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Impact of wave-dependent stress on storm surge simulations in the North Sea: ocean model evaluation against in situ and satellite observations

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Abstract

We investigate the impact of wave-dependent stress on surge modelling, from case studies in the North Sea, using a global ocean model forced with a wave-atmosphere coupled model. We select the storms with the largest surges and a range of sea state development from young to mature seas. The modelled surges are compared to tide gauges and altimeter data. The ocean model is able to accurately predict storm surges in coastal areas. The consistency of the model outputs, the altimeter, and the tide gauge data confirms the accuracy of altimeters for storm surge measurements. We show that using a wave-dependent rather than a wind-dependent only stress formulation gives more accurate surge simulations when the sea state is young and the sea rougher. Taking into account the waves in the stress formulation has a significant impact on the surges (up to 20 cm).

Keywords: air-sea exchanges, storm surges, wind stress, drag coefficient, wind-wave coupling, North Sea

1. Introduction

Storm surges are generated by atmospheric pressure gradient and wind stress. In coastal areas, the wind stress contribution is more effective due to shallow waters, water pileup along the coast, and resonant effects (Moon et al., 2009; Bertin et al., 2012). In addition, in nearshore areas, the radiation stress, which is the momentum flux carried by the waves, generates nearshore currents and wave setup (i.e. additional surge) when the waves dissipate (Bunya et al., 2010; Kim et al., 2010; Brown et al., 2010; Idier et al., 2012; Lee et al., 2013; Bertin et al., 2015; Thuy et al., 2017; Choi et al., 2018). Here, we focus on the impact of the wind stress on the surges. This study tackles the question ‘which impact has the sea state on the wind stress?’. To answer this, we determine if simulated surges are closer to observations when wind stress parameterization is wave-dependent.

The wind stress is usually parameterized using bulk formulae that express it as a function of the wind speed at a given height, generally 10 m above sea surface, and of a drag coefficient

$$\tau = \rho_a u_*^2 = \rho_a C_d U_{10}^2 \quad (1)$$

where u_* , C_d , and U_{10} are the friction velocity, the drag coefficient, and the wind speed at 10 m above the surface, respectively. Most formulations of the drag depend only on the wind speed (e.g. Moon et al., 2007; Edson et al., 2013; Peng and Li, 2015), whereas others include sea state

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51 parameters (e.g. Janssen, 1991; Moon et al., 2009). Despite years of research, the impact of sea
 52 state on drag remains inconclusive. Recently, Edson et al. (2013) concluded that the "COARE
 53 3.5 wind speed-dependent formulation matches the observations well without any wave
 54 information," whereas earlier works insisted on the importance of wave-enhanced drag for
 55 young waves (Mastenbroek et al., 1993).

56
 57 These contrasted results can be explained by the complexity of the problem. There are major
 58 issues that prevent a simple answer to the question ‘what is the wave impact on the wind stress,
 59 if any?’: (1) Wind stress measurements - and drag estimations - are difficult to perform,
 60 particularly at high winds. Uncertainties in observations could explain the variability of the drag
 61 at a given wind speed. Estimations from several recent field experiments based on direct in situ
 62 measurements (e.g. Black et al., 2007; Edson et al., 2013) or indirect ones (Powell et al., 2003;
 63 Jarosz et al., 2007; Holthuijsen et al., 2012), as well as laboratory tank measurements (e.g.
 64 Donelan et al., 2004; Takagaki et al., 2012) show differences in the drag up to a factor 2 at 30
 65 m/s (Pineau-Guillou et al., 2018). It is not known if these differences are due to measurement
 66 uncertainties or various environmental conditions (fetch, turning wind, bathymetry...). (2) Wind
 67 stress measurements are scarce. They generally come from short dedicated campaigns, with
 68 moderate winds often being lower than 20 m/s. There is clearly a lack of measurements at very
 69 high winds. (3) All the variables used to explain the drag variability are interdependent.
 70 Consequently, good correlations between the drag and variables such as wave age may mainly
 71 be due to self-correlation (Andreas, 2009). This contributes to a lack of confidence in these
 72 relationships in a part of the scientific community.

73
 74 The objective of this article is to investigate the impact of a wave-dependent stress on the surges.
 75 Here, we focus on the North Sea. The main idea is to simulate storms with various sea states
 76 (i.e. young and old) to estimate the sensitivity of surges to the sea state development (i.e. the
 77 wave age). Selected storms are simulated using wind-dependent only and wave-dependent
 78 stress parameterizations. The simulated surges are evaluated against observations (i.e. tide
 79 gauges and altimeters). The first part of the article describes the methods to compute the surges
 80 in models and observations. In the following part, we describe the case studies—namely, the
 81 storms selection, the numerical setup, and the validation data. Then, we analyze the results to
 82 estimate the impact of the wave-dependent stress on the storm surges. Finally, we discuss
 83 different points; among them is the difficulty of comparing the model with observations, as
 84 various processes contribute to the surges, particularly in coastal areas.

85 86 **2. Surges in models and observations**

87
 88 The surges are the differences between the water level and the tide prediction. Here, we describe
 89 the methods to compute the surges from the model and observations.

90 91 *2.1. Modelled surges*

92
 93 Surges are simulated with an ocean model forced with output from an atmosphere model. The
 94 ocean model resolves the classical Saint-Venant shallow water continuity and momentum
 95 equations in barotropic mode, formulated similarly as in Bertin et al. (2012)

$$96 \quad \frac{\partial \eta}{\partial t} + \vec{\nabla} \cdot \int_{-h}^{\eta} \vec{u} dz = 0, \quad (2)$$

$$97 \quad \frac{D\vec{u}}{Dt} = -f\vec{k} \times \vec{u} + \alpha g \vec{\nabla} \hat{\psi} - \frac{\vec{\nabla} P_a}{\rho} - g \vec{\nabla} \eta + \frac{\vec{\tau}_s - \vec{\tau}_b}{\rho(\eta+h)} \quad (3)$$

98
99

100 where η is the surface elevation, \vec{u} is the horizontal velocity, h is the bathymetry, f is the
 101 Coriolis parameter, α is the earth-elasticity factor, g is the mean gravitational acceleration, $\hat{\psi}$
 102 is the earth tidal potential, P_a is the sea level atmospheric pressure, ρ is the water density, τ_s is
 103 the surface stress, and τ_b is the bottom stress. The bottom stress is expressed as

$$104 \quad \vec{\tau}_b = \frac{C}{H} \|\vec{u}\| \vec{u} \quad (4)$$

105 where H is the mean local depth, and C is a dimensionless friction coefficient set as $2.5 \cdot 10^{-3}$
 106 (Lyard et al., 2006).

107 The water level variations are due to tide through the tidal potential ($\hat{\psi}$ in Eq. 3), wind through
 108 the wind stress (τ_s in Eq. 3), and atmospheric pressure through the pressure gradient ($\vec{\nabla}P_a$ in
 109 Eq. 3). Note that the wind stress effect is modulated with the bottom stress effect (τ_b in Eq. 3),
 110 whose influence is not studied here. We investigated the relative contribution of the wind stress
 111 and pressure gradient terms in Eq. 3 as $\vec{\tau}_s/\rho(\eta + h)$ and $\vec{\nabla}P_a/\rho$, respectively. In this case, the
 112 wind stress contribution increases in shallow waters, as it is divided by the water height.
 113 Comparison of the two terms shows that the wind stress term is largely dominant in the North
 114 Sea, that is, more than 90% of the sum of the two terms (Pineau-Guillou, 2018). This suggests
 115 that the currents are mainly driven by the wind, whereas the effect of the atmospheric pressure
 116 is negligible.

117 In the following, the surges are computed from simulations without tide (no tidal potential $\hat{\psi}$ in
 118 Eq. 3) and with atmospheric forcing only. As a consequence, the modelled surges correspond
 119 only to atmospheric surges; the surges due to wave breaking (i.e. wave setup) are not modelled
 120 here (see discussion in section 5). To compute the surges, we used the TUGO shallow water
 121 global ocean model, developed by LEGOS (Lyard et al., 2006). This model in barotropic mode
 122 resolves the classical shallow water continuity and momentum equations (Eqs 2 and 3).
 123 Following Lynch and Gray (1979), the model solves the generalized wave equation

$$124 \quad \frac{\partial(CE)}{\partial t} + c_0 CE = 0 \quad (5)$$

125 where $CE = 0$ is the continuity equation (Eq. 2), and c_0 is a relaxation coefficient towards the
 126 continuity equation. This gives more explicitly

$$127 \quad \frac{\partial^2 H}{\partial t^2} + \vec{\nabla} \cdot \frac{\partial H \vec{u}}{\partial t} + c_0 \left(\frac{\partial H}{\partial t} + \vec{\nabla} \cdot H \vec{u} \right) = 0 \quad (6)$$

128 where $H = \eta + h$ is the water height, and $\frac{\partial H \vec{u}}{\partial t}$ is formally replaced by using the momentum
 129 equations. This allows to make the elevation solver implicit, hence relaxing the CFL condition
 130 for time step. Actually, TUGO can use different solvers and discretization for the dynamical
 131 equations, but the most efficient so far in tides or storm surges modelling is the LGP1xLGP1
 132 (elevations and currents discretized at element vertices) generalized wave equations solver
 133 (hence being used in our study). TUGO is a reference model; it allowed the development of the
 134 tidal model FES2014 (Carrère et al., 2015), a worldwide reference model for tides, which is
 135 widely used in the scientific community. It also produces Dynamic Atmospheric Corrections to
 136 correct altimeter data from atmospheric effects (Carrère and Lyard, 2003). This correction is
 137 officially used by CNES and NASA for altimeter data processing.

147 2.2. Observed surges

148

149 Tide gauges measure the sea level. The surges are computed as the differences between the
150 observed and predicted sea level (Simon, 2007):

151

$$152 \quad \text{Surge}_{\text{TideGauge}} = \text{Observed sea level} - \text{Tide prediction} \quad (7)$$

153

154 We used 101 tide gauges in the North Sea (Figure 1), obtained thanks to Copernicus Marine
155 Environment Monitoring Service (CMEMS). Data temporal resolution is generally of 10
156 minutes (88% of the tide gauges) and sometimes 1 hour (12% of the tide gauges). We used the
157 Tidal ToolBox developed by LEGOS (Allain, 2013) to process the data and estimate the surges.
158 The method is the following; (1) a harmonic analysis is performed on the sea level observations
159 to estimate the harmonic constants (amplitude and phase) of the tidal constituents; (2) from
160 these harmonic constants, a prediction is computed over the same period as the observations;
161 (3) the surges are the differences between the observed and predicted sea levels (Eq. 7). Note
162 that this residual also includes the error associated with the prediction.

163

164 In addition to tide gauges, radar altimeters onboard satellites also measure the sea level—that
165 is, the instantaneous Sea Surface Height (SSH) above the ellipsoid. Data are processed by
166 providers to compute the Mean Sea Surface (MSS) and the Sea Level Anomaly (SLA). The
167 MSS corresponds to the mean of several years of SSH, eventually using several satellites. The
168 SLA is the difference between the instantaneous SSH and the MSS. Many geophysical and
169 environmental corrections are made to estimate SLA, and one among them is Dynamic
170 Atmospheric Correction (DAC) (Antony et al., 2014). This correction corresponds to the ocean
171 response to atmospheric forcing (atmospheric pressure and winds) and generally comes from
172 an ocean model elevations for high frequency part (e.g. TUGO, Carrère and Lyard (2003)) and
173 inverted barometer law for low-frequency part (e.g. using ECMWF atmospheric pressure
174 products). To be consistent with the model and the tide gauges, the surges from altimeters are
175 computed as the summation of the SLA and the DAC:

176

$$177 \quad \text{Surge}_{\text{Altimeter}} = \text{SLA} + \text{DAC}. \quad (8)$$

178

179 To compute the altimetric surges, we used the SLA and DAC from the JASON-2 1Hz X-
180 TRACK coastal product (Birol et al., 2016), developed by the Center of Topography of the
181 Ocean and Hydrosphere (CTOH/LEGOS, Toulouse). Along-track data have a temporal
182 resolution of 1 s, which corresponds to a spatial interval of about 6-7 km between points. X-
183 TRACK is a post-processing software which increases the SSH information derived from
184 satellite altimetry in the coastal ocean areas. Retrieved information along tracks come closer to
185 land, up to 5 km, against 10 km with the standard AVISO (Archiving, Validation and
186 Interpretation of Satellite Oceanographic data) product.

187

188 3. Case studies and modelling

189

190 Here, we describe the storms selection, the numerical setup, and the validation data.

191

192 3.1. Storms selection

193

194 The storms were selected according to the following criteria: 1) availability of good-quality
195 data at tide gauges as well as along tracks, 2) occurrence of maximum surges in the North Sea,
196 and 3) presence of various sea states. The objective was to select two storms with young sea

197 states, and two with old sea states. A common way to characterize the sea state is to consider
198 the wave age, following the expression $\xi = C_p/U_{10}$, where C_p is the phase velocity at the peak
199 of the wave spectrum, and U_{10} is the wind speed at 10 meters. When the wind has just started
200 blowing, the waves are short-period, steep, and short-crested. At this stage, the wind speed is
201 higher than the wave phase velocity, and the waves are growing. It is commonly considered
202 that the value of 1.2 separates the young and the old sea. In the North Sea during the storms,
203 the wave age is generally around 0.8 (Pineau-Guillou, 2018). In the following, sea state is
204 considered as "young" when the wave age is close to 0.8 and "old" when the wave age is greater
205 than 1.2 (swell). Note that the sea state quickly evolves spatially and temporally during the
206 storm; the reference to a "young" or "old" sea state for each storm characterizes the sea state
207 along the altimeter track during a short period of several minutes and is not representative of
208 the whole storm.

209
210 To select the storms, we analyzed tide gauges and JASON-2 altimeter data (Figure 1). The tide
211 gauge database consists of 101 tide gauges in the North Sea, from January 2012 to October
212 2017 (date of the extraction). The duration of observations depends on sites, and ranges from
213 less than 3 years to more than 5 years. Surges at tide gauges were computed following Eq. 7.
214 The altimeter database covers 8 years (2008-2015) of JASON-2 data along tracks. Surges along
215 tracks were computed following Eq. 8. Analysis of data led to select following storms: ex-
216 Gonzalo, Friedhelm, Felix, and Gunter (Table 1).

217
218 The tracks of the 4 selected storms are shown in Figure 2, and the surface winds during the
219 storms are shown in Figure 3. These storms were moving at a speed of around 15 to 20 km/h.
220 In December 2011, Friedhelm crossed the North East Atlantic. Its track is the southernmost
221 one. In the North Sea, the winds are very strong (up to 27 m/s), and the sea state is old, with
222 wave age larger than 1.2 along the altimeter track. In October 2014, ex-Gonzalo reached the
223 Northern part of the British Islands. This storm corresponds to the remnants of Category 4
224 Atlantic Hurricane Gonzalo. It is the strongest storm in terms of surges, but not in terms of
225 winds (no more than 22.7 m/s). The sea state is young with wave age around 0.9 along the
226 altimeter track. One of the characteristics of this storm is that the strong winds moved from the
227 west to the east as a nearly North-South front of around 1,000 km long (Figure 3 (b)). This
228 explains the discontinuity in the Mean Sea Level Pressure observed in the ex-Gonzalo track
229 over the North Sea, as the minimum moves along this front (red curve on Figure 2). In January
230 2015, Felix and Gunter crossed the North Atlantic. In fact, three storms succeeded: first Elon,
231 then Felix which crossed the North Sea on the 10th of January, and finally Gunter on the 12th of
232 January 2015. The winds were strong and reached 26.8 m/s (Table 1). For Felix, the sea state
233 was young with wave age close to 0.8 along the altimeter track, whereas for Gunter it was old
234 with wave age larger than 1.2. The old sea for Gunter can be easily explained by the succession
235 of three storms in a short period (few days). The Gunter track was farther north than the Felix
236 one, which may explain why the storm had less impact in term of surges.

237 238 *3.2. Numerical setup*

239
240 The ocean model is forced with output from a coupled wave-atmosphere model (Figure 4).
241 Here, we describe the ocean model, the atmosphere model, and the experiments.

242 243 *3.2.1. Ocean model*

244 We used the default configuration of the TUGO shallow water global ocean model (Lyard et
245 al., 2006), with FES2014 spatial grid (Figure 5). The unstructured space discretization allows
246 the increasing of the resolution in shallow waters, as well as along strong topographic gradient

247 areas. In the North Sea, the resolution varies from 10-15 km offshore to 4 km along the French
 248 and English coasts, and 2 km along the north of the Norwegian coasts. Note that the resolution
 249 is not refined in the southeast of the North Sea, due to a lack of detailed bathymetric information
 250 in this area (not enough available data).

251
 252 The ocean model is forced with (1) 10-m wind or (2) surface wind stress from the coupled wave
 253 atmosphere model (Figure 4). In the first case (1), the wind stress is computed from the ocean
 254 model bulk formula. The drag coefficient is expressed following Hellerman and Rosenstein
 255 (1983), which is a wind-only dependent formulation:

$$256 \quad 10^3 C_d = 0.934 + 0.788 \times 10^{-1} U_{10} + 0.868 \times 10^{-1} \Delta T - 0.616 \times 10^{-3} U_{10}^2 - 0.12 \times$$

$$257 \quad 10^{-2} \Delta T^2 - 0.214 \times 10^{-2} U_{10} (\Delta T) \quad (9)$$

259
 260 where U_{10} is the wind at 10 m, and ΔT is the air-sea temperature difference to take into account
 261 the stability effect. In the second case (2), the wind stress comes directly from the ECMWF-
 262 coupled wave-atmosphere model. Figure 6 shows the TUGO drag for $\Delta T = 0$ and the ECMWF
 263 drag, computed over the North Sea during two days for each storm: Fridhelm, ex-Gonzalo,
 264 Felix, and Gunter. For winds lower than 15 m/s, the TUGO drag is quite similar to the ECMWF
 265 one, but with no variability. For winds stronger than 15 m/s, the TUGO drag is generally lower
 266 than the ECMWF one. For a given wind speed, the variability of the ECMWF drag depends on
 267 the wave age (Figure 7), even if this dependency is not explicit (see section 3.2.2). Note that
 268 the TUGO drag is quite close to the ECMWF drag for old sea state, but lower for young sea
 269 state.

270 271 3.2.2. Atmosphere model

272 We used the ECMWF-coupled wave-atmosphere model - IFS (Integrated Forecasting System)
 273 - to generate atmospheric forcing (Figure 4). We conducted the simulations without data
 274 assimilation. The IFS CY41R1 cycle (ECMWF, 2015a) has a spatial resolution of around 16
 275 km (TL1279) and 137 vertical levels. It has been coupled with the spectral wave model
 276 ECWAM (ECMWF Wave Model, ECMWF (2015b)) since 1998. ECWAM uses a coarser
 277 horizontal resolution than IFS at around 28 km, with 36 directions and 36 frequencies
 278 exponentially spaced, with starting frequency 0.035 Hz and an increment of 1.1. In ECWAM,
 279 the source terms are $S = S_{in} + S_{nl} + S_{ds} + S_{bot}$, where S_{in} represents the wind input, S_{nl}
 280 represents the nonlinear wave-wave interactions, S_{ds} represents the dissipation due to
 281 whitecapping, and S_{bot} represents the bottom friction. The parameterizations of these source
 282 terms are discussed in ECMWF (2015b).

