

1 **Unstructured-grid model for the North Sea and Baltic Sea: validation against observations**

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24

25 **Abstract**

26 A new unstructured-grid model and its application to the North Sea and Baltic Sea are described.  
27 The research focus is on the dynamics in the two basins and in the multiple straits connecting  
28 them and more specifically on how the model replicates the temporal and spatial variability of  
29 physical processes. The comparison against observed data indicates the realism in the  
30 simulations of the exchange flows. The simulations demonstrated that in contrast to the tidal  
31 variability which decreases in the strait, the role of the barotropic forcing due to weather systems  
32 increases. In this zone reversal of transport is well manifested by the increased difference  
33 between the surface and bottom salinity values. Small sub-basins like Arkona and Bornholm play  
34 the role of reservoirs for denser water which under specific conditions cascades on its way to the  
35 Gotland Deep. Unlike the intermediate and deep water salinity in the Baltic Sea, which is  
36 strongly affected by fluxes in the straits, the simulated winter-refill and evolution of cold  
37 intermediate water are rather driven by surface cooling and processes in the upper mixed layer.

38

39 **Key words:** SCHISM; ocean circulation; strait processes; temperature; salinity/density; North  
40 Sea; Baltic Sea

41

42 **1. Introduction**

43 The North Sea and Baltic Sea (Fig.1) represent coupled tidal and non-tidal basins of  
44 approximately equal size connected through a system of straits. This straits system (hereafter the  
45 North Sea - Baltic Sea Transition Zone (NBTZ)) includes the Kattegat, Danish straits and the  
46 western part of the Baltic Sea. The surplus of fresh water in the Baltic Sea and the limited water  
47 exchange with the North Sea support a two-layer exchange flow in the NBTZ and explains the  
48 low salinity (brackish water) of the Baltic Sea. The deep part of the Baltic Sea known as the  
49 Baltic proper is separated from the straits area by a number of sills: the Darss Sill (depth of  
50 18m), the Drogden Sill (depth of 8m) and the Stolpe Channel (depth of 63-64 m) further east. In  
51 these shallow areas the inflowing North Sea water is subjected to substantial mixing with the  
52 highly stratified Baltic Sea water.

53 The North Sea - Baltic Sea system represents a challenge for the numerical modelling because of  
54 several reasons. The first one is that it consists of a tidal and non-tidal basin with different  
55 dominating balances in each individual basin. The tidal one (the North Sea) is very shallow,  
56 except for the Norwegian trench. The relatively high salinity values in the North Sea are typical  
57 for the ocean; the vertical mixing which is mostly due to tides dominates the hydrophysical  
58 fields. The second basin (the Baltic Sea) can be considered as a huge estuary with extremely  
59 strong vertical stratification, which inhibits the vertical mixing. The diffusion coefficients  
60 approach to their molecular values, resulting in the extremely weak mixing between the surface  
61 and deep layers. This specific vertical stratification is maintained by two major factors: (1) the  
62 river runoff and precipitation-evaporation balance at sea surface, and (2) periodic intrusions of  
63 saltier North Sea water triggered by extreme atmospheric conditions with an approximately  
64 decadal periodicity.

65 The second challenge in the modelling of Baltic Sea and the area of Skagerrak and Norwegian  
66 Trench stems from the fact that the processes there are strongly dependent upon the exchanges in  
67 the NBTZ. Compared to other similar transition zones (e. g. Mediterranean – Atlantic Ocean or  
68 Black Sea – Mediterranean; Sannino et al., 2009, Stanev and Lu, 2013), the NBTZ is more  
69 complex because of the presence of multiple straits. Under different weather or circulation  
70 conditions their individual contribution to the exchange between the basins varies (Stanev et al.,  
71 2015). Thus the partitioning of flows and recirculation in the belts (which includes 3 major  
72 waterways: Great Belt, Little Belt, and Oresund) are central to the problem of the ventilation of  
73 deep Baltic basins by the North Sea water (see also, Meier, 2005; 2007). Extremely high  
74 resolution is needed there in order to resolve the complex coastal line as well as the bottom  
75 topography found in the three straits (note that the Little Belt is only 1 km wide). Without an  
76 adequate resolution the hydrodynamics of the inflow-outflow system cannot be accurately  
77 simulated.

78 A number of numerical models based on primitive equations and finite-difference discretization  
79 have been used to simulate the circulation in the coupled North Sea – Baltic Sea; e.g., models of  
80 Funkquist and Kleine (2007), She et al. (2007), Fu et al. (2011; 2012), Zhuang et al. (2011).  
81 More detailed references are given in Meier and Kauker (2003), Lehmann et al. (2004), Schmidt  
82 et al. (2008), Leppäranta and Myberg (2009). However the horizontal resolution in these models  
83 did not allow sufficient resolution in the three straits. Burchard et al. (2005; 2009) used a

84 horizontal resolution of 0.5 nm to study the dominant dynamics in the Western Baltic Sea and  
85 validated the model performance with a focus on the mixing in the areas of Drogden Sill, Darss  
86 Sill and the Bornholm Channel. Although this resolution was not sufficient for the Sound and  
87 inadequate for the Little Belt, the idealized simulations of the authors allowed determination of  
88 the pathways of salt transport during medium-intensity inflow events (see also Meier, 2007),  
89 demonstrating a reasonable consistency with the observations. However in these publications  
90 some problems with the open boundary conditions at the Kattegat, and also with the initialization  
91 strategy have not been resolved; e.g., forced with the climatological condition, the model did not  
92 fully recover this condition in an annual simulation (Meier, 2007).

93 Motivated by the above challenges we (1) address the resolution problem by enabling sufficient  
94 resolution in the straits, and (2) avoid inconsistencies in some earlier studies. Some of these  
95 inconsistencies are associated with either the forcing being prescribed in the NBTZ, or with the  
96 one-way or two-way nesting techniques, which are not seamless. Therefore we describe in the  
97 present paper an application of an unstructured-grid model for the coupled basins starting from  
98 the English Channel in the South up to the Shetland Islands to the North. This configuration  
99 enables the seamless propagation of the large-scale forcing into the NBTZ (see Danilov, 2013 for  
100 a review of recent developments and practices on using unstructured meshes in ocean modelling  
101 and for more references). Unlike other applications of unstructured models for this region (e.g.  
102 Kliem et al. 2006), we use a 3D baroclinic set up. This presents a third challenge because it has  
103 never been shown before how unstructured-grid models can adequately simulate the complex  
104 thermohaline structure of two basins. If the adequacy is demonstrated, the model could be used  
105 also for other similar ocean areas.

106 It's our hope that our research will shed new light on the pros and cons of the current model vs  
107 other more traditional models (including both structured- and unstructured-grid models). The  
108 major differences include: implicit time step (which avoids splitting errors and enables efficiency  
109 and robustness), and treatment of momentum advection with Eulerian-Lagrangian Method  
110 (ELM, which further boosts efficiency and robustness) (although central-difference scheme has  
111 also been implemented). While the use of Galerkin Finite Element Method (GFEM) is not new,  
112 the combination of it with the previous two features seems to have achieved a good balance in  
113 terms of numerical diffusion and dispersion, as the numerical diffusion inherent in an implicit  
114 method and ELM is balanced out by the numerical dispersion inherent in GFEM. This is very

115 different from other earlier finite-volume models such as ELCIRC (Baptista et al. 2005) where  
116 numerical diffusion is dominant (another major advantage in this regard is SCHISM's ability to  
117 handle very skew elements and be completely free of orthogonality constraint). As a result, the  
118 model can be effectively used to simulate cross-scale processes with both accuracy and  
119 efficiency. However, as an unstructured-grid model, it's not immune to some common issues  
120 such as sensitivity to grid generation. The latter does not have an easy answer as the quality of  
121 model results is clearly tied to the grid used, and while there are some generic guidelines about  
122 mesh quality, ultimately the issue is application dependent. We are in the process of carefully  
123 testing and documenting this issue for a variety of barotropic and baroclinic applications  
124 including baroclinic instability (eddies and meanders), and comparing our results with those  
125 from structured-grid models (Zhang et al. submitted).

126 The research questions and novelties can be briefly summarized as follows.

- 127 1. Describe a new model and its application to a very specific region, which is  
128 dominated by tides and shelf processes, baroclinic processes driven by fresh water  
129 fluxes in the Baltic Sea and very specific (shallow) transition zone. Addressing all  
130 these needs good quality of simulation of both barotropic and baroclinic processes.
- 131 2. Quantify how well the model replicates the temporal and spatial variability of  
132 physical processes in the studied area.
- 133 3. Make available a reproducible reference set-up for this model to be used in further  
134 studies.

135 While deeper analysis of processes will be addressed elsewhere, this paper is focused on  
136 validating the new model. In section 2 we describe the model used. Section 3 describes the  
137 dynamics of sea level and the inter-comparisons with observations. Section 4 addresses the  
138 simulations of thermohaline fields, and Section 5 describes the quality of simulations of water  
139 mass structure. Short conclusions are formulated at the end. Because the processes in the two  
140 basins differ greatly, the validation presented here is not fully symmetric for the two basins; the  
141 baroclinic part of the Baltic Sea is presented in more detail along with the issue of water mass  
142 structure.

143

144 2. **The model**

145 **2.1 Model description**

146 SCHISM (Semi-implicit Cross-scale Hydroscience Integrated System Model; Zhang et al.  
147 submitted) is a derivative product of the original SELFE model (Zhang and Baptista 2008), with  
148 many improvements implemented by the first author at College of William & Mary and  
149 collaborators. It solves Reynolds-averaged Navier-Stokes (RANS) equation with transport of  
150 heat, salt and tracers in the hydrostatic form with Boussinesq approximation, using unstructured  
151 grids. Due to its highly flexible framework the model has found a wide range of cross-scale  
152 applications worldwide, from creeks to deep oceans: general circulation (Zhang et al. 2015),  
153 storm surges (Bertin et al. 2014), tsunami hazards (Zhang et al. 2011), water quality (Wang et al.  
154 2013), oil spill (Azevedo et al. 2014), sediment transport (Pinto et al. 2012), and  
155 biogeochemistry (Rodrigues et al. 2009). The model is being distributed as an open-source  
156 community-supported model under an Apache license (<http://www.schism.wiki>).

157 New development since the last writing (Zhang et al. 2015) includes: addition of mixed triangle-  
158 quadrangles grid, and 1D/2D/3D options all wrapped in a single model grid. These new additions  
159 greatly extend the capability of SCHISM and have been reported in Zhang et al. (submitted).

160 Currently no ice model is available within the SCHISM system, and this may explain some  
161 errors near the Bay of Bothnia.

162 **2.2 Description of the numerics**

163 SCHISM solves the hydrostatic RANS and transport equations using a hybrid Finite Element and  
164 Finite Volume approach grounded on unstructured grids in the horizontal dimension. The  
165 efficiency and robustness of SCHISM are mostly attributed to the implicit treatment of all terms  
166 that place stringent stability constraints (e.g. CFL) and the use of Eulerian-Lagrangian method  
167 for the momentum advection. The vertical grid has recently been extended from the original SZ  
168 (i.e. partially terrain-following S- and partial Z-coordinates) to a highly flexible LSC<sup>2</sup> (Localized  
169 Sigma Coordinates with Shaved Cell) grid. Zhang et al. (2015) demonstrated the superior  
170 performance of LSC<sup>2</sup> in cross-scale applications. LSC<sup>2</sup> is also applied in the current study and  
171 we'll demonstrate its capability to maintain sharp stratification in the Baltic Sea (cf. Fig. A1).

172

### 173 **2.3 The model grid**

174 Fig. 2a shows the grid size distribution of the unstructured grid covering North Sea & Baltic Sea.  
175 Altogether there are ~300K nodes and ~600K triangles, with some refinement along the German  
176 Bight and Danish straits, where a nominal resolution of 200m is used (Fig. 2b), with the  
177 minimum grid size of 60m found in the narrow Little Belt. Some comparison studies were  
178 conducted using two types of vertical grids supported by SCHISM: (1) 31  $S$  levels with  
179 stretching constants of  $\theta_b=1$ ,  $\theta_f=4$  (Zhang and Baptista 2008); (2) an LSC<sup>2</sup> grid with a maximum  
180 of 59 levels and an average of 29 levels in terms of computational cost (Zhang et al. 2015). The  
181 adoption of LSC<sup>2</sup> grid clearly has led to superior results especially in terms of stratification in the  
182 Baltic Sea (Fig. A1 in Appendix 1), and so this is the vertical grid used below. More examples of  
183 LSC<sup>2</sup> results can be found in Zhang et al. (2015).

184

### 185 **2.4. The model forcing**

186 On the open North Sea boundaries (Scottish Shelf and English Channel) time series of elevation,  
187 horizontal velocity, salinity and temperature are interpolated from MyOcean product  
188 (<http://www.myocean.eu>; last accessed in Jan 2015). The sea-surface boundary conditions use  
189 the output from the regional model COSMO-EU (wind, atmospheric pressure, air temperature  
190 and specific humidity) operated by the German Weather Service with a horizontal resolution of 7  
191 km. Heat fluxes (including solar radiation and downward long wave (infrared) radiation) used  
192 come from NOAA's CFSR product ([http://www.ncdc.noaa.gov/data-access/model-data/model-](http://www.ncdc.noaa.gov/data-access/model-data/model-datasets/climate-forecast-system-version2-cfsv2)  
193 [datasets/climate-forecast-system-version2-cfsv2](http://www.ncdc.noaa.gov/data-access/model-data/model-datasets/climate-forecast-system-version2-cfsv2); last accessed in Jan 2015). Monthly flow data  
194 at 33 rivers in the region are provided by the German Federal Maritime and Hydrographic  
195 Agency (Bundesamt für Seeschifffahrt und Hydrographie, BSH). Bathymetry data have been  
196 compiled from data provided by BSH, Danish Meteorological Institute and Baltic Sea  
197 Hydrographic Commission, 2013, Baltic Sea Bathymetry Database version 0.9.3  
198 (<http://data.bshc.pro/>, accessed on June 5, 2015). No bathymetry smoothing is done in our  
199 computational grid.

200

### 201 **2.5 Initialization and parameters**

202 The model is initialized with zero elevation and velocity, and the 3D profiles of temperature and  
203 salinity from the climatological data of Janssen et al. (1999). The heat exchange model inside  
204 SCHISM is Zeng's (1998) bulk aerodynamic model. We use a constant albedo of 0.06 and a  
205 Jerlov type III water (Paulson and Simpson 1977) for light attenuation. As this paper concerns  
206 large-scale process, we use a constant bottom roughness value of 0.5mm without any fine tuning.  
207 The turbulence model used in this paper is  $k-kl$  (Umlauf and Burchard 2003), with a background  
208 diffusivity of  $10^{-6} \text{ m}^2/\text{s}$ .

