the differences between both reanalyses used to drive the model (see section 2.1.2) are also noticeable in the
441 model outputs. The persistence of the Icelandic Low (the low-pressure center located usually between Iceland and
442 southern Greenland) is visible in both WRF experiments. The low pressure area generally has an elliptic shape and is
443 located near 90°N and 30-50°W, also described by Sahsamanglou (1990). The low pressure weakens and splits into
444 two centers in summer, one near Davis Strait and the other west of Iceland during the FMA period in the experiments
445 using the ASR data (left column), whereas this pattern is highly variable in the experiments using ERA-Interim
446 reanalysis for initialization and as lateral boundary conditions (right column). For the winter period FMA (Fig. 3), there
447 is a particularly strong average low pressure region in the year 2011, also reflected in high wind speeds at 850hPa (~17
448 m/s) over a large portion of the North Atlantic Ocean over the model domain. In the model experiments using ASR
449 data, wind speeds tend to be higher than in the experiments using ERA-Interim data during all years analyzed. High
450 wind speeds are also simulated over central Russia in both experiments during 2008. As for the summer period (JAS)
451 shown in Fig. 4, the low pressure region is less well defined. Notice that in summer 2010, cyclonic activity tends to be
452 high with a significant increase in wind speed, particularly over the Kara Sea, with 10m wind speeds as high as 11m/s



453 over a vast area of North-West Russia. The year 2010 was characterized by a strong heat wave that affected a large
454 part of North-West Russia. This event was linked to a combination of a very strong low pressure system near Iceland
455 and a less intense than usual low pressure system located near the North Pole (Trenberth and Fasullo, 2012). This is
456 also reflected in the position of the Polar Dome which is displaced further south in winter time, particularly in the
457 *noahmp_ecmwf* experiment (Fig. 3).

458

459 It has been noted in previous research that this anomalously cold winter in 2009/2010 was followed by an abrupt shift
460 to a warmer-than-normal early growing season (the Russian heat wave took place in summer 2010) which was
461 consistent with a persistently negative phase of the North Atlantic oscillation (NAO) (Wright et al, 2014). This cold
462 winter was characterized by unusually cold and snowy conditions over China and eastern Asia, the eastern United
463 States and Europe that lasted until February 2010. In contrast, other regions such as west Greenland and the Bering
464 Strait experienced anomalous warm temperatures (Seager et al., 2010).

465

466 3.1.2 Surface meteorology

467

468 Fig. 5 shows the spatial distribution of the planetary boundary layer height (PBLH) calculated by the model for all
469 sensitivity experiments for the winter (Fig. 5a) and for summer months (Fig. 5b). The PBLH from the model is
470 calculated by the PBL scheme and is based on the bulk Richardson number method. A detailed description of the
471 methods used to calculate the PBLH in the scheme can be found in Hong (2010). The Arctic atmosphere is stably
472 stratified most of the year with the lowest kilometer of the atmospheric column north of 70°N almost permanently
473 stably stratified. This is more pronounced during winter than in summer as seen, for instance, in the representation of
474 the planetary boundary layer height in Fig. 5. In winter the average modeled PBLH oscillates around 1000 m over the
475 ocean in the North Atlantic sector between the Svalbard Archipelago and the Nordic countries (Fig. 5a). However,
476 there are significant differences in the PBLH close to 90°N, where the model experiments using the ASR data simulate
477 a deeper PBLH (~250 m) compared to those obtained using the ERA-Interim reanalysis (~150 m). In summer (Fig. 5b),
478 the PBLH is deeper over the continents due to the large insolation during this period resulting in convective mixing.
479 The PBLH over the Arctic Ocean continues to be lower than in winter due to the increasing solar radiation over high
480 latitudes. In the experiments using the NoahMP scheme, the PBLH is generally deeper over the continent than in those
481 using the Noah LSM scheme in both seasons. Particularly over the Nordic countries and Russia in winter, the values
482 are between the 400-450 m compared with a shallower PBLH obtained with the Noah scheme (~200-350 m)
483 independent of the reanalysis used. The differences in PBLH between the two LSMs occurs at the time of day when the
484 PBL becomes a shallow layer after sunset as also reported by previous studies (e.g. Pino et al., 2006; Milovac et al.,
485 2016). Misenis and Zhang (2010) suggest that NoahMP simulates a later daily collapse of the PBL of at least 30 min
486 which might also be related to wind shear at the top and bottom of the PBL. This may have important implications on
487 other meteorological variables as described by Misenis and Zhang (2010).

488

489 The differences in the main meteorological variables analyzed here, i.e. sea level pressure (SLP), 2m temperature and
490 10m wind speed are summarized for the individual sensitivity experiments in Table 4. All sensitivity analyses tend to
491 show stronger cold biases in 2m temperature than the reanalyses used to create the initial and lateral boundary



492 conditions. These cold biases over the region have been also found by other authors (i.e. Déqué et al., 2007, Katragkou,
493 et al 2015). Katragkou et al (2015) suggest that coldest mean temperature bias in northern Europe is related to an
494 underestimation of SW radiation at the surface and an overestimation of cloud cover. This is particularly seen in WRF
495 model configurations using the Grell convective scheme (also used in our experiments). The spatial distributions (Fig.
496 6) of the modeled 2m temperature, 10m wind speed and SLP match those estimated from the observational data at the
497 24 stations dispersed throughout the domain (colored circles in Fig. 6) reasonably well for the winter months (left
498 column) and summer months (right column).

499

500 In order to investigate the temporal agreement between the simulated and observed meteorological variables from the
501 24 stations used to validate the model simulations, Taylor plots are used. These diagrams provide a statistical summary
502 of how well observed and simulated 3-month time series match each other in terms of their temporal correlation (R)
503 and normalized standard deviation (NSD) (Taylor, 2001). In the Taylor diagram R and NSD are given in this polar
504 projection by the angular and the radial coordinate, respectively (Fig. 7). The reference point (observations) is located
505 at $x=1, y=0$. The linear distance between a model and the reference point is proportional to the RMS error making it
506 easy to identify the stations for which the model experiments perform well. The correlation in 2m temperature ranges
507 between 0.45 and 0.85 depending on the model simulation and the station. In general, there is a large spread among the
508 different model experiments and measurement stations with no model configuration being clearly superior at all
509 measurement stations. Also the temporal standard deviation of the model experiments varies widely among the
510 different experiments and for the different stations. This large spread is even more pronounced for the time series of
511 10m wind speed. For 10m wind speed, the NSD from the model results is typically much larger than 1 with maximum
512 values of up to 4. This overestimation in 10m wind speed in the model experiments can also be seen in Fig. 6, with the
513 overestimation being particularly pronounced over the continent.

514

515 Given the sparsity of available long-term surface measurements, it is difficult to assess which model experiment
516 performed better compared to observations using only Taylor diagrams. We therefore also calculated biases and RMS
517 (shown in Table 4) in which we separate the stations in two groups, those below and those above the 60°N, and also
518 include both reanalyses (ERA-Interim and ASR) in the comparison with observations. In general for the studied
519 domain both reanalyses are slightly warmer in the winter months (FMA) with ERA being 0.40°C and ASR 0.18°C
520 warmer at stations above the 66°N and slightly negative biases in stations outside the polar circle with -0.12°C and -
521 0.28°C, respectively. A slightly negative bias in the summer months (JAS) in all the model experiments is persistent for
522 all the stations. In Bromwich et al (2016), the authors indicate that ASR may be little cold over sea ice. Our model
523 experiments also show a similar negative bias tendency as the ASR reanalysis for all variables. For 10m wind speed,
524 the ERA-Interim data show slightly positive biases of less than 1 m/s at all the stations in the two seasons. ASR biases
525 are quite small but tend to be negative. With regard to sea level pressure, both reanalyses generally show slightly
526 negative biases in all seasons ranging from -0.03 to -0.59 hPa.

527

528 3.1.3 Atmospheric vertical structure

529 Vertical gradients in potential temperature and wind speed are good diagnostics to estimate the transport pathway of
530 emissions as shown for example in Aliabadi et al. (2015). Fig. 8 shows the radiosonde profiles at three stations around
531 our domain for temperature, wind speed and relative humidity at 12UTC (the most data are available at this time of the



532 day) averaged of the winter months (February, March and April) for the five years of study. The overall performance
533 statistics for these meteorological variables is also summarized in Table 5. Radiosonde observations of the vertical
534 structure of temperature and wind speed typically show stable profiles with a temperature inversion caused by the lack
535 of surface heating by the sun. Temperature inversions inhibit convection so emissions are not lifted to higher altitudes
536 and pollutants from the surface can become trapped close to the ground. The comparison of the vertical temperature
537 profiles shows generally good agreement of the model experiments with observations. However for some stations (i.e.
538 Ostrov Dikson) the modeled profiles when using simple Noah scheme show closer proximity with observations than
539 when using Noah-MP. The overestimation of the vertical temperature may be caused by a possible overestimation
540 of the turbulent sensible heat flux towards the surface causing premature warming along the vertical column.
541 This may not be the case of stations located in urban areas such as Leningrad station which shows more consistency in
542 modeled profiles compared to the observations. Wind speeds tend to be overestimated in the model. These positive
543 biases in wind speed have been reported in other locations with complex terrain and can introduce systematic errors to
544 the simulations (i.e. Jiménez and Dudhia, 2012; Santos-Alamillos, et al 2013). For example, Jiménez and Dudhia
545 (2012) reported a positive bias in wind speed simulated with WRF over plains and valleys whereas negative biases may
546 be present over hills and mountains. The unresolved topographic features can produce additional drag, which is not
547 parameterized in the WRF, in addition to the drag generated by vegetation. The magnitude of this bias is related to the
548 differences between the actual and the elevation of the stations in WRF (Santos-Alamillos, et al 2013). Again, there is
549 no particular model configuration that clearly outperforms the other WRF sensitivity simulations. The radiosonde
550 observations are close to the lateral boundaries of the model domain. As measurements are very sparse in this region,
551 also the reanalysis data used as lateral boundary conditions are subject to large uncertainties translating directly into the
552 model results and can possibly overcompensate the effect of other model configuration options such as the choice of
553 LSM.