283 The wind stress is represented by classical bulk formulae (Eq. 1). We assume that the wind
 284 stress is in the wind direction—that is, the effects of wind-wave misalignment are not accounted
 285 here. The drag coefficient is expressed following Janssen (1991), which is a wave-dependent
 286 formulation. In neutral conditions, the drag coefficient can be expressed as

$$287 \quad C_d = \frac{\kappa^2}{\left[\log\left(\frac{10}{z_0}\right)\right]^2} \quad (10)$$

289
 290 where z_0 is the roughness length, and κ is von Kármán's constant (0.4). The roughness length
 291 is expressed as

$$292 \quad z_0 = \frac{0.11\nu}{u_*} + \alpha \frac{u_*^2}{g} \quad (11)$$

294

295 where ν is the kinematic viscosity, and α is the Charnock's parameter (Charnock, 1955). Note
 296 that the ECWAM wave model uses Eqs 10 and 11, but Eq. 11 is reduced to the second term,
 297 that is, roughness associated with an overall form drag of the wave field; the first term, that is,
 298 roughness associated to the viscous properties of the flow, is computed in the IFS atmosphere
 299 model. The modification of the roughness length z_0 impacts the drag coefficient and the wind
 300 stress (Eqs 10 and 1), then the sea level (Eqs 2 and 3), but also the wind profile. Indeed, when
 301 roughness increases, friction also increases, and this slows down the wind (Pineau-Guillou et
 302 al., 2018). Janssen (1991) parameterized the quasi-linear wave growth effect as an effective
 303 larger Charnock parameter, expressed as a function of the wave-induced stress τ_w

$$304 \quad \alpha = \frac{\alpha_0}{\sqrt{1 - \frac{\tau_w}{\tau}}} \quad (12)$$

306 with $\alpha_0 = 0.006$. The wave-induced stress τ_w is the momentum flux transferred from the
 307 atmosphere to the waves. It can be related to the wind-wave growth parameter β and the
 308 directional wave spectrum $E(f, \theta)$

$$309 \quad \tau_w = \rho g \int_0^\infty \int_0^{2\pi} \beta(f, \theta) E(f, \theta) / (c(f)) df d\theta \quad (13)$$

312 where θ is the direction, f is the relative wave frequency, and $c(f)$ is the phase speed, which
 313 is a function of frequency (Janssen, 2004). The wave-growth parameter is expressed as $\beta =$
 314 $(\beta_m / \kappa^2) \mu \ln^4(\mu)$, $\mu \leq 1$, where β_m is a constant (1.2), and μ is the dimensionless critical
 315 height (ECMWF, 2015b). In Eq. 13, the frequency f is integrated from 0 to a high-frequency
 316 limit f_c (prognostic part of the wave spectrum) using the discretized spectrum. Beyond f_c
 317 (diagnostic part of the wave spectrum), the shape of the spectrum is assumed, and the resulting
 318 integral can be evaluated using a simple integration scheme (ECMWF, 2015b). It is assumed
 319 that the diagnostic part of the wave spectrum is given as $E(f, \theta) = E(f_c, \theta) (f/f_c)^{-5}$ for $f >$
 320 f_c . The high frequency limit f_c is set as $f_c = \min(f_{max}, 2.5 f_{windsea})$, where f_{max} is the
 321 maximum discretized frequency, and $f_{windsea}$ is the mean frequency of the modelled wind sea
 322 (ECMWF, 2015b).

324 The ocean model is forced by the atmospheric stress τ . However, a part of the atmospheric
 325 stress is going into the waves. The momentum flux going into the ocean τ_{oc} is the sum of two
 326 contributions (Figure 8): the part of the atmospheric flux which was not used to generate the
 327 waves $\tau_o = \tau - \tau_w$, and the momentum flux transferred from the waves to the ocean by
 328 dissipation τ_{diss} (ECMWF, 2015b)

$$329 \quad \tau_{oc} = \tau_o + \tau_{diss} = \tau - \tau_w + \tau_{diss}. \quad (14)$$

332 A more correct approach would be to force the ocean model with τ_{oc} rather than τ . The
 333 normalized stress going into the ocean corresponds to the ratio τ_{oc}/τ (output parameter of the
 334 ECMWF operational version, but not available in the research version of IFS we used). The
 335 normalized stress is lower than 1 when the waves are growing, and greater than 1 when they
 336 are dissipating. It is globally close to 1 but can reach values as high as 1.5 under extreme
 337 conditions, such as with a passing front (Janssen, 2012). In the ex-Gonzalo case study, it could
 338 locally be greater than 2 when the front was passing (Pineau-Guillou, 2018). The strong
 339 gradients suggest a potential impact on the ocean model. Curcic (2015) also investigated the
 340 ratio between the oceanic and atmospheric stress in tropical cyclones; he found typical values
 341 between 0.85 and 1 depending on the wave state.

343

344 3.2.3. Experiments

345 We simulated the 4 selected storms (ex-Gonzalo, Friedhelm, Felix, and Gunter) with two stress
346 parameterizations—that is, the wind-dependent and the wave-dependent parametrizations. Note
347 that Felix and Gunter are in the same simulation as they follow each other. For each storm, the
348 model was initialized for at least 15 days, with winds and atmospheric pressure coming from
349 the ECMWF operational 1 h forecasts (i.e. combining hourly operational forecasts computed
350 twice a day at 00:00 and at 12:00). Once initialized, each storm simulation lasted 5 days and
351 was forced by (1) the 10 m-wind or (2) directly by the wind stress, and the atmospheric pressure,
352 with a 1h temporal resolution. In this case, the wind stress comes from the atmosphere model.
353 When forced by the 10 m-wind (1), the drag is a wind-dependent formulation, computed by
354 TUGO (Eq. 9 with $\Delta T = 0$, Hellerman and Rosenstein (1983)), whereas when forced by the
355 wind stress (2), the drag is a wave-dependent formulation, which has seen the waves through
356 IFS/WAM coupling (Eqs 10, 11, 12 and 13, Janssen (1991)).

357

358 3.3. Validation data

359

360 Among the 101 tide gauges, 22 tide gauges were selected for comparison with the model (Figure
361 1). The following were the criteria: 1) the tide gauges must have data available during the
362 storms, 2) the tide gauges must open up to the ocean, rather than at the end of a bay, 3) a
363 maximum number of the tide gauges must be offshore, where processes are different from
364 harbors, and 4) a maximum number of tide gauges must be located along the tracks. The
365 JASON-2 tracks with the maximum surges are tracks 170 for Friedhelm, 61 for ex-Gonzalo, 94
366 for Felix, and 170 for Gunter (Figure 1). All the data (tide gauges and JASON-2) were processed
367 as described in section 2.

368

369 4. Results

370

371 To investigate the impact of the waves, we compared the surges with wind-dependent
372 (Hellerman and Rosenstein, 1983) and wave-dependent (Janssen, 1991) parameterization
373 during four storms—two with a young sea state (ex-Gonzalo and Felix) and two with an old sea
374 state (Friedhelm and Gunter). Note that Felix and Gunter followed each other, and correspond
375 to the same simulation. We compared the simulated surges with observations, namely tide
376 gauges and JASON-2 altimetric data.

377

378 Figure 9 shows the surge comparison between the model and 3 tide gauges (representative of
379 the whole) for ex-Gonzalo, Felix, and Gunter (no data were available for Friedhelm in 2011, as
380 the CMEMS tide gauge database starts only in 2012); Table 2 summarizes the corresponding
381 errors between the model and the tides gauges: bias, Root Mean Square Error (RMSE), and
382 Peak Error (defined as the difference between the maximum observed and modelled surge).
383 Finally, Figures 10 and 11 show the surge comparison between the model and JASON-2
384 altimetric data, for ex-Gonzalo and Friedhelm. Note that the grey shaded area corresponds to
385 deep waters, where the wind stress effect is lower.

386

387 Analysis of ex-Gonzalo (young sea) and Friedhelm (old sea) gives the following results.

388

389 The first result is that globally, the model matches very well with the observations. For the
390 wave-dependent parameterization, on average, the bias between the model and all the tide
391 gauges is close to zero, the RMSE is 0.12 m, and the Peak Error is -0.09 m (Table 2). The errors
392 between the model and JASON-2 are lower than the errors between the model and the tide

393 gauges: the bias is close to zero, and the RMSE is 0.08 m, whereas the surge ranges up to 1.40
394 m (Figure 10). Note also the very good agreement between the model, JASON-2 data, and the
395 tide gauge situated along the track (D151TG, see Figure 1 for the tide gauges location).
396 Unfortunately, the tide gauge Cromer located on the northeast coast of England at the end of
397 the track did not record data during this storm. For Friedhelm, the model also matches the
398 altimeter quite well, but not as well as for ex-Gonzalo. For the wave-dependent
399 parameterization, the bias and RMSE reach -0.12 m and 0.08 m, respectively (Figure 11). The
400 differences could be due to uncertainties in altimeter corrections, such as geophysical
401 corrections (for example tide). However, note that the agreement between the model and the
402 altimeter is very good in shallow waters, where the wind stress effect is the most significant
403 (Figure 11 (b)). To conclude, the errors between the model and the observations (tide gauges
404 and altimeter data) for these two storms are small enough to confirm the capability of a global
405 model to accurately predict storm surges, even in the coastal areas, when its spatial resolution
406 is fine enough to catch the storm size. These results also confirm the capability of altimeters to
407 accurately measure surges (Antony et al., 2014).

408
409 The second result is that the wave-dependent parameterization yields higher surges only when
410 the sea state is young (ex-Gonzalo, Figures 9 (a) and 10 (b)). Otherwise, the surges are similar,
411 regardless of the parameterization (Friedhlem, Figure 11 (b)). Physically, this is not surprising
412 as old sea corresponds to a situation where waves are no longer rapidly growing resulting in
413 sharp reduction of the momentum flux from the atmosphere to the waves. When the sea state is
414 old, the drag coefficients from the two parameterizations are close to each other (Figure 7), and
415 the surges are then similar. However, in the presence of young and steep waves, the drag
416 increases with Janssen's parameterization due to higher values of the Charnock parameter
417 (Figure 12 (c) and (d)). This yields higher drag than the wind-dependent formulation, and later
418 higher wind stress (Figure 12 (e) and (f)) and higher surges. The differences between these two
419 parameterizations corresponds to the effect of the waves on the surges. This difference reaches
420 25 cm at Lowestoft (Figure 9 (a)) and 20 cm along Jason-2 track (Figure 10 (b)). We note the
421 very good agreement in terms of Significant Wave Height (SWH) between the ECMWF
422 coupled wave-atmosphere model and JASON-2 altimeter (Piolle et al., 2019) for ex-Gonzalo
423 (Figure 12 (b) and 13 (b)), and for Gunter (Figure 13 (d)). The very strong SWH gradient along
424 the track for ex-Gonzalo corresponds to the passage of the front (Figure 13 (b)).

425
426 The third result is that the wave-dependent parameterization is closer to the observations than
427 the wind dependent one. These results are consistent with those previously obtained by
428 Mastenbroek et al. (1993), Nicolle et al. (2009), and Bertin et al. (2015). Along the JASON-2
429 track, the RMSE is reduced from 0.13 m to 0.08 m (Figure 10 (b)). On average, in the 21 tide
430 gauges, the Peak Error is reduced from -0.21 m to -0.09 m (Table 2). However, in some tide
431 gauges, the surges are still underestimated. The tides gauges can be separated into three groups:

- 432 • Group 1: a first group of 11 tide gauges (in blue in Figure 1), where the surges with the
433 wave-dependent stress match well with observations, such as Lowestoft and
434 EuroplatformTG in Figure 9 (a). This corresponds to the 4 offshore tide gauges
435 (F3platformTG, D151TG, EuroplatformTG, VlakteVdRaantG) as well as 7 other tide
436 gauges onshore.
- 437 • Group 2: a second group of 5 tide gauges (in green in Figure 1), where the surges with
438 the wave-dependent stress are still underestimated, such as Whitby in Figure 9 (a).
- 439 • Group 3: a third group of 5 tide gauges (in black in Figure 1), where the effect of the
440 parameterization is not significant; this corresponds to tide gauges located in the
441 northern part of the North Sea, where surges are smaller than 0.50 m. In this part, the
442 bathymetry ranges from 50 to 200 m, and the effect of wind stress is smaller than in the

443 southern part, with shallow waters. That may explain the non-significant differences
444 between the two parameterizations.

445 Note that in the second group where surges are still underestimated, there are no tide gauges
446 offshore. This underestimation is probably partly due to processes taking place in the very
447 nearshore, and not modelled by TUGO (e.g. wave setup, see the discussion in section 5). This
448 could suggest that comparisons with the altimeter are better, as the tracks offshore are not
449 contaminated by coastal processes such as wave setup. Another reason that could explain this
450 negative bias is the lack of spatial resolution near the coast (only 2 km). Experiments show that
451 atmospheric surges may increase from 0.1 to 0.3 between 10 m isobath and the shoreline
452 (personal communication from X. Bertin); here, the spatial resolution is not fine enough to
453 correctly represent this increase.

454
455 After the analysis of ex-Gonzalo (young sea) and Friedhelm (old sea), analysis of Felix (young
456 sea) and Gunter (old sea) give partly similar results. (1) The model still matches quite well with
457 the tide gauges (bias, RMSE, and Peak Error are respectively 0 cm, 19 cm, and -10 cm for the
458 wave-dependent parameterization; see Table 2). (2) The wave-dependent parameterization still
459 yields to higher surges when the sea state is young. The impact of the waves on the surges
460 reaches around 11 cm at Europlatform and around 20 cm along Jason-2 track (not shown). (3)
461 We find also that the wave-dependent parameterization is closer to the tide gauge observations,
462 reducing on average the Peak Error from -0.14 m to -0.10 m. However, comparison between
463 the model and the altimeter is not as good as for ex-Gonzalo and Friedhelm (not shown), and it
464 is difficult to conclude which parameterization is the most appropriate. This suggests that the
465 number of case studies should be increased, to give more confidence in our conclusions.

466 467 **5. Discussion**

468
469 Here, we discuss the influence of the wind direction, the processes contributing to the surges,
470 and the impact on the altimetric corrections.

471 472 473 *5.1. Impact of the wind direction*

474
475 Analysis of tide gauge and altimeter data revealed that ex-Gonzalo was the storm with the
476 highest surge, whereas it was the one with the weakest winds (only 23 m/s in the North Sea,
477 against 27 m/s for Friedhelm, see Table 1). This is mainly due to the wind direction. Figure 14
478 shows wind roses in the middle of the North Sea (4°E 56°N), during the 5-day simulations of
479 the storms. For ex-Gonzalo (Figure 14 (a)), strong winds were mainly from the northwest
480 direction, pushing the waters along the southern coast of the North Sea, whereas for other
481 storms, strong winds came mainly from the west (see also Figure 3).

482 This is probably not the only explanation for ex-Gonzalo high surges. The track of this storm
483 is the southernmost one; the storm crosses the south of the North Sea, and the shallow waters
484 may enhance the wind stress contribution.

485 486 *5.2. Processes contributing to the surges*

487
488 Even if the wave-dependent parameterization yields higher surges, the modelled surges are still
489 underestimated compared with some tide gauges (5 over 21 for ex-Gonzalo storm). The
490 comparison between modelled and observed surges is complicated by the different processes
491 that contribute to the surge and that are not always modelled (see Table 3). Generally, the
492 dominant effect is the atmospheric forcing (Mean Sea Level Pressure and winds). This

493 contribution is commonly of the order of 50 cm, but can exceed 1 m in severe storms and
494 hurricanes, as in the case of Xynthia in February 2010 at La Rochelle (Pineau-Guillou et al.,
495 2012; Bertin et al., 2012). In case of progression in very shallow waters, as for example in the
496 South of the North Sea, the surges can reach up to 2 or 3 m. This atmospheric contribution is
497 taken into account in the ocean model through the atmospheric forcing.

498
499 Another important contribution to the total surge - which is here not taken into account - is the
500 wave setup, that is, the surge due to wave dissipation (mainly by wave breaking) in the
501 nearshore areas (Bunya et al., 2010; Kim et al., 2010; Brown et al., 2010; Idier et al., 2012; Lee
502 et al., 2013; Bertin et al., 2015; Thuy et al., 2017; Choi et al., 2018). To model the wave setup,
503 the radiation stress has to be introduced. The radiation stress is the momentum flux carried by
504 the waves. When the waves dissipate, such as by wave breaking or strong bottom friction, this
505 generates nearshore currents and an additional surge, the wave setup. Its contribution to the
506 total surge may be significant. For instance, values of 0.5 to 1.5 m were reported in Liverpool
507 Bay (Brown et al., 2010), more than 0.5 m (i.e. 50% or more of the total surge) in the southern
508 part of the Bay of Biscay (Idier et al., 2012), and 10 to 20 cm in the central part of the Bay of
509 Biscay (Bertin et al., 2015). In coastal areas, it can contribute up to 80% of the total storm surge
510 (Pedreros et al., 2018). During typhoons, it can contribute up to 15% following Thuy et al.
511 (2017) and even 40% following Kim et al. (2010). In the North Sea, Choi et al. (2018) reported
512 contributions of around 20 cm and 10% of the total surge during 1953 Big Flood. To capture
513 the wave setup well, (1) the ocean model has to be coupled one-way or two-way with a wave
514 model to take into account the radiation stress, and (2) the grid spatial resolution has to be high
515 enough, that is, around 10 m. As a consequence, due to a too coarse grid (3 to 15 km in the
516 North Sea), the wave setup is not modelled in this study (as is generally done in global and
517 regional models). This is a limitation for comparison with coastal tide gauges, but not with
518 altimeter, as wave setup is close to zero far from the coast. Note that the present study would
519 benefit from more tide gauge observations offshore (where wave setup is negligible).

520
521 Other contributions are the waves whose signature may be significant at the surface when
522 propagating in coastal areas, such as meteo-tsunami, infragravity waves - only few cm in deep
523 ocean (Aucan and Ardhuin, 2013) but can reach more than 1 m in coastal areas (Sheremet et
524 al., 2014), internal solitary waves (can reach 20 cm in coastal areas) as well as internal waves,
525 rogue waves, and tsunamis - surge due to an earthquake, landslide, or volcanic eruption. Finally,
526 seiches - resonance phenomena in closed or semi-closed basins - also contribute to the surges.
527 Their amplitude can be significant in harbors and sometimes reach several tens of centimeters.

528 529 *5.3. Impact on the altimetric corrections*

530
531 The accuracy of the simulated storm surges is essential, as it directly impacts the accuracy of
532 Sea Level Anomaly products through the Dynamic Atmospheric Correction (Eq. 8). Figure 15
533 shows the differences between the default SLA from CTOH (blue curve) and the new
534 reconstructed one (red curve) with the DAC taking into account the waves, that is, from the
535 TUGO simulation forced with the ECMWF wind stress. The new reconstructed SLA is on
536 average closer to zero. The SLA has been improved by removing some surge residual due to
537 atmospheric effect. The difference between the two SLA (the native and the reconstructed one)
538 reaches 40 cm near the coast. Note that in this 40 cm, around 20 cm may be attributed to the
539 waves impact, and the other 20 cm are probably due to a better ocean model resolution and a
540 better temporal atmospheric forcing (1h versus 6h in the DAC product from CTOH).