209 A time step of 120s is used in the simulation which translates to a maximum CFL number of  
210 16.7 (found in the NBTZ and some parts of the Norwegian North Sea coast). The transport of  
211 salinity and heat is done using a 2<sup>nd</sup>-order TVD method for depths deeper than 8m, and a more  
212 efficient upwind method for shallower depths (i.e., the limiter functions are zeroed out for  
213 shallower depths); the finer resolution used in shallow-depth areas effectively mitigates the  
214 numerical diffusion in upwind method. The simulation period presented in this paper is from  
215 July 1, 2011 to Nov. 13, 2012, or 500 days in total.

216

## 217 **2.6 Overall model computational performance**

218 The simulations were conducted on Sciclone cluster of College of William & Mary  
219 (<http://www.hpc.wm.edu/SciClone/Home>; last accessed in Jan 2015), NSF's Stampede, as well  
220 as NASA's Pleiades. On 144CPUs of Sciclone, it runs ~180 times faster than real time. About  
221 30% faster performance was achieved on the other two clusters.

222

## 223 **3. Sea level**

224 There are some important differences and similarities between the dynamics of sea level in the  
225 North Sea and Baltic Sea. In the North Sea the range of tidal oscillations is from more than 9 m  
226 in the English Channel to ~20 cm in the Kattegat. The narrow Danish straits damp the tidal  
227 oscillations and they propagate with measurable amplitudes no further than the Belts. The two  
228 basins, the North Sea and the Baltic Sea are in most of their areas shallow basins and therefore  
229 the atmospheric forcing creates a strong barotropic response. In the NBTZ the direction of the  
230 barotropic pressure gradient alternates (from/to the Baltic Sea), which is accompanied by a flow

231 reversal in the straits. In contrast to the deep parts of the Atlantic Ocean and the other two semi-  
232 enclosed European seas (the Black and Mediterranean Seas) the winter anomalies of sea level are  
233 positive in the shelf seas (Baltic Sea and northwest of European shelf) and negative in the deep  
234 basins (see Fig. 5.3 of Stanev and Lu 2013). This reflects the specific appearances of the fresh  
235 water budget, surface momentum and heat fluxes.

236 Because the variation of sea level in the North and Baltic seas are of utmost importance for the  
237 complex physical processes there and because their adequate simulations could guarantee the  
238 realism of interpretation of simulations, extensive validation of numerical simulations against  
239 tide gauge data has been performed. In the following we will focus on some specificity of sea-  
240 level dynamics in the both basins, as well as in the NBTZ. The latter zone is of particular  
241 importance not only because tidal amplitude reduces strongly (this is known for a long time) but  
242 also because tidal pumping can play the role of a possible driver affecting the straits exchange  
243 (Feistel et al. 2003b). Bendtsen et al. (2009) found that the tidal mixing was primarily  
244 concentrated to shallow areas around Kattegat and in the Great Belt. However these authors  
245 focused only on the NBTZ prescribing the tidal conditions at its boundaries. Furthermore, the  
246 horizontal resolution of their model was  $2' \times 2'$  minute and 30 vertical sigma layers, which is  
247 much coarser, compared to the resolution in the present study, thus justifying the revisit of this  
248 modelling issue using a coupled North Sea-Baltic Sea model with a very fine resolution in the  
249 straits.

250

### 251 **3.1 The North Sea**

252 The numerical simulations reveal the well-known anti-clockwise rotation associated with the  
253 Kelvin waves. The simulated tidal range reaches  $\sim 9\text{m}$  in the English Channel. One representative  
254 illustration of the spring-neap tide modulation is shown in Fig. 3a for the station of Jersey  
255 (Fig.1). At other stations, such as Helgoland this modulation is much less pronounced (Fig. 3b).

256 The Taylor diagram (Fig. 4) demonstrates in a systematic way the model skill. Most of the North  
257 Sea stations cluster in a small vicinity close to the purple star representing the “perfect model-  
258 performance”. While the correlation coefficient of all analyzed stations is 0.89 the ones in the  
259 North Sea (black dots in the Taylor diagram) show much better correlation. The model skill  
260 measured as the ratio between RMSE and the standard deviation of observation is about 0.1 (the

261 same number for all analyzed stations is 0.19). The deviations from this good performance at  
262 some stations may be associated with some dubious outliers in the observations.

263

### 264 **3.2 The Baltic Sea**

265 The variability of Baltic Sea level is represented in Fig. 3 at the stations of Tejn, Kronstadt and  
266 Kemi. The first station is in the Bornholm Basin, the second is the eastern-most part of the Gulf  
267 of Finland and the third on the Finish coast in the Bothnian Bay (see Fig. 1 for their positions). In  
268 all three stations the temporal variability in the simulations agrees well with that in the  
269 observations; the response to the atmospheric forcing very clearly demonstrates that the latter  
270 governs the ocean variability. However the correlation between the individual stations is not  
271 particularly strong in either model or observations, indicating large differences in the regional  
272 responses. At the three stations the model slightly overestimates the amplitudes associated with  
273 the weather systems, in particular during periods of extreme winds, e. g. at day 75 in Fig. 3d. At  
274 Kronstadt the high-frequency basin modes are underestimated while at Tejn the amplitudes of  
275 basin modes are comparable with the observed values. The variability in the Bothnian Bay which  
276 is in the northernmost part of the Baltic Sea also demonstrates large wind driven response (Fig.  
277 3e).

278 The overall skill of the model to predict the variability of sea level in the Baltic Sea is quantified  
279 by the blue circles in Fig. 4. In most of the stations the correlation coefficient is about 0.95; the  
280 scaled difference between model and observations is about 0.20. The model slightly over-  
281 predicts the range of oscillations. The performance is not satisfactory at Tallinn where the  
282 correlation with observation is  $\sim 0.72$ . Maximum over-predictions of the range of oscillations  
283 appear at Soru, St Petersburg and Tallinn, which are coastal locations in the Gulf of Riga and  
284 Gulf of Finland, possibly due to local bathymetric errors and or bottom friction. Note that the  
285 bathymetry used may not have fine enough resolution in many parts of Baltic. More detail about  
286 the quality of model performance is given in Appendix 2 where time series are shown  
287 exemplarily for 44 tide gauge stations around the Baltic Sea coast.

288

### 289 **3.3 The NBTZ**

290 The variability in the Danish straits is represented below with two stations: Kattegat  
291 (Frederikshavn, which is close to the North Sea boundary of the NBTZ) and Drogden which is at  
292 the southern entrance of the Sound (see Fig. 1 for their positions). Although very small  
293 (amplitudes less than 20 cm) the tidal oscillations in Frederikshavn are clearly pronounced (Fig.  
294 3f). However unlike most stations in the open part of North Sea (Fig. 3a&b) their modulation  
295 does not reveal a clear spring-neap cycle, but is rather dependent on the meteorological  
296 conditions (weekly to 10-days oscillations are very clearly seen in Fig. 3f).

297 The tidal signal is substantially reduced on the other side of the Baltic Sea straits (Drogden). As  
298 seen in Fig. 3g, the tidal amplitudes there reach ~10 cm, but change strongly over time. The  
299 model replicates well all substantial oscillations seen in the observations; the range of oscillation  
300 reaches ~1m at times. The difference between Fig. 3f and Fig. 3g also indicates very strong  
301 amplitudes and is instructive for the barotropic driving force in the straits, mostly due to the  
302 atmospheric forcing. The temporal evolution of sea level in Fig. 3c and Fig. 3g looks  
303 qualitatively more similar than between Fig. 3f and Fig. 3g. The most pronounced difference  
304 between Fig. 3c&g is the further reduction of the tidal range as one approach the Baltic proper.

305 At most of the stations in the NBTZ presented in Fig. 4 (red circles) the correlation coefficients  
306 are about 0.9, but there are some stations with the correlation  $< 0.9$ . At some stations the model  
307 over-predicts the variability (e. g. the normalized RMSE is relatively high at Fredericia).

308 The contrast between the points in the three different areas in Fig. 4 demonstrates that the straits  
309 are the most difficult part to simulate correctly even with the high resolution used here.

310 However, the temporal evolution of sea level presented in Appendix 2 demonstrates that the  
311 overall performance of the model is good also in the NBTZ. Note that some ‘errors’ are due to  
312 some suspicious outliers in the observations, which remained in the validation data set despite  
313 our effort to eliminate obvious outliers; Fig. A2 shows some of these suspicious data at some  
314 stations. We did not filter out those data because their quality is dubious.

315

## 316 **4. Thermohaline fields.**

### 317 **4.1 Surface-to-bottom salinity difference in the shallow channels of the NBTZ**

318 The temporal change of surface-to-bottom salinity difference is very instructive for the regime of  
319 vertical mixing and re-stratification. The numerically simulated temporal variability of salinity in  
320 the Fehmarn Belt matches relatively well the similar variability seen in the observations (Fig. 5).  
321 The skill of the model is quantified by the RMS errors normalized by the standard deviations of  
322 observation: 0.18 at the surface and 0.18 at the bottom layer (the non-normalized RMS errors are  
323 2.03 and 1.68; the standard deviations of observation are 11.27 and 9.34, respectively). The  
324 normalized model biases are 0.16 (surface) and 0.16 (bottom). The correlation coefficients are  
325 0.80 (surface) and 0.76 (bottom), and the Wilmot scores (Wilmot 1981) are: 0.82 (surface) and  
326 0.72 (bottom).

327 The maximum salinity up to 22 occurs in winter when the vertical stratification is weak. The  
328 stratification disappears during 1-2 months, with the sea-surface salinity (SSS) reaching ~20  
329 (compare also with Feistel et al., 2003a; Feistel et al., 2006). Both the model and the  
330 observations clearly showed several small outflow events revealed by the decrease of salinities  
331 with weakly time scales (e. g. between days 187 and 194 in Fig. 5). Although the model does not  
332 exactly replicate these events, their phase and duration agree well with the observations. The  
333 outflow-dominated conditions are well illustrated by the increased difference between the  
334 surface and bottom salinity values during days 120-150. This difference decreased dramatically  
335 by the end of fall (day 150), which is interpreted as a result of the vertical mixing due to cooling  
336 from the atmosphere, which changes the properties of waters entering the straits from  
337 neighboring basin. During such conditions the outflow signatures of salinity are a combination of  
338 variability with weekly periods, which is triggered by meteorological forcing and small short-  
339 periodic oscillations over-imposed.

340

#### 341 **4.2 Temporal evolution of salinity and temperature in the Arkona Basin**

342 The penetration of mixed North Sea water into the Baltic Sea follows a pathway along channels,  
343 sills and small deeper basins such as the Arkona and Bornholm Basin (Jakobsen 1995). The latter  
344 two basins play the role of reservoirs for denser water which under specific conditions  
345 propagates to the next (deeper) basin on their way to the Gotland Deep. Therefore it is  
346 worthwhile to analyzing the simulation results in these basins.

347 With the increasing distance from the straits, the inflow increasingly resembles the gravity flows  
348 (Fig. 6b, d). During most of the time the dominant two-layer salinity stratification is very clear  
349 both in the observations and simulations: about 5-10 m thick saltier bottom layer with salinities  
350 of ~14-18 is overlain by a thicker surface layer with salinity less than 8. Although the  
351 simulations appear slightly more diffuse than the observation and the bottom salinity is slightly  
352 overestimated, the timing of the individual bottom salinity maxima matches well, suggesting that  
353 the individual inflow events are captured by the model.

354 We remark that Fig. 6 displays more than four months of data during the summer and fall of  
355 2012. At the beginning of this period (summer) the cold intermediate layer (CIL) is very  
356 pronounced with an axis at ~30-35 m reaching at some times the bottom at the analyzed location  
357 (Fig. 6 a, b). The model produces slightly weaker vertical stratification, and the depth of the  
358 thermocline is ~5m too shallow; this indicates an overall reasonable model performance. The  
359 cold water content is reduced several times due to strong warming from above. After the day 435  
360 this layer collapsed, as seen in the observations and numerical simulations, as a result of the  
361 inflow of warm and high-salinity surface water from the straits. After day 460 (i.e., fall) the  
362 depth of surface cooling increases, and the upper layer is cooled more than the bottom. However  
363 the strong salinity stratification of deeper layer (Fig. 6 c, d) does not allow a complete mixing of  
364 the entire water column.

365 The correlation coefficients between simulated and observed temperature and salinity are 0.87  
366 and 0.77, respectively. Although the RMSE and bias for the bottom salinity seem relatively large  
367 (1.35 and 0.97 PSU, respectively), it is small compared to the standard deviation of data (6.8  
368 PSU), demonstrating a good skill.

369 The large differences between presented simulations and observations are explained by the large  
370 contrasts in the NBTZ and their rapid changes. However in comparison with other estimates they  
371 are not so large. For example, Fu et al. (2012) reported a salinity bias of plus 3-4 psu in the  
372 Bornholm Basin. In contrast to the present study these authors used data assimilation. In a  
373 similar study Liu et al (2013) demonstrated that compared to the free run, temperature and  
374 salinity had been improved significantly with data assimilation. Our results demonstrate that  
375 there is still a potential to get reasonable results even without data assimilation if more adequate  
376 resolution in the NBTZ is used.

377

### 378 **4.3 The Sea Surface Temperature (SST)**

379 Recall that the thermohaline part of the model forcing includes heat and water fluxes. In the  
380 following the Operational Sea Surface Temperature and Sea Ice Analysis (OSTIA) data will be  
381 used as the validation data set. The OSTIA system produces a high resolution analysis of the  
382 current SST for the global ocean using satellite data provided by the GHRSSST (The Group for  
383 High-Resolution SST) project, together with in-situ observations to determine the sea surface  
384 temperature (Donlon et al., 2009). The analysis was performed using a variant of optimal  
385 interpolation described by Bell et al. (2000) and is provided daily at a resolution of  $1/20^\circ$  (~5km).

386 The comparison between OSTIA and model results is presented in the following for simulation  
387 days 30-500. The RMSE between model and OSTIA data is small over most of the North Sea  
388 (less than  $1^\circ\text{C}$ ) (Fig. 7a). The low-error zone extends from the northern open boundary to the  
389 south mostly following the western coast. The simulations in the English Channel show errors  
390 reaching locally  $\sim 1.5^\circ\text{C}$ , which is mostly due to some boundary condition errors there (note  
391 some large errors right at the boundary). The comparison with the range of temperature  
392 fluctuations in this region demonstrates that the relative errors are small. The local maximum in  
393 the northeastern part of the basin is due to errors in OSTIA data (which can be seen from the fact  
394 that the SST from OSTIA there is mostly stationary over time). However the larger errors in the  
395 western part of Skagerrak and along the Norwegian Trench could present a motivation to further  
396 improve the performance of the model in these areas. It's encouraging that the errors in NBTZ  
397 are generally small.