554

555 For the model simulations with WRF-Chem, we therefore use the configuration that includes the most processes
556 relevant to modeling BC transport into the Arctic, the configuration using the ASR reanalysis and the NoahMP LSM
557 scheme. Even though the ASR/NoahMP configuration shows slightly smaller biases in simulated wind speed than the
558 other configurations, the evaluation of the simulated meteorology did not show a clearly superior configuration for all
559 variables analyzed. We regard this slightly small bias in wind speed as an advantage because of the importance to
560 simulated transport pathways of pollutants.

561

562 3.2 Case studies using WRF-Chem in the Eurasian Arctic

563 Atmospheric chemistry models have struggled for a long time to capture the observed distribution of aerosols in the
564 Arctic (Shindell et al., 2008; Koch et al., 2009, Eckard, et al 2015). The concentrations of BC during the Arctic Haze
565 season (winter to early spring) are typically underestimated whereas summer concentrations are sometimes
566 overestimated, in some cases by more than an order of magnitude (Shindell et al., 2008). In this study, several cases
567 covering the periods of maximum and minimum BC over this region coinciding with the availability of measurement
568 data for the model evaluation were selected within the five years used to evaluate the meteorology. The aim is to assess
569 the performance of the model in calculating concentrations of specifically BC over our model domain and to analyse
570 possible linkages with the modeled atmospheric circulation during this period.



571

572 **3.2.1 Aerosol Optical Depth**

573 We first assess the model AOD at 550 nm against observations from AERONET sun-photometers for the four case
574 studies using WRF-Chem to investigate the model's performance in representing spring/summer high/low aerosol
575 loadings. The aerosol loadings are expected to include a significant contribution from BC, especially at sites close to
576 urban or industrialized centers. Fig. 9 shows the modeled spatial distribution of AOD for the four cases April and July
577 2008 and April 2009 and July 2010. High aerosol loadings over northern Europe and western Russia are modeled with
578 monthly mean values typically in the range of 0.1-0.2. Maximum AOD values are modeled in April 2008 with values
579 exceeding 0.35 particularly over parts of western Russia. Several authors have reported that the Arctic AOD is mostly
580 dominated by sulfate in spring and organic carbon (from fires) in summer (Breider et al., 2014). Our model results
581 agree reasonably well with AOD observations in July 2008 at most of the selected stations (colored circles in Fig. 9a),
582 but overestimates AOD by about 0.1 in April 2008 and underestimates AOD by up to 0.1 in July 2010. Besides the
583 lateral boundary conditions for aerosols, also the emissions play a crucial role in the modeled aerosol loadings resulting
584 in differences between modeled and observed AOD. The modeled AOD also strongly depends on the relative humidity
585 as particles taking up water are more efficient in scattering light. Particularly in the range of high relative humidities,
586 even small errors in the modeled relative humidity can result in significant differences in the modeled AOD.

587 Also shown in Fig. 9b is a Taylor diagram comparing the observed time series of AOD at the AERONET stations with
588 the model results. The temporal correlation between model results and observations depends strongly on the month
589 simulated and ranges between 0.2 for April 2008 and 0.75 for July 2010. In most cases, the model overestimates the
590 temporal standard deviation of AOD which is related to a general overestimation of AOD at these stations, The only
591 exception is for July 2010, where the model tends to underestimate AOD values, as also observed in the spatial
592 distribution in Fig 9a. This fact can be highly dependent on the fire emission data which causes the model to under
593 predict amounts of burned areas, particularly for those stations located in west Russia which results in model
594 underestimating non-dust AOD. This problem could potentially be solved by testing another more advanced biomass
595 inventory of monthly burned area (e.g. the latest Global Fire Emissions Database, Version 4 (GFEDv4.1) biomass
596 inventory, also described in Randerson et al 2015).

597

598 **3.2.2 Black Carbon**

599

600 The highest modeled BC loadings in Russia typically correspond to the large urban areas (i.e. Moscow and St.
601 Petersburg) and less to locations close to the Barents Sea where oil exploration and exploitation sites are located (Fig.
602 10). Koch et al. (2009) found that many models underestimate BC at northern latitudes because of an overestimation of
603 BC scavenging (especially wet deposition) and vertical mixing during poleward transport. Observed high
604 concentrations of BC in April are characteristic for the high aerosol loadings typically found in late winter and early
605 spring, which is qualitatively reproduced by the model (Fig. 10). During this period of the year, conditions are
606 generally favorable for the transport of air pollutants into the Arctic forming the Arctic haze. For example, in April
607 2009 the atmosphere over the European and North American Arctic was often characterized by hazy conditions with
608 transport of air pollutants mainly from Eurasia (Stone et al., 2010). In Fig. 10 we show how concentrations for BC vary
609 in the different months simulated. The high BC concentrations observed in Russia are currently not captured by model
610 suggesting an underestimation of BC emissions in Russia in current emission inventories (Stohl et al., 2013; Eckardt et



611 al., 2015). Warneke et al. (2009) reported a substantial contribution from Russian forest fires and central Asian
612 agricultural burning to atmospheric pollution over the Arctic.

613

614 Long-term measurements of BC in the Arctic are very sparse, but data from measurements performed at Zeppelin
615 station (475 m a.s.l), near Ny-Ålesund are available for comparison with the model results for April and July 2008. The
616 modeled BC is averaged over the 3×3 nearest grid points surrounding the measurement station and calculated from
617 the concentrations in the lowest two model layers, centered around about 30 m above the ground.

618

619 The Zeppelin station observations (Fig. 11a) show 3-5 times higher BC concentrations in April compared to July. This
620 difference is qualitatively reproduced by the model. The vertical distribution of BC (Fig. 11b) shows that
621 concentrations larger than 20 ng/kg are simulated in April up to 200 hPa, whereas there is little to no BC simulated
622 above 500 hPa in July. This is consistent with the idea that BC from source regions outside the Arctic is transported
623 most efficiently into the Arctic in spring, whereas there is little to no long-range transport in summer. The model results
624 suggest that a considerable amount of BC aerosols originating from Europe and East Asia have been lifted into higher
625 altitudes (middle and upper troposphere) and transported over the northern oceans (Fig. 11b). BC aerosols in these
626 high-latitude areas are relatively long-lived due to lower wet scavenging compared to the mid- and low latitudes. The
627 WRF-Chem simulations performed in this study show a mean BC concentration in spring of $0.1 \pm 0.1 \mu\text{g}/\text{kg}$ north of
628 60°N , which is consistent with model simulations done by other authors (Breider et al., 2014 with $0.12 \pm 0.05 \mu\text{g}/\text{kg}$).

629

630 Fig. 11c shows the probability density functions (PDFs) calculated from the observed and modeled BC concentrations
631 at Zeppelin station. The model clearly overestimates the cases with very low BC concentrations in April and in July
632 and does not reproduce the highest observed BC concentrations. In April, the shape of the PDF for values above 0.02
633 $\mu\text{g}/\text{m}^3$ is qualitatively reproduced by the model even though the absolute values are underestimated by a factor of 5.
634 This is, however, not the case in July where the observations and modeled PDF are relatively flat for BC concentrations
635 no larger than $0.015 \mu\text{g}/\text{m}^3$ for the model. This might indicate a missing close-by BC source in the model that could be
636 the reason for the observed larger occurrences of BC up to $0.045 \mu\text{g}/\text{m}^3$ compared to those in the model.

637

638 Measuring the atmospheric concentration of BC is subject to higher uncertainties than, for example, sulfate
639 measurements (e.g. Sharma et al., 2004). Because of this, among other reasons, only few atmospheric BC
640 measurements in the Arctic are currently available. Cavalli et al. (2010) pointed out that the BC atmospheric
641 concentration can vary by a factor 5 depending on the observation protocol; on the model's part, both, errors in
642 emissions and in the deposition parameterizations can strongly affect the results. In our simulation, we used monthly
643 emissions from fossil fuel combustion and daily emissions from large forest fires. This relatively coarse time resolution
644 of the emissions certainly contributes to our simulation biases. In the real world, emissions can be quite variable, with
645 substantial implications for the BC concentration observed throughout the atmosphere. Other complex physical
646 processes are coarsely represented in our model and are expected to also contribute to our model biases.