541 542 **6. Conclusions**

543
544 We investigated the impact of a wave-dependent stress on surge modelling. During ex-Gonzalo
545 and Friedhelm storms, which are characterized respectively by young and old sea states, we
546 compared simulated surges with wind-dependent and wave-dependent stress (Hellerman and
547 Rosenstein, 1983; Janssen, 1991). We compared the results with tide gauges and altimetric data.
548

549 We showed that the global ocean model accurately predicts storm surges in coastal areas
550 (RMSE of 0.12 m). This can be attributed partly to the unstructured grid, which allows
551 increasing the resolution in the shallow waters. The consistency between the model, the
552 altimeter, and the tide gauges also confirms the capability of altimeters to accurately measure
553 surges (RMSE of 0.08 m along the track).
554

555 We showed that when the sea state is old, the wind-dependent formulation Hellerman and
556 Rosenstein (1983) is appropriate. However, when the sea becomes younger and rougher, the
557 wind stress increases, and a wave-dependent formulation (here, Janssen (1991)) is more
558 appropriate. This reduces the Peak Error significantly (e.g. from 0.21 m to 0.09 m). The waves
559 effect on the surge can reach 20 to 25 cm. This result is consistent with previous studies
560 (Mastenbroek et al., 1993; Nicolle et al., 2009; Bertin et al., 2015). However, the number of
561 case studies should be increased to confirm these conclusions. Indeed, taking into the waves
562 allows to obtain surges closer to observations, but could be a way to compensate other errors.
563 For example, Moon et al. (2009) concluded that Mastenbroek et al. (1993) obtained good
564 simulated surges with overestimated drag by compensating surge error due to a too coarse grid.
565

566 This work underlines the lack of consistency of the drag between the wave, atmosphere, and
567 ocean models (van Nieuwkoop et al., 2015). One recommendation could be to force the ocean
568 model with the wind stress from a coupled wave-atmosphere model, which has seen the waves
569 (e.g. ECMWF model). This would yield to (1) more consistency between the drag from the
570 ocean and the atmosphere models and (2) improvement of the storm surges taking into account
571 the wave effect. To go even further, we would recommend forcing the ocean model with the
572 stress going into the ocean τ_{oc} rather than the atmospheric stress τ .
573

574 In this study, we showed that the ocean model better matches with tide gauges offshore
575 (compared with onshore) and altimeter tracks. This is probably due to coastal effects in tide
576 gauges (e.g. wave setup) that are not modelled in the TUGO ocean model and that are not seen
577 with the altimeters (as the tracks are offshore). This suggests that tide gauges should not always
578 be considered as a reference, and that what we generally call "errors" between model/tide
579 gauges or altimeter/tide gauges also includes local coastal processes. Consequently, we should
580 not always want the model (or altimeter) to match perfectly with tide gauges.
581

582 Finally, we should mention several limitations in this study. (1) Results are specific to the
583 ECMWF and TUGO parameterizations used here (Janssen, 1991; Hellerman and Rosenstein,
584 1983). (2) The wind stress is mainly supported by the capillary and short gravity waves, that is,
585 the tail of the spectrum, which is crudely represented in wave models today (Kudryavtsev et
586 al., 2014; Peureux and Ardhuin, 2016). Dedicated efforts are being made to improve wave
587 breaking parameterization, in order to also improve the tail of the spectrum. There are also
588 many uncertainties of the wave growth parameter β . (3) We assume that the wind stress is in
589 the wind direction, as it is mainly supported by high-frequency waves which respond quickly
590 to changes in the wind direction (Janssen, 2004). Possible wind-wave misalignment is not
591 accounted here. (4) The ocean model is forced by the atmospheric stress τ , whereas a more
592 correct approach would be to force it with the stress going into the ocean τ_{oc} . (5) The wave

593 setup, that is, the surge generated by the wave dissipation (e.g. wave breaking), is not accounted
594 here. It may reach up to 20 cm in the North Sea (Choi et al., 2018). (6) We have investigated
595 the wind stress sensitivity on the surges, but other factors impact the surges, such as grid spatial
596 resolution, bottom friction, bathymetry, and coastal geometry (Mao and Xia, 2017). Despite we
597 obtain modelled surges closer to the observations, there are still some possible compensation
598 of errors due to uncertainties on other modelling terms. Note that we did not go further in
599 investigating how these other factors may impact the surge dynamics, as this is not within the
600 scope of the present paper. (7) The ocean model is here forced by the atmosphere model. Two-
601 way ocean-atmosphere coupling would probably affect the resulting water levels, as both the
602 wind U_{10} and the atmospheric stress τ will be modified to account for the moving ocean surface
603 (Hersbach and Bidlot, 2009).

604
605

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607

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615 careful reading and their constructive comments, which allowed us to improve this article.

616

617 Figure 1: Tide gauges and JASON-2 tracks. The tracks numbered 094, 170, and 061 are the ones with maximum
618 surges for storms Felix, Friedhelm/Gunter, and ex-Gonzalo (see Table 1). The 4 groups of tide gauges correspond
619 to the 22 tide gauges used for this study. For ex-Gonzalo storm, group 1 matches quite well with observations,
620 group 2 shows an underestimation, group 3 is not sensitive to the parameterization (deep waters), and group 4 has
621 no data during this storm.

622

623 Figure 2: Tracks of the minimum of the Mean Sea Level Pressure for the selected storms (from ECMWF
624 simulations). The asterisk on the figure corresponds to the time for the wind plot (Figure 3).

625

626 Figure 3: 10 m winds during the storms (a) Friedhelm, 8 December 2011, 22:00 (b) ex-Gonzalo, 21 October 2014,
627 17:00 (c) Felix, 10 January 2015, 10:00 (d) Gunter, 12 January 2015, 17:00 (from ECMWF simulations).

628

629 Figure 4: Ocean model forced with a coupled wave-atmosphere model.

630

631 Figure 5: FES2014 grid over the North East Atlantic.

632

633 Figure 6: Comparison of drag coefficient for the TUGO parameterization (Hellerman and Rosenstein, 1983) and
634 the ECMWF (CY41R1) parameterization. For the ECMWF model, computation is made over the North Sea over
635 two days.

636

637 Figure 7: Comparison of drag coefficient for the TUGO parameterization (Hellerman and Rosenstein, 1983) and
638 the ECMWF (CY41R1) parameterization. Vertical bars correspond to one standard deviation. For the ECMWF
639 model, computation is made over the North Sea over two days.

640

641 Figure 8: Schematic representation of momentum fluxes at the air-sea interface (adapted from Janssen et al.
642 (2013)). The momentum flux going into the ocean τ_{oc} is the sum of τ_o and τ_{diss} .

643

644 Figure 9: Observed and modelled surges with two parameterizations (wind- and wave-dependent) at tide gauges
645 Lowestoft, EuroplatformTG, and Whitby during (a) ex-Gonzalo, (b) Felix, and (c) Gunter storms.

646

647 Figure 10: Surges from TUGO model, JASON-2 altimeter, and tide gauges during ex-Gonzalo (young sea) (a)
648 over the North Sea (b) along Jason-2 track. The grey shaded area corresponds to deep waters (>100 m), where
649 wind stress effect is lower.

650

651 Figure 11: Surges from TUGO model and JASON-2 altimeter during Friedhelm (old sea) (a) over the North Sea
652 (b) along Jason-2 track. The grey shaded area corresponds to deep waters (>100 m), where wind stress effect is
653 lower.

654

655 Figure 12: (a) Wind, (b) Significant Wave Height (SWH), (c) Wave age, (d) Charnock parameter, (e) wind stress
656 from the ECMWF coupled wave-atmosphere model, and (f) wind stress from TUGO ocean model, during ex-
657 Gonzalo storm (young sea), October 21 2014, 23:00. The colored dots on (b) correspond to the observed
658 Significant Wave Height from JASON-2 altimeter.

659

660 Figure 13: Significant Wave Height from JASON-2 altimeter and the ECMWF model during (a) Friedhlem, (b)
661 ex-Gonzalo, (c) Felix, and (d) Gunter storms.

662

663 Figure 14: Wind roses (m/s) in the middle of the North Sea (4°E 56°N) during the 5-day simulations of storms (a)
664 Friedhelm (b) ex-Gonzalo (c) Felix/Gunter. The color corresponds to the wind speed (in m/s), and the bar size
665 corresponds to the distribution (in %).

666

667 Figure 15: Comparison of default SLA (blue curve) and reconstructed SLA (red curve) taking into account the
668 waves for track 061 during ex-Gonzalo.

669

Name	Track date	Track number	Type of sea state	Rank from JASON-2 analysis	Rank from TGs analysis	Max. wind (m/s)
Friedhelm	2011-12-10 02:43	170	old sea	5	/	27.0
ex-Gonzalo	2014-10-21 22:39	061	young sea	2	1	22.7
Felix	2015-01-10 12:44	94	young sea	6	5	26.8
Gunter	2015-01-13 11:55	170	old sea	20	/	25.5

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Table 1: Storms selected for this study. The type of sea state (column 4) indicates if the sea state is young or old along the track. The rank (columns 5 and 6) corresponds to maximum surges. Maximum winds (column 7) are computed over the North Sea (4°E 10°W 50°N 65°N) during two days from the ECMWF simulations.

Tide Gauge	Bias (m) Wind-dep.	Bias (m) Wave-dep.	RMSE (m) Wind-dep.	RMSE (m) Wave-dep.	Peak Err. (m) Wind-dep.	Peak Err. (m) Wave-dep.
EX-GONZALO						
Group 1						
D151TG	0.007	0.002	0.084	0.070	-0.124	-0.024
EuroplatformTG	0.016	0.024	0.130	0.120	-0.165	0.061
F3platformTG	0.007	0.000	0.084	0.061	-0.129	-0.005
Harwich	-0.068	-0.045	0.161	0.155	-0.169	0.106
IjmuidenTG	0.022	-0.001	0.192	0.162	-0.336	-0.175
Leith	-0.018	-0.012	0.119	0.114	-0.141	-0.080
Lowestoft	-0.029	-0.028	0.126	0.111	-0.307	-0.077
StavangerTG	0.018	-0.003	0.095	0.072	-0.103	-0.024
TregdeTG	0.011	-0.005	0.126	0.106	-0.248	-0.146
VlakteVdRaanTG	-0.012	-0.015	0.160	0.163	-0.161	0.047
Whitby	-0.005	-0.021	0.106	0.095	-0.234	-0.147
Group 2						
DunkerqueTG	-0.020	-0.041	0.189	0.181	-0.473	-0.319
HelgolandTG	0.083	0.072	0.202	0.173	-0.433	-0.207
HoernumTG	0.103	0.071	0.230	0.184	-0.437	-0.219
NorderneyTG	0.029	-0.000	0.261	0.218	-0.539	-0.328
VlielandHavenTG	0.014	-0.009	0.201	0.174	-0.374	-0.224
Group 3						
Aberdeen	0.026	0.011	0.084	0.074	-0.026	-0.003
Lerwick	0.010	0.015	0.038	0.038	-0.007	-0.004
MaloyTG	0.031	0.009	0.064	0.052	-0.065	-0.041
NorthCormorantTG	0.039	0.036	0.052	0.048	0.028	0.026
Wick	0.030	0.003	0.078	0.065	0.014	-0.026
Group 4						
Cromer	-	-	-	-	-	-
Mean	0.014	0.003	0.132	0.116	-0.211	-0.086
FELIX/GUNTER						
Mean	0.011	0.001	0.198	0.190	-0.138	-0.104

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Table 2: Bias, RMSE, and Peak Error for wind- and wave-dependent parameterization during the 5-day simulation for ex-Gonzalo and Felix/Gunter storms. For ex-Gonzalo, Group 1 matches quite well with observations (see the

681 last column for the Peak Error), Group 2 shows an underestimation (see also the last column for the Peak Error),
 682 Group 3 is not sensitive to the parameterization (deep waters), and Group 4 has no data during this storm. See
 683 Figure 1 for the location of the 4 groups.
 684

Process	Location	Computed in hydrodynamic models
Atmospheric surge due to wind and atm. pressure	Everywhere	Yes, if atmospheric forcing is introduced
Wave setup, i.e. surge due to wave breaking	In nearshore areas	Yes, if resolution is fine enough, and if radiation stress is introduced
Tide surge interaction	Significant in very shallow waters	Yes, if meteorological and tide forcing are introduced
Meteo-tsunami		Yes, if the space resolution is fine enough
Infragravity waves		Yes, if the hydrodynamic model is coupled with a wave model
Internal waves	Everywhere	Yes, in a baroclinic model
Rogue waves		No
Tsunamis i.e. surge due to an earthquake, landslide, or volcanic eruption		No
Seiches i.e. resonance phenomena in closed or semi-closed basins	Mainly in harbors or bays	Yes, if the spatial resolution is fine enough; this is rarely the case in global or regional models

685
 686 Table 3: Various processes contributing to the surge
 687
 688

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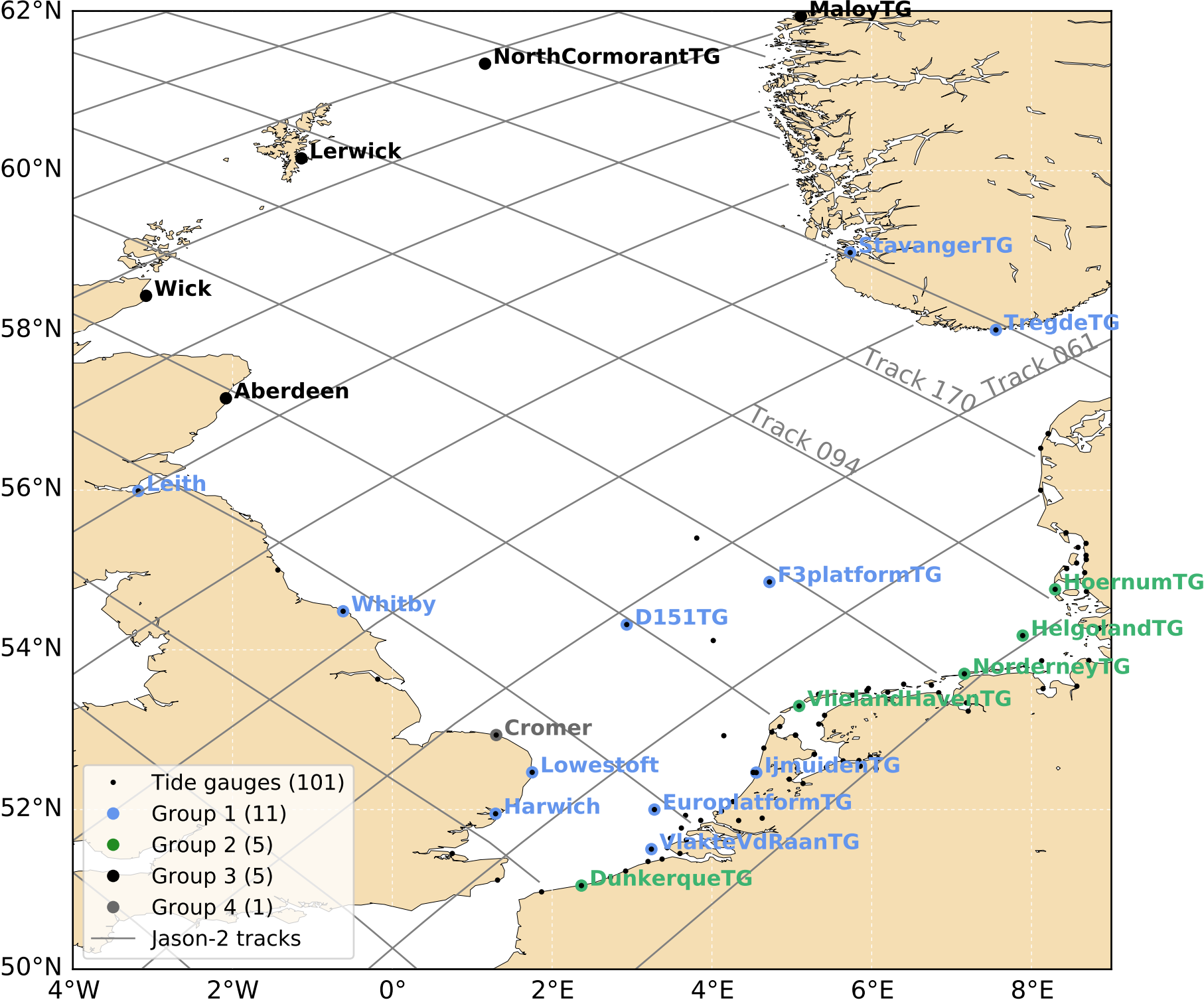
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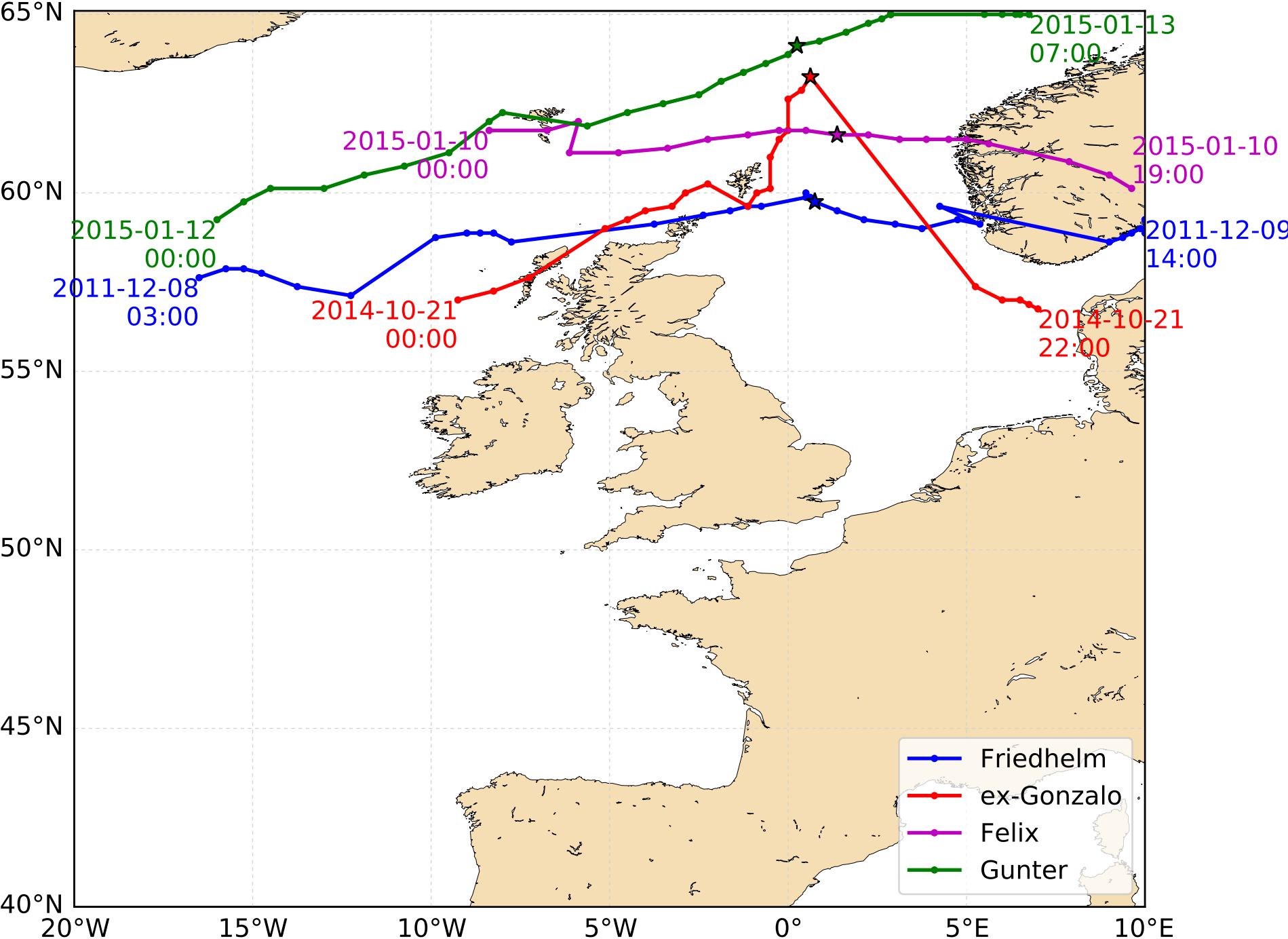
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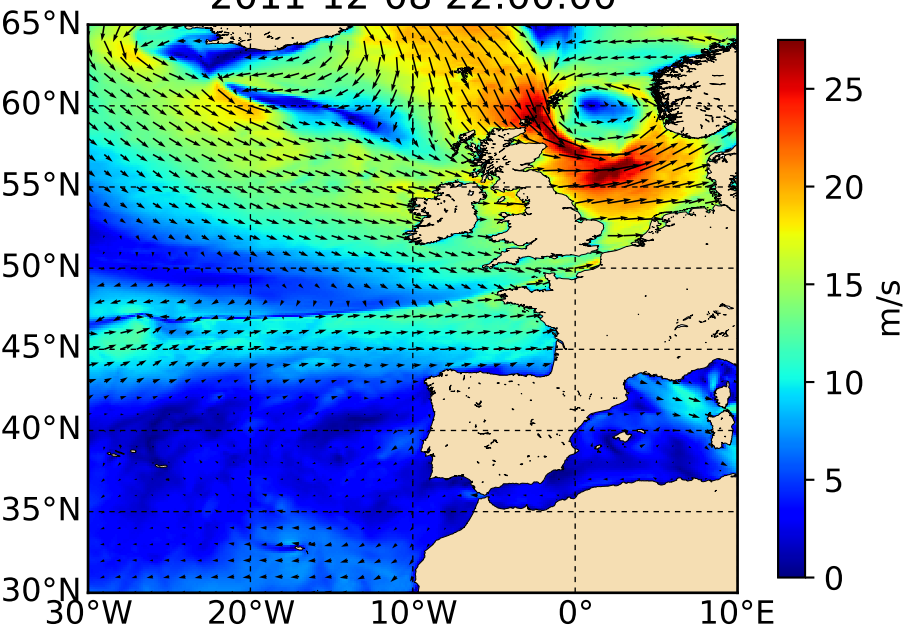
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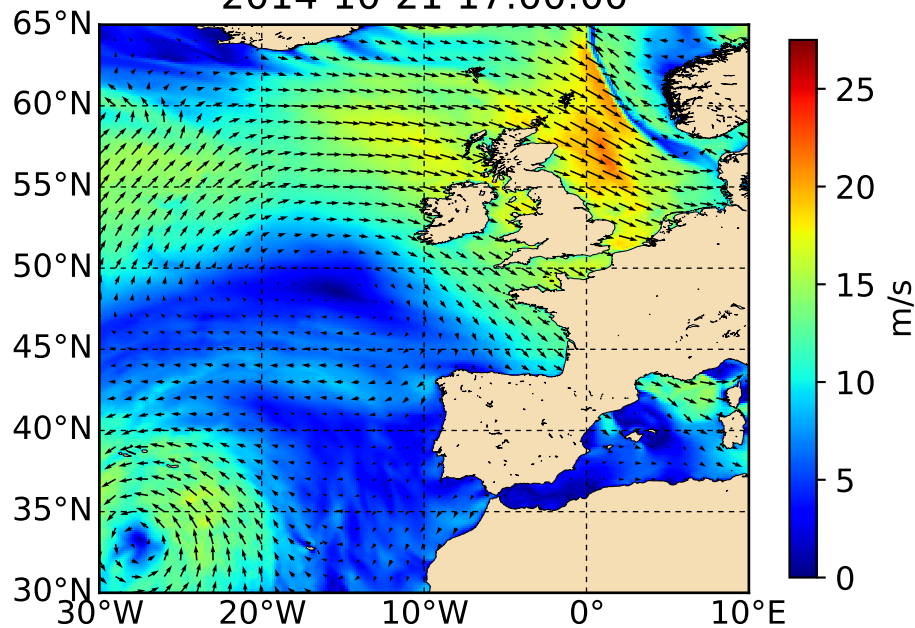