398 The RMSE pattern in the Baltic Sea could give a possible explanation of the large differences  
399 between the simulation and observation in the Skagerrak. High errors occur usually along the  
400 west coasts of Baltic Sea and south of them. In this neighborhood OSTIA data are sometimes  
401 unavailable (see the blank zones). Upwelling events along the Swedish coast explain the  
402 substantially cooler temperatures than in the interior and southern part of the basin. The  
403 accompanied tilting of isotherms toward the Swedish coast and upwelled CIL there are clearly  
404 seen in the model results also (not shown). Therefore the bias in these areas is clearly negative  
405 (Fig. 7b). Other potential error sources include the lack of an ice model (especially for the Bay of  
406 Bothnia), or the wind stress parameterization.

407

## 408 **5. The Water Masses**

409 The question of whether or not the simulated water mass structure is correct under the variable  
410 horizontal resolution as found in unstructured grids is a critically important issue to be addressed  
411 when estimating the performance of a new (unstructured-grid) model setup for such a complex  
412 area. In the following 3 subsections, three aspects of this issue will be analyzed using the  
413 example of the formation of CIL, the two-layer exchange in the NBTZ and the propagation of  
414 bottom water into the Gotland Basin.

### 415 **5.1 The Cold Intermediate Water**

416 The vertical stratification in the Baltic Sea is dominated by salinity (Fig. 8b), which provides  
417 enormous stability, decreases vertical diffusion (Stigebrandt, 1987; Meier, 2001; Stanev, 1997;  
418 Omstedt, 2014) and therefore prevent the permanent halocline from being destructed by the  
419 winter cooling. The vertical salinity gradient at ~150 m only changes slightly, mostly in winter,  
420 as a result of the convective cooling. The small-scale oscillations at the interface, separating  
421 surface and deep waters, reveals the model representation of the oscillations of pycnocline.

422 The cold water mass formation starts in fall, first in a shallow part of the upper layer (Fig. 8a); in  
423 January-February the cooling reaches depths of ~60 m. This value varies in space reaching in  
424 some areas depths of ~65 m. The 25 years-long simulations of Meier (2007) showed also a  
425 pronounced inter-annual variability of winter convection. In March-April the re-stratification  
426 starts to form and the cold water is overlain by waters from the seasonal thermocline. As the time  
427 progresses after the maximum cooling, the cold water content of the CIL starts to decrease,  
428 which is seen in Fig. 8a as a decrease of its thickness. The refill and thickening of the CIL in  
429 winter and the decrease of its cold water content in summer repeats in an almost periodic way  
430 every year. These results demonstrate that the model replicates some of the known features and  
431 processes associated with the temporal evolution of thermohaline circulation in the Baltic Sea.

432

### 433 **5.2 The Two-layer Exchange in the NBTZ**

434 The widths of the straits in the NBTZ, i.e. the Great Belt, the Sound and the Little Belt, are 16-32  
435 km, 4 km and 1 km respectively. The comparison of the straits widths with the internal

436 (baroclinic) Rossby radius of deformation (~5 km in the Kattegat) indicates that the Great Belt  
437 could be rotationally dominated. However all Baltic Sea straits are shallow and mostly friction-  
438 dominated. Furthermore the narrower ones are shaped by sills and contractions, which make the  
439 understanding of processes there difficult. In the following the simulations results are illustrated  
440 along three transects in the straits (Fig. 9a) for temperature and salinity sampled from the model  
441 outputs at 00:00, Jan. 17, 2012 and at 00:00, Feb. 18, 2012, for outflow and inflow conditions  
442 respectively. The bottom relief changes very differently along the three sections. The main sill  
443 for the Great Belt is the Darss Sill (383 km on the section). However it is not only this sill, which  
444 is practically outside the strait, but also the narrows that constrain the two-layer exchange. In the  
445 Little Belt the shallowest depths along the sections are close to the corresponding narrowest  
446 section. However in the Drodgen Sill is located at the southern entrance of the Sound whereas  
447 the narrowest section is located in the northern part of Sound between Helsingör and  
448 Helsingborg.

449 The overall conclusion from the cross-sections in Fig. 9 is that the straits decouple the water  
450 masses, which is better seen in the field of salinity under the outflow condition (Fig. 9 b, d, f).  
451 This is well pronounced in the Great Belt and the Sound by the tongue of Baltic Sea water  
452 propagating into the direction of Kattegat. In the Sound this tongue propagates with relatively  
453 low values north of the strait while the mixing along the wide and shallow Great Belt reduces  
454 substantially the salinity contrast. Therefore the salinity distribution is smoother in this strait and  
455 the isohalines rather vertical and the gravity flow-like pattern (high salinity at the bottom, not  
456 reaching the sea surface) is seen only beyond the Darss Sill. Unlike this strait, the Sound shows  
457 more classic estuarine patterns with far reaching intrusions in the Kattegat, and a gravity flow-  
458 like pattern just after the sill only during inflow situation (Fig. 9l). The situation in the Little Belt  
459 is rather different from the two wider straits. Here the water south of the sill is more stagnant,  
460 which enables higher salinity values south of the sill. However the salinity contrast peaks around  
461 the sill (Fig. 9b). Note that because the very narrow section of this strait the back-and-forth  
462 oscillations of salinity front in the Little Belt are not in phase with the ones of the main straits.

463 For the outflow period analyzed above the temperature in the Baltic Sea is higher or comparable  
464 with the one in Kattegat (Fig. 9c, e, g). The absolute maximum is below the sea surface, which  
465 again demonstrates that the strong salinity stratification is the main stabilizing factor. Another  
466 clear evidence of the dominant role of salinity is that the along-channel separation of temperature

467 follows the maximum extension of the Baltic plume. During the inflow phase the North Sea  
468 water is clearly identifiable by the warmer temperature; the temperature patterns (Fig. 9i, k, m)  
469 are very similar to the ones of salinity (Fig. 9 h, j, l).

470

## 471 **5.3 The Water Mass Formation**

### 472 **5.3.1 The Cascading of Mixed North Sea Water into the Gotland Basin**

473 The water mass formation in the Baltic Sea reflects the balance between the heat and water  
474 fluxes at the surface and in the straits. The latter is illustrated in Fig. 10a by the cascading of  
475 inflowing water from the straits area down into the Baltic proper (an animation of the gravity  
476 flow can be viewed at:

477 [http://ccrm.vims.edu/yinglong/TMP/anim\\_transect\\_TS\\_Darss\\_to\\_Gotland\\_new.zip](http://ccrm.vims.edu/yinglong/TMP/anim_transect_TS_Darss_to_Gotland_new.zip)). Physically  
478 this process has much in common with the major Baltic flow events. The situation shown in Fig.  
479 10(a&b) happened in February, 2012, which is consistent with the overall occurrence of the  
480 major inflow events. However it is weaker and there is no sufficiently big volume of mixed  
481 North Sea water that reaches the deepest layers of the Gotland Deep. The analysis of the 500-day  
482 simulations indicates that the situation shown in Fig. 10 is repeated many times during the  
483 analyzed period but usually with a lesser intensity.

484 During most of the time the Bornholm Basin is filled with mixed North Sea water up to its sill  
485 depth. The halocline reveals seiches-like oscillations at the interface in these basins triggered by  
486 the inflowing water, which supports the concept of Jakobsen (1995). Under strong wind  
487 conditions these denser waters overshoot the sill and cascade into the next basin. Beyond the  
488 Stolpe Channel the inflowing bottom water undergoes strong mixing over the sloped bottom.  
489 Traces of this saltier and warmer water can be found as deep as 150-200m. In the deeper layers  
490 its penetration is better illustrated by temperature, which acts as a tracer in this salinity-  
491 dominated environment.

492 As far as the large-scale surface salinity patterns are concerned, the continuous decrease of  
493 salinity to the east and the thinner diluted layer to the west give a manifestation that the  
494 numerical model captures well the main characteristic of this sea being a huge estuary. The  
495 patterns of temporal variability illustrated in Fig. 10 demonstrate that the simulated water mass  
496 formation undergoes large transformations basin-wide.

497

### 498 **5.3.2 The Ventilation of the upper layer**

499 The surface ocean processes are presented by the temperature distribution in winter (Fig. 10b)  
500 and summer (Fig. 10d). The seasonal thermocline starts to form at the beginning of April as the  
501 warmer surface water overlies the cold surface water formed by previous winter cooling. The  
502 temperature continuously increases up to 22° C in August, but this surface layer never gets  
503 thicker than 20-40 m. Very close to its upper boundary, at about 40-50 m, is the core of CIL with  
504 temperatures lower than 3 °C. It extends towards the NBTZ where it flattens. In this area the  
505 CIL is subject to periodic back-and-forth oscillations caused by the inflowing-outflowing straits  
506 circulation. The CIL mixes with saltier water in the straits, and depending on the season  
507 propagates with the inflow as the warm bottom water in summer (Fig. 10d) or cold bottom water  
508 in winter (Fig. 10b). This mixed water mass cascades from one to the next “retention” basin on  
509 its way to the Baltic proper.

510 The formation of cold intermediate water along the analyzed cross-section starts first at the  
511 surface of the Gotland Deep and at the end of September the CIL outcrops. Local traces of  
512 anomalous (warmer or colder) waters entrapped below the CIL propagate down-slope as seen by  
513 the temperature pattern in the area between the Bornholm Basin and Gotland Deep (Fig. 10  
514 b&d). Noteworthy is the sequence of cold and warm water intrusions at different depths in winter  
515 (Fig. 10b). Obviously the dominant stratification caused by salinity limits the depth of  
516 propagation of waters from the CIL. Below these cold intrusions, warm intrusions are  
517 propagating with the gravity currents at the bottom. The variety of intrusions of temperature  
518 compared to the patterns of salinity demonstrates again that temperature over most of the basin  
519 can be roughly considered as a passive tracer.

520 The refill of the Baltic Sea with cold water continues until the end of February. During this time  
521 part of the cold water is formed not only in the interior of the sea. In the shallower zone  
522 convection reaches the bottom and cold plumes propagate towards the deeper part of the basin.  
523 Because of the low salinities on the shelf their density is not high enough to reach deeper levels.  
524 Therefore these cold waters propagate in the direction of Gotland Deep as horizontal intrusions  
525 above the halocline contributing to additional cooling of the basin interior.

526

527 **6. Conclusions**

528 We have described here a new unstructured-grid model and its application to the North Sea and  
529 Baltic Sea, a region dominated by tides, inflows, shelf and baroclinic processes. The focus was  
530 on the NBTZ (North Sea - Baltic Sea Transition Zone) which is the most difficult area to  
531 simulate in the studied region because it requires very good quality of simulations of both  
532 barotropic and baroclinic processes. This zone is crucial for the processes developed on either  
533 side of straits (the Skagerrak and Norwegian Trench on one side, and the Baltic Sea on the other  
534 side). The specificity of the NBTZ in comparison to other similar zones is that its dynamics are  
535 controlled by multiple straits, the narrowest one being only 1km wide and the widest being about  
536 16 km wide. With the setup of the presented model we resolved in a reasonable way these straits  
537 which ensured realism in the simulations of the exchange flows.

538 The model overcame some inconsistencies in earlier simulations that either covered only one sea  
539 (only Baltic or only North Sea) or used non-seamless one-way or two-way nesting techniques.  
540 The placement of the open-boundary conditions far from the straits made it possible that the  
541 model developed its own dynamics which appeared consistent with the tide gauges and  
542 temperature-salinity observations. Thus the presentation of the model validation appears one  
543 substantial part of the present research because this is a prerequisite for establishing some  
544 confidence in the interpretation of numerical simulations.

545 The statistical analysis of the simulated sea levels demonstrated that most of the analyzed North  
546 Sea stations tend to tightly follow the observations; the tidal variability is very well simulated.  
547 The dynamics of sea levels in the Baltic Sea are mostly driven by the evolution of weather  
548 systems although the responses greatly differ over different areas of the sea. Eigen oscillations  
549 also seem to have an important contribution to the Baltic Sea dynamics. The model captured all  
550 these important responses well.

551 The NBTZ presents an interesting case of reduction of the tidal range when the tides propagate  
552 through the straits. On the other hand, in this zone, the relative contribution from the weather  
553 systems (which is part of the barotropic forcing) increases because the straits amplify the sea-  
554 level contrasts between the Baltic Sea and North Sea. The reversal of the atmospherically driven  
555 barotropic gradient triggers the temporal change of surface-to-bottom salinity difference. This is  
556 clearly revealed in the variability of the vertical mixing and re-stratification, and is very well

557 validated by the similar variability seen in the observations. It is noteworthy that in this zone  
558 reversal of transports (inflow-outflow conditions) is well illustrated by the increased difference  
559 between the surface and bottom salinity values. With increasing distance from the straits other  
560 topographic features such as small basins (Arkona, Bornholm) start to play the role of reservoirs  
561 for denser water which under specific conditions propagates to the next (deeper) basin on its way  
562 to the Gotland Deep. In these areas the inflow resembles the gravity overflows. The individual  
563 inflow events were well captured by the model.

564 For the Baltic Sea modelling, simulating water mass formation is of particular importance  
565 because the distribution of water masses gives the major representation of the estuarine dynamics  
566 of this basin. The simulated refill and thickening of the CIL (cold intermediate layer) in winter  
567 and the propagation of cold water in summer demonstrated that the model successfully replicates  
568 the known features of the temporal evolution of thermohaline circulation in the Baltic Sea. Of  
569 particular interest is that the simulated SST demonstrated a good agreement with the OSTIA  
570 data. Furthermore, the evolution of CIL near the straits was also realistically simulated. This was  
571 illustrated by the realistic simulation of the mostly salinity-dominated vertical stratification and  
572 the cascading of inflowing water from the straits area down to the Baltic proper; the cascading  
573 inflow process is one important mechanism in maintaining the vertical conveyor belt (Döös et  
574 al., 2004).