647

648 **4 Summary, Conclusions and Outlook**



649 In this work, we assessed first the role of the choice of the land surface model (Noah and NoahMP), which is part of
650 the physical parameterization suite of WRF together with the choice of reanalyses used for initialization and to create
651 lateral boundary conditions. Biases in the modeled meteorology (e.g. wind speed and direction, boundary layer height)
652 are expected to influence the modeled BC concentrations. As the modeled meteorology strongly depends on the model
653 configuration used, the impact of the choice of data used for initialization and lateral boundary conditions and the
654 choice of LSM scheme was systematically investigated in several model experiments. The results showed that no
655 particular model configuration clearly outperforms the other model configurations when compared to available Arctic
656 observations. Usage of the higher resolution ASR data or a more sophisticated formulation of land surface processes
657 (NoahMP) did not systematically improve model performance in the Arctic. The results suggest that there are other
658 factors to consider such as, for example, the accurate simulation of the Arctic sea ice decline, stratosphere-troposphere
659 interactions, atmosphere-ocean interaction, and boundary layer processes. Improvements of surface albedo, for
660 example, could be achieved by taking more advantage of continuous and reliable satellite and ground-based data for
661 calibration and comparison purposes. This study can therefore also be seen as a starting point for further, systematic
662 sensitivity studies to assess the impact of individual factors such as meteorological model performance, model
663 resolution, boundary conditions, soil characteristics, and the various parameterizations within the physical schemes on
664 the model results for air pollutants in the Arctic.

665

666 This study is a contribution to ongoing work aiming to evaluate the capability of WRF-Chem to simulate BC around
667 the Eurasian Arctic. While the modeled BC concentrations are qualitatively consistent with observations in terms of
668 spatial distribution and time series, the modeled concentrations of BC are too low in the Russian Arctic as noted by
669 (Eckardt *et al.*, 2015). Several factors are likely to influence the modeled BC concentration besides biases in modeled
670 meteorology. For example, Kuik *et al.* (2015) and Eckard (2015) found that underestimations in modeled BC
671 concentration in WRF-Chem are also related to the low quality of the emissions data. This has important implications
672 for the projected changes in Arctic climate when model natural dynamics are interacting with anthropogenic emissions.
673 Reliable emission inventories with a high temporal and spatial resolution are critical to improve the modeling of
674 aerosols and air chemistry. Especially for this region, model simulations are an important tool to assess the
675 contributions of individual source categories to the total BC concentrations (such as energy or industrial origin). Also,
676 for deepening the analysis of the different impacts of anthropogenic BC it is crucial to have good observational data,
677 e.g. for BC and particulate matter, vertical profiles of temperature and humidity. Other problems leading to systematic
678 errors include those already found in the model. For example, Marelle, *et al.* (2016) found that a bug in older versions
679 of the Noah LSM scheme that has been fixed in the current version used for this study (WRF3.7) in combination with
680 an underestimation lack of aerosol wet removal by subgrid-scale clouds, may be one of the main factors contributing to
681 the simulated low BC concentrations over areas dominated by a sea-ice/ atmosphere interface.

682

683 **5 Data availability**

684 The two external model data sets used in this work, ERA-Interim is available from <http://apps.ecmwf.int/> (Dee *et al.*,
685 2011) and Arctic System Reanalysis (ASR) available from <http://rda.ucar.edu/> (Bromwich *et al.*, 2016). The Met
686 Office Integrated Data Archive System (MIDAS) is available at the British Atmospheric Data Centre,



687 <http://catalogue.ceda.ac.uk/>. The radiosonde data from the Integrated Global Radiosonde Archive (IGRA) are available
688 at <ftp.ncdc.noaa.gov/pub/data/igra/> (Durre, et al 2006).

689

690 6 Acknowledgements

691 All model simulations used in this study were performed at at the high performance cluster computer of the German
692 Climate Computing Center (DKRZ). For data processing and figures we used the open-source software NCAR
693 Command Language (NCL) also available at <https://www.ncl.ucar.edu/>.

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1004 List of Tables

1005 Table 1. Model configuration and sensitivity experiments.

a) Model physical set up	
Model	WRF (ARW) version 3.7.1
Simulation period	Years 2008-2012 Winter: January, February, March (JFM) Summer: July, August, September (JAS)
Domain	270 X 252 grid points (175°W-175°E and 45°N-90°N)
Projection	Polar stereographic, center at 70°N, 40°E
Horizontal Resolution	15 km x 15 km
Vertical levels	30 σ -levels (11 levels below $\sigma = 0.9$)
Physical configuration^a:	
<i>Shortwave radiation:</i>	Goddard scheme
<i>Longwave radiation:</i>	RRTM scheme
<i>Cloud microphysics:</i>	Purdue Lin scheme
<i>Cumulus scheme:</i>	Grell-3D ensemble scheme
<i>PBL scheme:</i>	Yonsei University Scheme (YSU)
<i>Surface Layer:</i>	MM5 Similarity
<i>Land Surface Model (LSM):</i>	* Unified Noah * NoahMP
Data for initialization and lateral boundary conditions:	* ERA-Interim * Arctic System Reanalysis (ASR)
b) Experiments and Description:	
<i>Noah_ecmwf</i>	Noah LSM + ERA-Interim reanalysis
<i>NoahMP_ecmwf</i>	NoahMP LSM + ERA-Interim reanalysis
<i>Noah_asr</i>	Noah LSM + ASR reanalysis
<i>NoahMP_asr</i>	NoahMP LSM + ASR reanalysis

1006 ^a See text for references

1007 * Sensitivity analyses

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1010 **Table 2. Model set up for the WRF-Chem runs with chemistry and aerosols (physical parameterizations and**
 1011 **domain settings as for the meteorology-only WRF runs given in Table 1).**

Model	WRF-Chem version 3.7.1
Simulation period	April 2008 July 2008 April 2009 July 2010
Chemistry	RADM2+CMAQ scheme
Aerosols	MADE/SORGAM scheme
Dust	GOCART on-line dust emissions (AFWA modification)
Input data	
Boundary conditions chemistry	MOZART (Global CTM)
Anthropogenic emissions	EDGAR HTAP v2.2 (resolution 0.1°x0.1°)
Fire emissions	NCAR FINN v1.0
Biogenic emissions	MEGAN (online)
Time profiles	Yearly cycle (monthly data)

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1019 **Table 3. List of stations used for model evaluation (in bold, stations located north of 66°N).**

Station	Code	Latitude	Longitude	WMO
1. Krenkelja, RU *	KRE	80.64 °N	58.049 °E	20046
2. Ostrov Golomjannyj RU	GOL	79.55 °N	90.617 °E	N/A
3. Ny-Alesund, NO	ALE	78.92 °N	11.93 °E	01004
4. Danmarkshavn GRE	DAN	76.76 °N	18.66 °W	04320
5. Ostrov Dikson, RU	DIK	73.50 °N	80.42 °E	20674
6. Popova RU *	POP	73.33 °N	70.05 °E	N/A
7. Malye Karmakuly, RU	MAL	72.38 °N	52.73 °E	20744
8. Murmansk, RU	MUR	68.98 °N	33.12 °E	22113
9. Sojna RU	SOJ	67.87 °N	44.17 °E	22271
10. Narian Mar, RU	NAR	67.65 °N	53.02 °E	23205
11. Bodo, NO	BOD	67.26 °N	14.366 °E	01152
12. Kandalaksa, RU	KAN	67.15 °N	32.35 °E	22217
13. Sale-Khard , RU	SAL	66.53 °N	66.66 °E	23330
14. Turukhansk, RU	TUR	65.78 °N	87.95 °E	23472
15. Lulea-Kallax, SWE	LUL	65.55 °N	22.13 °E	02185
16. Arhangel'Sk, RU	ARH	64.58 °N	40.5 °E	22550
17. Orland, NO	ORL	63.70 °N	9.60 °E	01241
18. Timra, SWE	TIM	62.51 °N	17.45 °E	02365
19. Syktyvkar, RU	SYK	61.66 °N	50.85 °E	23804
20. Samarovo, RU	SAM	60.97 °N	69.07 °E	23933
21. Jokioinen, FIN	JOK	60.82 °N	23.50 °E	02963
22. Ivdel, RU	IVD	60.68°N	60.43 °E	23921
23. Leningrad, RU	LEN	59.95 °N	30.70 °E	26063
24. Moscow, RU	MOC	55.75 °N	37.57 °E	27612

* No radiosonde data for analyzed period

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Table 4. Biases and RMS error of 6-hourly values from the ERA-Interim and ASR reanalyses and the 4 WRF experiments compared against surface meteorological observations from AWS.