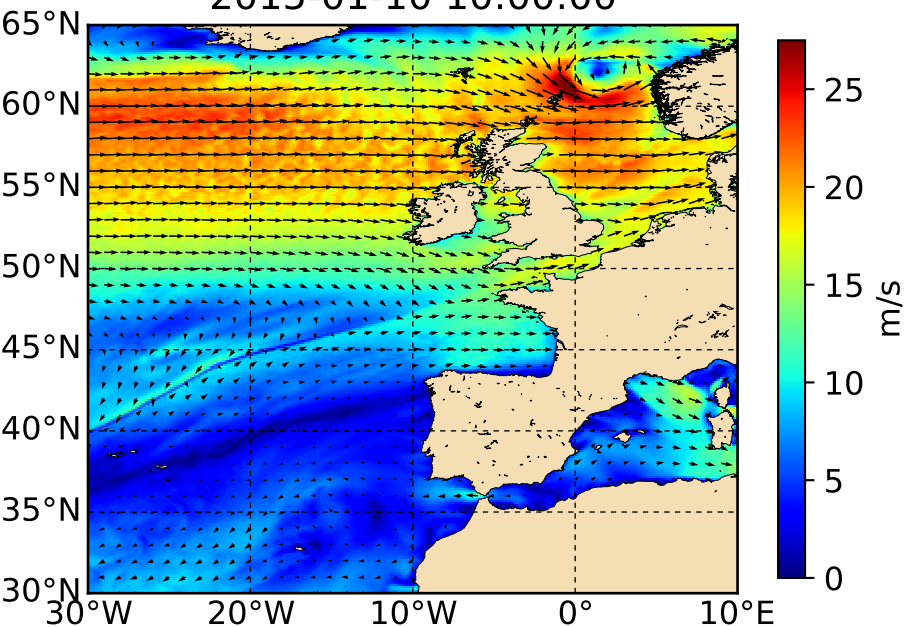
(a) Friedhelm
2011-12-08 22:00:00



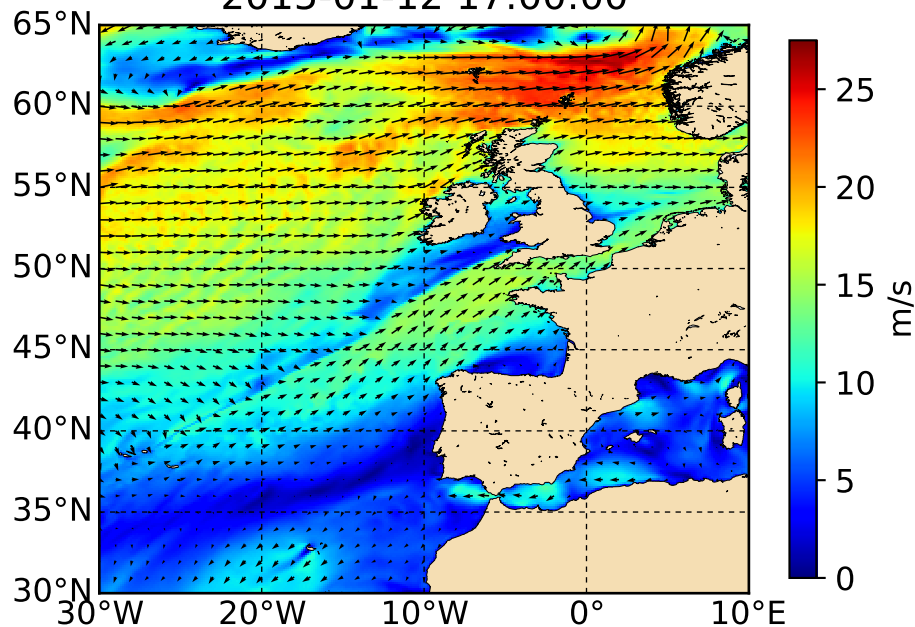
(b) ex-Gonzalo
2014-10-21 17:00:00



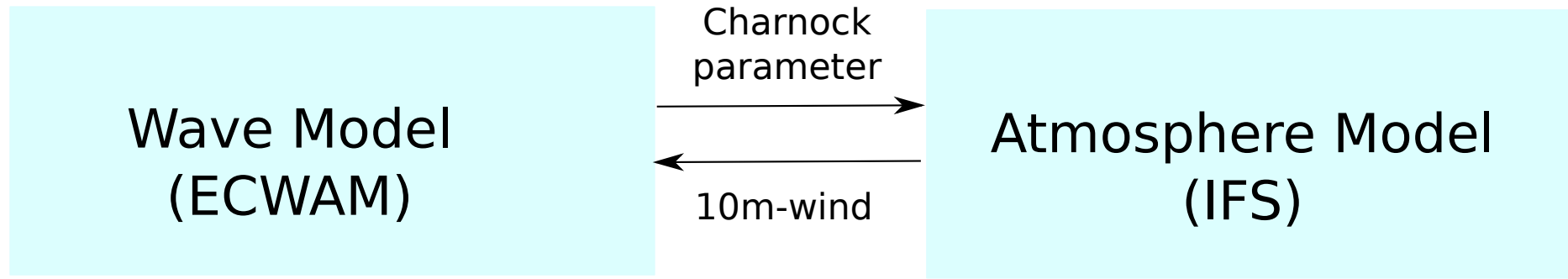
(c) Felix
2015-01-10 10:00:00



(d) Gunter
2015-01-12 17:00:00



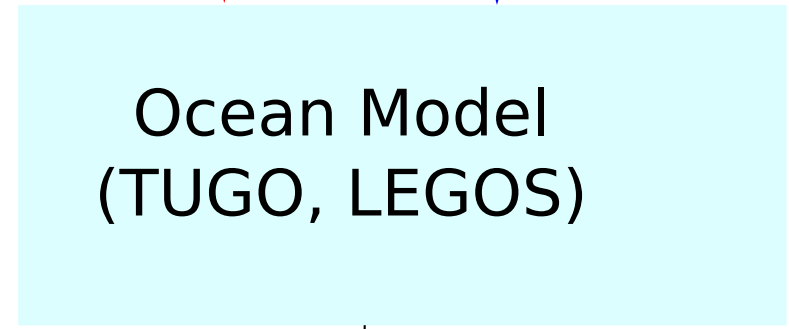
ECMWF Coupled Wave-Atmosphere Model (CY41R1)



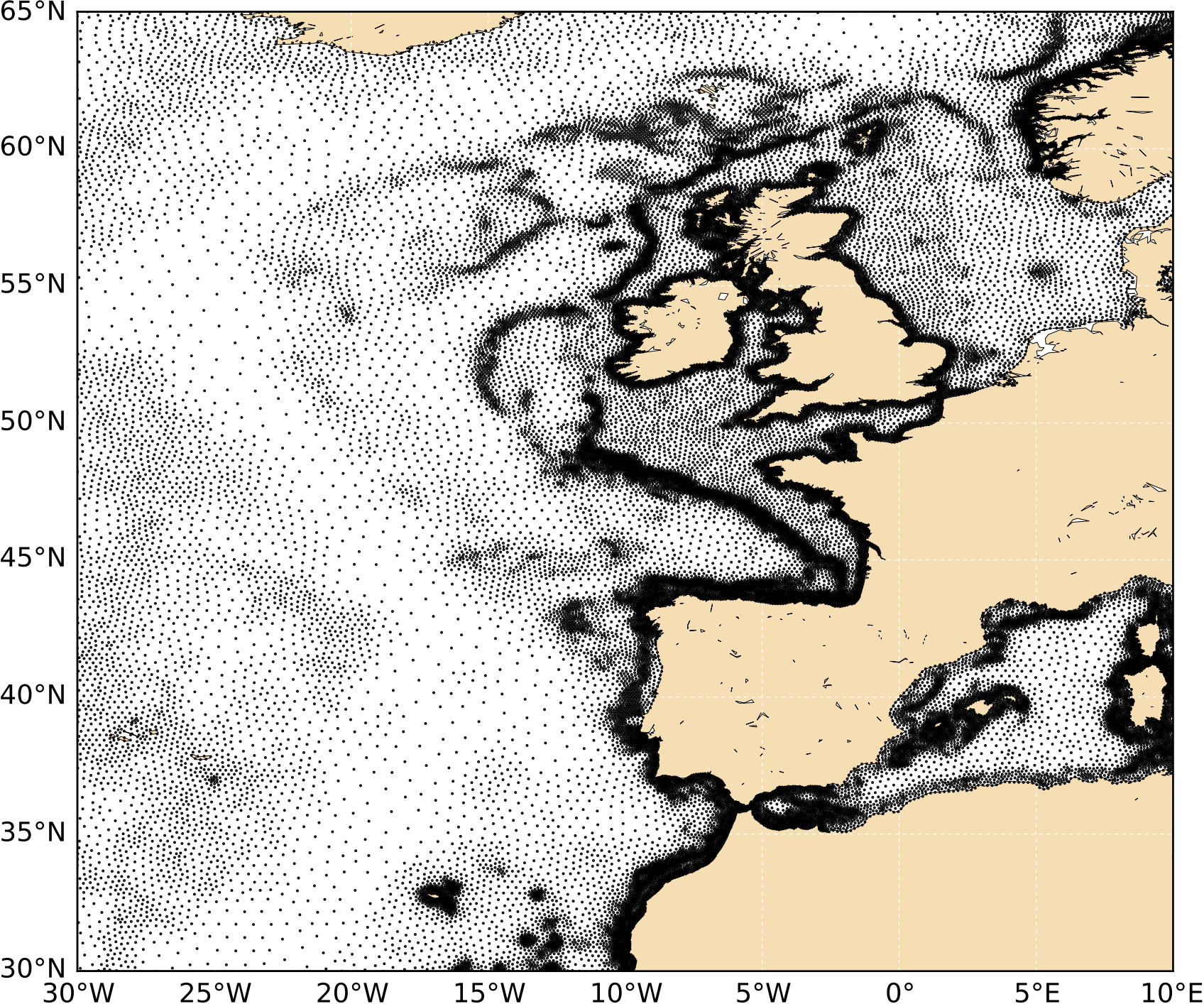
1 10m-wind

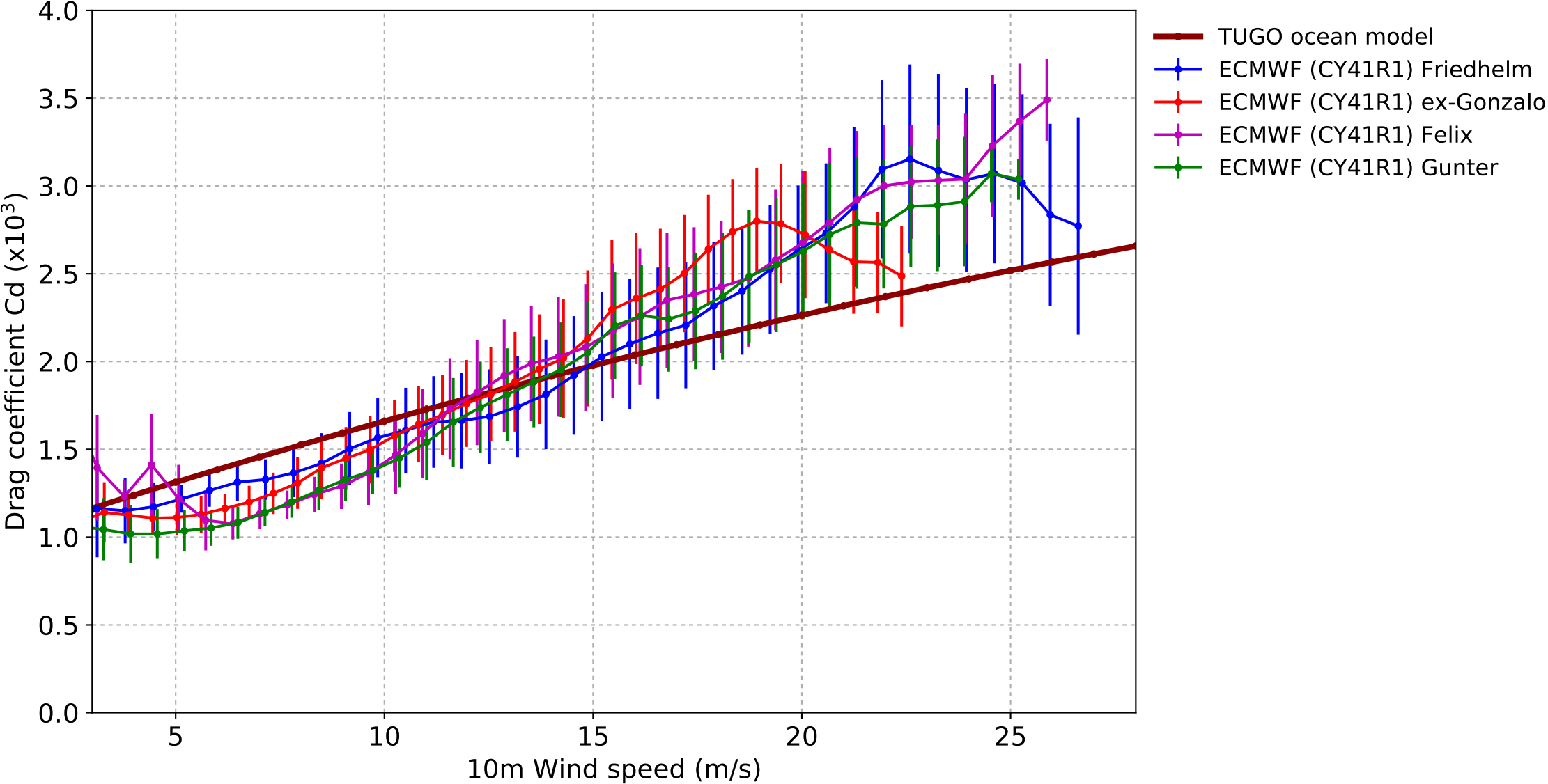
2 Wind stress from coupled wave-atmosphere model
WAVE-DEPENDENT DRAG

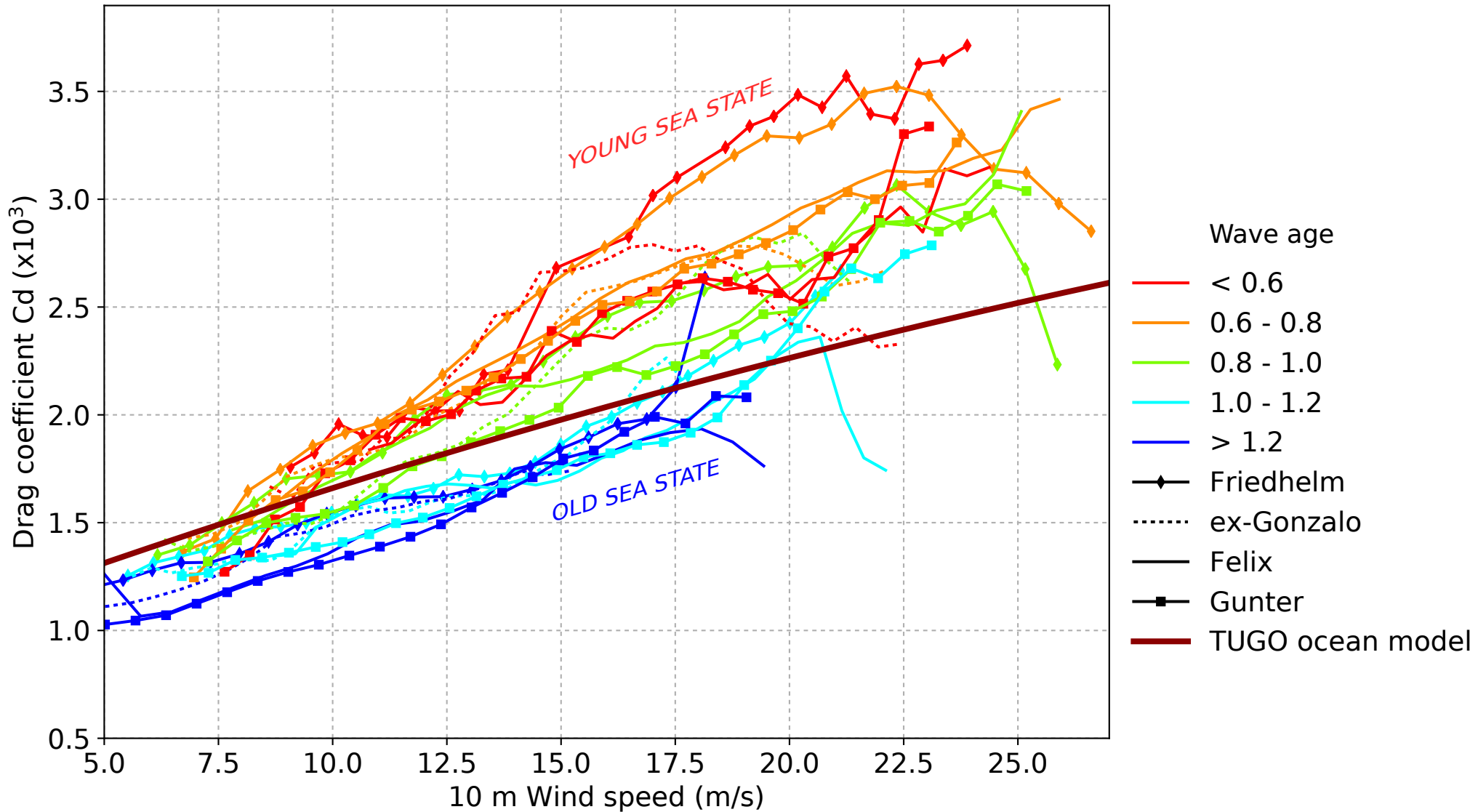
Bulk to compute the stress
(Hellerman and Rosenstein, 1983)
WIND-DEPENDENT DRAG