575

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592

### 593 **Appendix 1. Choice of vertical grids**

594 The important role of vertical grids in ocean models has long been recognized, and traditional  
595 choices ( $Z$ , terrain-following, or isopycnal coordinates) all suffer from certain shortcomings.  
596 Therefore we have recently proposed a new type of vertical grid that maximizes the flexibility in  
597 resolving processes of contrasting scales as found in the ocean, river and lake modeling (Zhang  
598 et al. 2015). Dubbed Localized Sigma Coordinates with Shaved Cell ( $LSC^2$ ), it allows flexible  
599 placement of vertical grid nodes at each horizontal position while minimizing non-smoothness  
600 among neighboring nodes. The latter is ensured by the Vanishing Quasi Sigma (VQS) of  
601 Dukhovskoy et al. (2009). Moreover, the staircases found in VQS are eliminated with a novel  
602 and simple shaved cell technique that effectively shuts down the diapycnal mixing and in the  
603 same time, reduces the coordinate slope and thus the classic Pressure Gradient Errors (PGE;  
604 Haney 1991). Many tests have been conducted to demonstrate the superiority of the new  $LSC^2$   
605 grid (Zhang et al. 2015).

606 Fig. A1 further illustrates the superior results obtained from the  $LSC^2$  grid. With the 31-level  $S$   
607 grid, the physically stable stratification as found in the Baltic proper is gradually eroded over a  
608 few months due to a combination of PGE and diapycnal (numerical) mixing. Increasing the  
609 number of vertical levels to 51 only delayed the erosion by a few months, and adjusting the  
610 stretching parameters used in  $S$  did not significantly affect the results either (not shown). On the  
611 other hand, the new  $LSC^2$  grid, despite having fewer number of levels on average, is able to  
612 maintain the stable stratification over multi-year simulations (also cf. Fig. 8b), which is crucial  
613 for Baltic Sea because the Major Inflow events only happen roughly over a ten-year scale.

614

### 615 **Appendix 2. Validation of sea levels**

616 We group the validation into 2 regions: NBTZ and Baltic Sea. The station coordinates can be  
617 seen in Table 1. Exemplary time series over a 14-day period and low-pass filtered signals are  
618 shown in Fig. A2. Some error statistics at 14 NBTZ stations are shown in Table 2, which can be  
619 compared with those in Fu et al. (2012). Tidal harmonics at North Sea stations are reported in  
620 Table 3. A mini-storm can be seen at the end of the 14-day period at most NBTZ stations, as a  
621 response to a weather system that dominated the region. No pronounced spring-neap cycle is  
622 found anywhere in the 2 regions, and while NBTZ stations exhibit some similarities among each  
623 other, the Baltic stations show greater disparity suggesting localized responses. For example,  
624 many Baltic stations did not witness the set-down near Day 148, and some show distinctive  
625 period of oscillation of a few days (#26, 30, 39, 73). Overall, the model captured both system-  
626 wide and localized responses pretty well.

627

628 Table 1. NBTZ and Baltic stations location

Station	Lon	Lat	Station	Lon	Lat	Station	Lon	Lat	Station	Lon	Lat
Aarhus	10.2167	56.15	Hornbaek	12.4667	56.1	Marviken	16.8371	58.5537	Smogen	11.2178	58.3537
Althagen	12.4194	54.3769	Juelsminde	10.0167	55.7167	Neustadt	10.8128	54.0967	Soru	22.5229	58.6938
Bagenkop	10.6778	54.7528	Kalix	23.0962	65.697	NordreRose	12.6867	55.6361	Spikarna	17.5312	62.3632
Ballen	10.6444	55.8167	Kalkgrund	9.8881	54.8247	Olands NorraUdde	17.0971	57.3662	StPetersburg	30.2667	59.9333
Barhoeft	13.0328	54.4397	Kappeln	9.9381	54.6644	Oskarshamn	16.4779	57.2749	Stenungsund	11.8325	58.0933
Barseback	12.9034	55.7565	Kaskinen	21.2167	62.3333	Paldiski	24.0796	59.3349	Stockholm	18.0818	59.3243
Daugavgriva	24.0167	57.05	Kemi	24.5167	65.6667	Pietarsaari	22.7	63.7	Stralsund	13.0986	54.3153
Degerby	20.3833	60.0333	KielHoltenau	10.1569	54.3722	Ratan	20.895	63.9861	Tallinn	24.7637	59.4444
Drogden	12.7117	55.5358	KielLT	10.2733	54.4997	Ringhals	12.1126	57.2498	Tejn	14.8333	55.25
Eckernfoerde	9.8361	54.4747	Klagshamn	12.8936	55.5223	Rodby	11.35	54.65	TimmendorfPoe	11.3756	53.9919
Flensburg	9.4331	54.795	Klaipeda	21.0833	55.7333	Rodvig	12.3728	55.2542	Travemuende	10.8722	53.9581
Forsmark	18.2109	60.4085	Kobenhavn	12.6	55.7	Rohukula	23.4248	58.9049	Triigi	22.7173	58.5914
Fredericia	9.75	55.5667	Kolka	22.5833	57.7333	Rostock	12.155	54.0831	Ueckermuende	14.0664	53.7503
Frederikshavn	10.5667	57.4333	Korsor	11.1333	55.3333	Sassnitz	13.6431	54.5108	Viken	12.5793	56.1421
Furuogrund	21.2306	64.9157	Koserow	14.0008	54.0603	Schleimuende	10.0367	54.6733	Virtsu	23.5081	58.5761
Fynshav	9.9833	55	Kronstadt	29.75	59.9667	Schleswig	9.5692	54.5114	Visby	18.2845	57.6393
Gedser	11.9333	54.5667	Kuivastu	23.3935	58.5742	Sillamae	27.7401	59.4227	Warnemuende	12.1033	54.1697
Goteborg Torshammen	11.7907	57.6846	Kungsholmsfort	15.5893	56.1052	Simrishamn	14.3577	55.5576	Wismar	11.4581	53.8989
Greifswald	13.4461	54.0928	Kungsvik	11.1274	58.9966	SjaellandsOdde	11.3722	55.975	WittowerFahre	13.245	54.5575
Grena	10.9306	56.4111	LandsortNorra	17.8589	58.7688	Skagen	10.5858	57.7194	Wolgast	13.7703	54.0417
Heiligenhafen	11.0056	54.3731	Langballigau	9.6542	54.8233	Skagsudde	19.0124	63.1905			
Helsinki	24.9667	60.15	Lehtma	22.6969	59.069	Skanor	12.8296	55.4168			

Heltermaa	23.0466	58.8664	Luebeck	10.7031	53.8931	Slipshavn	10.8333	55.2833			
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629

630 Table 2. Elevation error statistics at the 14 NBTZ stations used in Fu et al. (2012).

	Correlation coefficient	Unbiased RMSE (m)	Bias (m)
Aarhus	0.875	0.073	0.045
Frederikshavn	0.919	0.063	0.136
Slipshavn	0.800	0.067	0.046
Korsor	0.806	0.057	-0.003
Hornbæk	0.983	0.046	-0.030
Rodby	0.893	0.043	0.077
Gedser	0.908	0.043	0.017
Tejn	0.915	0.043	0.157
Kalix	0.966	0.049	0.329
Klagshamn	0.878	0.051	0.108
Kungsholmsfort	0.891	0.063	0.149
Kungsvik	0.925	0.030	0.112
Ratan	0.958	0.061	0.307
Visby	0.873	0.060	0.201

631

632 Table 3. Comparison of **M2** (which accounts for over 80% of the total tidal energy) amplitudes  
 633 and phases at North Sea stations.

Station name	Amp_Obs (m)	Amp_Mod (m)	Pha_Obs (degr)	Pha_Mod (degr)
Aberdeen	1.24	1.09	21.75	21.39
Bournemouth	0.41	0.52	271.82	273.00
Cromer	1.56	1.29	187.23	193.60
Dover	2.24	2.08	330.17	324.30
Harwich	1.32	1.32	323.84	320.15
Immingham	2.23	1.96	160.63	165.80
Jersey	3.29	3.30	176.57	175.29
Lowestoft	0.68	0.62	259.15	262.13
Newhaven	2.21	2.12	319.04	303.65
North Shields	1.60	1.34	84.66	88.08
Wick	1.01	0.91	320.81	326.75
Borkum	1.11	1.16	273.67	263.63
Helgoland	1.11	1.17	308.42	296.86
Hirtshals	0.15	0.20	107.10	113.21

634

635 **Figure captions**

636 **Fig. 1.** North Sea and Baltic Sea topography. The color scheme was chosen to better illustrate the  
 637 bottom relief in the deepest parts of the two basins. The positions of the following stations used  
 638 in model validation are also shown: (1) Jersey, (2) FINO-1, (3) Helgoland, (4) Frederikshavn, (5)  
 639 Fehmarn Belt, (6) Drogden, (7) Arkona Basin, (8) Tejn, (10) Kronstadt and (11) Kemi. Analyses

640 of numerical simulations are presented further in the paper at location (9) and along the transect  
641 line starting in the western Arkona Basin and ending in the northern Gotland Deep (grey line).

642 **Fig. 2.** Model grid size (a), and illustration of the fine grid resolution in the Great Belt (b), which  
643 is a zoom-in view of the rectangular box in (a).

644 **Fig. 3.** Modelled and observed sea level in Jersey (a), Helgoland (b), Tejn (c), Kronstadt (d),  
645 Kemi (e), Frederikshavn (f), Drogden (g). Red: observations; green: model (hourly time series).

646 **Fig. 4.** Taylor diagram of the modelled sea levels. The distance from the origin measures the  
647 RMSE of model results, scaled against standard deviation of the data. The correlation  
648 coefficients are shown on the full sector line (increasing from “North” to “East”). The purple star  
649 is the ‘truth’. Black circles are the North Sea stations, red ones the NBTZ and the blue ones the  
650 Baltic Sea stations. The average correlation is 0.8837, and average scaled RMSE is 0.19. The  
651 stations used for this validation are shown in Table 1.

652 **Fig. 5.** Time series comparison between the simulated (blue lines) and observed (red dotted  
653 lines) near-surface and near-bottom salinities at the station Fehmarn Belt (see Fig. 1 for its  
654 positions). The near-surface values are lower than the near-bottom ones.

655 **Fig. 6.** Time versus depth diagrams of temperature (a, b) and salinity (c, d) in Arkona Basin. (a)  
656 and (c) are from observation and (b) and (d) are from model.

657 **Fig. 7.** The RMSE (a) and mean bias (b) of SST between model and OSTIA data. The blank  
658 areas in Baltic Sea are due to missing data at some time instances, and the local extrema in the  
659 northern part of North Sea is due to OSTIA issue. The black isoline in (b) is the boundary  
660 between positive and negative biases.

661 **Fig. 8.** Time versus depth diagrams of temperature (a) and salinity (b) at the Gotland Basin (see  
662 location 9 in Fig. 1). The isolines in (a) represent 3, 3.5, 4, 4.5°C.

663 **Fig. 9.** Vertical cross-sections of salinity (left panels) and temperature (right panels) along 3  
664 transects in Little Belt (b, c, h, i), Great Belt (d, e, j, k) and Sound (f, g, l, m). (b-g) represent the  
665 outflow conditions at 00:00, Jan. 17, 2012, and (h-m) represent the inflow conditions at 00:00,  
666 Feb. 18, 2012. The black dashed line shows where the transect coming from Little Belt joins the  
667 transect coming from Great Belt. In all plots,  $x=0$  corresponds to the starting location in the  
668 north. The locations of the three transects are shown in (a).

669 **Fig. 10.** Water masses in the Baltic Sea along the section shown in Fig. 1. (a) is salinity on Feb.  
670 29, 2012, (b) is temperature on Feb. 29, 2012; (c) is salinity on Aug. 02, 2012, and (d) is  
671 temperature on Aug. 02, 2012.

672 **Fig. A1** Comparison of salinity profile at Gotland Deep (see Fig. 1 for location), from Nov. 1,  
673 2013 to Feb. 1, 2014, using (a) 31  $S$  layers, and (b)  $LSC^2$  grid. See Section 2.3 for more details of  
674 the two types of vertical grids used.

675 **Fig. A2** Comparison of elevations at (a) NBTZ and (b) Baltic Sea stations. The station locations  
676 are shown in Table 1. The red lines are observation, and green lines are model results. The  $x$ -axis  
677 is time in days from July 1, 2011. (c) and (d) show the low-pass filtered signals (with 30-hour  
678 cut-off) for (a) and (b) over a longer period.