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	<i>Stations above 66°N stations</i>				<i>Stations below 66°N</i>			
Temperature (°C)	<i>FMA</i>		<i>JAS</i>		<i>FMA</i>		<i>JAS</i>	
	Bias	RMS	Bias	RMS	Bias	RMS	Bias	RMS
Reanalysis								
<i>ERA-Interim</i>	0,40	2,40	-0,25	1,81	-0,12	1,78	-0,10	1,81
<i>ASR</i>	0,18	1,53	-0,25	1,06	-0,28	1,18	-0,19	0,85
WRF Experiments								
<i>noah_ecmwf</i>	-0,70	6,08	-1,74	3,15	-3,21	5,80	-1,64	2,94
<i>noahMP_ecmwf</i>	0,53	6,19	-1,57	3,11	-1,23	5,62	-1,03	2,99
<i>noah_asr</i>	0,83	6,79	-0,87	3,09	-1,77	5,75	-0,99	3,20
<i>noahMP_asr</i>	1,99	6,96	-0,84	3,05	-0,01	5,88	-0,59	3,15
Wind Speed (m/s)								
	<i>FMA</i>		<i>JAS</i>		<i>FMA</i>		<i>JAS</i>	
	Bias	RMS	Bias	RMS	Bias	RMS	Bias	RMS
Reanalysis								
<i>ERA-Interim</i>	0,81	2,57	0,38	1,89	0,78	2,01	0,90	1,85
<i>ASR</i>	-0,17	1,99	0,02	1,43	-0,32	1,63	-0,15	1,38
WRF Experiments								
<i>noah_ecmwf</i>	0,74	3,31	0,63	2,78	0,84	2,42	0,94	2,19
<i>noahMP_ecmwf</i>	0,55	3,26	0,38	2,58	0,61	2,23	0,61	1,99
<i>noah_asr</i>	0,89	3,75	0,67	2,91	0,79	2,54	0,93	2,22
<i>noahMP_asr</i>	0,70	3,66	0,50	2,83	0,62	2,41	0,61	2,04
SLP (millibars)								
	<i>FMA</i>		<i>JAS</i>		<i>FMA</i>		<i>JAS</i>	
	Bias	RMS	Bias	RMS	Bias	RMS	Bias	RMS
Reanalysis								
<i>ERA-Interim</i>	-0,06	0,56	-0,01	0,49	-0,03	0,46	-0,09	0,41
<i>ASR</i>	-0,59	0,63	-0,24	0,52	-0,43	0,60	-0,08	0,42
WRF Experiments								
<i>noah_ecmwf</i>	0,28	5,79	-0,19	4,94	0,57	4,65	-0,39	3,38
<i>noahMP_ecmwf</i>	0,12	5,74	-0,11	4,20	-0,03	4,56	-0,54	2,93
<i>noah_asr</i>	-2,69	7,80	-1,45	5,57	-0,99	5,82	-0,59	3,73
<i>noahMP_asr</i>	-2,93	7,89	-1,17	5,51	-1,68	5,64	-0,37	3,43

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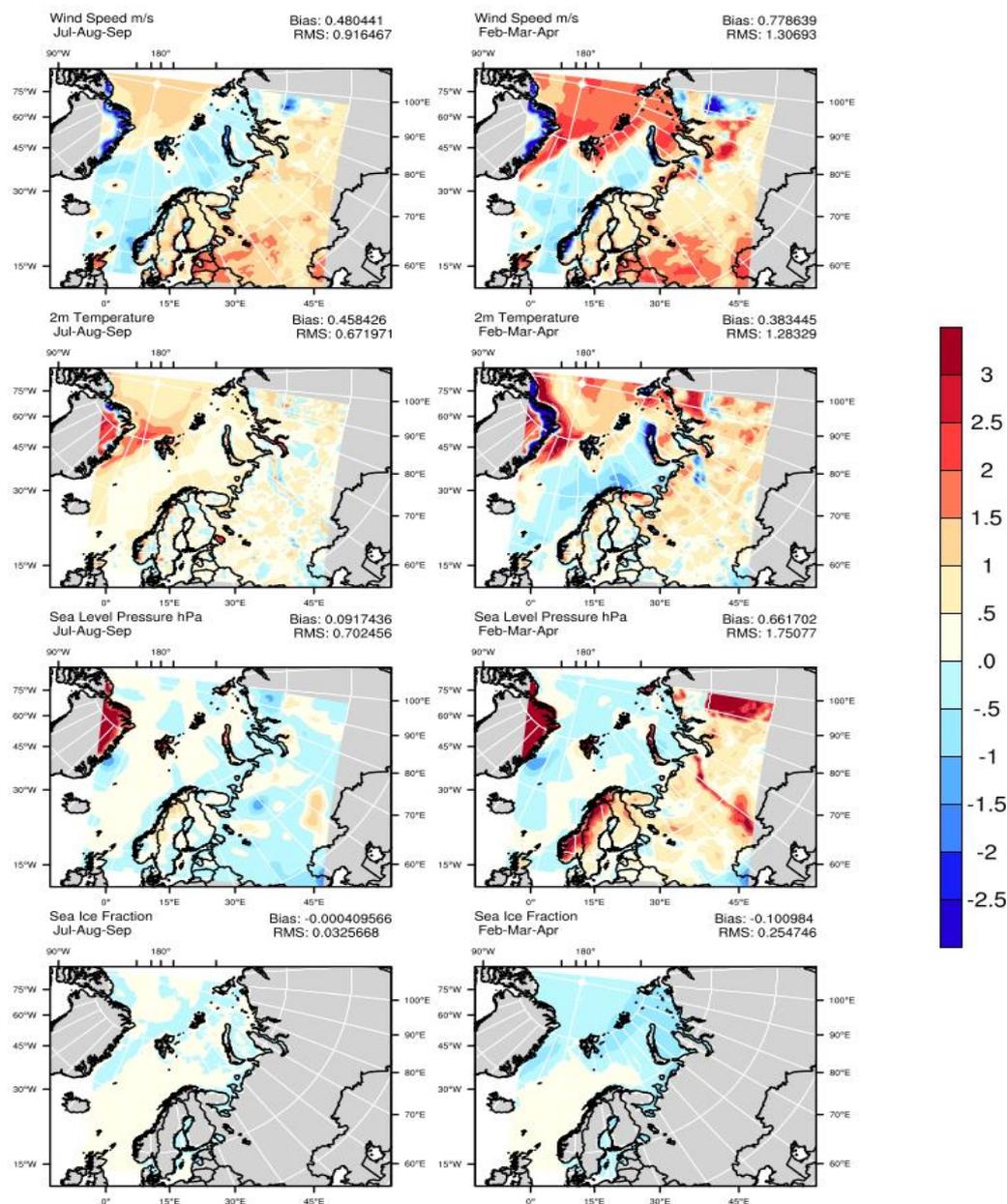
1037 **Table 5. Mean vertical profiles statistics: biases and RMS for the ERA-Interim and ASR reanalyses and the 4**
 1038 **WRF experiments against radiosonde observations from the IGRA database.**

	Stations above 66°N				Stations below 66°N			
Temperature (°C)	FMA		JAS		FMA		JAS	
	Bias	RMS	Bias	RMS	Bias	RMS	Bias	RMS
Reanalysis								
<i>ERA-Interim</i>	1,15	1,50	0,42	0,58	1,73	2,10	0,33	0,66
<i>ASR</i>	0,46	0,73	-0,58	1,49	0,73	0,97	-1,58	2,47
WRF Experiments								
<i>noah_ecmwf</i>	0,26	0,92	0,27	0,97	0,06	1,19	0,27	0,83
<i>noah_asr</i>	0,51	1,43	0,43	1,21	0,82	1,35	0,58	0,91
<i>noahMP_asr</i>	1,07	2,04	0,43	1,24	1,60	1,91	0,78	1,06
<i>noahMP_ecmwf</i>	0,89	1,33	0,40	1,01	0,97	1,34	0,54	0,89
Wind Speed (m/s)	FMA		JAS		FMA		JAS	
	Bias	RMS	Bias	RMS	Bias	RMS	Bias	RMS
Reanalysis								
<i>ERA-Interim</i>	-0,67	1,00	-0,42	0,67	-0,34	0,92	-0,36	0,73
<i>ASR</i>	-0,45	0,82	0,00	0,71	-0,26	0,645	-0,10	0,84
WRF Experiments								
<i>noah_ecmwf</i>	0,97	1,55	0,52	1,01	0,89	1,388	0,40	1,04
<i>noah_asr</i>	1,34	1,94	0,55	1,11	0,93	1,493	0,38	1,10
<i>noahMP_asr</i>	1,18	1,82	0,53	1,03	0,68	1,351	0,19	1,04
<i>noahMP_ecmwf</i>	0,85	1,47	0,33	0,86	0,47	1,176	0,31	1,01
% Humidity	FMA		JAS		FMA		JAS	
	Bias	RMS	Bias	RMS	Bias	RMS	Bias	RMS
Reanalysis								
<i>ERA-Interim</i>	7,05	8,94	2,24	4,41	5,71	8,07	1,76	4,06
<i>ASR</i>	-2,15	4,15	1,22	3,78	-0,64	3,28	2,75	5,38
WRF Experiments								
<i>noah_ecmwf</i>	-3,92	7,43	-0,22	5,97	0,20	7,13	0,07	6,77
<i>noah_asr</i>	-1,54	8,00	1,69	6,56	0,08	6,96	1,1	5,92
<i>noahMP_asr</i>	-3,58	8,93	1,84	6,78	-3,24	6,64	-0,4	5,73
<i>noahMP_ecmwf</i>	-5,96	8,38	-0,77	5,84	-3,56	6,70	-1,63	5,99