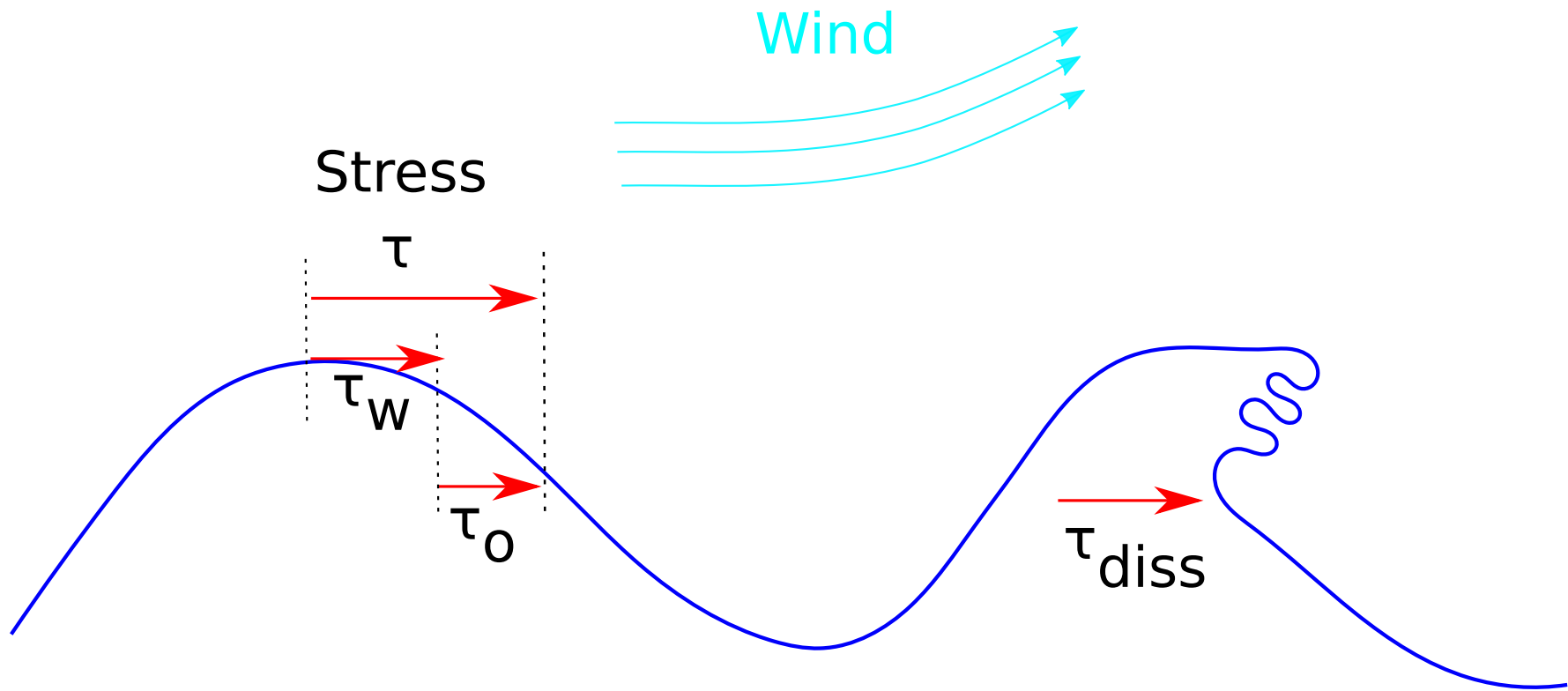


Sea levels, Surges, Currents

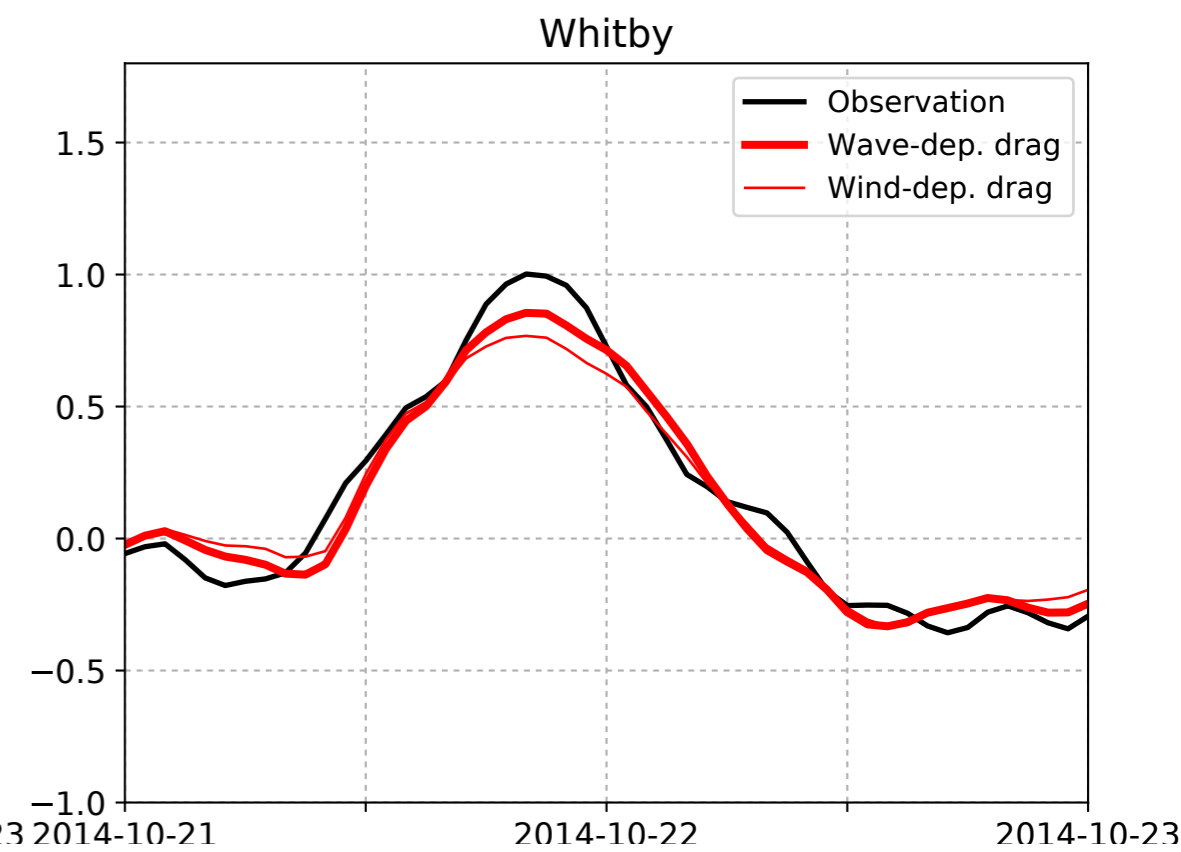
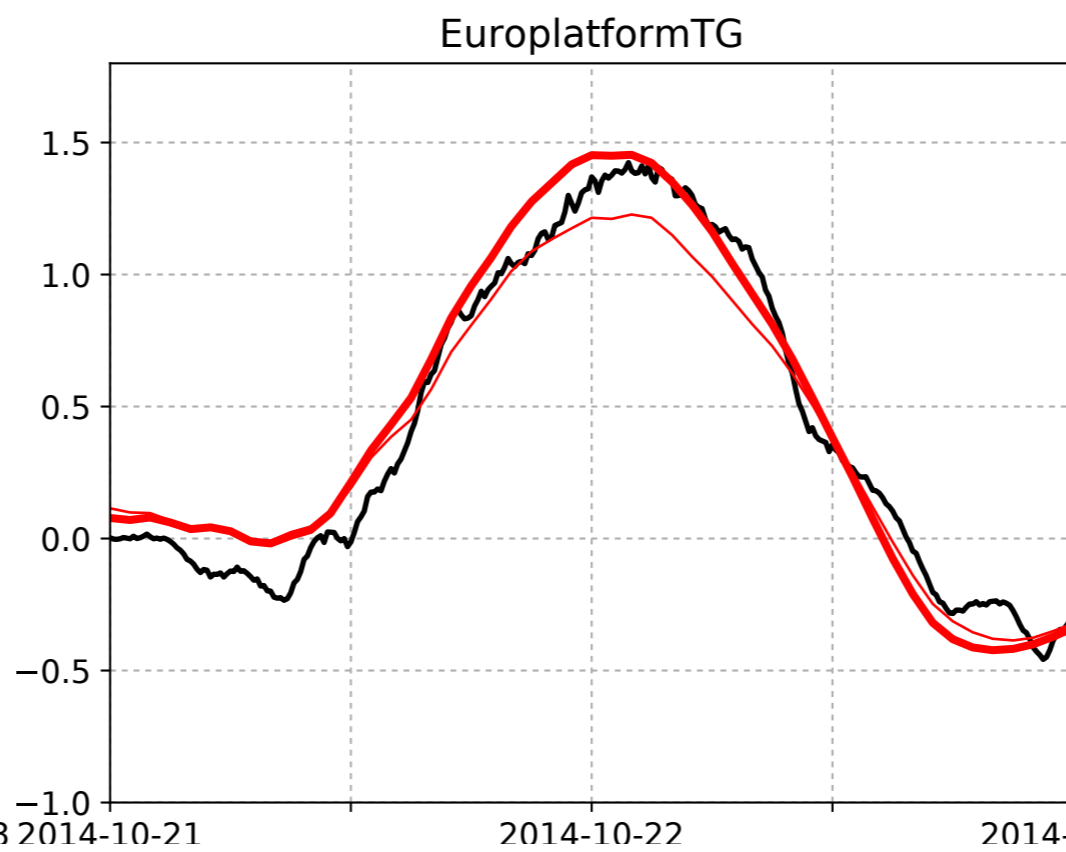
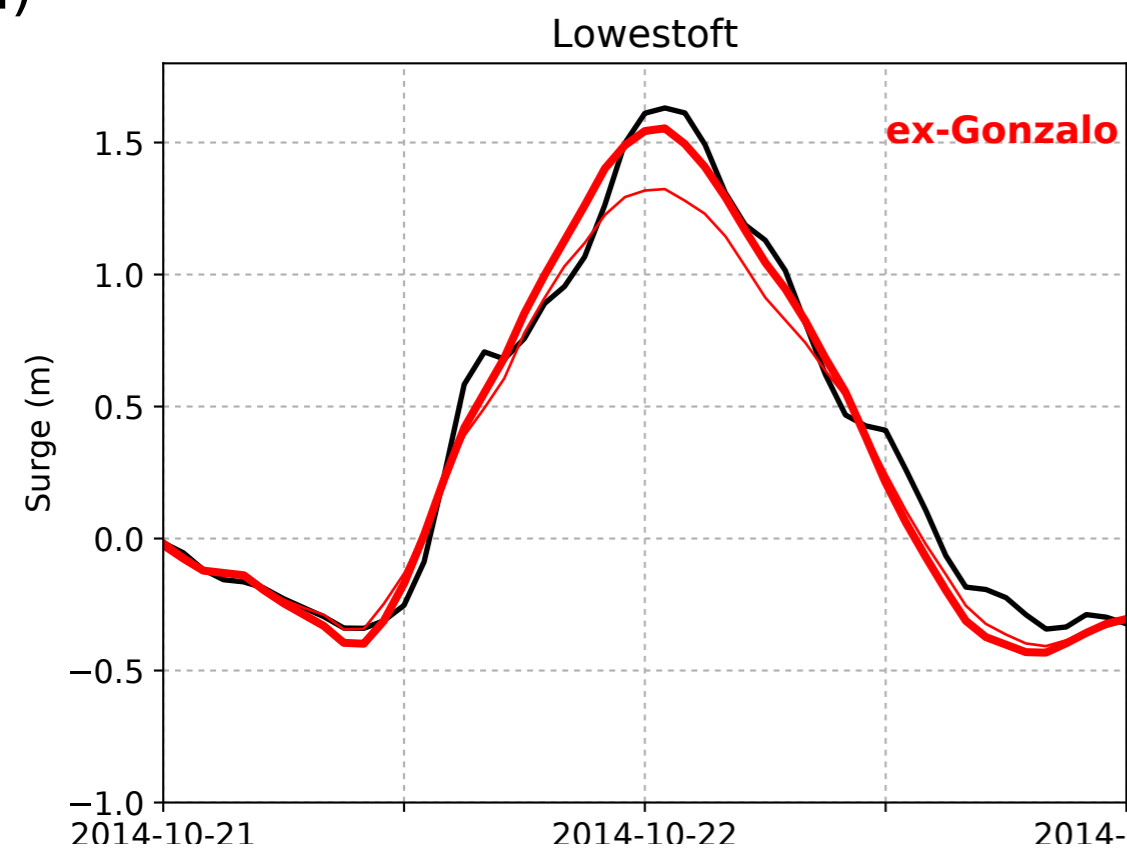




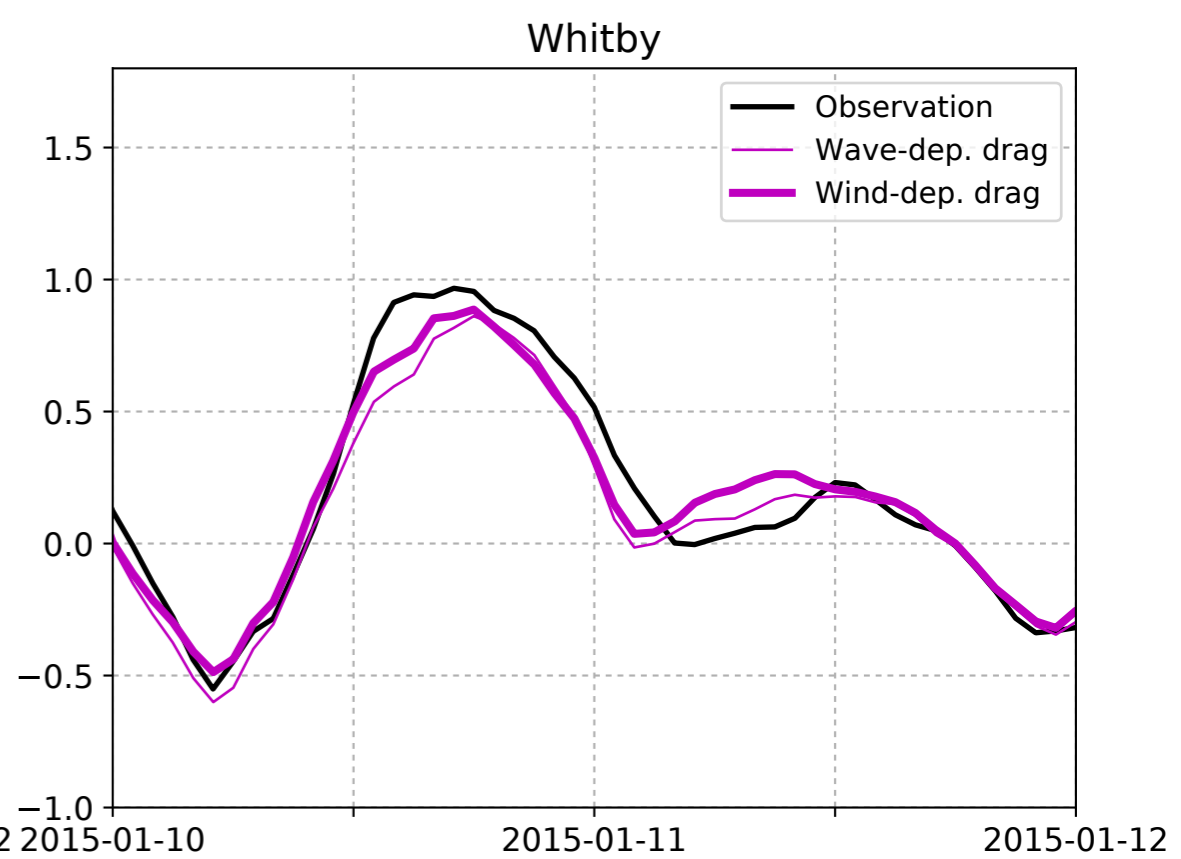
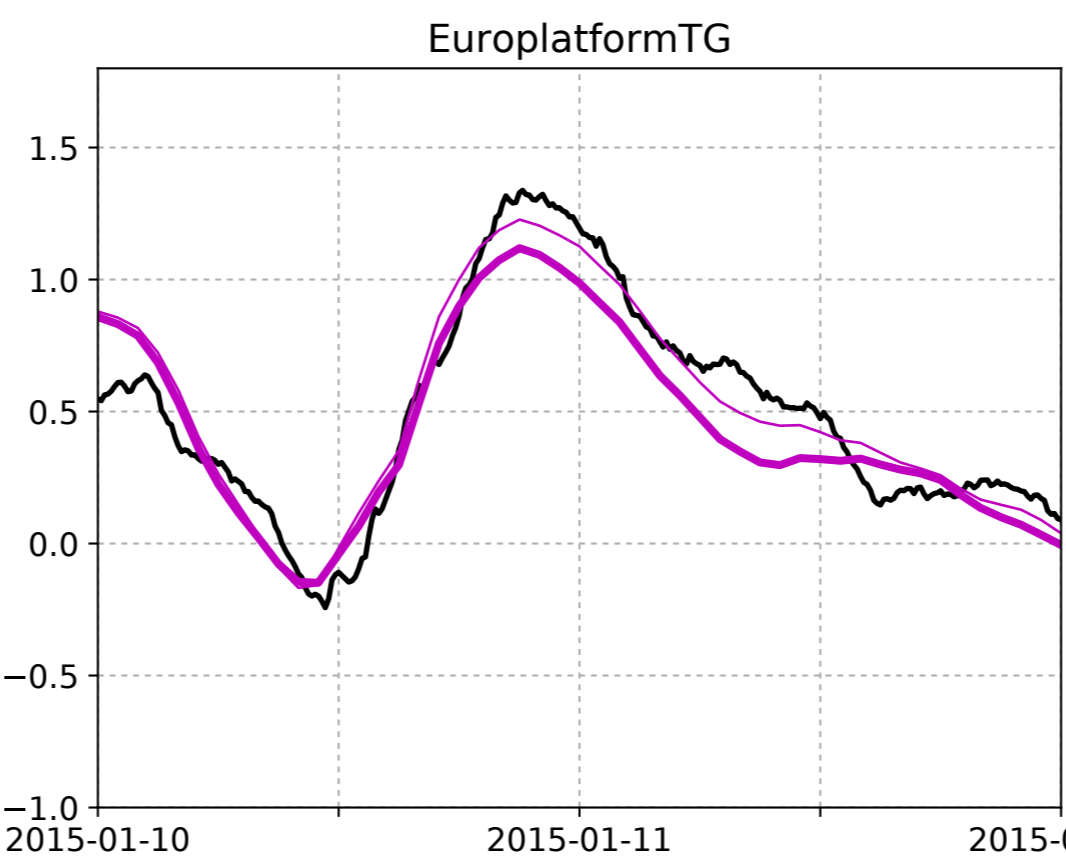
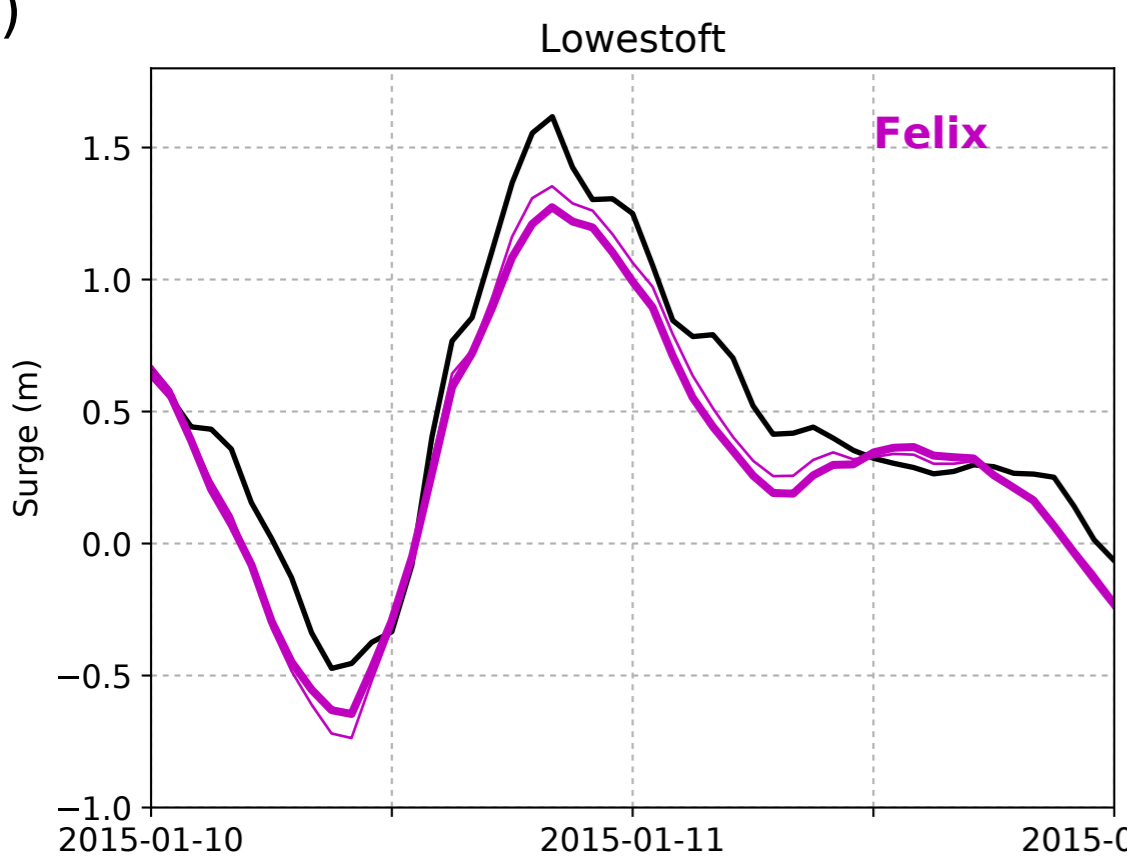