679

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Figure1

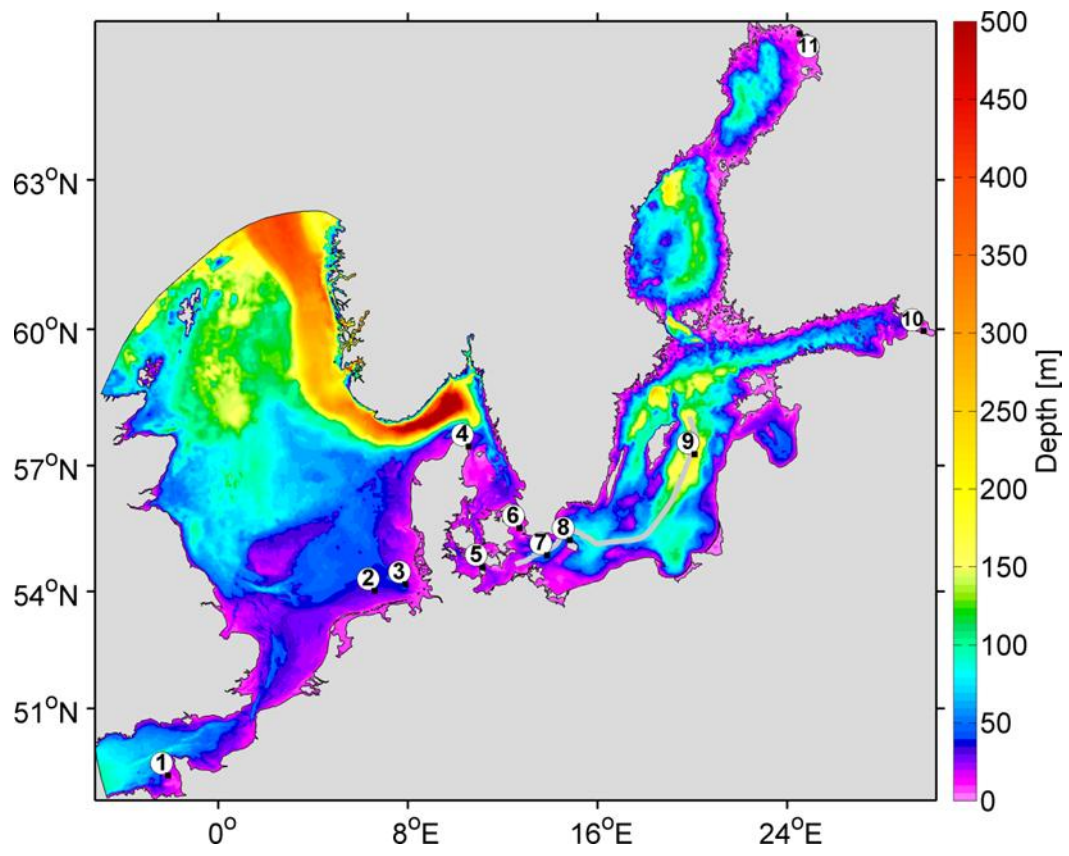


Fig. 1

Figure2

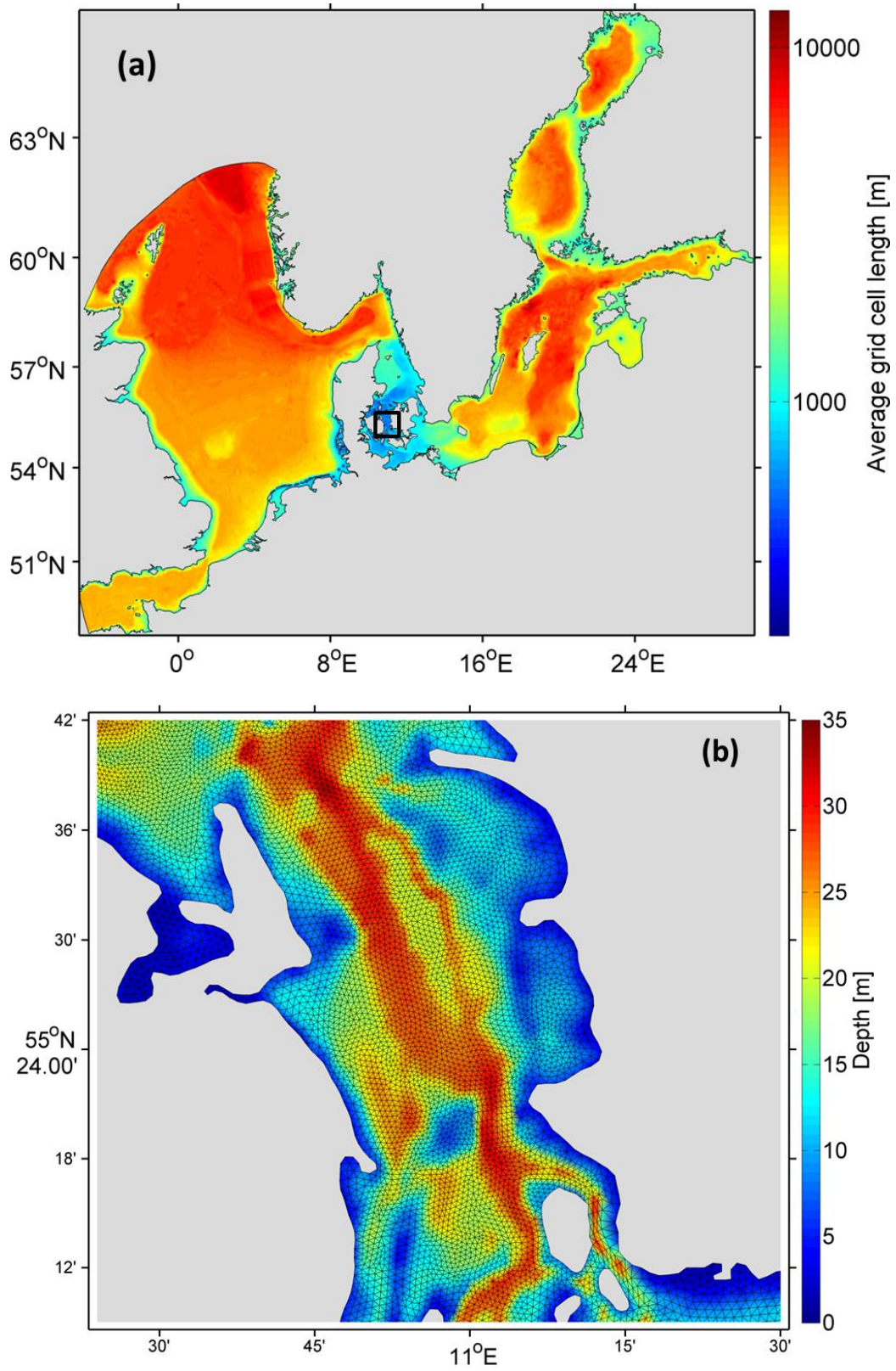


Fig. 2

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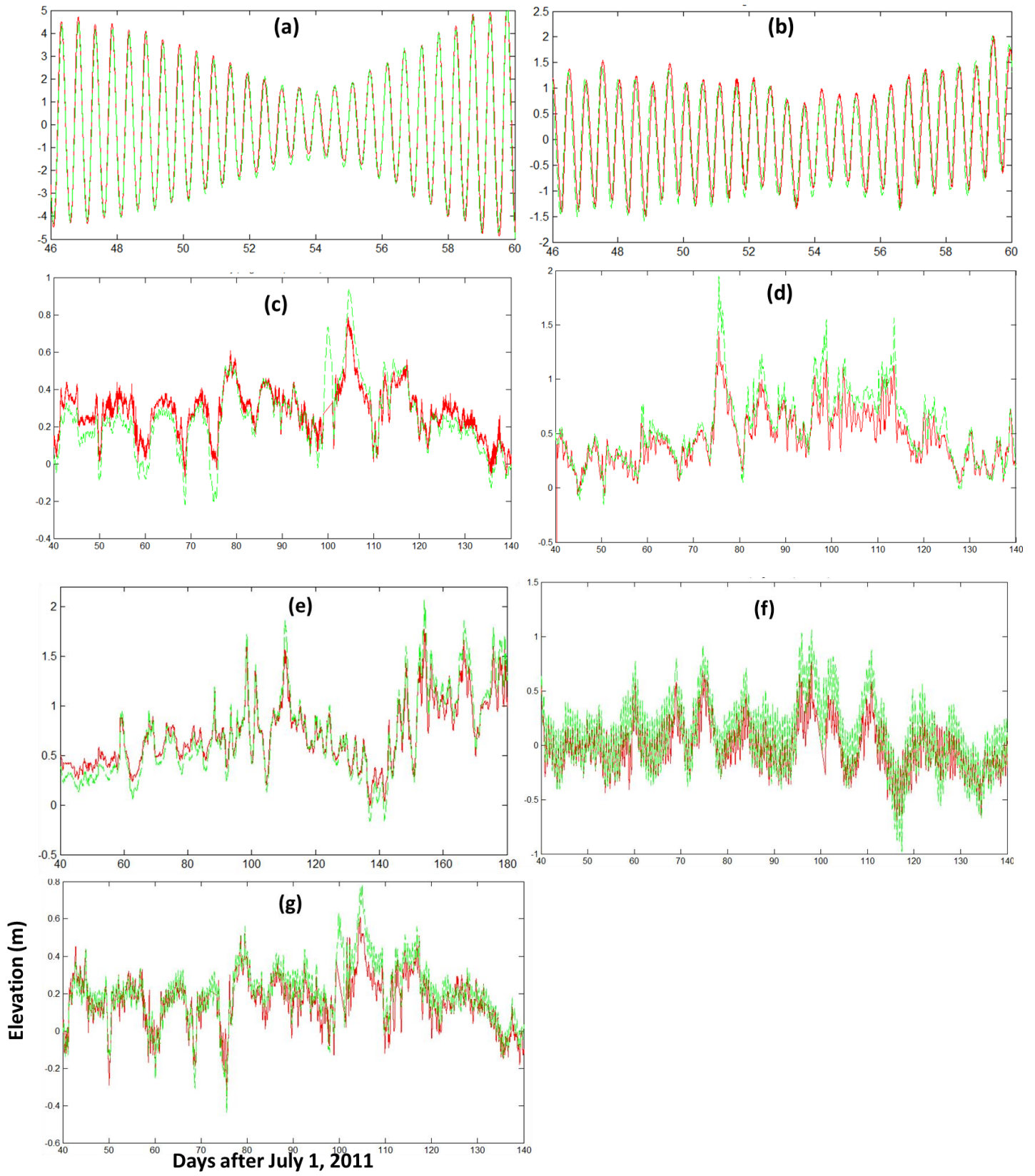


Fig. 3

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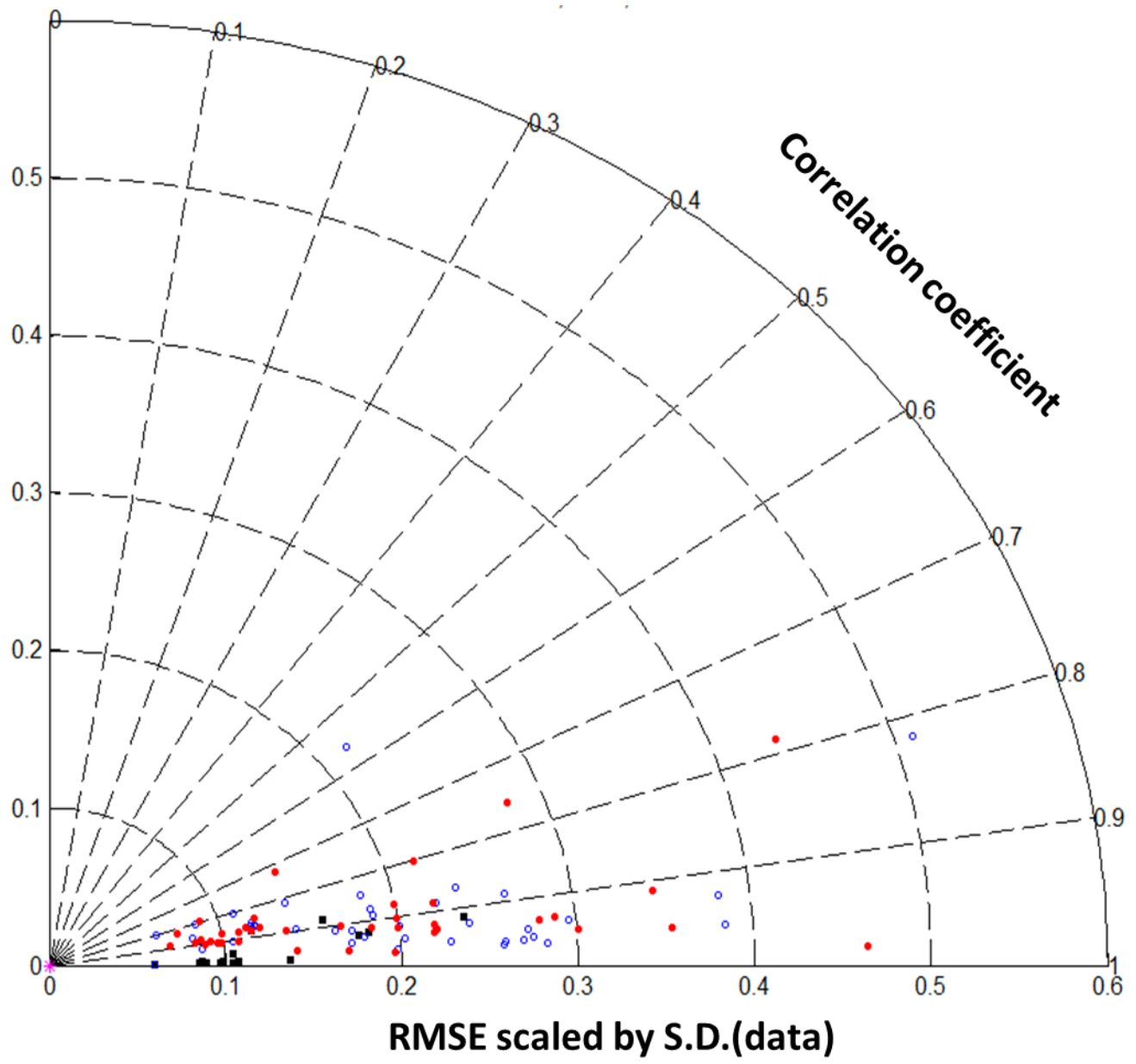


Fig. 4

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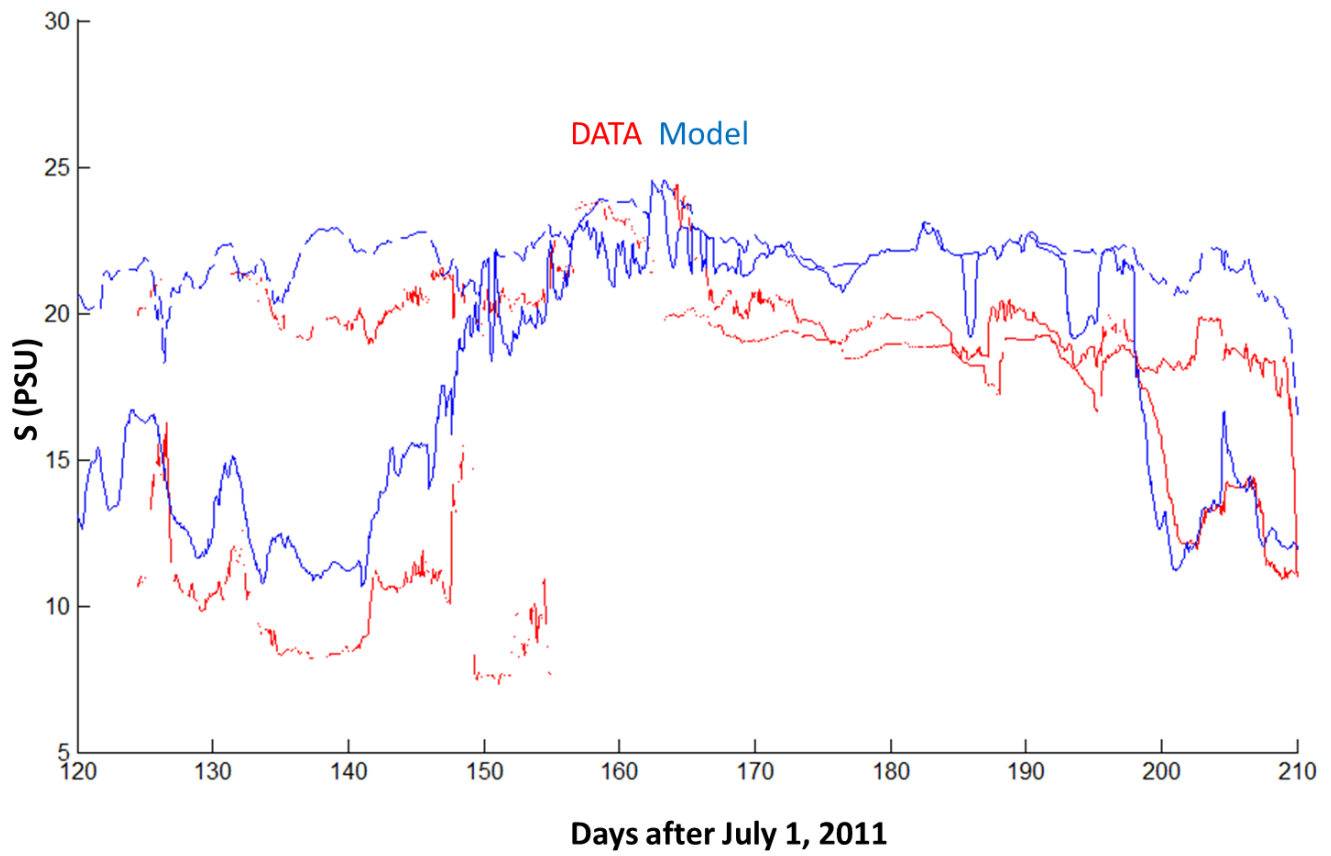