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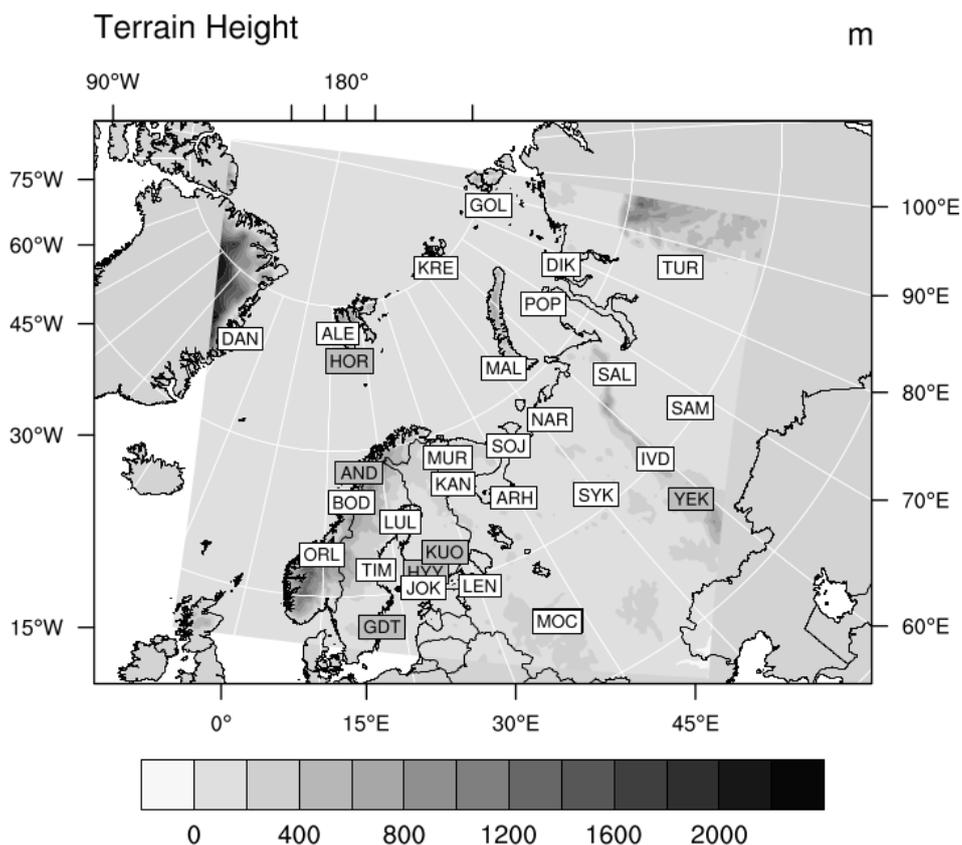
1041 List of Figures



1042

1043 Figure 1. Differences between the ERA-Interim and the ASR reanalyses for a) 10m wind speed in m/s, b) 2m
1044 2m temperature in °C, c) sea level pressure in hPa, and d) sea ice in fractional units for summer (JAS; left column)
1045 and winter (FMA; right column) averaged over five seasons (2008-2012). The root mean square (RMS) error
1046 and mean bias are given above each panel as a measure of the differences between the two reanalyses.

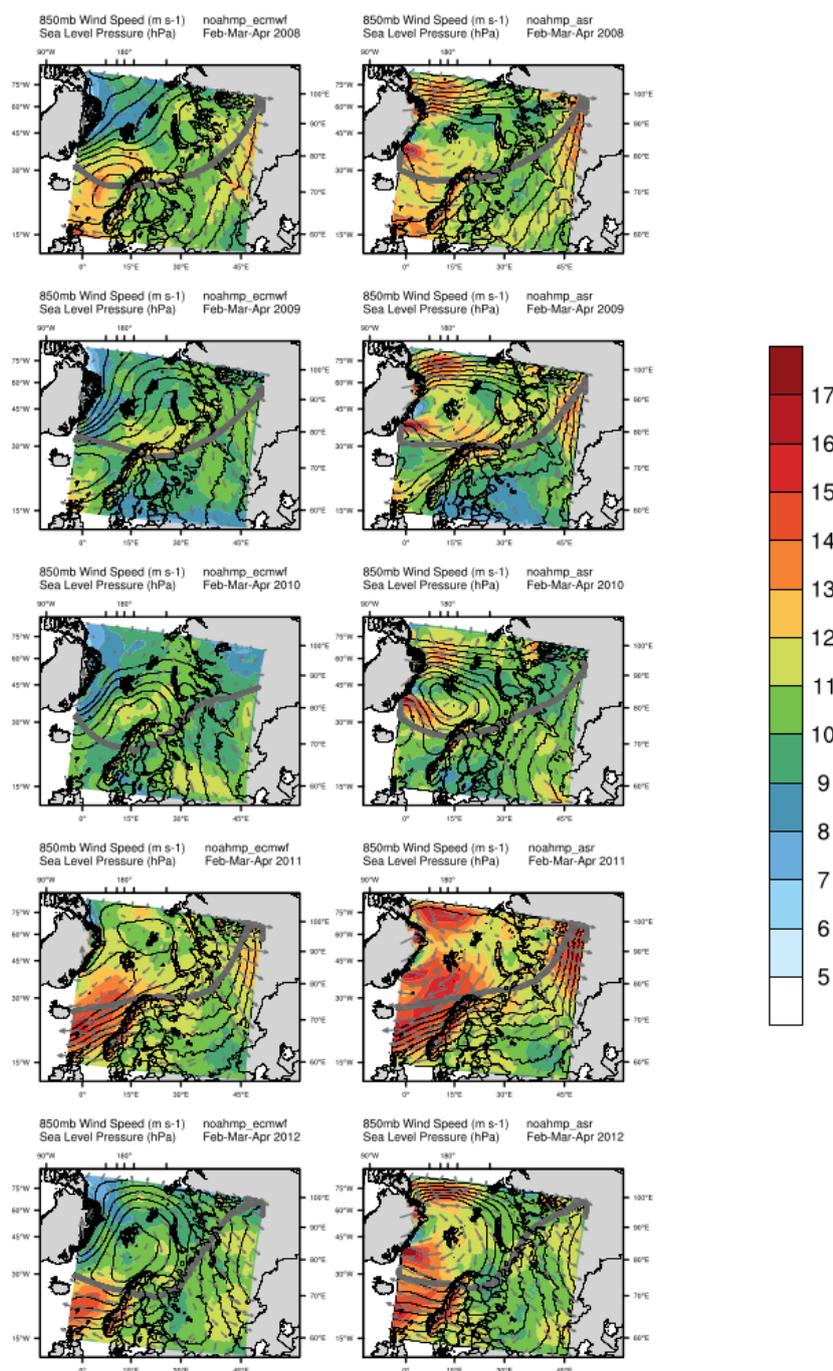
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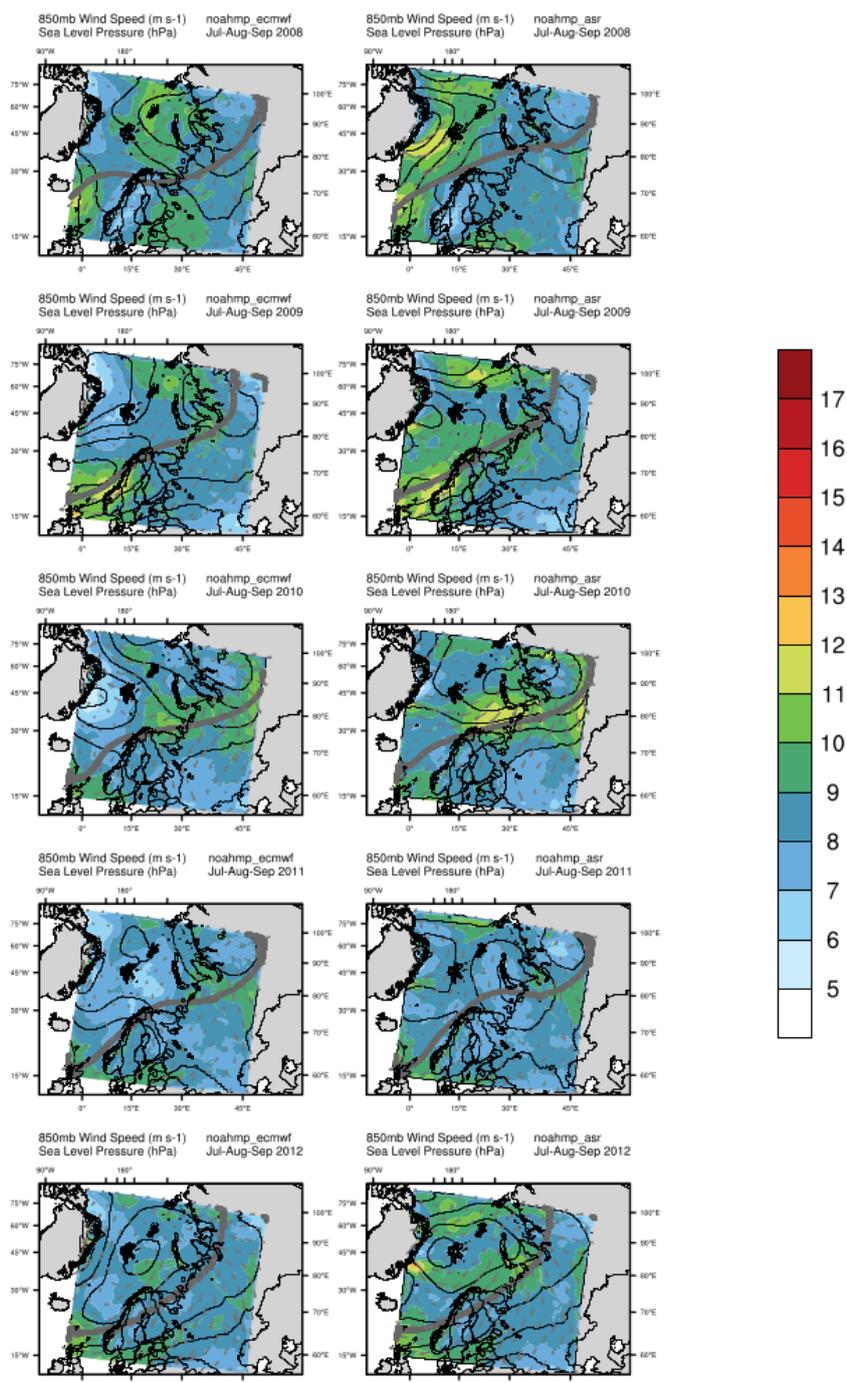
1049 **Figure 2. Model domain, topographic height (m), and location of the observational stations from WMO (white**
1050 **labels) and AERONET (grey labels) used for the model evaluation**