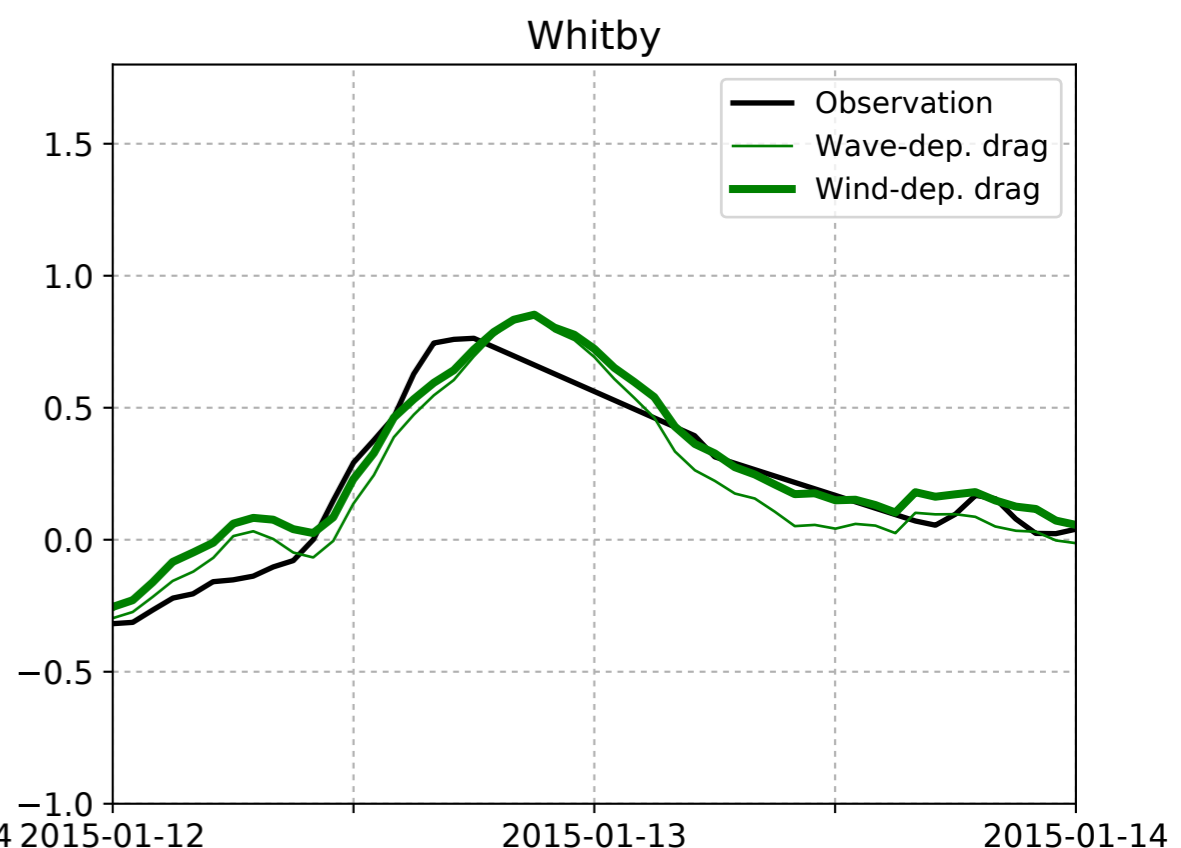
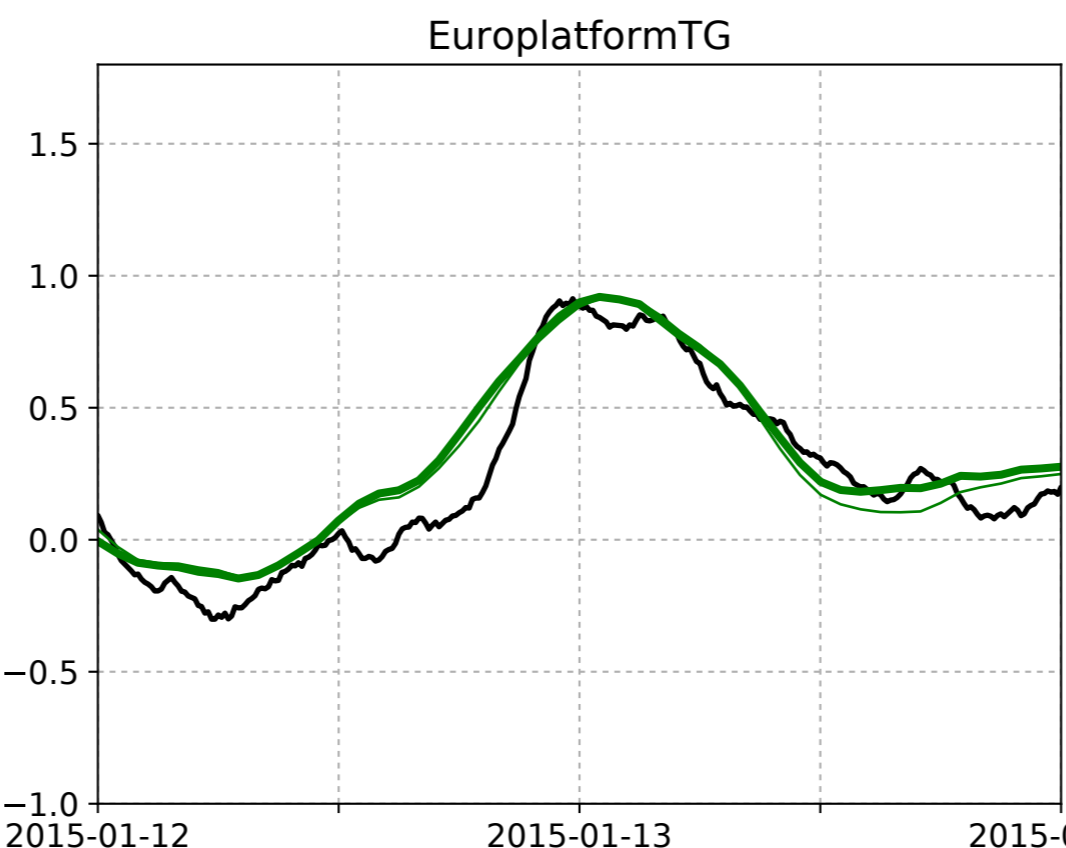
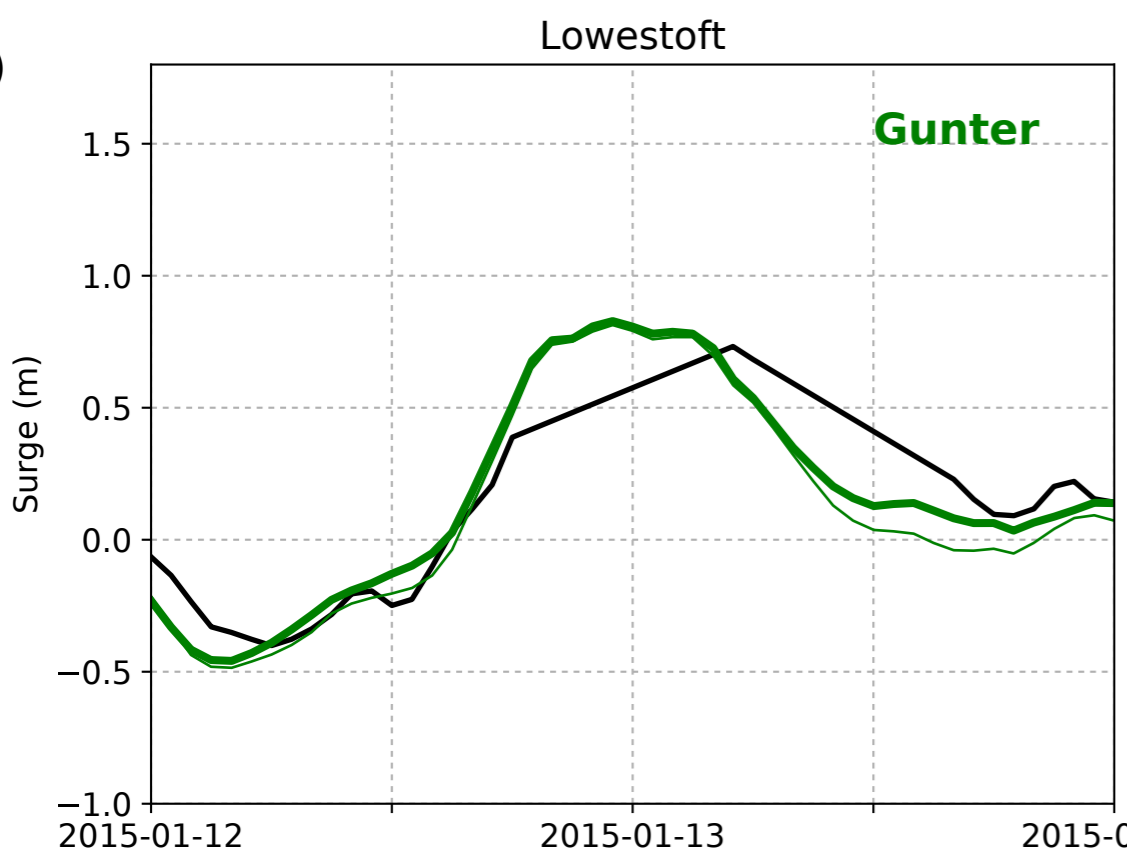
(a)



(b)

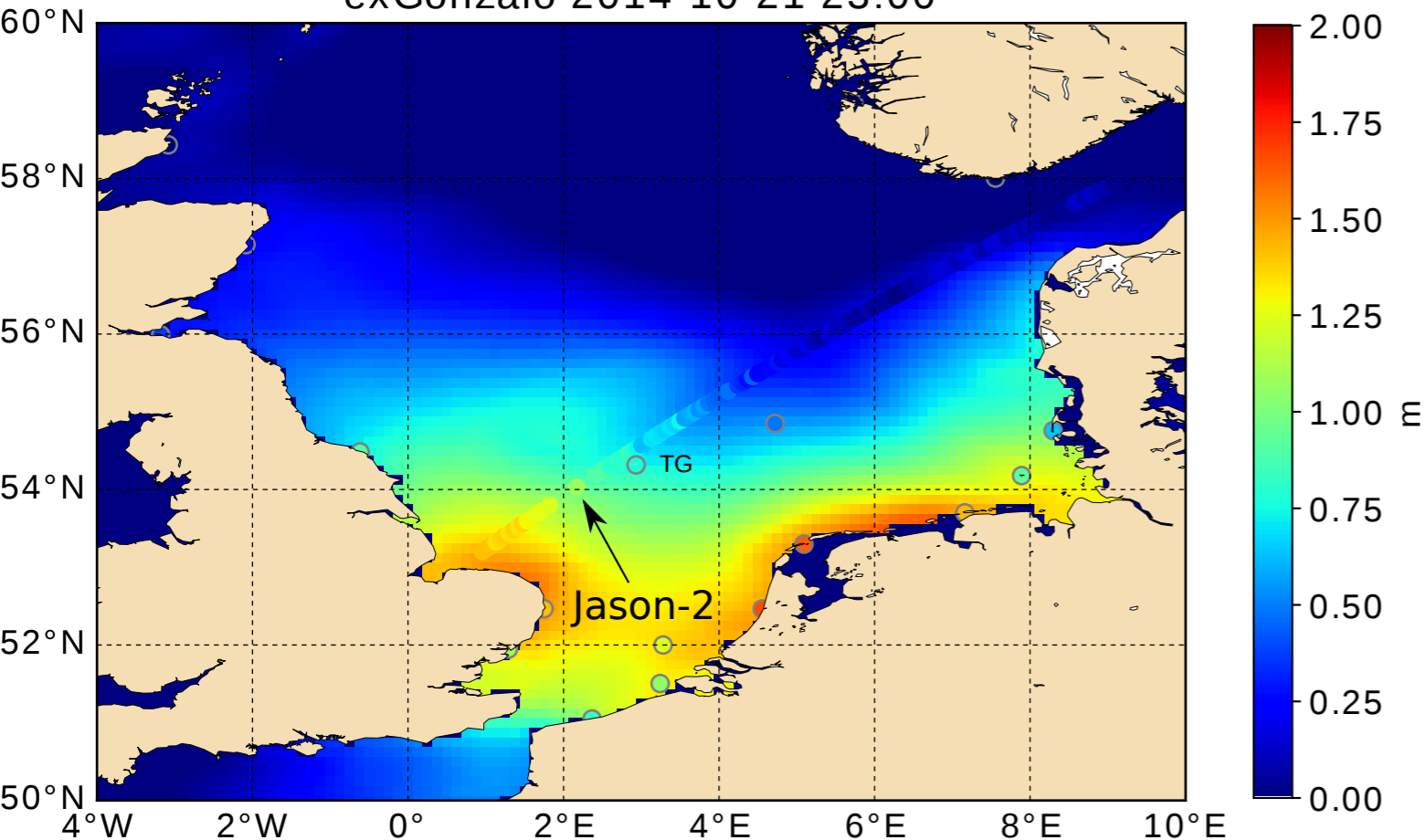


(c)