Fig. 5

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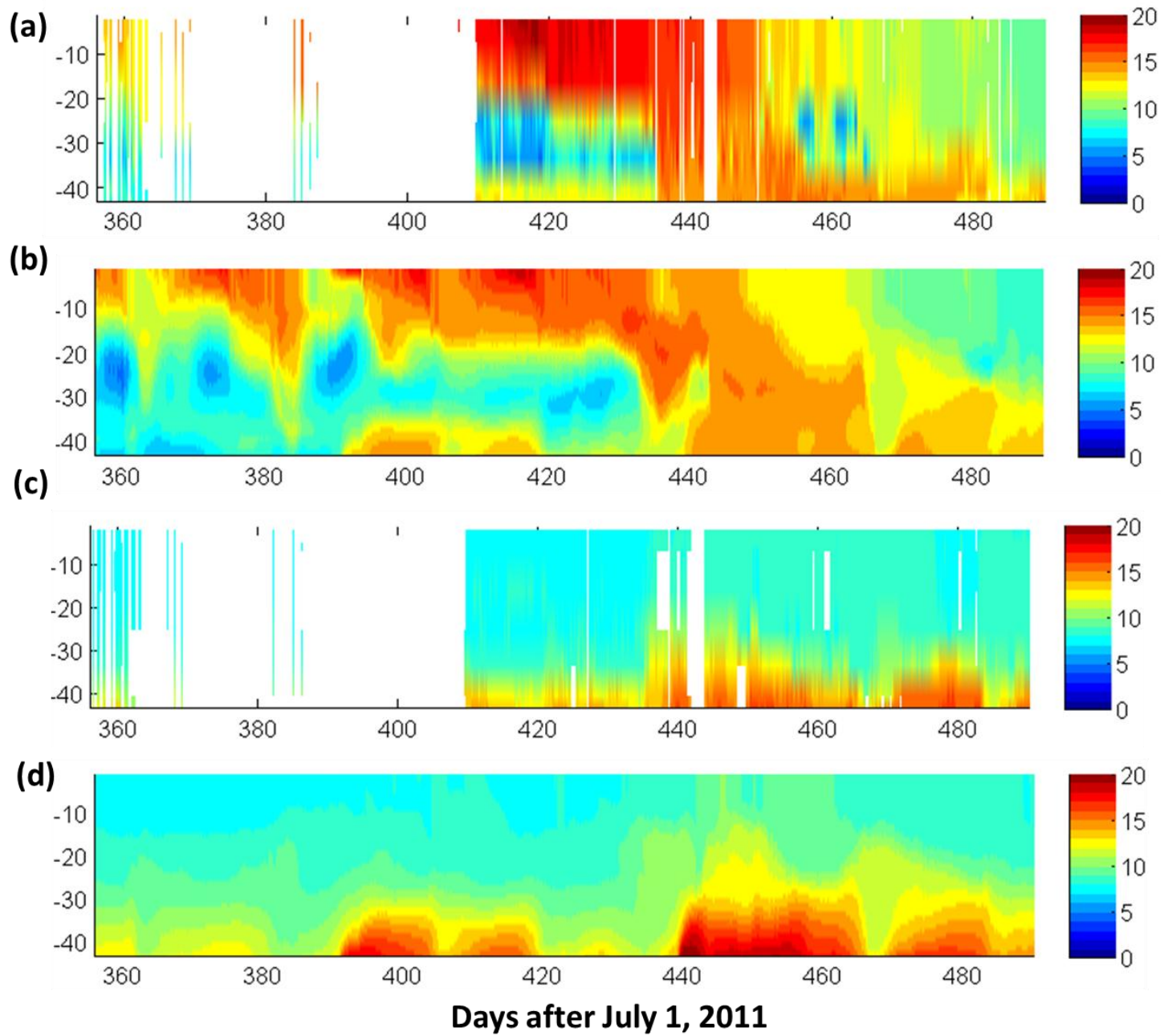


Fig. 6

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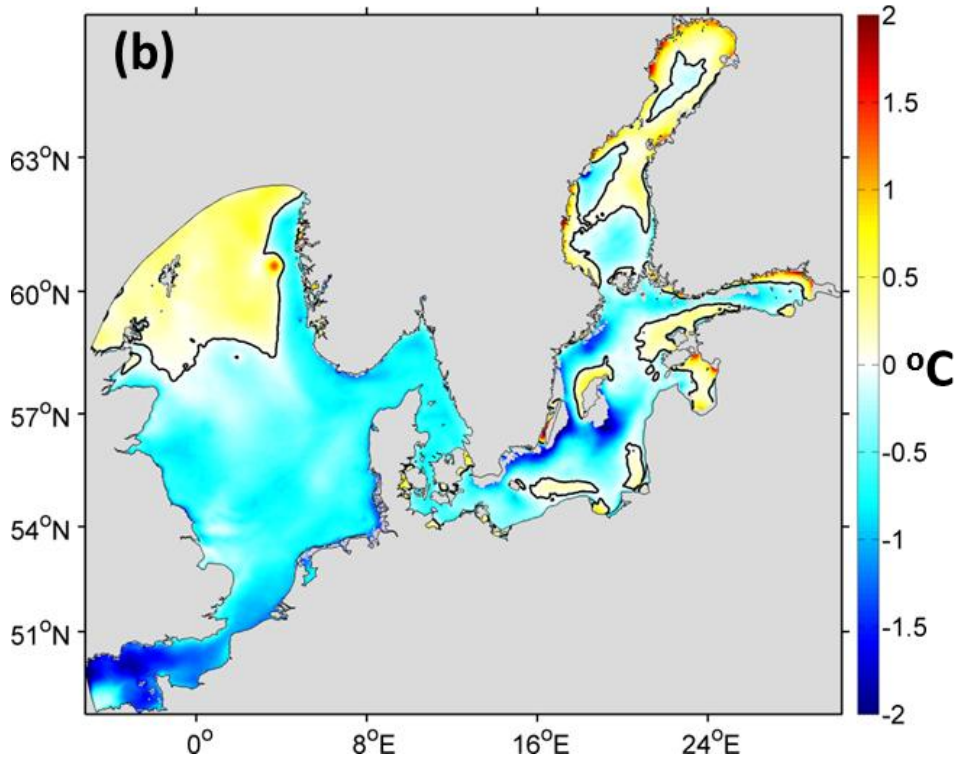
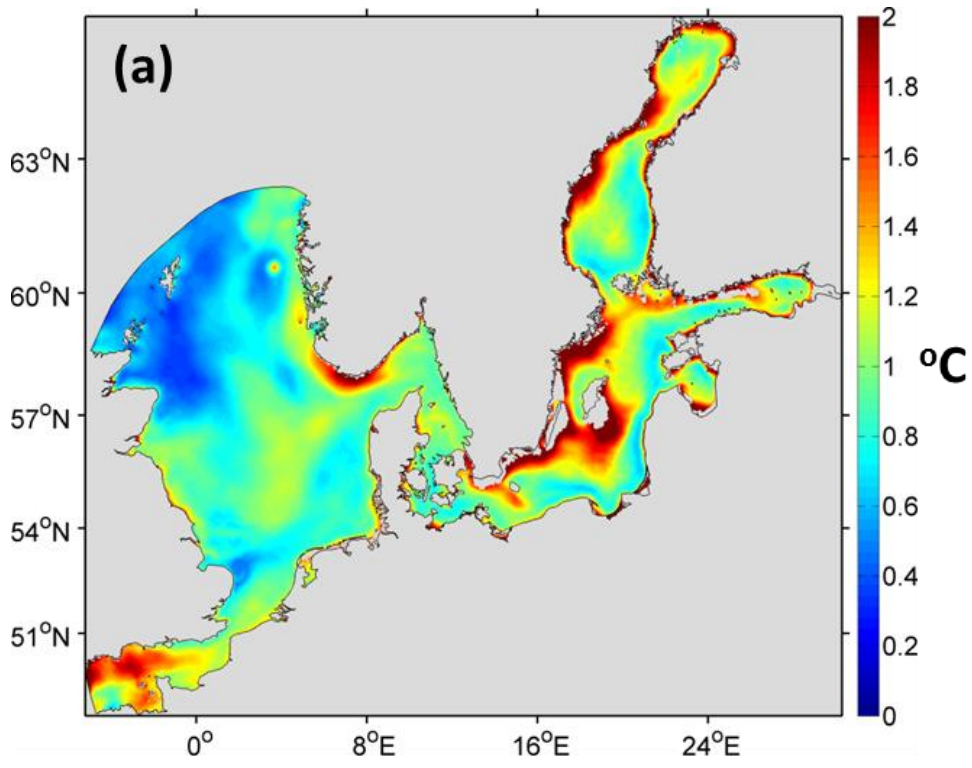