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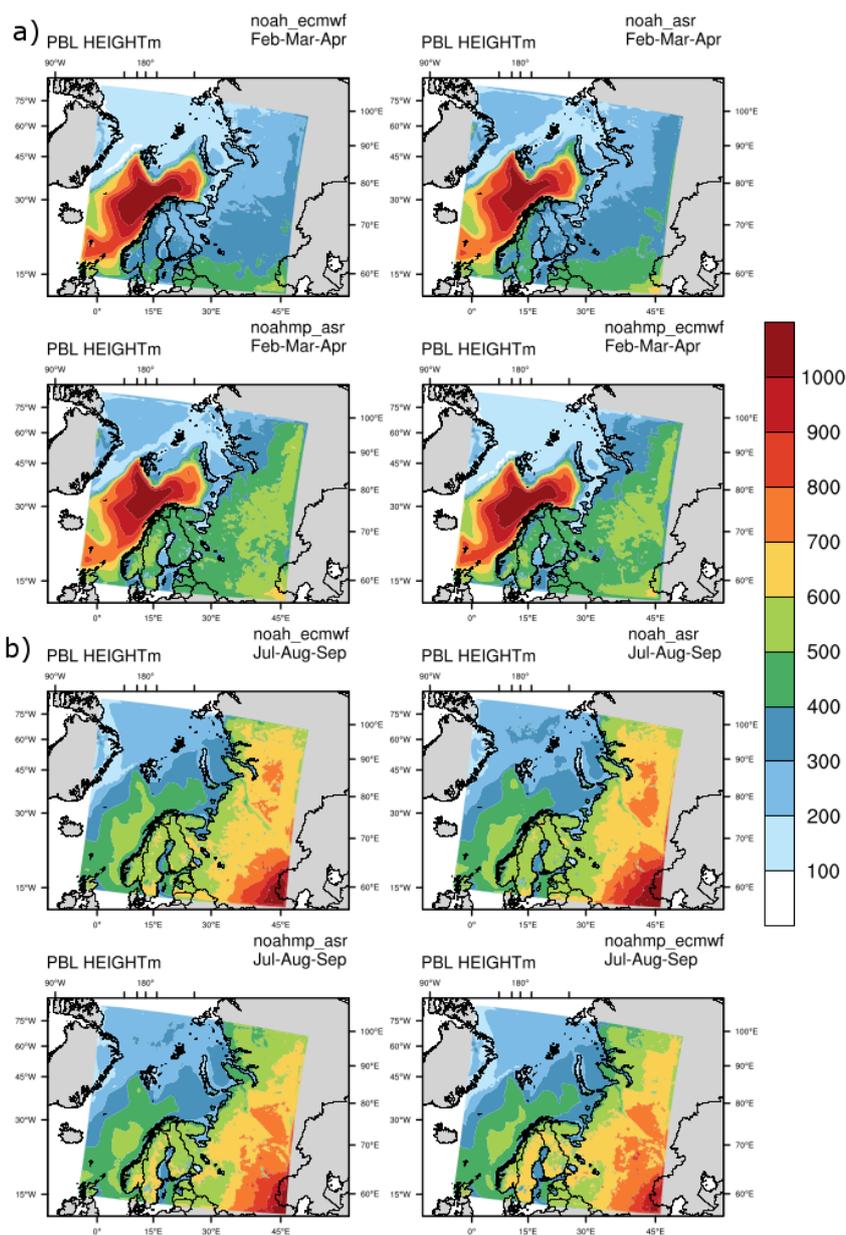
1053 **Figure 3.** Modeled 3-month means of the sea level pressure and 850 hPa winds for winter months (February,
 1054 March, April) from 2008-2012. Left column represents experiment *noahmp_ecmwf* and right column for
 1055 experiment *noahmp_asr*. Solid lines are isobars (contours from 1004-1020 hPa), color shading is wind speed at
 1056 850 hPa in ms^{-1} and gray arrows represent wind direction. The maximum gradient in geopotential height at 500
 1057 hPa (thick gray contour line) is used here as an approximation of the Arctic dome.



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1059 **Figure 4.** Same as Figure 3 but for summer months (July, August, September).

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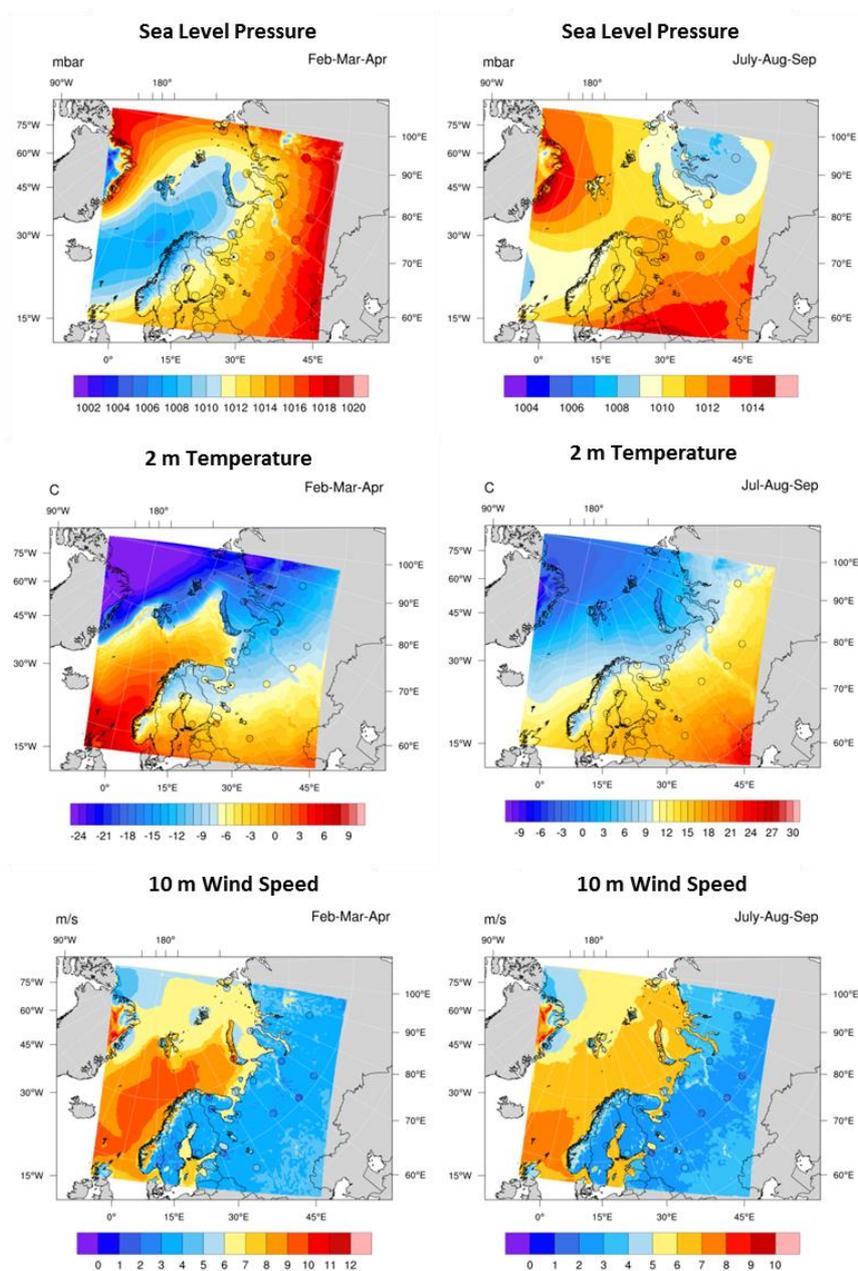


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1062 **Figure 5. Five year averages of the planetary boundary layer height for a) winter months (February, March and**
1063 **April) and for b) summer months (July, August and September) from the four meteorology-only WRF**
1064 **experiments (*noah_ecmwf*, *noah_asr*, *noahmp_ecmwf*, and *noahmp_asr*).**