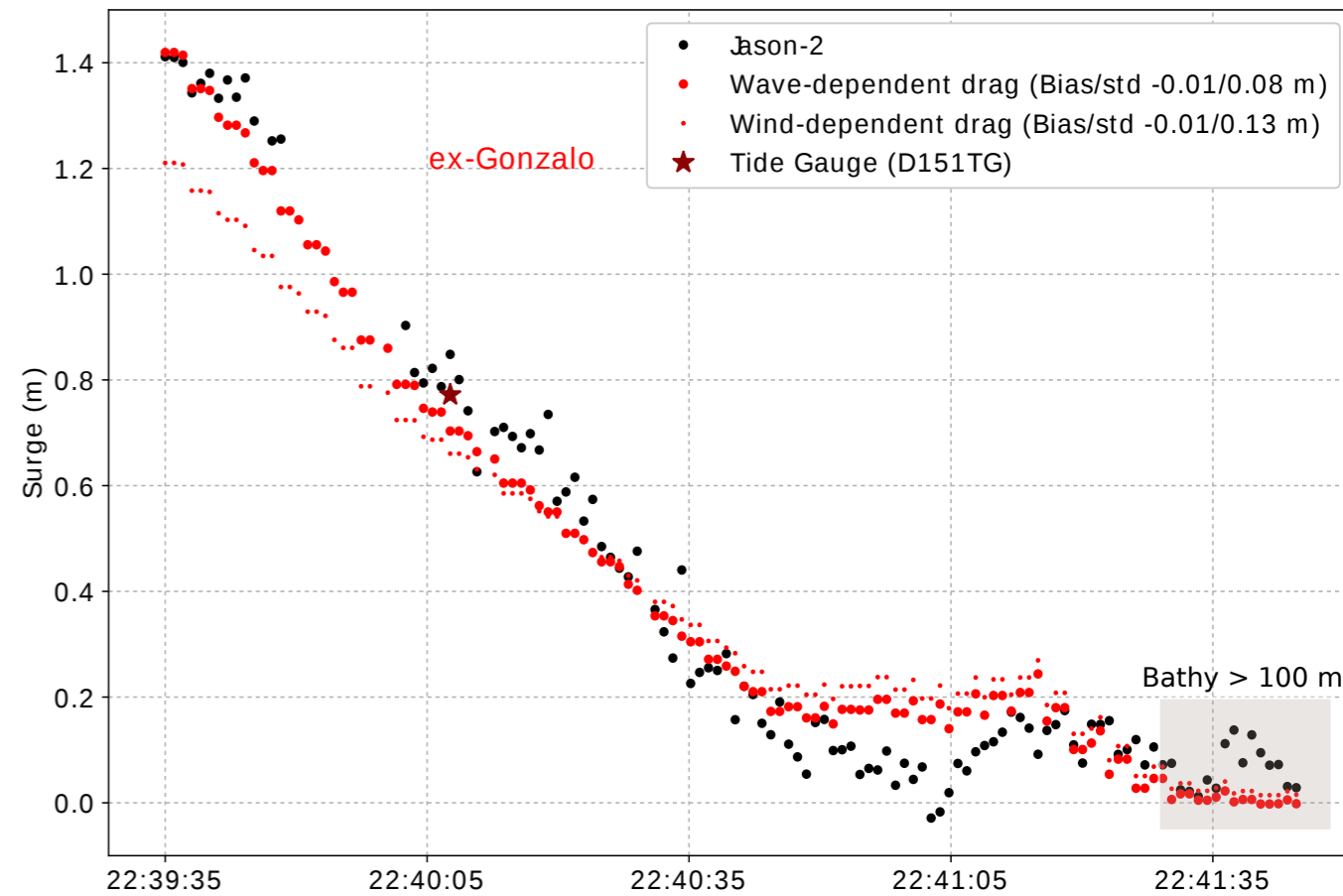
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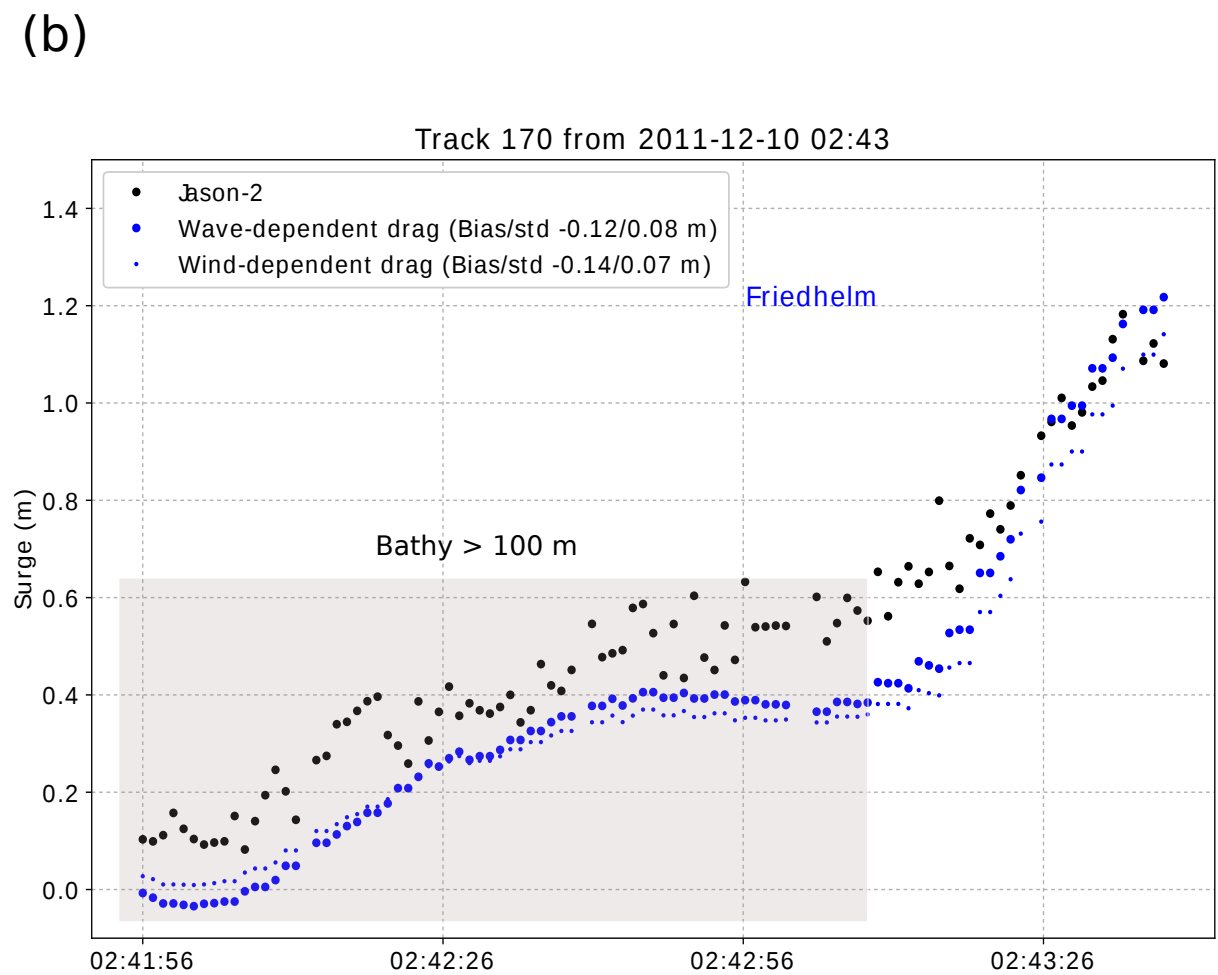
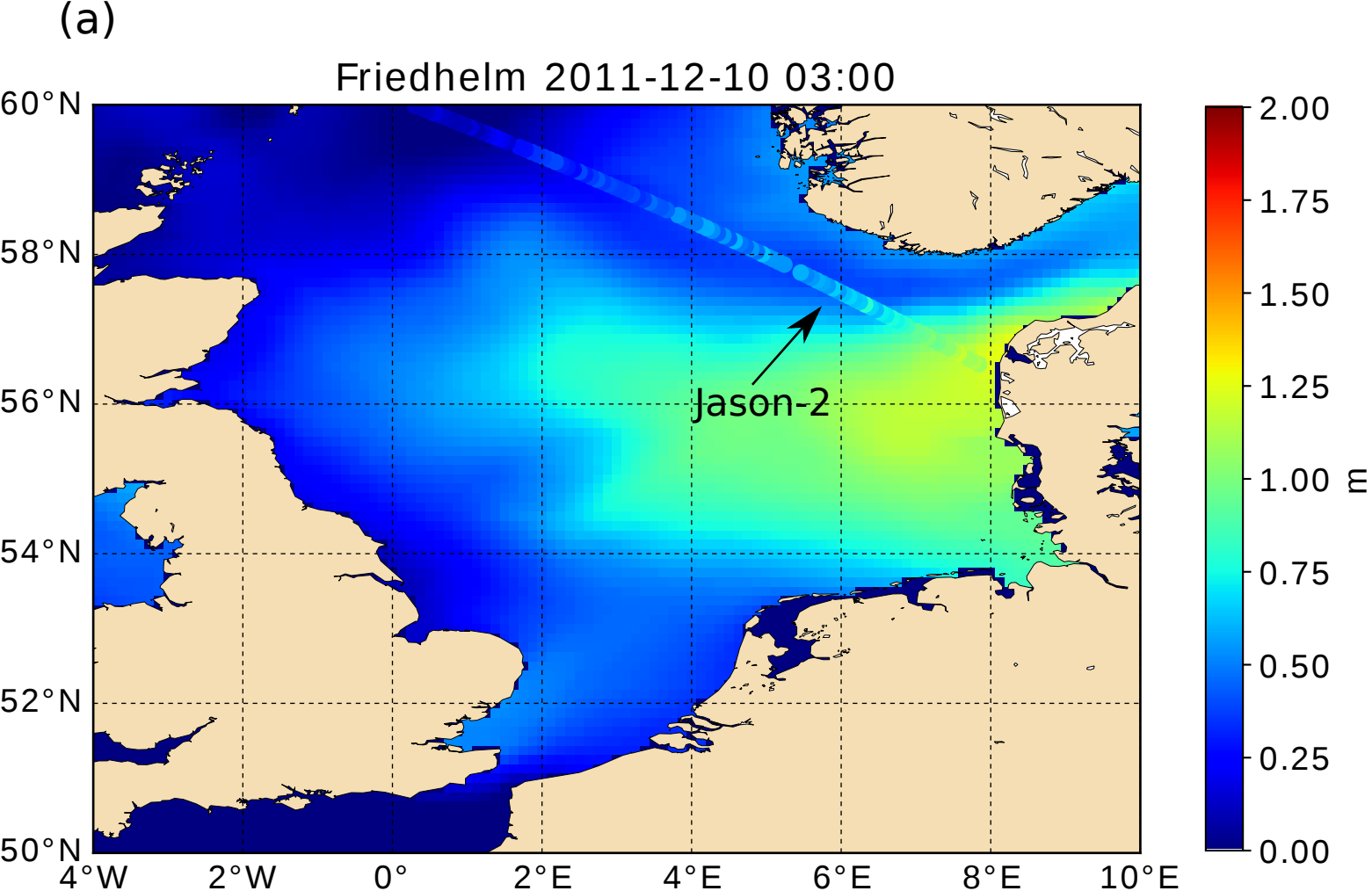
exGonzalo 2014-10-21 23:00

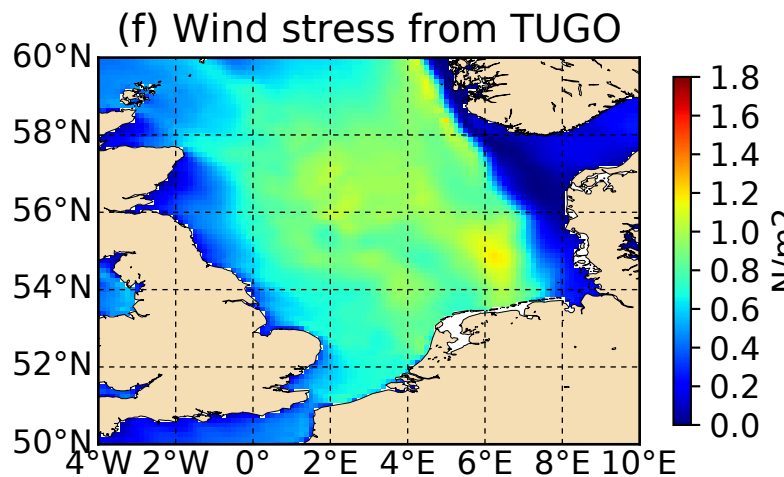
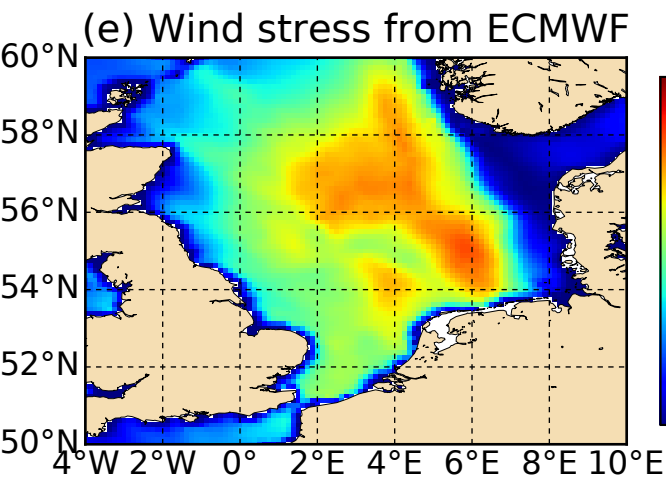
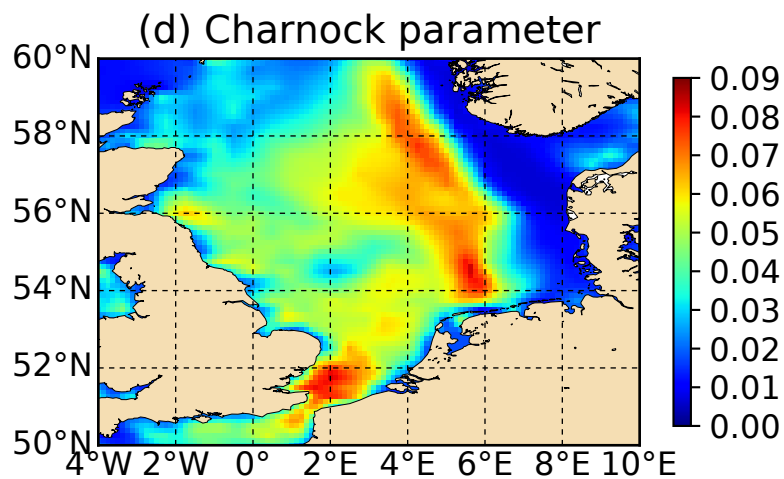
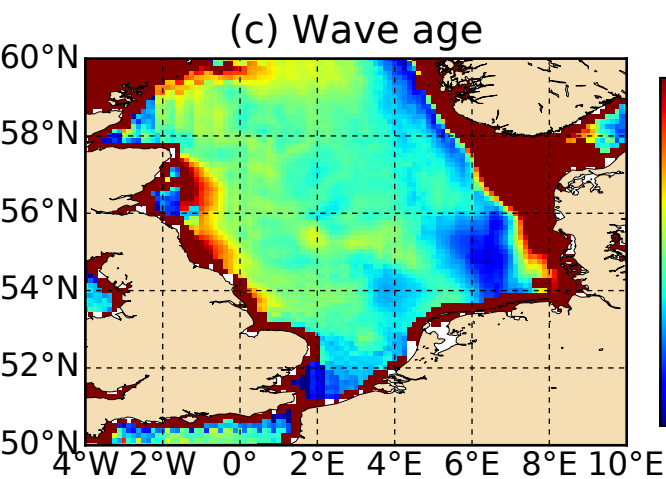
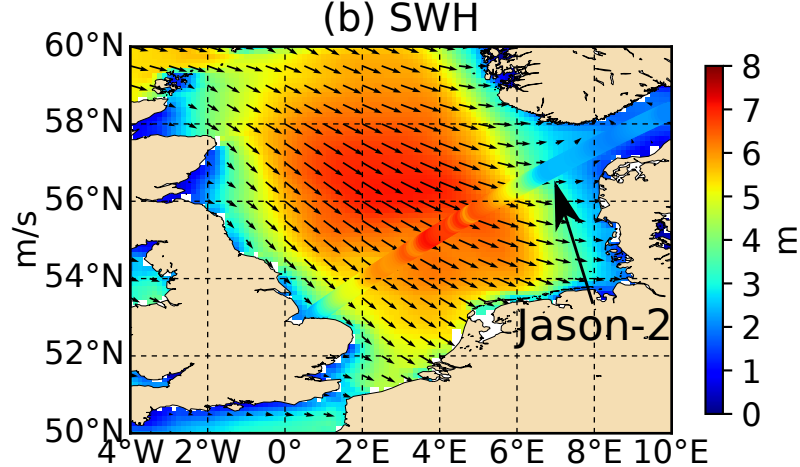
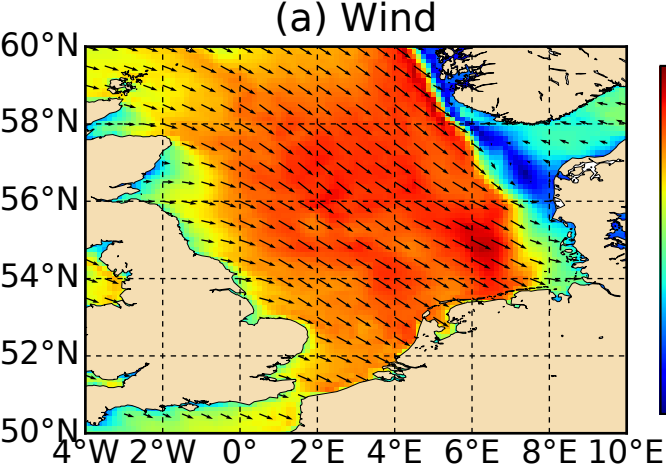


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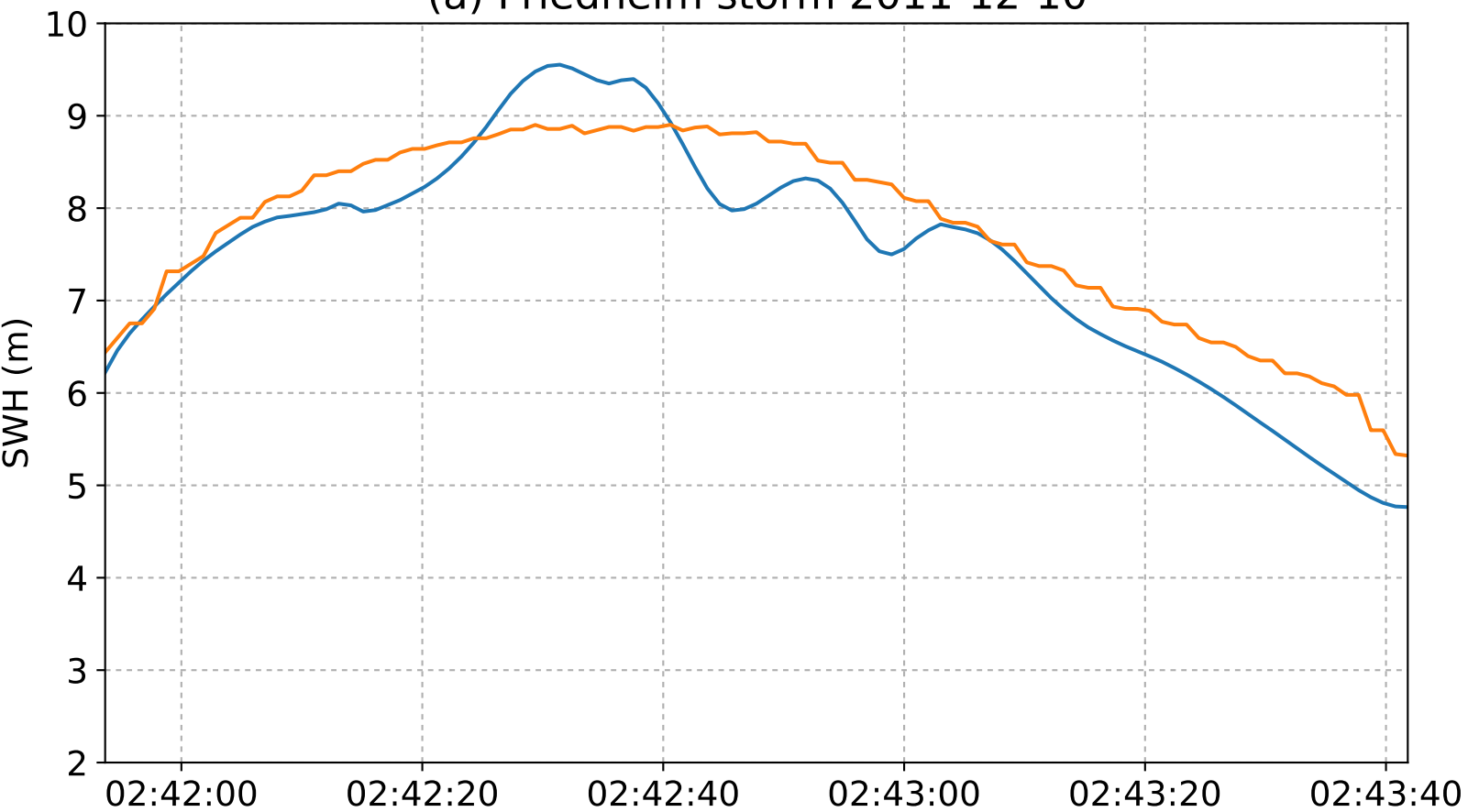
Track 061 from 2014-10-21 22:39