Fig. 7

Figure8

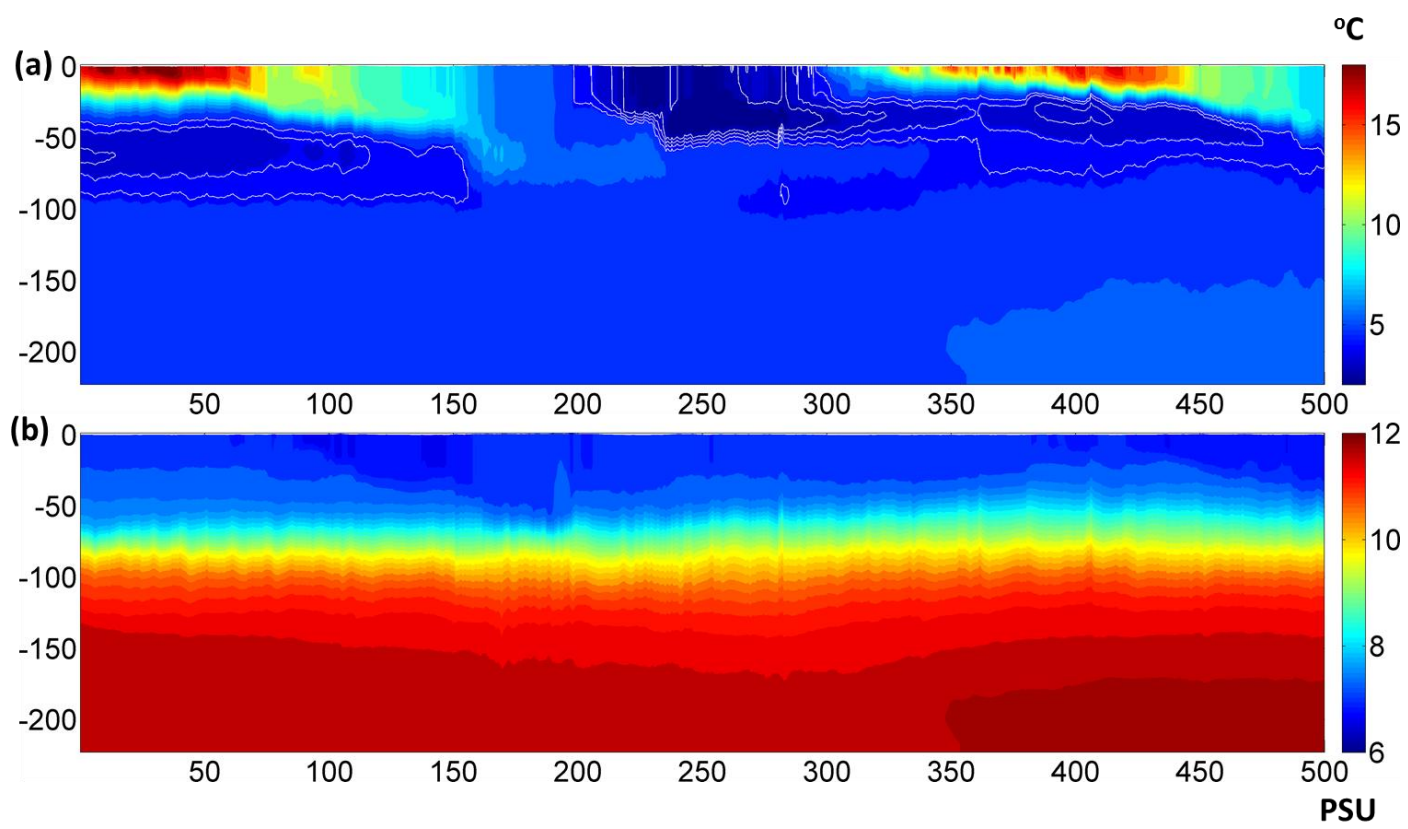


Fig. 8

Figure9

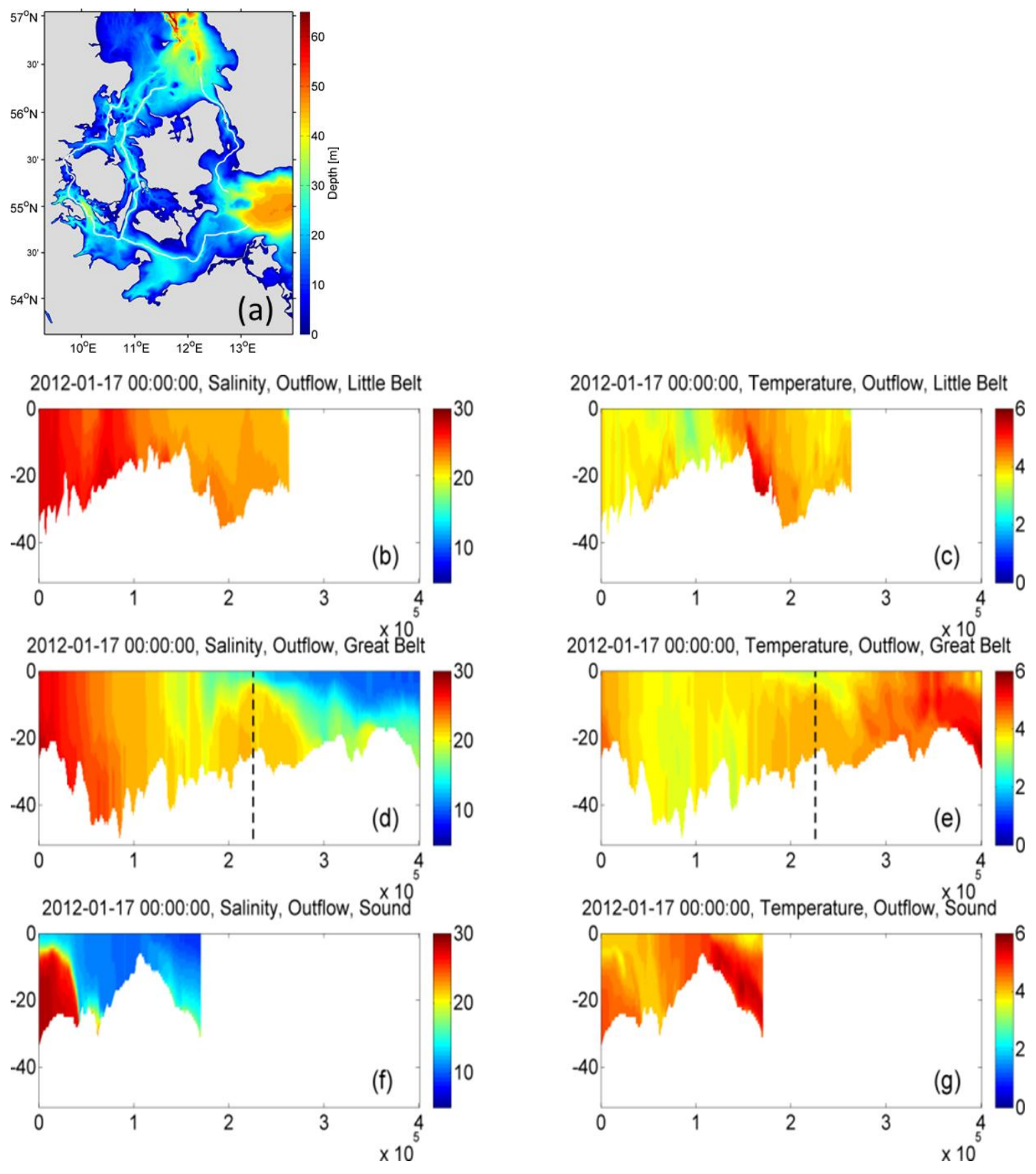


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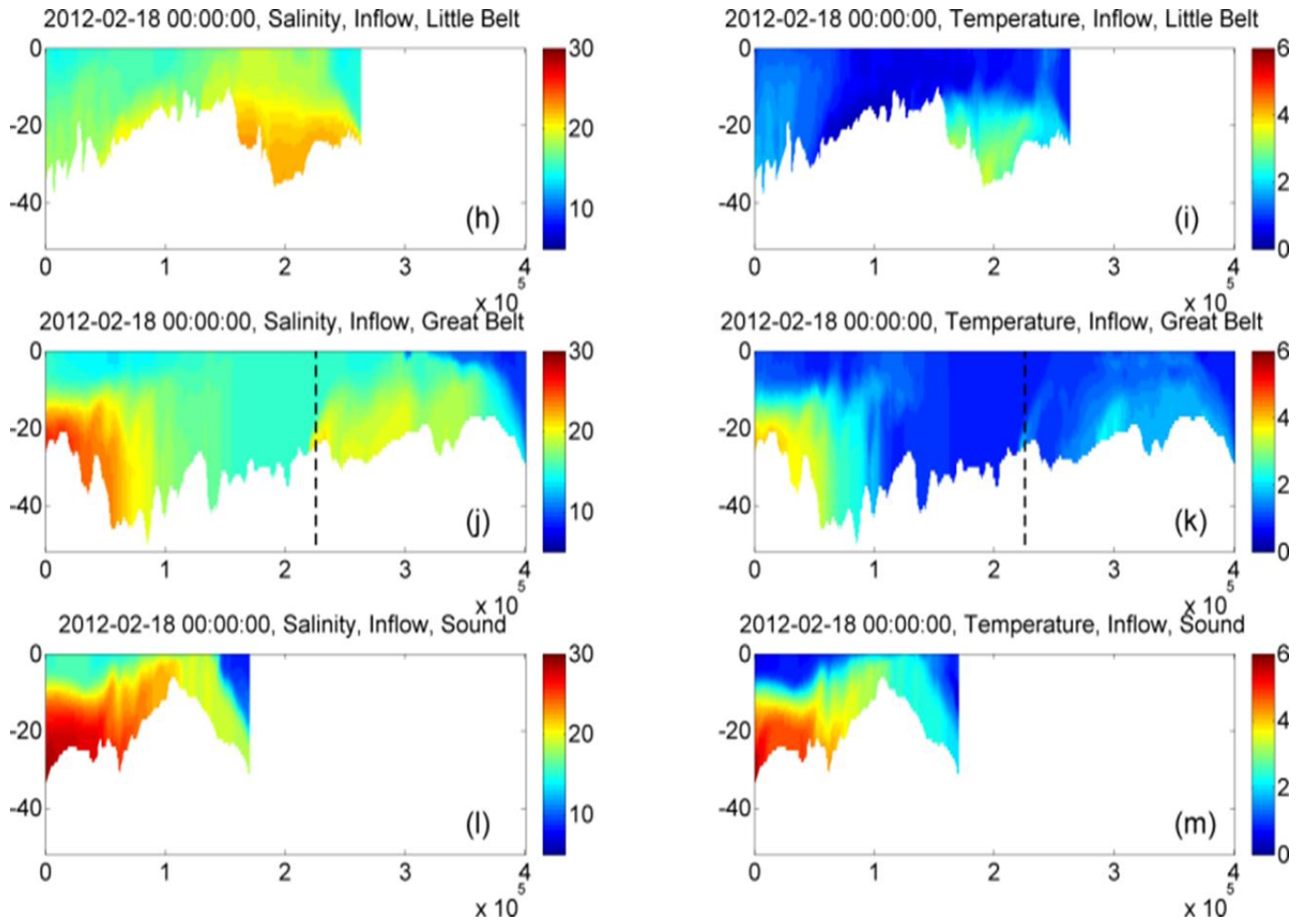