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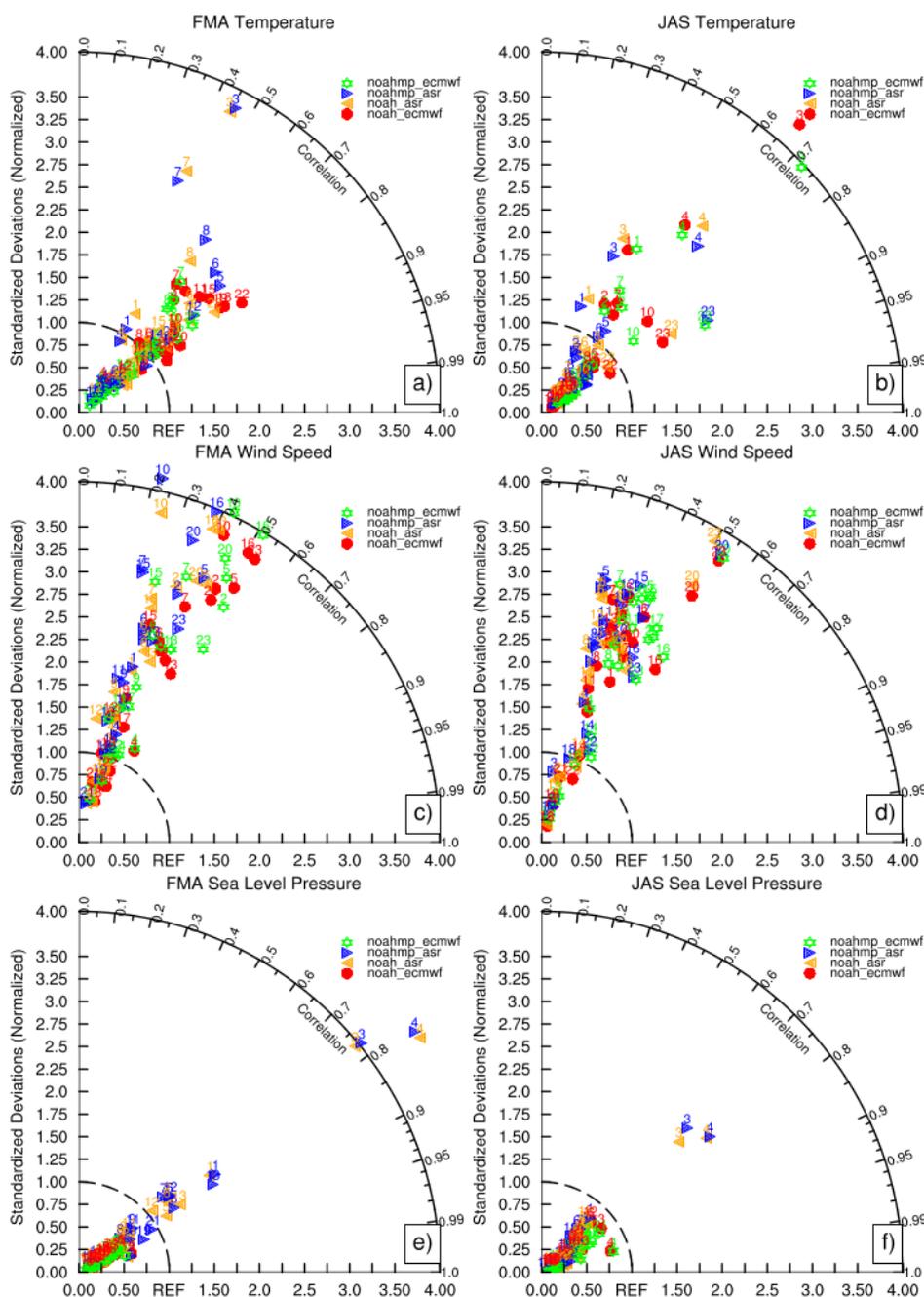
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1068 **Figure 6. Mean modeled spatial distribution (colored fields) and station observations (colored circles) of sea level**
1069 **pressure (hPa), 2m temperature (°C) and 10m wind speed (m/s) from from the meteorological experiment**
1070 **noahmp_ecmwf for winter (left column) and summer months (right column).**

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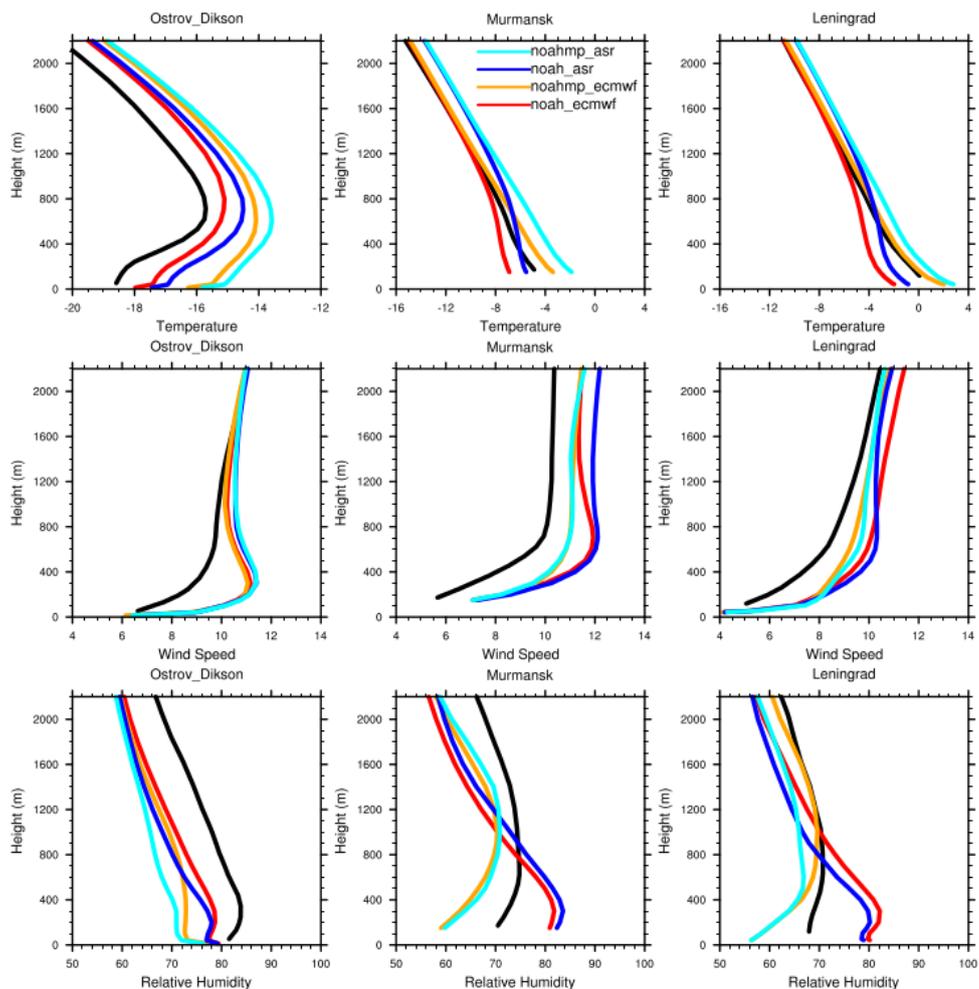
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Figure 7. Taylor diagrams comparing surface observations from 22 stations with the four meteorology-only model experiments calculated from time series of daily means over the five year period 2008–2012 for the winter months (FMA; left column) and summer (JAS; right column) for 2m temperature (a–b), 10m wind speed (c–d) and sea level pressure (e–f).