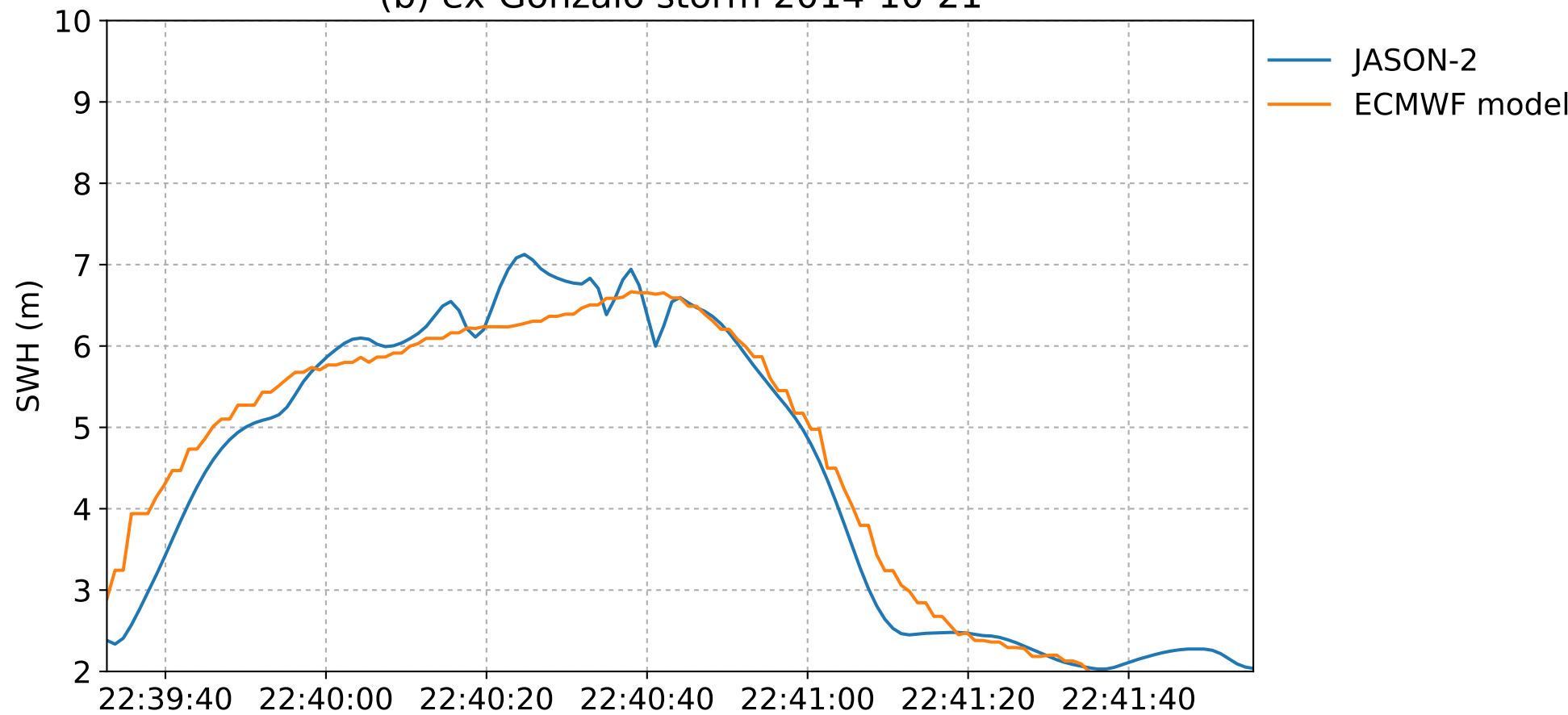




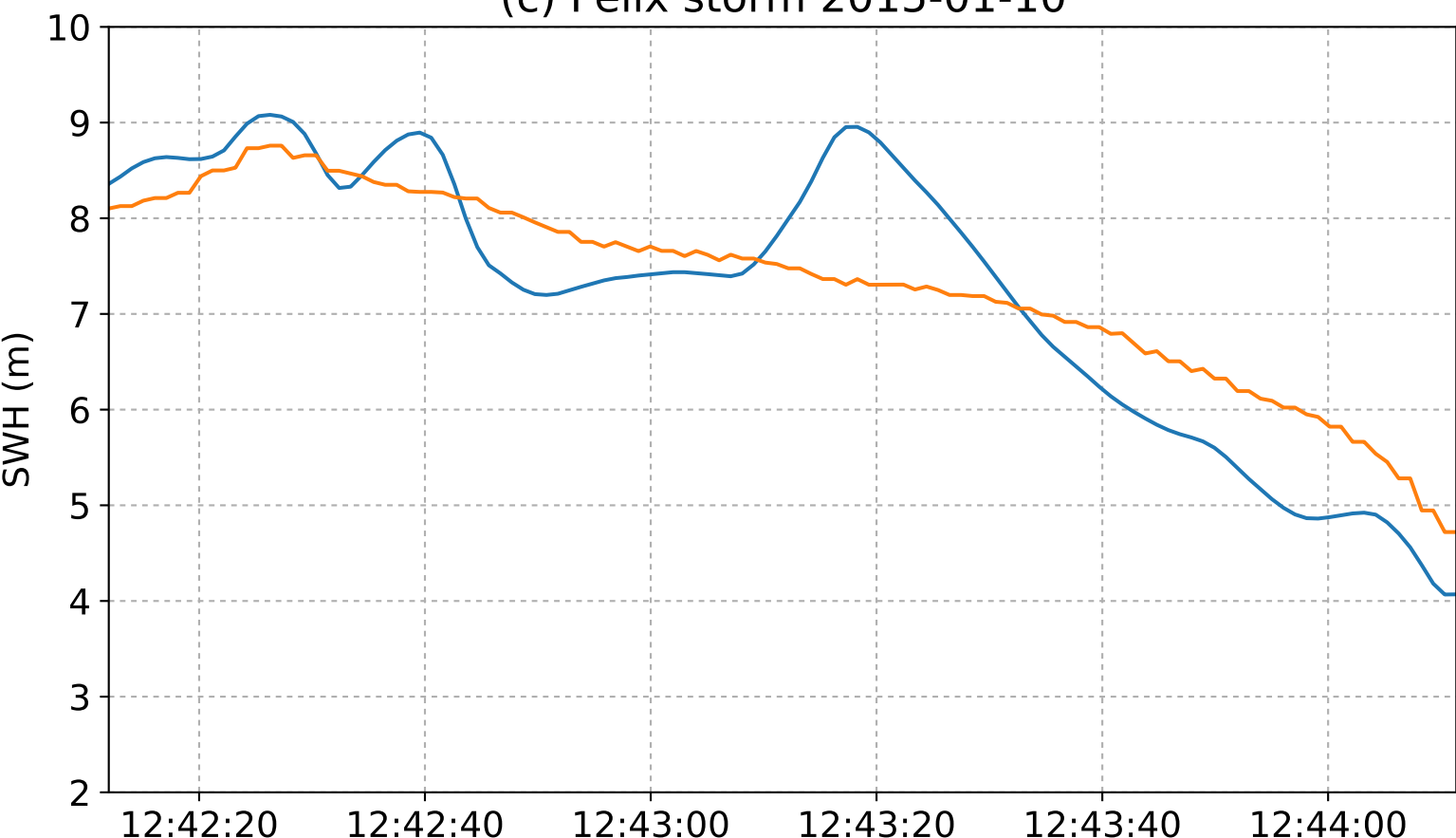
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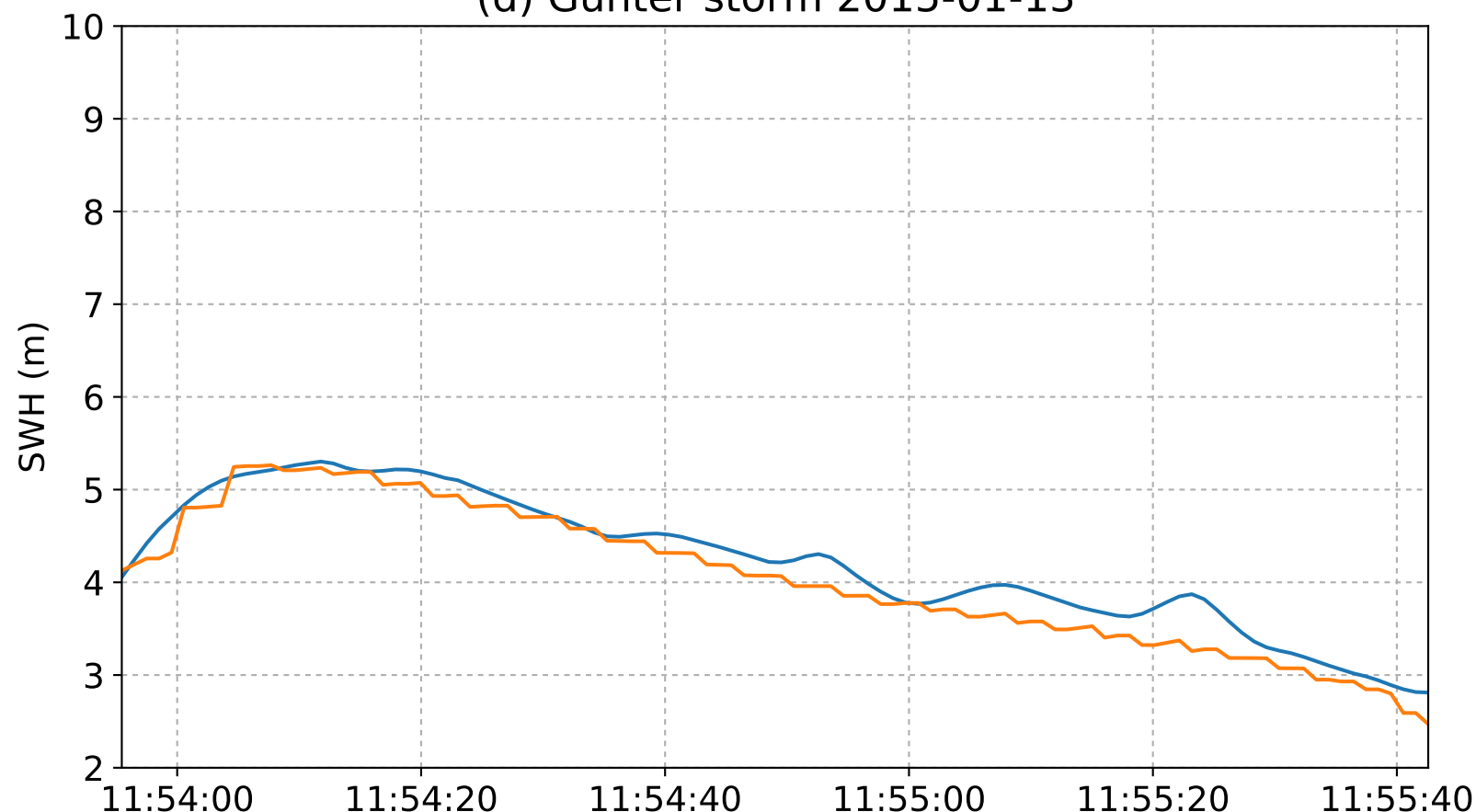
(b) ex-Gonzalo storm 2014-10-21



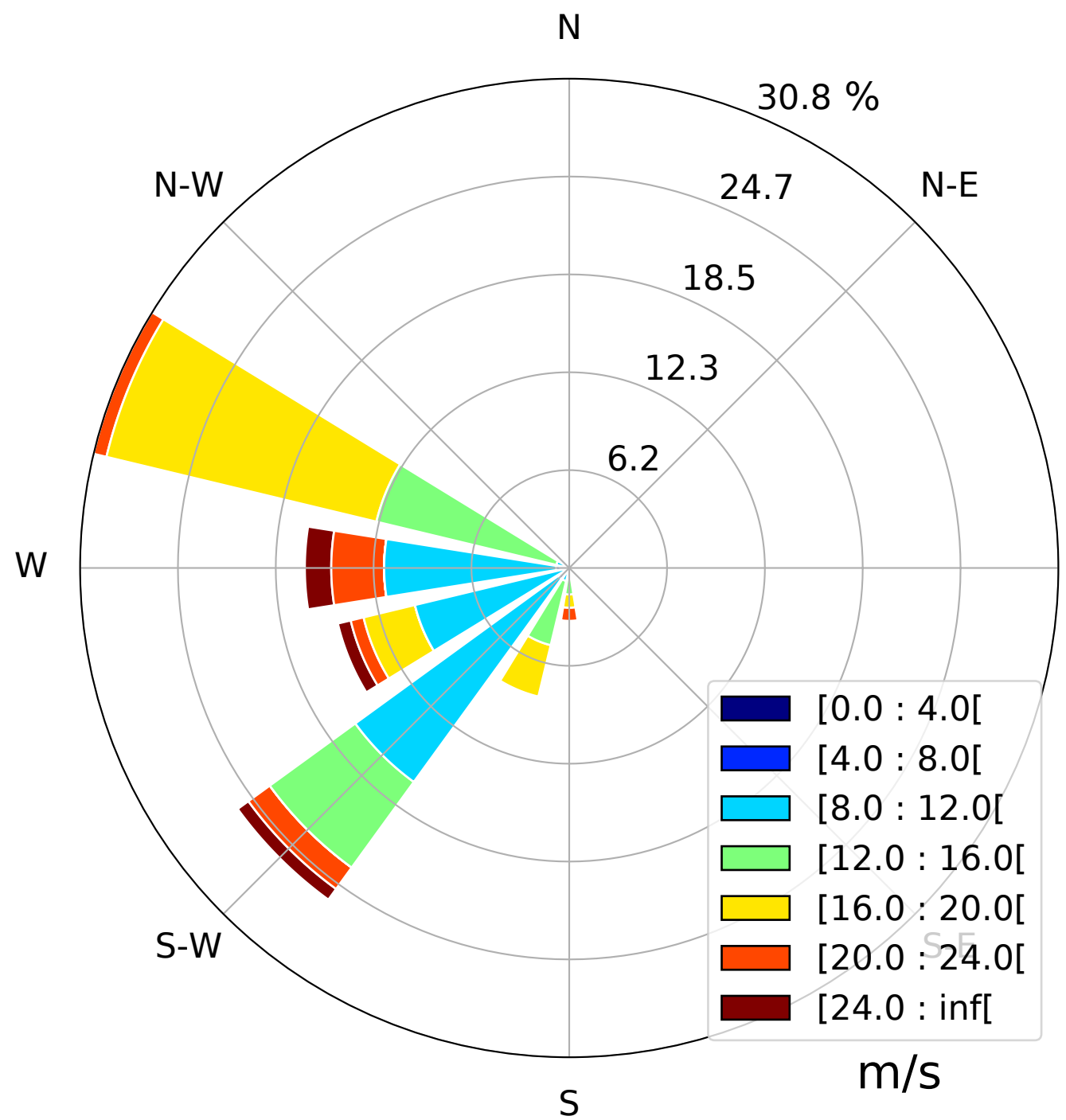
(c) Felix storm 2015-01-10



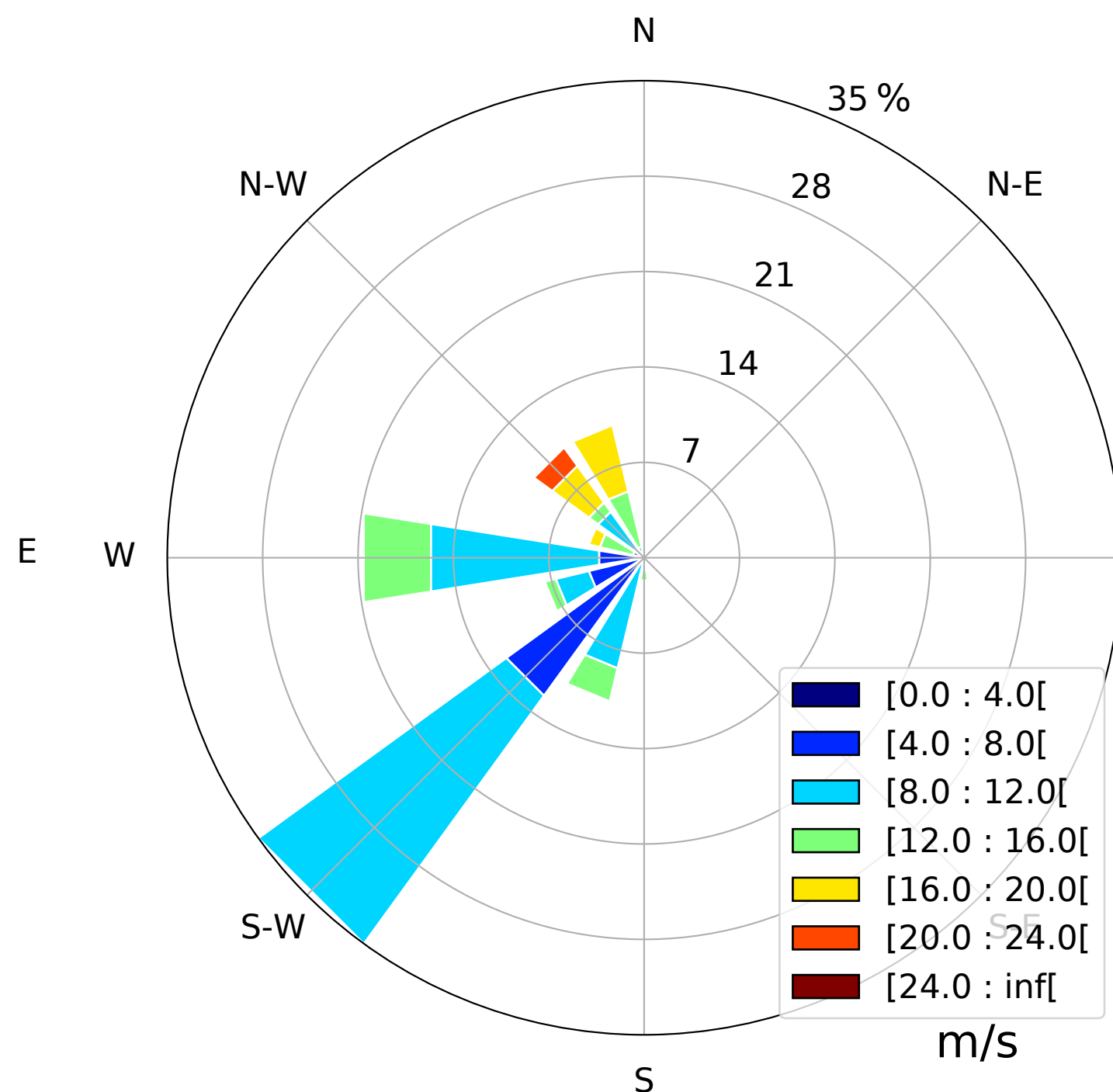
(d) Gunter storm 2015-01-13



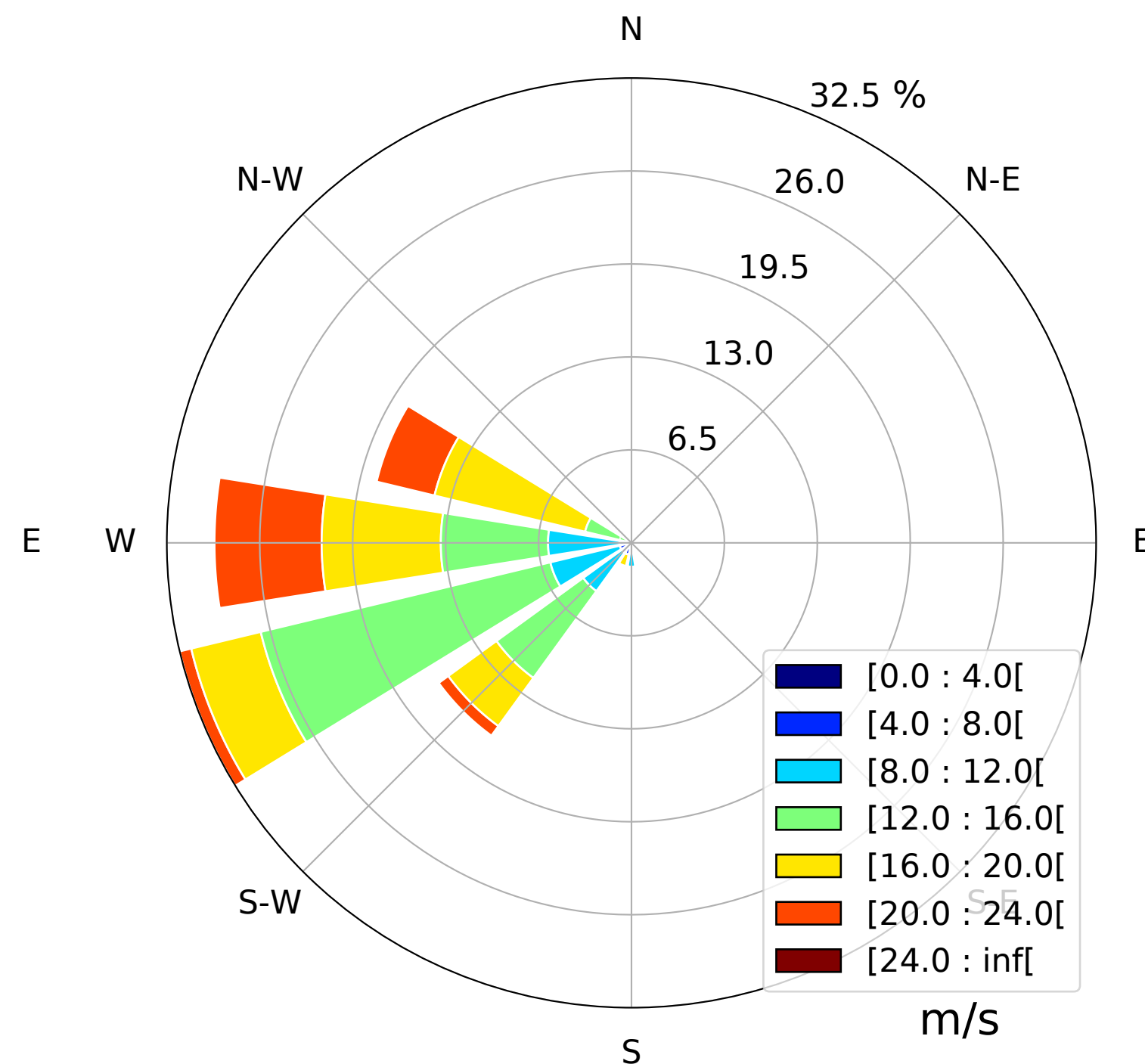
(a) Friedhelm



(b) ex-Gonzalo



(c) Felix/Gunter



Track 061 from 2014-10-24 21:51