Fig. 9

Figure10

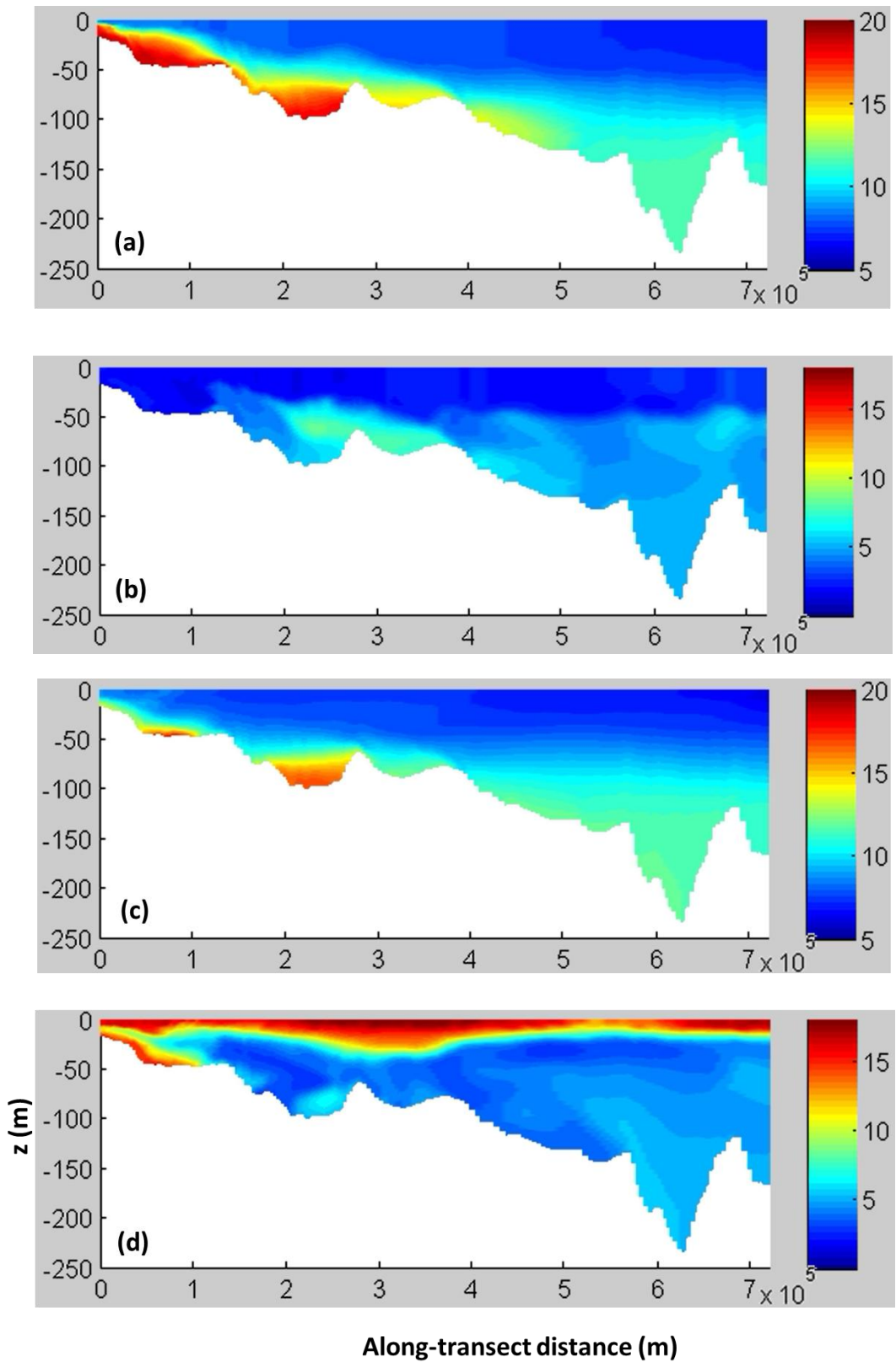


Fig. 10

FigureA1

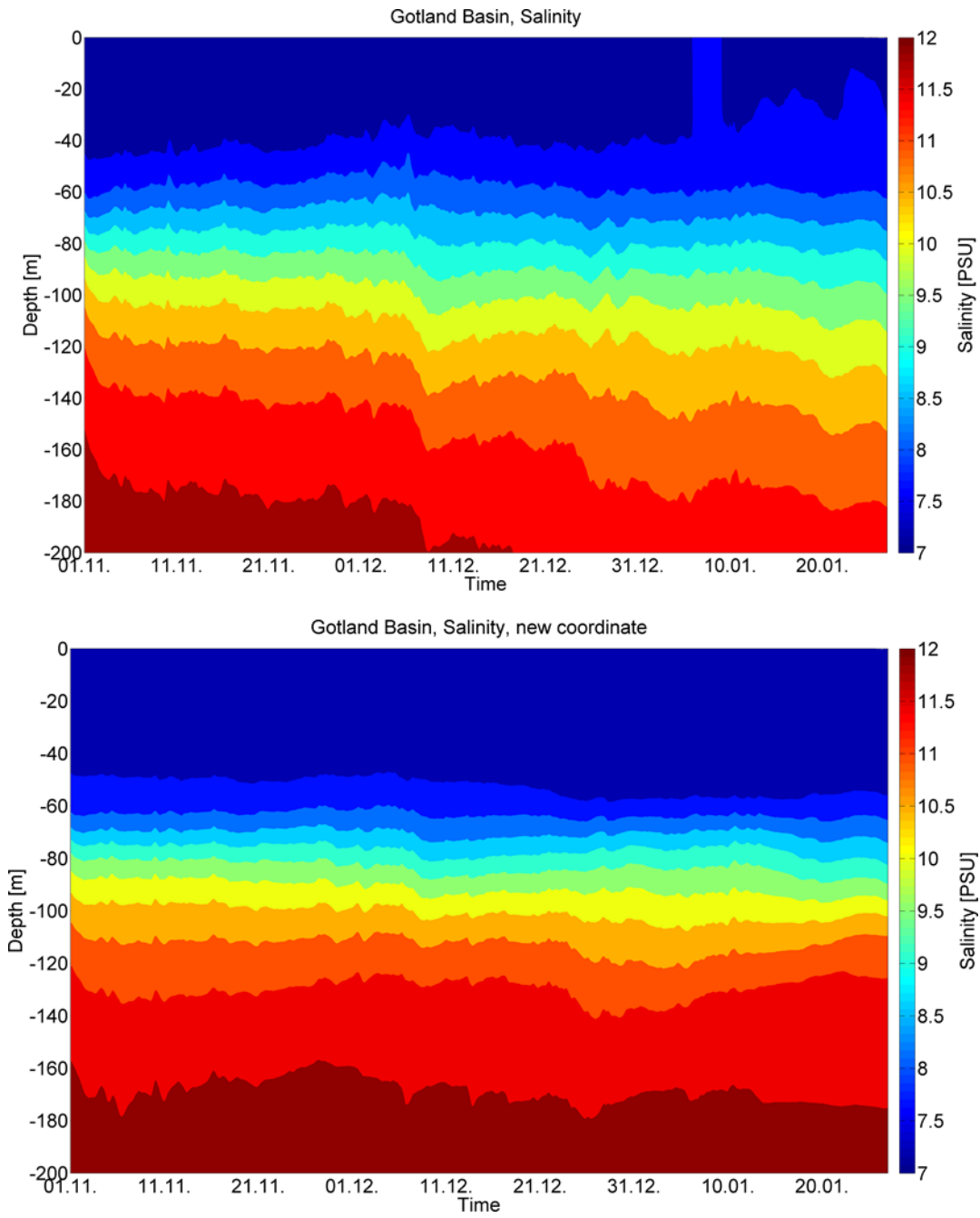


Fig. A1

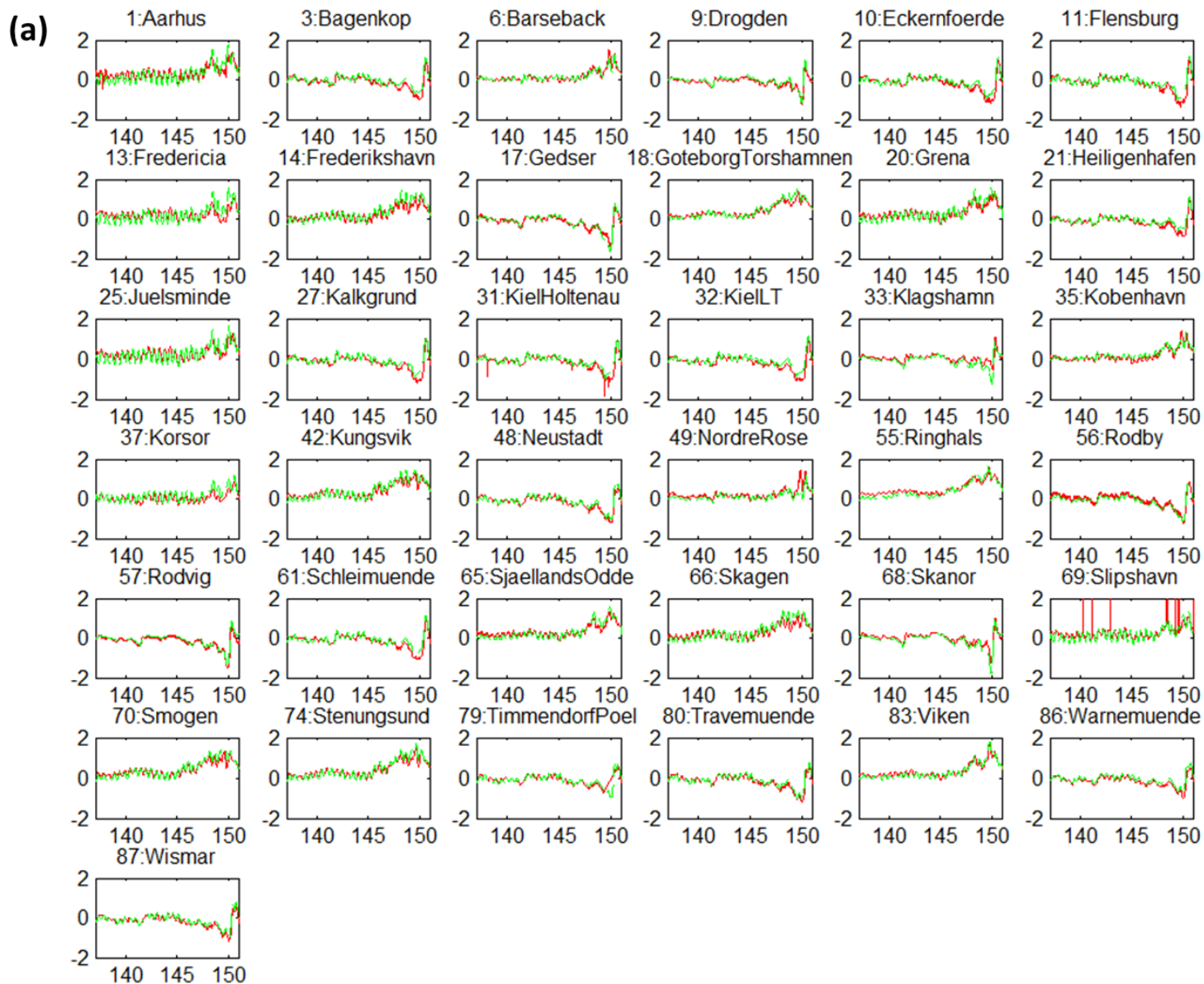


Fig. A2

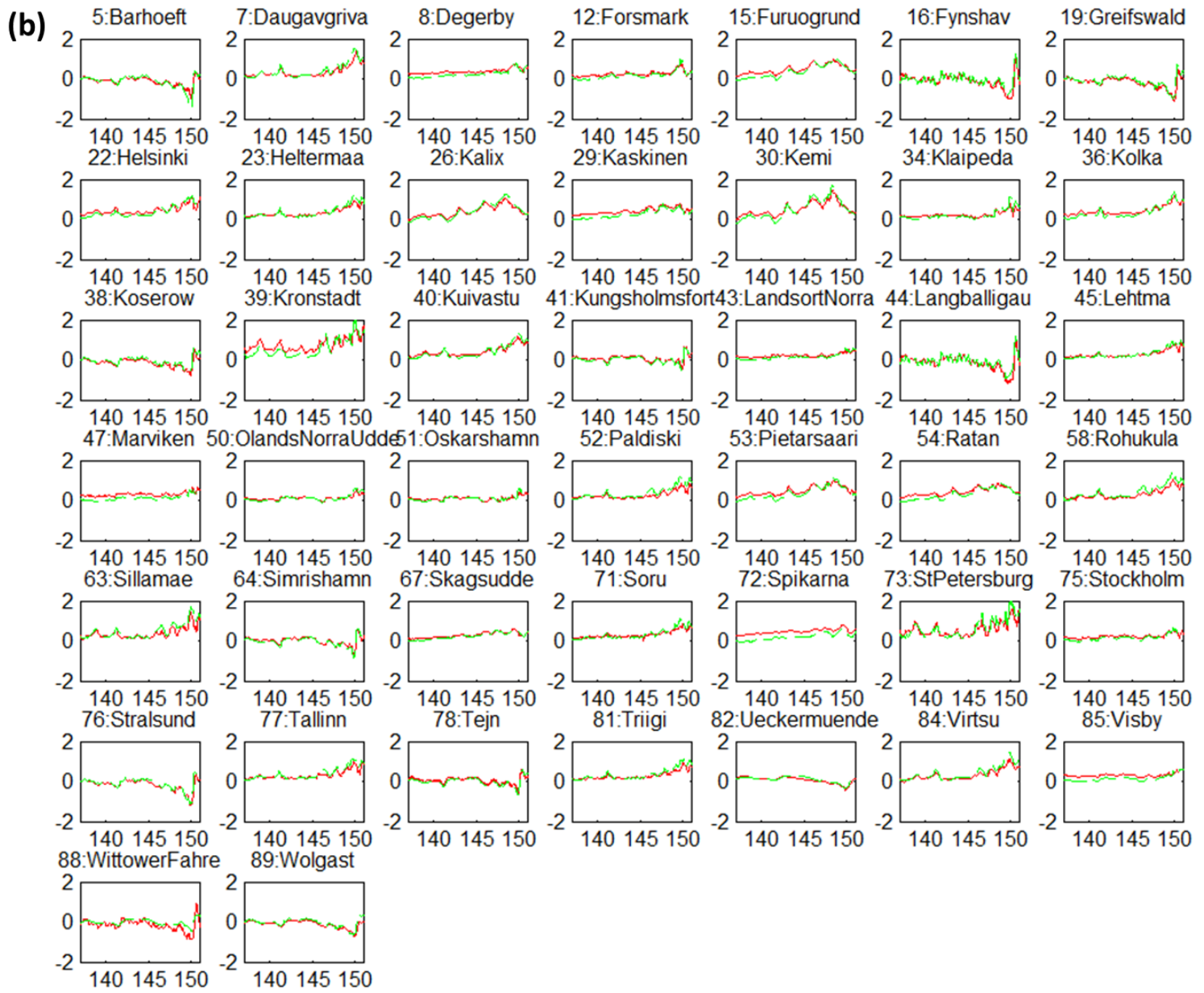


Fig. A2

(c)

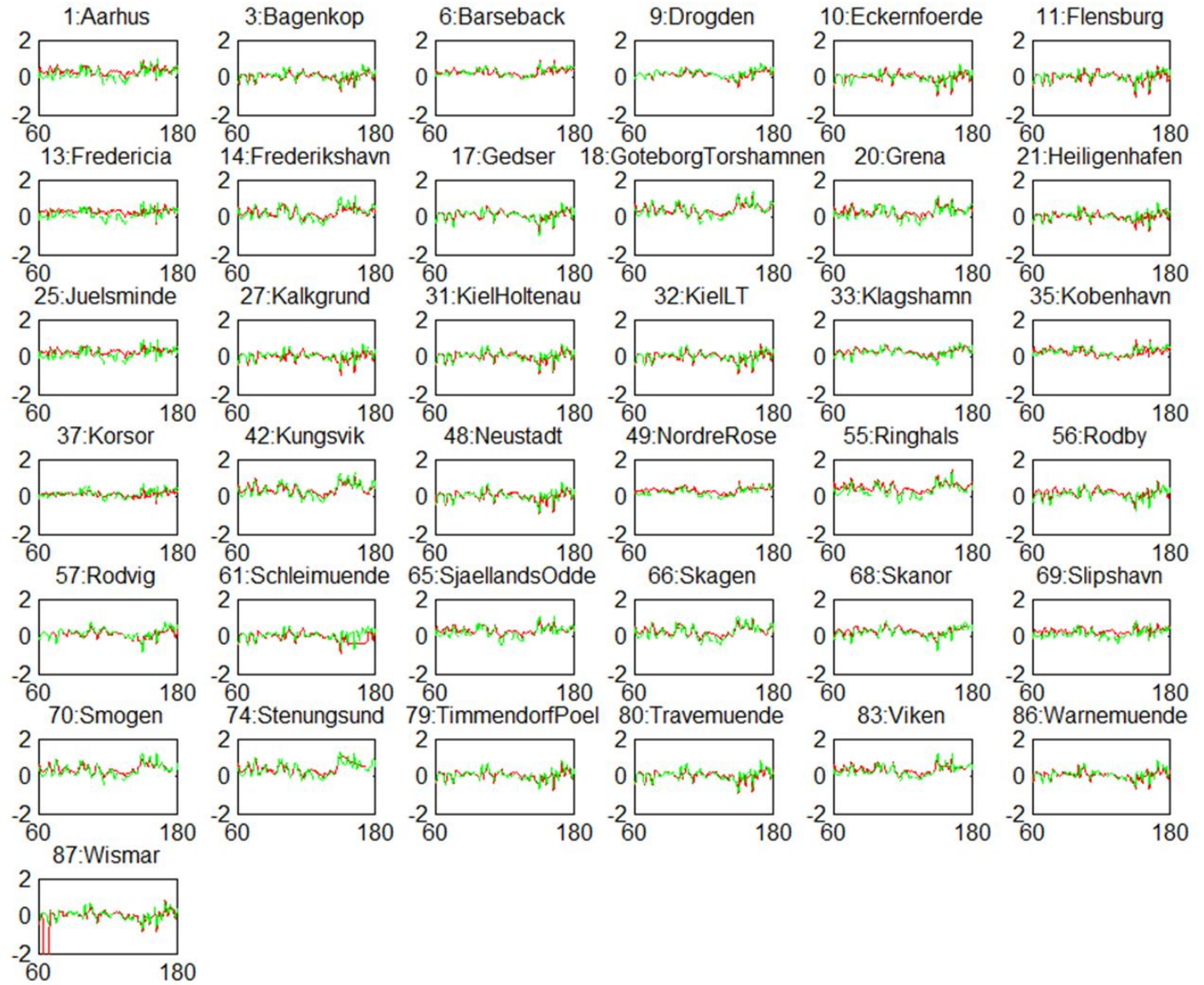


Fig. A2

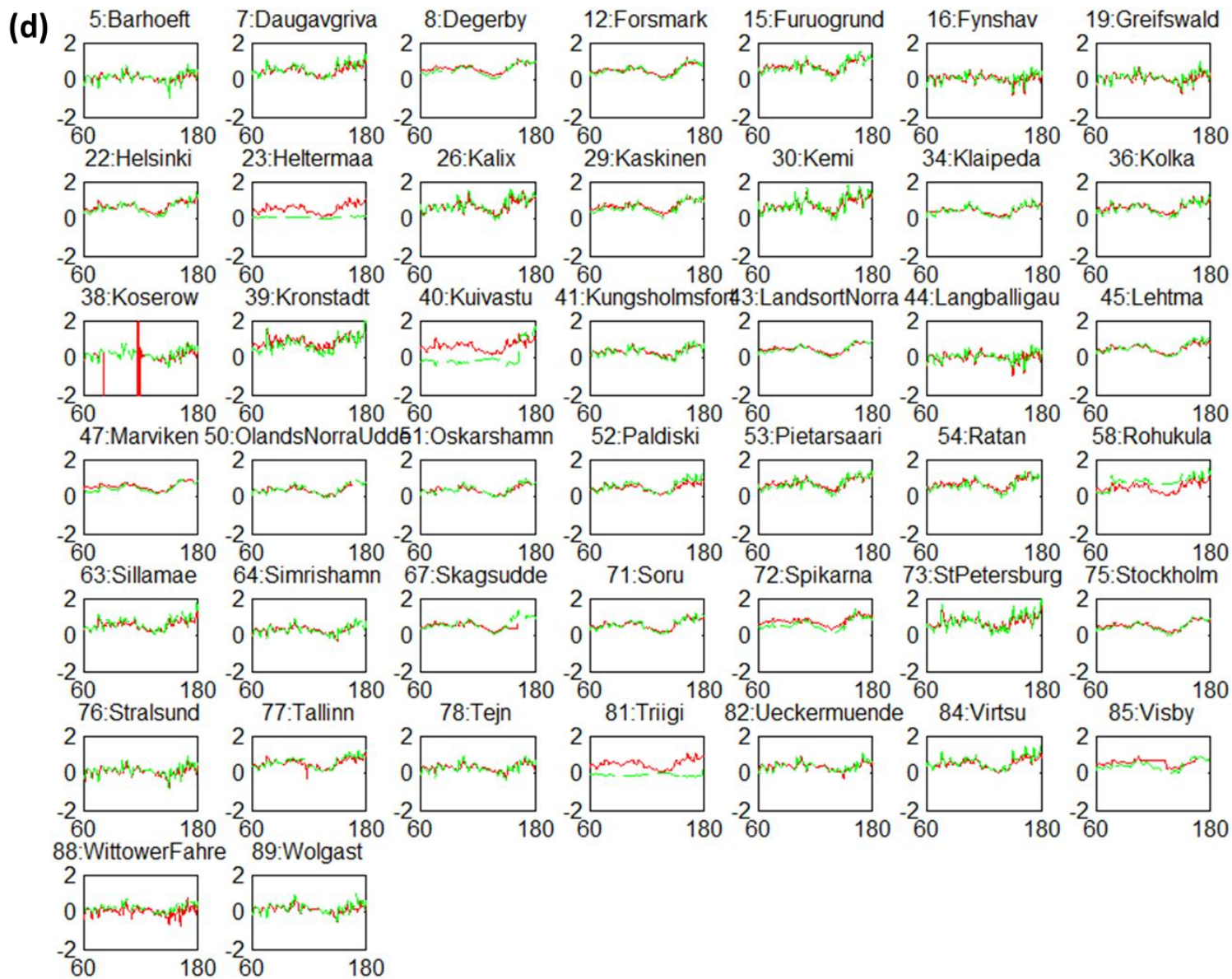


Fig. A2