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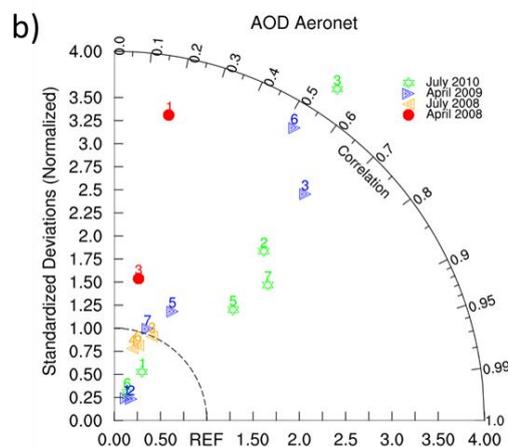
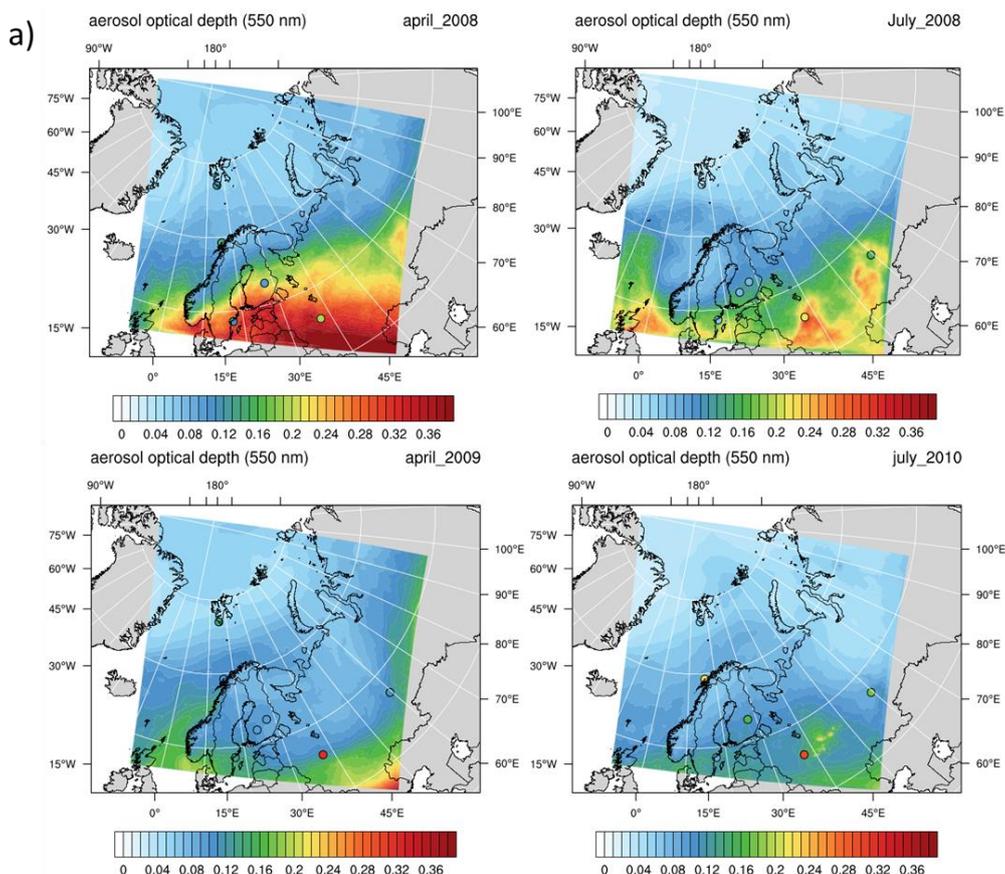
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1080 **Figure 8. Radiosonde and modeled mean profiles at midday readings from the 4 meteorology-only WRF model**
1081 **runs compared with radiosonde measurements (black line) at three selected stations for winter months**
1082 **(Febrero, March and April) for temperature in °C (first row) wind speed in m/s (middle row) and relative**
1083 **humidity in % (last row).**

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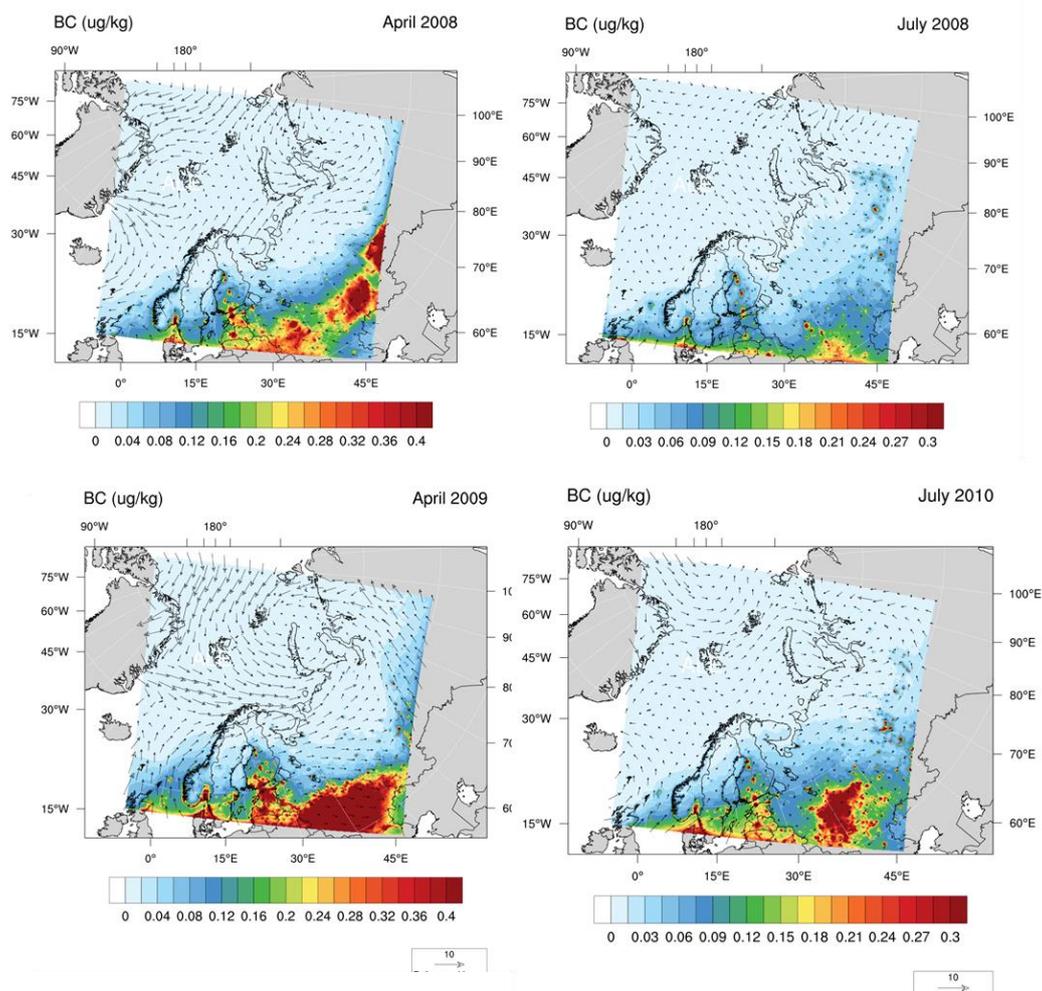
Aeronet stations:

- 1 Gustav_Dalen_Tower
- 2 Kuopio
- 3 Moscow
- 4 Hornsund
- 5 Andenes
- 6 Yekaterinburg
- 7 Hyttiala

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Figure 9. Comparison of the modeled mean AOD from the experiment with WRF-Chem and AOD from Aeronet observations, a) spatial distribution (colored fields) and AERONET mean observations (colored circles) and b) Taylor diagrams displaying a comparison between timeseries of observational AOD at different Aeronet stations versus modeled AOD.

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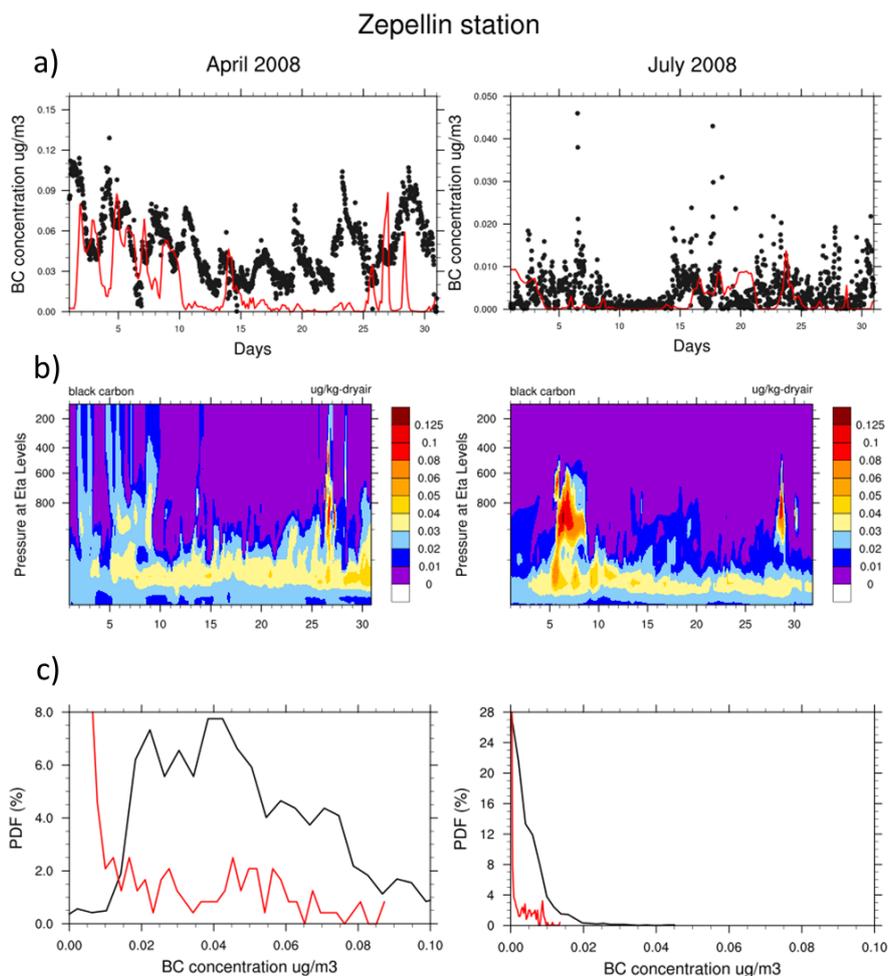


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1095 **Figure 10. Simulated monthly mean BC concentrations near the surface (colored fields) and 10m wind (arrows)**
1096 **from WRF-Chem for April 2008, July 2008, April 2009 and July 2010.**

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1100 **Figure 11.** Time series of April and July 2008 a) modeled (red line) and observed (black dots) surface BC
1101 concentrations at Zepellin station in Ny-Ålesund, b) vertical distribution of modeled BC and c) probability
1102 density function of BC calculated from hourly values of BC concentrations for the studied periods for the model
1103 (red line) and observations (black line).

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